Submarine Landslides: Selected Studies in the U.S. Exclusive Economic Zone

Edited by W.C. SCHWAB, H.J. LEE, and D.C. TWICHELL

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FOREWORD

With each passing year, the United States becomes increasingly dependent upon resources from the oceans. Historically, the sea has provided food resources, recreational resources, and a defensive barrier in times of war. During the last few decades, we have turned to the sea to provide other resources such as oil, gas, and construction aggregate which had been almost exclusively extracted from onshore areas. With an ever-growing population demanding more resources to sustain a high quality of life, the United States will have to turn more frequently to the oceans to meet its increasing requirements for fresh water, energy minerals, industrial minerals, agricultural minerals, waste disposal, recreational diversions, military exercises, and other activities. All this must be accompanied and balanced with environmental concerns.

Recognizing the growing importance of the oceans to the future of the United States, President Reagan in 1983 expanded the United States to include those parts of the oceans extending 200 miles off our shores. This enormous new territory, the Exclusive Economic Zone (EEZ), is over 3 million square nautical miles in size which is about 30 percent large than the land area of the United States.

The United States Geological Survey (USGS) through its Offshore Geologic Framework program was directed by the President and the Congress to explore the sea floor of this new frontier. In partnership with colleagues from other Federal agencies, the States, academia, industry, and scientists from the international community, the USGS began a program in 1983 to map the EEZ. To date, these reconnaissance-scale studies have mapped approximately 2 million square nautical miles of the EEZ and unveiled a fascinating, but extremely complex, sea floor. These discoveries hold the promise of potential resources that ultimately will help meet the needs of all Americans. This promise will be realized, however, only through continued research, technology advancement, and informed decision making.

Although our studies discovered significant potential resources, they also identified geologic processes that not only threaten the availability of these potential resources but also threaten human life directly. This volume is dedicated to increasing the understanding of one such process—submarine landslides. The authors focus on documenting our new understanding of the geologic processes involved in submarine landslides and the consequences of these processes.

Submarine landslides fall within a general category marine geologists call geohazards. To evaluate fully the significance of any geohazard, it is necessary to develop an understanding of the basic geological processes involved as well as the geographic distribution, frequency of occurrence, scale, magnitude, and consequences of the processes. All these topics and more are covered as the authors share their insights into submarine landslides occurring in diverse geologic settings throughout the EEZ from New England to Hawaii, from the arctic to the tropics, and from shallow water to deep ocean basins.

Without question, submarine landslides involve significant geologic processes which must be understood and documented for policy makers to make informed decisions as to when and how the marine resources within the United States EEZ are to be exploited. This volume is a substantial and unique contribution to expanding the knowledge base of submarine landslides.

Gary Hill

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Submarine Landslides: An Introduction

By H.J. Lee, W.C. Schwab, *and* J.S. Booth

INTRODUCTION

Consider these events: In the late afternoon of March 27, 1964, one of the most violent earthquakes of all time rocked southern Alaska. In the community of Valdez, shortly after the shaking began, a cargo ship unloading at the town dock began to toss violently, with vertical motion in excess of 10 m (meters) and rolls of 50°. Next, the docks vanished into turbulent water. The entire waterfront followed, along with warehouses, a cannery, heavy equipment, and 30 people. Ten-meter waves propagated back and forth across the bay on which Valdez is situated, surging over 500 m inland and flooding much of the town. For months following these catastrophic events, the shore area near Valdez continued to subside.

In August 1969, Hurricane Camille, one of the century's most intense hurricanes, struck the coast of Louisiana. At an oceanographic station near the South Pass portion of the Mississippi Delta, wave heights of 21 to 23 m were recorded before the instruments used to measure wave height failed. Three offshore drilling platforms were badly damaged or destroyed. Later, one of the platforms was found, half-buried in mud and displaced 30 m downslope.

On November 19, 1929, a major earthquake occurred on the continental slope south of the Grand Banks, off the eastern coast of the United States and Canada. During the earthquake, many submarine communications cables failed near the earthquake epicenter. For 13 hours following the earthquake, cables continued to fail in a systematic sequence: with the passage of time, cable breaks occurred farther and farther to the south.

On the Hawaiian island of Lanai, limestone-bearing gravel blankets the coastal slopes and extends to an altitude of 326 m. The characteristics of the gravel show that it originally formed at sea level. The island is sinking so rapidly that the deposit could not have been deposited during previous worldwide highstands of sea level. Most probably, the deposit was placed by a giant wave.

These four examples share a common element: they all show the devastation that can result from *submarine landslides*, a general term used to refer to downslope movement of sea-floor material en masse (also termed "mass movement"). At Valdez, a large portion of a submarine delta-front failed during the Alaska earthquake and slipped into the bay, carrying with it the town's waterfront and generating massive waves. During Hurricane Camille, the large waves generated by the storm loaded the weak sea-floor muds, causing them to fail and move downslope. On the Grand Banks, the magnitude 7.2 earthquake caused slope failures on the continental slope and rise. Some of these slope failures transformed into high-velocity turbidity currents that traveled over 720 km (kilometers) from the source area, breaking communication cables along the way. Off the coasts of the Hawaiian Islands, the remains of giant slope failures have been found using a modern side-looking sonar system. Some are more than 200 km long with volumes up to 5,000 km^3. The great displacement of the landslide blocks from their original source suggests that some of the failures were cataclysmic events, capable not only of removing landmass from the islands' coastlines but also of generating tremendous sea waves. One of these giant landslides may have been the cause of the wave that carried gravel from sea level up to an elevation of 326 m on the island of Lanai.

Most submarine landslides are not so spectacular. In fact, most undersea slope failures probably occur without any person being aware of them, but as man reaches farther into the oceans to extract oil and gas and mineral resources, searches for environmentally safe areas to dispose of waste, and occupies more coastal areas, the potential impact of undersea landslides on humanity increases. Major loss of property and life has already occurred; more losses may occur in the future. Even when particular landslides have no direct impact on man's development, they are still of basic interest to geologists in understanding how sediment and rock are transported to deeper water. Mass instability acts as a fundamental process in the formation and modification of continental and insular margins. The occurrence of such features tells us about the environments in which they form and allows us to use fundamental geologic knowledge to exploit resources and increase our knowledge of the importance of environmental hazards such as earthquakes and hurricanes.

The purpose of this report is to introduce the subject of submarine landsliding through selected studies that have been partly or fully supported by the U.S. Geological

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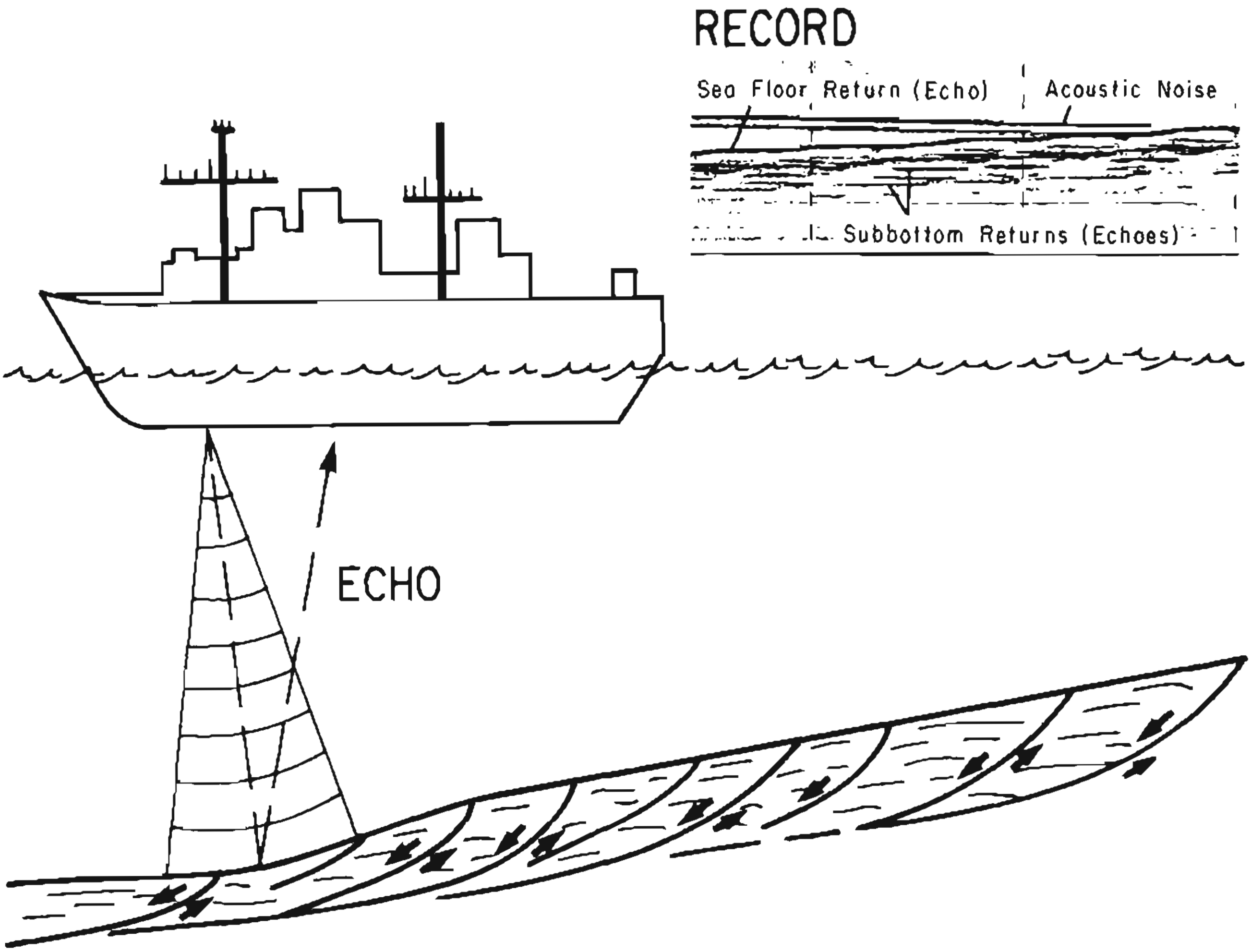

Figure 1. High-resolution profiling technique. The ship is crossing a series of slump blocks (arrows showing the direction of relative movement). The record in the upper right is a real profile showing what this idealized mass movement looks like on a seismic-reflection profile.

Survey. The report is not meant to be comprehensive but rather is an attempt to provide a cross section of the type of research going on within the academic, government, and ocean-engineering communities. The selected studies clearly demonstrate that marine investigators are just beginning to appreciate that submarine landslides are a common, widespread feature of the U.S. Exclusive Economic Zone (EEZ) and impose many constraints on the future of sea-floor utilization. The title of each section reflects the major point exemplified by the landslide being described. The sections are mainly visual in format and are intended for individuals with a general interest in the offshore areas of the United States, either from a basic scientific or from a seabed utilization point of view.

METHODS FOR FINDING SUBMARINE LANDSLIDES

Submarine landslides are occasionally recognized when they directly involve the shoreline. The catastrophic failure at Valdez is one of the best examples of this type. Other landslides disrupt or destroy offshore facilities, for example, the offshore platform lost during Hurricane Camille or the submarine cables broken off the Grand Banks. Some landslides are identified when major changes in water depth or sea-floor relief are observed, perhaps by fishermen or divers. The detection and recognition of most submarine landslides, however, relies on remote observation of sea-floor and subbottom morphology. That is, does the sea floor have the appearance of slope failure? Are features observed that look like landslides that have been seen on shore?

Development in offshore remote-sensing technology over the last 3 to 4 decades (such as precision depth recorders, high-resolution seismic-reflection systems, and sidescan-sonar systems) has allowed marine investigators to determine the appearance of the sea floor accurately and map areas of landslides. The basic principle of seismic-reflection techniques involves detection of sediment stratigraphy and structure as indicated by variations in acoustic-reflection characteristics (fig. 1). In general, the higher the frequency of the signal sent out by the shipboard seismic system, the higher the resolution and the smaller the depth of penetration below the sea floor. For example, a seismic system that uses an outgoing signal with a frequency of 12 kHz (kilohertz) is typically used as a precision depth recorder (high resolution with little, if any, subbottom penetration) while a 3.5-kHz system, although yielding a slightly lower resolution, is capable of penetrating over 100 m of soft sediment and defining subbottom landslide geometry. High-power, low-frequency systems (for example, air guns and sparkers) can penetrate up to 15 km beneath the sea floor and resolve reflecting surfaces 100 to 200 m apart.

Probably the most significant recent development in mapping sea-floor morphology is the evolution of sidescan-sonar systems (fig. 2). The principle of sidescan sonar is relatively simple. An instrument towed by a research vessel sends out two sound beams which are directed to either side of the towed vehicle. The sound beams are tuned to create a beam that is very narrow in the fore and aft direction but very broad in the across-track direction. The transmitted sound beams image, or "insonify," a narrow slice of sea floor perpendicular to the ship's track. The acoustic energy returned from the sea floor (acoustic backscatter) is then sensed, or "heard," by the instrument and transmitted to a shipboard recorder which displays the data in two-dimensional form as an image. As the instrument is towed along, it insonifies adjacent slices of the sea floor and, in this way, builds an acoustic image. The level of acoustic backscatter is a function of, among other things, the sea-floor topography, roughness, and composition. Using computer-processing techniques, it is now possible to produce true-scale mosaics of large areas of sea floor similar in appearance to aerial photographs.

Conventional sidescan-sonar systems can insonify swaths of sea floor typically less than 1 km wide. These high-resolution systems are inappropriate for use as a reconnaissance tool because of the huge amount of time required to survey large areas. Two other systems, SeaMARC I (Sea Floor Mapping And Remote Characterization) and GLORIA (Geological LOng-Range Inclined Asdic) are able to obtain much broader swaths of the sea floor (up to 5 km for SeaMARC and 45 km for GLORIA). The GLORIA system is towed only 50 m below the sea surface at relatively fast speeds (8 knots) and can cover an area of as much as 27,700 km^2 per day. Thus, GLORIA can obtain true large-scale reconnaissance views of the sea floor which are needed in assessing the extent and style of mass movement processes in the deep-water sections of the EEZ. Many examples of the use of both seismic-reflection and sidescan-sonar techniques in the investigation of submarine slope instability are presented in the sections of this report.

HISTORY

Our awareness of the importance and extent of submarine slope failure and mass movement has evolved over a long period. It has become apparent to marine geologists and ocean engineers that the largest landslides on Earth have occurred on the bottom of the sea. As recently as the 1960's, however, submarine landslides were thought to be rare in ocean basins except on steep slopes, on the soft sediment of river deltas, or in seismically active areas. The first evidence for large-scale submarine slope failure and mass movement came primarily from studies of anomalous sea-floor topography (for example, see Shepard, 1955; Moore, 1961; Inderbitzen, 1965) or displaced material (for example, see Heezen and Ewing, 1952, 1955) and from studies of ancient strata from many locations (Jones, 1937; Beets, 1946; Kuenen, 1949; van Straaten, 1949). Development of seismic-reflection, sidescan-sonar, and sea-floor sampling techniques since the 1950's coupled with increasing interest in the continental margins as a resource area has resulted in the accumulation of a large amount of data pertinent to the evaluation of submarine slope failure and mass movement. Large volumes of sediment and rock are involved in these slope failures, and they occur on seismically active and inactive submarine slopes of varied steepness (Dill, 1964; Heezen and Drake, 1964; Menard, 1964; Kelling and Stanley, 1970; Moore and others, 1970; Lewis, 1971; Booth and Dunlap, 1977; Carlson and Molnia, 1977; Embley and Jacobi, 1977; Hampton and Bouma, 1977; Embley, 1980; Booth and others, 1984).

CAUSES OF SLOPE FAILURE AND ANALYSIS TECHNIQUES

As recently as the mid 1970's, the few actual observations of submarine slope failure in the literature were mainly qualitative, because slope stability analysis requires knowledge of the slope topographic profile, the shape of the major slip surface (failure plane), and a variety of engineering properties of the material that composes the sea floor—information that is seldom available in the case of submarine landslides. Early investigators recognized from qualitative studies that several mechanisms can induce or "trigger" submarine slope failure. Heezen (1956) suggested that "over-steepening" of the slope due to rapid deposition of sediment was responsible for submarine slope failure near the mouth of the Magdalena River in Colombia. Dill (1964) suggested that the generation of gas associated with the decomposition of organic matter could lead to slope failure. Submarine slope failure triggered by earthquakes first was recognized from qualitative observations (for example, see Gutenberg, 1939; Heezen and Ewing, 1952, 1955; Ryan and Heezen, 1965). Henkel (1970) showed that large storm waves could load the sea floor and cause slope

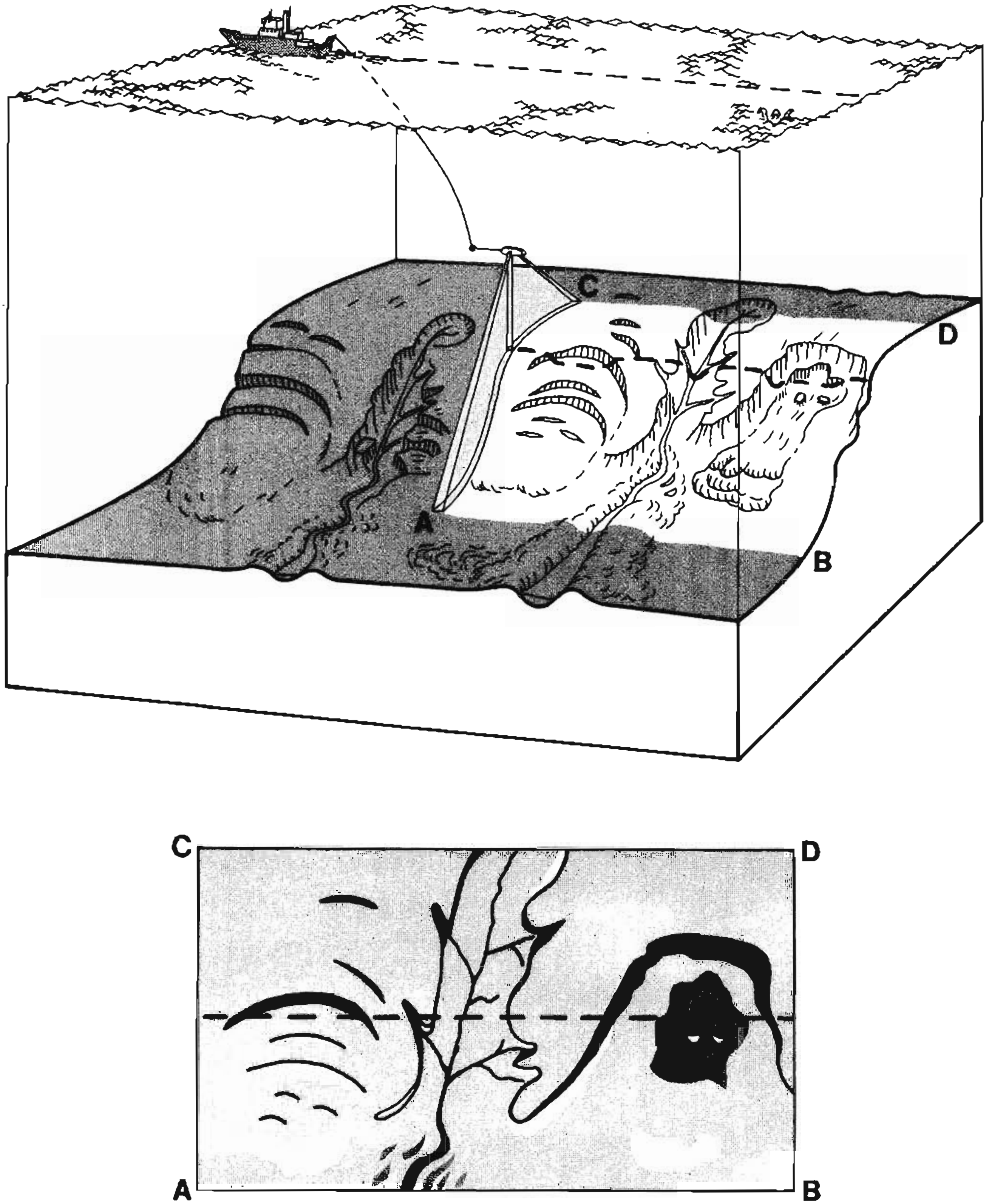

Figure 2. Principles of sidescan-sonar for mapping sea-floor landslide deposits. The top panel illustrates how sidescan sonar operates, and the bottom panel sketches the resulting sidescan-sonar image.

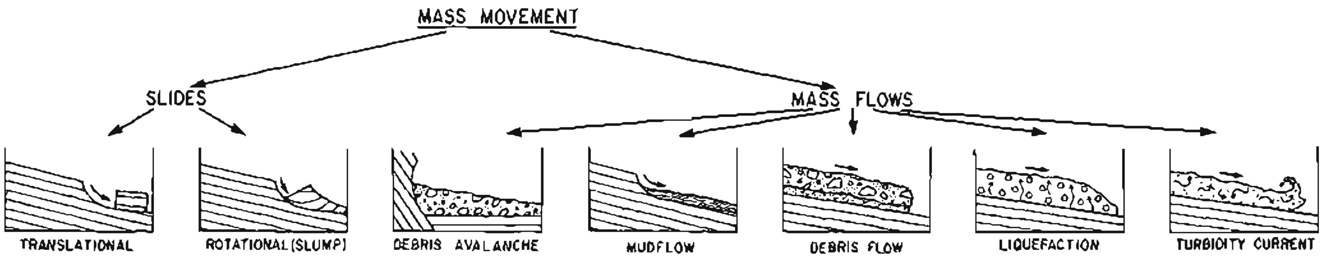

Figure 3. General landslide classification (modified from Varnes, 1958).

failure. This failure mechanism was observed clearly during the passage of Hurricane Camille (Bea and others, 1983).

Quantitative analysis of the mechanics of submarine slope failure requires making assessments of the strength of sediment and the magnitude of stresses acting downslope. An important quality of sediment is that the strength decreases as the excess pore-water pressure (that is, in excess of hydrostatic pressure) increases. Excess pore-water pressures commonly exist in deposits that are accumulating rapidly and can be evaluated using simplified sedimentation models (Terzaghi, 1956; Gibson, 1958). Excess pore-water pressures can also be caused by bubble-phase gas and by repeated loading from earthquakes or storm waves.

Downslope stresses result from gravity, earthquakes, or storm waves. Because, typically, so little is known about the properties and geometry of submarine slopes, investigators use analysis methods that are very simplified. The slopes are generally approximated by a tilted plane that extends for considerable distance in all directions (the so-called "infinite slope"). Earthquakes are represented by a constant lateral force that is proportional to the peak horizontal acceleration resulting from the shaking (Morgenstern, 1967). Wave-induced stresses are related to wave height, water depth, and other factors (Henkel, 1970; Seed and Rahman, 1978).

By evaluating the magnitude of downslope stresses relative to sediment strength, investigators have shown how steep slopes, high deposition rates, earthquakes, and storm waves can lead to slope failure (for example, Morgenstern, 1967; Almagor and Wiseman, 1977; Hampton and others, 1978; Schwab and Lee, 1983; Booth and others, 1984; and Lee and Edwards, 1986).

More sophisticated models for evaluating sediment deformation on slopes have also been developed. The quantitative accuracy of any deformation prediction is, however, limited by the quality of the sediment or rock stress-strain relation. Choosing the appropriate stress-strain relation for a seabed material poses major problems. For example, a significant change in the stress-strain behavior occurs when a sediment is subjected to repeated loading (Esrig, Ladd, and Bea, 1975; Anderson, 1976; Wright, 1976). Thus, the dominant influence of repeated loading (stresses associated with earthquakes or storm waves), along with other factors such as poor sample quality (Lee, 1985), severely limit the accuracy of these more advanced methods for offshore slope stability analysis, as a common practice. For site-specific studies, however, in which adequate resources are available for field borings, laboratory testing, and advanced computer modeling, accurate analyses of sea-floor deformations can be made using the finite-element approach (Arnold, 1973; Wright, 1976).

TERMS USED TO DESCRIBE SUBMARINE LANDSLIDES

Studies of sedimentary structures associated with slope failure and mass movement have shown that during and following slope failure, sediment can achieve a broad range of styles of mobility, from rigid block motion to turbulent flow. These styles of mobilization involving various types of material can be referred to by different terms. In this report, we follow the terminology recommended by Varnes (1958) with some modification (fig. 3).

Slope failure occurs when the downslope driving forces acting on the material composing the sea floor are greater than the forces acting to resist major deformations. Following slope failure, the failed mass moves downslope under the influence of gravity and possibly other forces. Thus, *mass movement* (also known as *mass wasting*) is defined as the movement of the failed material driven directly by gravity or other body forces, rather than tractive stresses associated with fluid motion. If the moving sediment takes a form that resembles a viscous fluid, the process is termed *mass flow*. Such a failure has considerable internal deformation with innumerable invisible or short-lived internal slip surfaces. *Slides* are translational or rotational movements of essentially rigid, internally undeformed masses along discrete slip planes. In the literature, all forms of mass movement are occasionally referred to as slumps. Correctly, *slumps* are a special kind of slide in which blocks of failed material rotate along curved slip surfaces (rotational movement). In each of these types, movement can be fast or slow. Extremely slow movement is called *creep*.

