United States Department of the Interior
GEOLOGICAL SURVEY

REGIONAL AND OTHER GENERAL FACTORS BEARING ON EVALUATION OF EARTHQUAKE AND OTHER GEOLOGIC HAZARDS TO COASTAL COMMUNITIES OF SOUTHEASTERN ALASKA

Open-file report

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REGIONAL AND OTHER GENERAL FACTORS BEARING ON EVALUATION OF EARTHQUAKE AND OTHER GEOLOGIC HAZARDS TO COASTAL COMMUNITIES OF SOUTHEASTERN ALASKA

By

Richard W. Lemke and Lynn A. Yehle

ABSTRACT

The great Alaska earthquake of March 27, 1964, brought into sharp focus the need for engineering geologic studies in seismically active regions. As a result, nine communities in southeastern Alaska were selected for reconnaissance investigations as an integral part of an overall program to evaluate earthquake and other geologic hazards in most of the larger Alaska coastal communities. This report gives background information on the regional and other general factors that bear on these evaluations.

Southeastern Alaska, about 525 miles long and averaging about 125 miles in width, consists of a narrow mainland strip and numerous islands. For the most part, it is a region of rugged relief with numerous glaciers capping many of the higher mountainous areas and with long linear fiords forming the inland waterways. A maritime climate prevails with mild winters and cool summers. The southeastern part of the region receives the highest precipitation in the continental United States. Ketchikan, with a population of 6,994 in 1970, is the largest city. Geology and structure of the area are complex. Igneous, metamorphic, and sedimentary rocks crop out and range in age from Paleozoic to Tertiary. Surficial deposits of Pleistocene and Holocene age mantle many areas.

All of southeastern Alaska, except probably the highest peaks, was covered by glacier ice advances of late Pleistocene age. Major deglaciation was well advanced by 10,000 years ago--a time which approximately marks the end of the Pleistocene and the beginning of the Holocene. There followed a period of warm climate called the Hypsithermal, which in southeastern Alaska began 7,000-8,000 years ago and ended about 4,800-3,500 years ago. Glaciers in most places receded back of their present positions. The Hypsithermal was followed by an interval (termed Neoglaciation) of cooler climate and resurgence of glacier ice which continues to the present, although most glaciers are now rapidly receding.

During the past 10,000 years worldwide sea level has risen about 100 feet, but during the past 4,000 years it has risen only about 10 feet or about 0.03 inch per year. With sea level used as a datum, the amount of sea-level rise must be added to the apparent uplift of land for the time under consideration to determine the actual amount of land uplift.

The widespread presence of emergent marine deposits, several hundred feet above sea level, demonstrates that the land in southeastern Alaska

has been uplifted since the last major deglaciation. The greatest known uplift is in the vicinity of Juneau where glaciomarine deposits are present 750 feet above present sea level. Part of southeastern Alaska is presently undergoing one of the most rapid rates of uplift of any place in the world. The fastest emergence is occurring in the Glacier Bay area where the land is being uplifted relative to sea level approximately 3.9 cm per year. Most or all of the uplift appears to be due to rebound as a result of deglaciation.

Southeastern Alaska lies within the circum-Pacific earthquake belt, one of the world's greatest zones of seismic activity. During historic time, there have been five earthquakes in the region with magnitudes of 8 or greater, three with magnitudes of 7 to 8, eight with magnitudes of 6 to 7, more than 15 with magnitudes of 5 to 6, and about 140 recorded earthquakes with magnitudes smaller than 5 or of unassigned magnitudes. All of the earthquakes with magnitudes 8 or greater, and a large proportion of the others, appear to be related to the active Fairweather-Queen Charlotte Islands fault system or its western extension, the Chugach-St. Elias fault. Earthquake epicenters on the Denali fault system, the other major fault system in southeastern Alaska, are few in comparison. However, because high microearthquake activity has been recorded recently on this system and earthquakes of moderate size have occurred on some of its segments, the Denali fault system probably should not be dismissed as a relict fault system of no current tectonic importance. There are numerous other known faults, as well as lineaments that may be faults of varying degrees of tectonic activity in southeastern Alaska, adjacent Canada, and eastern Alaska. One of these elements is the Totschunda fault system, which connects with the Denali fault system in eastern Alaska; it has been very active during Holocene time but few historical earthquake epicenters appear to be related to it.

Both historical seismicity and geologic conditions, such as frequency and recency of faulting, must be considered together to permit an assessment of the future earthquake probability of an area. Data are too few for both factors for an accurate evaluation to be made of earthquake probability in southeastern Alaska. However, information compiled in the form of strain-release and seismic-zone maps permit some generalizations. Thus, it is tentatively concluded that most, if not all, of southeastern Alaska should be placed in seismic zone 3, a zone in which earthquakes of magnitude greater than 6 will occur from time to time and where there may be major damage to manmade structures.

Inferred effects from future earthquakes in southeastern Alaska include: (1) surface displacement along faults and other tectonic land-level changes, (2) ground shaking, (3) compaction, (4) liquefaction in cohesionless materials, (5) reaction of sensitive and quick clays, (6) water-sediment ejection and associated subsidence and ground fracturing, (7) earthquake-induced subaerial slides and slumps, (8) earthquake-induced subaqueous slides, (9) effects on glaciers and related features, (10) effects on ground water and stream flow, and (11) tsunamis, seiches,

and other abnormal water waves. Because of the reconnaissance nature of our studies in the coastal communities and the sparsity of laboratory data on physical properties of geologic units in each area studied, the inferred effects must be largely empirical and generalized. Therefore, the inferences are based in large part upon the effects of past major earthquakes in Alaska and elsewhere, particularly upon the well-documented effects of the Alaska earthquake of March 27, 1964.

Buildings, highways, bridges, tunnels, harbor facilities, pipelines, canals, and other manmade structures may be severely damaged or destroyed by fault displacement or related tectonic land-level changes in southeastern Alaska. Direct damage from fault rupture would be restricted virtually to structures built directly athwart the fault. In California and Nevada, fault rupture almost always accompanies shocks of magnitude 6.5 or greater. The Alaska earthquake of March 27, 1964, and the Chilean earthquake of May 22, 1960, dramatically illustrated the severe adverse effects that can result from uplift or subsidence over a wide area.

The variable most responsible for the degree of shaking at any epicentral distance is the type of ground. Generally, shaking is considerably greater in poorly consolidated deposits than in hard bedrock, particularly if the deposits are water saturated. Severe shaking of alluvial deposits and manmade fill, with resultant heavy damage, is well documented from the records of many past earthquakes.

Damage commonly has been heavy as a result of ground settlement caused by compaction of loose sediments by shaking during an earthquake. This has been especially true where compaction was accompanied by tectonic downdrop of land, such as occurred during the Chilean earthquake of 1960 and the Alaska earthquake of 1964. Loosely emplaced manmade fill, deltaic deposits, beach deposits, and alluvial deposits may be susceptible to compaction in southeastern Alaska during a severe earthquake.

Liquefaction of sand and silt is a fairly common effect of large earthquakes. It was well illustrated at Niigata, Japan, during the earthquake of June 16, 1964, and resulted in extensive damage. When part of a sloping soil mass liquefies, the entire mass can undergo catastrophic failure and can flow as a high-density liquid. In southeastern Alaska, deltaic deposits probably would be most susceptible to liquefaction.

Sensitive and quick clays, which lose a considerable part of their strength when shaken, commonly fail during an earthquake and become rapid earthflows. Extensive studies were made of the sensitivity of the Bootlegger Cove Clay at Anchorage because of the marked loss of shear strength and dramatic failures of the deposits during the Alaska earthquake of 1964. If similar sensitive clays are present in some places in southeastern Alaska, they most likely are in some of the emergent fine-grained marine deposits; supporting data to confirm their presence, however, are largely lacking.

Records of some 50 major earthquakes show that in at least half of the instances water and sediment have been ejected from surficial deposits. Water-sediment ejection and associated subsidence and ground fracturing commonly cause extensive damage to the works of man. Ejecta may fill basements and other low-lying parts of buildings. Agricultural land can be covered with a blanket of infertile soils, and small ponds can be filled or made shallow. In southeastern Alaska these phenomena are most likely to occur on valley floors, deltas, tidal flats, alluvial fans, swamps, and lakeshores.

Earthquake-induced sliding on land generally is confined to steep slopes but may take place in fine-grained deposits on moderately to nearly flat surfaces if the deposits are subject to liquefaction. A large rockslide triggered by the Lituya Bay, Alaska, earthquake of July 10, 1958, generated a wave that surged up the opposite wall of the inlet to a record height of 1,740 feet. During the Hebgen Lake, Montana, earthquake of August 17, 1959, a spectacular rockslide plunged into the Madison River canyon, buried 28 people, dammed the river, and created a large lake. Earthquake records are replete with accounts of sliding of surficial deposits during moderate to large earthquakes. Most or all of the general factors that favor subaerial landsliding are present in southeastern Alaska.

Earthquake-induced subaqueous slides can produce adverse effects both nearshore and some distance offshore. Nearshore sliding may progress shoreward and destroy harbor facilities and other structures, commonly with substantial loss of life. Disastrous large submarine slides occurred along the fronts of deltas in Seward and Valdez during the Alaska earthquake of 1964. In similar fashion, the largest submarine slides in southeastern Alaska likely will be triggered along the larger delta fronts. Sliding farther offshore can constitute a threat to navigation because of changes in water depths. Also underwater sliding can break communication cables.

Glaciers were not greatly affected by the Alaska earthquake of 1964 despite the fact that about 20 percent of the area that underwent strong shaking is covered by ice. In contrast, the cataclysmic avalanche of ice and rock that fell from a high glacier-covered peak in Peru during the earthquake of May 31, 1970, produced devastating effects downvalley on man and his works in the form of mudflows. Most towns in southeastern Alaska are sufficiently distant from glaciers so as not be to directly affected.

Both the Alaska earthquake of 1964 and the Hebgen Lake, Montana, earthquake of 1959 significantly affected ground- and surface-water regimens. Water levels in some wells declined whereas in others flow increased. Some springs discharged at a rate three times as much as normal; flow of others decreased or stopped. Discharge of many streams increased markedly. Most or all of the effects described above could occur in parts of southeastern Alaska during future large earthquakes.

Tsunamis, seiches, and other abnormal water waves associated with large earthquakes commonly cause vast property damage and heavy loss of life. Tsunami effects can be devastating to coastal areas as far as many thousands of miles from their generation source. Seiche effects generally are confined to inland bodies of water or to relatively enclosed coastal bodies of water. Abnormal waves generated by submarine sliding or by subaerial sliding into water generally produce only local effects but may be highly devastating. Tsunami waves resulting from the Chilean earthquake of 1960 inflicted extensive damage and loss of life on coastal communities throughout a large part of southern Chile, and significant runups and damage were recorded in many places throughout the Pacific Ocean area. The tsunami waves generated by the Alaska earthquake of 1964 struck with devastating force along a broad stretch of the Alaska coast and produced heavy property damage and loss of life as far away as Crescent City, Calif. Seiche waves generated by that earthquake reached runup heights of 20-30 feet on some lakes in Alaska, and water-level fluctuations were recorded on streams, reservoirs, lakes, and swimming pools in States bordering the Gulf of Mexico. Waves generated by submarine sliding struck violently at a number of places during or immediately after the quake and were the major cause of loss of life and damage to property. Slide-generated waves probably would have a higher destructive potential in southeastern Alaska than either tsunami waves or seiche waves because of their possibly higher local runups and because they can hit the shores almost without warning during or immediately after an earthquake.

Nonearthquake-related geologic hazards, although generally far less dramatic than those related to earthquakes, tend to occur so much more frequently or persistently that their aggregate effects can be significant. Three kinds of geologic hazards of this type are discussed: (1) nonearthquake-induced landsliding and subaqueous sliding, (2) flooding, and (3) land uplift.

The potential for nonearthquake-triggered landsliding in southeastern Alaska ranges widely from place to place. Past sliding generally furnishes the clue in the prediction of where and in what materials future sliding will occur. Fast-moving rockslides, debris slides, and mudflows can be expected to occur from time to time on steep slopes and be highly destructive to highways, power plants, pipelines, buildings, and other facilities located on a slope or at its base. Present slow downslope movement of talus can be expected to continue at the same general rate unless conditions are changed by man or there are climatic changes. Snow and debris avalanches can be especially hazardous during winter months. Long-inactive landslides may be triggered into renewed activity or new slides may be created by man-induced modifications. Accelerated slope erosion and debris flows may follow large-scale clearing and cutting of timber. Subaqueous sliding can be expected to occur periodically along fronts of deltas and on other oversteepened underwater slopes.

Floods have been common in parts of southeastern Alaska because of heavy precipitation and rapid runoff from steep slopes with resulting heavy damage to roads and other facilities. Continued damage can be expected in the future unless more remedial measures are taken.

Current uplift of land in southeastern Alaska, although probably not affecting man significantly in a short period of time, may have some adverse long-term effects. These long-term effects should be borne in mind when facilities such as docks and boat harbors are constructed on or near the shore, where there is a critical relation between height of land and water.

INTRODUCTION

Purpose and scope of study

The great Alaska earthquake of March 27, 1964, brought into sharp focus the need for engineering geologic studies in seismically active regions. As a result, nine communities in southeastern Alaska were selected by us for reconnaissance investigation as part of an overall program of earthquake studies recommended by the Federal Reconstruction and Development Commission for Alaska. Initiation of the studies was based on the premise that some Alaskan communities, which were too far from the area of strong ground motion to be affected by the 1964 earthquake, may have geologic settings similar to those of towns heavily damaged by that quake. The earthquake history of Alaska strongly argues that some of these communities will be adversely affected by future large earthquakes.

Reconnaissance studies were completed by us in the following community areas in southeastern Alaska: (1) Haines, (2) Hoonah, (3) Ketchikan, (4) Metlakatla, (5) Petersburg, (6) Sitka-Mount Edgecumbe, (7) Skagway, (8) Wrangell, and (9) Yakutat (see fig. 1 for locations). Detailed studies have been made in Juneau and vicinity by R. D. Miller (1972). These studies attempt to evaluate future effects of earthquake hazards, as well as other geologic hazards in these communities and immediately adjacent areas. The resulting evaluations should be useful in city and regional land-use planning so that new facilities can be built, as nearly as possible, in the best geologic environment and can be designed so as to minimize future damage and loss of life.

This report gives background information on the regional and other general factors bearing on evaluation of earthquake and other geologic hazards to the coastal communities that we have studied in southeastern Alaska. As such, it is our intent that it be used in conjunction with the reports on the individual communities. A comprehensive presentation of the geology and structure of southeastern Alaska is outside the scope of this report. Thus, descriptions of these elements are brief and provide only general background data pertinent to the purpose of the report. More local detail is given, however, in the reports of the specific coastal communities.

Because of the reconnaissance nature of our studies in southeastern Alaska, we have depended heavily upon the published and unpublished work of others. A comprehensive list of references is given at the end of this report. Additional references are given in the reports on specific coastal communities. In addition, we gratefully acknowledge the many sources of unpublished information and the complete cooperation from Federal, State, and city organizations and individuals.

Geographic and geologic setting

Southeastern Alaska, known as the Alaska Panhandle, extends along the Pacific Ocean from the vicinity of Icy Bay (fig. 1) southeastward

to the Portland Canal. It is about 525 miles long and, except for a very narrow northwestern strip, is in most places about 125 miles wide. The region consists of a narrow mainland strip adjoining Canada, several large outlying islands, and very numerous smaller islands. The islands, together with similar islands along the Canadian coast to the southeast, enclose the well-known "Inside Passage" waterway from Seattle to Juneau and other towns in southeastern Alaska.

The region, for the most part, is one of rugged relief with numerous glaciers capping many of the higher mountainous areas or occupying valleys and long linear fiords forming inland waterways. In the northeastern part of the region the snow- and ice-clad mountains of the Coast Mountains rise sharply to altitudes of 8,000-10,000 feet. In the northwestern part of the region, the spectacular Fairweather Range, paralleled on the northeast by the equally spectacular St. Elias Range, rises to 12,000-15,000 feet in altitude only 15 miles from tidewater. Peaks are high ice-clad pyramids having steep-cliffed walls, sharp ridges, and spirelike summits (Wahrhaftig, 1965). Most of the Fairweather Range, as well as the St. Elias Range, is covered by glaciers, many of which descend to or nearly to tidewater. Somewhat less high and rugged topography characterizes most of the rest of southeastern Alaska although the partly glacier-covered Chilkat Range (west of Haines), as well as mountains on some of the larger islands, are highly dissected and rise sharply from the shores of the fiords and the outer coast. A low plain, formed mostly of glacial deposits, extends along the coast in the vicinity of Yakutat Bay. Also, more subdued topography, with local relief of 300-500 feet and reflecting glaciation, characterizes Kupreanof Island and northern parts of Prince of Wales Island (Wahrhaftig, 1965). Chatham Strait and its northward continuation as Lynn Canal, Chilkoot Inlet, and Taiya Inlet is about 240 miles long and, according to Shepard and Wanless (1971), is the longest fiord in the world. It ranges from about 2 to 15 miles in width and mostly from 1,000 to 2,500 feet in depth.

Part of southeastern Alaska receives the highest precipitation of any place in the continental United States. Precipitation ranges from approximately 160 inches annually at the southern end of Prince of Wales Island, southwest of Ketchikan, to approximately 40 inches in the vicinity of Skagway (Watson, 1959). The greatest single annual precipitation occurred at Little Port Walter (near south end of Baranof Island) in 1943, with a total of 269.30 inches (Searby, 1968). A maritime climate prevails with mild winters and cool summers. Mean daily minimum temperatures in January range from approximately 32°F at the southern end of Prince of Wales Island to 0°F north of Skagway; mean daily maximum temperatures in July range from approximately 60°F along most of the length of the outer coast to approximately 68°F north of Skagway (Watson, 1959). No permafrost is known to be present anywhere in southeastern Alaska although it possibly exists in some of the higher mountains.

Ketchikan, with a population of 6,994 (U.S. Bur. Census, 1971), is the largest city in southeastern Alaska. Juneau, with a population of 6,050, is the second largest. Most travel is by water or air. The Haines Highway, which extends northwestward from Haines for 159 miles

to connect with the Alaska Highway in Yukon Territory, Canada, is the only connecting road to the rest of the mainland. Short stretches of road extend out from the main towns. The Marine Highway Ferry System connects with most of the larger towns. The White Pass and Yukon Railroad extends between Skagway and Whitehorse in the Yukon Territory.

As depicted on the geologic map of Alaska (Dutro and Payne, 1954) and in many other maps and reports, the geology and structure of southeastern Alaska are complex. Rocks differ widely in composition and consist of igneous, metamorphic, and sedimentary rocks ranging in age from Paleozoic to Tertiary (Brew and others, 1966). Surficial deposits of Pleistocene age or Holocene age mantle many areas. Although much geologic mapping has been done by many workers, a number of areas, particularly away from the coast and the inner waterways, are imperfectly known. As will be discussed later, better regional knowledge of the locations of faults and recency of fault movement would be helpful to us in evaluating earthquake probability and resulting seismic risk. On the other hand, knowledge of bedrock types is much less important to us in this respect than knowing the distribution and physical characteristics of the surficial deposits because of a considerably smaller range in earthquake effects between different types of bedrock. Thus, bedrock descriptions in our coastal community reports generally are brief.

Glaciation and associated sea- and land-level changes

Regional glaciation

All of southeastern Alaska, except probably the highest peaks, was covered by glacier ice during advances of late Pleistocene age (Heusser, 1960). Contours drawn on inferred glacier-ice surfaces of these advances indicate that the ice attained an altitude of about 2,000 feet along coastal parts of southeastern Alaska and increased in elevation toward the northeast to an altitude of about 6,000 feet in the Coast Mountains (fig. 1) along the Alaska-British Columbia boundary (Coulter and others, 1965). If southeastern Alaska was glaciated prior to late Pleistocene time, evidence for such a glaciation has not been recognized.

Although major deglaciation did not occur everywhere in southeastern Alaska at the same time, it probably was well advanced in most places by 10,000 years ago, a time that approximately marks the end of the Pleistocene and the beginning of the Holocene. Glacier ice had started to withdraw from the Glacier Bay area (fig. 1) by 10,000 years ago and had receded back of its present position by 7,500 years ago (Goldthwait, 1963, 1966). It probably had largely disappeared from the Juneau area somewhat before 10,000 years ago (Lawrence, 1958; Heusser, 1960; Miller, R. D., 1972; Miller, R. D., oral commun., 1967). Radiocarbon dating (sample W-1734) of marine shells, collected by us 212 feet above present sea level in the Petersburg area (fig. 1), indicates deglaciation of that area prior to 12,400 years ago.

After major deglaciation, there followed a period of warm climate called the Hypsithermal interval. This interval probably began in southeastern Alaska 7,000-8,000 years ago and ended between about 4,800 and 3,500 years ago (Heusser, 1960; Goldthwait, 1963, 1966; Porter and Denton, 1967). In the Glacier Bay area, glaciers were present in the mountains during this interval but were largely absent in the fiords and lowland valleys. Thus, the glaciers in most places in that area and probably elsewhere in southeastern Alaska had receded farther back than their present positions (Goldthwait, 1963; Péwé and others, 1965; Goldthwait, 1966).

The Hypsithermal, in turn, was followed by an interval of cooler climate and a resurgence of glacier ice. The term Neoglaciation was applied (Porter and Denton, 1967) to this interval, which in southeastern Alaska began about 4,800-3,500 years ago and continues to the present. It has been marked by several fluctuations in glacier activity. An early widespread ice advance in the Glacier Bay area started about 4,800 years ago and continued until about 2,900 years ago before the glaciers again began to diminish in size (Goldthwait, 1966). A period of milder climate followed and persisted until about 270 years ago when glaciers again began to advance and appear to have attained their maximum Neoglacial positions in at least the Glacier Bay area of southeastern Alaska (Goldthwait, 1966; Porter and Denton, 1967). Glacier recession began again about 200 years ago and has continued with minor fluctuations until the present day (Péwé and others, 1965). Most glaciers in southeastern Alaska have receded very rapidly during this time. This has been dramatically evidenced in the Glacier Bay area where recession locally has exceeded 60 miles of which 18 miles has occurred in the Muir Inlet part of the bay during the period 1880-1960 (Lawrence, 1958; Goldthwait, 1963). Some trunk glaciers, such as in the Juneau area, are advancing today, but each at the expense of an adjacent shrinking trunk glacier (Maynard Miller, 1963).

Sea-level changes

Eustatic sea-level changes are those changes that pertain to worldwide changes of sea level that affect the entire ocean. Although a number of factors can account for the rise or fall of sea level, most probably reflect fluctuations of glaciers. During periods of large accumulation of glacier ice on land, sea level falls. Conversely, during periods of glacial melting and recession, sea level rises. Thus, there have been a number of fluctuations of sea level during Pleistocene and Holocene time--low levels during glacial stages and high levels during interglacial stages and postglacial time.

Eustatic sea-level studies have been made by many workers. The results of most of these studies have been summarized by Flint (1957), Fairbridge (1961), Curray (1961, 1965), Wexler (1961), Shepard (1963), Hicks and Shofnos (1965), Higgins (1965), Kenney (1965), Shepard and Curray (1967), Emery and Garrison (1967), and Bloom (1971). There is

divergence of opinion among these workers as to the amount of eustatic sea-level fluctuation during different periods of time. This divergence arises because of the difficulty of determining whether the land mass that is used as a sea-level datum has been stable (neither uplift or subsidence has occurred) during the period of sea-level measurements. Because of tectonism, isostatic adjustments due to glacier fluctuations, and adjustments caused by sediment load, most land masses, especially along the coast, have undergone some changes in altitude in fairly recent time. However, in spite of these difficulties, sufficient data based upon radiocarbon dates now seem to be available to determine the right order of magnitude of the sea-level fluctuations during approximately the past 16,000-18,000 years. This time interval, particularly the past 10,000 years, encompasses the period that is critical to our consideration of sea- and land-level changes in southeastern Alaska.

Approximately 15,000-18,000 years ago sea level stood about 430 feet (130 m) lower than today (Shepard and Curray, 1967; Milliman and Emery, 1968; Emery, 1969; Bloom, 1971). This low level probably reflected maximum glaciation during late Pleistocene time. Since then, sea level has been rising.

By 10,000 years ago, sea level had risen until it stood about 100 feet lower than at present[1/] (Shepard, 1963; Shepard and Curray, 1967; Redfield, 1967). However, 5,000-7,000 years ago the rate of rise started decreasing markedly until about 4,000 years ago the sea stood only about 10 feet below its present level (Redfield, 1967; Shepard and Curray, 1967). Thus, between 12,000 and 4,000 years ago, the rate of rise was 0.13 inch or 3.4 mm per year, but during the past 4,000 years it was 0.03 inch or 0.76 mm per year (Redfield, 1967). These rates of rise of sea level are important to the calculation of actual amount of land uplift; with sea level used as datum, the amount of eustatic sea-level rise must be added to the apparent uplift of land for the time under consideration.

Land-level changes

The widespread presence in southeastern Alaska of emergent shore, glaciomarine, and marine deposits, tens to hundreds of feet above sea level, demonstrates that the land has been uplifted in respect to sea level during late Pleistocene and Holocene time (McConnell, 1913; Chapin, 1918; Schofield and Hanson, 1922; Woodworth and Haight, 1927; Buddington and Chapin, 1929; Twenhofel, 1952; Heusser, 1960; Pierce, 1961; Goldthwait, 1963, 1966). Tidal records show that differential uplift of land is currently taking place (Hicks and Shofnos, 1965). The time and apparent rate of land emergence have been determined in places by radiocarbon dating of shells and other datable material in the emergent marine deposits.

1/ The 100-foot figure is taken from constructed curves drawn through a fairly wide scatter of data points; thus, some data indicate that sea level 10,000 years ago was higher than 100 feet whereas other data indicate a lower level.

The greatest known amount of land uplift in southeastern Alaska, since major deglaciation, has been in the vicinity of Juneau (fig. 1). Here, glaciomarine deposits have been uplifted as much as 750 feet above present sea level (R. D. Miller, 1972). If, as seems likely, the oldest deposits are about 12,000 years old, then about 150 feet should be added to the present height of the deposits to account for eustatic rise of sea level (Shepard, 1963; Shepard and Curray, 1967). Therefore, the land actually has been uplifted about 900 feet.

