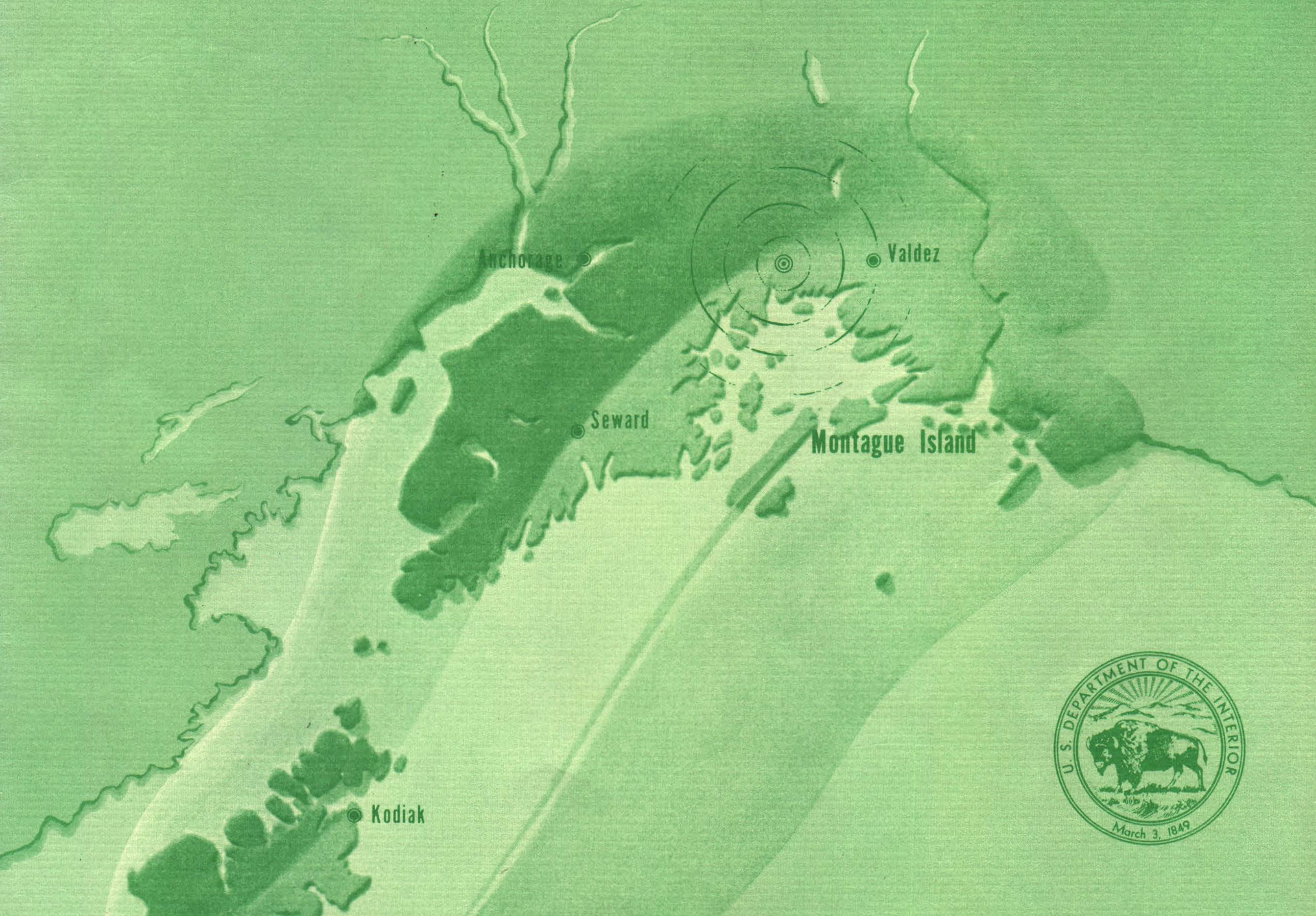
The Alaska Earthquake
March 27, 1964

Regional Effects



Erosion and Deposition on a Raised Beach, Montague Island

GEOLOGICAL SURVEY PROFESSIONAL PAPER 543-H

THE ALASKA EARTHQUAKE, MARCH 27, 1964:
REGIONAL EFFECTS

Erosion and Deposition on a Beach Raised by The 1964 Earthquake Montague Island, Alaska

By M. J. KIRKBY and ANNE V. KIRKBY

A quantitative study of geomorphic modifications of uplifted coastal features

GEOLOGICAL SURVEY PROFESSIONAL PAPER 543-H

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THE ALASKA EARTHQUAKE SERIES

The U.S. Geological Survey is publishing the results of investigations of the Alaska earthquake of March 27, 1964, in a series of six professional papers. Professional Paper 543 describes the regional effects of the earthquake. Other professional papers describe the history of the field investigations and reconstruction effort; the effects of the earthquake on communities; the effects on the hydrologic regimen; and the effects on transportation, communications, and utilities.

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EROSION AND DEPOSITION ON A BEACH RAISED BY THE 1964 EARTHQUAKE, MONTAGUE ISLAND, ALASKA

By M. J. Kirkby [1] and Anne V. Kirkby [2]

ABSTRACT

During the 1964 Alaska earthquake, tectonic deformation uplifted the southern end of Montague Island as much as 33 feet or more. The uplifted shoreline is rapidly being modified by subaerial and marine processes. The new raised beach is formed in bedrock, sand, gravel, and deltaic bay-head deposits, and the effect of each erosional process was measured in each material. Fieldwork was concentrated in two areas—MacLeod Harbor on the northwest side and Patton Bay on the southeast side of Montague Island. In the unconsolidated deltaic deposits of MacLeod Harbor, 97 percent of the erosion up to June 1965, 15 months after the earthquake, was fluvial, 2.2 percent was by rainwash, and only 0.8 percent was marine; 52 percent of the total available raised-beach material had already been removed. The volume removed by stream erosion was proportional to low-flow discharge raised to the power of 0.75 to 0.95, and this volume increased as the bed material became finer. Stream response to the relative fall in base level was very rapid, most of the downcutting in unconsolidated materials occurring within 48 hours of the uplift for streams with low flows greater than 10 cubic feet per second. Since then, erosion by these streams has been predominantly lateral. Streams with lower discharges, in unconsolidated materials, still had knickpoints after 15 months. No response to uplift could be detected in stream courses above the former pre-earthquake sea level.

Where the raised beach is in bedrock, it is being destroyed principally by marine action but at such a low rate that no appreciable erosion of bedrock was found 15 months after the earthquake. A dated rock platform raised earlier has eroded at a mean rate of 0.49 foot per year. In this area the factor limiting the rate of erosion was rock resistance rather than the transporting capacity of the waves.

The break in slope between the top of the raised beach and the former seacliff is being obliterated by debris which is accumulating at the base of the cliffs and which is no longer being removed by the sea. Current cliff retreat by rockfall, mudflows, and landslides was estimated at 0.7 to 2.0 feet per year, and in parts of Patton Bay the accumulation of debris has obliterated 78 percent of the original break in slope in 15 months.

Evidence of two relative sea-level changes before 1964 was found in Patton Bay. At a high stand of sea level lasting until about 2000 B.P. (before present), an older raised beach was formed which, over a distance of 5 miles, shows 40 feet of deformation relative to the present sea level. Peat deposits exposed by the 1964 uplift also record a low sea level that lasted until at least 600 B.P.

The 1964 raised beach was used to test the accuracy of identification of former sea-level elevations from raised beach features. The pre-1964 sea level could be accurately determined from the height of the former barnacle line, so an independent check on high-water level was available. The most reliable topographic indicator was the elevation of the break in slope at the top of a beach between a bedrock platform and a cliff. Even here, the former sea level could only be identified within 5 feet. The breaks in slope at the top of gravel beaches were found to be poor indicators of former sea level.

On Montague Island, evidence of former high sea levels appeared to be best preserved (1) as raised bedrock platforms on rocks of moderate resistance in slightly sheltered locations and (2) as raised storm beaches where the relief immediately inland was very low.

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INTRODUCTION

On March 27, 1964, an earthquake shook parts of Alaska and caused great devastation in inhabited areas. However, maximum uplift took place, and large bedrock faults were active, on Montague Island, an uninhabited island off the southern coast of Alaska.

The southern end of Montague Island was uplifted as much as 33 feet or more during the earthquake (fig. 1), and two active faults several miles long caused many landslides and broke or displaced many hundreds of trees (Plafker, 1967). Around the coast, marine deposits and a marine-abraded bedrock platform were lifted above sea level and exposed to subaerial processes.

The uplift associated with the earthquake has provided physiographic conditions on a scale much larger than could be simulated in a laboratory and with the added advantage that the features and changes were natural. The bay-head deposits of MacLeod Harbor (figs. 1, 2) on the northwest coast of the island provided an undissected surface, about 1 square mile in area and with a slope of only 20′, which had been suddenly uplifted 33 feet. On this surface, consequent drainage channels were initiated and were rapidly eroded vertically and laterally; thus, a unique opportunity was provided to study the effects of rapid uplift on the mode and rate of fluvial processes and the resultant valley forms and long profiles.

The earthquake also brought to view a bedrock platform and associated deposits, which allowed observation and measurement of features that are usually under water and therefore difficult to study.

On both sides of the island the amount of subaerial modification of the 1964 raised beach could be measured. Especially important was the opportunity to study the area only 15 months after the earthquake. In the MacLeod Harbor area, rivers in soft sediments have adjusted so swiftly that evidence of the postearthquake history of the area will soon be removed by erosion. Subaerial degradation of the raised beaches and abandoned seacliffs is so rapid that within a few years little of the original marine form will be left unchanged, and the growth of vegetation, which has already begun on the beach, will hasten its obliteration.

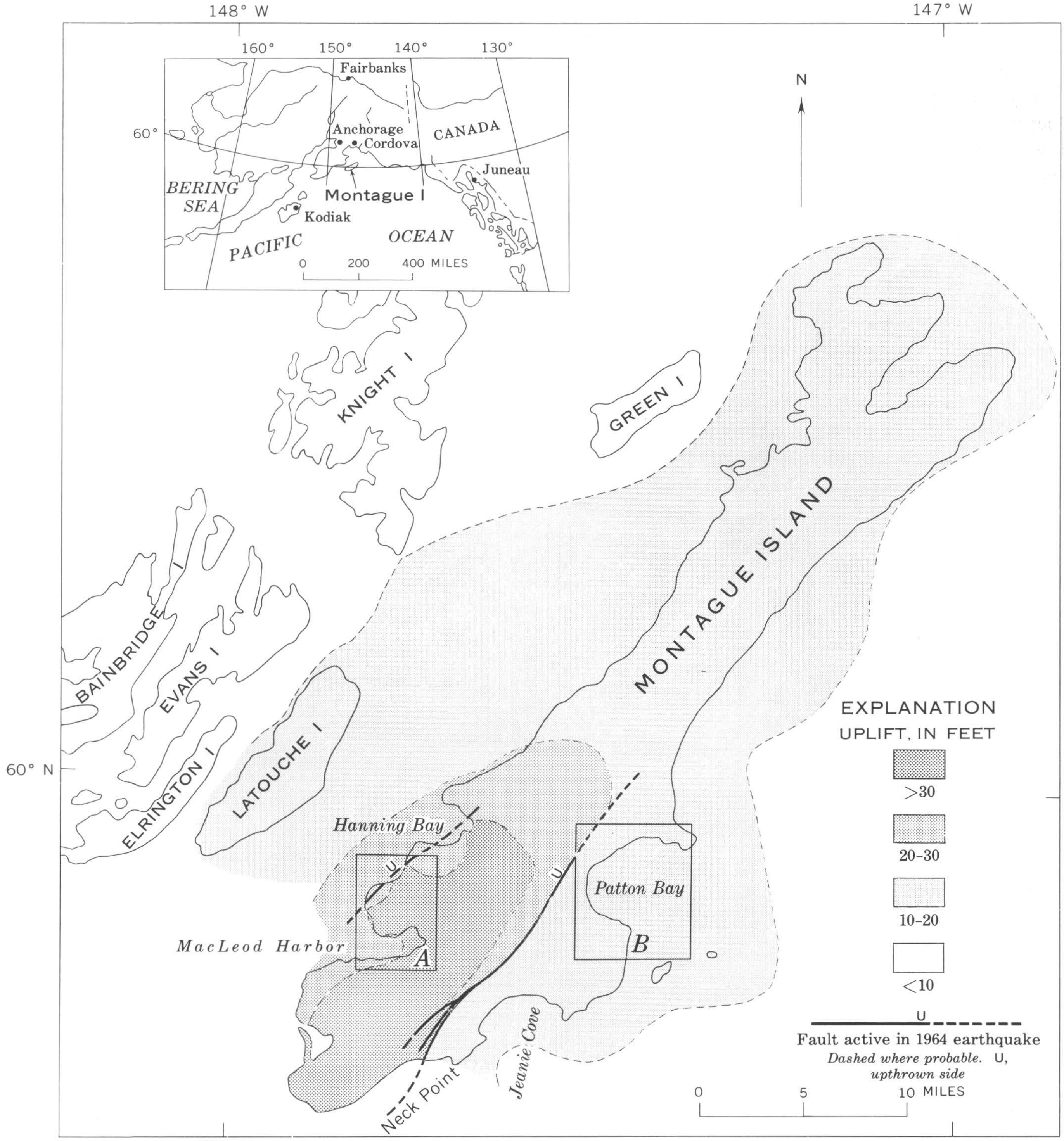
Fieldwork was done in 1965 at two locations, MacLeod Harbor and Patton Bay (fig. 1). The project was a reconnaissance and, although the study was concerned with postuplift changes in raised beaches, differences in emphasis result from physiographic and tectonic differences between the two bays. The work was supported by a grant from the National Science Foundation, administered by The Johns Hopkins University. We would like to thank Professor M. G. Wolman of The Isaiah Bowman Department of Geography, The Johns Hopkins University, for his continuous scientific and administrative help; and A. T. Hohl, Department of Geology, Princeton University, and Thomas Dunne, Department of Geography, The Johns Hopkins University, for their assistance in carrying out the fieldwork. Mr. C. LaBounty, a longtime seasonal resident of Montague Island, provided valuable data on pre- and postearthquake conditions in the bay-head deposits.

PHYSICAL SETTING

Montague Island, one of the outermost islands of Prince William Sound, lies off the southern coast of Alaska at lat 147.5° W. and long 60° N. It is about 50 miles long, and has a maximum width of 15 miles and a minimum width of less than 5 miles. The island is rugged and mountainous, and a chain of peaks forms its backbone at an average altitude of about 2,500 feet. Cliffs which occur almost all around the island make the coast dangerous. The 1964 earthquake elevated and exposed a flat bedrock platform which encircles the island; the former marine cliffs, now beyond the reach of the sea, were broken by a few inlets and sandy beaches that allowed access to the island.

Travel across Montague Island is difficult. It is uninhabited except for seasonal hunters and loggers; there are neither roads nor even paths, except those made by bears. The upper parts of the mountains are rocky and are snow-covered all year. Below the peaks the land is either steep and covered with alders or stunted conifers, or the slope becomes so gentle that waterlogging prevents tree growth, and patches of wet peat interspersed with pools of water form muskeg.

At altitudes below about 1,000 feet, the coniferous forest forms a dense cover down to sea level. Only where the slope is gentle enough for muskeg to form or where it is so steep that it is totally unstable is there any break in the forest. Most of the island has not been logged and is covered with apparently virgin forest containing conifers more than 6 feet in



1.—Index map of Montague Island showing amount of uplift (after Plafker and Mayo, 1965). *A*, area of figure 12; *B*, area of figure 17.

diameter and over 400 years old. The trees are as much as 200 feet high, and there is a thick, usually thorny, undergrowth. The forest floor is covered with shallow ponds and bogs; large fallen tree trunks, many of them rotten, lie across one another in a chaotic pattern and provide unstable footing.

Even on the postearthquake raised beaches, plant growth is rapidly establishing itself; only 15 months after the earthquake, conifer seedlings were growing on surfaces that were formerly below sea level (G D. Hanna, 1965, oral commun.).

Most of the rivers on Montague Island are narrow, steep torrents that rush down the slopes in a series of rapids and waterfalls. Only the Nellie Martin River has achieved any substantial floodplain development, and it flows through a wide valley, whose floor near the coast is covered with thick peat deposits that form vertical riverbanks as much as 10 feet high.