Submarine slides can become mass flows as the sliding mass progressively disintegrates and continued downslope movement occurs (Morgenstern, 1967; Hampton, 1972). End-member products of disintegrating slides have been given special names. *Debris flows* are flows of sediment in which the sediment is heterogeneous and may include larger clasts supported by a matrix of fine sediment. *Mudflows* involve predominantly muddy sediment. *Turbidity currents* involve the downslope transport of a relatively dilute suspension of sediment grains that are supported by an upward component of fluid turbulence. Turbidity currents are often generated by the disintegration and dilution of slides or debris flows, although they may be generated independently from other mass-movement events. *Liquefaction* occurs when a loosely packed sediment collapses under an environmental load. Here, the grains temporarily lose most contact with one another, and the particle weight is temporarily transferred to the pore fluid. Excess pore-water pressures are induced by this behavior. They eventually dissipate, although sometimes very slowly, and the material may flow downslope under the influence of gravity or spread laterally under the influence of stresses induced by earthquakes or perhaps storm waves.

Recent surveys using modern remote-sensing equipment have revealed truly giant submarine landslides that involve the failure of thousands of cubic-kilometers of rock and sediment (for example, see Moore and others, 1990). When these landslides have disintegrated into relatively smaller pieces (which may still be quite large) and have clearly moved rapidly, they are referred to as *debris avalanches*. Debris avalanches may occur on a smaller scale, too. In particular, when some sandy sedimentary deposits fail, most commonly as a result of slope oversteepening, they avalanche downslope as a thin *grain flow*.

ENVIRONMENTS

Submarine landslides are not distributed uniformly over the U.S. EEZ. Instead, they tend to occur most often where there are thick bodies of soft sediment, where the slopes are steep, and where the environmental loads are high. These conditions are met in fjords, deltas, and submarine canyons and on the continental slope.

Fjords

Of all the environments of U.S. EEZ, Alaskan fjords (and areas within Puget Sound, Washington) with high sedimentation rates are likely the most susceptible to slope failure, both in terms of the proportional areal extent of deposits that can become involved in mass movement and also in terms of the recurrence interval of slope failures at the same location. Fjords are glacially eroded valleys that have been inundated by the sea. They are typically fed by sediment-laden rivers and streams that drain glaciers. These factors lead to environmental conditions that are highly conducive to slope failure. The submerged sides of glacial valleys are commonly very steep and may extend to great depths, perhaps 1,000 m or more. There is typically a delta at the head of the fjord formed by streams that may drain the glacier that initially eroded the valley. These fjord-head deltas have beds that dip at 5° to 30° between 10 and 50 m water depth. In greater water depths, the submerged part of the delta dips at angles of 0.1° to 5° to the flat basin floor, typically at depths between 100 and 1,000 m (Syvitski and others, 1987). The glacial streams feeding these deltas carry rock flour and coarse sediment whose deposits easily lose strength when shaken by an earthquake. The sediment may also be deposited so rapidly that pore-water pressures cannot dissipate completely as sedimentation continues. The resulting underconsolidated state of the sediment causes abnormally low strength. Organic matter brought down with the glacial debris may be abundant, and it can decay and produce bubble-phase gas that also may lead to elevated pore-water pressure and low strength. Some fjord-delta deposits are so near instability that they fail during particularly low tides, during which the supporting forces of seawater are temporarily removed from the sediment. Some of these steep slopes with weak sediment fail seasonally or semicontinuously yielding numerous small-scale slope failures. Slope failures of a seasonal or semicontinuous nature have been reported in many fjords in Canada; for example, Bute Inlet (Syvitski and Farrow, 1983; Prior and others, 1986), Knight Inlet (Syvitski and Farrow, 1983), and North Bentinck Arm (Kostaschuk and McCann, 1983) and are probably common in Alaska, as well. Fjord-head delta slope failures occasionally produce catastrophic effects, such as occurred in Valdez (Coulter and Migliaccio, 1966; Hampton and others, this report), Seward (Lemke, 1967; Hampton and others, this report) and Whittier (Kachadoorian, 1965) in the 1964 Alaska earthquake or in Kitimat Arm, Canada (Prior, Bornhold, and others, 1982; Prior, Coleman, and others, 1982).

The sidewalls of fjords can also be unstable. Deposition of suspended sediment on the steep (10° to greater than 90° overhangs) submerged valley sides can frequently lead to small slope failures (Farrow, Syvitski, and Tunnicliff, 1983). Even more important are slope failures on side-entry deltas that build out rapidly onto the sidewall slopes. Failures in Howe Sound, Canada (Terzaghi, 1956; Prior and others, 1981), are of this type.

Finally, the deep fjord basins, which tend to have slopes of less than 0.1, commonly receive failed sediment masses and flows from the side walls and fjord-head deltas. If these landslides incorporate enough water during their movement, they can progress through mass flows into turbidity currents. These mass flows and turbidity currents can be fed into and across the basins by channels (Syvitski and others, 1987).

Slope failures in fjords can generate large waves which cause major damage to coastal communities. During the 1964 Alaska earthquake, much of the damage and many of the fatalities in Valdez (Coulter and Migliaccio, 1966; Hampton and others, this report), Seward (Lemke, 1967; Hampton and others, this report) and perhaps Chenaga (Plafker and others, 1969) resulted from giant waves generated by submarine landslides in fjords. Waves 8.2 m in height were generated during the major sea-floor failure at Kitimat Arm, Canada, in 1975 (Murty, 1979).

Because fjords are found in rugged, mountainous terrain, the fjord-head deltas and side-entry deltas are frequently the only flat land available for development. These marginally stable to unstable locations become the sites of coastal developments. Not only do these developments become vulnerable to natural slope failure, which might occur frequently even if the developments were not present, but man's activities can also lead to additional slope failures. For example, a river channel stabilization program at Howe Sound (Terzaghi, 1956) caused rapid delta growth to be localized and probably contributed ultimately to slope failure.

Active River Deltas on the Continental Shelf

Active river deltas are the next most likely sites for slope instability in the U.S. EEZ. Many rivers contribute large quantities of sediment to relatively localized areas on the continental margins. Depending on a variety of environmental factors, including wave and current activity and the configuration of the continental shelf and coastline, thick deltaic deposits can accumulate fairly rapidly. These sediment wedges and blankets can become the locations of sediment instability and landsliding partly because of their thickness. To create large, deep-seated landslides, a thick deposit of comparatively low-strength sediment is needed. Because most of the continental shelves were subaerially exposed during the last glacial cycle, most sediment on the shelves from that time or before has been eroded or desiccated. Accordingly, the strengths of these older deposits are commonly high enough to resist downslope gravitational stresses on the gentle shelves, and all but the very greatest storm- and earthquake-induced stresses as well. Only in areas of continued modern deposition where significant thicknesses (tens of meters) of younger sediment have been deposited above the strong, old sediment do landslides occur. These younger deposits tend to have relatively low strength because of rapid deposition rates. In addition, decaying organic matter can produce bubble-phase gas that can further reduce strength. These locations may fail under gravitational loading (due to the slope steepness alone) or during storms or earthquakes.

The locations of the major sedimentary depocenters provide some information on where undersea landslides might be expected on the continental shelf. For example, the Yukon River is a major source of sediment (60 million tons per year) in the northern Bering Sea. Although the prodelta is nearly flat, there are indications of localized liquefaction of sandy silt (Clukey and others, 1980) during large storms.

The occurrence of slope failures on active deltas of the continental shelf off southern Alaska is not surprising. Glacially fed rivers debouching into the Gulf of Alaska or adjacent sounds and inlets contribute 450 million tons of sediment per year (Milliman and Meade, 1983). Slope failures have been identified in modern sediment all along the Alaskan margin (Reimnitz, 1972; Molnia and others, 1977; Carlson, 1978; Schwab and Lee, 1983, 1988, this report; Schwab and others, 1987; Lee and Edwards, 1986). The landslides are likely induced by either storm waves or earthquakes (Schwab and Lee, 1988, this report); the high incidence of landsliding arises because of the intensity of earthquake and storm-wave loading, the thickness of modern sediment, and the tendency of the glaciomarine sediment to lose strength when cyclically loaded.

The west coast of the United States has relatively short, steep rivers with small drainage basins. The sediment yield of these rivers can be high but is extremely variable (Milliman and Meade, 1983). There are indications of slope instability on some of the deltas, for example, that of the Klamath River (Field and others, 1982; Field, this report).

The Mississippi River contributes the most sediment to the EEZ of any single river within the United States (2 to 7×10^8 tons of sediment per year; Coleman and others, 1980; Milliman and Meade, 1983). Most of this sediment is deposited on the shelf fronting the modern bird-foot delta, a delta-lobe that has been in existence for only 600 to 800 years (Fisk and others, 1954). The distributary mouths of the modern delta build seaward at rates varying from 50 to 100 m per year. Offshore of these distributaries, the sediment accumulation rates are very high, ranging to 1 m per year or more (Coleman and others, 1980). The sediment deposited consists mostly of clay-sized particles and is also rich in organic matter that is rapidly degraded to gas (mainly methane and carbon dioxide). Rapid sedimentation and gas charging lead to high excess pore-water pressures and a state of extreme underconsolidation. Although the slopes of the delta-front are very low, less than 1.5°, evidence of slope failure is widespread. Submarine gullies were first described by Shepard (1955). More recently, sidescan sonar mosaics have shown extensive fields of sediment instabilities all along the delta-front (Coleman and others, 1980; Coleman and others, this report).

Elsewhere along the coast of the Gulf of Mexico and Atlantic Ocean, the rate of sediment accumulation on the continental shelf is low and landslide features are rare. Most modern deposition occurs in bays that formed when valleys were drowned by rising sea level after the most recent glacial cycle (ice age).

Deltas of some of the major rivers that deposit large amounts of sediment along the margins of the United States do not display geomorphic evidence of submarine landsliding. Such lack of evidence does not automatically preclude previous slope failure in these deposits because waves and burial can erase the failure effects. The lack of slope failure features likely indicates, however, that the occurrence of slope failure is less frequent or less extensive in these deltas. Most notable is the delta of the Columbia River (Nittrouer and Sternberg, 1981) whose surface appears to be devoid of failures. Evidently, a combination of circumstances is needed for slope failure to occur on the continental shelf. High sedimentation rates are needed so that a sufficiently thick bed of modern sediment can develop. Relatively low permeability (fine-grained sediment) can allow high pore-water pressures to be retained and produce low sediment strength. Also needed, however, are environmental factors, such as storms or earthquakes, that can generate stresses in the sediment that exceed the strength. These can be augmented by strength degradation during cyclic loading, such as seems to be particularly common with glacially derived sediment. Finally, the configuration of the continental shelf, including its slope, can influence the stability of deltas. The interaction of all of these factors, rather than one factor alone, ultimately determines whether mass movement will occur.

Submarine Canyon and Deep-Sea Fan Systems

Submarine canyon systems serve as conduits for passing large amounts of sediment from near the continental shelf to deep-sea fans. Although more active at times of lowered sea level during periods of extensive continental glaciation, the presence of extensive, thick sediment deposits on the continental rise surrounding the United States testifies to the importance of mass movement mechanisms associated with these systems. Mass movement mechanisms are capable of bringing sand-sized and even coarser particles to locations hundreds of kilometers from shore. Landsliding appears to be one system element that allows massive deposition on submarine fans to occur. According to one model (Hampton, 1972), sediment accumulations in canyon heads begin to move as coherent landslide blocks following some triggering event, such as an earthquake or storm. As the blocks move downslope, the resulting jostling and agitation cause disintegration and subsequent incorporation of ambient water. The debris flow produced displays increasingly fluidlike behavior. As the debris flow continues on its path, further dilution by surrounding water occurs, particularly as sediment is eroded from the front of the flow. Ultimately, a dilute turbulent cloud of sediment, or turbidity current, is created. The turbidity current has a density slightly greater than the surrounding seawater and can flow for long distances (up to hundreds of kilometers) at moderate to high velocities (1 to 8 m/s, Shepard, 1963; Bowen and others, 1984; Reynolds, 1987). When the turbidity current enters a depositional phase, it leaves deposits that have distinctive textural characteristics; namely sandy sediments at the base that decrease in grain size to silt and clays at the top of the deposit (Bouma, 1962; Middleton and Hampton, 1976).

Landsliding, particularly within submarine canyons (Carlson and others, this report), appears to be an important, if not essential, part of the process of building deep-sea fans, which are among the most extensive sedimentary features of the Earth's surface. The circumstances surrounding these slope failures and their subsequent conversion into turbidity currents are poorly understood, however. Storms cause sediment movement in canyons, perhaps by inducing slumping near the canyon heads (Marshall, 1978) or perhaps by introducing or resuspending enough sediment to form a density current directly (Shepard and Marshall, 1973; Reynolds, 1987). Earthquakes also cause landslides in canyon heads and subsequent turbidity current flow (Malouta and others, 1981; Adams, 1984), but details of this process are lacking. Major earthquakes and other shocks do not always cause canyon-head landslides (Shepard, 1951; Dill, 1969). Landsliding in canyon heads and turbidity current mobilization were likely more common during glacial cycles (Nelson, 1976; Barnard, 1978) because of lowered sea level, increased sediment supply and possibly, increased storm wave loading. Thus, many of these submarine canyons and deep-sea fans, unlike fjords and river deltas, are not active areas of slope failure today.

The Open Continental Slope

A final common environment for undersea landsliding in the U.S. EEZ is the area between submarine canyons on the continental slope. Landslides have been reported all around the U.S. margin along slopes removed from submarine canyon systems. Included are slopes off southern California (Buffington and Moore, 1962; Haner and Gorsline, 1978; Field and Clarke, 1979; Nardin, Edwards, and others, 1979; Nardin, Hein, and others, 1979; Ploessel and others, 1979; Field and Edwards, 1980; Field and Richmond, 1980; Hein and Gorsline, 1981; Thornton, 1986; Edwards, Lee, and Field, this report; Field and Edwards, this report), central and northern California (Field and others, 1980; Richmond and Burdick, 1981; Field and Barber, this report; Gutmacher and Normark, this report), off Alaska (Marlow and others, 1970; Hampton and Bouma, 1977; Carlson and others, 1980; Hampton, this report; Kayen and Lee, this report), in the Gulf of Mexico (Lehner, 1969; Woodbury, 1977; Booth, 1979; Booth and Garrison, 1978; Twichell, Valentine, and Parson, this report; and McGregor and others, this report), and along the Atlantic coast (Embley and Jacobi, 1977; McGregor, 1977; Knebel and Carson, 1979; McGregor and others, 1979; Malaboff and others, 1980; Booth and others, 1984, 1988; Cashman and Popenoe, 1985; O'Leary, 1986; Booth and others, this report; O'Leary, this report; Popenoe, Schmuck, and Dillon, this report; Dillon and others, this report). The failures are found near river mouths and far removed from them, as well as in both arid and humid

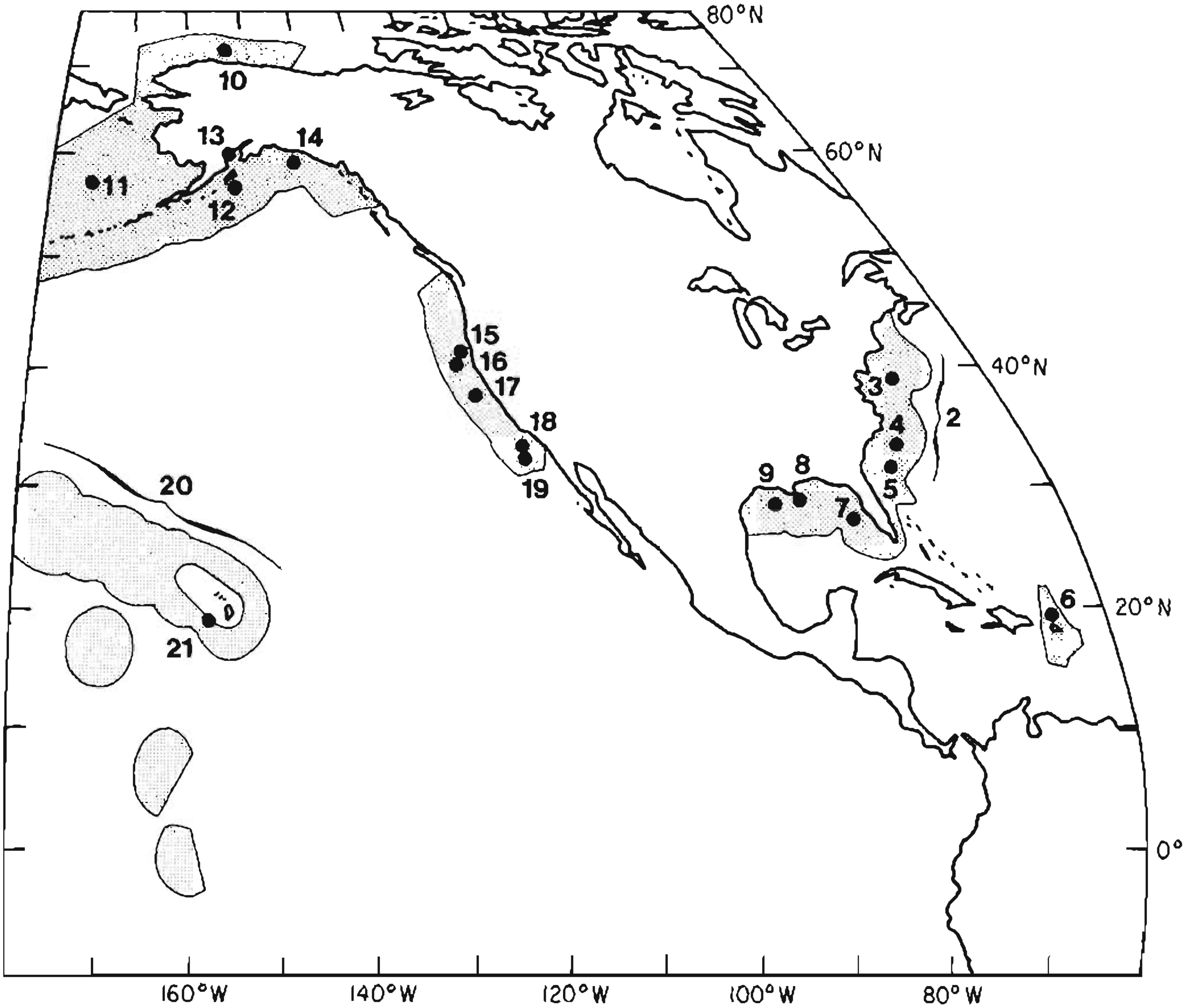

Figure 4. Locations of landslide studies described in this report. Numbers refer to sections listed in table 1. The shaded area is the U.S. EEZ.

climates. Ages of the slope failures are seldom known, so we cannot determine whether they occurred under glacial or interglacial conditions, nor can the recurrence interval be estimated with any degree of accuracy. Many were probably seismically induced because typical continental slopes have gradients of 5° or less which would be expected to be statically stable. Storm-wave loading is seldom a major factor much below the shelf break (Lee and Edwards, 1986). Occurrence of slope failures seems to correlate with sedimentation rate, slope declivity, seismicity, and presence of bubble-phase gas and gas hydrate, but the relationship is complex (Field, 1981).

SECTIONS IN THIS REPORT

This report contains 20 sections that discuss landslides within the U.S. EEZ (fig. 4, table 1). The locations of these landslides are distributed fairly uniformly around the United States. This compilation is not exhaustive, however. Rather, the goal is to show the variety of landslides that have been found, the associations of the landslides with other features, methods that have been used to understand the causes and mechanisms of the landslides, and the potential that some of these features have as a hazard to coastal and offshore development. One of the

Table 1. Sections included in this report

No.	Authors	Study area	Major theme of section
2	Booth and others	Continental slope, Atlantic margin	Variety and distribution of Landslides on Atlantic margin
3	O'Leary	Continental slope, southeast New England	Large-scale landsliding is a principal formative process of the New England continental margin
4	Popenoe, Schmuck, and Dillon	Cape Fear landslide, continental slope, North Carolina	Slope failure induced by salt tectonics and decomposition of gas hydrate during low sea level
5	Dillon and others	Lion's Paw landslide, Blake Escarpment, Florida	Erosional processes cause carbonate escarpments to fail
6	Schwab, Danforth, and Scanlon	Insular slope, Puerto Rico	Debris avalanche controlled by structure and regional tectonic setting
7	Twichell, Valentine, and Parson	Continental shelf and and West Florida Escarpment, Florida	Erosional processes cause carbonate slopes to fail
8	Coleman and others	Mississippi River Delta, continental shelf, Louisiana	Slope failures are common on active deltas and can affect offshore development
9	McGregor and others	Continental slope, western Gulf of Mexico, Texas	Salt diapirism related to the style of slope failures
10	Kayen and Lee	Continental slope, Beaufort Sea, Alaska	Slope failures induced by decomposition of gas hydrate during low sea level
11	Carlson and others	Beringian margin, submarine canyon Bering Sea Alaska	Mass movement is the principal process dominating the submarine canyon environment
12	Hampton	Kodiak continental margin, Alaska	Slope failures controlled by structure and regional tectonic setting
13	Hampton, Lemke, and Coulter	Fjords, Valdez and Seward, Alaska	Effects of earthquake-induced slope failures on communities along fjords
14	Schwab and Lee	Continental shelf, Gulf of Alaska	Different slope failure morphologies result from different failure mechanisms
15	Field	Continental shelf, northern California	Earthquake-induced liquefaction
16	Field and Barber	Continental slope, northern California	Slope failures associated with gas-charged sediment
17	Gutmacher and Normark	Sur submarine landslide, continental slope, central California	Large slope failure on a margin with a low sedimentation rate
18	Edwards, Lee, and Field	Gaviota mud flow, basin slope, southern California	Processes controlling earthquake-induced mud flow
19	Field and Edwards	Southern California borderland	Regional distribution of landslides in a well-studied area
20	Normark, Moore, and Torresan	Hawaiian Ridge, Hawaii	Giant slope failures are a primary process in the formation of many volcanic islands
21	Normark and others	Insular slope, Oahu, Hawaii	Storm-wave-induced slope failure of carbonate sediment

sections in this report (Hampton, Lemke, and Coulter) considers sediment failures in fjords, three sections consider failures in rapidly deposited sediment on the continental shelf (Coleman and others; Field; Schwab and Lee), one section (Carlson and others) considers mass instability in submarine canyons, and the remaining fifteen sections are concerned with failures of varying size on continental, insular, or basin slopes. The slope environment is emphasized because the largest and most spectacular landslides occur here and more nearly complete coverage with sidescan-sonar mapping has been obtained during the U.S. Geological Survey EEZ-SCAN Project (Gardner, 1986). More landslide deposits are preserved in the slope environment through glacial stages when the continental shelves are subaerially exposed and coastal fjords are filled with glaciers. That is, more landslide deposits are preserved on continental and insular slopes even though the frequency of landslide recurrence may be much greater in deltaic deposits on the continental shelf and in fjords.

The sections in this report provide insights into the presence of mass movement features and processes that control submarine mass movement. Although marine researchers and engineers are attempting to resolve the problems associated with slope instability through the use of more quantitative methodologies, the results of recent sea-floor mapping surveys provide a constant reminder that there is much yet to be known. The marine researcher will be tasked with the responsibility to solve these unknowns in the 1990's and beyond as the economic significance of the sea floor increases. As many of the sections in this report suggest, however, marine research will not only provide solutions but will continue to reveal new mysteries.