The maximum amount of land uplift north and northwest of Juneau is not known with any certainty. In the Glacier Bay area, evidence of maximum emergence is difficult to obtain because the uplifted deposits are largely buried under deposits of two later ice advances or under presently existing ice (Goldthwait, 1963). A total uplift of 300 feet or more in the Glacier Bay area, between 10,000 and 7,000 years ago is suggested by Goldthwait (1966). In the Haines area, the land has been uplifted at least 300 feet in respect to present sea level since major deglaciation and may have been uplifted as much as 600 feet or more (Lemke and Yehle, 1972). Marine shells have been found in the Haines area up to a present altitude of 50 feet. Those at a present altitude of 50 feet were dated at 11,020±400 years (sample W-2294).

Emergent fossiliferous marine sediments have been reported from numerous localities south and southeast of Juneau (McConnell, 1913; Chapin, 1918; Hansen, 1923; Buddington and Chapin, 1929). At Petersburg (fig. 1), we collected marine shells from an altitude of 212 feet that were dated at 12,400±800 years (sample W-1734); at Wrangell, we collected shells at altitudes of approximately 65 and 100 feet. Near Ketchikan, marine fossils were found at an altitude of about 85 feet on Gravina Island (Chapin, 1918). We also found marine shells at an altitude of 83 feet in the city of Ketchikan. Marine deposits at higher levels well may be present in the Ketchikan area but may not have been discovered because of the fewness of exposures. In the Portland Canal area in British Columbia near the southeastern part of southeastern Alaska (fig. 1), McConnell (1913) found shells in marine blue clays along the Bear River at between 343 and 348 feet above present mean tide. Hansen (1923) found similar clays, but without shells, at an altitude of 450 feet in the Portland Canal area on the Kitsa't River in British Columbia.

Part of southeastern Alaska is presently undergoing one of the most rapid rates of uplift of any place in the world (Whitten, 1970). The fastest rate of emergence is occurring in the Glacier Bay area (fig. 2) where the land is being uplifted at a rate of approximately 3.9 cm (1½ inches) per year (Hicks and Shofnos, 1965). Rates of uplift decrease progressively to the southeast with the possible zero line of uplift just south of Ketchikan. These data are based upon long and short series of observations made during 75 years (1887-1962) when differences in elevations of tidal bench marks (from Ketchikan to Glacier Bay) in respect to sea level were calculated. As discussed previously, eustatic sea level is believed to be rising during the past 4,000 years at a rate of

approximately 0.76 mm per year. Therefore, this amount per year probably should be added to the apparent uplift of land based upon tidal measurements to obtain the actual land uplift.

In attempting to determine the cause of land-level changes in southeastern Alaska since major deglaciation about 10,000 years ago, one finds it difficult to distinguish the amount of change that is attributable to rebound as a result of deglaciation from that which might be due to tectonism. If the maximum thicknesses of the last ice sheet of an area were known, as well as the time when deglaciation of that ice sheet started, then the amount of rebound resulting from unloading during deglaciation could be calculated for that region (Bloom, 1967, 1971; Crittenden, 1967; Andrews, 1968; McGinnis, 1968; Brotchie and Silvester, 1969). Any deviation of land-level change could then be attributable to tectonism. Unfortunately, thicknesses of the last major ice sheet or of succeeding lesser ice advances are imperfectly known. An additional critical complication is that most areas have been deglaciated for 10,000 years or more, whereas in other areas there have been one or more minor advances since then. Moreover, large areas are now covered by ice, which, in places like the Glacier Bay area, is rapidly receding.

In spite of the difficulties of differentiating the causes of land uplift in southeastern Alaska, it is concluded that most of the uplift has been due to rebound as a result of deglaciation. Areas of greatest known uplift are some distance inland from the outer coast where ice thicknesses are believed to have been the greatest and, thus, where greatest rebound would be expected. Certainly, the greatest present uplift is in areas where the most rapid deglaciation and unloading of the land is now taking place.

Much or most of the current uplift probably is the result of deglaciation during the last few hundred years rather than from major deglaciation about 10,000 years ago. As pointed out by Andrews (1968), the rate of glacioisostatic uplift in Arctic Canada was most rapid immediately following deglaciation with about one-third of the total rebound occurring in the first 1,000 years, 86.5 percent occurring in 5,000 years, and 100 percent being accomplished in 10,000 years. More recently, Bloom (1971) stated that rate of land recovery "decreases exponentially with a half-life on the order of 1,000 years." Using either rates of recovery, there should be little or no uplift of land occurring today in southeastern Alaska that is reflecting deglaciation that occurred 10,000 or more years ago. Radiocarbon dating in the Petersburg area supports Andrews' findings that much of the rebound occurs during the early stages of deglaciation. Marine shells collected by us from an altitude of 212 feet gave a date of 12,400±800 years (sample W-1734), whereas shells in growth position collected at an altitude of 29 feet were dated at 9,970±300 years (sample W-1738).

REGIONAL FACTORS BEARING ON EVALUATION OF EARTHQUAKE HAZARDS

Introduction

Most large shallow-focus earthquakes appear to be caused by sudden rupture of the earth's crust (Cluff and Bolt, 1967). Some earthquakes, generally small, are the result of volcanic activity or, more rarely, of other causes. The rupturing of the earth's crust is due to a sudden release of accumulated strain energy. The rupture, termed a fault, generally is accompanied by vertically and(or) horizontally displaced blocks of ground on opposite sides of the fault. The sudden rupturing along a fault generates seismic waves, which cause ground shaking. During larger earthquakes, this shaking along and in the region of a fault may be sufficiently strong to damage manmade structures. Shaking may also cause disastrous secondary effects such as landslides, liquefaction of sediments, and ground compaction. Structures built athwart a fault may be additionally damaged by vertical or horizontal movement on opposite sides of a fault.

The focus of an earthquake is the point below the earth's surface where rupturing on a fault first occurs. Focal depth is the distance of the focus from the earth's surface. Shallow-focus earthquakes are classified as those whose depth of focus is less than 70 km (43.4 miles); intermediate-focus earthquakes are those whose depth of focus is between 70 km and 300 km (186 miles); and deep-focus earthquakes are those whose depth of focus is greater than 300 km (Richter, 1958).

The epicenter of an earthquake is the point on the earth's surface directly above the focus and, for accurate location, must be determined instrumentally. It commonly, but not always, is the location of the most pronounced earthquake effects. The hypocenter is the subterranean point directly below the epicenter where rupturing first starts and, therefore, is a synonym for focus (Richter, 1958).

Earthquake size generally is measured by magnitude and intensity. Duration of strong ground motion (shaking) and maximum acceleration attained during shaking also are factors that determine the severity of the earthquake as it relates to seismic risk to man.

Magnitude is based upon the seismic records of an earthquake as recorded on seismographs. As originally defined by Richter (1958), the magnitude figure is the "logarithm of the maximum amplitude on a seismograph written by an instrument of a specified standard type at a distance of 100 kilometers (62 miles)* * * *. Because the scale is logarithmic, every upward step of one magnitude unit means multiplying the recorded amplitude by 10." The largest earthquake possible probably cannot exceed a magnitude of about 9.0--the largest recorded since the scale was devised had a magnitude of 8.9 (Cluff and Bolt, 1967). Unlike intensity scales, magnitude is a single figure from which the approximate energy released by the earthquake can be determined. It has been calculated for shallow-focus earthquakes, such as occur in California, that an

increase of a unit in magnitude indicates an increase in energy of approximately 32 times (Cluff and Bolt, 1967). Thus, a 2-unit increase in magnitude, such as from 5 to 7, results in approximately a 1,000-fold increase in release of energy. Although geological conditions in southeastern Alaska may be sufficiently different from those in California to alter these relations somewhat, the relations probably are generally applicable.

Intensity is a measure of the severity of ground motion (shaking) at a specific location during an earthquake. Therefore, unlike magnitude, which is a constant for any one earthquake, intensity varies from place to place during an earthquake. Although intensity commonly is highest at or near the epicenter, differences in ground conditions and other factors may markedly affect the degree of shaking and, hence, the intensity. At least 44 scales have been developed to measure intensity (Barosh, 1969), but in the United States the Modified Mercalli scale is in general use. The Modified Mercalli scale is a 12-point scale, generally denoted by Roman numerals, to describe the varying degrees of sensation to man (necessarily subjective in part), the dislocation or damage to manmade structures, and the effects upon the earth's surface (see table 1 for an abbreviated description of effects for each unit of the scale; more detailed descriptions are given by Richter, 1958). In relation to engineering aspects, the scale is best adapted to describing the behavior of ordinary construction of small to moderate size, especially masonry. Effects of shaking on tall buildings and on elevated structures, such as water tanks, are commonly of different character, and the effects extend to greater distances in earthquakes of large magnitude (Richter, 1959). An isoseismal map, which shows different intensities by contours, commonly is compiled after an earthquake to show the geographical distribution and range of intensities.

It is important to know how much intensity decreases with distance from the epicenter of an earthquake in trying to evaluate the effects from a distant earthquake. Unfortunately, the decrease in intensity with increasing distance does not produce a simple curve, owing to a number of other variables that affect intensity (Barosh, 1969). Therefore, average-intensity curves are constructed by drawing a smoothed curve through a scatter of points. Such a curve shows that average intensity decreases fairly regularly with increasing epicentral distance beyond an epicentral distance of 20-100 km (12.4-62 miles); at lesser distances, the intensity increases greatly with decreasing epicentral distance. Average-intensity-epicentral-distance curves for southern California prepared by Barosh (1969) show, for example, that: (1) an earthquake having an intensity of XI at the epicenter would have an intensity of between VI and VII at a distance of 100 km, an intensity of III at a distance of 400 km (248 miles), and between I and II at a distance of approximately 750 km (465 miles); and (2) an earthquake having an intensity of approximately VI at the epicenter would have an intensity between I and II at a distance of approximately 100 km (62 miles). A curve by Gutenberg and Richter (1956) shows the radius of perceptibility as a function of epicentral intensity. This curve shows, for example, that: (1) there would be perceptible

intensity to a distance of 600 km (372 miles) for an earthquake whose intensity was XI at the epicenter, and (2) perceptible intensity to a distance of 100 km (62 miles) for an earthquake whose intensity was VI at the epicenter. Neumann (1954) has constructed distance-intensity graphs for a number of earthquakes.

Another factor affecting the severity of an earthquake is the maximum ground acceleration that takes place during shaking. Acceleration, in this sense, is defined as the rate of change of ground velocity (modified from Hodgman and others, 1956, p. 2838). It is commonly expressed as a percent of the acceleration of gravity (about 981 cps^2 or 32.2 fps^2; adopted as a standard by the International Committee on Weights and Measures). Thus, it is common to refer to ground acceleration during an earthquake as 0.3g (30 percent) acceleration of gravity. At an acceleration of 1.0g, objects would be flung into the air.

Very approximate relations between earthquake magnitude, energy release, acceleration of gravity, and intensity are shown in table 1, as modified from the U.S. Atomic Energy Commission (1963). These relations are believed to apply fairly well in southern California where the average focal depth of an earthquake is about 10 miles (16 km) and may apply reasonably well for some earthquakes in southeastern Alaska and vicinity. It should be strongly emphasized, however, that these relations are only rough approximations and may vary widely for some earthquakes. This is particularly true in showing the relation between magnitude and intensity. As pointed out by Hodgson (1966) and by Cluff and Bolt (1967), intensity depends on a number of factors, such as the focal depth of the earthquake, distance from the epicenter, duration of shaking, the nature of the soil and the types of manmade structures, and the care with which the structures have been built. Differences in soil types particularly cause great variations in intensities in the same general area. Water-saturated alluvium, for example, may have an intensity 2 to 3 units higher on the Modified Mercalli scale than nearby igneous rock. Also, earthquakes of relatively small magnitude may have high accelerations such as the Parkfield-Cholame earthquake of 1966 in California. One of three shocks of this earthquake, which had a magnitude of 5.5, produced a short-duration horizontal acceleration of 0.5g (Cloud, 1966). According to table 1, the expectable acceleration from an earthquake of that magnitude would have been only about 0.07g. Perhaps even more divergence from the normal occurred during the San Fernando, California, earthquake of February 9, 1971. This earthquake, of magnitude 6.6, produced accelerations of as much as 1g (Maley and Cloud, 1971). According to table 1, the expectable acceleration would have been slightly less than 0.3g.

As yet it is not possible to predict when or where the next destructive earthquake will strike in the world or what its size and other characteristics will be. We do know, however, that some regions are much more likely to have destructive earthquakes in the future than other regions. The degree of likelihood is based upon two factors: (1) the seismicity

Table 1.--Approximate relations between earthquake magnitude, energy, ground acceleration, acceleration in relation to gravity, and intensity (modified from U.S. Atomic Energy Commission, 1963)

M[1]	E[2]	a[3]	a/g[4]	I[5]	
	10^{14}			I	Detected only by sensitive instruments
	10^{15}	2		II	Felt by a few persons at rest, especially on upper floors; delicate suspended objects may swing
3		3			
	10^{16}	4		III	Felt noticeably indoors, but not always recognized as a quake; standing autos rock slightly, vibration like passing truck
		5	.005g		
		6			
		7			
	10^{17}	8		IV	Felt indoors by many, outdoors by a few; at night some awaken; dishes, windows, doors disturbed; motor cars rock noticeably
4		10	.01g		
	10^{18}	20		V	Felt by most people; some breakage of dishes, windows, and plaster; disturbance of tall objects
		30			
5	10^{19}	40		VI	Felt by all; many frightened and run outdoors; falling plaster and chimneys; damage small
		50	.05g		
		60			
		70			
	10^{20}	80		VII	Everybody runs outdoors; damage to buildings varies, depending on quality of construction; noticed by drivers of cars
		90			
		100	.1g		
6	10^{21}	200		VIII	Panel walls thrown out of frames; fall of walls, monuments, chimneys; sand and mud ejected; drivers of autos disturbed
		300			
		400			
7	10^{22}	500	.5g	IX	Buildings shifted off foundations, cracked, thrown out of plumb; ground cracked; underground pipes broken
		600			
		700			
		800			
		900			
		1000	1g	X	Most masonry and frame structures destroyed; ground cracked; rails bent; landslides
8	10^{23}				
		2000		XI	Few structures remain standing; bridges destroyed; fissures in ground; pipes broken; landslides; rails bent
		3000			
	10^{24}	4000		XII	Damage total; waves seen on ground surface; lines of sight and level distorted; objects thrown up into air
		5000	5g		

These relations until 1971 are believed to have applied fairly well in southern California where the average focal depth of earthquakes has been about 10 miles (16 km). (See Gutenberg and Richter, 1956; Hodgson, 1966.) However, revisions of these relations may be necessary because of the exceptionally high accelerations resulting from the San Fernando, Calif., earthquake of February 9, 1971, when the earthquake, of magnitude 6.6, produced accelerations of as much as 1 g (Maley and Cloud, 1971).

1/M, magnitude scale, according to Richter (1958). 2/E, energy, in ergs. 3/a, ground acceleration, in centimeters per second^2. 4/a/g, ground acceleration shown as a percent of the acceleration of gravity (about 981 cps^2 or about 32.2 fps^2; adopted as a standard by the International Committee on Weights and Measures). 5/I, Modified Mercalli intensity scale (abridged from Wood and Neumann, 1931); complete description of scale units given in Richter (1958).

or historical record of earthquakes in a certain region, and (2) the degree of tectonic activity of the region, as indicated chiefly by the recency of fault movement. Assessment of these two factors affords a means of determining the earthquake probability of a region and the resultant seismic risk.

Seismicity

Introduction

In many parts of the world it has been customary to assess the earthquake probability of a region more or less solely on the basis of the seismicity of the region. Seismicity is defined as the historical frequency of earthquakes occurring in a certain region or area.

The historical seismic record in Alaska, however, is far too short to assess adequately, on this basis alone, the earthquake probability in any particular part of the State. There are only about 20 recorded accounts of earthquakes in or near the State that occurred more than 100 years ago. In fact, accounts are few until about the beginning of the present century. This is not because of a greater number of earthquakes during the later years but because of better detection and reporting methods. Thus, in the 10 years preceding the large Alaska earthquake of March 27, 1964, about 300 hypocenters were located in or offshore from Alaska--more than in the previous 50-year period (Tobin and Sykes, 1966).

Seismicity of Alaska

A large part of Alaska lies within the circum-Pacific earthquake zone, one of the world's greatest zones of seismic activity. The State is the most seismically active region in the United States (Tobin and Sykes, 1966; Tobin and Sykes, 1968; Pitman and Hayes, 1968) and accounts for 6 percent of the world's shallow earthquakes (St. Amand, 1957; Wood, 1966). The earthquakes have occurred chiefly in two seismic zones. One zone, known as the Aleutian arc zone, is about 2,500 miles long and 200 miles wide and includes the Aleutian Islands, the Aleutian trench, Alaska peninsula, Alaska Range, and eastward as far as the Copper River (see fig. 3 for locations). The Aleutian trench part of this zone has very frequent shocks and a significant part of the world's great earthquakes. The other zone is in southeastern Alaska and along the coast in British Columbia, particularly the Queen Charlotte Islands area (fig. 3); its southern terminus is near the west coast of Vancouver Island (Wood, 1966).

According to Tobin and Sykes (1966; 1968), about 80 percent of the earthquakes in Alaska have been shallow earthquakes and the remainder have been intermediate depth. Most of the shallow earthquakes in Alaska were in a belt that extends from the Alaska Peninsula and adjacent offshore areas into central Alaska. All the earthquakes in southeastern

Alaska were shallow. Earthquakes of intermediate depth were confined to a narrow belt extending from the Aleutian Islands, through the Alaska Peninsula, and into central Alaska. The northern limits of this belt approximately coincide with the Denali fault system (fig. 3). The deepest of these intermediate-depth earthquakes was about 170 km. No deep-focus earthquakes have been detected in Alaska (Tobin and Sykes, 1966).

An idea of the frequency and size of earthquakes in Alaska can be gained from the following figures. During the period 1899 through 1964, there were eight earthquakes of magnitude 8 or greater, 56 between magnitudes 7 and 8, and 234 between magnitudes 6 and 7 (Wood, 1966). In the period July 1788 to August 1961, there were approximately 880 earthquakes of magnitude approximately 5 or greater (Davis and Echols, 1962). Between 1786 and December 31, 1964, approximately 1,560 felt earthquakes were reported (Wood, 1966).

Seismicity of southeastern Alaska and adjacent parts of Canada

Southeastern Alaska and adjacent parts of Canada lie in the second most active seismic zone in Alaska. The major historic seismic activity of this zone has been concentrated in an area about 300 miles wide, which extends about 200 miles offshore and parallels the coastline from the vicinity of Cape Yakataga (long 142°W., lat 60°N.) southeastward along the western coast of Baranof Island and continues along the western coast of Queen Charlotte Islands and Vancouver Island (in British Columbia).

Figure 4 shows the distribution of earthquakes that have occurred through 1969 in southeastern Alaska, adjacent Canada, and vicinity offshore, which have known epicenters. The epicenters and magnitudes have been determined from various sources listed in figure 4. These sources indicate that there have been five historical earthquakes of magnitude 8[2] or greater, three of magnitudes 7 to 8, eight of magnitudes 6 to 7, more than 15 of magnitudes 5 to 6, and about 140 with magnitudes less than 5 or of unassigned magnitudes. In addition, there undoubtedly have been many additional unrecorded earthquakes since 1899.

Three of the five earthquakes of magnitude 8 or greater occurred in 1899 and 1900 in the general area of Yakutat Bay. The first, on September 4, 1899, had a magnitude of 8.3 (Wood, 1966), and its epicenter was probably about 30 miles northwest of Icy Bay (fig. 1), in an unsettled area. The second and largest earthquake was on September 10, 1899, the epicenter being near Yakutat Bay. It had a magnitude of 8.6, an intensity of XI, and was widely felt (Wood, 1966). Because of the

[2] The Lituya Bay earthquake of July 10, 1958 (designated as "P" on fig. 4), was assigned a magnitude of 8 by Davis and Echols (1962) and Davis (1964). However, the U.S. Coast and Geodetic Survey (Wood, 1966) assigned it a magnitude of 7.9. Because the magnitude is either 8 or nearly 8, we show it in the category of earthquakes of magnitude 8 or greater.

sparsity of people in the area, there was neither loss of life nor property damage. Topographic changes, however, were profound. Investigations 6 years later by Tarr and Martin (1912) showed uplift and subsidence of shorelines in Yakutat Bay and contiguous waterways. A maximum uplift of 47 feet 4 inches was measured on the west shore of Disenchantment Bay (at the head of Yakutat Bay) and 1-7 feet of subsidence in nearby unconsolidated deposits. This uplift is the largest recorded in the world for a single seismic event except for an indicated displacement of 50 feet or more offshore from Montague Island (fig. 3) that may have resulted from the Alaska earthquake of March 27, 1964 (Malloy, 1964). The epicenter of the third earthquake, which occurred October 9, 1900, is uncertain but apparently it was in the same general area as the one of September 4, 1899. It was felt strongly as far away as Kodiak (fig. 3) where there were nearly continuous aftershocks for 6 hours (Wood, 1966).

The epicenter of the fourth historical earthquake of magnitude 8 or greater was offshore from the northern end of Graham Island, northernmost of the Queen Charlotte Islands in British Columbia. It occurred August 22, 1949, had a magnitude of 8.0 or 8.1, and is the largest recorded earthquake in Canada (Gutenberg and Richter, 1954; Davis and Echols, 1962; Hodgson, 1966; Milne, 1966). Because it was located offshore from a sparsely inhabited area, damage was minor.

The epicenter of a fifth earthquake of magnitude 8 or 7.9 is the well-documented and spectacular Lituya Bay earthquake of July 10, 1958. Although its epicenter (designated as P on fig. 4) is plotted by Tobin and Sykes (1966) as 20-25 miles southeast of Lituya Bay, the most awesome effects of the quake were most evident in the Lituya Bay area. For this reason it commonly has been referred to as the Lituya Bay earthquake (Wood, 1966). Probable movement along the Fairweather fault, accompanied by severe ground motion, started rockslides and earthslides on high slopes above the bay in Gilbert Inlet, the northwest arm of Lituya Bay. One of these slides started from an altitude of about 3,000 feet, and 40 million cubic yards of rock and earth slid into the bay. This generated a giant wave whose runup on the opposite slope reached an elevation of 1,740 feet (D. J. Miller, 1960). Within the next 4 minutes, the wave traveled down the bay stripping soil and vegetation from 4 square miles of the shore, sinking two of three fishing boats near the mouth of the bay, and drowning two people. As much as 1,300 feet of ice sheared off the front of Lituya glacier near the rockslide in Gilbert Inlet. Gilbert Inlet and Crillon Inlet, the southeast arm of Lituya Bay, and the upper 2½ miles of Lituya Bay were almost solidly covered by floating ice blocks (D. J. Miller, 1960). At Crillon Lake, about 5 miles southeast of Lituya Bay, the southwest side of the Fairweather fault moved relatively northwest 21½ feet and up 3½ feet (Tocher, 1960). Lesser displacements have been noted along the entire length of the onshore segment of the fault. Tocher (1960) reported fresh surface breaks near some of the displacements observed by Tarr and Martin (1912) after the 1899 earthquakes, and Davis and Sanders

(1960) reported subsidence near Yakutat near some of the subsided areas of 1899. Great numbers of sand boils and fissures were formed on a low coastal plain southeast of Yakutat and people and animals were thrown from their feet (Davis, 1964). Near Yakutat, part of an island subsided into the sea and three people plunged to their deaths. The earthquake, which was assigned a maximum intensity of XI on the Modified Mercalli scale, was felt over an area of 400,000 square miles, which encompassed all of southeastern Alaska and extended as far south as Seattle, Wash., and as far northeast as Whitehorse, Canada (Wood, 1966).

Most epicenters for earthquakes of magnitude less than 8 are located offshore or are on land fairly near the coast (fig. 4). Of earthquakes with magnitudes between 7 and 8, two have epicenters near or on the coast and the third one is offshore; all are north of lat. 57°. Four of the earthquakes with magnitudes between 6 and 7 were in the vicinity of the Queen Charlotte Islands or somewhat to the northwest, one was offshore from the southern end of Baranof Island and three were inland northwest of Haines of which two were in the Yukon. Of the 15 earthquakes of assigned magnitudes between 5 and 6, all have epicenters approximately on or north of lat. 58°. There probably have been a considerably greater number of earthquakes in this magnitude range but data are insufficient to permit locating of epicenters and calculation of magnitudes. This may account for the lack of earthquakes south of lat. 58°N. in this range of magnitude. Epicenters of earthquakes of magnitude less than 5, or of unassigned magnitude, are somewhat randomly distributed offshore and along or fairly close to the coast. An exception is an area approximately between lat 55°N. and lat 56°N., where there is nearly an absence of epicenters of earthquakes with magnitudes less than 5, or with unassigned magnitudes, and where there are no epicenters of earthquakes with greater assigned magnitudes.

Major faults and recency of fault movement

General description and inferred fault connections

Southeastern Alaska lies within the active tectonic belt that rims the northern Pacific Basin. It has been tectonically active since at least late Paleozoic time and the bedrock outcrop pattern is largely the result of late Mesozoic and Tertiary deformational, metamorphic, and intrusive events (Brew and others, 1966). Large-scale faulting with right-lateral strike-slip movement[3], was common during Tertiary

[3] There are several kinds of faults: A normal fault is one in which the hanging wall (the block above the fault plane) has moved downward in relation to the footwall (the block below the fault plane); on a vertical fault, either side may have moved down in relation to the other side. A thrust fault is a low-angle fault on which the hanging wall has moved upward relative to the footwall. A strike-slip fault is a fault on which there has been lateral displacement approximately parallel to the strike of the fault. (If the movement is such that, when an observer looks across a fault, the block across the fault has moved relatively to the right, then the fault is a right-lateral strike-slip fault; if the displacement is such that the block across the fault has moved relatively to the left, then the fault is a left-lateral strike-slip fault.)

time. Horizontal movement, as well as vertical movement, has continued in places during Quaternary time. Studies by St. Amand (1957), Twenhofel and Sainsbury (1958), Gabrielse and Wheeler (1961), Lathram (1964), Wilson (1965), Grantz (1966), Brew, Loney, and Muffler (1966), Loney, Brew, and Lanphere (1967), Tobin and Sykes (1968), Plafker (1969, 1971), Souther (1970), and Richter and Matson (1971), as well as those of earlier workers, have contributed substantially to the regional structural knowledge of southeastern Alaska and adjacent areas. The structural patterns of large areas, however, are incompletely known. As a result, there is considerable divergence of opinion as to age of faulting, amount of displacement, and whether or not certain faults connect. These differences in interpretation between the workers on the problem, as well as a generalized description of the tectonic setting of the region, are summarized here. More detailed descriptions of the major fault elements are discussed later.