METHODS

Mapping was done with a planetable and telescopic alidade. Beach and stream profiles were surveyed with a hand level or Abney clinometer, and distances were measured by pacing. Preearthquake mean high water was assumed to be at the top of the abandoned barnacle line. (Plafker, 1965, p. 1–2). The level of postearthquake mean high water was obtained from the U.S. Coast and Geodetic Survey predicted tide tables. These agreed, within 1 foot, with the position of the new barnacle line where it was observed.

Grain sizes of the MacLeod Harbor bay-head deposits were determined by counts (Wolman, 1954) of 100 pebbles in stream beds and by sieve analyses of fine materials. On beach profiles we measured the size of the largest stones in a section and estimated visually the percentage of material finer than 1 inch in diameter. Permeabilities were determined approximately by measuring the time for 1 inch of water to infiltrate; the water was contained by a metal ring.

Discharge of streams in McLeod Harbor was calculated as

Surface velocity in center of stream × mean depth × width.

This calculation gives values approximately twice the true value, because the centerline surface velocity is approximately twice the mean velocity of the whole cross section. Discharges were measured at a time of relatively low flow (see table 1, p. H10) but were corrected for daily variations by multiplying by a standardizing factor:

$$\frac{\text{Discharge of stream 1 on June 20, 1965}}{\text{Discharge of stream 1 on day of measurement}}$$

To avoid confusion, the expression "1964 sea level" has not been used in this paper. Instead, the expressions "preearthquake sea level" and "postearthquake sea level" are used. A sea level higher than the post-earthquake sea level is called a raised sea level, and a sea level lower than the preearthquake sea level is called a low sea level. However, inasmuch as the earthquake took place in 1964, it is correct, and gives rise to no ambiguity, to refer to the uplifted beach exposed by the earthquake as the 1964 raised beach. The changes in relative sea level are probably caused more by movement of the land than by absolute change of sea level, but in the absence of positive evidence it is customary (Sparks, 1960, p. 211) to refer to the land as fixed in relation to sea-level changes.

GEOLOGY

BEDROCK

Little is known about the bedrock of Montague Island; most exposures are covered by forest and muskeg or, at higher elevations, by scree and snow. The recent uplift has improved accessibility to the coast and has brought nearly continuous exposures of bedrock into view along the shore. These strata are thought to be of Tertiary age (Plafker and MacNeil, 1966, p. 66–67).

A study of the coastal exposures indicates that the bedrock of the island consists of a sequence of interbedded thin layers of black argillite, thick beds of graywackelike sandstone, and occasional beds of conglomerate and volcanic rocks. The sedimentary rocks are uniformly bedded and have flat parallel bedding planes. The coarser clastic rocks are of two types: (1) conglomerate having coarse rounded quartz and rock-fragment clasts set in a gray sandstone matrix and (2) breccia of black angular argillite chips, also in a sandstone matrix. The volcanic rocks were not seen in an outcrop, but hillside boulders and beach cobbles indicate that they consist of buff and green medium- to fine-grained agglomerate. In the Patton Bay region, light-gray volcanic interbeds occur.

Two major faults, both active in the 1964 earthquake, have been mapped on the island (Plafker, 1967, fig. 2). The faults trend roughly parallel to the island axis and have dips of 50°–85° NW. The Hanning Bay fault is well exposed for about 4 miles between Hanning Bay and MacLeod Harbor; it has a maximum vertical displacement of 13 feet in bedrock and 17 feet in unconsolidated beach material (Plafker, 1967). Inland, the fault can be traced by landslides, fallen trees, and earth fissures several feet deep and 1 foot wide.

The Patton Bay fault can be traced for about 22 miles, but for most of that distance it is concealed beneath landslide debris, its maximum measured vertical displacement is 20 to 23 feet (Plafker, 1967). This fault follows and accentuates a preexisting topographic break in slope. Near the Nellie Martin River, the Patton Bay faultline is characterized by a linear scarp covered with debris, including fallen trees, and modified by numerous landslides. The faultline scarp varies in height from 100 to 300 feet and has an average gradient of 40° made up of a series of irregular steps. At the top of the debris slope a clean exposure of rock 5 to 20 feet high slopes at an angle of 65° to 75°. There is no evidence of mineralization and no gouge.

Surficial deposits comprise recently uplifted shallow-water marine silt and sand, coarse beach sand and gravel, and fluvial gravel. Extensive peat and some glacial deposits are also present.

RECENT UNCONSOLIDATED DEPOSITS IN MacLEOD HARBOR

At the head of MacLeod Harbor, uplift has exposed unconsolidated sediments which formerly occupied beach, intertidal, and subtidal zones. Three main sediment types are recognized: sand, silt, and gravel. The areal distribution of these types is shown in figure 2. The deposits occupy a roughly rectangular basin that deepens seaward; they themselves also thicken seaward, reaching a maximum exposed thickness of 18 feet, but their total thickness is probably much greater.

The sand occurs mainly on the south side of the bay and has a gradational contact with the silt. The silt occupies an area which was formerly occupied by quiet water in the deepest part of the bay head. The upper surfaces of the silt and sand were formerly partly intertidal and partly below spring low water level.

SAND

The sand of MacLeod Harbor is dark gray and uniform throughout the deposit. It stands in vertical cliffs as much as 20 feet high; for the most part it is well bedded, the thickness of individual beds ranging from 6 inches to 2 feet. The beds are continuous for a minimum of 200 feet and show the internal crossbedding that is characteristic of a mainly fluvial sediment.

Scour channels and filled channels about 4 feet wide are common. Convolute bedding occurs at all levels in the sand, and some convolutions are truncated by overlying uncontorted beds as much as 12 feet below the present surface. The cause of the convolutions is not known, but the possibility that they were formed by thixotropic transformation produced by earthquakes prior to the 1964 earthquake best fits the known facts.

Occasional well-rounded pebbles as much as 3 inches in diameter are found within the sand. Throughout the deposit there are discontinuous stringers up to 2 inches thick of small wood pieces, shells, seaweed, and pine needles, similar to present high-tide swash accumulations. These stringers probably indicate former positions of high tide. The permeability of the unsaturated sand is about 25 inches per hour.

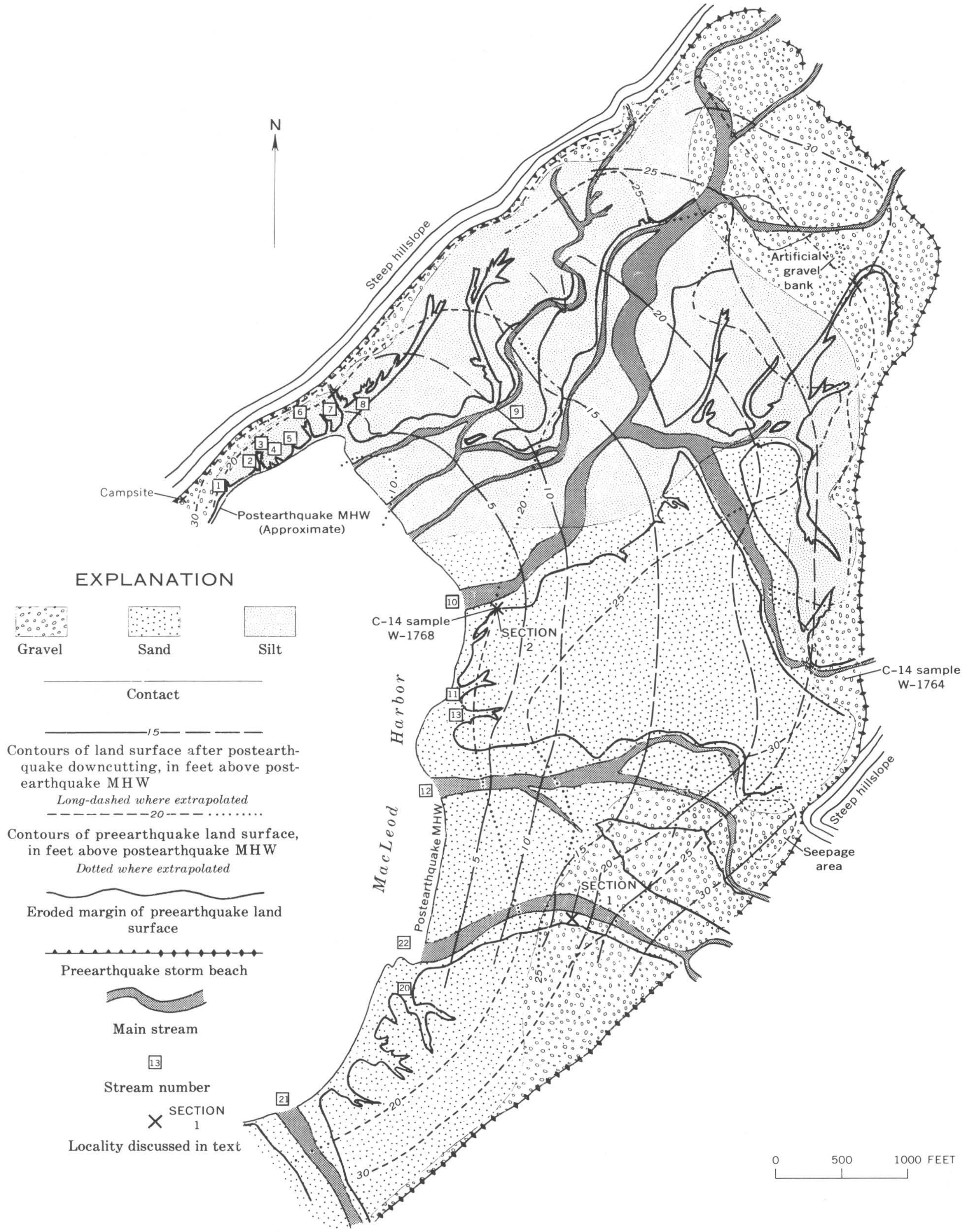
SILT

The silt is black and uniform in composition, except for the percentage of sand present, which increases toward the contact with the sand. The silt shows an apparent lack of bedding and internal structure, although occasionally thin parallel bedding was suggested by differential erosion along streambeds. Clam shells (mostly *Macona nasuta*) that are found throughout the deposit are concentrated through erosion on the silt surface as a lag deposit. The permeability of the silt is about 0.01 inch per hour.

GRAVEL

Gravel deposits occur in three forms—as beaches around the inland edge of MacLeod Harbor, as fans around the former mouths of streams entering the bay, and as lenses and discontinuous sheets at or near present sea level.

The beach gravel consists mainly of local rock types and occasional exotic pebbles of granodiorite, granite, or hornblende gneiss. The gravel is very well rounded and well sorted and has only minor amounts of sand matrix. The mean size of the pebbles varies widely on the beaches around the bay; on some beaches more than 80 percent



2.—Bay-head deposits in MacLeod Harbor and June 1965 elevations of their upper (preearthquake) and lower (postearthquake) surfaces.

are smaller than an inch in diameter, but on other beaches 80 percent are larger than 4 inches in diameter. On any individual beach the size range is much less, although generally pebble size decreases from the storm beach toward the sea, and sorting becomes better in the same direction. Permeabilities are about 450 inches per hour.

Behind the gravel beach around most of the bay is a storm beach which is characterized by little or no sand matrix, poorer sorting, coarser grain sizes, steeper gradients on the upper surface both landward and seaward, and extremely high permeabilities.

The river fan gravel deposits also consist of local rock types and vary in mechanical composition from their apexes in gaps through the storm beach to their seaward edges. At their apexes they are similar in composition to the beach gravel. The percentage of sand matrix increases and the size of the pebbles decreases toward the seaward edges of the fans. Near the lower part of the fans (section 1, fig. 2) the deposit is of interbedded gravel and sand, the gravel beds ranging in thickness from one pebble (½ inch) to 12 inches. Individual pebbles vary from ⅛ to 5 inches in diameter; many are disk shaped and most are well rounded but have low sphericity. The gray-sand matrix makes up 40 percent of an individual gravel bed; wood fragments and abraded logs are common. No graded bedding or cross bedding was observed. The upper and lower contacts with adjacent sand beds are uneven, and the gravel lenses out, the thinnest beds having the least lateral continuity. The sand beds are discontinuous and from 1 to 10 inches thick; planar crossbeds are present in a few beds. The total section at this location (fig. 2) consists of 60 to 70 percent gravel beds and 30 to 40 percent sand beds, not counting sand matrix in the gravel beds.

The gravel lenses and discontinuous sheets occur only at or below 1965 mean sea level. They are found beneath the sand and silt deposits and have been partly eroded and redistributed by postearthquake stream action. They form a more or less continuous sheet where the streams have been eroded down to 1965 sea level. Near the present high-water level where the streams flow into the sea, gravel lenses occur beneath the streambeds. These lenses are as much as 3 feet thick, and in places a sand bed lies between them and the gravel sheet on the present surface. The lower gravel lenses have a sand matrix and a maximum pebble size smaller than that of the present stream gravel.

BEACH DEPOSITS IN PATTON BAY

Bay-head deposits are not present in Patton Bay. Instead, uplift has merely widened the existing beach deposits or bedrock platforms. The distribution of the coastal forms is discussed on page H20 in relation to individual beach profiles.

The coastline around Patton Bay consists mainly either (1) of wide gently sloping sand beaches with gravel storm beaches surmounted by large driftwood accumulations at their inland edges or (2) of cliff-backed bedrock platforms with or without narrow sand or gravel beaches and driftwood at their inward margins. Between zones of these two main coastal types, small areas of large boulders lie on bedrock.

EROSION AND DEPOSITION OF BAY-HEAD DEPOSITS IN MACLEOD HARBOR

FORMATION OF BAY-HEAD DEPOSITS

Both the surface composition and the topography of the preearthquake deposits at the head of MacLeod Harbor (fig. 2) seem consistent with conditions of normal intertidal and shallow-water deposition by rivers at the head of an inlet. At their seaward limit the deposits seem to have formed a steeper slope. The evidence for this statement, although indirect, is fairly conclusive: (1) the U.S. Geological Survey 1:63,360 topographic map and the U.S. Coast and Geodetic Survey charts show depths close inshore that indicate a steep offshore slope, and (2) before the earthquake local fishermen were able to anchor close inshore in areas that are still close to the seaward edge of the silt. A steeper slope, therefore, very probably existed at the seaward edge of the deposits, although whether its form was erosional or depositional is unknown. It is also considered probable that this slope was partly exposed by the uplift associated with the earthquake, because major streams very rapidly trenched the newly uplifted deposits (C. LaBounty, oral commun., 1965); such trenching would be favored by the exposure of a steep slope by uplift above sea level.

Figure 2 shows elevations of the bay-head deposits in June 1965. Surveyed contours of the remaining parts of the upper surface (in interfluve areas) have been interpolated in stream-eroded areas to reconstruct the preearthquake topography of the deposits. Surveyed contours of areas of postearthquake stream erosion have similarly been interpolated to estimate the final postearthquake topography of the bay-head deposits. The upper surface could still be fairly well defined in June 1965, but the former seaward limit of the deposits is less definite. Four independent levels from interpolated 1965 mean high water to the preearthquake barnacle line give a mean measurement of 33 feet for the uplift. The mean tidal range in MacLeod Harbor is 9.2 feet, and spring low tides were formerly 19 feet above the 1965 mean high water. The lowest elevations, which are still preserved on remnants of the preuplift surface, are 13 feet above the postearthquake mean high water on sand and 9 feet above on silt. Thus, the top of the seaward steep slope apparently was at least 10 feet below spring low water (1963) on the silt and at least 6 feet below on the sand. The top of the steep slope may be close to the approximate position of the +10-foot contour of the upper surface shown in figure 2.

The presence of a gravel sheet or sheets, both below the 1965 stream courses and in the outliers of silt and sand, suggests that a former stream system similar to the one which is now developing produced these gravel sheets. Since the gravel sheets could be traced down to the 1965 mean sea level, they very probably indicate a former relative sea level that was at least as low as the 1965 one and that predates the finer deposits. Cores taken in the present intertidal zone (Barrett, 1966, p. 997) show a series of pebbly layers extending down as low as 1 foot above mean lower low water. These sections seem to show deposition of 25 to 30 feet of sand in and close to the intertidal zone, and imply a fairly steady rise of relative sea level before the earthquake. A similar steady rise has been reported for other parts of Prince William Sound (Plafker and Rubin, 1967).