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Slope Stability in Regions of Sea-Floor Gas Hydrate: Beaufort Sea Continental Slope

By R.E. Kayen *and* H.J. Lee

INTRODUCTION

The continental slope of the Alaskan Beaufort Sea is disrupted by an apparently continuous area of submarine landslides (Grantz and Dinter, 1980). The zone of slope instability extends from water depths of 200 to 400 meters (m) at the shoreward edge, to depths in excess of 2,000 m (fig. 1). The thickness of displaced sediment typically is 100 to 400 m. The area of slope failure largely lies in a region underlain by marine gas hydrate (sediment bonded by a solid mixture of water and methane). Gas hydrate has been interpreted from seismic-reflection profiles (Grantz and Greenberg, 1981; Grantz and others, 1981), and extends down from near the sea floor to at least 700 m beneath the sea floor between the 400- and 2,800-m water depths (fig. 2). For many of the landslides, the basal shear surface corresponds approximately with the base of the gas hydrate. For example, the mass-movement deposit in figure 3, profiled in the eastern Alaskan Beaufort Sea, is bounded upslope by a scarp and appears to share its basal shear surface with a strong gas-hydrate reflector. Many of the

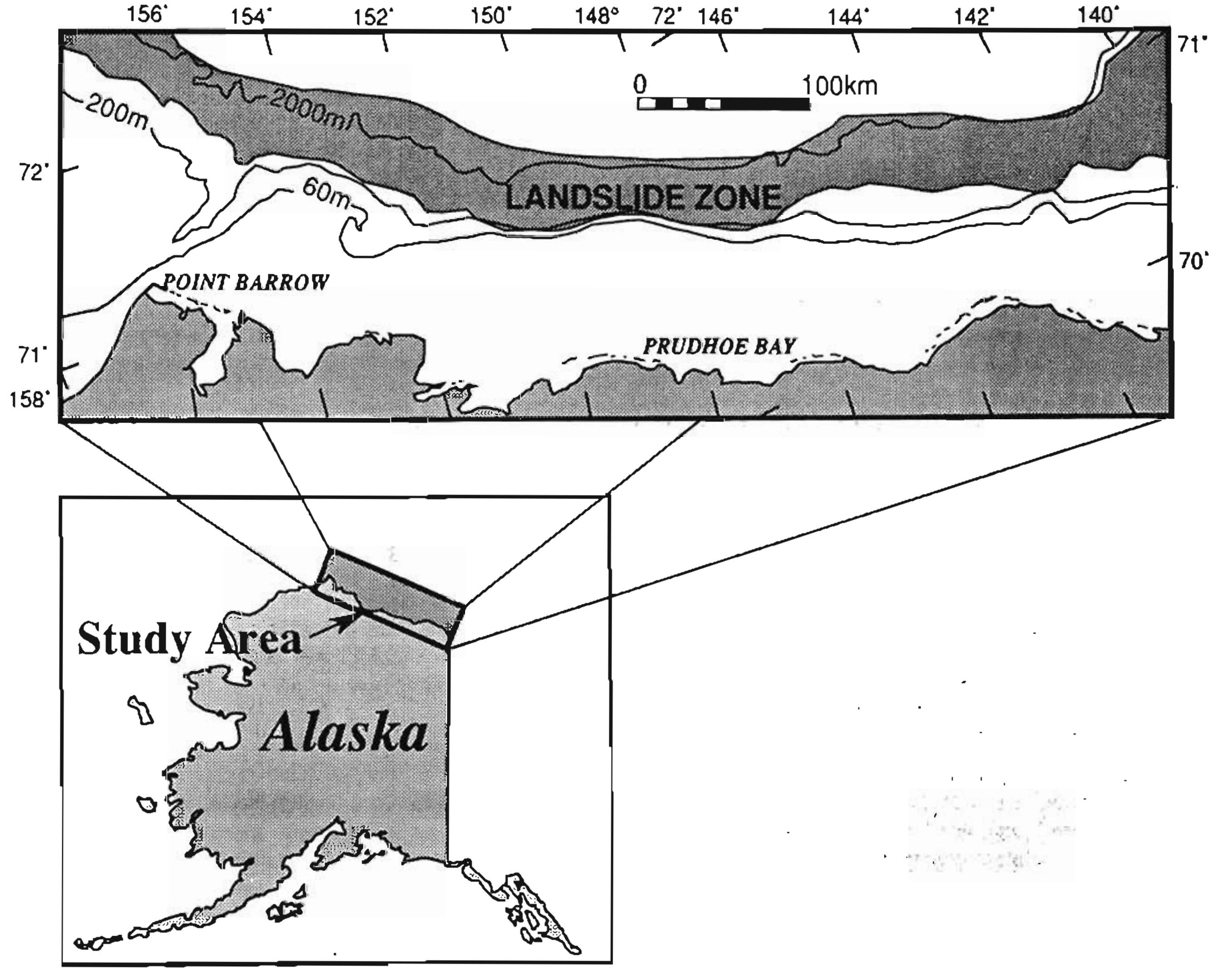

Figure 1. The zone of large landslides on the continental margin of the Beaufort Sea, interpreted from seismic-reflection profiles (from Grantz and Dinter, 1980).

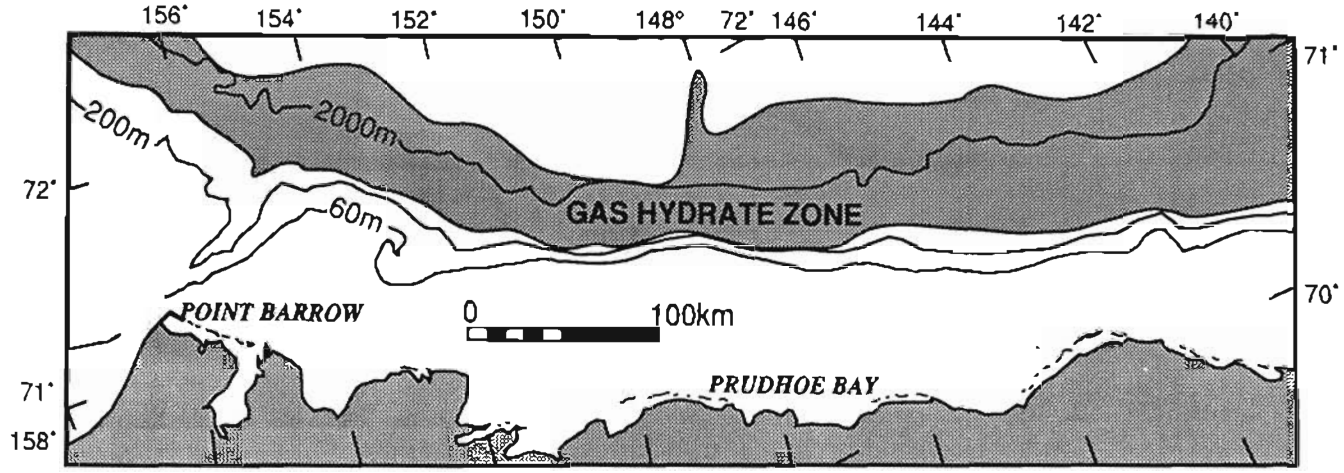

Figure 2. The region of sea-floor gas hydrate on the continental slope extends from approximately 400 meters water depth near shore to over 2,000 meters. The presence of gas hydrate within sea-floor sediment is interpreted from a strong bottom-simulating reflector in seismic-reflection profiles (from Grantz and Dinter, 1980).

observed slope failures probably were triggered during Pleistocene sea-level fall in response to partial disassociation of the gas hydrate.

THE FORMATION OF GAS HYDRATE

Solid-solution gas hydrate forms beneath the sea floor from natural gas and water within a stability field determined by temperature and pressure (fig. 4). Formation can occur in both polar and nonpolar regions due to the normally cold bottom-water conditions (typically 2° to 5°C) found on continental slopes throughout the world. Hence, most slopes found in water deeper than 400 to 500 m have a sea-floor pressure and temperature environment within the stable phase for gas-hydrate formation. To produce a marine gas hydrate, water molecules bond to form a cubic lattice structure within which individual gas molecules, primarily methane, are caged (Kvenvolden and McMenamin, 1980). The resulting structure densely packs natural gas molecules in a configuration that is far tighter than would exist in free gas at the same pressure and temperature. For example, at standard temperature and pressure approximately 170 volumetric units of free methane gas are compressed and stored in a single volumetric unit of gas hydrate.

Conditions conducive to formation of gas hydrate lie below a phase boundary (left side of fig. 4). The actual depth and temperature conditions at a representative location on the Beaufort Sea slope (profile *A–A'*) intersect the phase boundary diagram and show where gas hydrate is stable beneath the sea floor. Because temperature increases relatively rapidly with depth in the sediment due to the geothermal gradient, gas hydrate can form only within the upper section beneath the sea floor. The depth beneath the sea floor at which the pressure and temperature conditions reach the phase boundary determines the location of the base of the gas-hydrate zone, which can be seen readily in seismic-reflection profiles (fig. 3, also see Carpenter, 1981). The reflection from the base of the gas hydrate approximately mimics the bottom bathymetry and commonly is termed a *bottom-simulating reflector* or BSR (fig. 3).

EFFECT OF SEA-LEVEL FLUCTUATIONS ON GAS HYDRATE

During ice ages, a worldwide drop in sea level that occurs in response to the formation of massive glaciers on land causes reduced total pressure within sediment beneath the sea floor. Normally, pore pressures within the hydrate-free sediment dissipate very rapidly during a drop in sea level to maintain essentially hydrostatic conditions (fig. 5*A*). In regions of gas hydrate, however, the lowering of sea level reduces pressure along the gas-hydrate base, initiating disassociation (melting) and consequently releasing large volumes of bubble-phase gas into the sediment directly beneath the hydrate base. The dense concentration and large quantity of natural gas within a gas hydrate is such that disassociation of only a small part of the base elevates pressures back into equilibrium phase-boundary conditions. That is, as sea level drops, gas-augmented pore-water pressures near the phase boundary remain almost as great as they were when sea level was high.

Therefore, as sea level falls, the gas-hydrate base tends to disassociate and retreat upward in the sediment column to a new position of lower equilibrium pressure and temperature (Field and Kvenvolden, 1985; 1986), but elevated pore-water pressures at the base, trapped beneath

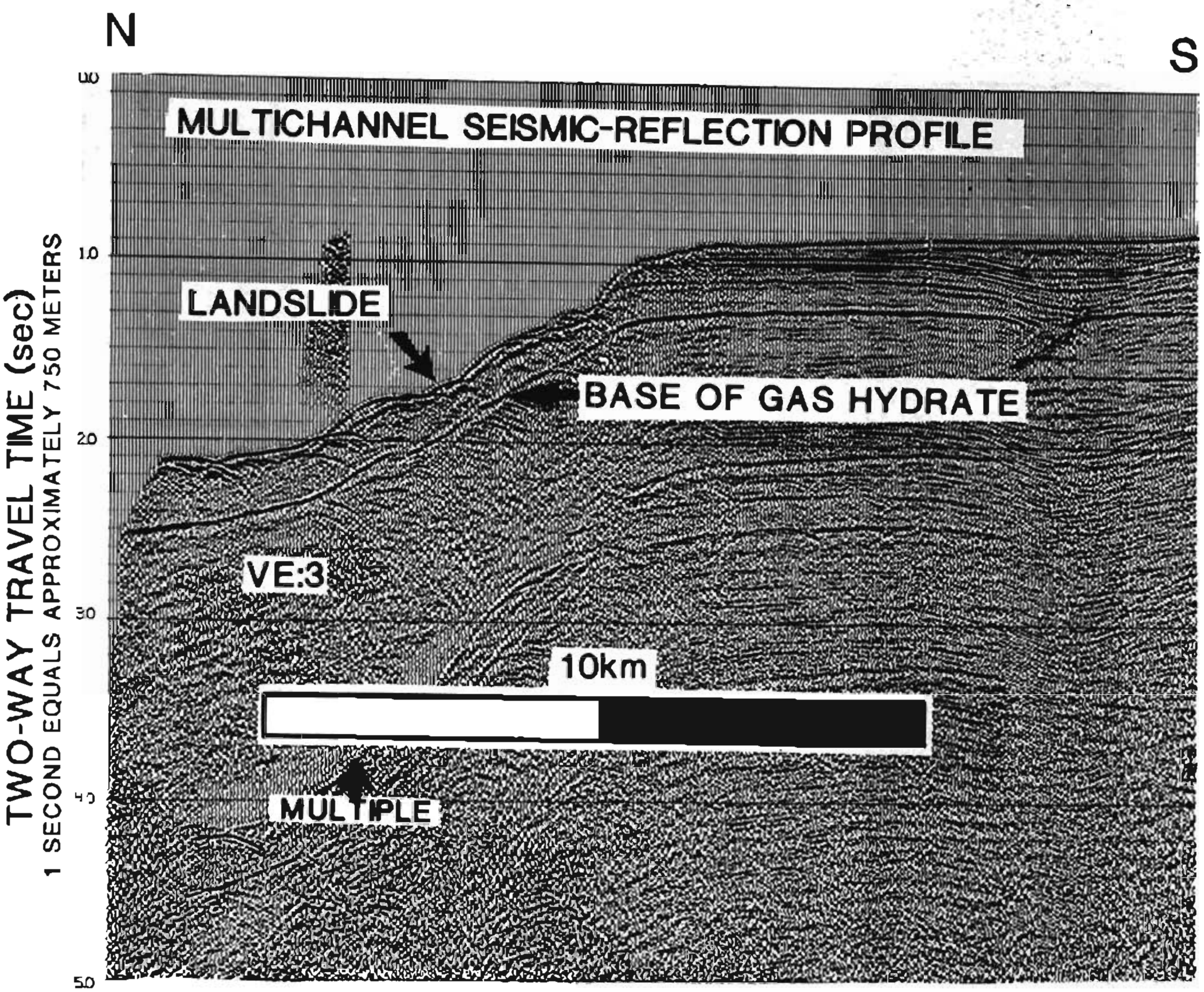

Figure 3. The landslide revealed on this seismic-reflection profile in the eastern Beaufort Sea displays a 100-meter-high headwall scarp at the shelf-edge break that defines the landward edge of the slope failure (data from Grantz and Greenberg, 1981). Seaward, the hummocky irregular topography, characteristic of the disrupted surface of a mass-movement deposit, extends to a water depth of at least 1,900 meters. The basal shear surface of the slide appears to closely follow the strong bottom-simulating reflector that marks the base of the gas hydrate. Note: Vertical exaggeration (VE) is 3 times.

the largely impermeable gas hydrate above, oppose further disassociation. As a result, excess pressures preserve a base that is deeper in the sediment column than the base would be if these pressures did not exist.

A predicted pore-water pressure profile through the sediment column at the completion of a large drop in sea level is presented in figure *5B*. Because of disassociation, hydrodynamic pressures are elevated at the gas-hydrate base with respect to the hydrostatic profile developed in unhydrated sediment as seen in figure *5A*. These excess pressures compromise the stability of the slope by reducing the normal stress between individual sediment grains and, consequently, the sediment's resistance (sediment strength) to shear stresses that tend to move the sediment downslope. Because the shear stresses acting downslope in this region are primarily caused by gravity acting on a sea-floor slope, the excess pore-water pressure may cause the slope under static conditions to become unstable (Kayen, 1988).

SLOPE STABILITY DURING GAS-HYDRATE DISASSOCIATION

Evaluating the effect of gas-hydrate disassociation on slope stability requires quantifying the process of excess

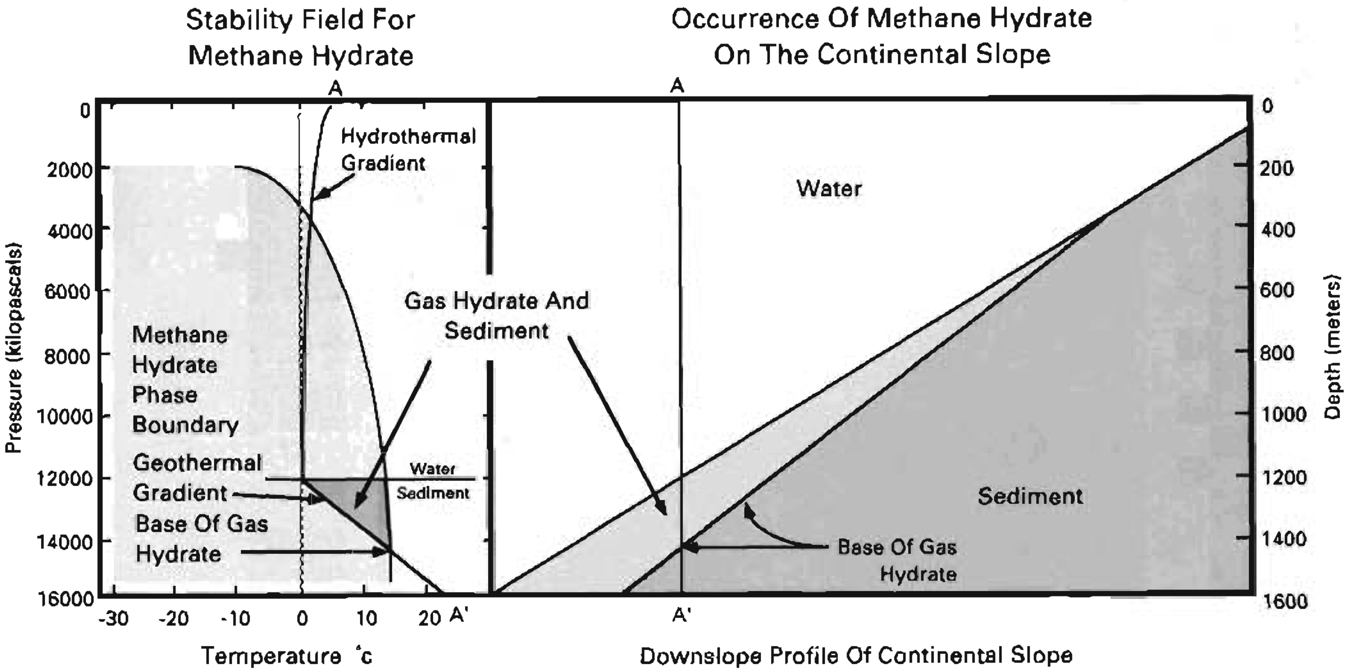

Figure 4. On the left is a gas hydrate pressure-temperature stability field (phase diagram) and on the right a schematic diagram of the continental slope. Formation of gas hydrate can occur only below the methane-hydrate phase boundary. Pressure (in kilopascals) and temperature (in degrees Celsius) conditions of profile A–A′ through the sediment slope on the right are drawn on the phase diagram. Environmental conditions are well within the stability field at the sea-floor surface but intersect the phase boundary with burial, which determines the thickness of the hydrate zone in sediment.

pore-pressure generation and dissipation. Numerical solutions were obtained by Kayen (1988) for the temporal and spatial distribution of excess pore-water pressures developed during gas-hydrate disassociation for a variety of possible sediment types, from coarse sand to fine-grained deposits of clay, beneath a gas-hydrate base. The numerical approach, a finite-difference model, was formulated around an averaged and linearized sea-level fall of 100 m, occurring over 9,100 years ago, representative of the last Pleistocene lowering of sea level as based on the work of Matthews (1973), Steinem and others (1973), Bloom and others (1974), and Chappell (1974). The model for predicting excess pore-water pressures beneath the base of the hydrate was constrained by previously determined gas hydrate thermodynamic data (Kvenvolden and others, 1984; Makogon, 1981; and Sloan, 1990).

The model shows that a continuous fall of sea level causes a slow degradation of the gas-hydrate base. The level of excess pore-water pressure that develops near the base depends largely on the ability of the sub-base sediment to dissipate excess pressures. For example, clayey deposits have a low pore-fluid dissipation capability: excess pressures do not dissipate and, therefore, suppress further disassociation (fig. 6). Accordingly, at the completion of a sea-level fall, a large excess pressure is present. If little disassociation occurs, the excess pressure at the end of the drop in sea level will be nearly equal to the total hydrostatic pressure change, from high stand of sea level to low stand. In contrast, sand deposits can readily dissipate excess pressures, so the rate of disassociation along the base is relatively high and largely controlled by heat flow into the base. At the end of a large drop in sea level, a marked upward retreat of the base and relatively low excess pressure are expected. The numerical solution for a sandy sediment supports this prediction (fig. 7). The two sediment types, sandy and clayey, can be considered end-members with regard to pore-fluid dissipation. Thus, a sub-base clayey sediment will tend to develop greater excess pore-water pressure than sandy sediment during gas-hydrate disassociation and would be more prone to slope failure.

SLOPE STABILITY OF THE BEAUFORT SLOPE: CONCLUSIONS

Studies of sediment sampled from the zone of slope failure of the Beaufort Sea continental slope indicate that the diffusion and physical properties are more like clayey sediment than sandy sediment (Kayen, 1988). Given the geometry and sediment physical properties of the contine-

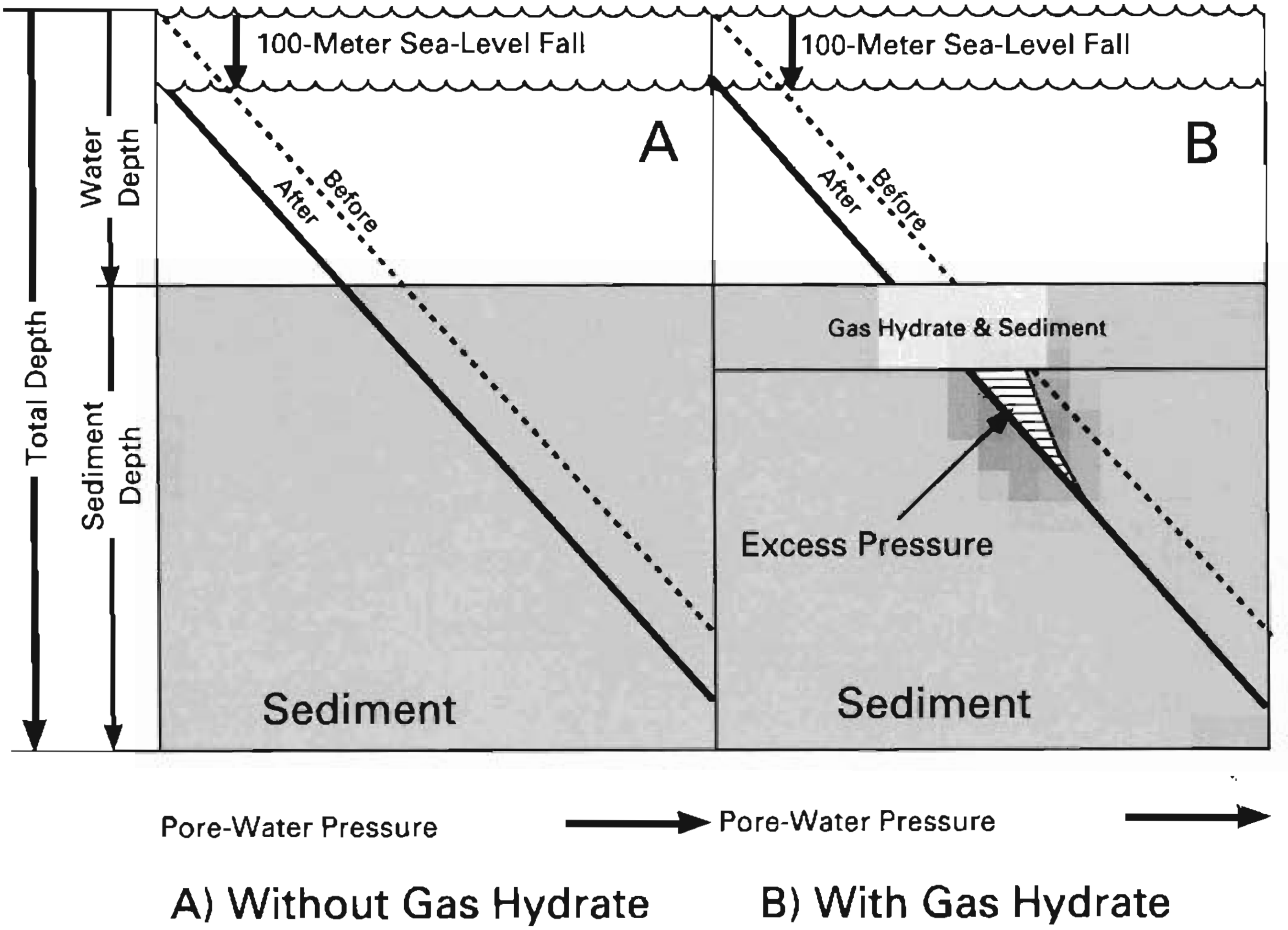

Figure 5. Pressure profiles before and after a 100- meter sea-level fall caused by a glaciation cycle are displayed for (*A*) normal sea floor without gas hydrate and (*B*) sea floor bearing gas hydrate. The dashed line marks the hydrostatic pressure profile before sea-level fall. The solid line marks the pressure profile at the completion of the sea-level fall. Without gas hydrate, sea-floor sediment can fully adjust to the lowering of sea level, and a hydrostatic pore-water pressure profile can be maintained. The presence of gas hydrate prevents the normal reduction of pressures with sea-level fall because the lowering of pressure initiates disassociation along the base. This excess pressure reduces slope stability if the sea floor is inclined.

tal slope, the amount of excess pore-water pressure that will cause slope failure can be calculated using basic soil mechanics principles (Kayen, 1988). The range of pressures likely to trigger slope failure of a 100- and 200-m-thick sedimentary deposit on a 5° sea-floor slope, typical of the Beaufort margin, is presented in figure 8. Superimposed on this critical range of pore pressures are the pressures likely to be generated at a gas-hydrate base during disassociation induced by sea-level fall.

For finer-grained sediment, dissipation of excess pressures through sediment pore space is probably sufficiently slow to cause slope failure on the continental slope of the Beaufort Sea. Stability is maintained only if excess pressures are rapidly vented away from a disassociating base, perhaps along fault planes. Stability may also be maintained if gas hydrate forms in the pore space of strongly cemented sediment or in the joint space (cracks) of a consolidated rock mass. In both cases, the materials have high inherent cohesive strengths. For many of the unlithified sedimentary deposits on the Beaufort Slope, however, excess pressure generation at the base of the gas-hydrate zone during Pleistocene drops in sea level was sufficient to initiate sea-floor landsliding (Kayen, 1988). This mechanism for triggering mass movement probably also applies to the numerous other regions of the world where gas hydrate is present within sea-floor sediment (Lee, Schwab, and Booth, this report; Popenoe, Schmuck, and Dillon, this report). Accordingly, times of lowered sea level during ice ages also are times of widespread instability of continental slopes that are underlain by gas hydrates.