The trends of many linear fiords of southeastern Alaska are believed to be controlled by major faults or fault zones (Twenhofel and Sainsbury, 1958); other fiords, such as the northeast-trending ones transecting Baranof Island (fig. 1), are believed to be controlled by joints (Brew and others, 1963). The fiords are formed along faults or joints chiefly as a result of outlet glaciers scouring and deepening preglacial river valleys whose courses followed, at least in part, the more easily erodible fault or joint planes. Many other linear features such as straight valleys, coastlines, and troughlike depressions reflect faults, shear planes, and joints. The more conspicuous of these lineaments, most of which are believed to be fault traces, are shown on figure 5.

Two of the most prominent fault systems in southeastern Alaska and surrounding regions are: (1) the Denali fault system, and (2) the Fairweather-Queen Charlotte Islands fault system. Also, of major tectonic importance is the Totschunda fault system, which appears to connect with the Denali fault system, and the Chugach-St. Elias fault, which joins the northwestern end of the Fairweather fault. These fault systems, as well as inferred connections between fault segments are shown in figures 3 and 5. Other faults are described in our separate reports of coastal communities in southeastern Alaska.

As first reported by St. Amand (1954) and by Sainsbury and Twenhofel (1954), and later described in more detail by St. Amand (1957), and Twenhofel and Sainsbury (1958), the "Denali fault" is a great arcuate series of related faults about 1,300 miles long. The "fault" was described as extending from the Bering Sea across the northern flank of Mount McKinley, through northernmost British Columbia, and thence down the Chilkat River valley and Chilkat Inlet to Lynn Canal and Chatham Strait (fig. 3). Thus, in Canada and southeastern Alaska, it included the Shakwak valley, Chilkat River, Chatham Strait, and Lynn Canal faults (fig. 5). Grantz (1966) called this series of related faults the Denali fault system but restricted the Denali fault itself to that part of the

fault system that extends from the Kuskokwim River drainage east of Bethel to northernmost British Columbia (a length of about 870 miles).

Some doubt was expressed by Hamilton and Myers (1966) on the continuity of the Denali fault into British Columbia and southeastern Alaska, as described by St. Amand (1957) and others. Instead, they suggested that the Denali fault system extends southeastward along a lineament that Richter and Matson (1971) named the Totschunda fault system. Furthermore, they noted that this fault is alined with the Fairweather fault to the southeast, and they assumed that the two faults connect. Richter and Matson (1971) left open the question of whether or not there is a connection between the Denali and the Totschunda fault systems [4] and only suggested that the Totschunda fault system might connect with the Fairweather fault. As discussed below, Plafker (1971; written commun., 1971) does not believe that the Totschunda fault system is a continuation of the Fairweather fault.

The Fairweather fault and the Queen Charlotte Islands fault probably are a part of the same tectonic element (St. Amand, 1957; Grantz, 1966; Tobin and Sykes, 1968; Page, 1969; George Plafker, written commun., 1971; and Richter and Matson, 1971). The onland part of the Fairweather fault is a segment about 125 miles long extending southeastward from Yakutat Bay to Icy Point (figs. 3 and 5). The offshore southeastern extension of the fault follows the continental slope off southeastern Alaska and probably joins the Queen Charlotte Islands fault off the coast of British Columbia. The Queen Charlotte Islands fault was inferred on the basis of the configuration of offshore topography and the presence of a belt of high seismicity (Menard and Dietz, 1951; St. Amand, 1957; and Wilson, 1965). As mapped by Plafker (1969; 1971), the northwestern end of the Fairweather fault joins the eastern end of the Chugach-St. Elias thrust fault (figs. 3 and 5).

These differences in interpretation as to the true trend of the Denali fault system, as well as possible connections between the Denali and Fairweather-Queen Charlotte Islands fault systems cannot be resolved by us. However, for purposes of discussion here, the Shakwak valley fault, Chilkat River fault, Lynn Canal fault, and the Chatham Strait fault (fig. 5) will be included as probably an older part of the Denali fault system and the Totschunda fault system as a younger part of that system. The Fairweather-Queen Charlotte Islands fault system will be considered to be more or less separate from the Denali fault system but one which merges on its northwest end with the Chugach-St. Elias fault.

The variable accuracy of locating earthquake epicenters, particularly during the 19th century and early 20th century, plus incomplete knowledge of fault locations make it difficult to directly relate

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[4] Field studies by Richter and Matson in 1971 indicate that the Denali and Totschunda fault systems do connect (N. A. Matson, Jr., written commun., 1971).

seismicity in southeastern Alaska to known and inferred faults. In spite of these difficulties, most of the larger and many of the smaller earthquakes probably can be related to faults shown in figures 3 and 5.

Denali fault system

The Denali fault system, as indicated above, is a profound structural feature, but kinds and amounts of movement are still incompletely known. Wright and Wright (1908, p. 40) were the first to suggest that the straightness of Lynn Canal and Chatham Strait (fig. 5) was controlled by a fault coincident to the long axes of these fiords. They believed this fault to be the largest in southeast Alaska and its presence was indicated both by geology and by topography. A length of more than 200 miles was assigned to the fault with the main Lynn Canal-Chatham Strait fault branching near the head of Lynn Canal (fig. 5). One of these branches (the Chilkat fault) presumably extends up Chilkat Inlet, and the other branch (the Chilkoot fault) presumably extends up Chilkoot Inlet (St. Amand, 1957; Twenhofel and Sainsbury, 1958; Brew and others, 1966; Grantz, 1966).

Most investigators believe that the principal movement on the Denali fault system has been horizontal although Muller (1967), from his work along the Shakwak segment (fig. 5) in Canada, concluded that movement on that segment has been chiefly vertical. St. Amand (1957) suggested, on the basis of indicated offset of glacial river valleys and on seismic evidence, that the fault is a right-lateral strike-slip fault with as much as 150 miles of strike-slip movement along some segments. Grantz (1966), however, pointed out that this amount of right-lateral offset forms as many topographic anomalies as it solves. Richter and Matson (1971) believe that strike-slip faulting has occurred since early Pliocene time (latest epoch of the Tertiary Period) and that movement during Holocene time on that part of the Denali fault system west of the Totschunda fault exceeds 2 cm per year. Twenhofel and Sainsbury (1958) stated that available geologic data indicate that a large amount of vertical movement took place, and they suggested that the west side of the Chatham Strait fault was upthrown but that the data neither support nor deny St. Amand's hypothesis of major strike-slip movement. However, on the basis of much additional mapping by a number of workers, Lathram (1964) supported the opinion that right-lateral separation of about 120 miles along the Chatham Strait fault is suggested by apparent displacement of major geologic features on opposite sides of the fault. As further pointed out by Lathram (1964), the Chatham Strait fault would line up with the Chilkat River fault although some problems in alining correlative bedrock units would arise. On the other hand, Brew, Loney, and Muffler (1966), on the basis of the realinement of facies of the major sedimentation units, suggest a right-lateral separation of about 50 miles on the Chatham Strait fault since Cretaceous time or earlier but with at least some of the movement taking place during the Cenozoic. Stratigraphic evidence and radiometric ages also indicate that Chichagof Island and Baranof Island (fig. 1) on the west side of the fault have been uplifted in respect to Admiralty Island on the east side of the fault. Baranof Island and possibly Chichagof Island have been uplifted several kilometers

since Miocene time whereas Admiralty Island has been relatively stable tectonically since Paleocene time (Loney and others, 1967).

As indicated by the location of earthquake epicenters, historical movement on the Denali fault system, other than possibly creep or other minor movement, has been relatively small. As shown in figure 4, and discussed in more detail in a report on the Haines area (Lemke and Yehle, 1972), the number of earthquakes related to the Shakwak Valley, Chilkat River, Lynn Canal, and Chatham Strait segments of the Denali fault system (those segments of most significance in evaluating movement on the fault system in southeastern Alaska) is small as compared to the number along the Fairweather and Queen Charlotte Islands faults. The reason for the relatively low seismicity on these fault segments may be due to the fact that this part of the Denali fault system may have been rendered relatively inactive in favor of movement along the newer Totschunda fault system. As suggested by Richter and Matson (1971), the Totschunda fault system may represent part of a new transform fault segment of the Denali fault system, which bypasses the Denali fault system in Canada and southeastern Alaska and leaves it as a more passive tectonic segment. However, Bostock (1952) believed that there might have been movement on a part of the Shakwak fault in Canada in recent centuries. Moreover, that this part of the Denali fault system is not everywhere currently passive is evident from the number of earthquake epicenters that appear to be related to segments of it. Also, Boucher and Fitch (1969, p. 6648) concluded, on the basis of their microearthquake studies, that "the Denali fault is active in some sense along its entire length east of Mount McKinley * * * and it probably should not be dismissed as a relic fault of no current tectonic importance."[5] The high microearthquake activity, they suggest, may document the occurrence of very large earthquakes in the past. A possible alternate explanation, given by them, is that the microearthquakes are signatures of some kind of background seismicity such as that associated with a creep phenomenon.

Fairweather-Queen Charlotte Islands fault system

As shown on figures 3 and 5, the Fairweather fault and the Queen Charlotte Islands fault are believed to be part of the same tectonic element (St. Amand, 1957; Grantz, 1966; Tobin and Sykes, 1968; Richter and Matson, 1971). However, each segment generally has been described more or less separately in the literature and this method will be followed here. It should be emphasized that the combined segments constitute a major tectonic element of western North America, especially

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[5]The term active fault is in common usage in the literature, but there is no general agreement as to the meaning of the term in relation to time. In general, an active fault is one on which continuous or, more likely, intermittent movement is occurring. As used in this report, an active fault is defined as one that has displaced the ground surface during Holocene time.

when considering southern and northern connections with other major fault systems.

All of the large and many of the moderate and smaller historical earthquakes in southeastern Alaska and adjacent areas appear to be related to the Fairweather-Queen Charlotte Islands fault system and the connecting Chugach-St. Elias fault or to their branches (fig. 5). Thus, most have epicenters on land close to the coast or offshore in the Gulf of Alaska and in the northern Pacific Ocean.

Fairweather fault.--The best known part of the Fairweather fault is a segment about 125 miles long extending southeastward from Yakutat Bay to Icy Point (figs. 3 and 5). Here, the fault lies largely in a linear valley partly filled by glaciers and separating crystalline rocks of the Fairweather Range from partly younger and less altered rocks of the coastal region (D. J. Miller, 1960). The block northeast of the fault has been uplifted more than 5 km (3.1 miles) and the fault has undergone right-lateral slip of unknown magnitude from late Pliocene or early Pleistocene to Holocene time (Grantz, 1966). The trace of the fault across a transverse ridge that forms North Dome (about 12 miles southeast of Lituya Bay) indicates that the fault at that location is vertical or dips steeply to the northeast (D. J. Miller, 1960).

Mapping in the Yakutat area (Plafker, 1969, 1971; written commun., 1971) indicates that the northwest end of the Fairweather fault merges with the east-west-trending Chugach-St. Elias fault rather than extending northwestward to join the Totschunda fault system. George Plafker (written commun., 1971) states " * * * our data preclude the possibility that the Fairweather fault or a major splay of it, extends northwestward along the Artlewis Glacier to link up with the Totschunda fault system. Strike-slip motion along such a fault for a million years or so at 2 cm per yr would have offset the contacts between Mesozoic rock sequences more than a kilometer; yet no such offsets are found. Thus, if a strike-slip fault does indeed extend up the Artlewis Glacier, it cannot possibly have been active for more than a few hundred thousand years even at extremely modest strain rates." The southeastern offshore extension of the fault follows the continental slope off southeastern Alaska and apparently connects with the Queen Charlotte Islands transform fault system (St. Amand, 1957; Grantz, 1966; Tobin and Sykes, 1968; Page, 1969; George Plafker, written commun., 1971). It appears to intersect the Lynn Canal-Chatham Strait fault system off the southeast coast of Baranof Island (Grantz, 1966).

The onland segment of the Fairweather fault, as well as probably its western extension, the Chugach-St. Elias fault (fig. 3), has been very active tectonically during Quaternary time (Grantz, 1966; Page, 1969; George Plafker, written commun., 1971). The epicenter of the great Yakutat earthquake of September 10, 1899 (estimated magnitude 8.6), was not accurately located but is believed to have been near the head of Yakutat Bay where there was movement on portions of the Fairweather fault

or on one of its western extensions (Tarr and Martin, 1912; Plafker, written commun., 1971). Likewise, during the Lituya Bay earthquake of 1958 (magnitude approximately 8.0), there was movement along the entire onland length of the Fairweather fault (George Plafker, written commun., 1971) with 21½ feet of right-lateral slip and 3½ feet of associated dip-slip (up on the south) measured in one place (Tocher and Miller, 1959).

That the southeastern offshore extension of the Fairweather fault, which is believed to connect with the Queen Charlotte Islands fault, is also active is indicated by the fairly large number of earthquake epicenters in that area (figs. 4 and 5). Although the assigned epicentral locations are not well alined, probably most of the earthquakes are related to movement along the Fairweather fault. Lack of alinement can be explained by inaccurately located epicenters or by the epicenters being along more than one branch of the fault system. That there is branching and splaying along at least part of the fault system is verified by the fault pattern on land in the Yakutat Bay-Lituya Bay area (Grantz, 1966; Plafker, 1969, 1971).

Queen Charlotte Islands fault.--This fault, named by Wilson (1965), was first inferred on the basis of the narrowness of the continental shelf and the steepness of its scarp off the west coast of the Queen Charlotte Islands, the presence of a closed depression lying parallel to the undersea scarp, and the straightness of the west coast of the islands themselves (Menard and Dietz, 1951; St. Amand, 1957). The southern extension of the fault is inferred to be some distance offshore from the western coast of Vancouver Island. As discussed by Wilson (1965) and by Tobin and Sykes (1968), and supported by Richter and Matson (1971), the fault is interpreted to be a major transform fault related to the ocean-ridge system and to other transform faults to the south that include the Juan de Fuca ridge, the Blanco fracture zone, the Gordo ridge, the Mendocino fracture zone, and the San Andreas fault.

High seismicity along the entire length of the concealed Queen Charlotte Islands fault (as far south as Vancouver Island) is well documented by the large number of earthquakes that appear to be related to the fault. These earthquakes have ranged in size from the large earthquake (magnitude 8.1) of August 22, 1949 (L in fig. 4), through several earthquakes of magnitude 6 to 7, to numerous earthquakes of smaller magnitude. Here, also, the epicenters are not well alined but they do fall, nevertheless, along a fairly definite offshore northwest-southeast belt that strongly suggests a relation to an active fault zone (Gutenberg and Richter, 1954; St. Amand, 1957; Wilson, 1965; Tobin and Sykes, 1968).

Related major faults

As discussed above, the Chugach-St. Elias fault is related to the Fairweather fault, and the Totschunda fault system probably connects with the Denali fault system. There are numerous other connecting faults,

many of considerable length, in southeastern Alaska (fig. 5). Those that relate to the structure of specific coastal community areas will be discussed in our separate reports of coastal communities and, therefore, are not discussed further here.

Chugach-St. Elias fault.--During late Tertiary or early Pleistocene time, the Chugach-St. Elias Mountains were uplifted along a north-dipping fault system, and the sedimentary rocks to the south were displaced along many high-angle thrust faults. The largest of these thrust faults is the Chugach-St. Elias fault, which dips 30°-60° N. and is estimated to have a stratigraphic throw of not less than 10,000 feet (D. J. Miller and others, 1959). It has been traced from Yakutat Bay to the delta of the Copper River (fig. 3), a distance of approximately 180 miles. Surface mapping in the Yakutat Bay area indicates that it merges with the Fairweather fault at the head of Yakutat Bay (Plafker, 1969, 1971). Thus, George Plafker (written commun., 1971) suggests that "The northwestward motion of the crustal block along the seaward side of the Fairweather fault seems to be taken up by relative oblique underthrusting beneath the Chugach-St. Elias fault system. By implication, therefore, these oblique thrust faults must be active, although it has not been possible to find evidence for recent fault scarps in the rugged alpine terrain traversed by the Chugach-St. Elias fault." In support of the supposition that the Chugach-St. Elias and other thrust faults in the area are probably active, all three earthquakes during 1899 and 1900 of magnitude 8 or greater, as well as many smaller ones since then, have been centered in the general area of these thrust faults (fig. 4). However, because the epicenters of the earlier earthquakes probably were poorly located and the faults are fairly close together, it is not possible to relate them to a specific fault.

Totschunda fault system.--This tectonic element, observed on aerial photographs by Hamilton and Myers (1966), was studied first in the field and named by Richter and Matson (1971). Richter and Matson are of the opinion that this is a major structural feature composed of a number of very active discrete segments having a known length of more than 90 miles and with the principal movement being right-lateral strike-slip. They have named the three largest individual faults in the fault system the Totschunda (northwesternmost), Cooper Pass, and Cross Creek faults. The faults overlap, branch, and are broadly curving. The northwest end of the fault system starts in a zone of high-angle reverse faults and tight folds within a few miles of, and parallel to, the Denali fault (fig. 3). Field studies during 1971 indicate that these two fault systems are connected (N. A. Matson, Jr., written commun., 1971).

Although the overall topographic expression is more subdued than for the Denali fault, detailed topographic expression delineating small scarps, sag ponds, landslides, offset glacial deposits, and drainage, are more apparent. The freshness of these features suggest to Richter and Matson (1971) that the Totschunda fault system is younger than the Denali system and could have been developed since the beginning of the

Pleistocene, or at the earliest, late Pliocene time. Movement during Holocene time is well defined as evidenced by the presence of small scarps, offset deposits of the last glacial stage of the Pleistocene, and offset young streams. Based upon amounts of horizontal displacement of Holocene deposits, the average rate of horizontal movement during Holocene time is 1.6 cm a year. About the only evidence of very recent movement is the presence of active landslides along the fault trace. Few earthquake epicenters appear to be related to the mapped part of the fault system. However, several earthquake epicenters are present in Canada more or less along the projected southeastward extension of the fault. This part of Canada has not been studied yet to establish whether a major fault is present. As noted previously, if the Totschunda fault system is projected southeastward across Canada, it is in direct alinement with the northwestward projection of the Fairweather fault. Richter and Matson (1971) state that "The Totschunda-Fairweather alignment probably represents the beginning of a new transform fault by-passing the southeast part of the Denali fault system." A different viewpoint is expressed by George Plafker (written commun., 1971), who states "I suspect that the Fairweather and the Totschunda fault systems occur along the margin of the Pacific Plate where it underthrusts Alaska and thus can be considered as giant tear faults. They are not true transform faults, or plate boundaries, because they do not take up all the differential motion between the Pacific and American plates in this region."

Consideration of other tectonic factors

Most larger earthquakes, as discussed above, are related to major throughgoing faults that have had extensive displacements during Quaternary time. However, other related tectonic factors must be considered in any assessment of the earthquake probability of a region. Some of these factors, which may help in an evaluation of earthquake probability in southeastern Alaska are: (1) faults that long have been inactive may suddenly become reactivated, (2) faults that were active during Quaternary time may have been inactive during most of historical time but suddenly become reactivated, (3) certain presently inactive segments of otherwise active fault systems can be expected to become active in the future, (4) the occurrence of small earthquakes is not necessarily an indication of where large earthquakes may occur, or vice versa, and (5) large earthquakes may occur in areas where there is little or no record of seismicity and no obvious tectonic structure that would result in an earthquake.

When consideration is given to long inactive faults that have suddenly become reactivated, the White Wolf fault in southern California commonly is cited as an example. This fault, considered a "dead" fault on the 1922 fault map of California (Allen and others, 1965), was the locus for the destructive earthquake of 1952 in Kern County, California. The earthquake had a magnitude of 7.7 and a local intensity of VIII (Modified Mercalli scale), 10 to 12 people were killed, and damage was estimated at nearly $40 million. In a strong aftershock (magnitude 5.5)

on August 22, two additional people were killed, 35 were injured, and structures weakened by the first quake were further wrecked by the aftershock and had to be torn down. Certain segments of the Denali fault system, such as parts of the Chatham Strait and Lynn Canal faults, which have been relatively inactive for a long time, might some day be examples similar to the White Wolf fault.

Faults that were active in Quaternary time may have been largely inactive during most of historic time but during some particular historic period were highly active. For example, the seismicity of the Owens Valley area in California has been relatively low during a 29-year period (1934-63) of study of the seismicity of southern California by Allen, St. Amand, Richter, and Nordquist (1965). However, probably the largest earthquake in the history of California centered in this area in 1872; fault scarps are still surprisingly fresh after 100 years. The Totschunda fault system, which has been active throughout the Quaternary, except that seismicity has been low along the fault during historical time, might be an example where activity might occur again at any time.

That presently inactive segments of otherwise active fault systems may become active again in the future is evident from the past historic and geologic record. Thus, it can be exceedingly hazardous to extrapolate long-term activity from short-term records. Three examples are given to illustrate this point.

Allen, St. Amand, Richter, and Nordquist (1965) cite an area of about 8,400 km^2 (210 km long and 40 km wide) cut by the San Andreas fault northwest of San Bernardino in southern California. On the basis of earthquakes recorded during a 29-year period (1934-63), this area has been one of the most seismically inactive areas in southern California, with an extrapolated recurrence rate for an earthquake of magnitude 8 or greater being only once in 18,300 years. As pointed out in the study cited, this recurrence rate is grossly misleading as is evident by the fact that the great Fort Tejon earthquake of 1857 (shaking comparable to the San Francisco earthquake of 1906 and a possible magnitude of 8 or greater) was centered in this area. Also, there is no doubt that numerous other great earthquakes have occurred here in the recent geologic past as evidenced by the abundance of relatively fresh fault scarps.

A second example is the expectable recurrence rate for earthquakes in the Los Angeles Basin area based upon short-term records. The Long Beach earthquake of March 11, 1933 (magnitude 6.3), on the basis of these records, would qualify as a "one-hundred-year earthquake" and the recurrence rate for an earthquake of magnitude 8 or greater would be 3,740 years (Allen and others, 1965). However, the San Fernando earthquake of February 9, 1971, in this area had a magnitude of 6.6 and, therefore, there have been two "one-hundred-year" earthquakes in the area in the past 38 years. Also, it is noteworthy that strain-release maps (Allen and others, 1965) indicate that the northern part of the San Fernando Valley, where the epicenter was located, was seismically less active

than other parts of the greater Los Angeles area. Moreover, the epicenter for the San Fernando earthquake, as shown by Wentworth and Yerkes (1971) is less than 10 miles from the southern edge of the San Andreas area (Allen and others, 1965) where, on the basis of the 29-year record, only one earthquake of magnitude 5.6 would be expected to occur in 100 years.

Another example was the great Chilean earthquake of 1960 (approximate magnitude 8.5), which was centered in an area of low seismicity on the basis of the 1904-52 period of observation. People interviewed in the Province of Cauten (southern Chile), one of the areas most heavily affected with aftershocks, had never before felt an earthquake (Allen and others, 1965). However, the fact that great earthquakes in 1575 and in 1835 preceded the 1960 earthquake in the region illustrates that a relatively short period of quiescence is, in itself, no assurance of the seismic inactivity of the region.

In applying the above examples to southeastern Alaska and adjacent areas, there are several areas that might be comparable. The area of presently low seismicity on the Fairweather-Queen Charlotte Islands fault system between approximate latitudes 55°N. and 56°N. might be one such area. Certain segments of the Shakwak valley fault also may be an example. Because of the very short seismic record, other faults exhibiting fresh scarps perhaps should be considered more active than is indicated solely from the historical record.

The occurrence of small earthquakes in an area is not necessarily an indication of where to expect large earthquakes, or vice versa. Most large earthquakes have their epicenters along large preexisting throughgoing faults. Smaller earthquakes, particularly aftershocks from large earthquakes, commonly are much more randomly distributed. The distribution of aftershocks from the great Chilean earthquake of 1960, if superposed on the map of California, would entirely encompass that State (Allen and others, 1965). Likewise, the aftershocks of the great Alaska earthquake of 1964 were distributed over a very large area, so a direct relation to tectonic structure is not obvious.

There is considerable disagreement at present whether small earthquakes or creep along a fault can act as "safety valves" in releasing sufficient stress to prevent large earthquakes. As indicated by Allen, St. Amand, Richter, and Nordquist (1965), cohesion across a fault may be so great that accumulating strain cannot be relieved by creep or small earthquakes and can be released only by a future large earthquake. Even if creep or small earthquakes do act to some degree as "safety valves," they may not occur often enough as shown by the fact that large earthquakes keep happening (Richter, 1964). Thus, the relatively high microearthquake [6] activity that was recorded along the Denali

[6] Microearthquakes are earthquakes that generally are too small to be felt by man and have to be detected instrumentally. The limit of magnitude of felt earthquakes generally is between 2 and 3; many microearthquakes, on the other hand, have magnitudes of less than 1.

fault system between Haines and Mount McKinley may not be releasing sufficient strain to preclude the triggering of a large earthquake along the fault segments. It also should be remembered that it takes approximately 1,000 earthquakes of magnitude 5 to release the same strain as one earthquake of magnitude 7 (Environmental Sci. Services Adm., 1967). On the other hand, when a great earthquake does occur, it releases sufficient strain along the newly reactivated segment of the fault to greatly lessen the recurrence probability of another great earthquake in the near future on that particular segment of the fault. This does not, however, lessen the probability of a large earthquake along other segments of the same fault where the strain has not been released. Thus, a quiescent segment of an otherwise seismically active fault system may well be regarded with more concern than a segment of the fault upon which a large earthquake has just occurred. It also should be borne in mind that large earthquakes, closely spaced in time, have been successively destructive to the same area. They may or may not be on the same fault system but, because of the great magnitude of the earthquakes, intensities are high over a large region. The three earthquakes in the Yakutat Bay area during the period 1899-1900, all with magnitudes greater than 8, are good examples of large earthquakes that have occurred in a fairly limited area in a short space of time.