Some indication of the rate of rise of relative sea level has been obtained by the dating of two wood samples. C–14 sample W–1768, from 4 feet above postearthquake mean high water (section 2, fig. 2), was interbedded with sand that probably was deposited above extreme low water level (14 feet below the 1965 mean high water). The radiocarbon age of this sample is 820±200 B.P. C–14 sample W–1764, from a tree in the position of growth approximately at the level of the preearthquake barnacle line, has a radiocarbon age of 380±200 B.P. and indicates that mean high water was then at least 5 feet lower than the preearthquake mean high water. Rates of fall of relative sea level calculated from samples W–1768 and W–1764 are 0.033±0.008 foot per year and more than 0.014±0.013, respectively.

STREAM EROSION

Bay-head silt, sand, and gravel were studied only at the head of MacLeod Harbor. Since the uplift of about 33 feet in March 1964, the area of intertidal and shallow-water deposits has been considerably dissected by rivers, few of which rise within the area that was formerly below sea level; the coastal perimeter of the uplifted deposits has been trimmed by wave action. By July 1965, about 50 percent of the available deposits had been regraded to a new level, accordant with the present sea level. Most of the regrading has been done by lateral erosion of the largest rivers (fig. 3).

Long profiles were made of 20 streams (numbered 1–17 and 20–22, table 1), particular attention being paid to the position of knickpoints. At least three cross sections were made of each stream in order to estimate the total volume removed from each by erosion. A grain-size analysis was made of samples from near the mouth of each stream, and a relative value of low-flow discharge was calculated. Additional information about the earlier course of erosion was obtained from U.S. Coast and Geodetic Survey aerial photographs taken 8 weeks and 21 weeks after the earthquake.

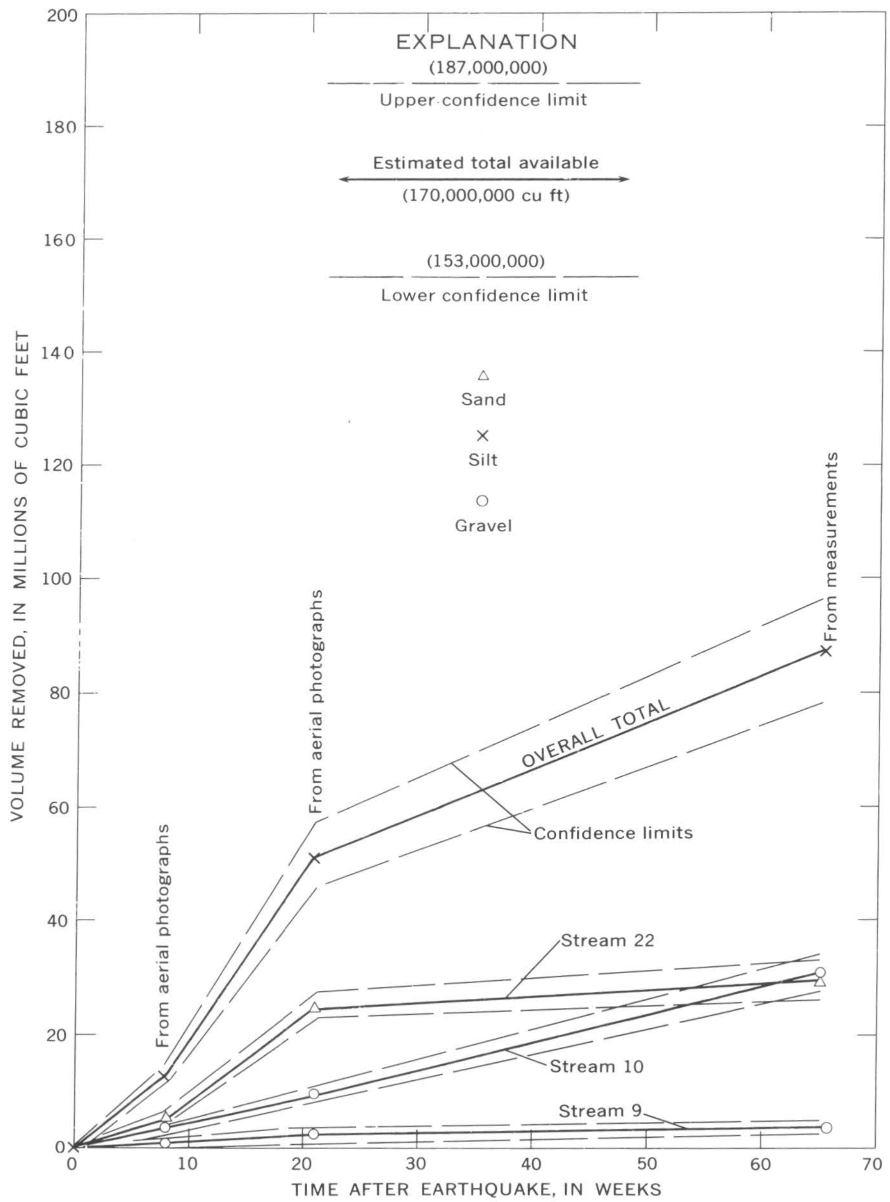
Interpretation of the information collected has been in terms of the overall course of erosion by the rivers, including the recession of knickpoints. This part of the interpretation is essentially a study of fluvial processes in an unusual environment. The data have also been studied in relation to the modification of a raised platform. The rapidity with which such a raised platform is altered, and the agents responsible for it, are discussed on pages H15–H16.

DATA COLLECTED

The banks of streams 1–9, 15, and 16 (table 1) are silt; the banks of streams 10–14 and 20–22 are sand; and the banks of stream 17 are gravel. Bed material is not infrequently coarser than bank material. Low-flow discharges ranged from zero for the smallest

3.—Panorama of eroded bay-head deposits in MacLeod Harbor.

streams in sand and 0.035 cfs (cubic feet per second) for the smallest stream in silt (stream 6) to 301 cfs (stream 10). The volume of material removed by all streams from the time of the earthquake to late June 1965 (fig. 4) was calculated from direct measurements. For the larger streams (streams 9, 10, 12, 21, and 22), estimates of volumes removed at earlier stages were calculated from aerial photographs, which showed the width of erosion, and from planetable maps made in June 1965, which gave the depth of erosion. For these large streams, errors arise mainly from incorrect interpolation of the initial elevation of the surface before stream erosion (fig. 2). This error in initial elevation is unlikely to be more than 1 foot near stream mouths where thicknesses removed are 10 to 15 feet; it will be correspondingly smaller upstream where the rivers are narrower and thicknesses of material removed are 0 to 5 feet. Measurements of widths of erosion, made from field measurements and from aerial photographs, are considered to be much more accurate than the depth measurements. For small streams, errors arise mainly from the intricacy of variation of width and bifurcation, but all major tributaries, were included in our calculations. An overall error of ±10 percent is considered a generous maximum allowance (95 percent



4.—Relationship of volumes removed by fluvial erosion to time elapsed since the earthquake.

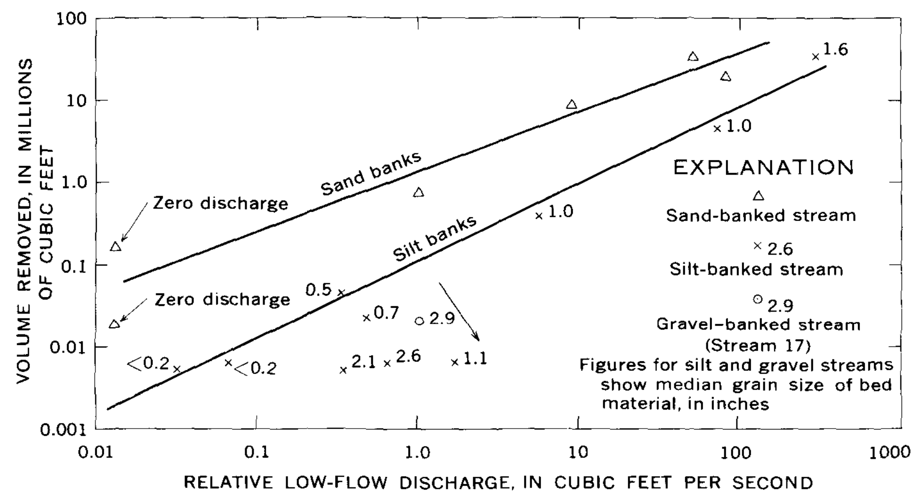
TABLE 1.—*Summary of data for streams in MacLeod Harbor*

[*, no knickpoints apparent in stream]

Stream number	Distance of knickpoint from mouth (feet)	Width at mouth (feet)	Relative low-flow discharge (cfs)	Cross-sectional area at mouth (sq ft)	Total volume of material removed from the time of the earthquake to June 1965 (cu ft)	Size of bed material D_{50} (in.)	Total volume of material removed (millions cu ft)		Type of bank material	Remarks
							May 1964 (8 weeks)	Aug. 1964 (21 weeks)		
1	76	25	0.37	97	5,025	2.1	------	------	Silt	
2	Many knick points.	22	1.73	180	6,160	1.06	------	------	do	
3	75	15	.67	103	6,250	2.64	------	------	do	
4	66	23	.051	128	6,300	Silt	------	------	do	
5	63	18	.035	78	5,250	do	------	------	do	
6	255	32	.35	198	41,500	.49	------	------	do	
7	335	20	.50	85	22,400	.74	------	------	------	
8	1,150	72	5.71	480	361,000	1.01	------	------	do	
9	0	190	73.0	2,600	3.20×10^6	.96	0.81	2.17	do	Gravel bed.
10	0	1,625	301.0	25,200	36.7×10^6	1.56	3.54	9.50	Sand and silt.	Do.
11	*	70	0	650	164,000	Sand	------	------	Sand	
12	0	665	80.8	9,650	18.0×10^6	1.27	3.18	12.8	do	Do.
13	*	130	1.10	1,220	710,000	Sand	------	------	do	
14	*	30	0	120	19,400	do	------	------	do	
15	50	------	.0129	------	------	------	------	------	Silt	Data incomplete.
16	50	------	.0094	------	------	------	------	------	do	Do.
17	110	39	1.09	176	19,600	2.9	------	------	Gravel	
20	*	------	2.16	------	------	------	------	------	Sand	
21	0	490	9.30	6,860	6.04×10^6	------	.92	2.31	do	Gravel bed.
22	0	1,450	52.3	21,700	29.0×10^6	------	5.20	24.7	do	Do.
Total	------	------	------	------	94.3×10^6	------	13.6	51.5	------	

confidence). Streams appear to erode at a more or less uniform rate until they approach the limits set by topography or by neighboring streams, at which time their rate of erosion becomes much less. Stream 10, for example, is continuing to erode at a uniform rate, whereas stream 22 appears to have neared its limit. The overall relationship of volume removed to time elapsed since the earthquake, for these streams and for all the streams combined, is shown in figure 4.

Volumes removed by erosion up to June 1965 have been calculated by summing cross-sectional areas removed along the length of the stream. Figure 5 and table 1 show volumes removed, low-flow discharge, bed and bank material, and other data. Each point in figure 5 has been distinguished as to bank material, and streams flowing between silt banks have been further distinguished as to bed-material size on the basis of a pebble count.



5.—Variations in erosion rate with low-flow discharge of streams in MacLeod Harbor. Arrow shows direction of increasingly coarse bed material.

INFLUENCE OF UNDERLYING MATERIAL ON STREAM EROSION

Three major characteristics of the underlying material apparently control stream erosion.

The first is the permeability of the deposit, which determines the drainage area required to produce surface runoff. For example, surface flow commonly appears on the silt immediately below its upstream contact with a gravel bed, and results from the large difference of permeability. Differences in permeability probably also cause large differences in the ratios of high-flow to low-flow discharge.

The second control is the variation in resistance to erosion. This control might be expected to result in marked differences in drainage densities and networks. A difference in networks is apparent (see fig. 7 for silt and fig. 8 for sand), but differences in drainage density are masked because many streams receive most of their water supply from streams outside the area of uplifted deposits. Their drainage density is therefore partly determined by the number and spacing of the feeder streams.

The third control is the effect of bed armoring. Armoring is widespread in larger streams, mainly with gravel deposits that are eroded less readily than the bank materials. Extensive armoring is usually associated with streams that have almost reached their limit of possible downcutting, so it is not immediately clear, from the data in this area, whether the slow rate of downcutting is a cause or an effect of the armoring.

Differences in erosion rates of streams flowing in silt and in sand banks may be clearly distinguished in figure 5. The distinction seems to depend mainly on the differences in the ratio of high- to low-flow discharges, but the greater erodibility of sand may be a contributory cause. The silt channels can be distinguished by the size of the bed material except for stream 2, in which the small volume of material removed is probably due to the fact that the stream's knickpoint had already reached gravel and it could not cut back farther. Note that the plot of the gravel-banked stream 17 falls with silt-banked streams with the same bed-material size. The data may indicate that armoring of the bed with coarser material tends to reduce the rate of erosion and that recession of the knickpoint is greatly slowed when it reaches the coarse gravel deposits which form the preearthquake storm beach.

6.—Typical stream with silt banks (stream 8, table 1), MacLeod Harbor. Note the conspicuous white clam shells on the silt surface.

STREAMS IN SILT

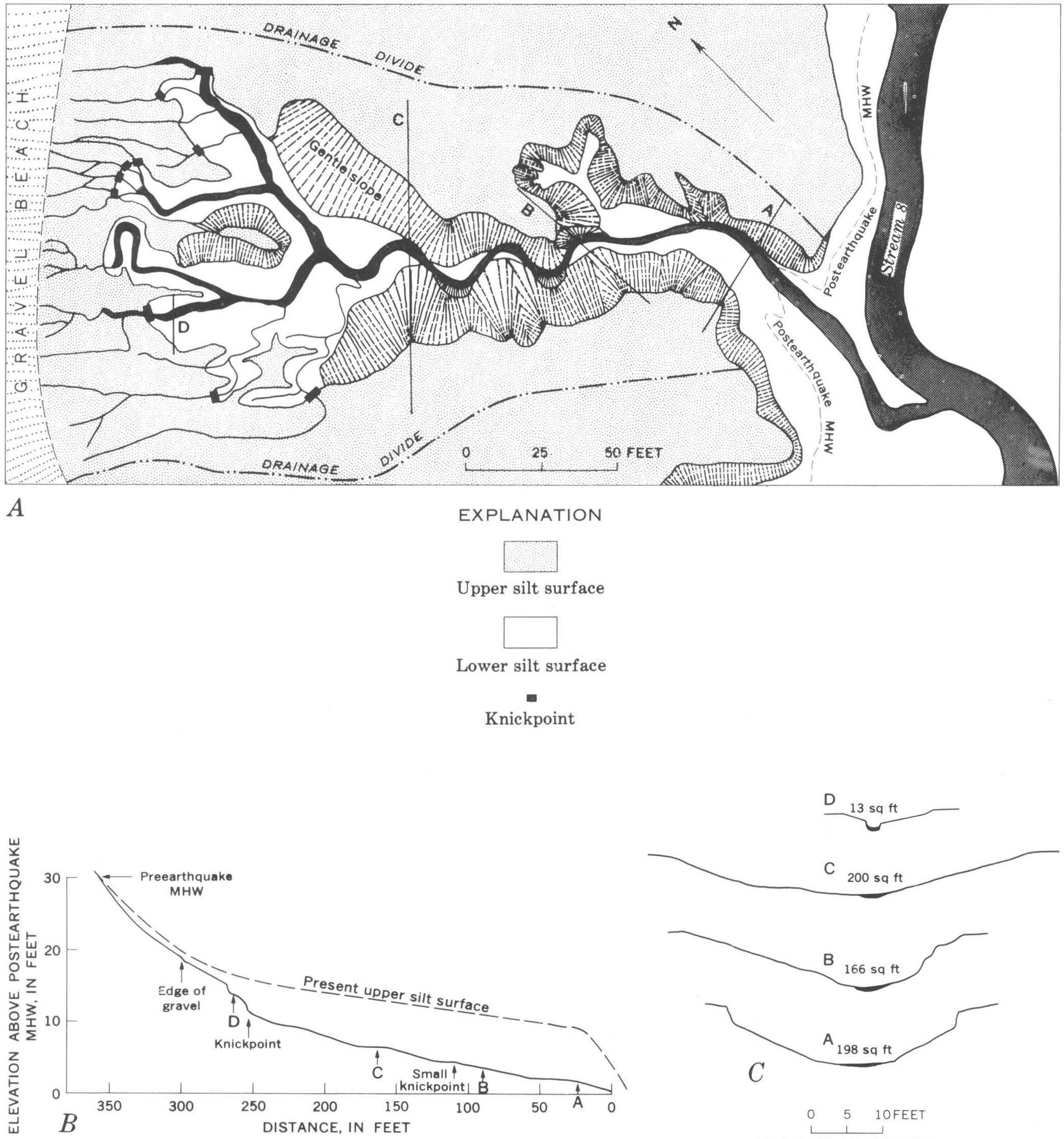
Initial stream courses in silt seem to have been consequent and to follow the directions of greatest slope, as exemplified by the streams that have not migrated laterally and that still follow the reconstructed consequent directions. The initial directions of larger streams, even though they have since moved laterally, also correspond to reconstructed consequent directions. Most of the streams in the silt are fed by runoff from the hillside along the north side of the harbor.