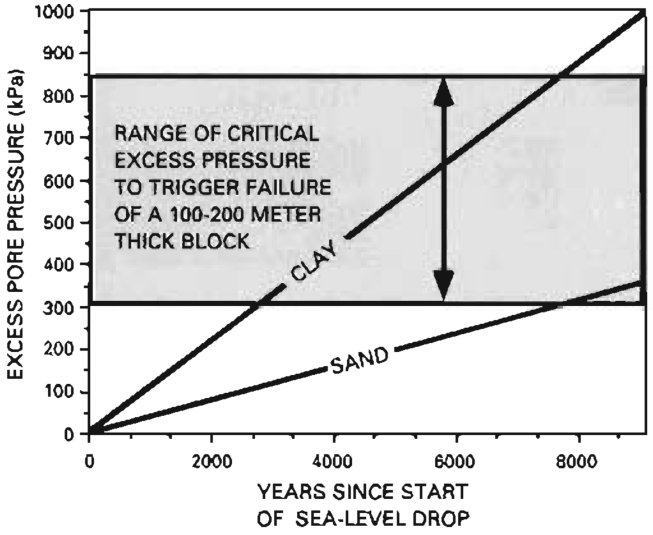

Figure 8. Predicted excess pore-water pressures (in kilopascals) generated at the base of a disassociating gas hydrate (diagonal lines) during a 100-meter sea-level fall are presented against, and exceed, a range of likely excess pressures required to initiate slope failure (horizontal lines). For sandy to clayey sediment, slope failure is predicted unless excess pore pressures can be rapidly vented to the sea floor through alternate conduits.

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Mass Movement Related to Large Submarine Canyons Along the Beringian Margin, Alaska

By P.R. Carlson, H.A. Karl, B.D. Edwards, J.V. Gardner, *and* R. Hall

INTRODUCTION

The 1,400-kilometer-long Beringian continental slope extends between the Aleutian Islands and the Soviet Far East and divides the Bering Sea into the wide, shallow Bering Sea shelf and the deep Aleutian Basin (fig. 1). The continental slope and outer shelf of the U.S. part of the Beringian margin is dissected by seven large submarine canyons, several of which are among the world's largest (Carlson and Karl, 1988). Recently collected GLORIA sidescan-sonar imagery has documented the importance of mass movement in shaping the entire Beringian continental margin.

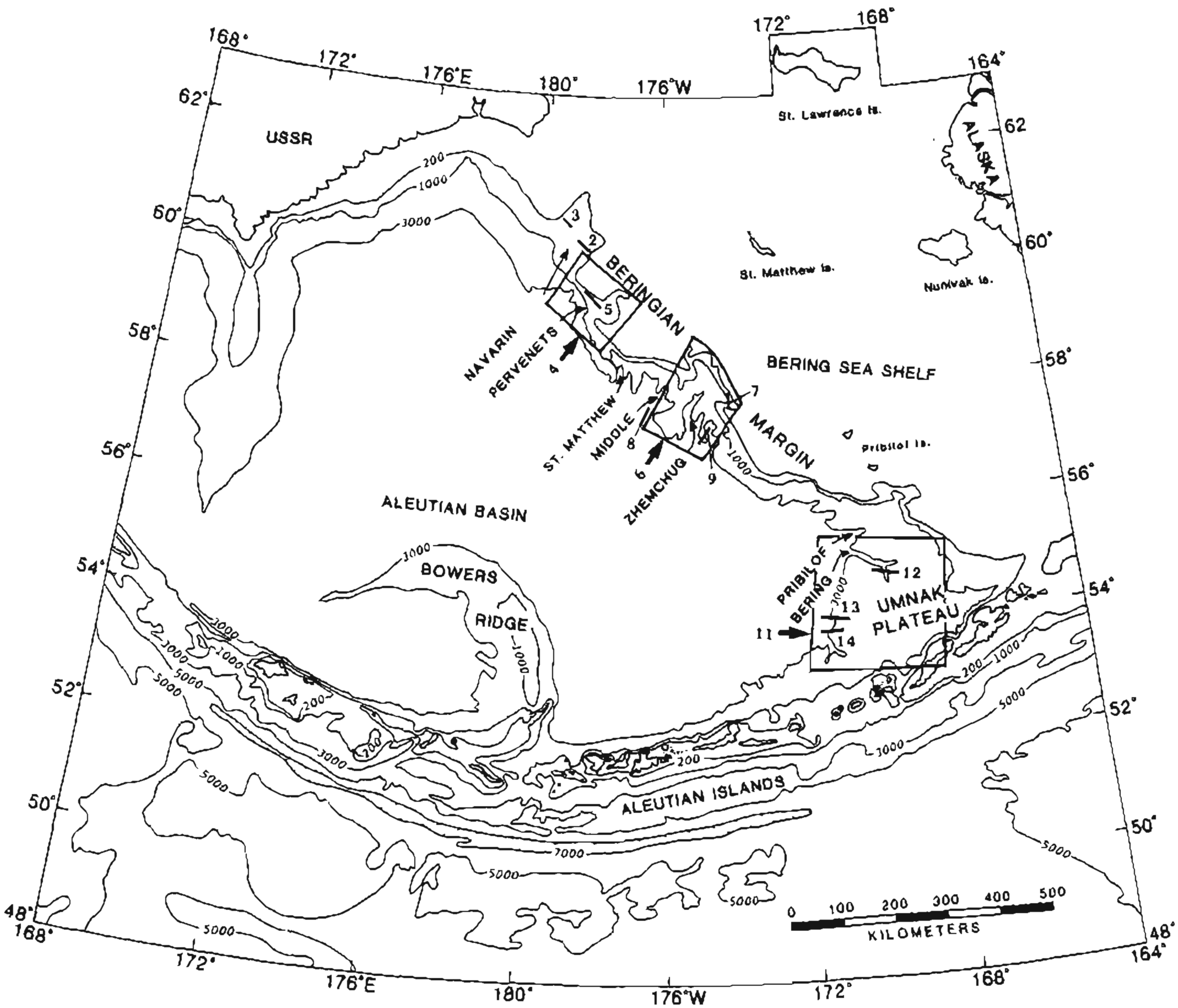


Figure 1. The Beringian margin, incised by several large canyons, lies between the Bering Sea shelf and the Aleutian basin. The seismic-reflection profiles and GLORIA images shown in figures 2 to 14 are indicated by numbered lines and outlined areas.

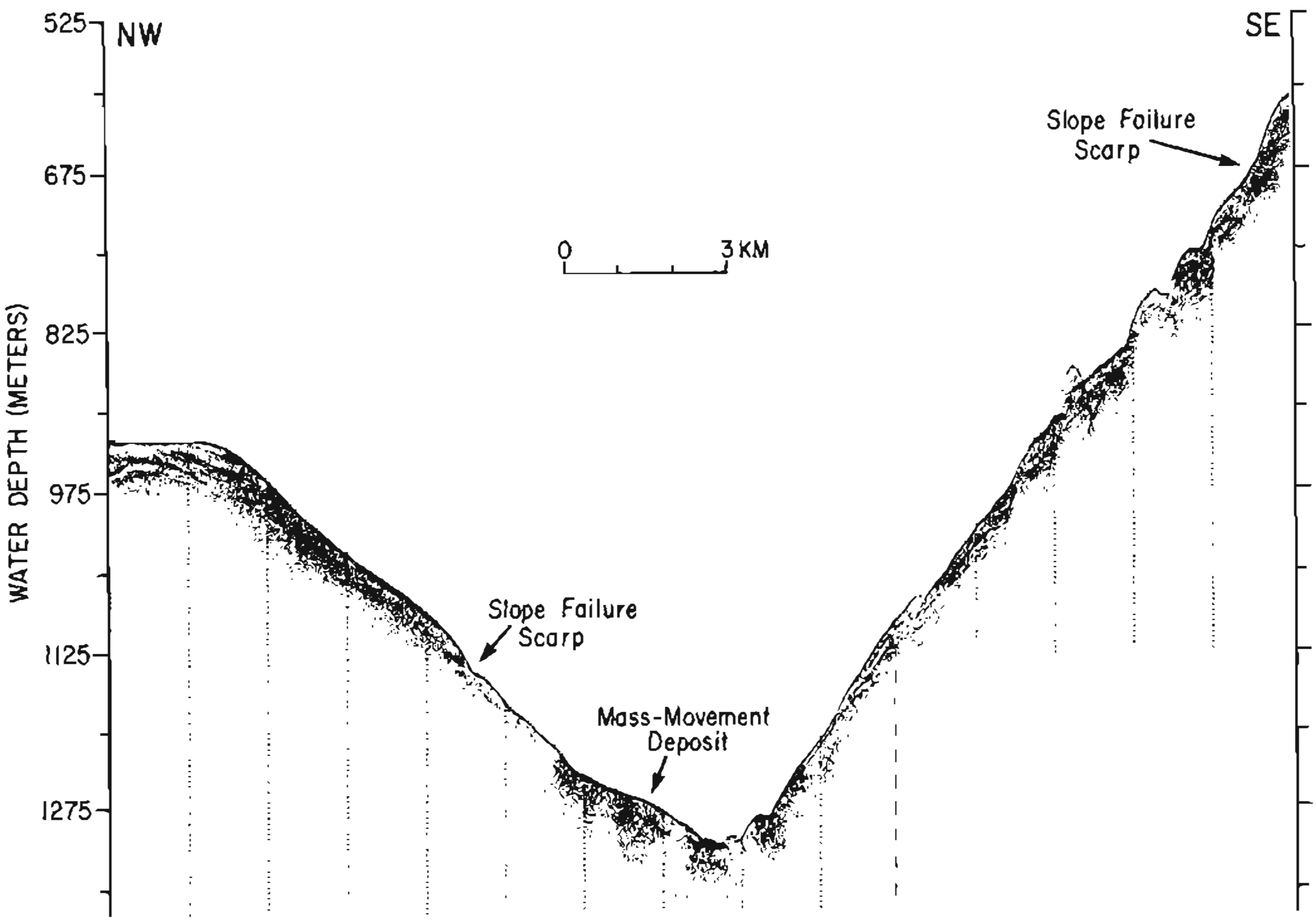


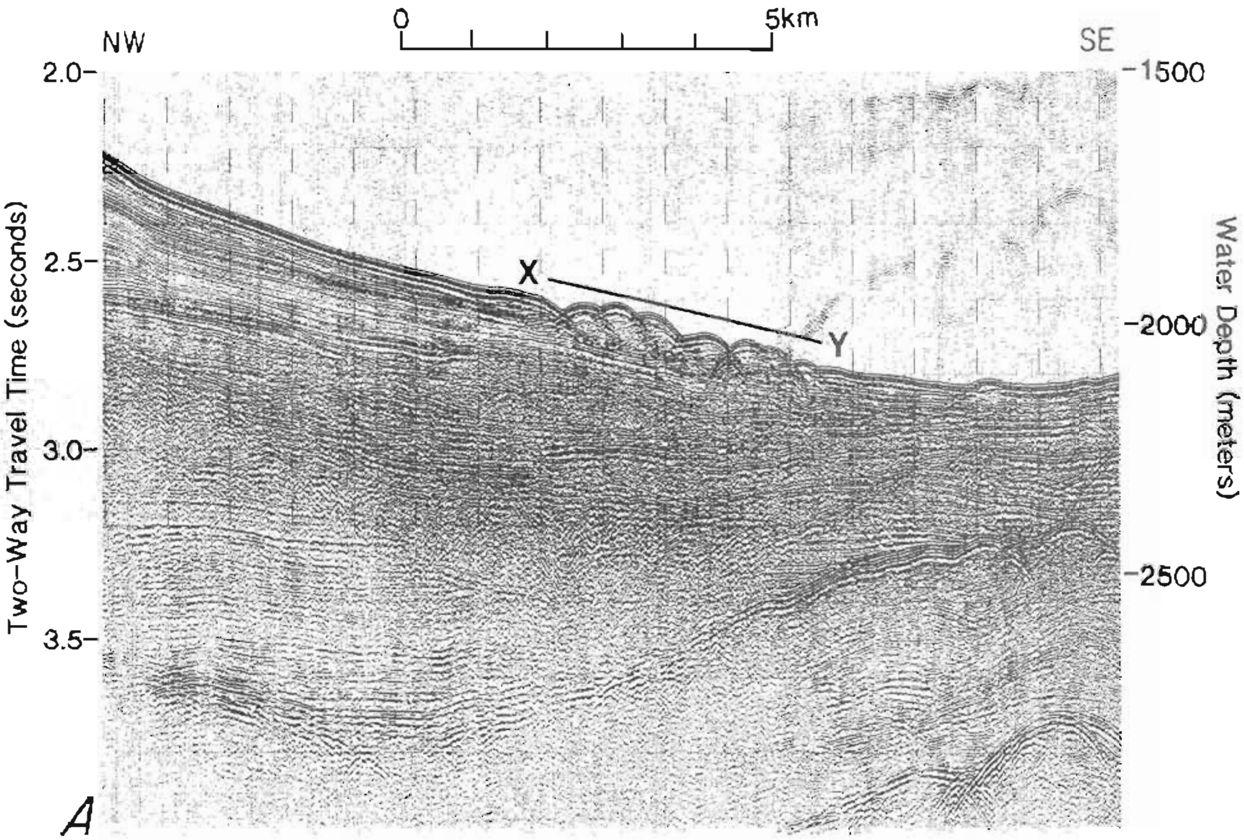
Figure 2. Seismic-reflection profile (3.5 kilohertz) across southeast branch of Navarin Canyon showing slope failure scarps, about 15 meters high, on both walls and mass-movement deposits several tens of meters thick on the canyon floor.

Soviet scientists were some of the first to study the Beringian margin and discovered several of the canyons (Kotenev, 1965). U.S. scientists began to study the area in the late 1960's (Scholl and others, 1968), and petroleum lease sales efforts in the 1970's and early 1980's intensified scientific interest. In 1986, the U.S. Geological Survey conducted three GLORIA (Geological LOng-Range Inclined Asdic) sidescan-sonar surveys that covered the entire Beringian slope and provided a reconnaissance image of this frontier region.

This report illustrates the large canyons that are incised in the margin and demonstrates that these canyons have been shaped by several styles of mass movement, including mudflows, debris flows, slumps, and massive slides. The illustrations show bathymetric, GLORIA sidescan-sonar, and seismic-reflection data. We discuss selected mass-movement features of the Beringian margin in three segments, beginning in the north with Navarin Canyon, which straddles the U.S.-U.S.S.R. 1867 convention line, then shifting to the central part of the margin, which is dominated by Zhemchug Canyon, and ending with the southern margin, which includes two large canyons and an oceanic plateau (fig. 1).

NORTHERN MARGIN

Navarin Canyon, the most northerly (fig. 1) and the second largest of the Beringian canyons, is 258 kilometers (km) long from its head to the base of the slope and has a width and relief, where it cuts the shelf edge, of 100 km and 1,150 meters (m), respectively (Carlson and Karl, 1988). Many of the seismic-reflection profiles across Navarin Canyon show slope-failure scarps on both walls and mass-movement deposits on the canyon floor (fig. 2). Mass-movement features were identified along 830 km of a total of 2,570 km of seismic-reflection profiles (32 percent) collected across the northern Beringian margin (Carlson and Karl, 1984/85). An example of a relatively thick landslide mass on the floor of Navarin Canyon is presented in figure 3. Figure 3*A* shows a seismic-reflection profile collected in 1980 that crosses the axis of Navarin Canyon transversely



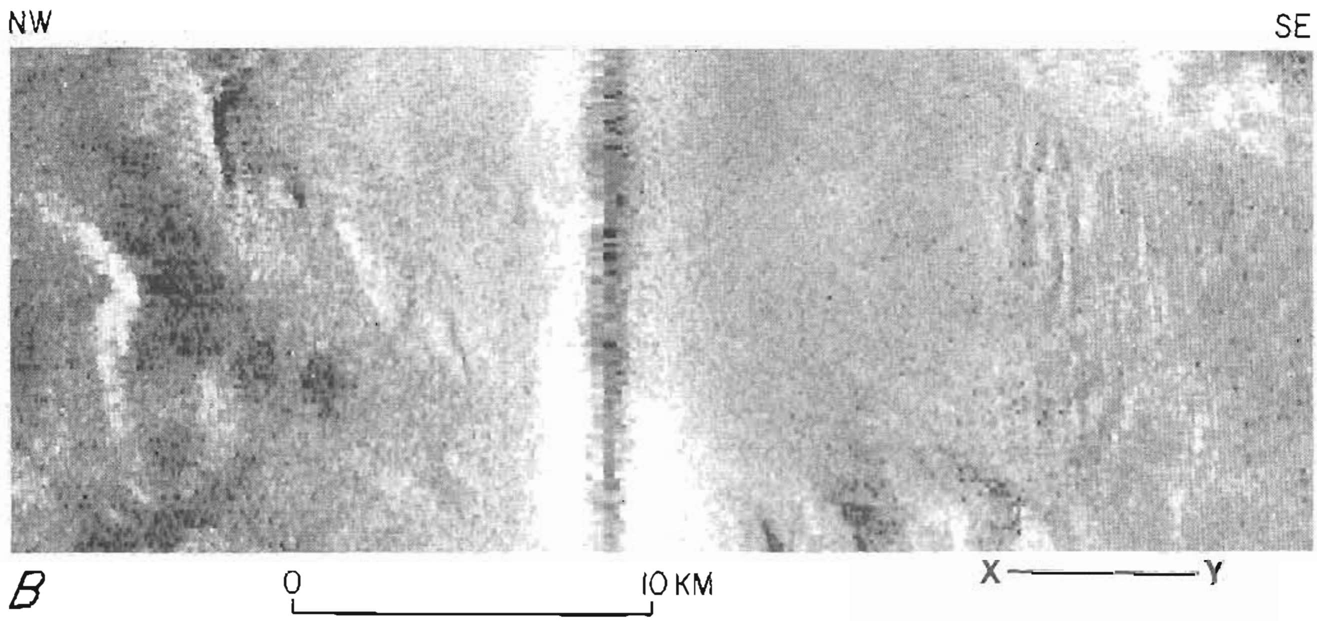


Figure 3. Two views of a mass-movement deposit on the floor of Navarin Canyon. (*A*) seismic-reflection profile showing hummocky, crumpled sediment constituting the toe of a mass-movement deposit. The ship track was run transverse to the axis of the canyon (vertical exaggeration is 10×). (*B*) GLORIA sidescan-sonar image showing same hummocky toe (X–Y). The width of this deposit is about 10 kilometers. The trackline of the GLORIA image is located on the canyon wall about 5 kilometers to the left (northwest) of the seismic-reflection profile.

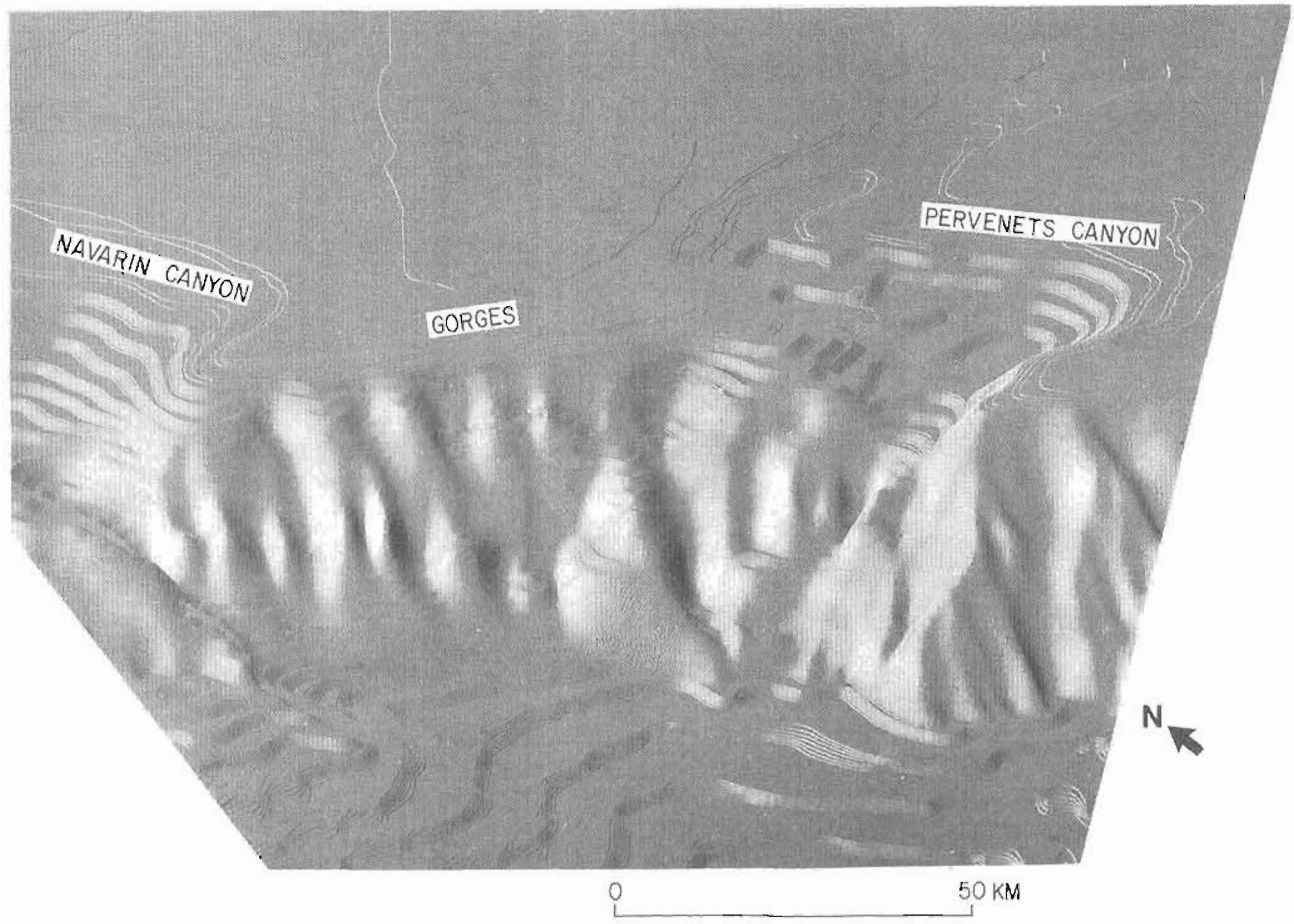


Figure 4. Three-dimensional shaded-relief image of bathymetry showing relatively small canyons (gorges) incised in the slope between Navarin and Pervenets canyons. This computer-processed image uses a look-angle of 45° above the horizon and has a vertical exaggeration of 5×.

and shows the hummocky nature of the 100-m-thick toe of a landslide (X–Y). In 1986, our GLORIA survey of this area imaged this same landslide deposit (fig. 3*B*). The seismic-reflection profile shows that the landslide deposit is about 100 m thick. By combining the map view provided by the GLORIA image and the sediment thickness provided by the seismic profile, we can calculate the volume of sediment disturbed to be at least 5 km^3.

Several relatively small canyons, 5 to 10 km wide, are incised into the 100-km-long upper slope between Navarin and Pervenets canyons (fig. 4). GLORIA imagery of the upper slope near the heads of these small canyons, suggests that these incisions may have been initiated by sliding of large sedimentary blocks (fig. 5). These blocks are as large as several kilometers across, thus, nearly the width of the small canyons.