That large earthquakes may occur in areas with little or no previous record of seismicity and where there are no obviously active tectonic structures that might cause a large earthquake is well illustrated by the great Charleston, South Carolina, earthquake of 1886. This earthquake, with a maximum intensity of XI, strongly shook the surrounding area for a distance of 100 miles and was felt over an area of 2 million square miles--as far away as Chicago, Boston, and Cuba (Dutton, 1889). Its magnitude is not known but it probably was one of the largest historical earthquakes to have shaken the United States. Sixty people were killed and damage to structures was heavy even though the area was sparsely settled. Previous to the earthquake, the Charleston area had no historical record of earthquakes, except for light tremors that began a couple of months before the main shock.

Although the epicentral area of the great Alaska earthquake (magnitude 8.5) of March 27, 1964, had not been completely quiescent in the years prior to 1964, no earthquakes of magnitude 8 or greater had occurred in that area in historic time. Moderate activity had been concentrated in a belt to the northwest.

The Koyna, India, earthquake of December 11, 1967, is another example of a moderate to large earthquake occurring in a supposedly geologically stable and earthquake-free area. This earthquake, of a magnitude variously calculated between 6.2 and 7.5 (Mathur, 1967; Narain and Gupta, 1968), struck the Koyna Dam area causing severe damage to manmade structures and heavy loss of life. The epicenter was located 5 km from the dam on the Deccan Plateau, traditionally regarded as a very stable mass of great rigidity and placed in seismic zone 0 (nonseismic) by a committee

of experts, which prepared a seismic map of India (Mathur, 1967; Science and Technology, 1968). Some geologists have attributed the earthquake to increased load caused by impoundment of 2.2 million acre-feet of water in the reservoir behind the dam (Engineering News Record, 1967; Narain and Gupta, 1968). Whether or not this was the cause, the fact remains that large damaging earthquakes can occur where there are no prior indications for such an event.

There are a number of areas in southeastern Alaska and vicinity that, except for microearthquake activity (Whitham, 1969), have been seismically quiet during historic time and probably for much longer but which cannot be ruled out as possibly the loci of future large earthquakes. For example, the area inland from the coast, extending from southeast of Prince Rupert to northwest of Juneau (fig. 1) may not be as seismically inactive in the future as the present record indicates.

Earthquake probability for southeastern Alaska and adjacent Canada

As discussed previously, both the historical seismicity and the geologic conditions, such as frequency and recency of faulting, must be considered together to assess the future earthquake probability of an area. When adequate data on these two factors are available, seismic regionalization maps, more appropriately called seismotectonic maps (Stepp, 1966) can be constructed. These maps show the geology of an area in related intensities combined with data on the seismicity of the area. These maps, in essence, are earthquake-probability or earthquake-risk maps. Unfortunately, lack of accurate data on the historical seismicity and tectonic movement of the region, and a lack of full understanding of earthquake mechanism in general, commonly makes it difficult, if not impossible, to fully assess all the factors that must be considered in making an earthquake probability evaluation. For these reasons, all attempts up to the present time to show earthquake probability in any particular part of southeastern Alaska must be regarded as approximations.

In spite of the difficulties mentioned above, some data are available for southeastern Alaska that permit generalizations on earthquake probability and consequent seismic risk. A considerable amount of this information has been compiled in the form of strain release and seismic zone maps, and is described below.

Map sources of information

Strain-release maps.--One way of expressing the seismicity of a region is to show the release of accumulated energy from instrumented earthquakes in the region by means of strain-release maps. This type of map shows the amount of strain that is released by means of contours drawn through points of equal strain release. As discussed by Allen, St. Amand, Richter, and Nordquist (1965), there are a number of ways of expressing strain release. For southeastern Alaska and adjacent parts

of Canada, strain-release maps have been prepared by St. Amand (1957) and by Milne (1967). Sources of data and methods of presentation are different for the two maps and, thus, the data presented are difficult to compare. The strain-release map prepared by St. Amand (1957) represents the strain released by earthquakes between the years 1898 and 1955, expressed in equivalent earthquakes of magnitude 4 per square degree of latitude and longitude per year. Because of a number of arbitrary assumptions that had to be made, St. Amand (1957) made no claim for the precision of his map. A strain-release map prepared by Milne (1967), which followed the methods of Benioff (1951) and also had the benefit of computerized calculations of earthquake data, shows the strain released by earthquakes for the period 1898-1960, expressed in equivalent earthquakes of magnitude 5 per unit area of 100X100 km using a minimum magnitude 3.7 earthquake as a zero contour line.

Although prepared in different ways, the strain-release maps by St. Amand and by Milne show similar strain-release patterns. The main difference between the two maps is that Milne shows a considerably greater strain release in the general area of Lituya Bay and the Fairweather Range, probably because he incorporated the strain release from the Lituya Bay earthquake of 1958. This earthquake postdated St. Amand's map. Another difference is that Milne shows a greater strain release in the Haines-Skagway area, presumably because of using data from more epicenters in that area.

Figure 6 is a strain-release map of southeastern Alaska and northwestern Canada as modified from Milne (1967). The largest strain release in the region, according to Milne, is in the Yakutat area owing mainly to the large Yakutat earthquakes of 1899-1900. For that area, Milne gives a maximum strain-release value of 700[7]. Therefore, on the basis of mathematical frequency, the number of earthquakes necessary to release the energy in the Yakutat area in 100 years is: 700 earthquakes of magnitude 5, or 120 earthquakes of magnitude 6, or 21 earthquakes of magnitude 7, or 3.5 earthquakes of magnitude 8, or 1 earthquake of magnitude 8.7 (fig. 6). If one were to assume that the strain release for the next 100 years will be the same as for the period 1898-1960, then the future earthquake probability for the Yakutat area would range from 700 earthquakes of magnitude 5 to one earthquake of magnitude 8.7. However, as will be discussed later, the frequency of earthquakes in the past in a particular area is not necessarily an indication of the number or size of expectable earthquakes in the future.

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[7] The contour to depict a strain-release value of 700 is not shown on Milne's map (1967) but the value is given in his report. The greatest strain-release value shown on Milne's map is 100, depicted by contour value 6. Therefore, three additional contours (7, 8, and 9) would need to be added to Milne's map to show the strain-release value of 700 in the Yakutat area. See Richter (1958) for method of calculating strain-release values.

According to figure 6, the Queen Charlotte Islands area has the second greatest strain release. This is due in large part to the Queen Charlotte Islands earthquake of 1949, which had a magnitude of 8.1. The maximum strain value for this area is 236 (Milne, 1967), which is equivalent to the energy release of one earthquake of magnitude greater than 8 in 100 years.

A second kind of map, based upon amounts of strain release, was prepared by Milne and Davenport (1969) to indicate peak earthquake accelerations per 100-year interval. This map is reproduced in part as our figure 7. Contours on the map show peak acceleration as a percent of gravity. Approximate relations between acceleration as a percent of gravity and magnitude and intensity are shown in table 1.

Seismic-zone maps.--A seismic probability map dividing Alaska into four seismic zones (0, 1, 2, and 3) was prepared in 1957 by the U.S. Army, Corps of Engineers, Alaska District (written commun., 1968). As noted on the map, it was adapted from a 1950 map prepared by the U.S. Coast and Geodetic Survey using data furnished by the U.S. Geological Survey. Although no source was given, an apparently identical seismic probability map for Alaska was published by the U.S. Departments of Army, Navy, and Air Force (1966) in connection with the seismic design of buildings. The map also corresponds to the seismic probability map in the Environmental Atlas of Alaska, as shown in Alaska Industry (1970).

In 1965 the U.S. Army, Corps of Engineers, Alaska District, revised their 1957 seismic probability map with the assistance of Commander Short, formerly with the U.S. Coast and Geodetic Survey (Warren George, written commun., 1968). This map differs from the 1957 version by substantially widening the land area of zone 3 (expected major earthquake damage) and decreasing the width of zone 2 (expected moderate earthquake damage). The net effect of this map is to place nearly all of southeastern Alaska into zone 3 and the remainder into zone 2. As a result, it shows a higher probability of earthquake risk and resulting damage from larger earthquakes in southeastern Alaska. Our figure 8, taken from the 1965 map of the Corps of Engineers, differs from that map only in the change of zone 0[8] in northern Alaska to zone 1. This change was made as a result of discussions in 1971 between Ernest Dobrovolny of the U.S. Geological Survey and members of the U.S. Army, Corps of Engineers, Alaska District (Ernest Dobrovolny, written commun., 1971). Also, the discussion indicated that further revisions of the map are planned by the Corps of Engineers, all tending to increase the areal extent of zones 3 and 2 in the general vicinity of the Alaska-Yukon boundary north of Yakutat Bay.

A seismic-zone map of Alaska, modified from the 1970 edition of the Uniform Building Code (International Conference of Building Officials,

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[8] No structural damage is expected from earthquakes in zone 0 and magnitudes are expected to be less than 3.

1970), is shown by us as figure 9. The southeastern Alaska part of this map appears to be virtually the same as the 1957 map of the Corps of Engineers. It is not clear to us how this map was derived and what factors were assessed in its compilation.

A new seismic zone map of Canada (Natl. Research Council Canada, 1970) shows all of the coastal region of British Columbia, the northwestern part of British Columbia, and most of the Yukon territory as being in zone 3 (fig. 10). On the basis of this map all of southeastern Alaska also would be in zone 3. The zones on the map are similar to those prepared in the United States for evaluating earthquake risk. They range from zone 0, where no damage is expected, to zone 3 where major destructive earthquakes may occur (Hasegawa, 1971). In zone 3 the estimated maximum intensity, as measured on the Modified Mercalli scale, is between VIII and IX, and the corresponding horizontal ground acceleration may be taken as 50 percent of gravity (Ferahian, 1970).

Summary evaluation of earthquake probability for southeastern Alaska

It is obvious that agreement is not close between authors of the maps described above in their evaluations of the earthquake probability in southeastern Alaska. At least part of this divergence is because some of the maps, such as the strain release map of Milne (1967) and the earthquake acceleration map of Milne and Davenport (1969), are based solely upon historical seismicity and, therefore, are not true probability maps. Differences in the assessment of seismic risk, based upon the Corps of Engineers map of 1965, the Uniform Building Code map of 1970, and the new seismic zone map of Canada may be more apparent than real because the three maps are not assessing all the same factors. The Uniform Building Code map sets up only minimum standards to be met by industry. The Corps of Engineers map of 1965, on the other hand, attempts to assess the overall earthquake probability of the area and, thus, expectedly would show a higher risk value. The zonal values for the new seismic zone map of Canada are similar to those that have been prepared for seismic zone maps in the United States, but apparently they are not identical because of somewhat different derivative factors.

The differences in methods of map evaluation, plus the great difficulty of assessing a large number of poorly known pertinent tectonic factors, makes it impossible to evaluate the earthquake probability of southeastern Alaska with any degree of accuracy. We tentatively conclude, however, that most if not all of southeastern Alaska should be placed in seismic zone 3, a zone where earthquakes of magnitude greater than 6 will occur from time to time and where there may be major damage to manmade structures. Also, on the basis of present knowledge, the largest earthquakes will occur most frequently near the coast or offshore, along a segment or extension of the active Fairweather-Queen Charlotte Islands fault system. In this seismic belt, it has already been noted that five earthquakes of magnitude 8 or greater have occurred since 1899. Farther inland, earthquakes of magnitude 8 or greater appear to be much

more unlikely but, for reasons discussed previously, cannot be ruled out. Thus, for purposes of discussions that are to follow, as well as for broad guidelines for our reports of specific coastal community areas in southeastern Alaska, figure 8 will form the main basis for our evaluation of seismic risk.

POSSIBLE EFFECTS OF FUTURE EARTHQUAKES INFERRED FROM PAST EARTHQUAKES

Introduction

Because of the reconnaissance nature of our studies in the coastal community areas of southeastern Alaska and the sparsity of laborabory data on physical properties of geologic units in each area studied, the inferences regarding geologic effects of future earthquakes must of necessity be largely empirical and generalized and must be based chiefly upon the effects of past major earthquakes on similar-appearing geologic units, in Alaska and elsewhere. The well-documented effects of the Alaskan earthquake of March 27, 1964, provide the basis for many of those inferences.

It should be emphasized, however, that the properties of the geologic units in a particular study area in southeastern Alaska may not be the same as those being compared from a past earthquake area, even though they superficially may resemble them. Moreover, the physical setting and other factors, which could markedly influence the effects, also may differ. For these reasons, our inferences on future earthquake effects should not be rigorously interpreted or applied. Rather, they are intended as broad guidelines useful both in assessing the kind and degree of hazard that may be present and in leading toward minimizing these hazards as they affect man and his structures. As such they are directed to structural and civil engineers, city and regional planners, public and private utility companies, and all other public and private groups or individuals who are responsible for the safety and welfare of the people in southeastern Alaska now and in the future. In no way are they intended to preclude the making of additional detailed geologic and engineering studies, particularly those studies that pertain to small tracts or individual sites. In fact, we recommend that such additional studies be made whenever possible.

As discussed previously, earthquakes greater than magnitude 6 are expected to occur fairly frequently in southeastern Alaska and adjacent areas (fig. 4), and earthquakes of magnitude 8 or greater may occur from time to time. Therefore, the examples of earthquake effects cited below from other areas are mostly from large earthquakes. It should be emphasized, however, that for every large earthquake that may occur, there will be many smaller earthquakes with few to minor effects of the types described here. Therefore the effects described below should generally be considered to be maximum or near maximum. As a guide to the reader in this respect, the magnitude of each earthquake used as an example is given whenever known.

Surface displacement along faults and other tectonic land-level changes

Buildings, highways, bridges, tunnels, harbor facilities, pipelines, canals, and other manmade structures may be severely damaged or destroyed by fault displacement or related tectonic land-level changes. Direct

damage from fault rupture is restricted chiefly to the structures built directly athwart the fault. Damage may result from shearing, compression, extension, or rotation caused by tilting or bending of the structure (Bonilla, 1970). Damage from related tectonic land changes, that is regional uplift, subsidence, or tilt of the land, may affect certain kinds of facilities (particularly waterfront facilities) over wide areas.

In northern California and Nevada, where it has been possible to adequately observe and study surface faulting, surface rupture in some degree almost always accompanies shocks of magnitude 6.5 or greater (Tocher, 1958). On the other hand, known surface faulting in that region is uncommon for shocks of magnitude 6 or less although there have been some exceptions in recent years. One such exception was the Parkfield-Cholame, Calif., earthquake of June 27, 1966, with a magnitude of 5.5, where surface tectonic fractures formed along a fault length of 23½ miles (Brown and others, 1967). The smallest California earthquake with observable fault displacement was the Imperial, Calif., earthquake of March 4, 1966, with a magnitude of 3.6; approximately 6 miles of the Imperial fault showed surface rupture (Brune and Allen, 1967). As pointed out by Bonilla (1967), with present-day techniques and more detailed field examinations, surface faulting probably will be found to accompany many future low-magnitude earthquakes.

Tables and curves showing interrelations between length of surface rupture, maximum surface displacement, and magnitude of associated earthquake have been compiled for historic surface faulting in the United States and adjacent parts of Mexico (Bonilla, 1967, 1970) and for worldwide historic faulting (Bonilla and Buchanan, 1971). Data from these sources show that the longest surface rupture (partly submarine) that is known occurred along 270 miles of the San Andreas fault during the San Francisco earthquake of 1906. Horizontal movements have ranged from 0.05 foot of strike-slip movement during the Imperial, Calif., earthquake of March 4, 1966 (Brune and Allen, 1967), to 29 feet during the Gobi-Altai earthquake (magnitude between 7.7 and 8.6) of 1957 (Florensov and Solonenko, 1963). During the Assam earthquake of 1897 there was as much as 35 feet of vertical displacement (Oldham, 1899), and there may have been as much as 47 feet 4 inches of vertical displacement during the Yakutat Bay, Alaska, earthquake of 1899 (Tarr and Martin, 1912; Bonilla, 1967). Both length of fault rupture and displacement generally increase with greater magnitudes. However, as pointed out by Bonilla (1970), the curves showing these relations should serve only as general guides and probably will need to be revised as more data become available.

The Alaska earthquake of 1964 (magnitude 8.5) dramatically illustrated the wide areal extent and amount of uplift and subsidence that can result from a large earthquake, as well as the corresponding severity of adverse effects. The land was warped over an area of at least 70,000 square miles and possibly over more than 110,000 square miles to constitute a deformed belt nearly 600 miles long and at least 200 miles wide (Plafker, 1969). South of a line extending from the southeast coast of Kodiak Island, and through the western part of Prince William Sound to

near Valdez, and thence eastward to about long. 142° W., the area was elevated; northwest of this line to as far as the Alaska Range (see fig. 3 for locations), the land was lowered. The maximum measured uplift on land was approximately 38 feet on Montague Island; there, the Patton fault relatively displaced blocks of land a maximum of 26 feet, the land on both sides of the fault being upthrown in respect to sea level (Plafker, 1969). Southwest of Montague Island and just north of the offshore extension of the Patton fault, there may have been as much as 50 feet of uplift (Malloy, 1964). Maximum subsidence was along the southwest coast of the Kenai Peninsula, where the land was lowered 7½ feet (Plafker, 1969). The tectonic land-level changes, both uplift and subsidence, caused heavy damage to the works of man and to the flora and fauna of the region. Most docks, piers, and other shipping facilities were placed beyond the reach of boats in areas of maximum uplift, and harbors and channels became too shallow to be navigable. In areas of subsidence, low coastal areas were inundated and many port facilities were rendered useless. Coastal forests, bird-nesting areas, fish-spawning grounds, and shellfish habitats also were adversely affected. Corresponding amounts of uplift or subsidence resulting from a large earthquake in or near southeastern Alaska would have similar adverse effects.

Large-scale land-level changes accompanied the Chilean earthquake of 1960 (magnitude 8.5) over an area in southern Chile more than 600 miles long in a north-south direction and more than 125 miles wide. A large part of the affected land area subsided; the maximum measured subsidence was 8.9 feet (Plafker and Savage, 1970). Large areas of fertile farmland were inundated by the ocean, and low-lying communities near mouths of rivers became subject to seasonal flooding. In the vicinity of the coastal town of Lebu, the land rose approximately 5 feet (St. Amand, 1960) and farther to the south, on the island of Isla Guamblin, uplift was approximately 18.6 feet (Plafker and Savage, 1970). Some coastal facilities were impaired by the uplift but on a minor scale as compared to the effects of the Alaska earthquake of 1964, mainly because of the smaller areal extent of uplift and fewer major facilities in the area. Uplift or subsidence of this degree in southeastern Alaska could have considerable adverse effects to coastal communities.

During the San Francisco earthquake of 1906 (magnitude 8.2) the San Andreas fault underwent a maximum horizontal displacement of 21 feet and a vertical displacement of 3 feet (Richter, 1958). Movement along the fault displaced and damaged buildings, roads, a reservoir, and other facilities. More importantly, the major pipeline carrying water to the city reservoirs of San Francisco was sheared, and exceptionally heavy loss from fire ensued because of lack of water. Damage to reservoirs and to pipelines carrying water into coastal communities in southeastern Alaska could severely affect the inhabitants in the event of a large earthquake.

The greatest vertical displacement on a fault from a single earthquake is generally attributed to movement during the Yakutat Bay, Alaska, earthquake of 1899 (magnitude about 8.6). A maximum uplift of 47 feet

4 inches was measured by Tarr and Martin (1912), determined from differences of land levels on opposing sides of Disenchantment Bay (northern end of Yakutat Bay). More recently, however, Bonilla (1967) has raised the question, on the basis of mapping by Plafker (1967), whether the displacement might represent movement along more than one fault or could be attributed to warping of the land. Regardless of the mechanism, there still has been an uplift effect of 47 feet 4 inches in respect to sea level. That amount of uplift in some coastal communities of southeastern Alaska could have extremely adverse effects.

Ground shaking

The direct effects of ground shaking during an earthquake commonly are difficult to differentiate from secondary effects that result from the shaking, such as landsliding, subsidence, ground fracturing, and water-sediment ejection. This is particularly true because the secondary effects generally are most common in the same type of ground that is susceptible to the greatest shaking.

The amount of ground shaking (intensity) at any one place is dependent upon a great many variables, which include magnitude of the earthquake, distance from the epicenter, acceleration, period, duration of shaking, amplitude of the seismic waves, physical properties of the ground, geologic structure, and ground water (Barosh, 1969). The variable most responsible for the degree of shaking at any epicentral distance, however, is the type of ground.

The pertinent literature describing the relation between type of ground and the degree of expectable shaking during an earthquake is so large that it is not feasible to cite it all here. It comes principally from those countries that have competent personnel in seismology and have seismically active regions (United States, Japan, U.S.S.R., and New Zealand). Two good reference sources for the literature up to 1957 are by Duke (1958a). A more recent comprehensive reference source is Barosh (1969).

Although much remains to be learned and many effects are not clearly understood, it generally has been found that shaking (ground vibration) is considerably greater in poorly consolidated deposits than in hard bedrock. Moreover, the thicker the beds and the finer the grain of these deposits, the more destructive generally is the ground motion, particularly if the deposits are water saturated. Certain types of manmade fill, and alluvium, generally are subject to the strongest shaking. Neumann and Cloud (1955) state that instrumental records show that a rock outcrop, such as granite, and an adjoining area of unconsolidated deposits may have as much as a ten- to fifteen-fold difference in acceleration, and, with certain assumptions, possibly as much as a 22-fold difference (Neumann, 1954). Gutenberg (1957), in comparing records of relative amplitudes of earthquake motion for different types of ground stated that amplitudes in water-saturated soft ground may be 10 times or more those in bedrock. He assigned the following amplitude factors, using

solid rock as 1: sandstone, as much as 3; dry sand, about $3\frac{1}{2}$; and marshy land, 12. Richter (1959) pointed out that alluvial deposits are particularly subject to strong shaking where seismic waves emerge directly from rock into alluvium.

Many examples can be given to demonstrate that manmade fill is subject to strong shaking and heavy damage during an earthquake. H. O. Wood's study (1908) of the San Francisco earthquake of 1906 showed that the most severe damage was on manmade fill, with damage successively less severe on alluvium, conglomerate, shale, and hard rock. Damage was 5 to 10 times greater on fill than on hard rock. Although some of the damage undoubtedly was due to secondary effects, such as ground fracturing and differential settlement, the shaking itself probably was the dominant cause. Greatest damage also has occurred on manmade fill during major earthquakes in eastern Canada, Germany, Chile, Turkey, Greece, Algeria, and China (Duke, 1958a). Likewise, during the great Chilean earthquake of May 22, 1960 (magnitude 8.5), damage to buildings constructed on manmade ground was conspicuously greater than that on any other type of ground (Dobrovolny and Lemke, 1961; Barozzi and Lemke, 1966). During the Alaska earthquake of 1964, damage was especially heavy to highway fills (Kachadoorian, 1968).

Severe shaking of alluvial deposits during a large earthquake and heavy damage to associated manmade structures also is well documented from the records of past earthquakes. Similar to the San Francisco earthquake of 1906, damage to buildings constructed on water-saturated alluvium in the city of Puerto Montt during the Chilean earthquake of 1960 was conspicuously greater than those constructed on all other types of ground except manmade fill (Richard W. Lemke, written commun., 1960). Ground motion in alluvium during the Alaska earthquake of 1964 apparently also was very high in comparison to most other types of deposits. However, the direct effects of shaking were hard to differentiate from heavy damage resulting from secondary effects such as landsliding, lurching, subsidence, fountaining, and ground fracturing. The very fact, though, that the most conspicuous secondary effects commonly were concentrated in alluvial deposits having a high water table strongly suggests that strong ground motion itself was correspondingly high. During the earthquake of 1966 in the Varta area, eastern Turkey, the heaviest damage was concentrated in an old alluviated river channel where the water table was very near or at the surface (Wallace, 1968). During the Caracas, Venezuela, earthquake of July 29, 1967 (magnitude 6.5), it was amply demonstrated that modern high-rise earthquake-resistive buildings located on alluvium can be extensively damaged even during a moderate earthquake (Steinbrugge and Cluff, 1968). These examples are only a few that could be cited that may be applicable to areas underlain by alluvium or similar deposits in southeastern Alaska.

The behavior of different kinds of earth materials during the Mexico earthquake of July 28, 1957 (magnitude 7.5), may furnish some insight to the degree of shaking expectable in possibly similar types of fine-grained deposits in southeastern Alaska. Ninety-nine percent of the buildings

in Mexico City damaged by this earthquake were founded on highly compressible lake clays and silts of Pleistocene ago, 32 feet to more than 700 feet thick, and underlain by more than 1,700 feet of coarser grained alluvium; the remaining 1 percent of the damaged buildings were on adjacent incompressible rocks of Tertiary age (Duke and Leeds, 1959; Rosenblueth, 1960). Differential settlement of the buildings, due to subsidence of the lake deposits where there had been extensive pumping of underground water, undoubtedly was a significant damage factor. However, shaking of the lake clays and silts, as indicated from isoseismal maps, also must have been much stronger in these deposits than on adjacent rocks. Although the elevated fine-grained marine deposits in coastal areas of southeastern Alaska are different in origin and age from the lake deposits in Mexico City, they may be similar enough in grain size and other physical properties to display corresponding earthquake intensities.

More and more instrumental records are becoming available to support the subjective conclusions that have resulted from observing the effects of shaking in different kinds of geologic material. As previously pointed out, instrumental records show great differences in acceleration and amplitude between rock outcrops, such as granite, and adjoining areas of unconsolidated deposits (Neumann and Cloud, 1955; Gutenberg, 1957). Richter (1959) also noted that "short-period motion in local earthquakes [Los Angeles Basin, California] recorded regularly with about 4 times the amplitude on the alluvium as on the rock." Seed and Idriss (1968) state "Strong ground motion records at several locations have clearly shown that the nature of ground motion is influenced to a large extent by soil conditions underlying the recording station; furthermore there appears to be a reasonably consistent relationship between the softness or hardness of the soil deposits and the ground motion characteristics."

Where sufficient quantitative data are available relative to intensity differences of ground types and the seismicity of a region has been determined, it is possible to construct seismic microregionalization maps. These maps show the expected variations of intensity within an area, and the susceptibility of a particular area to damage in relative terms, or in absolute terms if an absolute value is assumed to apply to a particular area of the map (Barosh, 1969). Thus, the maps show earthquake risk in terms of maximum intensity expected. As discussed by Barosh (1969), microregionalization maps have been made for many parts of Russia. Richter (1959) constructed a generalized microregionalization map of the Los Angeles area, California, as an illustration on how to compile a map of this kind.