A sharp division can be made between streams that can transport gravel effectively and those that cannot. Those that cannot carry gravel effectively are characterized by an abrupt ending of the channel at its upper end, at the lower margin of the gravel beach or at a gravel fan. The great difference in permeability between silt and gravel emphasizes this feature. In progressively larger streams, the bed material tends to change (provided suitable material is available) from silt to clam shells to gravel. The amount of coarse armoring material and the ability to move it thus increase together, although figure 5 shows that the armor has a net retarding effect on erosion.

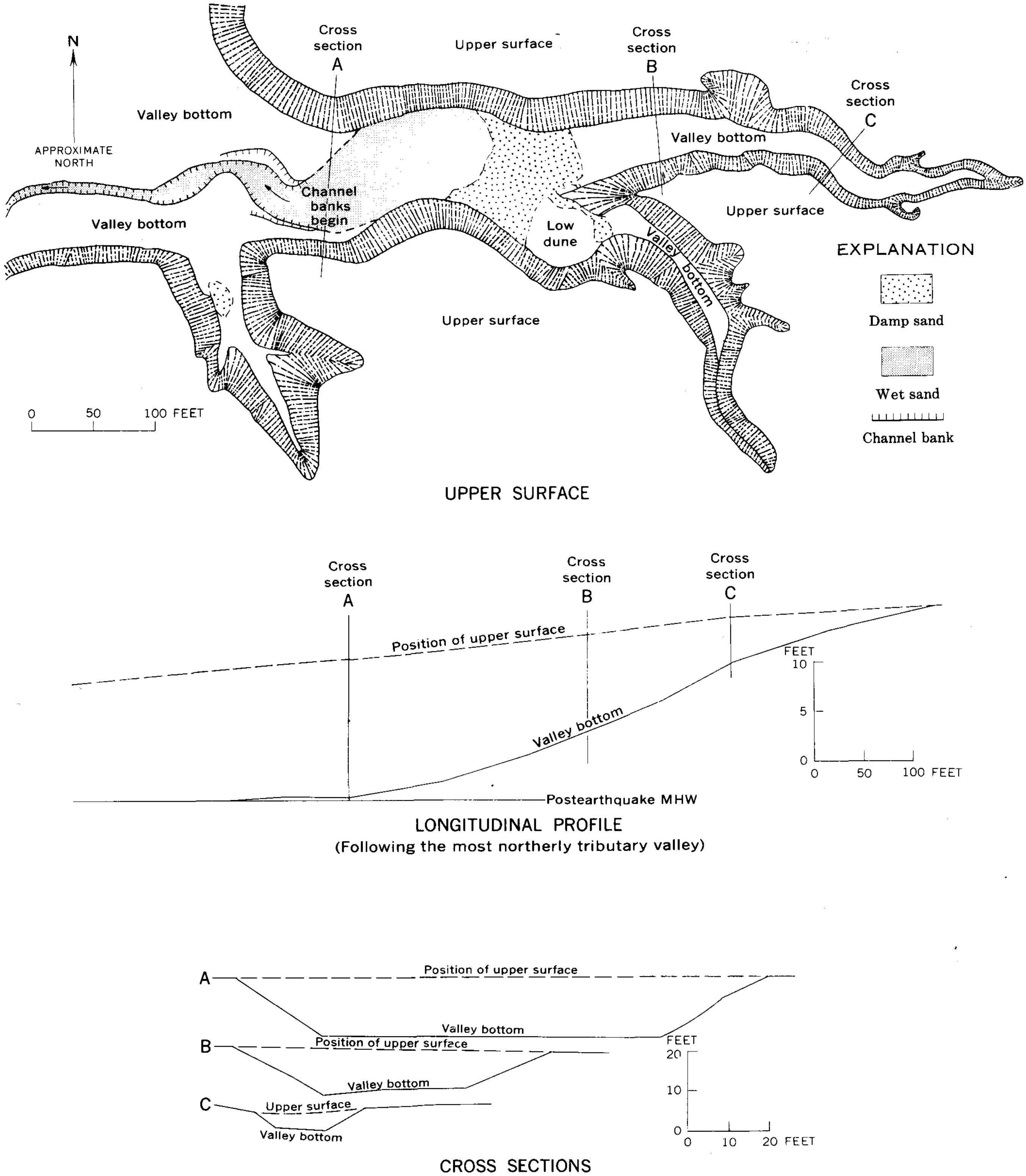
Small streams in silt (fig. 6) generally have steep banks, narrow valleys, and very marked knickpoints in their longitudinal profiles (fig. 7). Streambank gradients locally decline from the usual 60°–70° to 10°–20°, where the silt is saturated. This decline is associated with extensive slumping and earthflows. The saturation is usually caused by unchanneled seepage from the edge of the gravel beach. Figures 6 and 7 illustrate typical silt valleys.

STREAMS IN SAND

The drainage areas of many smaller streams in sand (fig. 8) lie entirely within the bay-head sand area. These streams have no gravel or clam shells on their beds. Clearly defined channels are small and are confined to the seaward ends of the valleys. In the smallest



7.—Map (*A*), long profile (*B*), and cross sections (*C*) of typical stream in silt (stream 6, table 1) in bay-head deposits of MacLeod Harbor. Letters on profile show position of cross sections. Numbers in cross sections refer to cross-sectional area eroded.



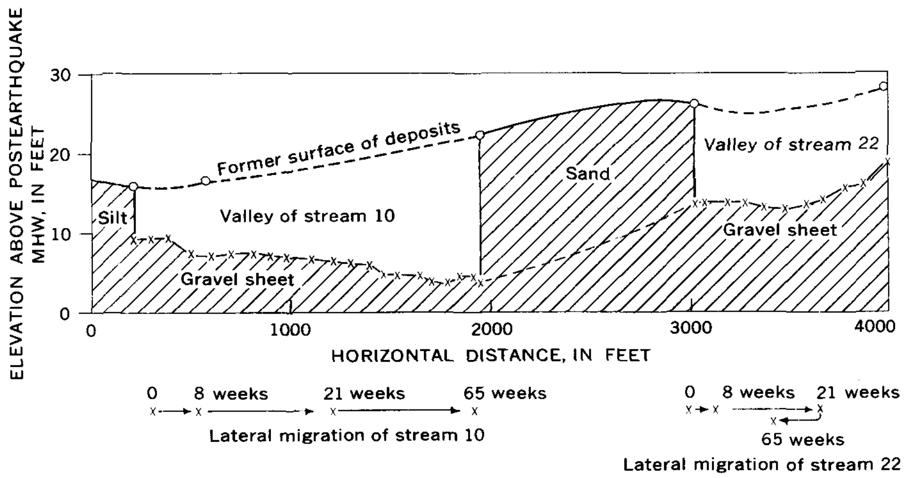
8.—Planetable map, long profile, and cross sections of a typical stream in sand (stream 13, table 1), MacLeod Harbor.

valleys, no channels at all are present. The downstream parts of the sand valleys are broad and have flat bottoms. These are joined, by sharp breaks in slope, to side slopes that are uniformly at the angle of repose. At the top of the side slopes there is a second break in slope, above which is an almost flat upper surface, which roughly corresponds to the original surface of the sand before and immediately after uplift. The whole flat section of the valley floor may be wet, even above the head and outside the banks of the channel; basal sapping of the side slopes apparently is chiefly responsible for valley widening. The upstream parts of the valleys are V-shaped in cross section and are dry on the valley bottom. The irregularity of the long profile suggests that there may be periods of infilling by wind between periods of fluvial erosion. Figure 8 illustrates a typical small stream valley in sand.

Large streams, fed by water from the hillsides above the former sea level, are gravel floored and have steep vertical banks in sand, or in sand and gravel in their upper reaches (but still below the preearthquake water level). Clear knickpoints are not present in the longitudinal profiles of either large or small streams flowing over sandbeds.

STREAM IN GRAVEL

One small stream (stream 17) flows down the hillside and then across a steep gravel beach directly into the sea. The gravel had some finer matrix and was able to stand, at least for a short period, in vertical banks. The stream had a steep course and a knickpoint zone less well defined than those of the streams in silt (fig. 14, p. H18). It is debatable whether placement of this single example with the streams in silt (fig. 5) is normal.

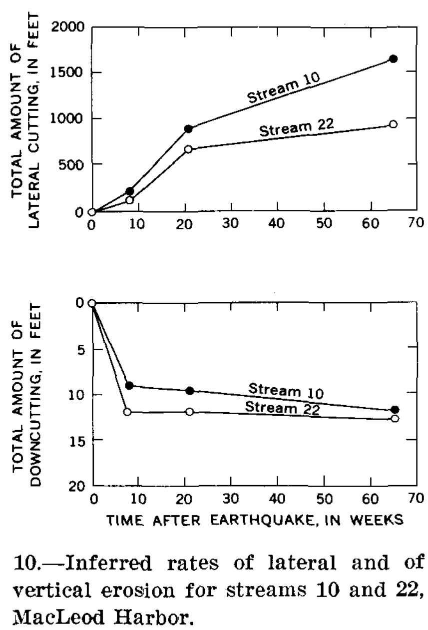


9.—Cross profile of gravel sheet and lateral migration of streams 10 and 22 (table 1), MacLeod Harbor. Arrows are drawn to scale, and show distances and direction of migration.

LATERAL AND VERTICAL EROSION

Figure 9 shows gravel sheets that crop out approximately at the surface of the present major streams. One or more gravel sheets at a similar level may extend beneath the remaining outliers of silt and sand. This preexisting gravel probably has some influence on the rate of downcutting of the streams. At one extreme, the control may be so complete that the present streams are simply being guided in their lateral erosion into hollows in the former gravel sheets and are thus prevented from further downcutting. At the opposite extreme, the gravel sheets may only slow downcutting temporarily, without tending to shift the streams laterally. The lateral migration of stream 10 shown in figure 9 might support either possibility, but the reversal of the direction of lateral migration of stream 22 shows that the stream does not simply move into a hollow and stay there; if streams did this, they would never change their direction of migration. Instead, it is concluded that lower positions of the gravel sheet represent later times and that the streams are steadily downcutting without rigid control of their lateral migration by the gravel. Figure 9 can thus be interpreted as a record of the rate of downcutting, which can be dated by reference to the lateral position of the streams in aerial photographs.

The rates of lateral and vertical erosion deduced in this way (fig. 10) tend to support the generally held view that lateral erosion mainly takes place after most downcutting is completed. Observations of Mr. C. LaBounty, (oral commun., 1965) indicate that the bulk of the vertical ero-



10.—Inferred rates of lateral and of vertical erosion for streams 10 and 22, MacLeod Harbor.

sion in the major streams was very rapid, having been almost completed within 48 hours, and that the lateral erosion has been proceeding ever since. Erosion in unconsolidated sediments probably is limited by the rate at which material can be transported by the stream. Thus vertical erosion and lateral erosion compete for available capacity, as figure 10 seems to show for the first period of 8 weeks. Vertical erosion will result in changes in stream gradient—perhaps the most independent control of transport rate in a given cross section—and gradient is itself controlled by the rate of sediment transport from upstream. Material eroded from the bed will be redeposited if the slope is lowered too much, and the result will be very little net vertical erosion. No such regulatory mechanism applies to lateral erosion, which may therefore continue unchecked until all available material is out of the possible meander belt of the river. Figure 10 illustrates this general course of events, but gives no information about possible mechanisms.

MODEL OF EROSION

The results of these erosional studies are valid only within a limited framework. It would for example be quite improper to apply them to conditions in which the erosion was limited mainly by the resistance of the materials. Within the framework, however, the overall course of erosion following a sudden relative lowering of base level may be approximated by an equation of the form:

$$V=A\cdot Q^{r}(1-e^{-kt})$$

where Q is the low-flow discharge, in cubic feet per second; V is the volume removed, in cubic feet; t is the time since the change of base level, in weeks; and A, r, and k are constants. For the MacLeod Harbor sediments, the constants A, r, and k have the following approximate values:

Sand: $A=1.9\times10^{6}$; $r=0.75$; $k=0.012$.

Silt: $A=1.7\times10^{5}$; $r=0.95$; $k=0.012$.

The differences apparently due to lithology may partly reflect differences in hydrology rather than in erodibility (see p. H10–H11).

Total volume removed seems to be the only consistent measure of the course of erosion. Recession of knickpoints (where present), cross-sectional area at the mouth of the stream, and width at the mouth of the stream all showed much more scatter in their relation to other variables.

RECESSION OF KNICKPOINTS

The knickpoints in the smaller streams in silt apparently are controlled mainly by slight variations in resistance of the material; the streams seem to pick out certain bedding planes which are not otherwise distinctive. The knickpoints recede until they finally become fixed at the junction with the gravel beach above; this junction has already occurred for stream 2. On the larger streams, the gradient of the stream above the former high-water mark and the gradient on the gravel spread below it seem to be more or less independent. No identifiable knickpoints were observed in the long profiles, the only recognizable feature being a local steepening, probably associated with the former beach or storm beach.

OTHER EROSIVE AGENTS IN REGRADING OF DEPOSITS

The three principal types of erosional processes that modified the bay-head deposits were fluvial, marine, and slope processes. The most effective slope processes probably were the slopewash, caused by raindrop impact, and erosion by wind.

In the broad expanse of deposits in MacLeod Harbor, fluvial erosion seems to be the most important process. The sea can only act along the perimeter, but rivers act over the whole surface. The volume of material removed by the rivers has been estimated from the planetable map of the deposits (fig. 2).

Along the margin of the deposits, marine erosion produces cliffs as high as 18 feet. Below these cliffs a narrow beach is being formed, coarse material accumulating at its top. Some intertidal sorting of material may also take place, but its total effect in MacLeod Harbor is thought to be slight.

On the silt deposits, clam shells protect small sloping pillars of material; the silt around them is detached by raindrop impact and washed away by surface runoff. At the time of measurement, all these pillars sloped at about 30° to the horizontal in the same direction, which is presumed to be the angle and direction of the driving rain. Such an angle is probably not characteristic of all storms, so the heights measured probably represent only the last major storm. The median vertical height of 50 of these pillars was 1.9 inches (0.16 ft).

The total effect of this rainbeat since the time of uplift could be seen as follows: The shipworm (*Bankia* spp.) bores into wooden pilings underwater, but it cannot live within the silt; pilings become honeycombed with these boreholes and eventually break off within an inch of the surface of the silt; a line of old broken-off pilings projecting out of the present silt beach therefore records the former surface level very accurately, and the total erosion of the surface may be measured. The rarity of surface runoff on the sand and gravel is thought to minimize erosion on these deposits by rainbeat and rainwash.