CENTRAL MARGIN

The central part of the Beringian margin is dominated by the massive Zhemchug Canyon (fig. 6). The submarine canyon, with a volume at 5,800 km^3 is the world's largest. It is 168 km long from its head to the base of the slope, and at the shelf edge, its incision has a relief of 2.6 km and a width of 100 km (Carlson and Karl, 1988). Most seismic-reflection profiles across the Zhemchug Canyon system show evidence of mass movement. Some mass-movement deposits are composites of multiple events that accumulated on the floor of the canyon and severely constricted the axis of the canyon (fig. 7). Such constrictions along the axis of the canyon inhibit water circulation as well as mass flows down the canyon. In the development of submarine canyons, as well as subaerial canyons, however, such constrictions or dammings are geologically short lived, as long

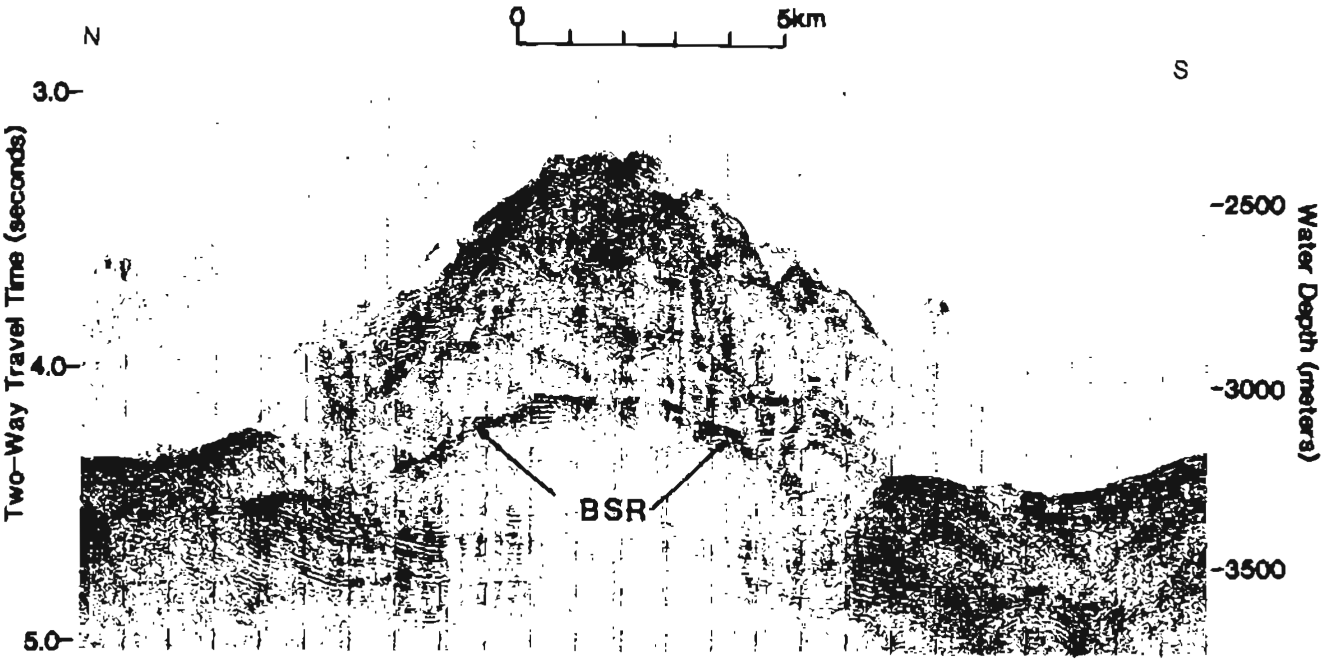


Figure 9. Seismic-reflection profile across large slide block near the mouth of Zhemchug Canyon. The bottom-simulating reflector (BSR) is probably a silica diagenetic boundary.

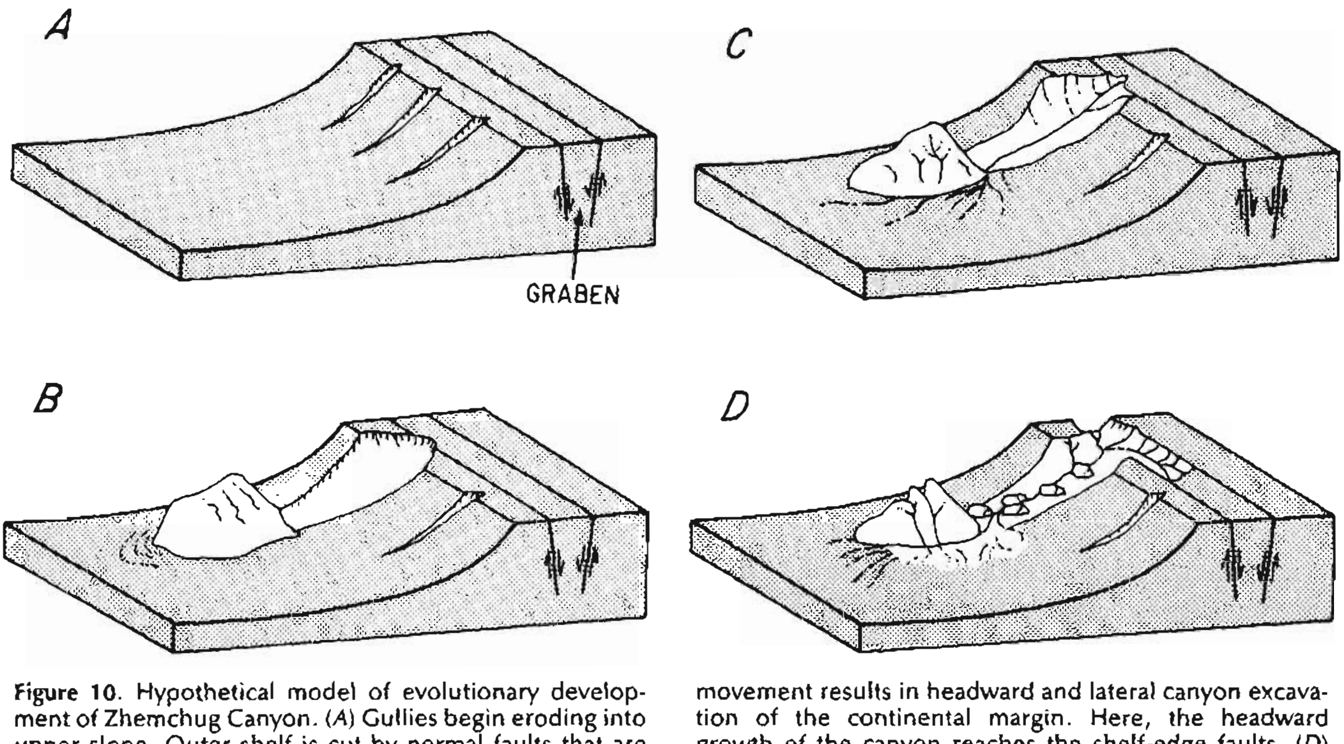


Figure 10. Hypothetical model of evolutionary development of Zhemchug Canyon. (*A*) Gullies begin eroding into upper slope. Outer shelf is cut by normal faults that are parallel to the shelf edge and a slight graben is formed. (*B*) Large blocks tear loose from upper slope initiating the breeching of the shelf-slope break. (*C*) Continuing mass movement results in headward and lateral canyon excavation of the continental margin. Here, the headward growth of the canyon reaches the shelf-edge faults. (*D*) Excavation of canyon progresses laterally along the fault scarps and canyon develops its modern morphology.

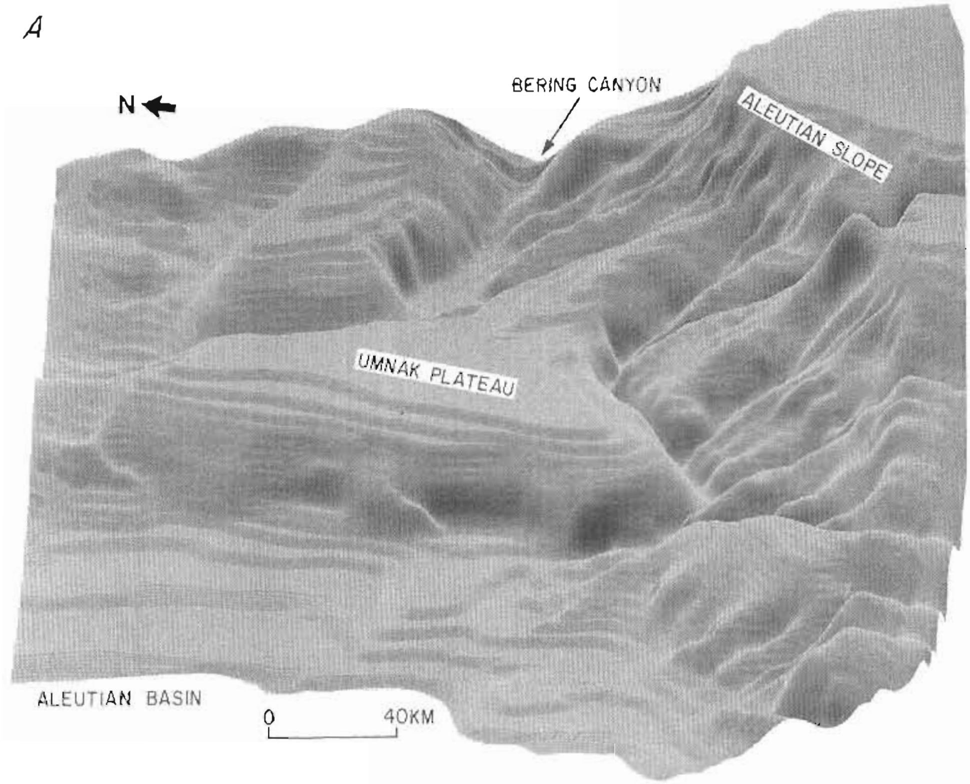


Figure 11. (*A*) Three-dimensional shaded relief image of Umnak Plateau and adjacent continental slope with a look angle of 20° from the horizontal and a vertical exaggeration of 10×.

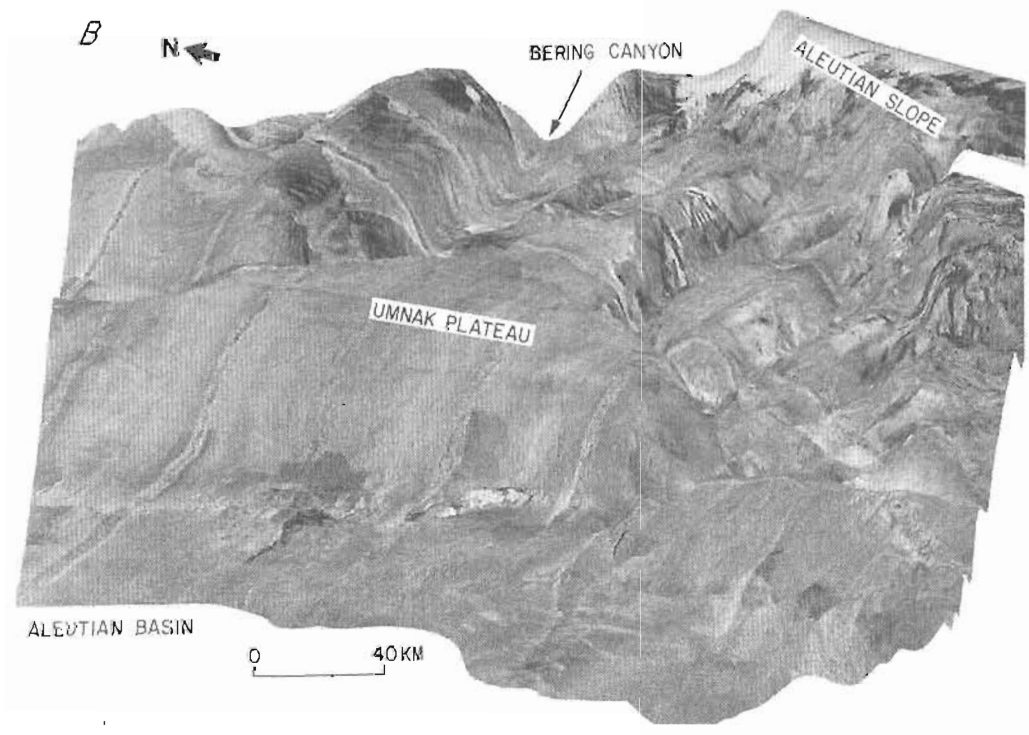


Figure 11. (*B*) GLORIA image of same area merged with bathymetry.

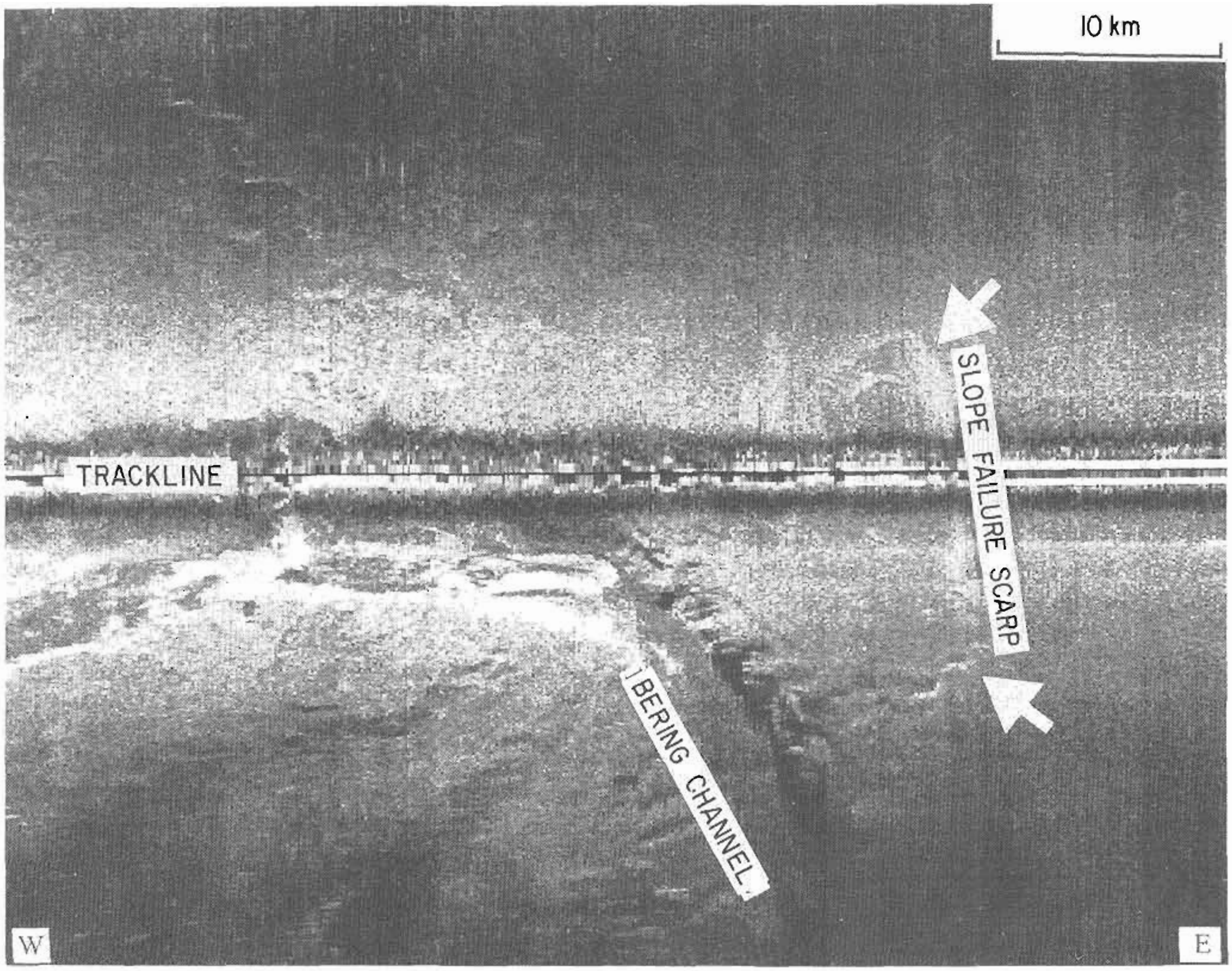


Figure 12. GLORIA sidescan-sonar image of a slope failure on eastern edge of Umnak Plateau, adjacent to Bering Canyon. Note that much of the debris from the slope failure has been removed from the channel. Based on the size of the slope-failure scarp, more than 18 cubic kilometers of sediment appears to have been incorporated in this mass movement.

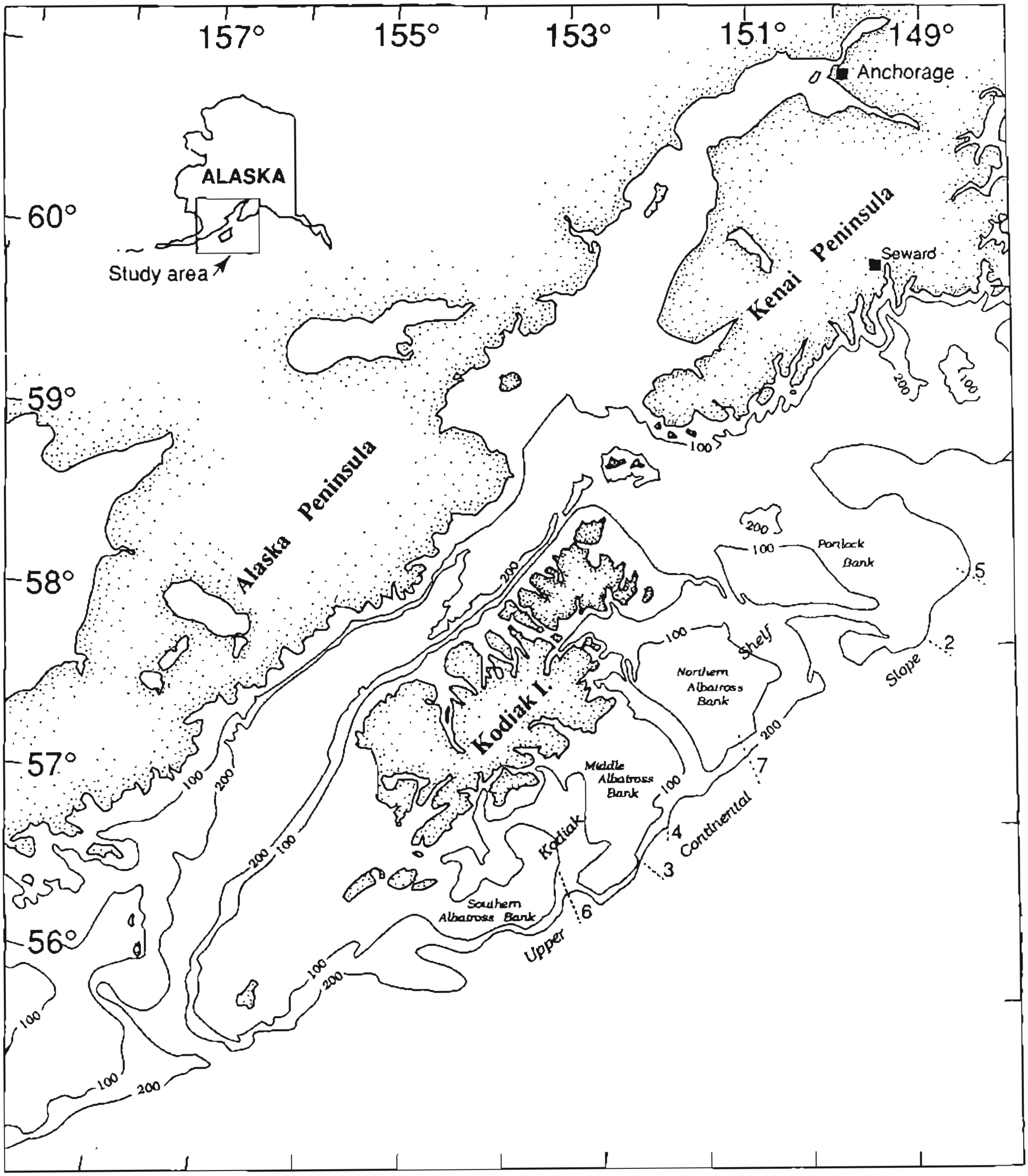


Figure 1. Location of the Kodiak upper continental slope seaward of the Kodiak shelf. Numbered dashed lines indicate location of the seismic-reflection profiles in figures 2 to 7.

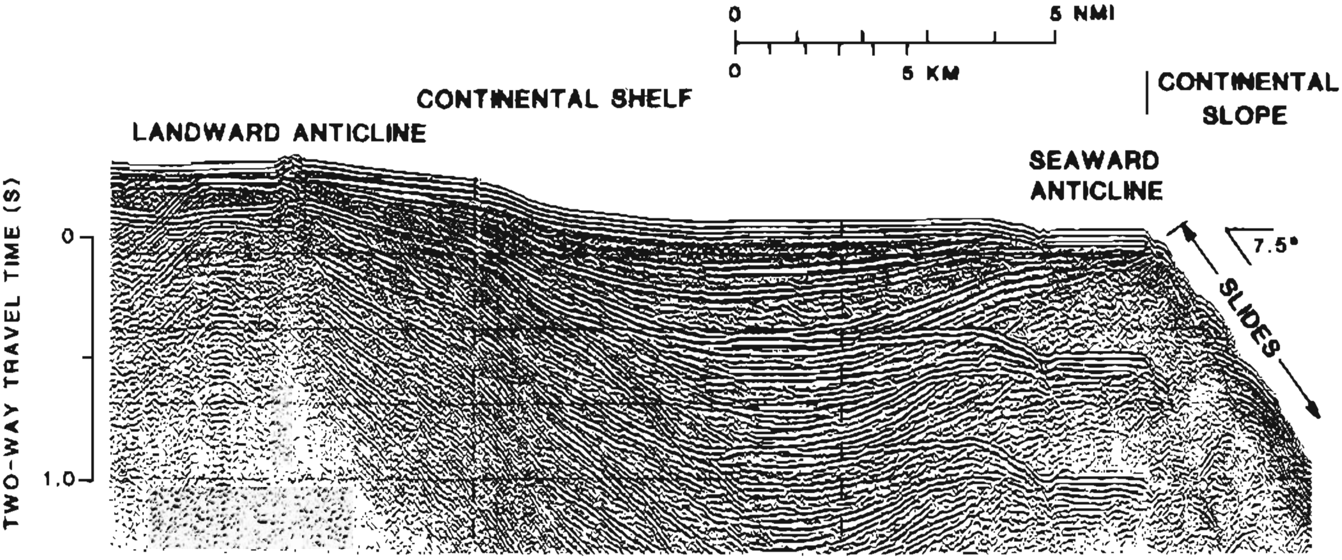


Figure 6. A more mature example of the multiple anticlinal folds shown in figure 5, where the seaward anticline has grown sufficiently to extend the shelf seaward and define a new *shelf-slope break.* (One second of two-way travel time indicates a water depth of approximately 750 meters.)

Figure 7. Relatively mildly deformed outer shelf and upper slope on northern Albatross Bank. The subsurface strata are broadly folded, the inclination of the upper continental slope sea floor is gentle, and no large slumps occur, at least to the seaward extent of the seismic profile. (One second of two-way travel time indicates a water depth of approximately 750 meters.)

that all but the steepest areas are stable under static, gravitational loading (downslope- directed forces of gravity acting on the sedimentary deposit due to the sea-floor slope angle), but that earthquake loading can cause slope failure (Hampton and others, 1978; Hampton, 1989). Earthquakes probably triggered many of the slides on the Kodiak upper continental slope.

Another factor associated with tectonism is removal of the buttressing support of slopes by earthquake-associated faulting. Such an example is shown on the seismic-reflection profile in figure 3. The slump occurs on the landward slope of a basin that is filled by a thick sequence of well-stratified sediment. Offset of the sea floor within the basin, just seaward of the toe of the slide mass, and discontinuity of acoustic reflectors beneath the offset suggest the presence of a recently active fault. Hampton and others (1978) proposed that repeated movement of the fault removes support from the toe of the slide mass and, thereby, promotes sliding.

SUMMARY

Geologic analysis of seismic-reflection profiles indicates that large slumps on the Kodiak upper continental slope occur only where tectonic deformation has occurred recently, with the implication that sea-floor steepening and removal of support by faulting control the location of the slumps. Smaller, shallow slides occur in all areas of the upper continental slope, implying that their location is independent of sea-floor deformation. By analogy with similar slides studied on land, their location probably is determined by the local presence of a buried weak sedimentary layer or an abrupt strength increase at the depth of failure. Limited geotechnical data support the contention that earthquakes actually trigger slides in the region.

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Submarine Landslides That Had a Significant Impact on Man and His Activities: Seward and Valdez, Alaska

By M.A. Hampton, R.W. Lemke[1], *and* H.W. Coulter[2]

INTRODUCTION

On the evening of March 27, 1964, the largest earthquake to occur in North America during this century struck southern Alaska. The epicenter of the magnitude (Mw) 9.2 shock was at the head of Prince William Sound near the foot of the Chugach Mountains (fig. 1). The duration of strong ground motion lasted an incredibly long 3 to 4 minutes, with tectonically induced elevation changes occurring over a land and sea-floor area exceeding 180,000 square kilometers (km^2). Of the several destructive earthquake-related effects, slope failure (landslides) probably caused the most property damage, and large sea waves (tsunami), most of which were generated by submarine landslides, took the most lives (Hansen and Eckel, 1966). These two phenomena were particularly devastating to the communities of Seward and Valdez (Lemke, 1967; Coulter and Migliaccio, 1966). Large landslides originated below sea level and retrogressed landward to destroy the waterfront of both towns, sinking the dock and harbor facilities that were their economic lifeblood. Sea waves generated by

[1] Consulting Engineering Geologist, Lakewood, Colo.
[2] Retired, Washington, D.C.