Pertinent data are far too few in southeastern Alaska to permit construction of seismic microregionalization maps. About the best that can be done at this time is to make rough comparisons of expected degree of shaking between geologic units in a particular study area. Thus, in the Haines area (Lemke and Yehle, 1972), the geologic units were tentatively divided into three general categories: (1) those units where the strongest shaking is expected during an earthquake, (2) those units where shaking is expected to be intermediate between the other two categories, and

(3) those units where the least shaking is expected. In some instances, one unit may fall into more than one category. Even by using these broad categories, the interpretations are subject to revision and should be used only as generalized guidelines. In some coastal communities in southeastern Alaska that we have studied, even these generalizations are not possible because of the paucity of relevant information.

It also should be emphasized that the amount of expectable damage to manmade structures may not be directly comparable to the amount of shaking of the geologic units upon which they are built. There are several factors, mostly pertaining to differing engineering aspects of the structures, that may markedly affect the amount of ensuing damage and even may be the dominant factor. Whether a structure is built of wood, cinder blocks, bricks, or concrete (reinforced or nonreinforced) can markedly affect the amount of damage. Likewise, the design of the building in relation to the foundation conditions and whether it is built to be earthquake resistive or not are significant factors. Also, the height and rigidity of the building, particularly in respect to the natural frequency of the building in comparison to the period of the seismic waves, are factors that may significantly affect the damage potential. Thus, a more or less direct comparison between amount of ground shaking from a particular earthquake and ensuing damage to structures can be made only between similar structures in the same locality. It is the responsibility of the engineer, therefore, to be as aware as possible of the comparative intensities of the different geologic units, to collect additional data on soil properties whenever necessary, and to design accordingly.

Compaction

When loose cohesionless soils (those with no significant clay content) are shaken by strong ground motion during an earthquake, there is a tendency for them to compact with associated settlement of the ground surface. Also, the resulting densification of the materials, under some conditions, produces liquefaction and water-sediment ejection. Only the associated effect of settlement will be discussed here.

Settlement of the ground surface, where the surface is underlain by loose cohesionless materials, probably has accompanied every severe earthquake. The amount of settlement generally is dependent upon: (1) the looseness of the material, (2) the intensity of shaking, (3) the length of time of strong shaking, and (4) the relation between the natural frequency of the material when vibrated and the frequency of the impulse vibrations of the seismic waves. Thus, the higher the void ratio of the material, the greater the intensity of shaking, the longer the length of time of strong ground motion, and the closer the frequency of the seismic waves to the natural frequency of the material, the greater will be the expectable compaction and resulting ground settlement. Although the vibration effects on clays are considerably less than for cohesionless materials like sand, even soft clay compacts to some extent when subjected to intense vibrations having a frequency close to the natural frequency of the clay (Terzaghi and Peck, 1948).

Damage commonly has been heavy as a result of ground settlement due to compaction of loose sediments by shaking during an earthquake. This has been especially true where compaction was accompanied by tectonic downdrop of the land. Such a combination of effects occurred during the Chilean earthquake of 1960. Valuable agricultural land was inundated (especially at high tide) by sea water that extended upvalley several miles from stream mouths. Roads, buildings, and other manmade structures also were rendered useless in several Chilean cities (Dobrovolny and Lemke, 1961). Differential compaction and accompanying liquefaction of sediments in Puerto Montt, Chile, caused widespread damage to harbor areas. Similar effects resulted in heavy damage during the Niigata, Japan, earthquake of 1964.

Effects of compaction were widespread as a result of the Alaska earthquake of March 27, 1964. In the Portage area southeast of Anchorage, silt, sand and gravel, several hundred feet thick, underwent 1.5-2 feet of compaction; the compaction was accompanied by slightly more than 5 feet of tectonic subsidence (Plafker, 1969). The resulting 6-7 feet of total subsidence caused the town of Portage and large stretches of the highway and railroad to be inundated at high tide. In places, the highway roadbed compacted as much as 3 feet and otherwise was heavily damaged by fissuring, lateral spreading, and water-sediment ejection. The most severe damage to highways was at the Snow River crossing on the Seward-Anchorage Highway. There, new bridge piers, resting on piles driven to depths of as much as 160 feet, subsided a maximum of 4 feet, owing either to compaction or to liquefaction of fine-grained sediments constituting the valley fill (Kachadoorian, 1968). A section of roadway fill, 4 feet thick before the earthquake, subsided to the extent that it was about 2 feet beneath the surrounding ground surface after the earthquake. At Homer, near the southern end of the Kenai Peninsula (fig. 1), it was possible to get a quantitative measurement on the amount of compaction that took place during the earthquake. A well casing, installed to bedrock through 468 feet of alluvium, projected 1 foot above the surface before the earthquake. After the earthquake, it was 3.5 feet above the surface, indicating 2.5 feet of compaction (Grantz and others, 1964) or about 0.5 percent.

Loosely emplaced manmade fill, deltaic deposits, beach deposits, and alluvial deposits in southeastern Alaska may be susceptible to significant compaction during a severe earthquake, with resultant settlement and destruction to facilities built thereon. Other types of deposits that are loose and cohesionless might be similarly affected. Greatest damage probably would occur along or near the shoreline, especially if the subsided areas were subject to inundation by the sea.

Liquefaction in cohesionless materials

If loose to medium-dense materials, which are saturated and virtually cohesionless and which are loaded and confined by impermeable material, are subjected to strong shaking, the resulting tendency to

compact increases the pore-water pressure. The resulting upward flow of water may turn the materials into a "quick" or liquefied condition (Seed, 1970). As a result of closer packing of the solid particles, there is an excess of water and the load is transferred from the solids to the fluid. If sufficient pore pressures are produced, effective stresses are minimal or reduced to zero and the material assumes the character of a viscous liquid, hence the term "liquefaction" (Schroeder and Schuster, 1968). Other factors being equal, fine sands and coarse silts are most subject to liquefaction (Terzaghi and Peck, 1948). Also, the higher the void ratio, the greater is the tendency for the materials to liquefy.

Liquefaction of sand and silt deposits is a fairly common effect of large earthquakes. Liquefaction of sand was well illustrated at Niigata, Japan, during the earthquake of June 16, 1964 (magnitude about 7.3), and resulted in extensive damage. Structures and other objects gradually settled into the ground as a result of upward flow of water through the sand. The most dramatic case was the settlement and tilting of some apartment buildings; one building tilted through 80° and the occupants were able to evacuate the building by walking down the side of the building (Seed and Idriss, 1968).

When part of a sloping soil mass liquefies, the entire mass can undergo catastrophic failure and flow as a high-density liquid. The resulting flow slides can move rapidly down even nearly flat slopes and can cause heavy loss of life and property destruction. This type of failure is common along the shoreline where the materials generally are saturated, where they commonly are of the size most favorable for liquefaction, and where there is a free face for movement. There are several examples of large nearshore landslides and close offshore submarine slides that were generated by the process of liquefaction during the great Alaska earthquake of 1964 (Coulter and Migliaccio, 1966; McCulloch, 1966; Lemke, 1967). Some of these examples will be discussed further in the sections on earthquake-induced slides.

In southeastern Alaska, deltaic deposits probably would be most susceptible to liquefaction during a major earthquake because of their generally small grain size (sand and silt) and because of their high degree of saturation. Other loose and saturated surficial deposits consisting largely of fine sand and coarse silt, such as valley alluvium and beach deposits, also may liquefy.

Reaction of sensitive and quick clays

A sensitive clay loses a considerable part of its strength when disturbed either by remolding (kneading or squeezing) by man or by being shaken by an earthquake. When shaken by an earthquake, such clays commonly fail and become rapidly moving earthflows that can cause heavy damage and loss of life.

Sensitivity of a clay is defined as the ratio of undisturbed shear strength of a clay sample to the remolded shear strength of the same

sample (Terzaghi and Peck, 1948). Thus, for example, a clay having an undisturbed shear strength of 0.4 tsf (tons per square foot) and a remolded shear strength of 0.02 tsf has a sensitivity of 20. Shannon and Wilson, Inc. (1964) have classified clays by sensitivity as follows:

<3	: low
3-5	: low to medium
5-7	: medium
7-11	: medium to high
11-14	: high
14-20	: high to very high
20-40	: very high
>40	: extremely high

The term "quick" clay is commonly used to denote a clay of such high sensitivity that it behaves as a viscous fluid in the remolded state (Mitchell and Houston, 1969). In a quick clay the natural water content is greater than the liquid limit and may greatly exceed it. As a result of a shock, such as produced during an earthquake, a quick clay may suddenly change from virtually a solid to a viscous liquid that will flow almost horizontally and have high transporting power. Such sudden flowage over a wide area can produce catastrophic results.

Sensitivity of a clay has been found to correlate fairly well with the water-plasticity ratio of the clay or the so-called liquidity index. This ratio, as defined by Casagrande and Fadum (1944), is equal to the natural water content of the soil minus the plastic limit divided by the liquid limit minus the plastic limit. A high water-plasticity ratio indicates a low shear strength and a high sensitivity. A graph showing the relation between sensitivity and the liquidity index was drawn by Shannon and Wilson, Inc. (1964), for the Bootlegger Cove Clay at Anchorage, Alaska. This graph indicates that samples of Bootlegger Cove Clay having liquidity indexes of 1 or less had sensitivities of about 15 or less, whereas those samples having a liquidity index of 1.5 had a sensitivity of about 50. Analyses of Norwegian marine clays plotted on this graph show sensitivities 25-45 percent lower for the same liquidity indexes (Houston and Mitchell, 1969).

It also was found by Shannon and Wilson (1964) that for most samples of Bootlegger Cove Clay there was a relation between the liquidity index and the liquid limit. For those samples in which the liquidity index was greater than 1.0, the liquid limit ranged from 20 to 40 percent, whereas in clays having a liquidity index greater than 2.0, the liquid limit was between 19 and 27 percent. Therefore, liquid limits decrease with increasing sensitivity for the Bootlegger Cove Clay; this also is true for Norwegian marine clays (Houston and Mitchell, 1969).

Most highly sensitive (quick) clays of the world are glacial marine clays (Mitchell and Houston, 1969) which have been uplifted above sea level by isostatic rebound of the land or by tectonic deformation. They have been described from Scandanavian countries, Canada, Alaska, and other parts of the world. The high salt content of the marine environment of deposition of the clays acts as an electrolyte causing the clay to floculate. The floculated clay particles arrange themselves in a "house-of-cards" type of fabric with salt water filling the intervening large voids (Mitchell and Houston, 1969). After the clays are elevated above sea level, the salt begins to be leached from the clay. This removal of salt causes the interparticle force balance to be so altered that the formerly existing internal structure is no longer the most stable arrangement of particles and there is a marked decrease in strength. When the clays are disturbed, a high sensitivity results.

Extensive studies were made of the sensitivity of the Bootlegger Cove Clay at Anchorage because of the marked loss of shear strength and dramatic failures of the deposits during the Alaska earthquake of March 27, 1964. The Turnagain Heights landslide was the most conspicuous failure of this type (Hansen, 1965). About 2,100 samples of the Bootlegger Cove Clay, which is a delicately laminated silty clay, were tested (Shannon and Wilson, Inc., 1964) for sensitivity. Of these, 302, or 14 percent, had sensitivities greater than 10; 125, or 6 percent, had sensitivities greater than 20; 40, or 1.9 percent, had sensitivities greater than 30; and 11, or 0.5 percent, had sensitivities greater than 40. The greatest sensitivity measured was 60 (Hansen, 1965). Salt (NaCl) content of samples analyzed ranged from 1.03 grams per liter to somewhat less than 6 grams per liter. Inasmuch as sea water contains about 34 grams of salt per liter, the salinity tests indicate that most of the salt has been leached from the sediments since their original deposition.

If sensitive clays are present in southeastern Alaska they probably are most likely to be in some of the emergent fine-grained marine deposits (see section "Glaciation and associated sea- and land-level changes"). Available data, though, are too few to definitely confirm their presence. However, at least some of the deposits, such as those in the vicinity of Haines, have nearly the same grain-size material as the Bootlegger Cove Clay at Anchorage, have other similar physical properties, and may have similar sensitivities (Lemke and Yehle, 1972). The oldest emergent fine-grained marine deposits in southeastern Alaska are nearly as old as the Bootlegger Cove Clay, which was deposited about 14,000 years ago (Ernest Dobrovolny, oral commun., 1971). Therefore, like the Bootlegger Cove Clay, much of the former salt content may have been leached from the deposits resulting in a marked decrease in strength of the deposits. Much additional data need to be obtained to support or negate these tentative conclusions.

Water-sediment ejection and associated subsidence and ground fracturing

Documented records of some 50 major earthquakes show that in at least half of the instances water and sediment have been ejected from the surficial deposits (Waller, 1968). The ejection phenomena have been called fountaining, sand spouts or sand boils, mud or sand vents, blowouts, craters, and other names, depending upon the type of ejection or the resulting landform. The ejections commence sometime after strong motion has started and may continue for as long as 24 hours after seismic shaking has ceased (Waller, 1968). Ejecta may range from clear water to mud containing material as large as coarse gravel. Sand, however, is a common size fraction. Emission generally is in a pulsating manner although it may be steady but in diminishing flows. Ejecta heights of several feet are common, but fountaining to a height of as much as 100 feet was reported during the Alaska earthquake of 1964 (Grantz and others, 1964).

Ejections have been attributed to special soil conditions (especially liquefaction), to ground fracturing, and in a few instances to structural control. In all instances, however, the ejections appear to be associated with surface or near-surface unconsolidated deposits where there is a high water table. It also seems necessary, in order to produce an ejection, that there be a restrictive layer so that there is a confined-water condition[9] (Waller, 1968). If there is not a confined water-table condition, the water and sediment merely wells up to produce a quicksand with no jetting or spouting. This phenomenon was common during the Niigata, Japan, earthquake of 1964. A frozen surface crust furnished the restrictive layer to produce a confined-water condition in many places during the Alaska earthquake of March 27, 1964.

Roughly circular vents or craters commonly are formed during emission of the ejecta. Most are less than 3 feet in diameter and generally are backfilled with ejecta to within a few feet of the surface. Some however, may be several tens of feet in diameter and remain open to depths of at least 25 feet. In places, mounds or cones of ejected material accumulate around the vents and attain heights of several feet. The resulting mound, with a depression in its center, resembles in form a minature volcano. More commonly, however, the ejected material is sufficiently fluid to spread out over the surrounding ground as a thin sheet.

Most ejecta are emitted along earthquake-induced fractures but some appear to have a point source unrelated to fracturing. On the other hand, not all the fracture systems contain ejection sources or the emissions may be confined only to a certain part of the fracture system.

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[9] A confined-water condition is one in which the top of the water zone is in contact with an overlying relatively impervious zone. In contrast, under normal water-table conditions the top of the water zone is free to rise or fall as the water moves through the system (Waller, 1968).

The ejecta may come from a considerable depth and the ejected material may differ considerably in grain size from the surface material. The San Francisco earthquake of 1906 reportedly produced a fountain in which the ejected sand was brought up from a depth of 80 feet (Byerly, 1942). Sand was extruded from a depth of more than 25 feet during the Alaska earthquake of March 27, 1964 (Foster and Karlstrom, 1967).

The associated fractures commonly form an intricate mosaic pattern of ground breakage. Individual fractures may intersect nearly at right angles and be remarkably linear, vertical, and well defined, especially if the ground surface is frozen (Lemke, 1967). In other places, where tension fractures have formed and there is lateral extension toward a free face, long roughly parallel fractures may form. Individual fractures commonly range in width from hairline cracks to 1 or 2 feet wide and are backfilled by slumping to an open depth of a few feet. Some, however, have been reported to be as much as 30 feet wide and open to a depth of 25 feet (Foster and Karlstrom, 1967).

Previous to the Alaska earthquake of 1964, most reported ejections seem to have been localized along valley floors or in low-lying coastal areas. During the great New Madrid earthquake of 1811, ejections were prevalent along the Mississippi River bottomlands but were lacking on nearby river terraces and other higher ground (Fuller, 1912). During the Charleston earthquake of 1886, ejections were confined to lowlands, particularly swamps (Dutton, 1889), and during the Assam earthquake of 1897, they were localized near river channels (Oldham, 1899). Ejections were concentrated in or near low-lying coastal sediments during the Yakutat Bay earthquake of 1899 (Tarr and Martin, 1912), the Lituya Bay earthquake of 1958 (Davis and Sanders, 1960), the Chilean earthquake of 1960 (Sergerstrom and others, 1963), and the Niigata, Japan, earthquake of 1964 (Seed, 1970). It is interesting to note that extensive sand flows, which resulted from the Huslia (north-central Alaska) earthquake of April 1958 presumably were restricted to sand-dune localities (Davis and Sanders, 1960). The ejections from the Huslia earthquake were produced under winter conditions in a permafrost region, so there was both a frozen surface-confining layer and a thick underlying permafrost layer.

During the Alaska earthquake of March 27, 1964, water-sediment ejection was most prevalent on deltas, particularly the outer portions, and generally was associated with intensive ground fracturing (Waller, 1968). Ejections, however, also were locally abundant on valley floors, alluvial fans, swamps, outwash plains, lakeshores, tidal flats, and other low-lying coastal areas. The abundance of ejections on alluvial fans during this earthquake seems unusual because of lack of reported ejections in this type of terrain during other large earthquakes elsewhere. In general, ejecta were not observed in dune or morainal areas or along hillsides (except where there were landslides), although ground-fissuring was severe in some of these areas (Waller, 1968).

Water-sediment ejection and associated subsidence and ground fracturing commonly cause extensive damage to the works of man. Ejecta may

fill basements and other low-lying parts of buildings and other structures. Also, agricultural land can be adversely affected by being covered with a blanket of infertile soil; small ponds can be filled or made shallow. Where there is only a general welling up of ejecta without fountaining, a quicksand condition is produced and structures sink into the liquefied material. Removal of ejecta material from beneath the surface can cause ground collapse, enlargement of ejecta craters, and general subsidence over fairly large areas with resultant damage to structures built thereon. Some cratering enlargement and subsidence may not be evident for several months after an earthquake (Waller, 1968). Associated ground fractures can cause heavy damage by displacing building foundations, breaking waterlines, sewerlines, and gaslines, and cracking streets, roads, and railroad grades. Damage of this type was severe in a residential area (Forest Acres subdivision) of Seward following the Alaska earthquake of 1964 (Lemke, 1967).

Water-sediment ejection and associated subsidence and ground fracturing might be fairly widespread in southeastern Alaska in the event of future large earthquakes. As noted from the examples cited above, these phenomena are most likely to occur in lowland areas such as on valley floors, river terraces, deltaic deposits, tidal flats, alluvial fans, swamps, and lakeshores. Low-lying coastal areas in general, underlain by unconsolidated deposits with a high water table probably should be considered likely loci for these phenomena.

Earthquake-induced subaerial slides and slumps

Landslides commonly are triggered both on land and offshore as a result of strong ground shaking or of ground rupture. Depending upon the material involved in sliding and the physical setting, slides may be of several different types[10] with accompanying diverse effects. Under this heading will be described large slides and smaller associated slumps that are generated on land. Subaqueous slides, although in places part of the same slide mass, will be described under a separate heading.

Earthquake-induced slides on land most commonly are confined to steep slopes, but sliding also may take place on moderate to nearly flat slopes. On steeper slopes, either bedrock or surficial deposits may be involved. On moderate slopes to nearly flat surfaces, sliding generally is confined to fine-grained surficial deposits that are moderately to highly plastic or are subject to liquefaction.

Examples of earthquake-induced slides on land are given below to illustrate the factors that contributed to the failures and to relate these factors to towns in southeastern Alaska. The first examples are those involving chiefly bedrock.

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[10] For description and classification of landslide types see Varnes (1958).

During the Lituya Bay, Alaska, earthquake of July 10, 1958, about 40 million cubic yards of rock, loosened either by displacement on the Fairweather fault zone or by shaking alone, plunged down a steep slope of an inlet at the head of Lituya Bay (fig. 1) from an altitude of about 3,000 feet (D. J. Miller, 1960). The rockslide caused water to surge over the opposite wall of the inlet to a maximum height of 1,740 feet and generated a fast-moving wave that sank two of three fishing boats near the mouth of Lituya Bay and killed two people.

During the Hebgen Lake, Montana, earthquake of August 17, 1959, a spectacular rockslide, involving 37 million cubic yards of bedrock and talus debris, slid into the Madison River canyon, dammed the river and created a large lake, and killed 28 people. Topographic and geologic conditions along the south canyon wall caused the instability (Hadley, 1964).

The Alaska earthquake of March 27, 1964, triggered numerous rock and debris slides (Post, 1967). One of the largest was the Sherman rockslide avalanche, which originated on the west side of a Matterhorn-like peak (now called Shattered Peak), and plunged 2,000 feet down a slope exceeding 40°. It then moved 3 miles at high speed possibly on a cushion of compressed air, across the terminal part of Sherman Glacier (Shreve, 1966). Preexisting conditions that contributed to the failure were a combination of: (1) steep unstable slopes (in places as much as 70°), and (2) potential surfaces of rupture along bedrock fractures and bedding planes dipping steeply toward the free face (Plafker, 1968). Also, a probable thrust fault, located where the slide scar developed, may have contributed to the failure (Marangunic and Bull, 1968). Numerous rock and debris slides in still other places blocked transportation routes or partially filled drainage courses, which caused a threat of water being impounded back of the slide materials and of the slide-dams suddenly rupturing (Grantz and others, 1964; Lemke, 1967).

Earthquake records are replete with accounts of sliding of surficial deposits during moderate to large earthquakes. The types of slides and the geologic conditions under which surficial materials slide are diverse. Several examples from other places are given so as to better assess the hazard in or near towns in southeastern Alaska.

Earth slumps, debris slides, earth lurch, and earthflows were common during the San Francisco earthquake of 1906 (Lawson and others, 1908). One of the larger earth slumps was about a mile long, half a mile wide, and distended into the ocean about a quarter of a mile. It apparently originated from an old landslide area and flowed as a partially plastic mass; incorporated ground water apparently was an important factor in facilitating the sliding. Sudden accession of water to incoherent materials was the major contributing slide factor in triggering an earthflow 9 miles south of San Francisco. During shaking, water suddenly emerged from near or along a concealed fault and poured onto a sandy surface. The mixture of sand and water flowed rapidly down an arroyo of of low gradient for a distance of 900 yards and attained a maximum

thickness of 13 feet. Earth lurching (horizontal translation of cracked ground toward a free face) was common on the flood plains of several rivers or tidal flats and on manmade fill.

The great Chilean earthquake of 1960 triggered numerous landslides. Most were in fine-grained material, particularly in manmade fill. Liquefaction of loose sandy soils caused heavy damage along the waterfront of Puerto Montt. Loose sandy hydraulic and dumped fills lost virtually all their shear strength and flowed into the sea (Duke and Leeds, 1963). There was total failure of harbor quay walls, heavy damage to a sheet pile seawall, and failure of building foundations. Throughout southern Chile, earthfills for highways and railways were heavily damaged by slumping, fissuring, and earthflow. Many failures were in sidehill construction, the earth moving laterally downhill (Duke and Leeds, 1963). Fill slumped in many bridge approaches and caused cracked abutment walls, rotated piers, and collapsed decks.

Landsliding of surficial deposits during the great Alaska earthquake of 1964 was widespread and resulted in tremendous destruction to manmade facilities. Damage was particularly heavy along the waterfronts in the cities of Anchorage, Seward, and Valdez (see fig. 3 for locations).

In the Anchorage area, as discussed by Hansen (1965), most of the destructive landslides moved on nearby horizontal slip surfaces after loss of strength in the Bootlegger Cove Clay. The Bootlegger Cove Clay is a fine-grained glacial estuarine-marine deposit, which contains zones of low shear strength, high water content, and high sensitivity under vibratory stress. Some failures were attributed to spontaneous liquefaction of intercalated sand layers. Translatory slides devastated part of downtown Anchorage and Turnagain Heights. There was virtually total destruction to homes in the Turnagain Heights slide area owing to complete disintegration of the prequake land surface as the slide moved seaward.

Of several types of landsliding in the Seward area during the 1964 quake, the most spectacular was large-scale subaerial and subaqueous landsliding of alluvial fan-delta deposits along the waterfront. Harbor facilities were almost completely destroyed and the entire economic base of the town was wiped out. As described by Lemke (1967), a strip of land, approximately 4,000 feet long and 50-500 feet wide, slid into Resurrection Bay. Ground behind the slide scarp was intricately fractured for a landward distance of as much as 800 feet owing to almost horizontal movement of the landmass toward a free face. The submarine sliding resulted in marked deepening of water in the vicinity of the former shoreline. Factors other than strong shaking that may have contributed to the sliding were (1) the long duration of strong ground motion, (2) the fineness of grain and texture of some of the material involved in the sliding, (3) the probability that the finer grained materials liquefied and flowed seaward, and (4) the added load of manmade facilities built on the edge of the shore.

A massive slide in outwash fan-delta deposits, similar in many respects to the one along the waterfront in Seward, was triggered at Valdez by the 1964 quake. The slide, which originated under water along the front of the delta but which extended some distance landward, involved approximately 98 million cubic yards of generally fine grained material and destroyed harbor facilities and nearshore installations. Several factors apparently combined to produce the slide (Coulter and Migliaccio, 1966). Shaking was of critical intensity to liquefy the loose- to medium-dense saturated sand of the deltaic deposits and to transform the sediment into a concentrated suspension with minimal shear strength. Also, withdrawal of water from the beaches during shaking, owing to generation of abnormal waves, may have been accompanied by a sudden decrease in hydrostatic head. This may have resulted in an immediate increase in weight of the upper soil layers and a corresponding decrease in support at the toe of the slide owing to the lowered sea level. The fact that sliding took place at low tide also may have been a factor.

In addition to landsliding along waterfronts, the Alaska earthquake of 1964 triggered landslides in surficial deposits in many other areas. Particularly prevalent were those that affected highways and railroads. Lateral displacement of unconsolidated sediments toward a free face caused severe damage to road fills and bridges. For example, all bridges on the Copper River Highway, except one on bedrock, were damaged by being placed in compression by lateral displacement of sediments (Kachadoorian, 1968). Landsliding and snow avalanching onto highways also caused damage. Damage to the Alaska Railroad from slides was similar in most respects to that for highways. Between Seward and Anchorage, 17 railroad bridges were damaged or destroyed as a result of direct shaking, landslides, subsidence, ground cracks and lurching, and inundation at high tides (Hansen and others, 1966).