The profile of the silt along the line of pilings is shown in figure 11. Both the volume of material removed since the earthquake by surface wash above the top of the low seacliff (an average thickness of 0.7 ft) and the cross-sectional area removed by marine erosion (175 square ft) can be calculated from figure 11. These computations can be compared with the volume removed by fluvial erosion. Volumes removed by the three main processes responsible for the regrading of the deposits are:

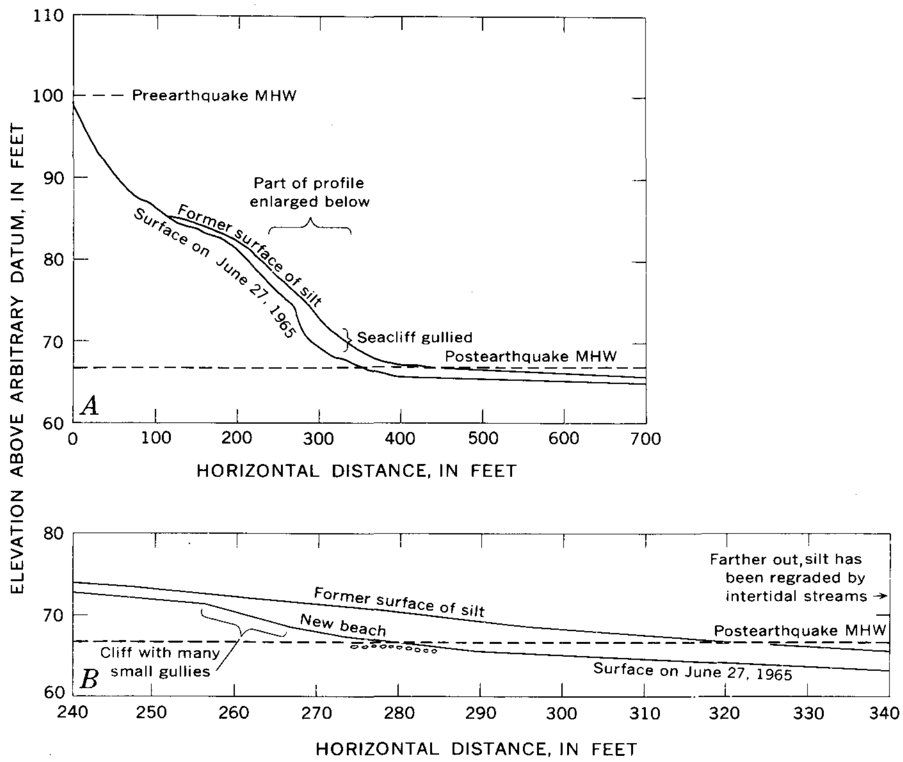
By rivers (see fig. 4) = 87 million cu ft

By rainwash = 0.7 ft × area of silt deposits = 2 million cu ft

By the sea = 175 sq ft × coastal perimeter = 700,000 cu ft.

Wind action is locally evident on channel sides in the sand areas only. The total effect probably is a rearrangement rather than a net removal of material, because the form of the upper surface of the sand is remarkably even and shows no hollows, such as might be expected if deflation had occurred.

It is concluded that, in MacLeod Harbor, the agents responsible for regrading are fluvial processes, rainwash, and marine action, in that order. However, a different order of importance might be obtained for, say, a narrow gravel beach.



11.—Postearthquake degradation of silt, by rainwash and marine erosion, along a line of pilings. *A*, Profile of entire beach; vertical exaggeration × 10. *B*, Part of *A*, without vertical exaggeration, showing retreat of seacliff.

EROSION AND DEPOSITION ON UPLIFTED BEACHES AND ROCK PLATFORMS

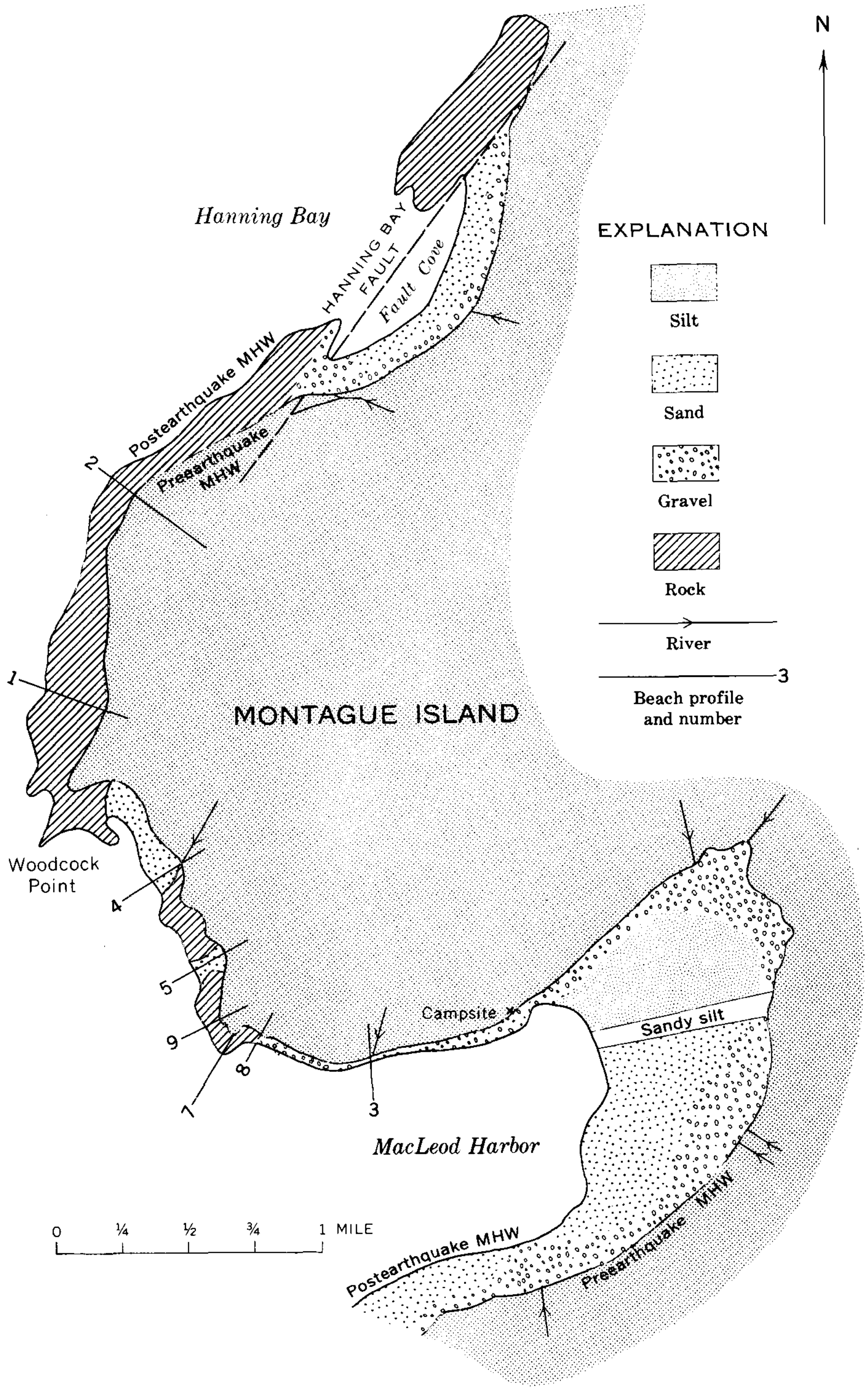
Beach studies in MacLeod Harbor and Patton Bay had three objectives. The first was to provide data on the detailed form of beaches and offshore platforms that are difficult to study under normal conditions; the second was to observe the rates of erosion and deposition on the raised beach as an aid to the general study of raised beaches; and the third was to clarify the tectonic history of the island. These three objectives are closely related and data were collected from (1) profiles perpendicular to the shoreline, (2) surveys of selected geomorphic features and of amounts of erosion and deposition, and (3) stratigraphic sections and their associated radiocarbon dates.

MacLEOD HARBOR BEACHES

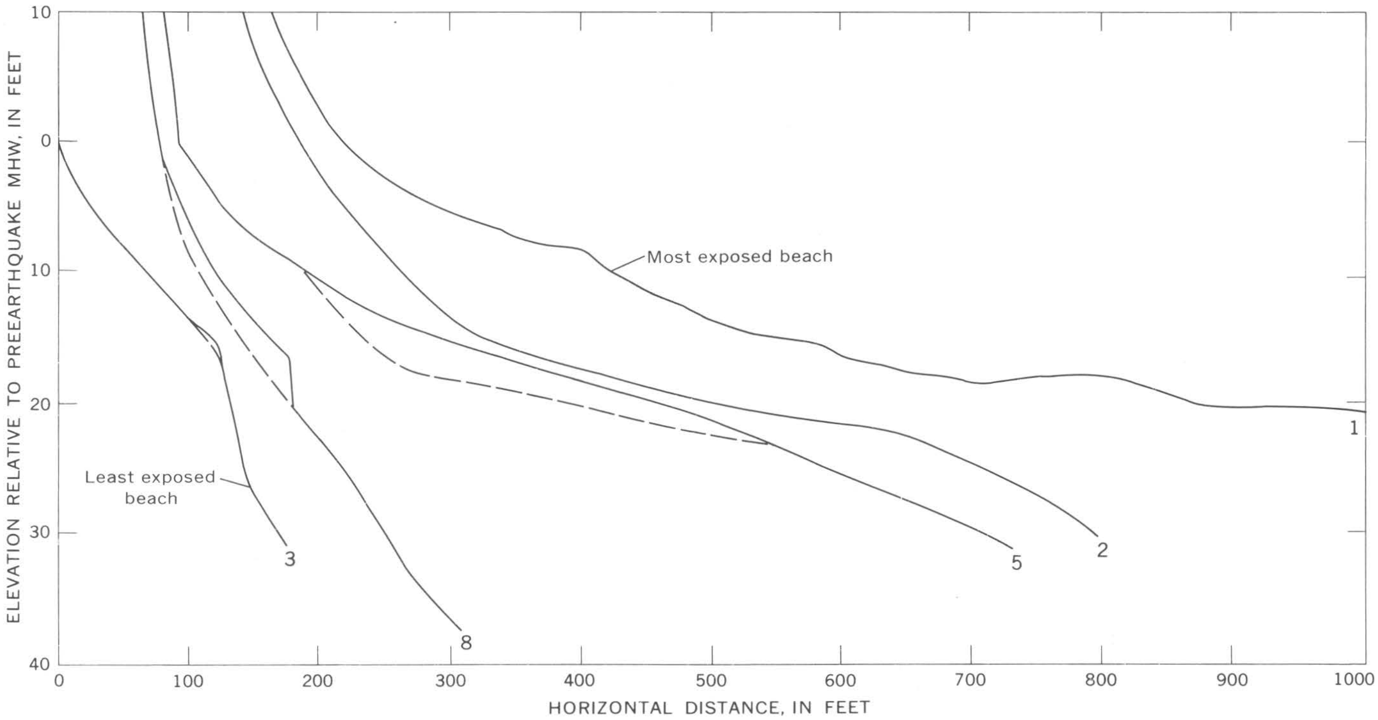
BEACH PROFILES

Eight longitudinal profiles of preearthquake beaches and uplifted sea floors were made in the MacLeod Harbor area (locations shown in fig. 12). Most profiles were on rock platforms, but some were on beaches that were composed wholly or partly of sand or gravel. Interpretation of the profiles has been made in terms of the relief found on rock platforms, in terms of marine regrading of beaches, and in terms of knickpoint recession and fluvial regrading.

The profiles seem to have gentler



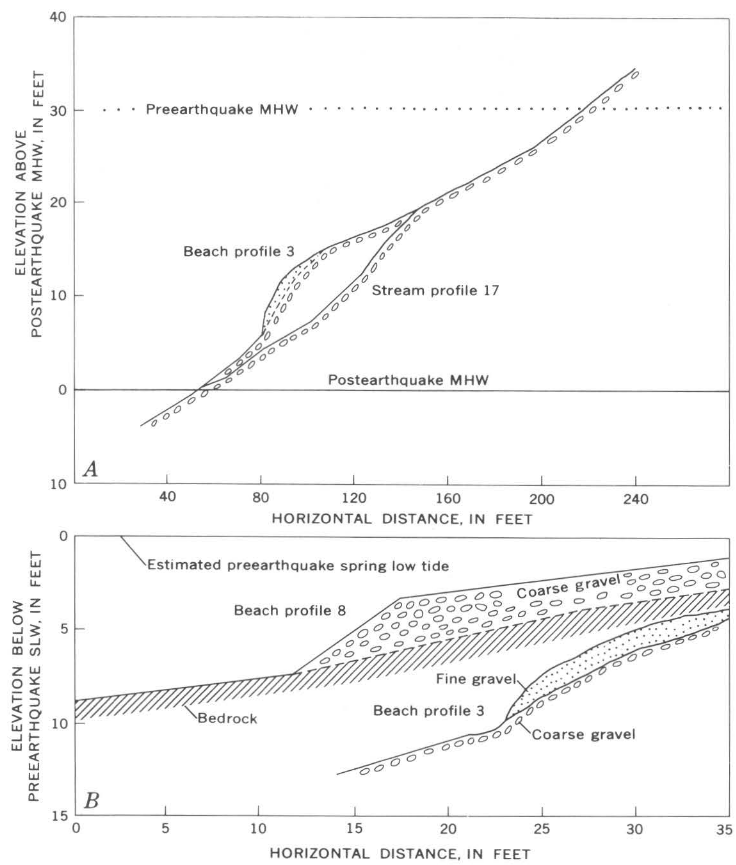
12.—Shoreline lithology and location of surveyed beach profiles between MacLeod Harbor and Fault Cove.



13.—Beach profiles around MacLeod Harbor showing the effect of increasing exposure to waves. Broken lines show effect of local modification by stream action.

overall slopes below the preearthquake mean high water and broader rock platforms where the beaches are more exposed to waves of long fetch. This relationship is shown by the series of five representative profiles in figure 13. How much of the differences in profiles is due to varying wave exposure and how much, for example, to varying initial slopes of the coast is unknown.

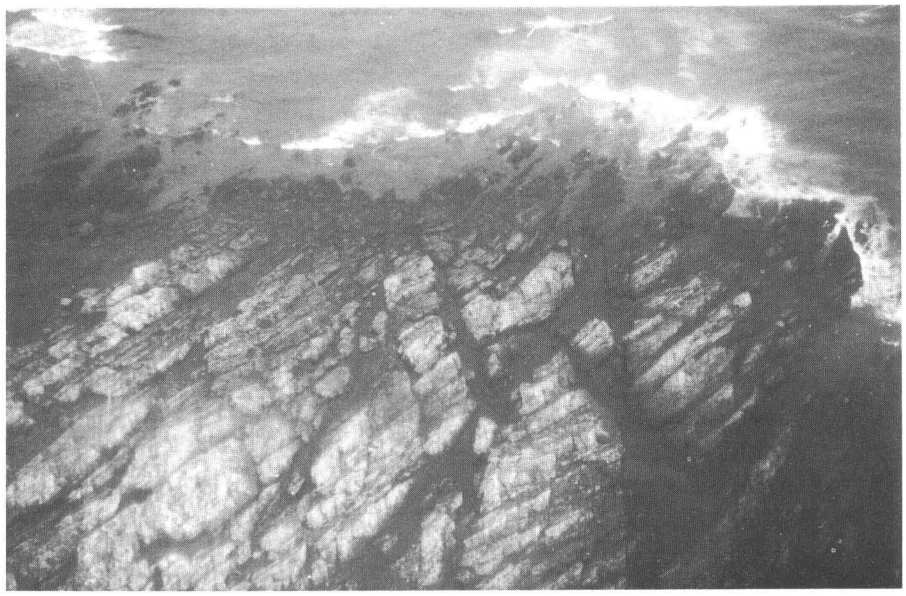
Sediments from the upper parts of both present and former beaches show a consistent change in grain size down slope; the finer material is found in deeper water and on lower gradients. In comparing profiles 3 and 8, a given grain size



14 (right).—Beach profiles 3 and 8 (figs. 12, 13), and stream profile 17 (table 1), MacLeod Harbor. *A*, Beach profile 3 and stream profile 17, vertical exaggeration × 4; *B*, parts of beach profiles 3 and 8 referred to preearthquake spring low-tide level, no vertical exaggeration.

on profile 3 is found on a steeper gradient and in shallower water than on profile 8. This difference may be partly explained by differences in sizes of material available, but it fits in well with the fact that profile 8 is the more exposed (fig. 13).

Profiles 3 and 8 seem to have a step composed of different materials deposited over an earlier surface. The top of the step of both profiles is about 3½ to 4 feet below the level of former spring low tides (fig. 14). The top of the steps seems to represent the shelf of deposits that was being built at the preearthquake sea level below the zone of effective wave transport. The relatively shallow depth compared with that for the bay-head sediments (about 10 ft below preearthquake spring low water) is consistent with the greater coarseness of the material in the beach profiles. On the more exposed beaches (profiles 1, 2, 5), the slope of the rock floor is probably too gentle for the extension of sediments from the preearthquake sea level to have produced a detectable step. The exposure of a former surface, on which this step rests, shows that at some previous time there must have been a sea level in the area at least as low as the postearthquake level. Such an earlier low sea level also explains the existence of gravel sheets beneath the bay-head deposits, although no dateable material was found.