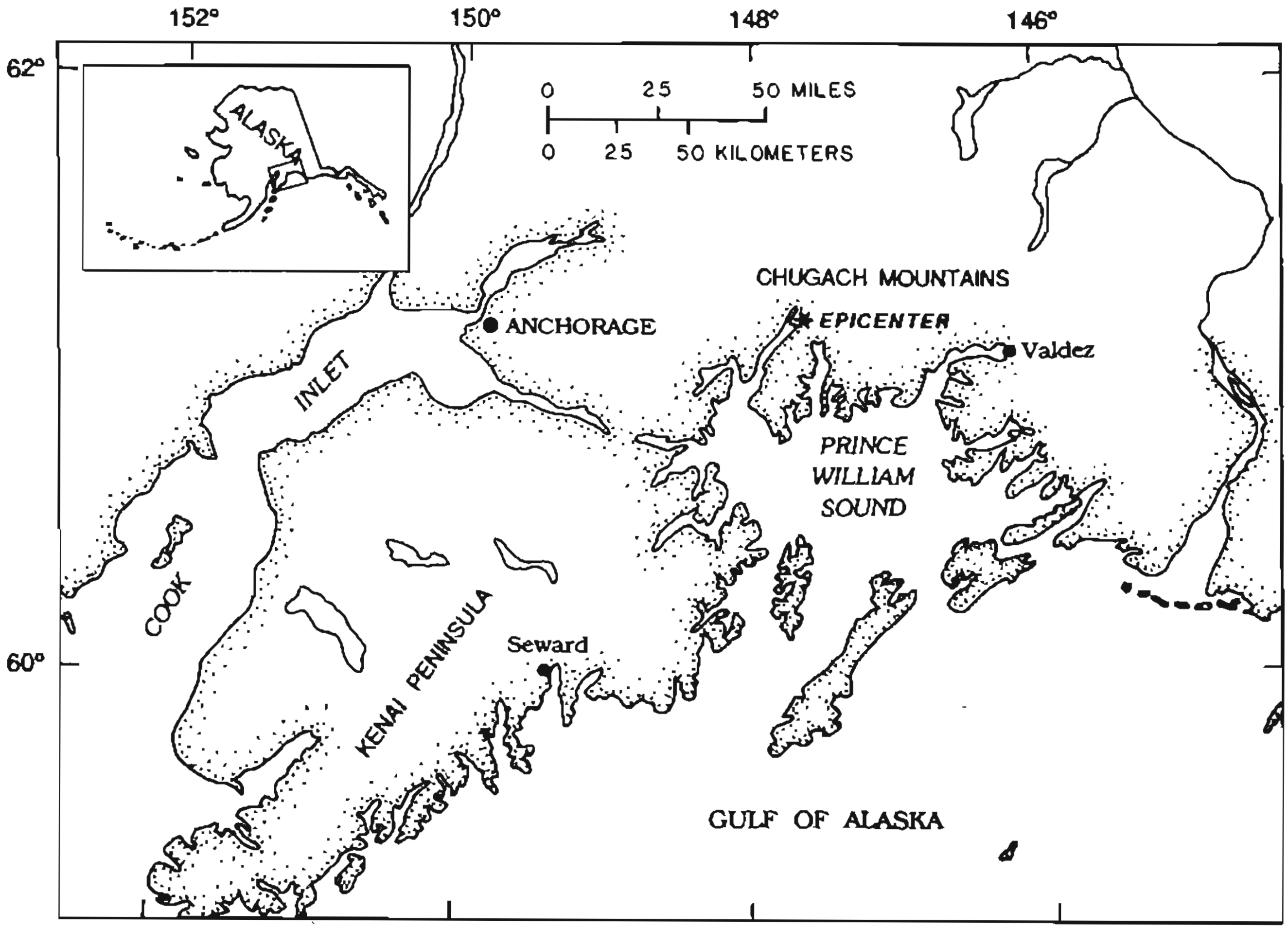
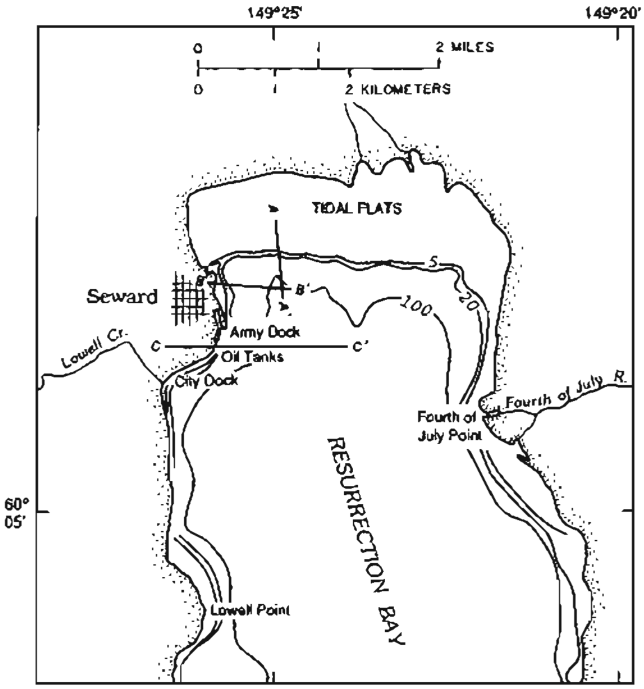


Figure 1. Location of the epicenter of the Alaskan earthquake of 1964 and the towns of Seward and Valdez.



◀ Figure 2. Northern part of Resurrection Bay showing Seward, Alaska, and vicinity. (Contours are in feet.)

the landslides caused further destruction, both by flooding and by spreading burning oil from local storage depots.

The geologic settings of Seward and Valdez are similar, a fact that explains the similarity of earthquake effects. Seward is located near the head of Resurrection Bay and Valdez at the head of Port Valdez, both steep-sided fjords (figs. 2 and 3). The landslide-affected areas are on deltas constructed of stream-transported sediment that derives from a nearby steep, glaciated mountain front. This type of sediment, especially when composed of sand- and silt-size grains, is particularly susceptible to liquefaction during earthquakes (Committee on Earthquake Engineering, 1985; Lee, Schwab, and Booth, this report; Schwab and Lee, this report). Landslide displacement of water in the restricted bays creates an ideal situation for the generation of destructive waves that reflect back and forth across the narrow fjords and repeatedly inundate the coast.

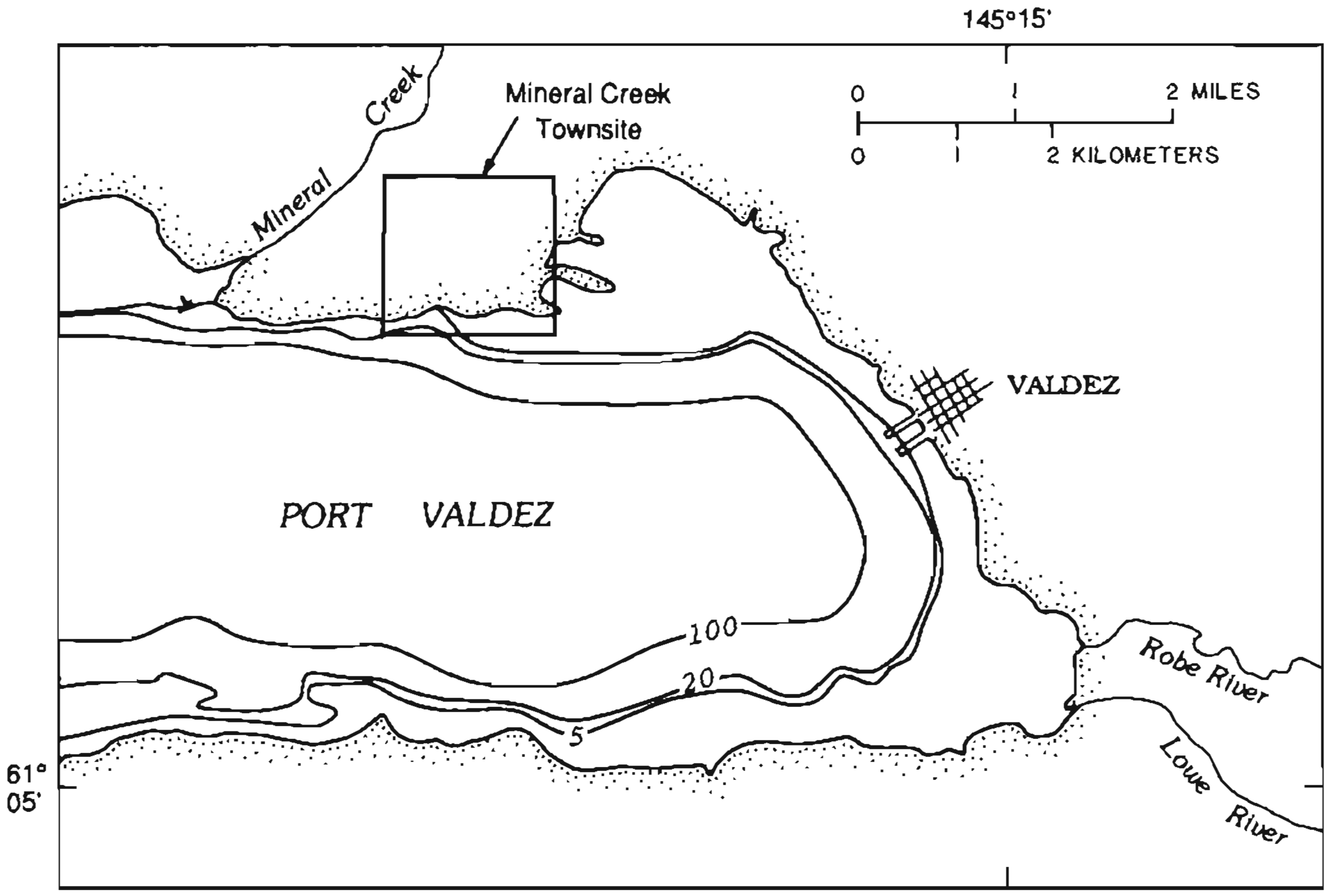


Figure 3. Eastern part of Port Valdez showing the town of Valdez, Alaska, and vicinity. Note the Mineral Creek townsite, a safer location where the town of Valdez was moved after the earthquake. (Contours are in feet.)

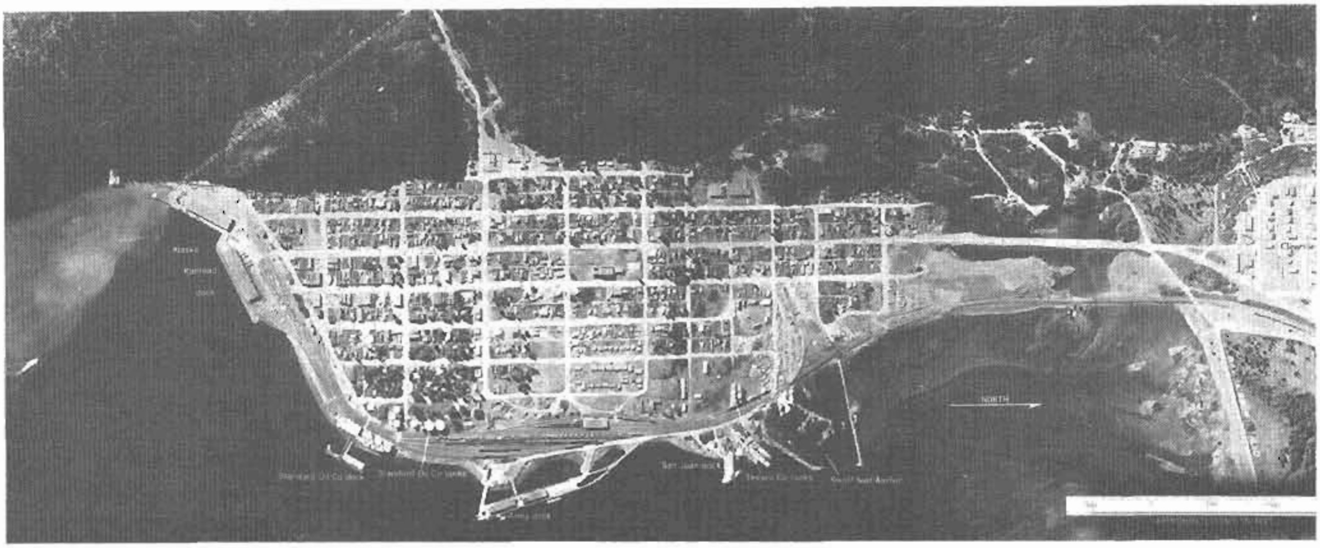


Figure 4. Seward before the earthquake (from Lemke, 1967). (Scale is in feet.)

Figure 5. Seward 1 day after the earthquake (from Lemke, 1967). Note the white line that indicates the landward limit of wave runup. The most obvious indication of the landslide effects at the scale of the photograph is the missing docks and other coastal facilities (see fig. 4).

THE EARTHQUAKE DAMAGE AT SEWARD

Seward is located principally on the Seward delta of Lowell Creek (fig. 2), about 150 km south southwest of the earthquake's epicenter (fig. 1). After about 35 to 40 seconds of earthquake shaking on March 27, 1964, a strip of waterfront 1,200 meters (m) long and 15 to 150 m wide started to subside slice by slice and eventually disappeared into the bay (figs. 4 and 5). This was a consequence of landward retrogression of a slope failure that initiated on the steep (20° to 35° to a water depth of 50 m) submerged delta-front. By the time shaking stopped, a zone of incipient slope failure marked by cracked ground had formed for a distance up to 250 m behind the landslide (figs. 6–8). Nearshore water depths increased more than 30 m in some places (fig. 9). Delta sediment in the northwest corner of the bay, deposited from the Resurrection River, also failed, and the sea surface deformed, presumably in response to the sudden underwater sediment movement. Within 30 seconds after the slope failure began, the first of three landslide-generated sea waves attacked the shoreline, continuing for about 15 minutes and surging to elevations of approximately 10 m above sea level. Wave runup resumed when another train of waves, generated by tectonic displacement of the sea floor outside Resurrection Bay, rather than by the landslides, hit the coast about 30 minutes after the earthquake.

The land that slid into Resurrection Bay took docks, railroad equipment, and oil storage tanks with it (fig. 10). Fire erupted when storage tanks overturned and pipes

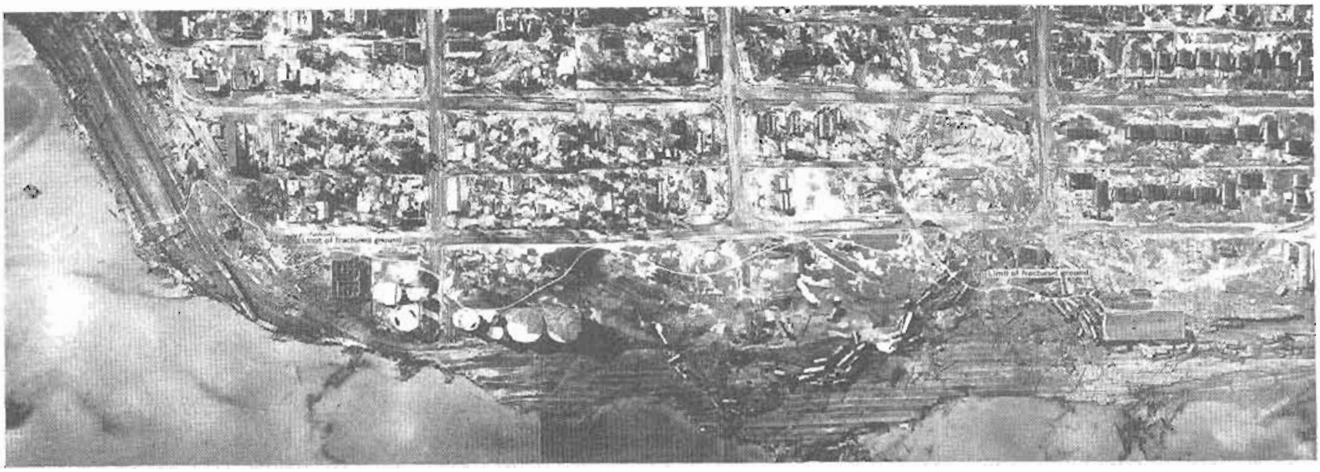


Figure 6. Southern part of the Seward waterfront showing the limit of fractured ground associated with landslides (from Lemke, 1967).

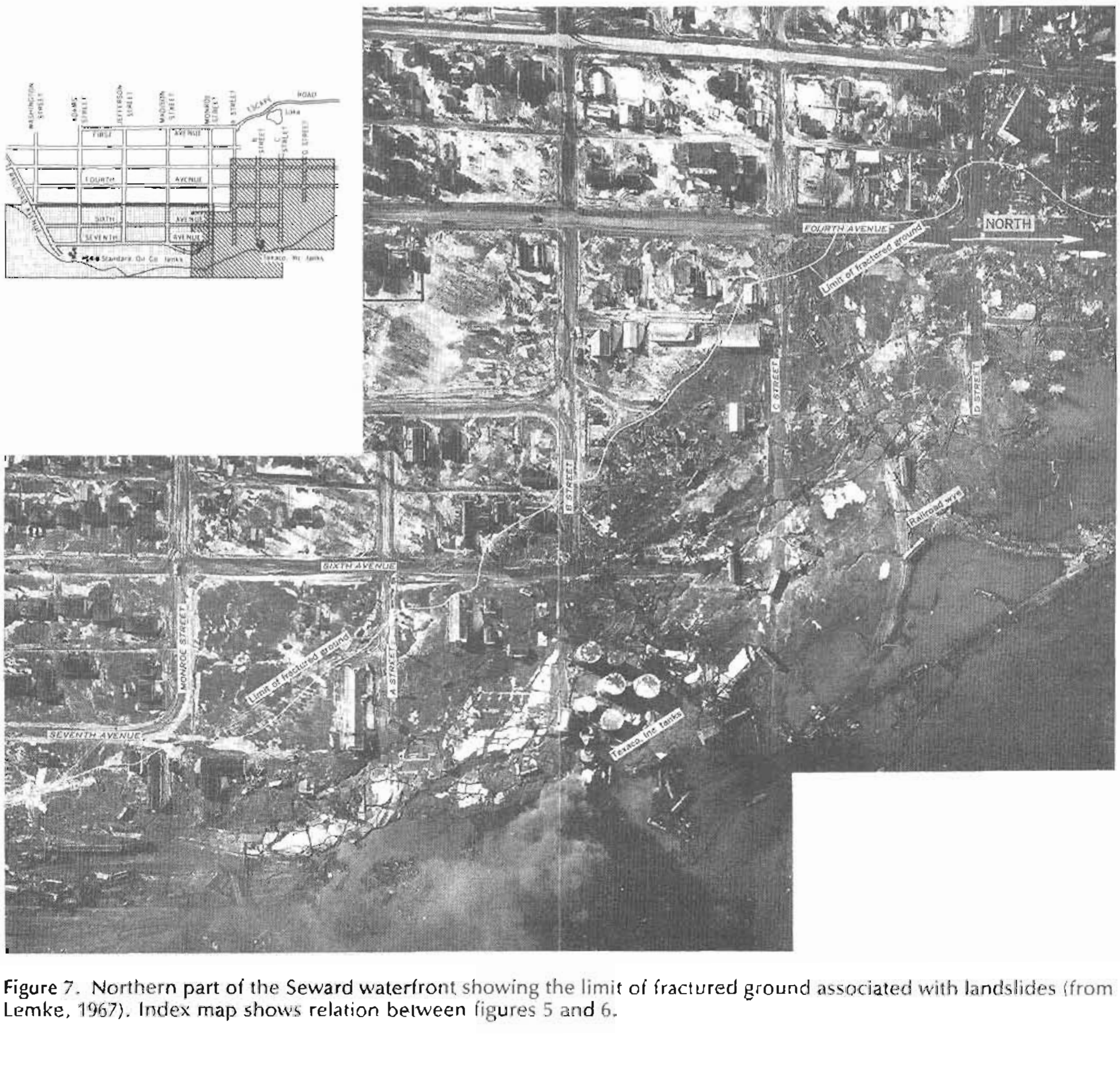


Figure 7. Northern part of the Seward waterfront showing the limit of fractured ground associated with landslides (from Lemke, 1967). Index map shows relation between figures 5 and 6.

Figure 8. An example of ground displacement (approximately 1.5 meters) and fractures associated with landslides, 30 meters from the shoreline (from Lemke, 1967).

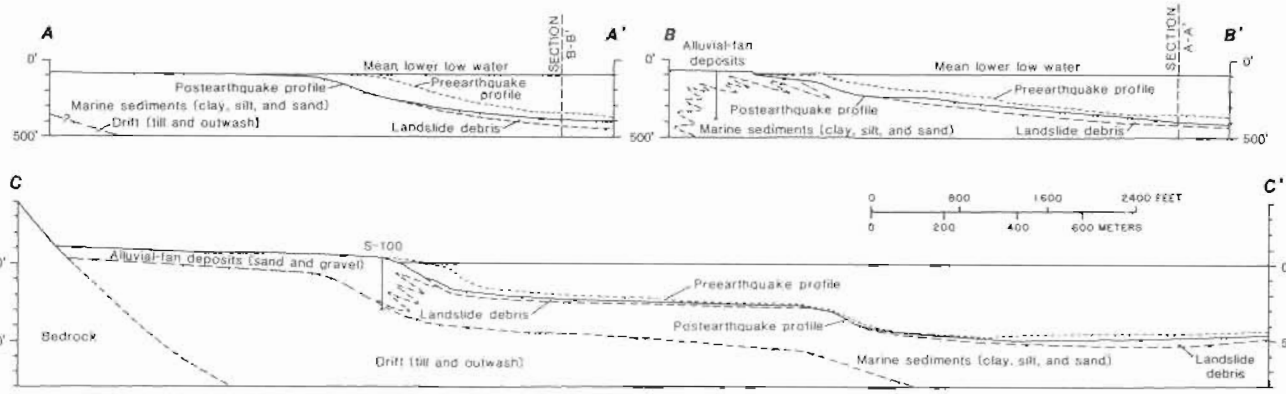


Figure 9. Cross sections showing pre- and postearthquake profiles of the land surface in Resurrection Bay (from Lemke, 1967). Note that the sea floor has been lowered up to 30 meters by the displacement of sediment by the landslides. The locations of these cross sections are shown on figure 2.

ruptured. Water that receded from the shore as a result of the initial landslide displacement carried the fire offshore, then sea waves returned it with each subsequent runup. The waves lifted railroad cars off the tracks, smashed earth-moving equipment, lifted boats over a breakwater, displaced houses from their foundation, and dumped a cargo

Figure 10. Landslide destruction of railroad and petroleum-storage facilities on the Seward waterfront (from Lemke, 1967).

dock into the bay (fig. 11). The death toll at Seward was 13 people, mostly as a result of wave runup. Damage to property was estimated at $22,363,349.

Geologic and engineering studies conducted after the earthquake identified several factors that contributed to the landslides at Seward (Shannon and Wilson, Inc., 1964a; Lemke, 1967). Onshore and offshore subsurface borings were made to identify sediment types, and physical properties were measured on sediment core samples. Offshore seismic-reflection profiling provided acoustic data for description of sea-floor morphology and subbottom stratigraphy. Obviously, the slope failures were related to the earthquake. Strong ground motion initiated the slope failures, and the long duration of the earthquake led to the great extent of mass movement. Some of the deltaic sediment in Resurrection Bay, such as rapidly deposited, interbedded fine- and coarse-grained layers, is known to build up pore-water pressures and liquefy under repeated, earthquake-type loading, even if the packing of the sediment grains is medium to high density (Committee of Earthquake Engineering, 1985). A large component of gravitational force acted downslope because of the steep delta fronts, and the weakness along sedimentary bedding planes that dip up to 30° seaward probably lessened the applied force necessary for failure.

The less obvious factors concern static pore-water pressure conditions within the sedimentary deposit (see Lee, Schwab, and Booth, this report). In particular, sediment borings revealed the presence of high artesian pressure in some confined, buried sedimentary beds, which decreased the stabilizing frictional stress ("effective stress") that acts between grains, thereby weakening the sediment irrespective of earthquake loading. Moreover, the low tidal level at the time of the earthquake, along with the rapid drawdown of water in the bay that accompanied the initial slope failure, also decreased the stability because the pore water could not drain from the sediment quickly enough to maintain hydrostatic conditions during the removal of external water pressure. Submarine landslides on fjord deltas elsewhere have occurred in conjunction with tidal drawdown, without earthquake loading (Terzaghi, 1956; Prior and others, 1981; Prior and others, 1982).

Man-induced loading in the form of artificial fill, dock facilities, oil storage tanks, and railroad cars aided the slope failure, as did the added surcharge due to wave runup. The magnitude of these man-induced effects relative to others is uncertain, however.

Seismic-reflection profiles point out that postearthquake sea-floor slopes are as steep as those before. A stability analysis based on measured solid properties implies that pre- and postearthquake slopes are stable under static conditions but could readily fail under earthquake accelerations of 0.15 g (Shannon and Wilson, Inc., 1964a). Contrary to normal expectations, the overall stability of the offshore sediment was not improved because of the slope failure.