The above examples indicate that the following factors, occurring singly or in combination, could contribute toward earthquake-induced landslide failures in southeastern Alaska: (1) steep slopes, (2) rocks whose foliation or beds dip toward a free face, (3) easily weathered bedrock and associated thick talus, (4) planes of weakness along preexisting faults, (5) actual tectonic displacement along a fault during an earthquake, (6) presence of earth materials that tend to liquefy and flow, even on nearly flat surfaces, (7) prior saturation of deposits or sudden emission of water onto deposits, (8) presence of plastic or sensitive clays and silts, (9) presence of stream channels and other open faces to facilitate slumping and lurching, (10) presence of loose fine-grained deltaic and fan delta deposits on unstable slopes, and (11) presence of loose manmade fills. Each of these factors probably is present in one or more places in southeastern Alaska.

Earthquake-induced subaqueous slides

Submarine slides triggered by large earthquakes, although particularly important in respect to navigation because of changed water depths, generally are imperfectly known. It is possible to define the areal

extent and size of the slides only in those places where accurate nautical charts existed before the earthquake and where equally or more accurate postearthquake soundings were taken. The amount of tectonic uplift and subsicence in the ocean is equally difficult to determine.

Postearthquake nautical charts, made by the U.S. Coast and Geodetic Survey, now the National Ocean Survey, (Environmental Science Services Administration, 1967) for a number of places in Alaska following the great Alaska earthquake of 1964, showed that most major sliding in the sea was concentrated along the foreslopes of fan deltas. At Seward, in addition to the sliding along the waterfront that has been described previously, large-scale submarine sliding of deltaic deposits at the head of Resurrection Bay shifted bathymetric contours several hundred feet landward. Water that was 20 feet deep before the earthquake was about 130 feet deep after the earthquake and the underwater slope steepened from approximately 15° to approximately 20° (Lemke, 1967). At Valdez, as previously described, a large submarine slide, which extended headward onto land, was triggered along the front of the outwash delta. It is not known whether other nearby fan delta fronts slid, but cable breaks 3 miles west of Valdez during the earthquake of Feb. 19, 1908, were attributed to earthquake-triggered slides at that time (Coulter and Migliaccio, 1966).

A number of other breaks in communication cables in southeastern Alaska also appear to have been caused by earthquake-induced submarine slides (D. Alford, Alaska Communication System, written commun., 1966; Heezen and Johnson, 1969). Cable breaks offshore from the Katzehin River delta (5 miles southeast of Haines) can be correlated in time with the Lituya Bay earthquake of July 10, 1958. A submarine slide that occurred along the face of the Katzehin River delta 8 hours after the Alaska earthquake of March 27, 1964, probably also was due to secondary slumping of the delta front and related to that earthquake. Considering the limited extent of fiord bottoms transected by these cables, there probably have been many undetected earthquake-induced slides in southeastern Alaska along delta fronts and in perched sediments along fiord walls.

In attempting to assess the potential for earthquake-induced subaqueous slides, it seems reasonable to assume that future sliding will occur most frequently in those types of deposits that have slid in the past. On this basis, the largest slides are likely to be triggered along the larger delta fronts. However, some of the smaller fan delta deposits and the perched sediments along fiord walls also are likely to slide during a large earthquake. If these nearshore deposits are subject to liquefaction, slides may be triggered on fairly gentle slopes. The resulting slurrylike mass can move a considerable distance from its source and spread out as a relatively thin blanket.

Effects on glaciers and related features

Strong ground motion and land-level changes resulting from earthquakes can affect glaciers and related features in several ways:

(1) cause thickening of glacier ice, (2) cause thinning of glacier ice, (3) disrupt glacier-fed streams, (4) change flow characteristics of glacier ice resulting in glacial advance, (5) cause accelerated calving of the termini of tidewater glaciers resulting in possible retreat of glaciers, (6) cause glaciers to advance or retreat as a result of land-level changes, and (7) produce long-term changes in mass or flow characteristics of glaciers because of tectonic displacement (Post, 1967).

Tarr and Martin (1914) attributed the sudden short-lived advances of nine glaciers in the Yakutat Bay region to the large Yakutat Bay earthquakes of 1899 and, as a result, proposed the Earthquake-Advance Theory. They theorized that the severe shaking caused large amounts of snow and ice to avalanche onto the glaciers from adjacent steep mountain slopes. The increased load, they speculated, upset the dynamic equilibrium of the glaciers and caused sudden short-lived advances. However, since Tarr and Martin's work, it is not known that several glaciers in Alaska, western Canada, and elsewhere undergo sudden advances called "surges" that are unrelated to earthquakes. No glacier surges have been reported in the immediate vicinity of any strong earthquake since 1899. This fact, plus the probability that significant avalanching did not take place on at least one of the glaciers observed by Tarr and Martin, render the Earthquake-Advance Theory suspect (Post, 1967).

The Lituya Bay earthquake of July 10, 1958, produced only minor evidence of avalanching onto glaciers other than small rockslides. However, as previously described (see section "Earthquake-induced subaerial slides and slumps"), a huge rockslide plunged down a steep slope at the head of Lituya Bay and generated a wave 1,740 feet high. The wave cut off part of the front of Lituya Glacier but did not cover its surface (D. J. Miller, 1960).

The Alaska earthquake of March 27, 1964, produced no snow and ice avalanches of large size or any that adversely affected man despite the fact that about 20 percent of the area that underwent strong shaking is covered by ice (Post, 1967). This is even more surprising because the earthquake occurred at a time when large quantities of snow were present and the avalanche hazard was high under nonearthquake conditions. Also, with few exceptions, hanging glaciers did not appear to have been affected and there was no unusual calving of glacier termini into tidewater (Ragle and others, 1965). The most important effect of the earthquake upon the regime of glaciers was the extensive rockfalls onto the glacier ice. The largest of these rockfalls avalanched onto Sherman Glacier on the east side of Prince William Sound near Cordova. It covered about 50 percent of the ablation area of the glacier to an average thickness of 5 m (Post, 1967) and was computed by Plafker (1965) to contain about 25 million cubic meters of rock debris. Possible future long-term effects of this rockslide onto the glacier, as well as effects of smaller rock avalanches onto other glaciers, include reduced melting, increased flow, and advance of the glaciers (Post, 1967). No evidence was found of changes in levels of glacier-dammed lakes by englacial drainage or other

newly established drainage courses as a result of the earthquake. A few minor changes in stream channels were noted near some glaciers, but there was no evidence of any permanent changes (Post, 1967). Changes in land levels due to tectonic deformation were too small to noticeably affect the regimen of the glaciers.

In contrast to the general lack of glacier effects on man related to the Alaska earthquake of 1964, the cataclysmic avalanche of ice and rock that fell from a high glacier-covered peak in Peru as a result of the Peru earthquake of May 31, 1970, produced devastating effects on man and his works. As described by Erickson and Plafker (1970), the avalanche fell from the face of a 5,500- to 6,500-meter-high cliff, moved downslope (probably part way on a cushion of air) for a distance of 14.5 km at an indicated average velocity of 400 km per hr and devastated a 15 km^2 area. As it moved downvalley, it took on the character of a highly fluid mudflow and was estimated to average about 3 m in thickness. The mudflow buried the town of Yungay, most of Ranrahirca, and several other communities. At Yungay, 320 people are known to have died and 15,000 were missing. At Ranrahirca, an estimated 1,800 of the 1,850 residents were killed.

In trying to evaluate the effects of future earthquakes on glaciers in southeastern Alaska, as the effects relate to man, it should be noted that present glaciers are limited chiefly to mountainous areas in the northwestern and northeastern parts of the region. Only in the Glacier Bay area and in a few other bays and inlets do glaciers actually reach tidewater. Therefore, most towns in southeastern Alaska are sufficiently distant from glaciers so as not to be directly affected by them in the event of a large earthquake. Adverse effects in areas near glaciers probably would be limited largely to (1) flooding of streams owing to sudden release of large quantities of water by the breaking of earthquake-induced dams of ice or the sudden draining of ice-walled lakes, (2) calving off of icebergs or avalanching of large quantities of ice and debris down steep slopes into bays and inlets causing navigation problems from floating icebergs or generating high waves, and (3) snow avalanching on steep slopes. The effects of tectonic changes resulting from most future earthquakes probably would be too small to significantly change the regimen of any of the glaciers. If, however, glaciers did begin to advance, because of tectonic uplift for example, then facilities such as roads near the present termini of glaciers might be adversely affected.

Effects on ground water and streamflow

It is well known that large earthquakes can significantly affect ground- and surface-water regimens. Examples from two earthquakes, the Hebgen Lake, Montana, earthquake of August 17, 1959, and the Alaska earthquake of March 27, 1964, are sufficient to illustrate most of the expectable effects.

The Hebgen Lake earthquake produced widespread effects on ground- and surface-water supplies in the area. As described by Swenson (1964),

water issued under pressure from at least two wells in the area, which had never been known to flow; water levels in several other wells declined noticeably. Flow began or flow pressure increased in wells within a 200-mile radius of the epicenter. Numerous sand spouts and some sinkholes were formed where the ground water reached the surface under pressure. Some springs discharged at rates three times as much as normal; the flow of others decreased or stopped. Water from many of the springs, which before the earthquake was remarkably clear, became brown and turbid. The water temperature of some springs reportedly was higher after the earthquake, and new thermal springs issued from an active fault zone. The discharge of many streams promptly increased 0.5-0.2 cfs per mile after the earthquake (Stermitz, 1964). Sediment load also increased on all streams tributary to Hebgen Lake; most of the material in suspension consisted of fine clay (Hadley, 1964). Removal of material from the large landslide on the Madison River became evident soon after the water started flowing through the new spillway across the slide. Shortly thereafter, large quantities of sediment were being transported downstream as the stream downcut its spillway channel. Aggradation in the river and bank erosion were very pronounced for a distance of about 2 miles downstream from the base of the slide (Hadley, 1964).

The effects of the Alaska earthquake of March 27, 1964, on the ground- and surface-water regimens were studied fairly extensively. Waller (1966a, b) noted that the short-term effects on ground water included (1) surging of water in wells, (2) extrusion of water, mud, and sand, (3) failure of well systems, and (4) turbidity of water in wells and springs. He ascribed long-term effects to physical changes in aquifers, which included temperature changes, chemical quality changes, and the lowering of the water level or artesian pressures. Ground water was drastically affected mostly in unconsolidated aquifers for at least 100 miles from the epicenter. Water in shallow wells oscillated as a result of ground motion. Water levels dropped in some wells after the earthquake but increased in others. Many well systems failed owing to sanding or silting of the pump column or to differential movement and breaking of well casing. Hydrostatic pressure in deep aquifers dropped as much as 15 feet. Lower water levels and artesian pressures presumably were caused chiefly by sediment compaction, which lowered the discharge rates or transmissibility of the aquifers. Salt-water encroachment to wells on an island near Anchorage seems to have increased. Increased pore-water pressure caused liquefaction in some granular materials and also caused disastrous landslides. Ice cover on many lakes and streams was extensively broken during the earthquake. Some streams increased their flow but this was not as common as diminished streamflow. Changes in streamflow were controlled by ground fracturing in and near streambeds and by snow and rock avalanches. Most avalanches did not block the stream flow for long but effects from some persisted for several months (Waller, 1966a; Lemke, 1967).

Most or all of the effects on ground water and streamflow described above probably also could occur in parts of southeastern Alaska during future large earthquakes. In general, ground-water effects would be

most pronounced in major stream valleys and in coastal areas where surficial deposits are most prevalent and are thickest. Streamflow, as previously indicated, may be high if a slide dam, holding large quantities of water, is breached suddenly. Water is rarely impounded behind a snowslide or a rockslide if there is an absence of fine-grained material, thus permitting the water to drain through the slide. On the other hand, if the slide material is fine grained and relatively impervious, the dam may fail catastrophically by overtopping, followed by very rapid erosion.

Tsunamis, seiches, and other abnormal water waves

Abnormal water waves associated with large earthquakes commonly cause vast property damage and heavy loss of life. Tsunami effects can be devastating to coastal areas many thousands of miles from their generation source. Seiche effects generally are confined to inland bodies of water such as lakes or to relatively enclosed coastal bodies of water such as bays. Other abnormal waves, generated by submarine sliding or by subaerial landsliding into water, generally produce only local effects but may be highly devastating. During an earthquake, two or all three types of waves may be present in an area. It may not be possible to differentiate the wave type and the resultant damage from one particular kind of wave. By some definitions, all three types of waves are called "tsunamis." However, for purposes of discussion here, each type and its possible effects will be discussed separately.

Tsunamis

Tsunamis, otherwise known as seismic sea waves, are gravity waves generated by exceptional underwater disturbances, such as those associated with earthquakes or from other causes (Wiegel, 1964, 1970). Waves of this type erroneously have been referred to as tidal waves but they are not directly related to tides. Tsunamis consist of trains of waves with periods of 5-60 minutes or longer and may be detectable for as long as a week on tide gages. Near the point of origin the first wave generally is the largest, but at greater distances one of the succeeding waves may be the greatest. Large tsunamis are believed to be generated by sudden tectonic displacement (uplift, subsidence, or tilting) of the ocean bottom. For noticeable tsunamis to occur, the magnitude of the earthquake must be greater than $6.3+0.01H$, where H is the focal depth in kilometers; for disastrous tsunamis to be generated, the magnitude generally must be greater than $7.75+0.008H$ (Wiegel, 1964). Also, for a tsunami to be of large size, a substantial vertical change of ocean bottom is necessary, such as could occur on a dip-slip fault. It is unlikely that a major tsunami would be generated as a result of movement on a strike-slip fault unless the fault offset a seamount, submarine cliff, or some other large underwater topographic disconformity (Wiegel, 1970).

In the open ocean, tsunami waves are long and low (of the order of 50 miles long and 2 feet high) and, therefore, are not easily detectable. In oceanic depths of 12,000-30,000 feet, they have speeds of 425 to 600

miles per hour. In shallower water, such as in bays and inlets, the speeds are considerably less. When tsunami waves approach a coast, they may greatly increase in height and (or) runup onto land. The wave height depends partly upon the offshore topography, the waves generally being higher where the offshore zone is gently shelving, where reefs border the coast, or where waves have been refracted around shoals, islands, or other intricate topographic configurations. Heights attained by tsunami waves running into the heads of funnel- or triangular-shaped bays may be considerably higher than at the mouths of the bays (Cox, 1961; Hodgson, 1966). This is not always so, however, because wave height appears to be a function of the ratio of the period of the tsunamis to the period of the bay (Wiegel, 1970).

Wave runup onto shore can range from a barely perceptible abnormal rise, generally detectable only on a tide gage, to heights of more than 100 feet (Wiegel, 1964). In most instances the waves come in as fast rising swells; in other instances they come crashing onto shore with great violence. It is characteristic of tsunami waves that the water level draws down along the coast between arrival times of the wave crests (Wiegel, 1970). In fact, one of the best indications that tsunami waves can be expected in an area is for the water to rapidly recede from shore following an earthquake. During past earthquakes many people have not recognized this phenomenon for the danger it portends. Thus, instead of fleeing to higher ground if they are near the shore, some people have actually gone out onto the newly exposed ocean floor to collect shellfish or for other reasons and have been drowned by the incoming tsunami wave.

Damage from tsunamis can be exceptionally great. In addition to heavy loss of life and damage to nearshore buildings, docks, and other harbor and coastal installations, moored boats can be extensively damaged by pounding against other boats and docks or by being carried ashore and beached.

One of the largest known tsunamis occurred June 15, 1896, in Japan. A wave ran up onto land to an elevation of 75-100 feet above tide level and killed more than 27,000 people and destroyed approximately 10,000 houses (Leet, 1948). This is the largest known loss of life from a tsunami.

The enormous tsunami generated by the large Chilean earthquake of May 22, 1960, demonstrated that disastrous wave effects can be produced thousands of miles from the generation source. The tsunami waves not only inflicted extensive damage on coastal communities throughout a large part of southern Chile, but waves traveled at high velocity across the Pacific Ocean with significant runups recorded in Hawaii, California, Alaska, Japan, New Zealand, Australia, and at many other coastal areas (Berkman and Symons, 1964).

Along the coast of southern Chile, waves reached a maximum reported height of 70 feet at one locality, about 50 feet at another locality, and 20-40 feet at several localities. The first wave commonly was not the highest. At many localities, the third wave obtained the highest runup whereas in still other places the second or even the fourth wave was the highest. Water drawdown commonly preceded the arrival of the first wave so that in places, for periods of 10-20 minutes or more, the offshore floor was exposed out to distances ranging from several hundred to 1,500 feet. The tsunami caused catastrophic destruction to the shoreline part of the town of Ancud in southern Chile. All structures (several hundred houses) to the maximum height reached by the waves (about 20 feet) were torn from their foundations and most were swept out to sea (Galli O. and Sanchez R., 1963). Farther north at the mouth of the Rio Valdivia, three powerful waves, each about 28 feet high, destroyed docks, buildings of wood and concrete, and installations built of reinforced steel (Sievers and others, 1963). Three ships anchored in the bay were first pounded against the coast, then carried seaward, only to be returned by incoming waves, which grounded and sank two of the vessels. Waves traveled up the Rio Valdivia a distance of about 15 miles. The depth of the Rio Valdivia near its mouth was increased by about 15 feet in places by the scouring action of the waves, so it became navigable for ships of average tonnage--one of the few examples where a tsunami produced a beneficial effect.

Tsunami waves generated by the Chilean earthquake of 1960 traveled across the 6,600 miles of ocean to Hilo, Hawaii, in 14.9 hours, an average velocity of between 440 and 445 miles an hour (Cox and Mink, 1963). The waves formed a bore at Hilo and reached a runup height of 35 feet. In spite of an elaborate tsunami warning system that was put into effect 5 hours in advance of the arrival of the tsunami at Hilo, 61 people were killed and 282 were injured. A few hours later, waves as high as 130 feet hit coasts in Japan causing extensive damage and drowning many people (Bird, 1969). In Alaska, maximum wave heights were 11.0+ feet in Massacre Bay (Attu Island in the Aleutian Islands), 5.2 feet at Yakutat, 3.0 feet at Sitka, and 1.2 feet at Skagway (Berkman and Symons, 1964).

The tsunami generated by the Alaska earthquake of March 27, 1964, was the largest since the Chilean tsunami of May 22, 1960(Spaeth and Berkman, 1969). It was caused by broad crustal warping along a hinge line which extended northeastward from near the southeast coast of Kodiak Island (fig. 3) and across the eastern part of the Kenai Peninsula, passed approximately through Valdez, and terminated in the vicinity of the Copper River (Plafker, 1969). The generation area was the uplifted ocean bottom southeast of this line, where uplift may have been as much as 50 feet (Spaeth and Berkman, 1969).

The tsunami waves generated by the Alaska earthquake of 1964 struck with devastating force along the Alaska coast from Kodiak Island to Cordova (fig. 3). In spite of the fact that the effects of the tsunami were less than they might otherwise have been, because the highest waves struck in most places in Alaska at low tide (Van Dorn, 1964), approximately

105 people lost their lives in Alaska (Doak C. Cox, written commun., 1970) from this type of wave and property damage was extremely high. On Kodiak Island and nearby islands, the waves killed 18 people, and about 45 million dollars worth of property was destroyed by a combination of wave action and tectonic subsidence (Plafker and Kachadoorian, 1966). The waves also were particularly destructive in other areas of tectonic subsidence, such as on the coast of the Kenai Peninsula. Most of the 13 people who lost their lives at Seward were killed by wave action; property damage could not be separated from loss incurred from landsliding and other causes but it was exceedingly high for a town of its size (Lemke, 1967). At Cordova, the waves caused some damage and one casualty (Plafker and others, 1969). About 5 million dollars worth of damage was sustained in the twin cities of Alberni and Port Albernia in British Columbia on the west coast of Vancouver Island. Minor damage was done along the Washington coast and moderate damage occurred along the Oregon coast where a family of four was drowned. The tsunami waves reached record heights along the coast of northern California. At Crescent City, Calif., the highest wave crested about 20 feet above tide level (Tudor, 1964); 11 people were killed and 35 were injured. Total damage in California amounted to almost 10 million dollars (Spaeth and Berkman, 1967). Damage was light in Hawaii and in Chile.

As one would expect, arrival time and periodicity of the waves, number of large waves, and height of waves of the 1964 tsunami varied from place to place. The first of seven large waves, with periods of 55-90 minutes, arrived at Cape Chiniak on Kodiak Island, Alaska, 34 minutes after the quake and crested about 23 feet above tide level (Plafker and Mayo, 1965). The first wave reached the southeast coast of the Kenai Peninsula about 20 minutes after the earthquake began. It was followed by a second wave, 10-12 minutes later, which ran up to an estimated height of 35-40 feet above lower low water. The first wave reached Seward (fig. 3), at the head of Resurrection Bay, about 25 minutes after shaking stopped; it and several succeeding waves were 30-40 feet high as they neared the head of the bay, the third wave probably was the highest (Lemke, 1967). Boats, houses, and railroad cars, collected from the waterfront, were carried over a railroad embankment and deposited inland. Rails were twisted and an overturned locomotive (weighing 85 tons) was moved several tens of feet inland, attesting to the energy of the waves. Water withdrew from the shores in the intervals between the tsunami waves but apparently not before the arrival of the first tsunami wave. The first wave was preceded, however, by a withdrawal of water (lowering of 18 feet) at Rocky Bay near the southwest tip of the Kenai Peninsula, as well as at Middletown Island (near Montague Island shown on fig. 3) and at several other places to the east (Plafker and Mayo, 1965; Wilson and Tørum, 1968). The Coast Guard vessel Sedge ran aground near Cordova during a 27-foot drawdown of the sea in the interval between two waves (Grantz and others, 1964). At Sitka (fig. 1), waves reached a maximum height of 14.3 feet above normal tide, at Yakutat they reached a height of 7.6 feet, and at Juneau they were 7.5 feet above normal (Spaeth and Berkman, 1967). An unusually high tide occurred at Haines about 5.75 hours after the earthquake, followed by five more waves or high tides

about 1 hour apart (Retha M. Young, Postmaster, Haines, written commun., 1964). However, arriving as they did a considerable time later than the tsunami waves arrived along the outer coast of southeastern Alaska, there is some doubt that these were tsunami waves.

The tsunamis generated during the Chilean earthquake of 1960 and the Alaska earthquake of 1964 demonstrated that, under certain conditions, tsunami waves can travel up long narrow inlets, similar to those in southeastern Alaska, and attain fairly high runups. During the Chilean earthquake, tsunami waves traveled approximately 80 miles up a relatively narrow island-studded inlet to Puerto Aquirre, where they attained a runup of approximately 10 feet; thence, they traveled 42 miles up a much narrower transverse inlet to Puerto Aysen, where they reached a height of about 5 feet (Sievers and others, 1963). During the Alaska earthquake, tsunami waves did heavy damage to the town of Port Alberni in British Columbia, located at the head of a long inlet about 40 miles from the entrance to Barkley Sound on the west coast of Vancouver Island. The inlet is 0.5-1 mile wide, about 600 feet deep at its mouth, and less than 180 feet deep at its head. Damaging waves may have reached a maximum height of 27 feet and had a periodicity of 1.72 hours. Wilson and Tørum (1968) believe that the waves must have gained their extraordinary height by some near-resonance local phenomenon inasmuch as the maximum waves at Victoria, only a short distance away, were only 3 feet high.

It is beyond the scope of this paper to try to predict the maximum height of tsunami waves and of runup on land that might be reached in different parts of southeastern Alaska. Most all of the earthquakes in southeastern Alaska and vicinity of large enough magnitude to produce a tsunami have been along the Fairweather-Queen Charlotte Islands fault system where movement has been chiefly horizontal. As noted previously, a considerable vertical displacement of the ocean bottom, such as would occur along a dip-slip fault, is required to generate a tsunami. However, this does not preclude tsunami waves from far-distant generation sources hitting the outer coastal areas and attaining high runup levels. The degree of attenuation reached by these waves in traveling up the long linear inlets in southeastern Alaska is largely conjectural at this time. As noted previously, tsunami waves, under certain conditions, will travel up long inlets and attain fairly high runups. However, unless a hydraulic model of the inlet is constructed and studied in relation to the characteristics of tsunami waves, it is not possible to estimate potential wave height in any particular inlet. However, theoretical travel-time charts for tsunamis have been made for 17 towns in southeastern Alaska as part of the Alaska Regional Tsunami Warning System (Butler, 1971). The system is part of the Pacific Ocean Seismic Sea Wave Warning System that detects and locates major earthquakes in the Pacific region and issues appropriate warnings to the coastal population when tsunami waves might constitute a hazard.

Seiches

Seiches are periodic oscillations (standing waves) in lakes, bays, inlets, reservoirs, and rivers produced by changes in wind stress, atmospheric pressure, or by earthquakes. The periods of the standing waves may be the natural period of the basin, or harmonics of the natural period as controlled by the physical dimensions of the basin (McCulloch, 1966). Seiches produced by earthquakes (seismic seiches) are believed to be caused by the horizontal acceleration of short-period seismic surface waves (McGarr and Vorhis, 1968). Seiching also may be caused by tectonic tilting of the basin causing periodic wave oscillations (McCulloch, 1966). Seiche waves generally are generated in enclosed or nearly enclosed bodies of water.

Because seismic surface waves travel much faster than tsunami waves, seiche waves are operative in an area before the tsunami waves arrive, and therefore, commonly can be distinguished from them. Generally, they hit the shore within minutes after the earthquake and have largely subsided by the time the first tsunami wave has arrived. However, they are more difficult to differentiate from locally generated waves caused by submarine sliding.

Seismic seiches have been recognized as long ago as the great Lisbon, Portugal earthquake of Nov. 1, 1755, and have been described as an earthquake effect many times since then. Seiching during only two earthquakes will be discussed here: (1) the Hebgen Lake earthquake of 1959, and (2) the Alaska earthquake of 1964.