15.—Rock platform showing differential erosion following the structure.

16.—Rock platform near beach profile 8 (fig. 12), MacLeod Harbor.

A marked break in slope, similar in form to a wave-cut notch, was seen on the seaward side of an abandoned stack near profile 8, and in the rocky promontory at profile 7. This break is, at both places, about 16 feet below the preearthquake barnacle line. The break, which presumably was formed at or near a former high-tide level since it does not seem to be structural, may mark a former sea level that is between the pre- and postearthquake levels, although it presumably predates them.

Bedrock platforms are notably smooth in overall outline, but in detail vary from highly irregular surfaces with as much as 5 feet of local relief to much smoother surfaces with less than a foot of local relief. All the differences seem to emphasize lithological and structural differences in the near-vertical beds (figs. 15, 16).

REGRADING OF THE BEACHES

The uplifted MacLeod Harbor beaches have been modified by both marine and fluvial processes, but slope processes seem to have had little erosional effect.

Marine processes act only in the zone below the storm beach. In the MacLeod Harbor area, storm beaches corresponding to the preearthquake sea level range from 9 feet (at the head of MacLeod Harbor) to 12 feet (profile 4) above mean high water. There is no evidence of new storm beaches beginning to form, but, since uplift, the sea has regraded a zone between mean high water and 4 feet above it. In the short period since the earthquake, all unconsolidated material (but no bedrock) in this zone has been regraded. Marine regrading probably will eventually extend up to the full elevation of a new storm beach. Marine processes have had a much greater effect than fluvial processes in the regrading of all beaches studied.

Fluvial processes are effective only where sand or gravel occurs on the beach. Beach profiles 4 and 5 (fig. 13) are along streams, and stream 17 (table 1) is very close to profile 3. Comparison of profile 4 with profiles 1–3 and 8 (fig. 13) shows the typical almost-straight beach profile that results from regrading by a competent stream flowing mainly through sand. The upper part of the course of the stream in profile 4 on the uplifted beach is erosional; lower down the stream spreads out into a fan and effectively regrades a width of several hundred feet of the beach by deposition. The low-flow discharge of the stream is 10.0 cfs.

The stream in profile 5 has a low-flow discharge of only 0.03 cfs. This stream is competent to erode sand from an area underlain by cobbles but it cannot carry the cobbles, so the fluvial regarding is limited to erosion and deposition of sand. Not far from the line of profile 5, the stream is divided into several small channels, each 4 to 10 feet wide and 1 to 2 feet deep, which do not appear to be cutting laterally; hence the area regraded here is much smaller than that regraded by the stream in profile 4.

Stream 17, cutting into beach profile 3, is a small stream with a low-flow discharge of 1.09 cfs, and a very steep course through unconsolidated gravel (fig. 14). This stream illustrates the tendency for a channel of uniform gradient to develop across a beach. Lateral erosion is at a minimum here, apparently because this stream is still actively downcutting. The area regraded is therefore very small.

On the beach, which is both narrower and steeper than the uplifted bayhead, some resorting of beach deposits has already taken place, and streams flow at much steeper gradients than in the bayhead deposits. Thus stream 17 is steeper on the beach than it is farther upstream, whereas the stream in profile 4 has the same gradient above and below the preearthquake water level. The stream in profile 5, however, which falls down a steep gully onto the beach, is less steep in its beach section than immediately inland. The gradient of a stream across the beach, therefore, seems to be unrelated to its gradient above the preearthquake beach; in other words, the relative drop in sea level caused by the earthquake will not necessarily lead to the propagation of a knickpoint inland along every stream course. The formation of a knickpoint appears to be least likely in small streams flowing from steep hillsides.

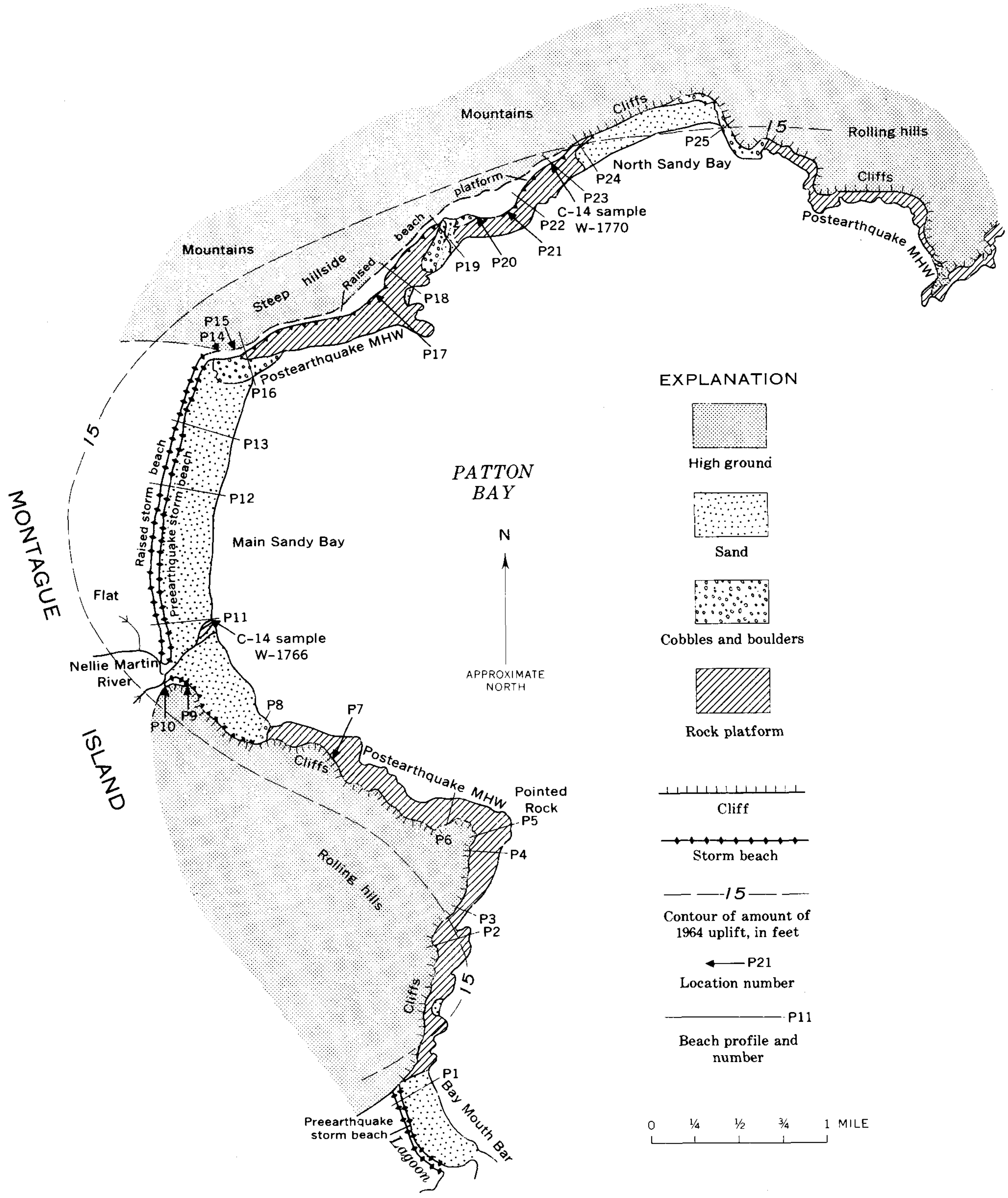
PATTON BAY BEACHES

Seventeen beach profiles, on all types of material, were made in Patton Bay, across the island from MacLeod Harbor. These profiles show the relationships, both vertically and areally, between the following features, if they are present: postearthquake high-water mark, postearthquake storm beach, preearthquake high-water mark (the former barnacle line), preearthquake storm beach, and breaks in slope at the foot of cliffs corresponding to both the preearthquake sea level and to raised sea levels. Regrading of the uplifted beaches was also studied.

Regrading of sand and gravel was similar to that on the other side of the island. An estimate was also made of the rate at which rock platforms on the Patton Bay side have been regraded since the abandonment of the older uplifted shoreline. The many seacliffs, some several hundred feet high, also provide an opportunity for examining the way in which the break in slope at the foot of a cliff is obliterated by subaerial processes.

Most of the coastline of Patton Bay now exhibits a rock platform several hundred feet wide, the landward edge of which is generally covered by beach deposits (fig. 17). This platform, which seems to accord with the preearthquake sea level, is backed, except in the center of the bay, by a steep seacliff about 100 feet high. Above this sea cliff the land is gently rolling and rises 200 to 300 feet above sea level. In the center of the bay coastline, however, the cliff is absent, and instead there is a low step and a raised beach platform 80 to 500 feet wide, which is backed inland by slopes of 40° to 50° leading up to the main 2,500-foot backbone of the island.

Sandy beaches occur in three main areas above the coastline studied. Just south of the center of the bay is the most extensive stretch of sand, called in this report "Main Sandy Bay." The largest river on the island, the Nellie Martin River, debouches at the south end of the bay. At this point the present beach is backed by two



17.—Coastal features and deformation caused by the 1964 earthquake, Patton Bay.

storm ridges, behind which is an extensive flat area drained by the Nellie Martin River and its tributaries. There is a second sandy area called in this report "Bay Mouth Bar" at the southern limit of the coast studied, where a lagoon is dammed up behind two storm-beach ridges. The third sandy area is to the north of the center of the bay. This area, called "North Sandy Bay," is backed by a cliff about 400 feet high, above which rise the main mountains of the island. There is also a small area of sand and gravel in the middle of the central rock platform, where two small rivers debouch across the platform. In several localities between a sand beach and an area of bare rock platform, a short section is covered by large boulders, 2 to 10 feet in diameter, lying on bedrock; their undersurfaces are angular but their upper surfaces are smooth. This contrast between smoothed upper surfaces and angular undersurfaces suggests that marine action has not been powerful enough to overturn them during the period required to smooth their upper surfaces by abrasion and solution. Smaller rounded cobbles, most of them as much as 6 inches in diameter, between the large boulders indicate that 6 inches is approximately the maximum size of stone that the waves were competent to transport.

The main coastal features of the bay and the positions of the 17 beach profiles and other key locations (indicated by series Nos. P1–25) are shown in figure 17.

STRATIGRAPHIC EVIDENCE FOR SEA-LEVEL CHANGES

HIGH SEA LEVELS

There is a continuous raised beach platform between Main Sandy Bay and North Sandy Bay that definitely represents a single former beach. Three profiles showing the raised beach deposits are described below.

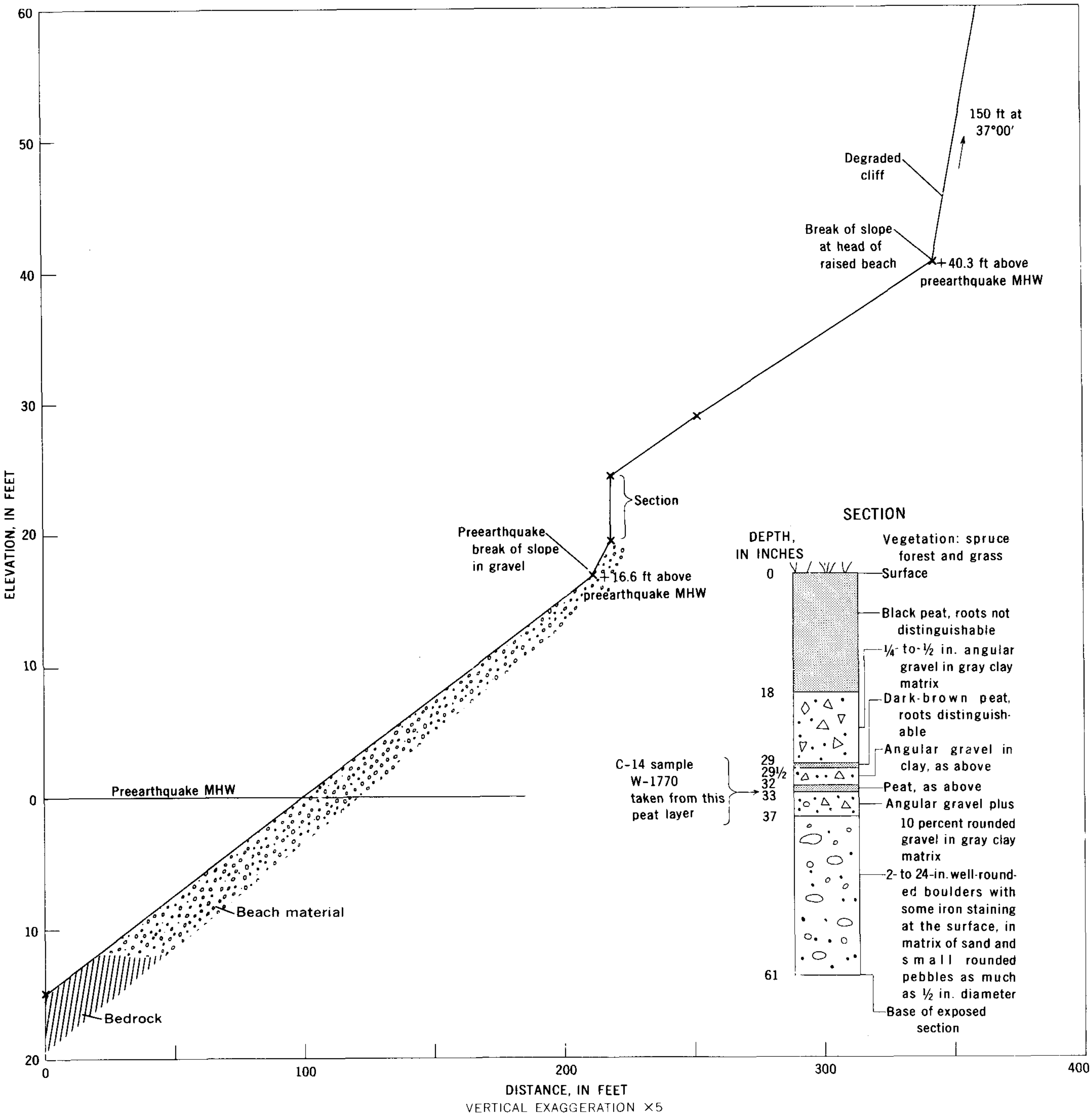
The position of beach profile P23 is shown in figure 17. Figure 18 shows the profile and a stratigraphic section located on it. The material in the bottom 24 inches of the section is similar to the modern cobbles in the upper part of the beach and is definitely identified as a gravel of the former beach. Its topographic situation might make it seem to be a part of the preearthquake beach, but the raised gravel is an integral part of the section above it. When sea level dropped from its raised level to the preearthquake level or lower, vegetation would have covered the former beach material, the cliff at the back of the beach would degrade, and the observed sequence of peat and angular gravel would be deposited. The date of the change of sea level must have been earlier than the deposition of the lowest peat in the section, which has been dated by radiocarbon as 2070±200 B.P. (W–1770).

The position of profile P19 is shown in figure 17. Sections D_1 and D_2 (fig. 19) are close to the bank of a small river, which has exposed a 4-foot-thick bed of angular to subrounded gravel in a finer matrix overlying bedrock. Sections D_1 and D_2 are close together and have been correlated on the basis of the position of the top of the angular gravel bed. The gravel is thought to be a former alluvial fan of the river, deposited while the sea was at a higher level. In section D_1 the level of the raised platform can be clearly seen in bedrock, and a small step in bedrock is exposed at the seaward edge of the vegetated platform. Section F (fig. 19) is close to this step, about 50 feet north of the river. In section F the material resting directly on bedrock is much more uniformly rounded, has less fine matrix, and contains a wider variety of rock types than the angular gravels of sections D_1 and D_2. This rounded material is interpreted as being a true beach deposit, beyond the edge of the alluvial fan.