A scientific and engineering task force was formed to study the earthquake effects and, thereby, aid the reconstruction effort. On July 24, 1964, the task force made recommendations to Seward city officials regarding the

Figure 11. Houses and other debris carried by earthquake-related waves into the lagoon at the north end of Seward (from Lemke, 1967).

stability of the Seward delta. The recommendations were based on the geologic studies of Lemke (1967), the geotechnical engineering studies of Shannon and Wilson, Inc. (1964a), and visits by the task force to Seward (Hansen and others, 1966). The delta area was divided into two categories: (1) nominal risk and (2) high risk (fig. 12). The high-risk category included much of the waterfront area and was essentially coincident with the area of fractured ground shown in figures 6 and 7. On the basis of these recommendations, the city officials restricted the high-risk area, formerly used for railroad marshalling and by the oil tank farms, to parkland use. There was little doubt that under resumed dynamic conditions of severe earthquake shaking, the area of fractured ground would again fail and begin sliding into Resurrection Bay.

THE EARTHQUAKE DAMAGE AT VALDEZ

At Valdez, located about 70 km east of the earthquake's epicenter (figs. 1 and 3), the greatest damage was associated with a delta-front landslide involving an estimated 75 million cubic meters of sediment. The landslide was first observed indirectly when, shortly after shaking began, a cargo ship unloading at the town dock began to toss violently, with vertical motion in excess of 10 m and rolls of 50°. The motion was caused by landslide-generated sea waves. Shoreward retrogression of the landslide soon reached the docks, causing them to vanish instantly into the turbulent water. The entire waterfront was lost, along with warehouses, a cannery, heavy equipment, and 30 people (figs. 13–15). The initial 10-m-high wave inundated the

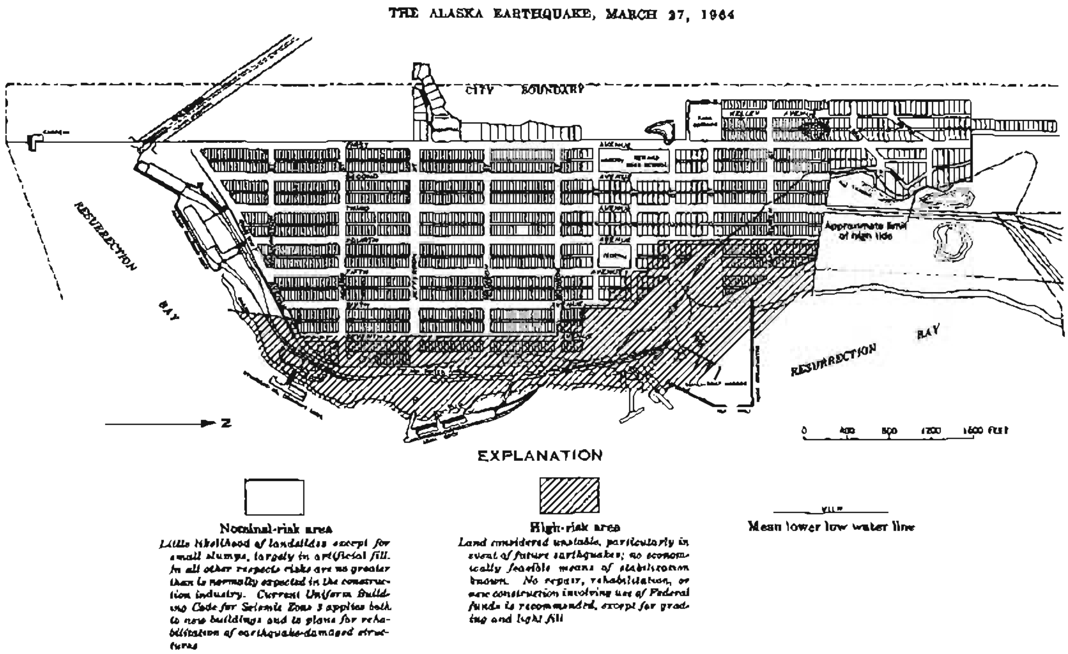


Figure 12. High- and nominal-risk areas of a part of Seward, as concluded by the scientific and engineering task force (from Hansen and Eckel, 1966).

waterfront and propagated westward down the bay where it surged to 50 m above sea level. A wave of similar size, probably reflected from the far end of the bay, struck Valdez about 10 minutes after the earthquake. Later that night, at 11:45 p.m. and 1:45 a.m., tidally augmented seiche waves (an oscillation of the surface of the bay) again advanced into the town. Waves surged a distance exceeding 500 m inland at Valdez. Parts of the shore area subsided because support was removed from the face of the delta. Subsidence continued for several months after the earthquake.

The sediment along the waterfront consists of a surficial 7- to 10-m-thick layer of sandy gravel fill that is underlain by an unknown thickness of gravelly sand outwash with interbedded layers of silt. Both sedimentary units have loose to medium density. Scientists and engineers who studied the slope failure at Valdez speculated that the delta-front liquefied in response to the earthquake and the drop in water level, then the landslide material transformed into a mobile, low-density turbidity current that spread much of the sediment as a thin deposit away from the delta.

At least five previous seismic events in Valdez, between 1899 and 1925, were accompanied by submarine landslides, as evidenced by the occurrence of phenomena such as communication cable breaks and sudden water-depth increases (Coulter and Migliaccio, 1966). There are reports of three small submarine landslides (in the early and late 1920's and in the early 1940's) that damaged parts of the dock facilities but were not associated with earthquakes. Instead, they probably were due to loading by the docks themselves or by localized sedimentation consequent to river-control projects.

As at Seward, the stability of the delta at Valdez did not seem to be increased by the landslide, and similar large slope failures are expected in future earthquakes. This threat prompted the decision to move the entire town of Valdez to a more stable site 5.5 km to the west on the Mineral Creek fan (fig. 3). An engineering and geological study of the fan pointed out the comparative desirability of this site (Shannon and Wilson, Inc. 1964b). The area is underlain by complexly bedded, medium to very dense sand and gravel that is buttressed by bedrock ridges and apparently was stable during the earthquake. The nearshore sea-floor slope is relatively gentle to a distance of 100 to 300 m offshore, where it steepens to about 25°. Offshore sediment in part of the area is a loose, fine sand of the type that is prone to slope failure when shaken, and comparison of pre- and postearthquake water depths indicates that a small mass of this sediment about 100 m offshore failed during the earthquake. The overall stability of the Mineral Creek fan during the earthquake indicates that it is a safer location for the town of Valdez, however.

CONCLUSION

The destruction at Seward and Valdez clearly demonstrated the devastating effects that offshore landslides can have on man and his activities. Most submarine slope

Figure 13. Submarine landslide area at Valdez (from Coulter and Miliaccio, 1966). The dashed lines indicate the dock area destroyed by the landslide.

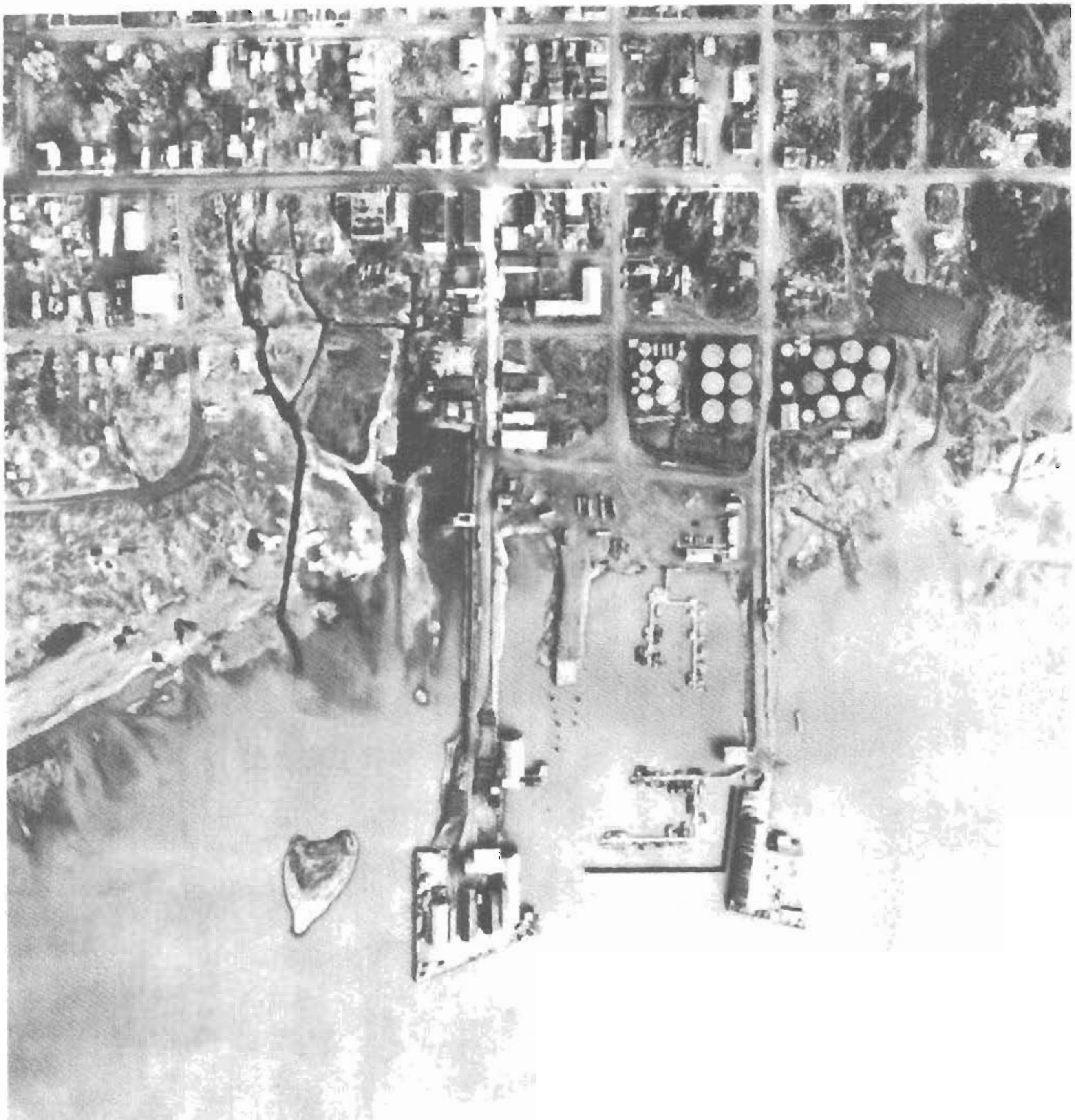


Figure 14. Valdez before the earthquake. Note the intact dock facilities extending out into Port Valdez (from Coulter and Migliaccio, 1966).

failures are benign in this regard; they occur far offshore and the range of their influence does not extend to the coastline. Moreover, the geologic conditions (coarse sediment, gentle sea-floor slope) in most nearshore areas do not permit slope failure even during the strongest seismic events; however, in those places where sandy and silty sediment is deposited rapidly on a steep sea-floor gradient, especially if earthquakes occur or if the tidal range is high, offshore landslides must be regarded as a potential hazard. This situation exists in nearly all Alaskan fjords.

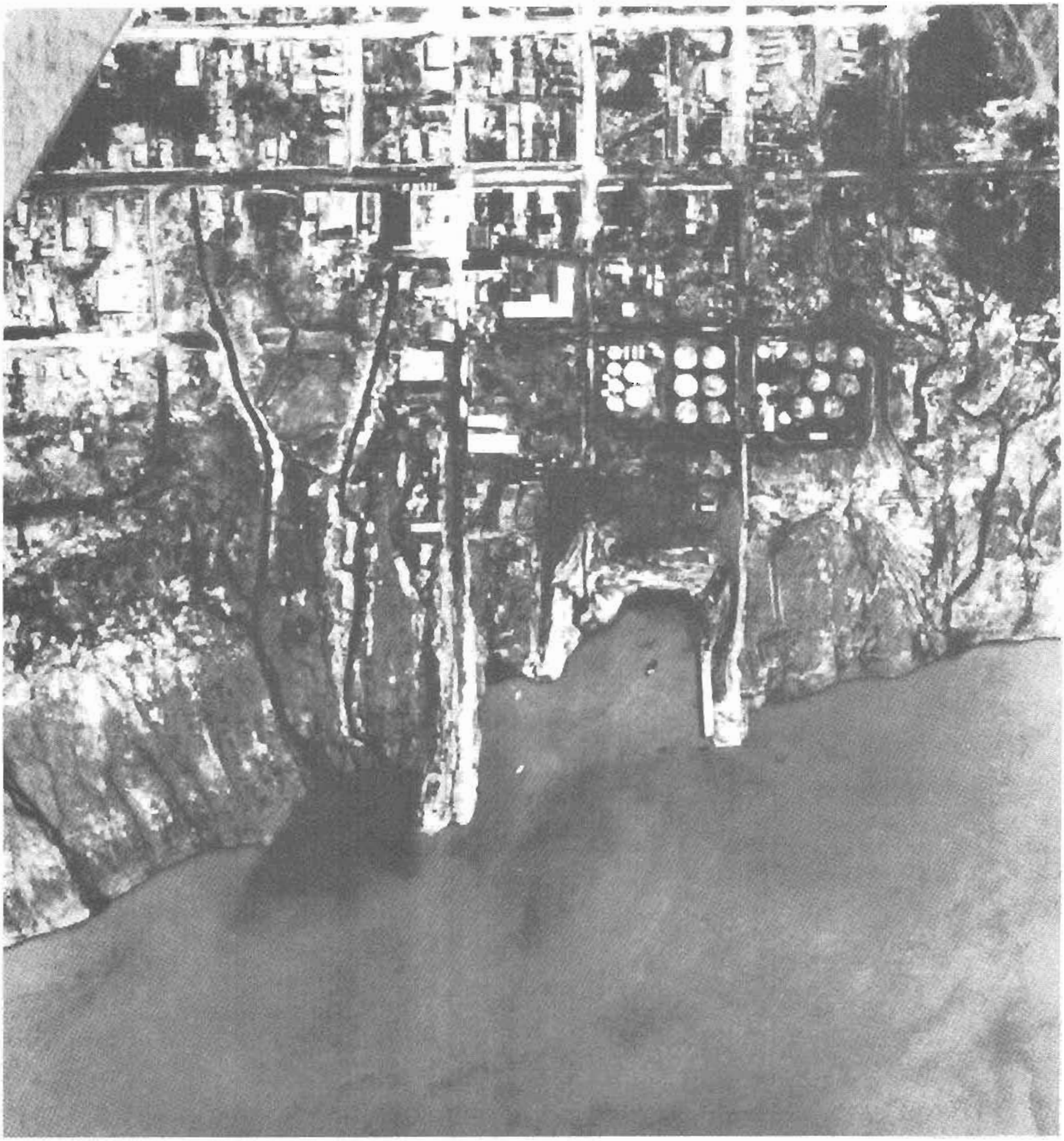


Figure 15. Valdez about 3 months after the earthquake (from Coulter and Migliaccio, 1966). Note the disappearance of the dock facilities and the devastation along the waterfront.

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Processes Controlling the Style of Mass Movement in Glaciomarine Sediment: Northeastern Gulf of Alaska

By W.C. Schwab *and* H.J. Lee

INTRODUCTION

The type of mass movement that occurs following slope failure can range from rigid block motion to fluidlike flow. As use of sea-floor resources and development of the sea floor continues to expand, it becomes increasingly desirable to develop methods that allow the ocean engineer to predict not only the relative stability of a slope but also the amount of sea-floor deformation that follows slope failure. We investigated several large areas of slope instability on the continental shelf of the northeastern Gulf of Alaska in an attempt to explain the different types of slope failure encountered (Schwab and Lee, 1983, 1988; Lee and Edwards, 1986; Schwab and others, 1987). Results of these studies are applicable to the development of a methodology that can predict the consequences of slope failure (also see Edwards, Lee, and Field, this report).

Glaciation is the dominant process contributing sediment to the northeastern Gulf of Alaska continental shelf (Molnia, 1983). Just as in the case of streams of water, glaciers perform erosion, transportation, and deposition of mineral matter. Blocks of rock being carried within the glacial ice are scraped and dragged along the rock floor, gouging and grooving the bedrock and chipping out fragments of rock. Much of the rock is ground by the glacier into extremely fine particles of silt- and clay-sized sediment: termed *rock flour*. Glaciers in the Chugach-St. Elias Mountains discharge their sediment into lakes, streams, and bays, and in turn, much of this material is transported by currents into the open-marine environment. This Holocene (deposited since the last ice age) glaciomarine sediment blankets most of the inner continental shelf (fig. 1), reaching a thickness of 200 meters (m) seaward of Icy Bay and Yakutat Bay (Carlson and Molnia, 1975) and 260 m

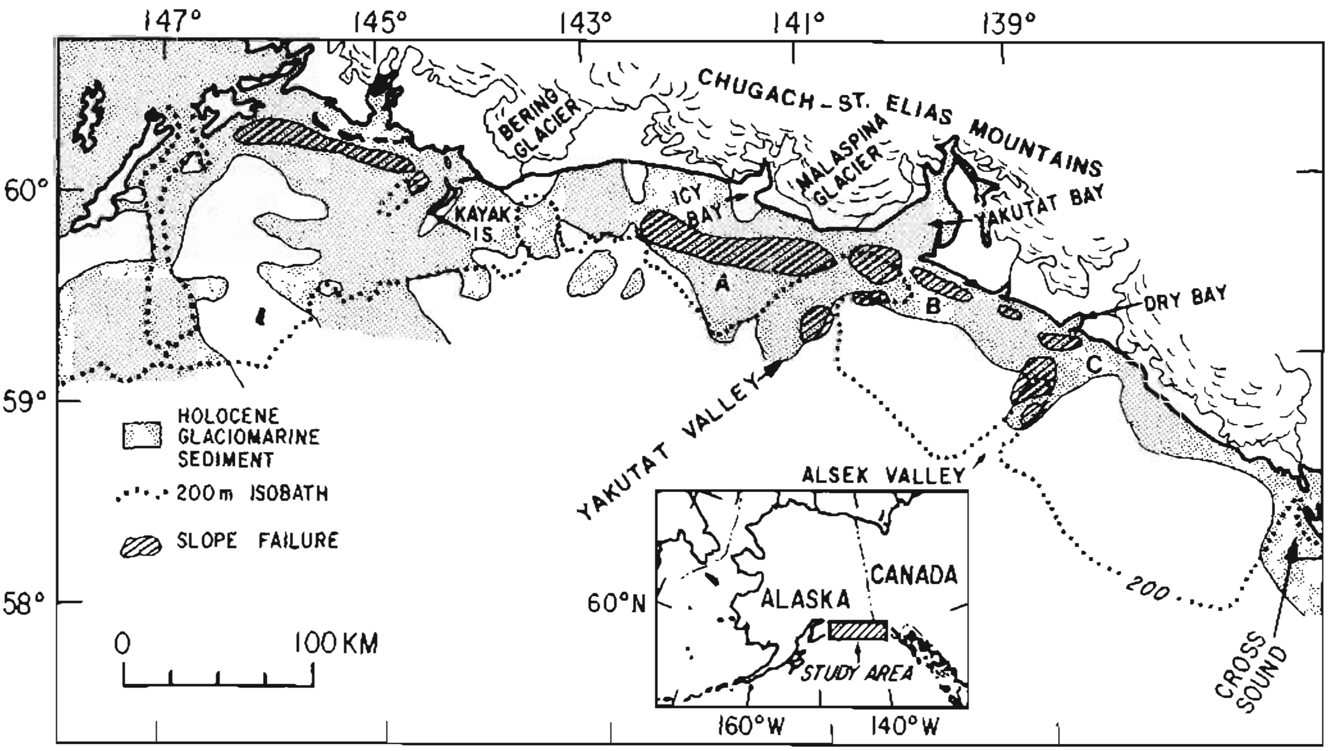

Figure 1. Northeastern Gulf of Alaska depicting the Holocene (deposited since the last ice age) glaciomarine sedimentary deposit and the location of areas of slope failure on the continental shelf, including (A) the Icy Bay-Malaspina slump, (B) the Yakutat slump, and (C) the Alsek prodelta slope instability area (from Schwab and Lee, 1988).

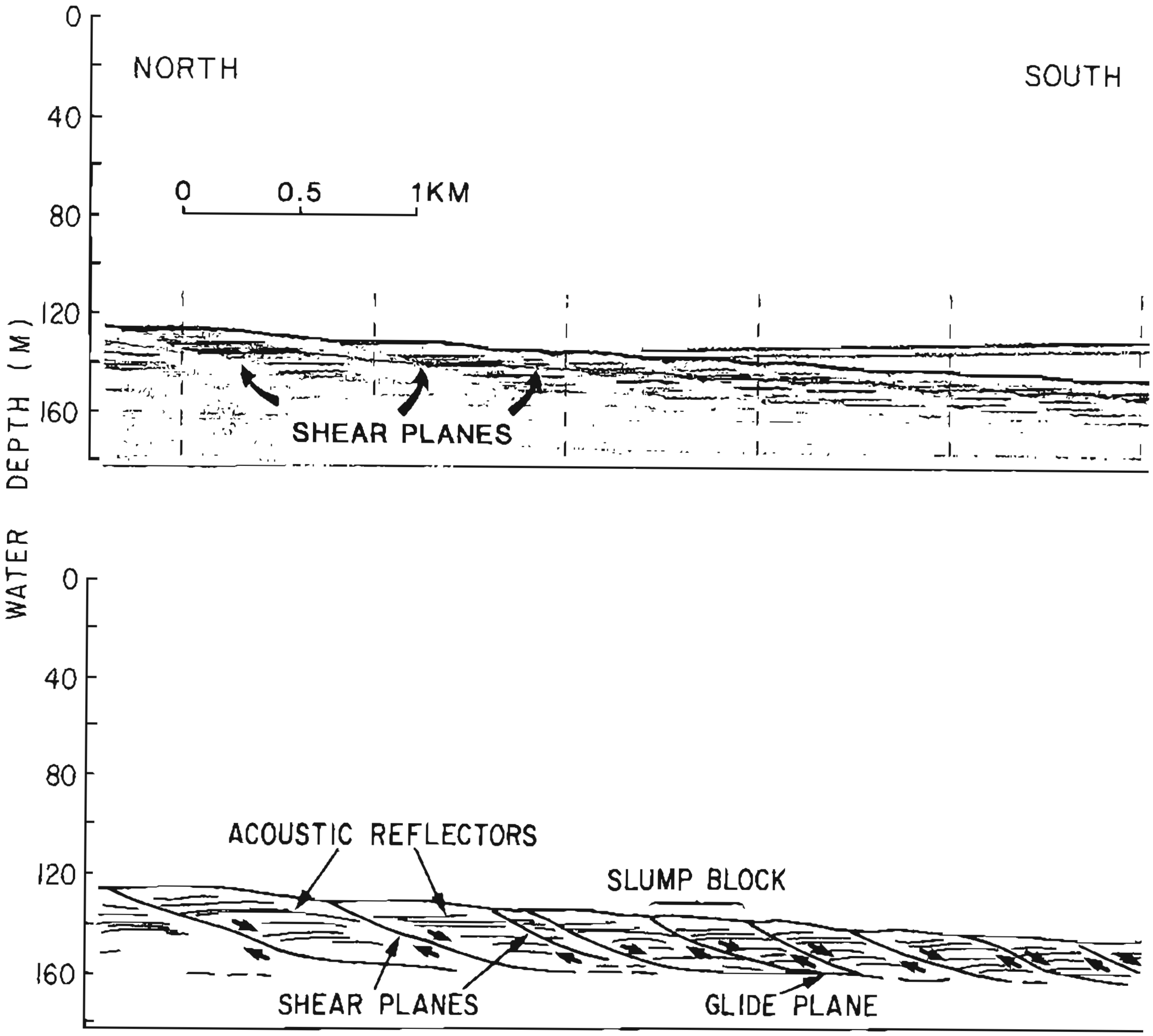

Figure 2. A 3.5-kilohertz seismic-reflection profile and interpretive sketch collected over part of the Icy Bay-Malaspina slump (modified from Lee and Edwards, 1986).

seaward of Dry Bay (Schwab and Lee, 1983). Slope failures on the continental shelf of the northeastern Gulf of Alaska occur entirely within this Holocene glaciomarine sedimentary deposit.

CASE STUDIES

The Icy Bay-Malaspina slump, the Yakutat slump, and mass flows on the Alsek prodelta (fig. 1) typify the two types of slope failures found on the continental shelf of the northeastern Gulf of Alaska. The region offshore from Icy Bay and the Malaspina Glacier is an area where the glaciomarine clayey silt has failed over an area of about 1,080 square kilometers (km^2) in water depths of 70 to 150 m on a sea-floor slope less than 0.5° (Carlson, 1978). Seismic-reflection profiles (for example, fig. 2) collected over the area of slope failure show broken acoustic reflectors in the subsurface (reflectors are indicative of sedimentary bedding) and scarplike surficial forms, suggesting that discrete slump blocks have undergone a slight backward tilting while moving approximately 18 m downslope. These slump blocks extend over the entire failure area and are about 0.5 km long and offset the sea floor from 2 to 5 m.