The Hebgen Lake earthquake of 1959 produced pronounced seiching in Hebgen Lake, a northwest-trending lake about 15 miles long, as much as 80 feet deep, and dammed at its northwest end. The lake area during the earthquake subsided differentially 22 feet (Myers and Hamilton, 1964). The water surface was warped so that water in a southeast arm of the lake was 5 feet higher than at the dam and 10-15 feet higher than in the main part of the lake a few miles southeast of the dam. As a result, a wave of water about 10 feet high was generated and overtopped the dam about 10 minutes after the earthquake. The first and highest wave was followed by a series of oscillations with gradually decreasing size for at least 12 hours after the earthquake. Wave periodicity was roughly estimated at about 10 minutes by eyewitnesses (Stermitz, 1964). Because of the large amplitude of the waves and their general violence, the waves were called "surges" by Myers and Hamilton (1964) rather than seiches. However, damage from the waves was relatively light. Although the earthdam (with a concrete-core wall) at the northwest end of the lake was overtopped by about 1.5 feet of water, it did not fail. However, gullies 3-4 feet deep were carved on the downstream face of the dam and the spillway was damaged. Some houses along the shore of the lake were floated off their foundations and boats and docks were washed away. The banks of Hebgen Lake slumped extensively into deeper water and greatly changed the bottom topography. How much of this slumping was due to bank undercutting from wave action and how much can be attributed to other earthquake effects is not known.

Seismic seiches caused by the Alaska earthquake of 1964 were of two kinds: (1) those caused by horizontal acceleration of water by seismic surface waves, and (2) those caused by tectonic tilting of a water basin.

Nearly all of the first type of seiche were recorded at teleseismic distance (600 miles or more from the earthquake epicenter). In the United States, where most of this type of seiche was noted, 763 of 6,435 surface-water gages of the U.S. Geological Survey registered seiches (McGarr and Vorhis, 1968). The greatest number of these gages that registered seiches were in States bordering the Gulf of Mexico. The seiches occurred on bodies of water having a wide range in depth, width, and rate of flow and included rivers, streams, lakes, reservoirs, ponds, and swimming pools. Waves as high as 6 feet were described in coastal regions of Louisiana and Texas (Donn, 1964). The maximum wave heights recorded on gages were (1) 0.87 foot in Missouri, (2) 0.68 foot in Louisiana, (3) 0.67 foot in Texas, and (4) 0.66 foot in Florida (McGarr and Vorhis, 1968). Seiching in the States bordering the Gulf of Mexico began between 30 and 40 minutes after the earthquake, or about the time that Love and Rayleigh types of surface seismic waves were passing through the area (Spaeth and Berkman, 1967). Wave periods generally were in the 5- to 15-second range. In general, the seiches having the highest amplitudes and longest durations (about 2 hours) were in reservoirs. Damage, although minor, was widespread in States bordering the Gulf of Mexico. Flooding occurred on both sides of a bayou in Louisiana, a number of boats broke loose in the same area, and several boats sank or were washed ashore. Mooring lines snapped and many barges were set adrift on a canal near New Orleans.

Seiches, caused by tectonic tilting of water basins and perhaps also in places by horizontal acceleration of water by seismic surface waves, were noted in several places in Alaska following the 1964 quake. They were best documented on several lakes. On Kenai Lake (about 20 miles north of Seward), at least nine distinct seiche waves, with wave periods between 1.40 and 36.36 minutes, were produced as a result of the lake being tilted so that the west end was about 3 feet lower than the east end (McCulloch, 1966). In constricted and shallow parts of the lake, seiche waves had 20- to 30-foot-runup heights. Although most of the damage along shore was due to slide-generated waves, minor additional damage was caused by seiching. Had the shore area been densely populated, damage would have been markedly higher. Seiche waves also were reported on Portage Lake, fronting Portage Glacier (about 30 miles southeast of Anchorage). Eyewitness accounts state that the ice on the lake first began to violently vibrate, then heave and crack; this was followed by about 5-foot-high oscillations of the ice, which continued for almost 2 hours (Waller, 1966a). Five-inch-thick ice on Tustumena Lake (western part of the Kenai Peninsula) reportedly cracked markedly along the shore and the lake level oscillated about 2 feet. Seiche action was noted in several other large lakes.

In attempting to evaluate the potential effects of seiching in southeastern Alaska, it should be noted that there are only a few lakes of significant size that are near enough to centers of population to affect directly those centers. However, as more facilities are built along lake shores for recreational or other purposes, the potential effects from seiching can be expected to increase. Whether significant seiching can take place in bodies of water that are open at one end, such as inlets and bays, cannot be ascertained by us. However, answers to questionaires submitted by the U.S. Geological Survey to postmasters and others suggest that seiching did take place in some of the long narrow inlets as a result of the Alaska earthquake of March 27, 1964. In at least four places in southeastern Alaska abnormal waves, as much as 5 feet high and possibly higher, were reported immediately following the earthquake and more than an hour before the arrival of the first tsunami waves.

Other abnormal waves

Local waves generated by submarine sliding or by subaerial landsliding can be highly destructive because they generally hit the shore suddenly during or immediately after an earthquake and because their occurrence and runup height at a particular locality is largely unpredictable. Therefore, a warning system, such as for tsunamis, is not possible.

As noted previously, a gigantic rockslide descended into an arm of Lituya Bay during the Lituya Bay earthquake and generated the highest wave runup ever described. This huge wave caused water to surge up the opposite wall of the inlet to a height of 1,740 feet (D. J. Miller, 1960). It then moved toward the mouth of the bay at a speed of 97-130 miles an hour, stripped soil and vegetation from 4 square miles of shore, and sank two of three fishing boats. Dating of trees in the area and configuration of the trimlines indicate that giant waves had been generated previously in the bay and had high runups in the years 1853 or 1854, about 1874, and 1899 (D. J. Miller, 1960). The 1899 wave probably was generated by a landslide triggered by the large Yakutat earthquake of September 1899. Still other waves between 1899 and 1958 can be documented from tree-dating and trimline studies.

Local slide-induced waves struck violently at a number of places during or immediately after the Alaska earthquake of March 27, 1964, and were the major cause of loss of life and property. Most of the known waves of this kind originated along the shores of fiords and bays in northern and western Prince William Sound, in bays on the south coast of the Kenai Peninsula, and along the shore of Kenai Lake. A map showing location, direction of wave motion, and height of runup of the larger waves in the Prince William Sound area was prepared by Plafker, Kachadoorian, Eckel, and Mayo (1969). The waves, which had runups of as much as 70 feet at Chenega (southwest side of Prince William Sound) and of more than 170 feet in several other inhabitated parts of the sound, caused most of the drownings in these areas. Highly destructive waves also hit Seward, Valdez, Whittier (about 50 miles southeast of Anchorage), and other communities.

The town of Chenega was virtually destroyed and 23 of its 75 inhabitants were lost (Plafker and others, 1969). Sixty to 90 seconds after shaking started, the first wave arrived and traveled about half way up the beach. It then receded rapidly and exposed the entire cove at the edge of town to a depth of 20 feet. Four minutes after the shaking started, a second wave ran up to a height of 70 feet. All the buildings (except one house and a school) were carried inland into the trees and smashed or were swept out to sea with the backwash.

At Seward, slide-generated waves probably did not attain the height of tsunami waves that followed but, because they were the first waves to strike shore, they caused most of the damage and loss of life (Lemke, 1967). Thirty to 40 seconds after violent shaking from the earthquake began, large-scale landsliding began along the Seward waterfront. A large mound of water was formed out in the bay owing to the sudden displacement of water near shore. Waves moved out in all directions from this mound and overran the shore in several places to heights of 30 feet. Two miles south of Seward, a nearly completed marineway was completely destroyed and heavy earthmoving equipment and other objects were smashed against trees. Along the Seward waterfront, a large swell broke over the railroad dock area lifting flatcars off the tracks and picking up several vehicles. Burning oil, carried from exploding oil tanks by waves, set fire to a freight train whose last 40 cars were filled oil tankers. This resulted in a chain-reaction of exploding cars down the tracks and of oil tanks along the shore. As wave crests reached the small-boat harbor, moored boats were lifted over a breakwater; others were driven wildly to and fro. A large dock was lifted by a wave and then dumped into the bay. Houses and other buildings were swept from the waterfront. Water continued to swirl and slosh back and forth in the bay for at least 15 minutes before the first tsunami wave arrived about 25 minutes after the earthquake.

At Valdez, sliding of a strip of deltaic deposits, 4,000 feet long and 600 feet wide, created a highly destructive wave (Plafker and Mayo, 1965; Coulter and Migliaccio, 1966). Shortly after the sliding, a "boil" of muddy water appeared offshore, and a wave 30-40 feet high rushed shoreward to inundate what remained of the waterfront and other low parts of town. Effects of the sliding and of wave action resulted in the deaths of 30 people and the destruction of all docks and associated facilities. A second wave, presumably a return surge with a runup nearly as high as the first wave, crossed the Valdez waterfront 10 minutes after the first wave and carried a large amount of floating debris. Much larger slide-induced waves were generated near Shoup Bay (about 10 miles west of Valdez) at about the same time and caused heavy damage along considerable lengths of coastline (Plafker and Mayo, 1965). At one locality, high-water marks reached 170 feet above mean sea level (Coulter and Migliaccio, 1966).

Multiple submarine slides at and near Whittier (about 50 miles southeast of Anchorage) triggered a series of waves, which inundated much of the shoreline at the head of Passage Canal to a maximum height

of 104 feet above mean lower low water (Plafker and Mayo, 1965). The first of three waves at Whittier, which was described as "glassy," arrived about 1 minute after the earthquake started and reached a height of 25-26 feet near town. A second wave, 1-1.5 minutes later and described as a breaking wave, carried much debris and reached a height of 40 feet at the railroad depot. A third wave, 0.5-1 minute after the second wave was a breaking wave and reached an altitude of about 30 feet near the depot (Kachadoorian, 1965). Thirteen people lost their lives as a result of wave action; in addition, lumber company buildings, the small-boat harbor, a cargo-barge dock, and several homes were destroyed.

Earthquake-induced slides from nine delta fronts in Kenai Lake caused two kinds of destructive local waves: (1) a "backfill" type of wave formed by water that rushed toward the delta to fill the void left by the sliding mass, and which overtopped the scarp and ran inland, and (2) a "far-shore" type of wave, which hit the opposite shore from the delta (McCulloch, 1966). Some backfill waves had runup heights of 30 feet and ran inland more than 300 feet, where they uprooted and broke off large trees. One far-shore wave had a runup height of 72 feet; trees were shorn of their limbs and bark and some, as large as 2.5 feet in diameter, were uprooted or broken off. A log house, one of the few manmade facilities along the shoreline, was carried more than 200 feet by a backfill wave.

Slide-generated waves probably would have a higher destructive potential in southeastern Alaska than either tsunami waves or seiche waves because of their possibly higher local runups and because they can hit the shores almost without warning during or immediately after an earthquake. The strong indication that slide-induced waves have been generated a number of times in the Lituya Bay area portends future abnormal wave action in that area. Other places where topographic conditions favor earthquake-induced sliding, particularly where there is active faulting, also should be considered as sites for the generation of future local waves.

Plafker (1969) suggested that water disturbances might have been generated in some lakes, fiords, and rivers during the Alaska earthquake of 1964 by inertial effects of the water bodies as the land mass was displaced horizontally beneath them. Such horizontal movement of the land, if it occurred fast enough, would tend to result in the water being piled above its original level along shores on the side of the basin opposite to the direction of displacement and at the same time to be lowered along shores in the direction of displacement. Plafker postulated that sudden rises of water level and highly destructive waves observed at numerous coastal localities during or immediately after the Alaska earthquake where there was no evidence of submarine sliding, strong tilting, or faulting might have been caused in this manner. We cannot evaluate the probability of waves being generated in southeastern Alaska in this manner or estimate their potential destructiveness.

REGIONAL FACTORS BEARING ON THE EVALUATION OF GEOLOGIC HAZARDS OTHER THAN THOSE CAUSED BY EARTHQUAKES

Nonearthquake-related geologic hazards, although generally far less dramatic than those related to earthquakes, tend to occur so much more frequently or persistently that their aggregate effects can be significant. Three kinds of geologic hazards in this category and their possible effects in southeastern Alaska are discussed briefly here: (1) effects of nonearthquake-induced landsliding and subaqueous sliding, (2) effects of flooding, and (3) effects of land uplift. More detailed discussions will be given in the reports of specific coastal communities in southeastern Alaska.

Nonearthquake-induced landsliding and subaqueous sliding

The potential for nonearthquake-triggered landsliding in southeastern Alaska ranges widely from place to place, depending upon such factors as differences in topography, precipitation, tree cover, rock and soil types, and tectonic structure. In a general way, past sliding furnishes the key to predicting where and in what materials future sliding will occur. However, it usually is not possible to predict when sliding will occur or the exact location.

Fast-moving rockslides, debris slides, and mudflows can be expected to occur from time to time on steep slopes. If the bedrock on a precipitous slope is strongly jointed or fractured and particularly if the process of freeze and thaw is operative, large masses of rock may break loose suddenly and be highly destructive to such manmade facilities as highways, power plants, pipelines, buildings, and other facilities on the slope or at the base of the slope. On the other hand, the slow downslope movement of talus, now taking place on many steep slopes, generally can be expected to continue at about the same rate unless conditions are changed by man. Snow and debris avalanches can be especially hazardous during winter months. Sequential avalanches commonly follow paths downslope that previous avalanches followed. Thus, the presence of preexisting avalanche-cone deposits at the base of slopes and of slopes stripped of vegetation or covered by young growth afford valuable clues as to where future avalanches might descend, and therefore, of areas to be avoided for building sites or other works of man. Also, moderately fast-to fast-moving mudflows and debris slides may develop in surficial deposits after periods of high precipitation. These commonly can move even on fairly gentle slopes and, if of large size, can cause heavy loss of life and property damage.

Man, by imposing change on his geologic environment, commonly accelerates or enlarges naturally occurring landslides. For example, landslides that have long been inactive may be triggered into renewed activity, or new landslides may be created by the building of a highway or a street through or at the base of a slide mass or by other man-induced modifications. Also, accelerated slope erosion and debris flows may follow large-scale clearing and cutting of timber. Swanston

(1969), who has made an extensive study of this cause-and-effect relation in southeastern Alaska, states that the erosion occurs mainly as mass movements of soil associated with steep slopes and high water levels in the soil. He noted more than 3,800 large-scale debris avalanches and debris flows in southeastern Alaska. Although most of these slides are the direct manifestation of natural mass wastage and slope reduction, some are the direct result of logging and logging-road construction. More sliding of this type can be expected as additional logging and clearing take place on steeper slopes.

Whether or not damaging waves might be created by nonearthquake-induced landslides moving down steep slopes into bodies of water, such as inlets, bays, lakes, and reservoirs, is difficult to evaluate. To our knowledge there have been no destructive waves generated in this manner in southeastern Alaska. However, the presence of fresh-appearing scars extending in places from the top of an inlet wall to the water's edge shows that fairly large rockslides have occurred there recently. Conceivably a sufficiently large slide-generated wave might be produced sometime in the future that would damage nearby or opposite-shore facilities. A large wave generated in a reservoir impounded behind a dam might cause the dam to fail. Or, the wave might overtop the dam and cause great loss of life and property below the dam such as occurred during the giant rockslide in the Vaiont valley in Italy in 1963.[11]

Subaqueous sliding can be expected to continue in those places where sliding has been most prevalent in the past. Thus, fronts of large deltas and other underwater slopes that have become oversteepened by normal depositional processes are the most susceptible to future sliding. That sliding from nonearthquake-triggered causes has been taking place off the front of the Katzehin River delta (5 miles southeast of Haines), and in nearby inlets, is indicated by cable breaks in those areas at times when no earthquakes have been recorded or felt. Fronts of other large deltas also probably will slide from time to time. Sliding in these places may disrupt underwater facilities and possibly create navigational problems. Local small-scale sliding also can be expected on steep underwater slopes along fiord walls where there may be perched bodies of glacially derived sediments. Such materials can fail suddenly when offshore structures, such as docks, are built upon them.

Flooding

Floods are common in southeastern Alaska. As noted previously, precipitation ranges from about 160 inches annually in the southeastern part of the region to about 30 inches at Skagway in the northeastern part. The precipitation runs rapidly off the steep slopes, causing

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[11] A rock mass having a volume that exceeded 240 million m^3 (312 million cubic yd) slid into the reservoir behind the Vaiont Dam and created the worst dam disaster in history (Kiersch, 1964). Waves of water swept over the dam to a height of 100 m (328 feet). Although the dam did not fail, everything was destroyed by flooding for miles downstream and nearly 3,000 lives were lost.

Alaska's highest known flood-discharge rates (Childers, 1970). Temperatures generally are mild, resulting in floods even in winter when combined snowmelt and rainfall occasionally produce exceptionally large floods. Both the short swift streams flowing off the steep mountain fronts and the larger through-going rivers are subject to flooding. Sudden draining of two glacier-dammed lakes in British Columbia, one in the Salmon River valley (Portland Canal area) and another 40 miles northeast of Juneau, have caused spectacular downstream floods in Alaska with very high peak discharges.

Flooding has caused heavy damage in the past in southeastern Alaska and can be expected to do so in the future, unless more remedial measures are taken. Roads and other facilities have been constructed or are being planned along rivers and streams where flood hazards exist. Flood damage to the city of Skagway, built on the flood plain of the Skagway River, has been high on several occasions. However, construction of dikes has alleviated to some extent the danger of flooding. If new communities and associated facilities, such as roads, are built on flood plains or present towns are extended onto flood plains, the hazard from flooding can be expected to increase.

Land uplift

Current uplift of land in most of southeastern Alaska, although probably not affecting man significantly in a short period of time, may have some adverse long-term effects. As noted previously, differential uplift of land in southeastern Alaska is presently taking place at a rate ranging from 3.9 cm (1.5 in.) per year in the Glacier Bay area to approximately zero just south of Ketchikan (fig. 2). Also, as discussed previously, it is unlikely that the rate of uplift during the next 50 years will be appreciably different than the present rate. Thus, for example, it can be expected that land in the Glacier Bay area will be uplifted 6.5 feet (in respect to sea level) during the next 50 years. These long-term effects should be borne in mind when consideration is given to constructing facilities on or near shore, such as docks and boat harbors, where there is a critical relation between heights of land and water.

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Figure 1.--Map of southeastern Alaska and adjacent Canada showing pertinent geographic features. CRV, Chilkat River Valley.

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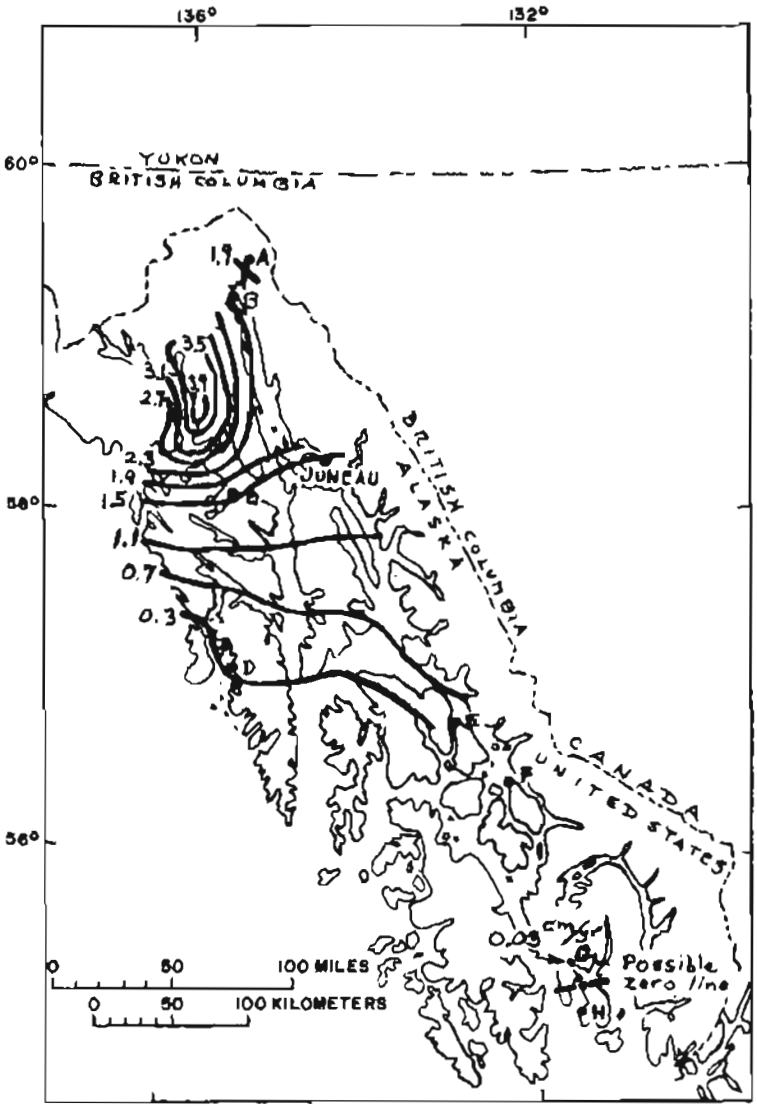
PART 2 of 2



EXPLANATION

Contour interval, 0.4 centimeters/year

cm	Ft
0.3	0.01
0.7	0.02
1.1	0.04
1.5	0.05
1.9	0.06
2.3	0.08
2.7	0.09
3.1	0.10
3.5	0.11
3.9	0.13

Possible zero line based on interpretation of data from Metlakatla by Woodworth and Haight (1927)

A SKAGWAY
B HAINES
C HOONAH
D SITKA
E PETERSBURG
F WRANGELL
G KETCHIKAN
H METLAKATLA

Figure 2. Rate of land emergence of part of southeastern Alaska relative to present sea level. Modified slightly from Hicks and Shofnos (1965).

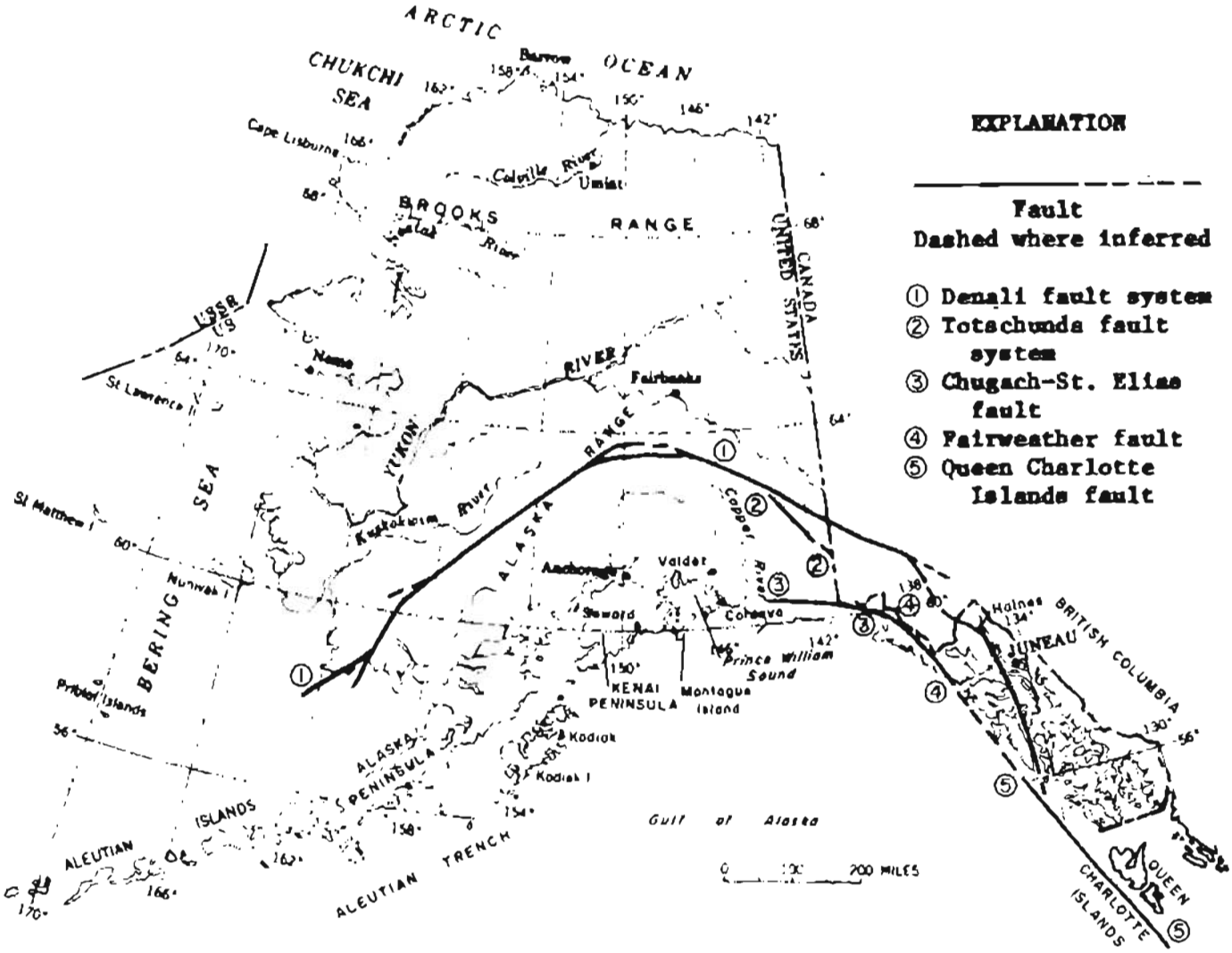


Figure 3.—Map of Alaska showing major elements of the Denali and Fairweather-Queen Charlotte Islands fault systems. Modified from Grantz (1966), Tobin and Sykes (1968), Plafker (1969); 1971), Richter and Matson (1971).