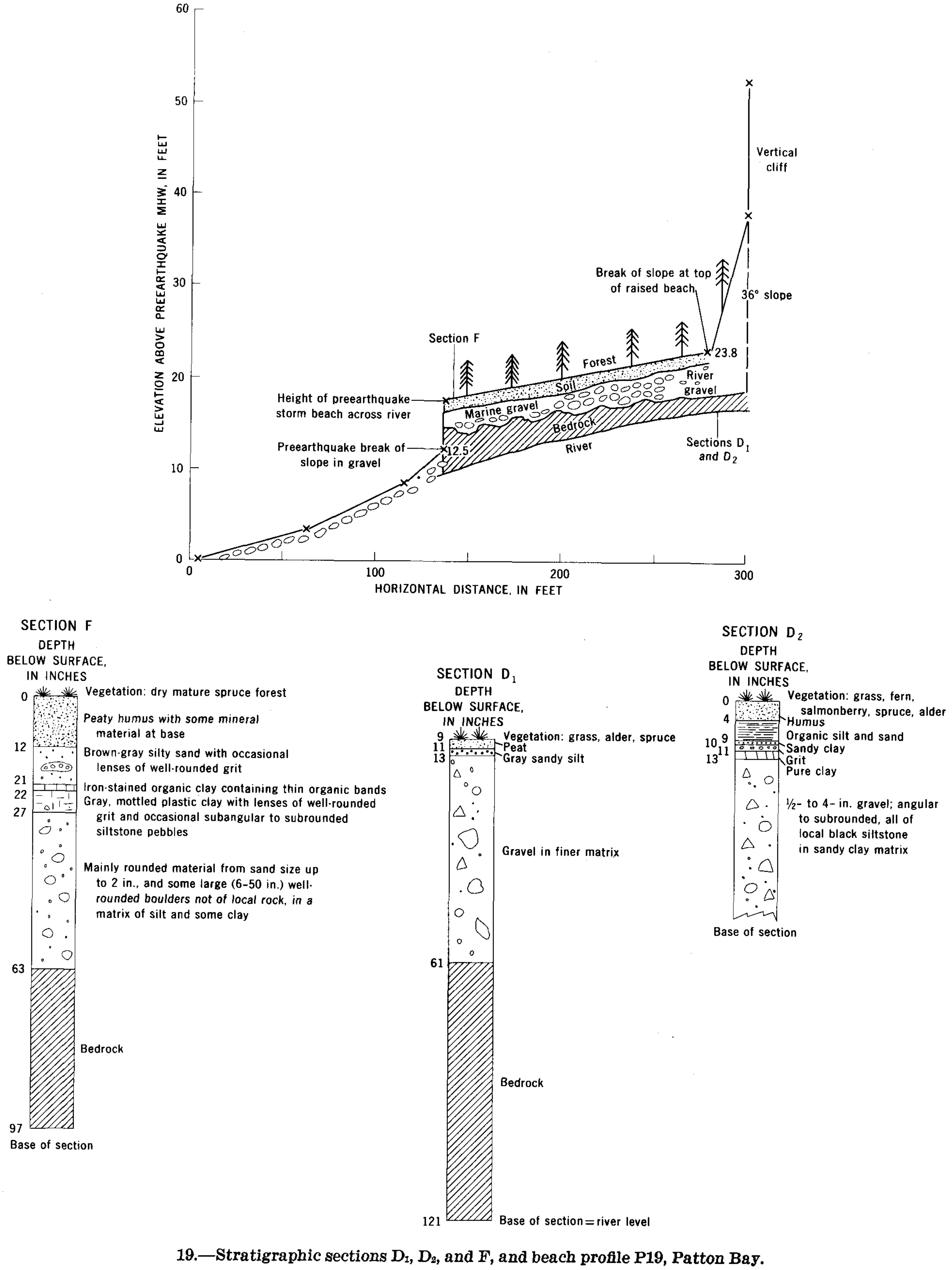
At profile P24 (fig. 17), at the south end of North Sandy Bay, the raised beach platform is much narrower; it is the northernmost exposure where beach gravel is clearly resting on a nonstructural rock platform. The position of the break of slope at the foot of the cliff can be located very accurately here (figs. 20, 21).

LOW SEA LEVELS

Possible evidence of a sea level slightly lower than the preearthquake level is shown by tree stumps, commonly found in the position of growth between preearthquake mean high water and the top of the preearthquake storm beach. These stumps all show some wave erosion. More substantial evidence of a former low sea level is found in peat layers in the banks of the Nellie Martin River. A continuous peat bed is exposed along the banks in the lower course of the river. The top of the bed slopes down toward the sea; it can be definitely identified as much as 1.8 feet below the postearthquake high-water mark and seems to continue farther out to sea (fig. 22). Under present topographic conditions such peat is found behind storm beaches, that is, about 12 feet above high-water mark. It therefore seems probable that this peat layer was formed at a time when sea level was at least 14 feet lower than the postearthquake level at this locality, or 25 feet below the preearthquake level. A sample taken from 2 feet below the top of the peat in section *B* (fig. 22) has a radiocarbon date of 600±200 B.P. (W–1766).



18.—Stratigraphic section and beach profile P23, Patton Bay.



19.—Stratigraphic sections D_1, D_2, and F, and beach profile P19, Patton Bay.

20.—Cobbles of raised beach resting on truncated bedrock platform, profile P24, Patton Bay.

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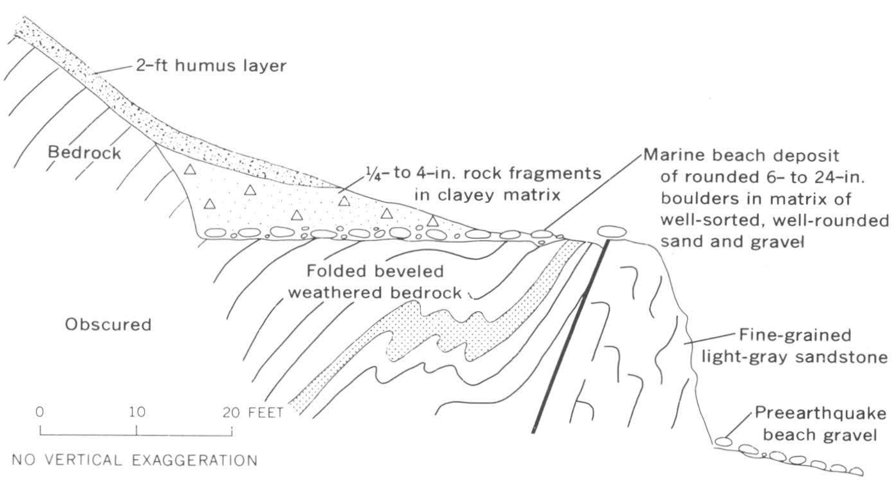
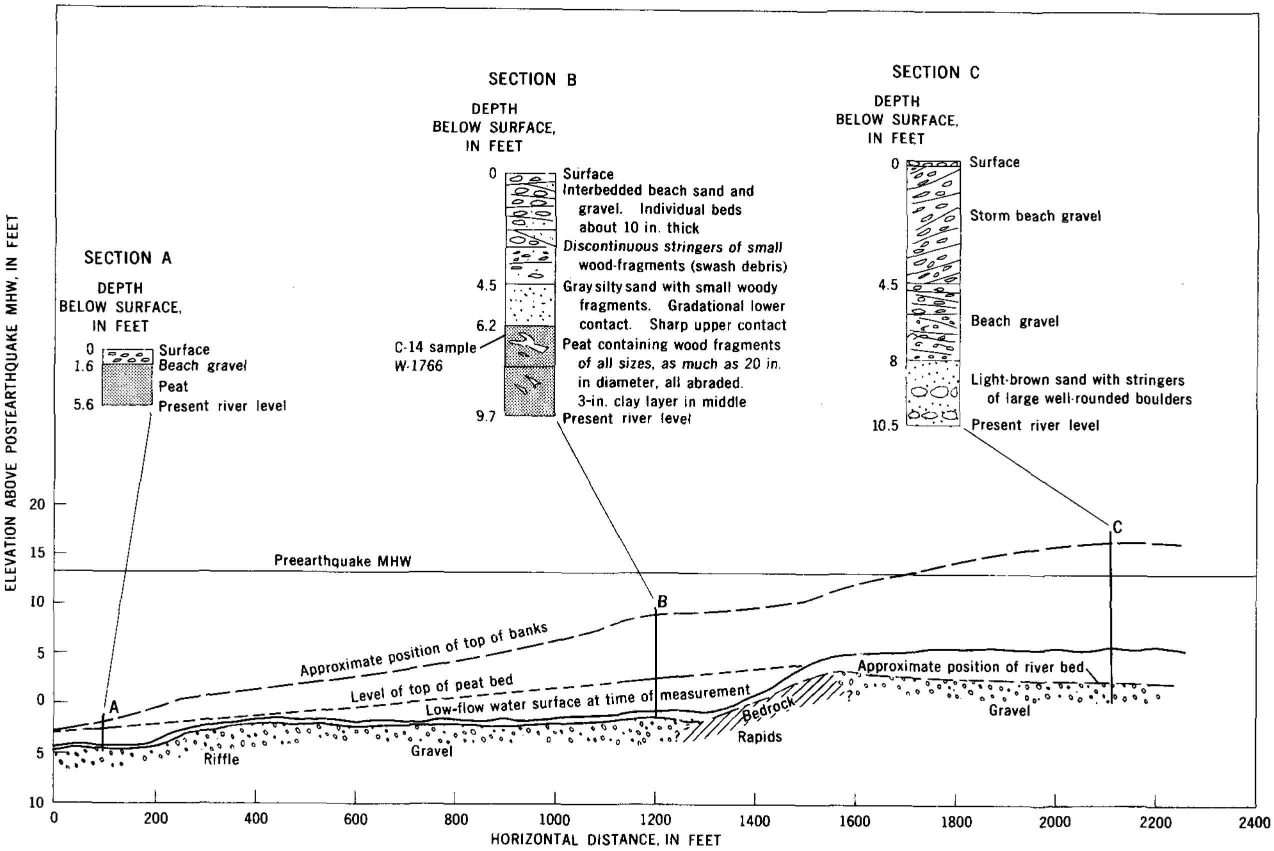


21.—Stratigraphic section showing old raised beach at P24, Patton Bay.



22.—**Stratigraphic sections and long profile of Nellie Martin River.**

It is concluded that there was an earlier raised sea level, higher than the sea level just before the earthquake, the exact elevation of which is discussed on pages H28 to H33. By 2070±200 B.P. the sea level was low enough relative to the land for a layer of peat to have grown upon the rock platform eroded at the higher sea level. A period of low sea level followed during which the sea was low enough for peat to form 1.8 feet below the 1965 high-water mark near the Nellie Martin River. The topographic position of the present peat layer behind storm beaches or higher indicates that the sea was locally at least 14 feet lower than at present relative to the land. The sea was at this low level at 600±200 B.P. Because the dated sample was near the top of the peat layer, it is assumed that relative sea level rose shortly after this time.

ELEVATIONS OF BEACH FEATURES

The sea level at which a raised beach was formed cannot be measured directly; fossil beach features are among the best available indicators of the former sea level. The range of elevation of several beach features, and the relationship of each to the high-water mark, has been studied in Patton Bay, mainly with reference to features formed at the preearthquake sea level. From these data the reliability of each feature as an indicator of a former high-water mark has been assessed, and some conclusions drawn about the causes of variation in elevation for each feature. The beach features chosen for this study were the break in slope between the foot of a cliff and a bedrock platform, the break in slope between a cliff and a gravel beach, and the top of the storm beach ridge.

The elevation of a break in slope between a cliff and a bedrock platform could be measured in only three localities. The elevations of the break in slope above the relevant (preearthquake) mean high water, measured as the barnacle line, were +2.0 (P7), +0.6 (P4), and −0.8 (P5) feet. These figures suggest that, for the conditions found on Montague Island, the elevation of a break in slope in bedrock is within 2 feet of mean high-water level and that this feature is the best indicator of a former sea level.

A gravel beach is a temporary

deposit whose thickness varies with time. A break in slope between a cliff and a gravel beach is therefore likely to show a greater range in elevation between localities than one between a cliff and bedrock platform, and the lower limit of this range should be close to high-water mark. Seven measured elevations of this feature (fig. 23*C*) ranged from 0 to 16 feet above mean high water, with a mean value of +9.3 feet. At location P7 (fig. 17) the elevation of the break in slope increased by as much as 8 feet up small reentrants in the cliffs. It might be expected that the break in slope of a cliff with a gravel beach would give the most reliable evidence of high-water elevation; however, the possible error of ±8 feet shows that this feature is not a reliable indicator of this altitude.

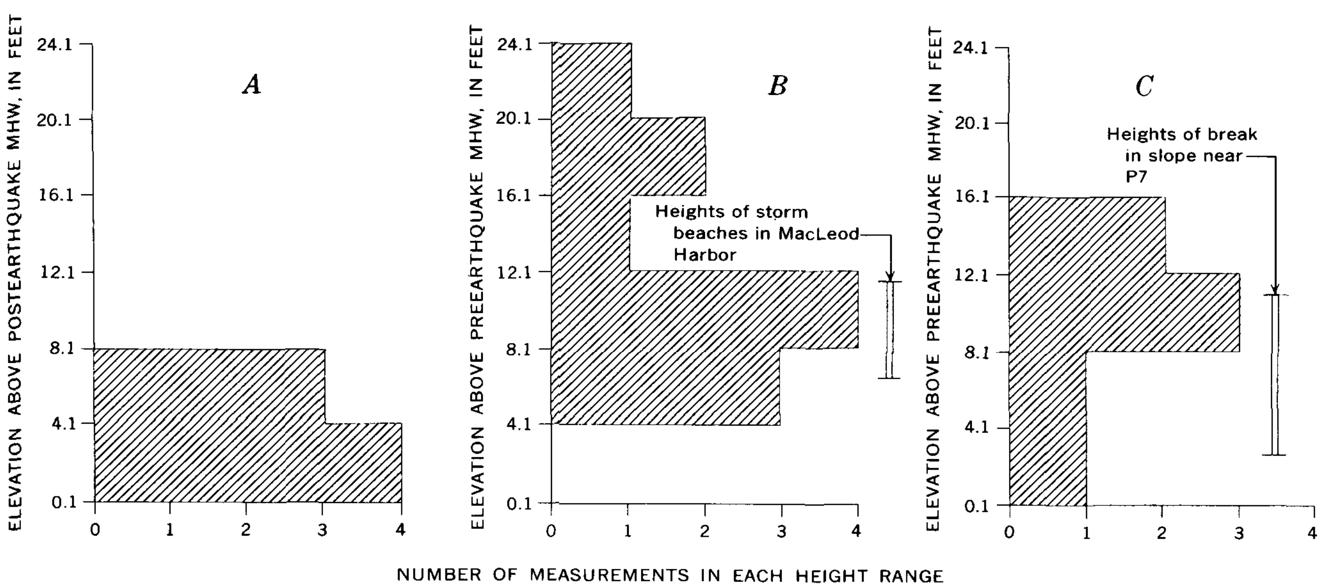
The difference in elevation between the top of the preearthquake storm beach and the preearthquake mean high-water level was measured at 11 points in Patton Bay. The values ranged from 6 to 21 feet above mean high water, with a mean of 12.0 feet (fig. 23*B*). These values are consistent with the idea that a gravel beach banked against a cliff is in effect an incomplete storm beach; therefore, elevations of breaks in slope in gravel beaches should be lower than the elevations of storm beaches.