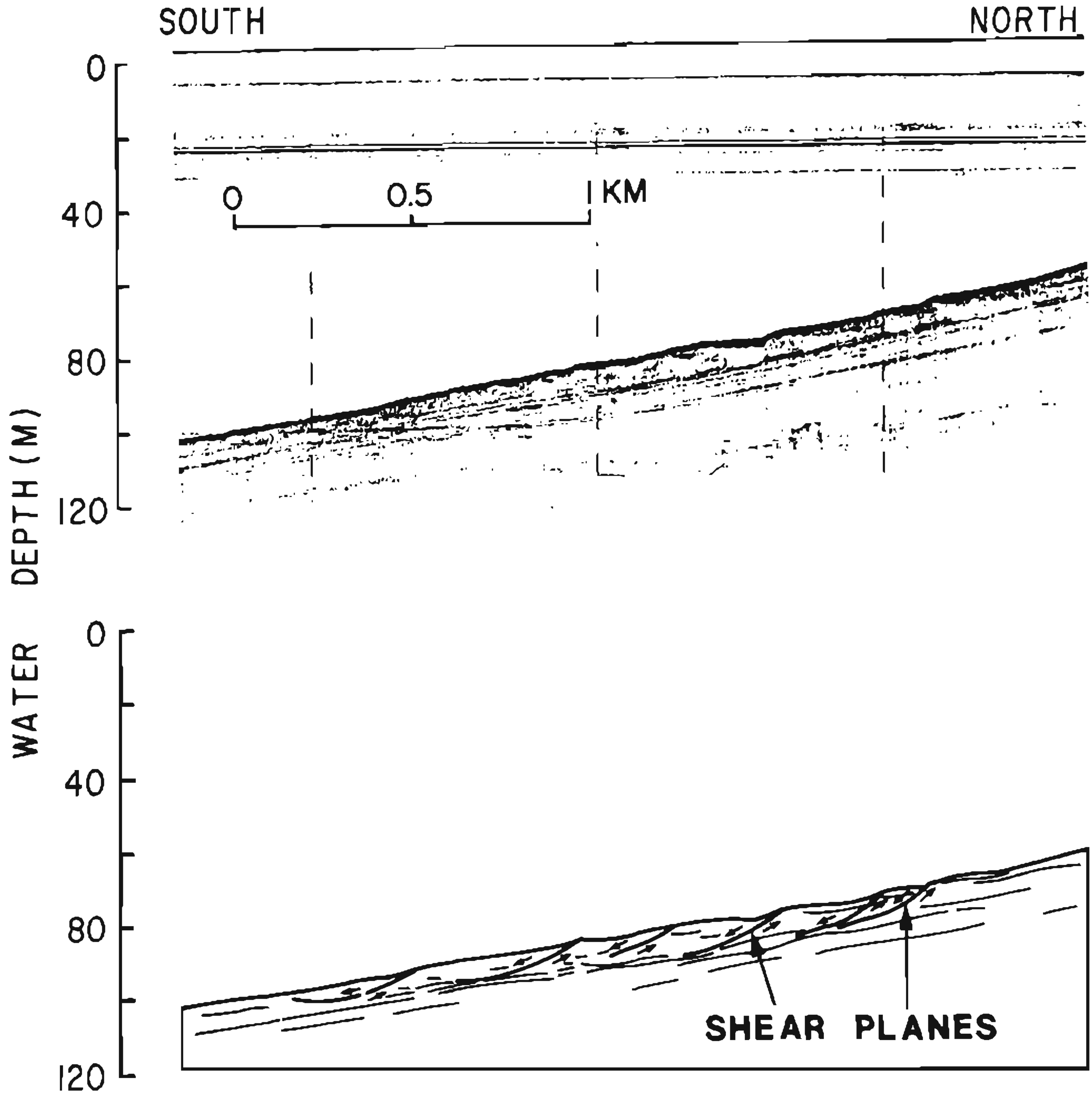

Figure 3. A 3.5-kilohertz seismic-reflection profile and interpretive sketch collected over the Yakutat slump (Schwab and Lee, 1988).

The individual failure planes (shear planes) extend down below the sea floor to a depth of 15 to 40 m, where they form a well-defined glide plane parallel to the sea floor.

The Yakutat slump is characterized by backward-tilted blocks of glaciomarine clayey silt that have undergone about 13 m of downslope movement in water depths of 65 to 90 m (Carlson and others, 1980). Seismic-reflection profiles show that the slump blocks, although not as well developed as in the Icy Bay-Malaspina slump, extend over an area of about 260 km^2 and are about 100 m long and have sea-floor relief of 3 to 4 m (fig. 3). The failure planes extend 10 m below the sea floor. The nearshore part of the Yakutat slump has a slope angle of about 1°. The sea-floor slope decreases to about 0.5° at the seaward edge of the failure area.

Slope failures (fig. 4) cover an area of at least 150 km^2 on the Alsek prodelta (Molnia, 1982). These slope failures originate in a water depth of about 35 m on a

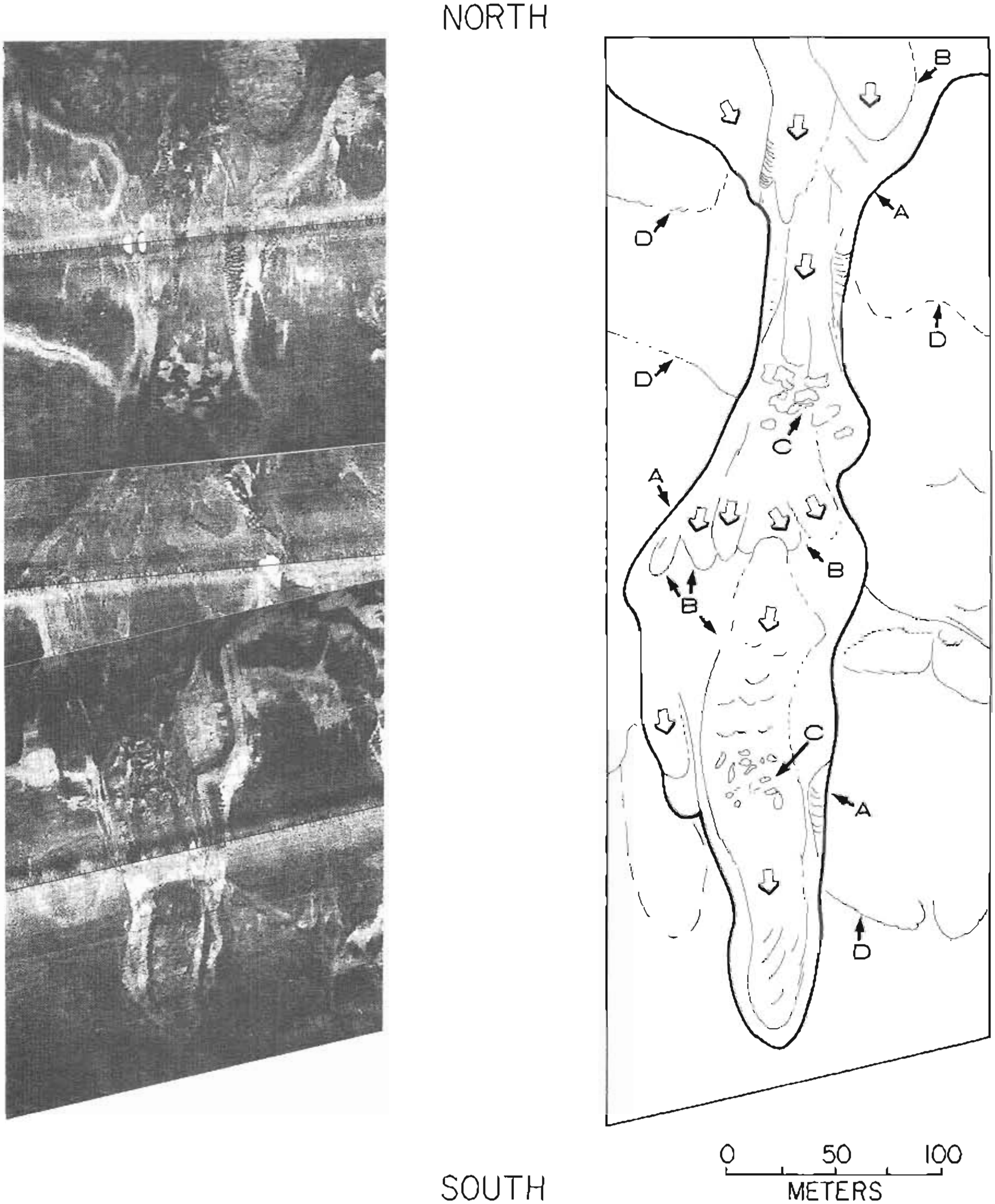

Figure 4. A mosaic of 100-kilohertz sidescan-sonar imagery and interpretive sketch of an area of the Alsek prodelta showing: (A) boundaries of an elongated mass flow, (B) coalescing mass-flow deposits, (C) collapsed, blocky sea floor or "flake-type" slides, and (D) headwall escarpments of slope failures that predate the elongate mass-flow deposit. Highlighted arrows indicate the direction of flow.

sea-floor slope of about 0.5° and extend to a water depth of at least 80 m and possibly as far as the floor of the Alsek Valley (fig. 1) which has a sea-floor slope of 1.3°. Unlike the Icy Bay-Malaspina slump and Yakutat slump, individual slope failures on the Alsek prodelta are not composed of discrete slump blocks but display a morphology on sidescan sonographs indicative of sediment that has flowed (fig. 4). These mass flows show downslope movement ranging from a few meters to about 2 km. The poor quality of the seismic-reflection profiles collected over the Alsek prodelta precludes determination of the maximum depth below the sea floor that is affected by slope failure; however, it is estimated to be less than 20 m (Carlson and others, 1980).

Using engineering analyses of the strength characteristics of the glaciomarine sediment and the environmental stresses affecting this sedimentary deposit, Schwab and Lee (1983) determined that earthquake-shaking and the passage of storm waves are the two triggering forces that caused the slopes to fail. Downslope stress resulting from gravitational forces (that is, sea-floor slope angle) are a less important factor, although they determine the ultimate direction of movement. Some uniformity in the style of mass movement following slope failure on the continental shelf of the northeastern Gulf of Alaska might be expected because the sediment in the failure areas is all of the same type, glaciomarine clayey silt. However, variation in the morphology of the mass movement deposits implies that there are major differences in the styles of mass movement.

ANALYSIS

To determine what caused the sediment found on the northeastern Gulf of Alaska continental shelf to deform in two different ways, we compared the environmental factors affecting the slumps and mass flows (Schwab and Lee, 1988). Our three study areas are suited for this type of comparison because the seismic and storm-wave environments, as well as the sediment, are similar in each. For example:

1. Due to its tectonic setting (Bruns, 1985), the northeastern Gulf of Alaska is the most seismically active region in the United States, apart from the Aleutian Islands. All three study areas are located in a region suggested as having a uniform ground-shaking intensity; that is, each study area will, over a period of time be subjected to earthquake movements having about the same maximum intensity (Stephens and Page, 1982).
2. All three areas are influenced by the same storm-wave climate. Large storm waves, which commonly have a wave height of at least 15 m, roll across the continental shelf throughout the winter (Royer, 1975). The probable maximum height of winter storm waves in the area is 37 m (Quayle and Fulbright, 1975; Bea, 1976).

Table 1. Sediment textural data for the three slope failure study areas

	Sand (>0.063 mm) (%)	Silt (%)	Clay (<0.004 mm) (%)
Icy-Bay-Malaspina slump	1.23±1.94	66.15±9.94	32.60±10.11
Yakutat slump	7.74±6.09	54.32±6.59	37.93±6.20
Alsek prodelta	1.70±2.41	59.26±7.33	39.16±6.83

3. The grain-size distribution of samples collected in the three study areas varies locally but similarly in each area (table 1). The glaciomarine sediment on the continental shelf of the northeastern Gulf of Alaska is dominantly clayey silt.
4. The mineralogy of the sediment from the three study areas is similar (Molnia and Hein, 1982; Schwab and Lee, 1988).
5. Sediment cores in the three study areas that contain a higher proportion of sediment with water content in the range of 35 to 45 percent of the dry sediment weight correlate with the locations of submarine slope failure (Schwab and Lee, 1983). The water content of this sediment is a simple index property that correlates with more advanced engineering properties and also more basic sediment properties such as grain-size distribution.
6. Sediment pore-water pressures in the three study areas do not appear to be in excess of hydrostatic pressure, an important condition that relates to slope stability (Schwab and Lee, 1983; Lee, Schwab, and Booth, this report).

Our slope-stability analysis (Schwab and Lee, 1983) described the influence of storm waves and earthquake forces that lead to slope failure of the glaciomarine sediment. These forces were defined by laboratory testing in which only a relatively small amount of sediment deformation was found to occur. This level of deformation, although small, is sufficient to be seen in high-resolution seismic-reflection profiles collected over the Icy Bay-Malaspina and Yakutat slumps (fig. 2). The slope failures on the Alsek prodelta (fig. 4), however, include elements of flow that extend well beyond the level of deformation that typically is observed in a laboratory-test-based slope stability analysis. Therefore, although our initial slope-stability analysis most likely predicts the initiation of slope failure, it provides little information regarding its consequences.

Two end-member consequences of slope failure have been described: disintegration and nondisintegration (Whitman, 1985; Schwab and others, 1987). *Disintegrative failure* is a condition in which a sediment mass is so weak following slope failure that it deforms continuously under downslope gravitational stresses even after the stress causing the slope failure has been removed (for example, following an earthquake). This deformation ceases only after large displacement has occurred or the slope gradient is reduced. *Nondisintegrative failure* involves permanent displacement during slope failure, but the sedimentary deposit stops deforming after the triggering stress is removed. The Icy Bay-Malaspina slump and the Yakutat slump are nondisintegrative failures whereas the mass flows in the Alsek prodelta are disintegrative failures.

To achieve a disintegrative failure, the sediment must lose a good deal of its strength during a transient loading event (earthquake or major storm). In the case of the continental shelf in the northeastern Gulf of Alaska, the sediment strength must be reduced to the point that continued downslope sediment movement can occur on slopes that are less than 1°. Our laboratory testing of sediment samples from the area of the slope failures show that the strength of the sediment cannot be reduced sufficiently, as long as the natural water content of the sediment remains constant. That is, the sediment in place has a water content that is too low to allow the development of a mass flow. If the water content were increased during a transient loading event, however, the sediment's strength could be lowered enough to allow a mass flow to develop (Schwab and others, 1987; Schwab and Lee, 1988).

How can sediment water content be increased during an earthquake or major storm? One mechanism proposed by Whitman (1985) involves local pore-water movement within the sediment column; that is, pore water is redistributed within the sediment but does not flow out of the overall sediment system. A unique characteristic of the response of fine-grained sediment to cyclic stresses (the stresses induced by storm waves or earthquakes) is the tendency to develop sediment pore-water pressures that equal lithospheric stress (the stress induced by the overlying sediment which, in turn, increases with increasing depth in the deposit). If the level of cyclic stress is high enough, pore-water pressures that are equal to the lithospheric stress develop over a substantial thickness of the sedimentary deposit. The resulting pore-pressure gradient leads to an upward flow of pore water; from the area of high pressure to the area of relatively lower pressure (fig. 5). The base of the part of the sediment column affected by the cyclic stresses becomes more dense while the upper sections loosen, resulting in a continually decreasing strength of the upper sedimentary section. Once the sediment strength falls below a critical level, a disintegrative failure occurs.

Another mechanism for converting a nondisintegrative failure into a disintegrative failure was described by Hampton (1972). He suggested that water is incorporated from the overlying sea water into a submarine slump by the jostling and agitation it received as it begins to move. Sediment deformation during a transient loading event can become large. Some of the sea water may be incorporated into the sediment simply as a result of these rapid movements and agitations, thus significantly reducing the strength of the sediment.

On the continental shelf in the northeastern Gulf of Alaska, we suggest that these two mechanisms are most effective during storm-wave loading and that storms are the principal cause of mass flows in this glaciomarine sediment. Sediment framework expansion (and strength reduction) during a major storm event, enhanced relative to what would occur in an earthquake, can be justified in three ways.

First, a long series of loading cycles causes the sediment strength to fall to a low level. Our experimental results on the behavior of Gulf of Alaska glaciomarine clayey silt suggest that the strength of sediment samples is about 73 percent lower after 1,000 stress cycles are applied than after 10 (Schwab and Lee, 1983, 1988). Major earthquakes produce from 10 to 25 representative loading cycles (Seed and Peacock, 1971), whereas a major storm would likely produce 1,000 cycles or more. With such large reductions of sediment strength during major storms, the level of deformation experienced during each cycle of storm-wave loading should be greatly enhanced. The mechanism proposed by Hampton (1972) could then come into play; the increased level of jostling would cause increased amounts of sea water to be incorporated into the sediment.

Second, the long duration of stress application during a major storm allows increased pore-water redistribution. Thus, by Whitman's (1985) mechanism, water begins to flow from deeper strata into shallower strata (fig. 5). If the duration of cyclic stress application is short, the process is limited, and pore-water pressure drops in the deeper strata, thus, halting the upward flow. With a long period of cyclic stress application, however, the pore-water pressures in the deeper strata are renewed continually. A steady pumping action, drawing water from deeper sediment toward steadily expanding sediment near the sea floor occurs.

Third, storm waves generate stresses that propagate from the sea floor down, whereas earthquakes generate stresses that propagate from the bedrock up. Because weakening and failure of sediment occurs after application of a number of cycles of stress, the ability of the failed sediment to transmit stresses is degraded. In the case of an earthquake, the slump blocks, once they have formed, become partially isolated from the source of the stress application below. Significant stresses cannot propagate upward across the failure plane (Lee, 1976). On the other hand, with storm-wave-induced failures, it is the zone under the failure blocks that become partly isolated from the source of stress. The blocks themselves are still openly exposed to wave-induced stresses propagating down from the sea floor. The processes leading to framework expansion are free to proceed, unhindered by the development of a barrier to stress propagation.

The suggestion that storm waves are more likely to cause mass flows than are earthquakes in this environment is also supported by field and laboratory evidence. We determined the water depths in which major earthquakes are most likely to cause a slope failure and those in which peak storm waves are the most critical factor (Schwab and Lee, 1983). We found that earthquake shaking dominates in water depths greater than 76 m and storm-wave loading dominates in water depths less than 35 m. Sediment in depths between 35 and 76 m have an equal chance of failure by either mechanism. Using these criteria, we suggested that the Icy Bay-Malaspina slump, in water depths between 70 and 150 m, was caused by earthquake shaking. The Yakutat slump, with water depths between 65 and 90 m, was also probably seismically induced. However, the Alsek

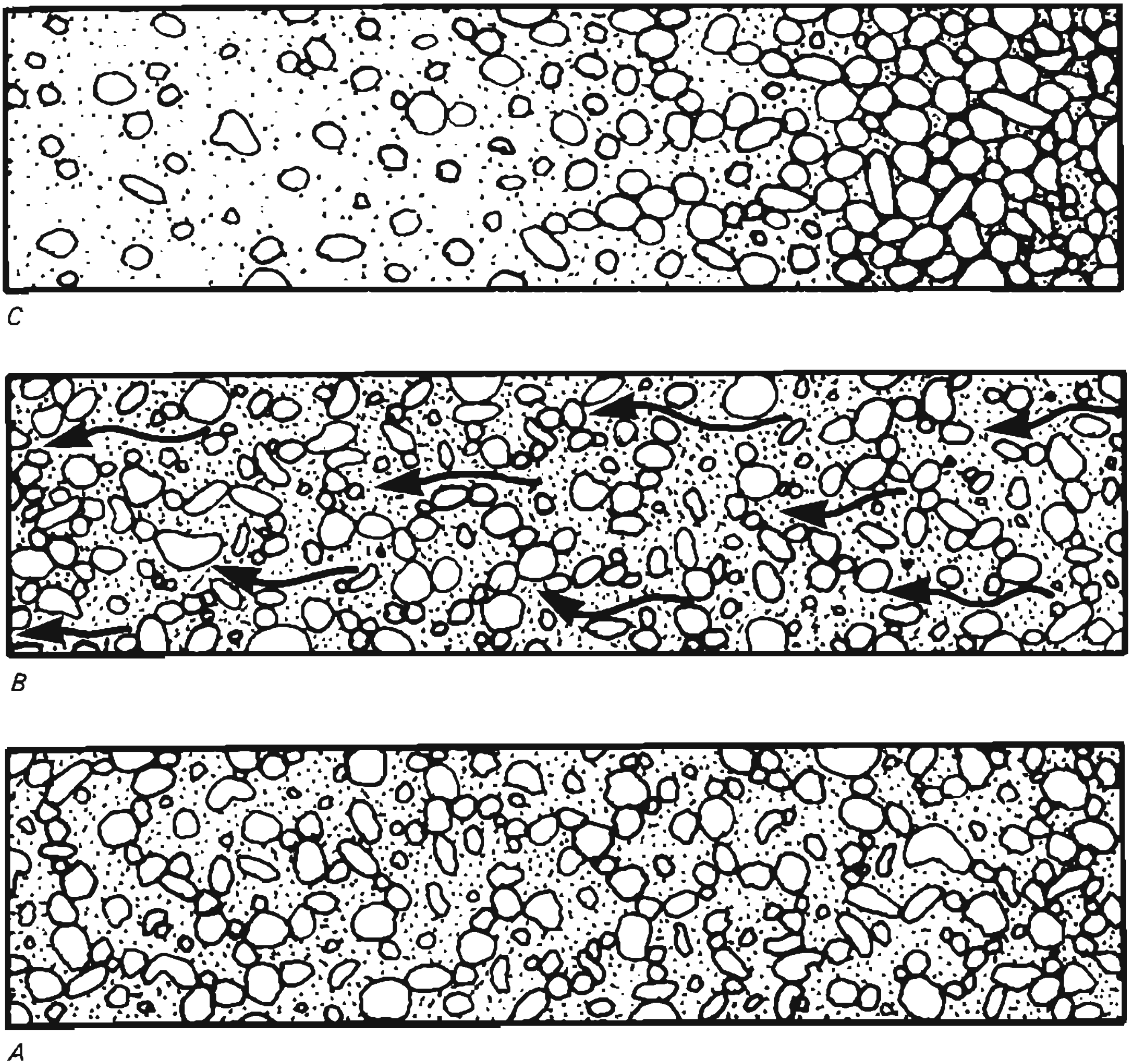

Figure 5. Illustration of the possible redistribution of pore water during the application of cyclic stresses: (*A*) the sediment column at rest, (*B*) pore water flowing upward due to the pore-water pressure gradient induced by the application of cyclic stresses, and (*C*) the base of the sediment column affected by the cyclic stresses contracts due to the redistribution of pore water while the upper section expands.

prodelta failures, in water depths of 35 to 80 m, occurred in a depth range where storm waves could cause slope failure as readily as earthquakes.

Disintegrative failures on the Alsek prodelta are found within an environment in which either winter storm waves or earthquakes can cause failure. On the other hand, nondisintegrative failures in the Icy Bay-Malaspina slump and Yakutat slump are associated with environments in which storm waves are probably insufficient to cause failure, but in which earthquake shaking is clearly capable of producing the slumps. We suggest that storm waves, which represent the application of cyclic stresses for a long duration, were responsible for generating disintegrative failures on the northeastern Gulf of Alaska continental shelf, and that earthquake shaking, which is characterized by the application of cyclic stresses for a relatively short duration, led to nondisintegrative failures in this environment.

CONCLUSIONS

Three differences—(1) enhanced jostling, (2) enhanced redistribution of pore water, and (3) the absence of a partial barrier to stress propagation—separate the application of cyclic stresses related to storm waves from those related to earthquakes. These factors are sufficient to allow disintegrative failures to form during storms. On the other hand, only nondisintegrative failures are likely to occur during earthquakes on the northeastern Gulf of Alaska continental shelf. These differences explain the varied types of slope failure found in the region.

These findings regarding the cause of mass flows are not universally applicable to other sedimentary deposits. That is, we cannot state that slumps are caused by earthquakes and disintegrative failures by pore-water redistribution during major storms. For example, earthquake-induced mass flows may be common wherever relatively loose (low-density) sedimentary deposits are found. Recent extensive liquefaction on a 0.25° slope following an earthquake off northern California (Field, this report) illustrates such a situation. A mudflow on a 4° slope in Santa Barbara Basin, Calif., is another illustration (Edwards, Lee, and Field, this report). Likewise, disintegrative failures in fine-grained deltaic deposits can probably occur without a significant redistribution of pore water during major storm-wave-loading events. Disintegrative failures on very gentle slopes (0.01° to 0.45°) of the Mississippi Delta (Coleman and others, this report) illustrate the behavior of sediment that flowed without needing to expand. In each new situation, therefore, a full evaluation of the factors relating to slope failure is needed. For the case of the northeastern Gulf of Alaska however, we have shown that the unique loading environment, sediment type, and sediment conditions lead to limited possibilities in terms of resulting failure types.

ACKNOWLEDGMENTS

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