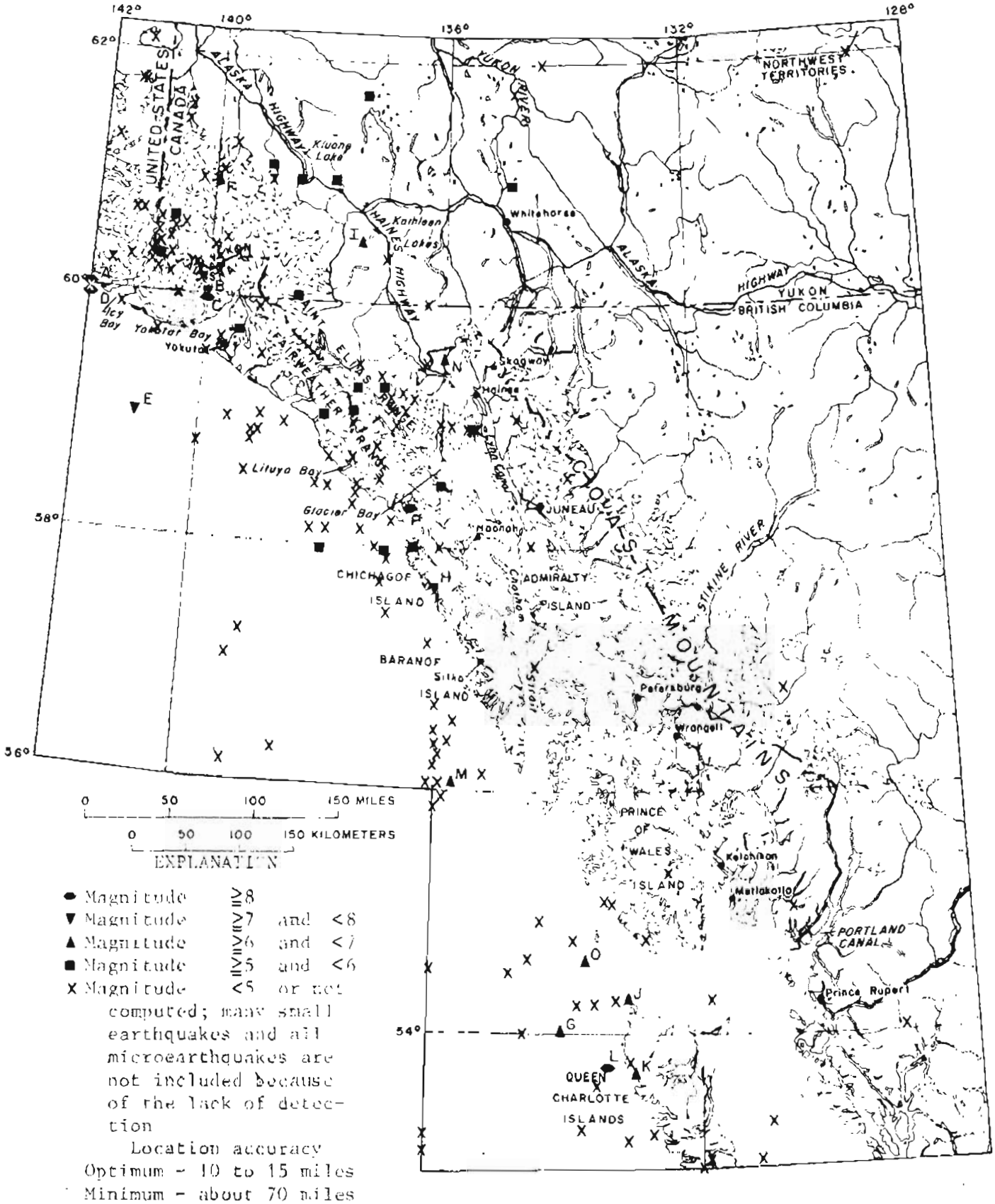


Figure 1.--Map showing locations of epicenters and approximate magnitude of earthquakes in southeastern Alaska and adjacent areas for period 1899-1969 inclusive. Data from Canada Dept. of Energy, Mines and Resources, Seismological Service (1953, 1955, 1956, 1961-63, 1966, 1969, 1970), Davis and Echols (1962), International Seismological Centre (1967-1970), Milne (1963), Tobin and Sykes (1968), U.S. Coast and Geodetic Survey (1930-1969, 1964-1970, 1969), and Wood (1966)

Dates and magnitudes of some earthquakes of magnitude ≥ 6

Designation on map	Date (Universal Time)	Magnitude
A	September 4, 1899	8.2-8.3
B	September 10, 1899	7.8
C	September 10, 1899	8.5-8.6
D	October 9, 1900	8.3
E	May 15, 1908	7
F	July 7, 1920	6
G	April 10, 1921	6.5
H	October 24, 1927	7.1
I	February 3, 1944	6 1/2
J	August 3, 1945	6 1/4
K	February 28, 1948	6 1/2
L	August 22, 1949	8.1
M	October 31, 1949	6 1/4
N	March 9, 1952	6
O	November 17, 1956	6 1/2
P	July 10, 1958	7.9-8.0

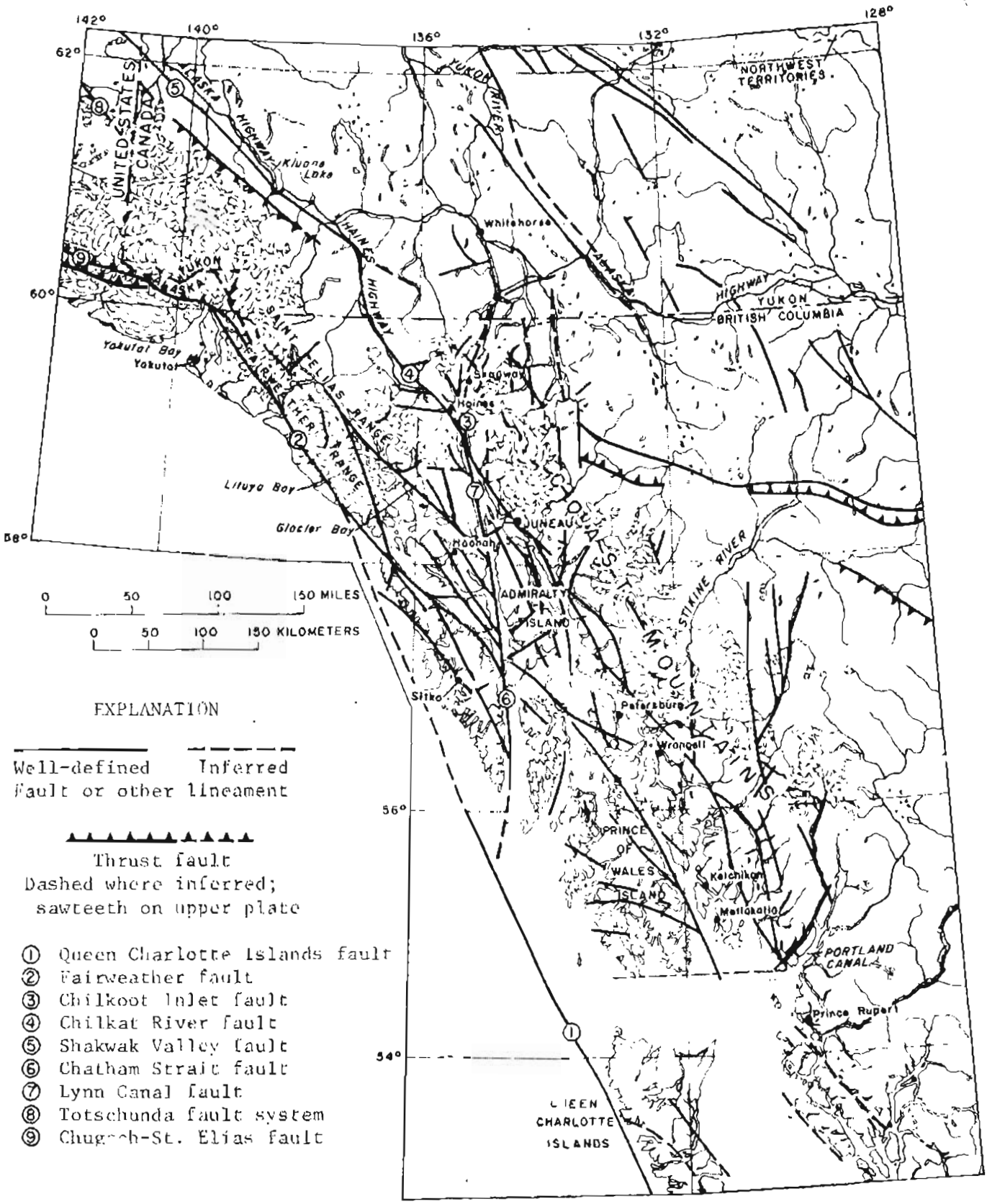


Figure 5.--Map of southeastern Alaska and adjacent Canada showing major faults and selected other lineaments interpreted to be probable or possible faults, shear zones, or joints. Taken from St. Amand (1957), Twenhofel and Sainsbury (1958), Gabrielse and Wheeler (1961), Brew and others (1966), Tobin and Sykes (1968), Geological Survey of Canada (1969a; 1969b), King (1969), Plafker (1969), Souther (1970), Plafker (1971), and Richter and Matson (1971); and with additions and modifications by the writers.

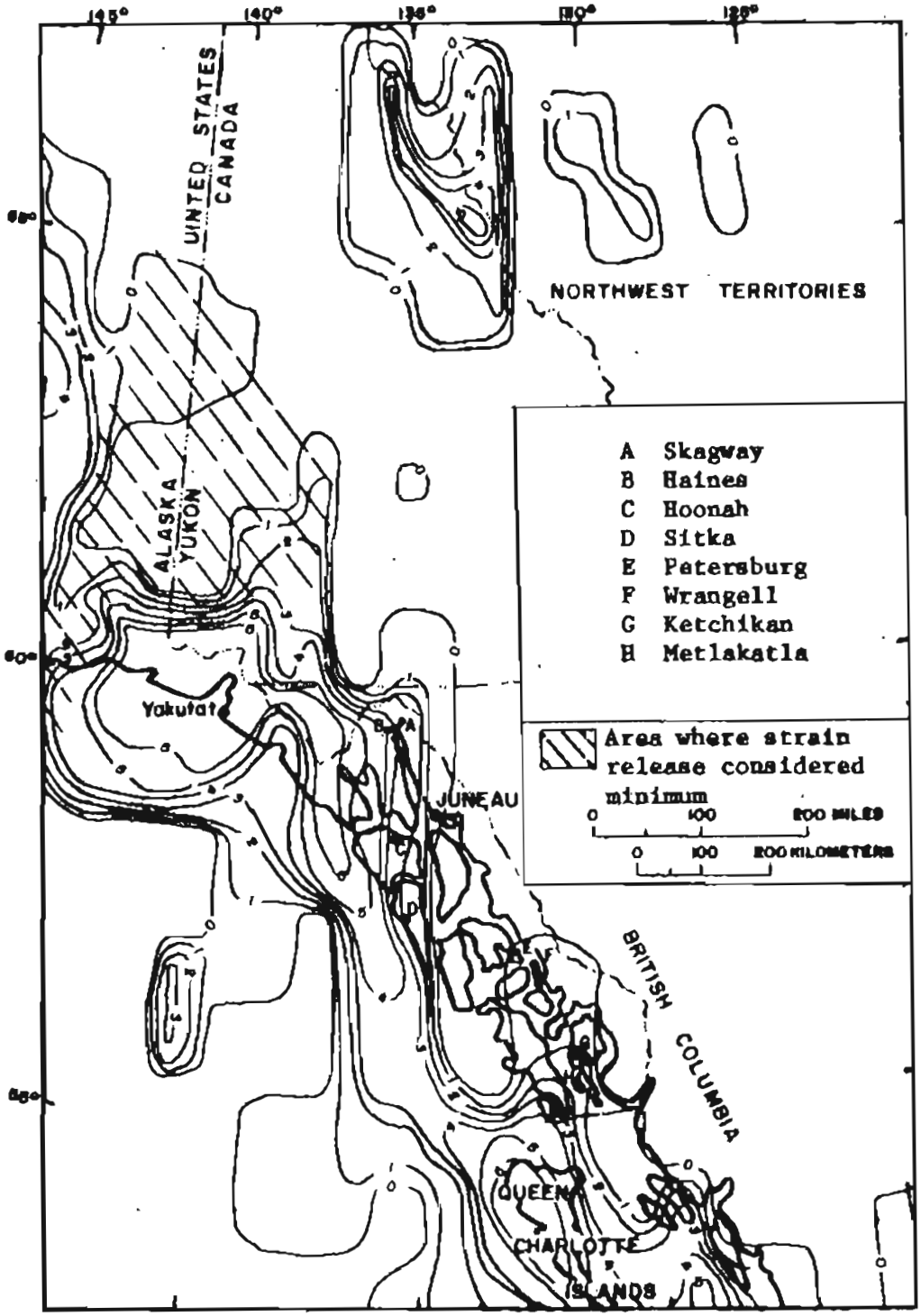


Figure 6.--Strain-release map of seismic energy 1898-1960, inclusive, in southeastern Alaska and part of adjacent Canada with explanation showing interpreted frequency of energy release. Modified from Milne (1967).

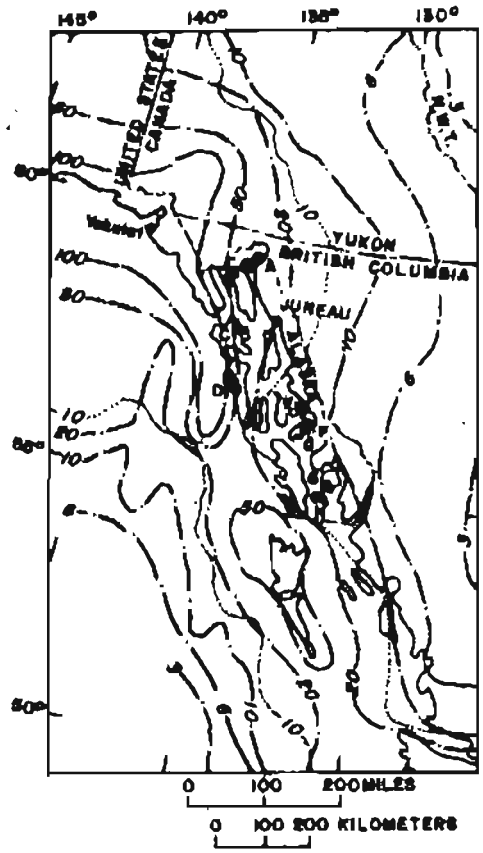
EXPLANATION FOR FIGURE 6

Map contour	Energy level in strain release units[1]	Interpreted frequency per 100 yrs of certain magnitudes (M)[2] necessary to release all of energy level				Interpreted magnitude necessary to release all of energy level in a single event per 100 yrs
		M 5	M 6	M 7	M 8	
0	0.1	---	----	----	----	3.7
1	1	1	----	----	----	5.0
2	5	5	----	----	----	5.9
3	10	10	1.8	----	----	6.3
4	20	20	3.5	----	----	6.7
5	50	50	8.9	1.6	----	7.3
6	100	100	17.8	3.1	----	7.7
[3] 7	200	200	36	6.5	1.03	8.1
8	500	500	90	15	2.5	8.6
9	700	700	120	21	3.5	8.7

[1]Energy level, strain-release (Benioff, 1951) unit here defined in terms of energy of a magnitude 5 earthquake ($10^{1.5(M-5)/2}$) per area (10^4 km^2) based on earthquakes 1898-1960 inclusive, extended to a 100-year base.

[2]A one-unit increase in magnitude is about a 30-fold increase in energy release and a two-unit increase is a 900-fold increase (Steinbrugge, 1968).

[3]Northern area of contour 6 has a maximum energy of 700 strain-release units; southern area of contour 6 has 236 units. Contours 7, 8, and 9 are not shown on map; tabular data for 7, 8, and 9 have been extended by the writers.



EXPLANATION

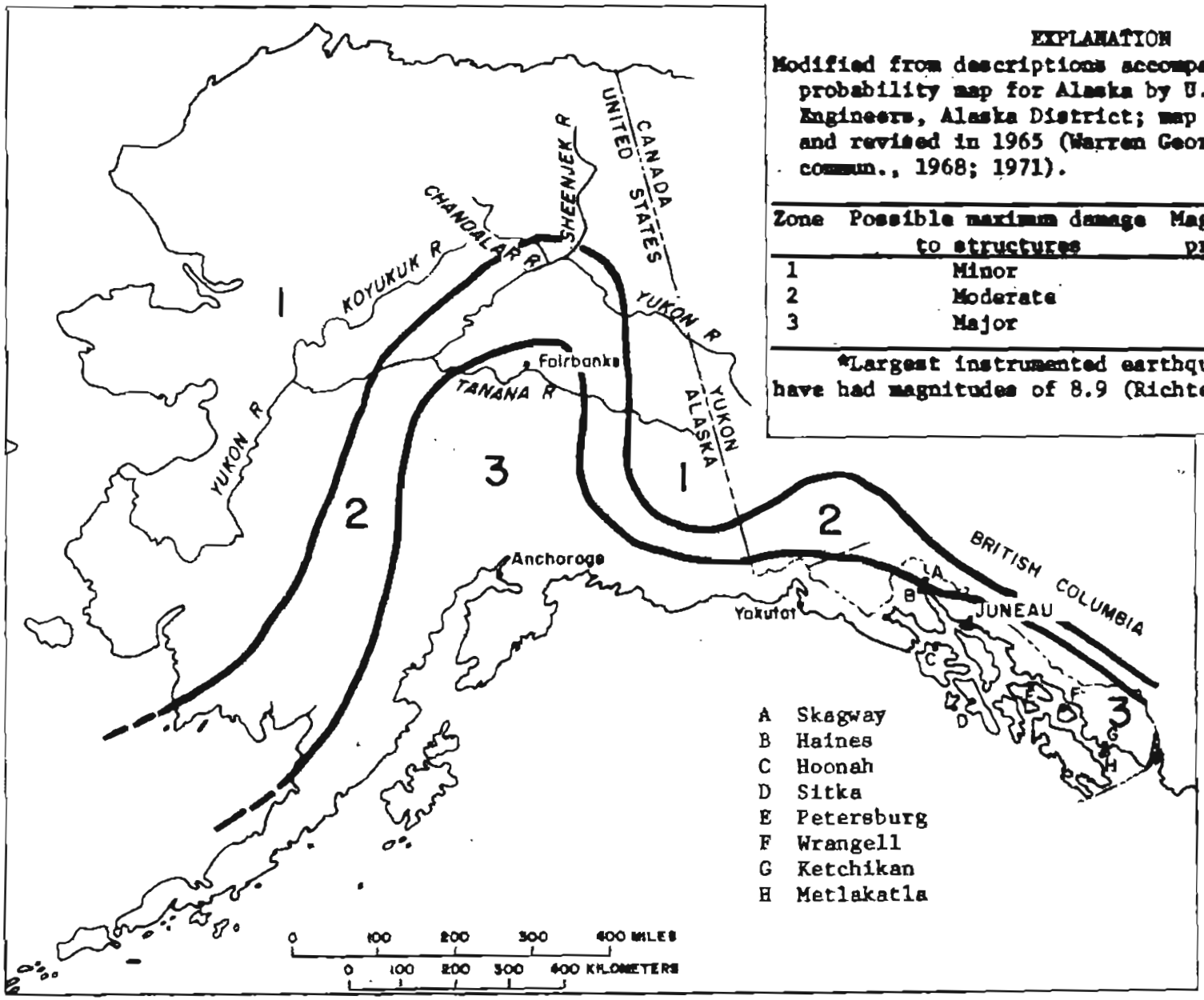
Average-value method	3 6 10 20 50 100	Extreme-value method

Contours showing peak earthquake acceleration as a percent of gravity (about 982 cm/sec^2 for southeastern Alaska at sea level*)

A	Skagway	E	Petersburg
B	Haines	F	Wrangell
C	Hoonah	G	Ketchikan
D	Sitka	H	Metlakatla

*See table 1 showing relations between acceleration units, energy, magnitude, and intensity.

Figure 7.—One-hundred-year probability map showing peak earthquake accelerations for southeastern Alaska and part of adjacent Canada. Modified from Milne and Davenport (1969). Based upon earthquake strain release from 1898-1960 (extended to a 100-year interval) as interpreted by an extreme-value method and using data from all instrumented earthquakes. For comparison of method, another interpretation is offered through an average-value method (dotted contour on map) which uses only earthquakes having an acceleration of 10 percent gravity.



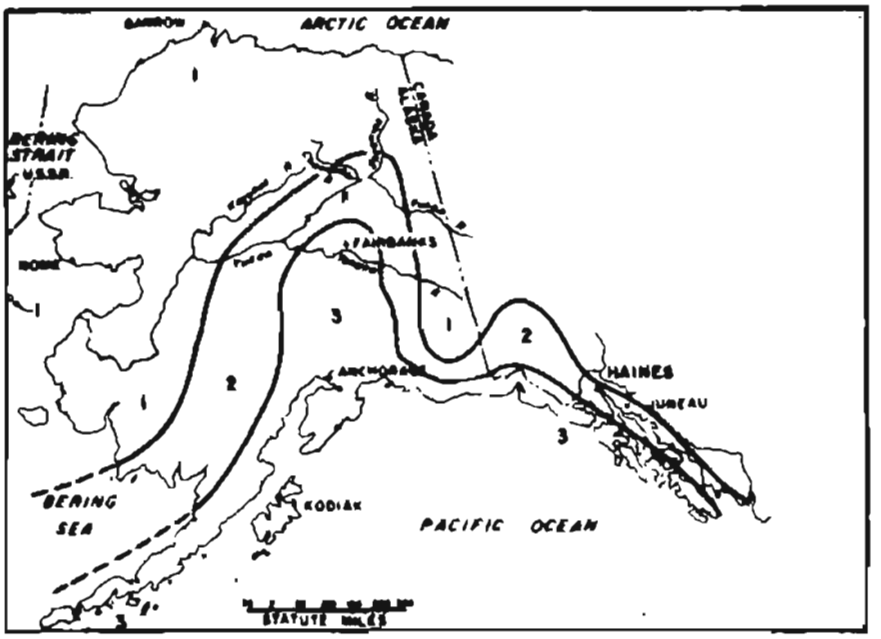
EXPLANATION

Modified from descriptions accompanying the seismic probability map for Alaska by U.S. Army, Corps of Engineers, Alaska District; map prepared in 1937 and revised in 1965 (Warren George, written commun., 1968; 1971).

Zone	Possible maximum damage to structures	Magnitude of largest probable earthquake
1	Minor	3.0–4.5
2	Moderate	4.5–6.0
3	Major	>6.0*

*Largest instrumented earthquakes of the world have had magnitudes of 8.9 (Richter, 1958).

Figure 8.--Seismic probability map for most of Alaska as modified from U.S. Army Corps of Engineers, Alaska District.



ZONE 1 - Minor damage: distant earthquakes may cause damage to structures with fundamental periods greater than 1.0second; corresponds to intensities V and VI of the MM* Scale

ZONE 2 - Moderate damage: corresponds to intensity VII of the MM* Scale

ZONE 3 - Major damage: corresponds to intensity VIII and higher of the MM* Scale

*Modified Mercalli Intensity Scale of 1931

Figure 9. Seismic zone map of Alaska. Modified from the 1970 edition of the Uniform Building Code (International Conference of Building Officials, 1970).

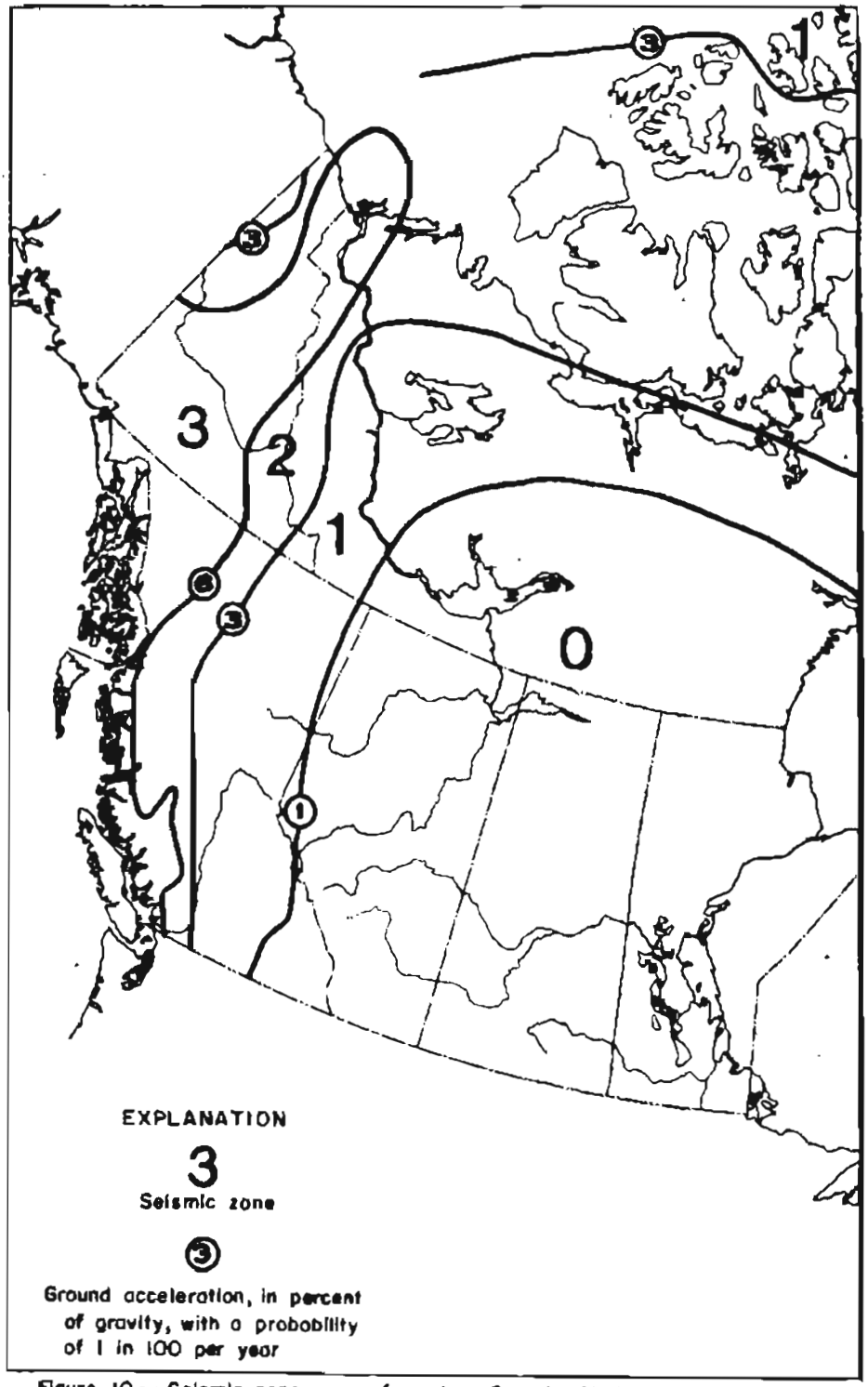


Figure 10.--Seismic zone map of western Canada. Modified from National Research Council of Canada (1970).