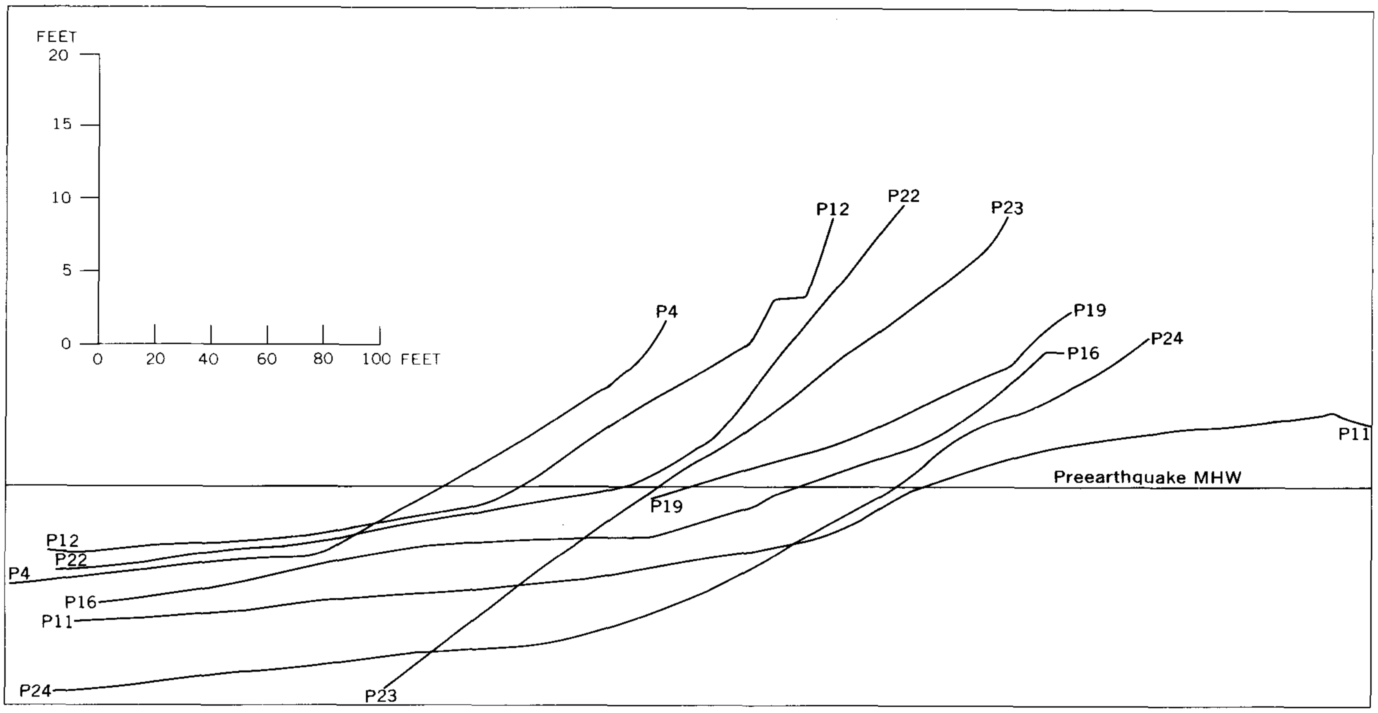
In several locations a new storm beach is probably beginning to form in response to the postearthquake sea level. Seven measurements show that the ridges formed are as much as 6 feet above mean high water (1965), with a mean value of 4.1 feet (fig. 23*A*). The present size and position of the ridges illustrates the initial stages in the formation of a mature storm ridge.

The height of the top of a storm beach is dependent upon wave and shore environment, but it will approach a maximum elevation which is stable for its own environment. Mean high-water level can be determined from the elevation of the top of a storm beach only within ±8 feet, so this feature is also a poor indicator of sea level. However, the difference in elevation between two mature parallel storm beaches should be an accurate measure of the elevation difference in sea levels when the two ridges were formed. This difference has been used in calculating the elevation of the raised sea level in Main Sandy Bay.

In MacLeod Harbor the level of mean high water could be determined from the form of the beach profile, but no satisfactory correlation was found between beach form and water level in Patton Bay (fig. 24). The gradient of the beach, the depth from high-water mark to the seaward edge of the beach gravel, and grain size of the beach material were all found to be of little use for the precise location of sea level. In MacLeod Harbor a consistent relationship was found between the preearthquake water level and the steep seaward edge of the gravel beach (fig. 14), but such a step is only to be expected where the gradient of the

23.—Distribution of elevations of beach features; *A*, above postearthquake mean high water, of postearthquake storm beaches; *B*, above preearthquake mean high water, of preearthquake storm beaches; *C*, above preearthquake mean high water, of the break in slope at heads of beaches.





24.—Beach profiles in Patton Bay in relation to the preearthquake mean high water, showing that the shape of the profile is not a reliable indicator of the position of the mean high water. See figure 17 for location of profiles.

seaward-building gravel is appreciably lower than the angle of the rock platform on which it is built. The evidence from MacLeod Harbor (p. H19) suggests that these conditions are found only in areas well sheltered from wave attack. Because the whole of Patton Bay is more exposed than any of the beaches in MacLeod Harbor, no steps similar to those in MacLeod Harbor are to be expected in Patton Bay.

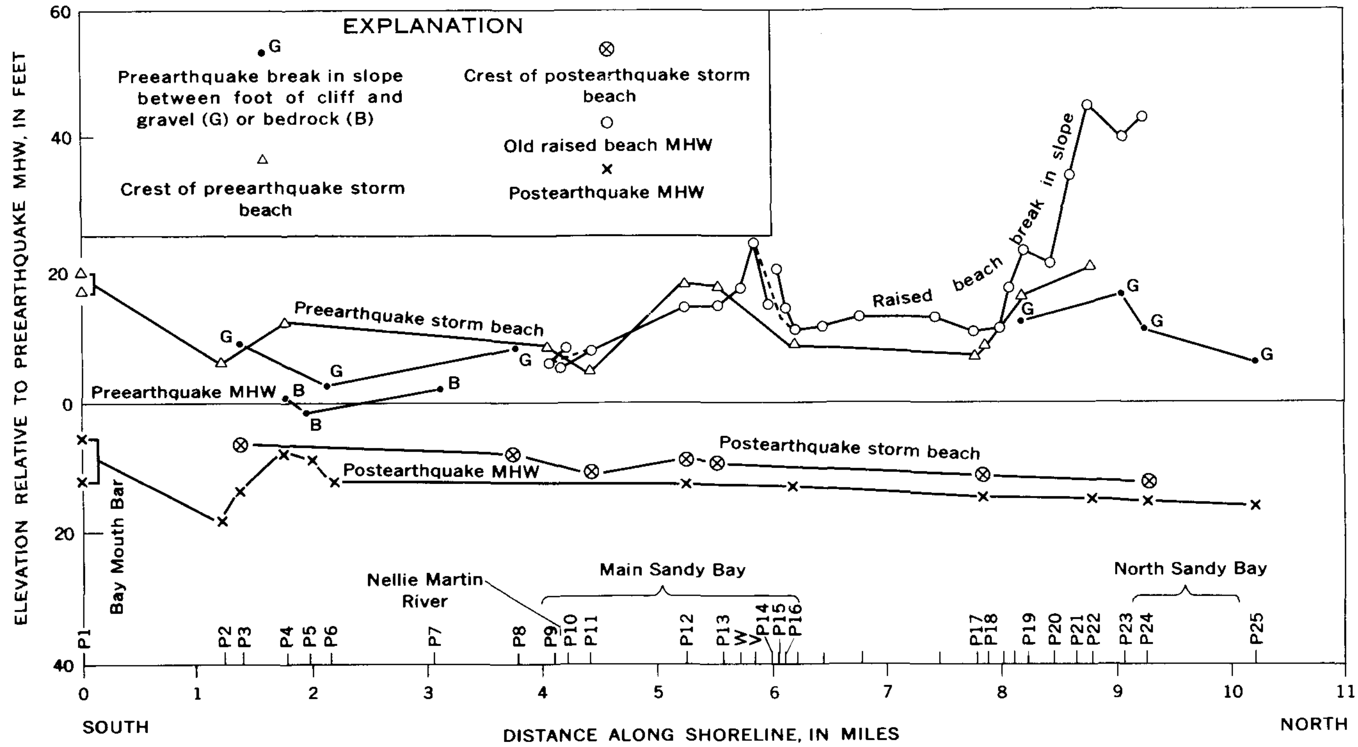
In summary, where a bedrock platform against the foot of a cliff is not covered by unconsolidated material, then the elevation of the wave-cut break in slope is considered to be the best available topographic indicator of mean high water, and the difference in elevation between two breaks in slope formed at different sea levels is an excellent measure of the difference between the two sea levels.

Where unconsolidated material lies at the foot of a seacliff, the shape of the underlying bedrock, the amount of gravel available, and random factors probably control the height to which the gravel is accumulated. At one extreme there is no gravel, and at the other extreme a full storm beach is formed against the base of the cliff. The elevation of the break in slope between the gravel and the cliff therefore varies from near high-water mark to full storm-beach elevations. Such breaks in slope are not accurate indicators of former water levels.

Storm-beach elevations increase with time to a maximum, which takes several years to reach. This maximum elevation is only constant for a given wave-exposure environment. Storm-beach elevations are therefore poor indicators of the absolute water level, but the difference in elevation between two mature parallel storm beaches should provide a good measure of the difference of elevation between the two sea levels at which they were formed.

TECTONIC DEFORMATION OF BEACHES

Elevations of beach features were measured at many localities over a 10-mile stretch of coastline in Patton Bay. The elevations of storm beaches, high-water marks, and breaks in slope were measured using the preearthquake high-water mark as a standard. In figure 25, each symbol represents a measured point, and the lines show the interpolated elevations of each feature between measured points, so each line shows the change of elevation of a beach feature along the coast, in relation to the preearthquake high-water mark. The figure shows the elevations of pre- and postearthquake shoreline features and of raised shoreline features and the variation of height of these features along the coastline of Patton Bay. All elevations were mea-



25.—Variation in differences of elevation among coastal features in Patton Bay.

sured relative to the preearthquake mean high water (which is taken as the top of the former barnacle zone). Postearthquake mean high water was estimated from U.S. Coast and Geodetic Survey tide tables. Where storm-beach crests and breaks in slope between beaches or rock platforms and the cliffs behind them are present, their elevations are shown.

Elevations are shown for features which correspond to pre- and postearthquake sea levels and also to a raised sea level. The elevation of the raised mean high water has been calculated in one of the following ways: (1) in Main Sandy Bay, as the difference in elevations between the crest of the preearthquake storm beach and the crest of the raised storm beach, or (2) elsewhere in Patton Bay, as the difference in elevation between the top of the preearthquake barnacle zone (preearthquake mean high water) and the break in slope between the raised beach platform and the steep slope behind it (raised mean high water). Figure 25 shows the tectonic deformation of the shoreline from the time of the earlier raised beach until just before the 1964 earthquake, and also the deformation caused by the 1964 earthquake. The deformation due to the 1964 earthquake is plotted in figure 17. These data are in general agreement with the smaller scale map by Plafker (1965, fig. 7), but they add detail for the Patton Bay area.

The evidence for the raised sea level consists of: (1) a continuous platform from just north of location P24 (fig. 17) to the north end of Main Sandy Bay; (2) a continuous raised storm-beach ridge across Main Sandy Bay; and (3) a platform at two points immediately south of the Nellie Martin River, close to the southern end of the storm beach (locs. P9 and P10, fig. 17). The continuity of the landward edge of the platform and of the raised storm beach shows that they represent a single sea level throughout their extent.

At Bay Mouth Bar an earlier storm beach within 2 feet of the elevation of the younger ridge has been preserved behind the younger ridge. This abandoned beach was presumably formed at a sea level almost at the preearthquake level, but no positive basis was found for correlating it with the raised beach deposits described above.

Figure 17 shows not only the distance over which the raised beach can be detected but also the width of the platform which has has not yet been destroyed. At the north end of the coastline studied, the raised beach platform gradually narrows and finally disappears in North Sandy Bay where the preearthquake seacliffs are several hundred feet high. The absence of the raised beach in this section can be attributed to the nature of the local rock, which offers very little resistance to marine erosion; a preearthquake rock platform would have been eroded by marine action and its surface covered with sand and gravel, in contrast to neighboring more resistant bare-rock platforms. The lack of resistance of the rock is also indicated by (1) the relatively gentle gradient of the cliffs (40°–50°), (2) the tendency of the cliffs to degrade by landslides, mudflows, and gullies rather than by rockfalls and talus accumulation (this tendency indicates a high clay content in the rock or in its primary weathering derivatives), and (3) the considerable amount of erosion of the cliffs since the earthquake (fig. 29, p. H35).

The coastline north of North Sandy Bay was not examined in any detail. The south side of Box Point (north of the area of fig. 17) has continuous cliffs 50 to 100 feet high; here low raised platforms apparently have not been preserved. The surface of the peninsula is relatively low and flat at altitudes of 100 to 200 feet, but there is too much relief for a raised platform. It is therefore tentatively assumed that, within the area studied, only in the central part of Patton Bay was wave attack sufficiently attenuated and the rock sufficiently resistant to allow the preservation of a raised beach platform.

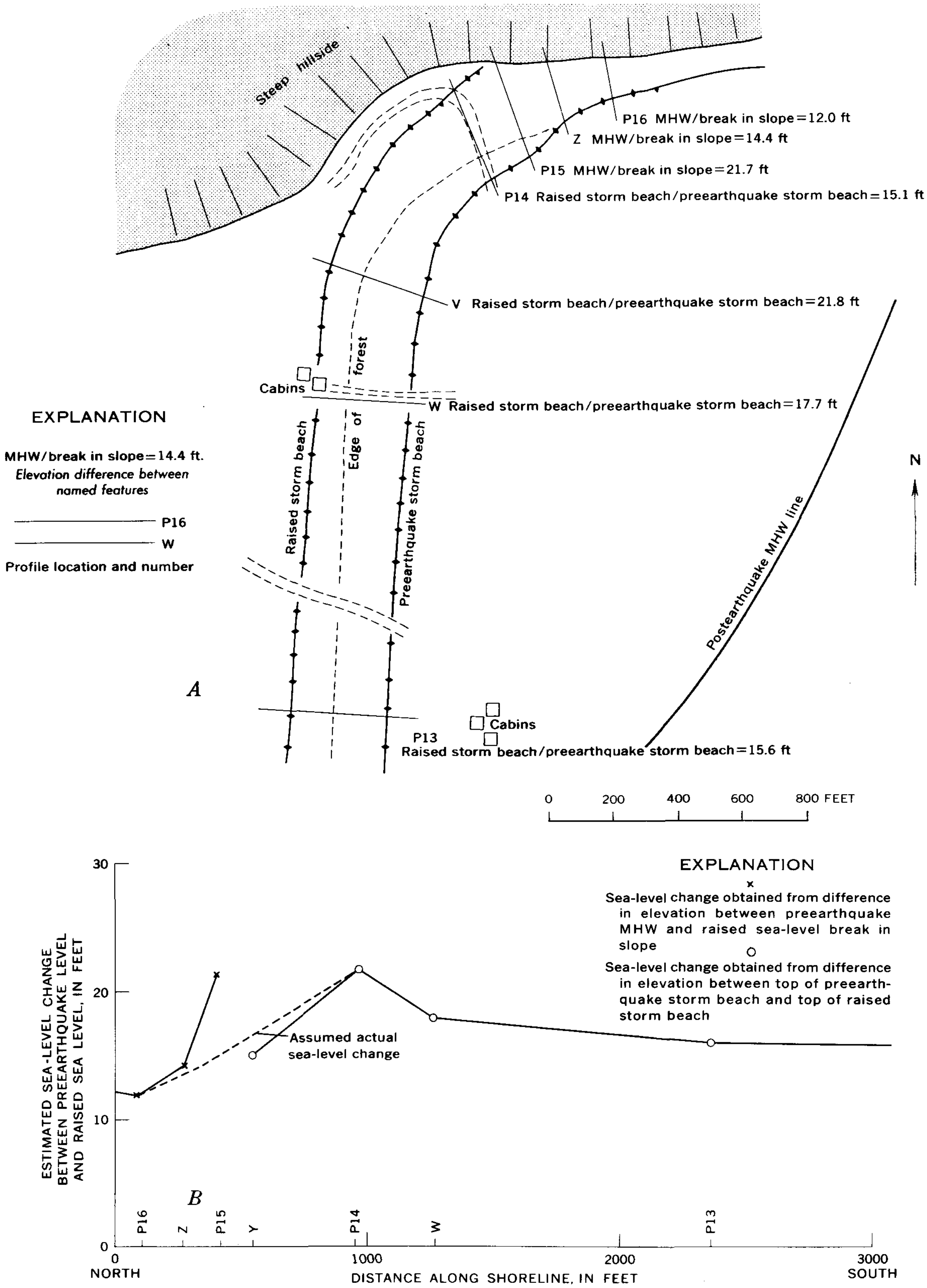
Since measurements of the relative heights of the raised beaches are based on the elevations of breaks in slope and of storm beaches, it must be recognized, for the reasons discussed above, that some measurements may be in error by 5 to 10 feet. Nevertheless, the amount of the warping is much greater than the amount of error inherent in the measurements, so the general pattern of the uplift can be accepted. For example, near the northern end of the raised beach (fig. 25), its elevation above the preearthquake mean high water rises from 10.2 feet (location P17) to 45 feet (profile P22) in a distance of 1 mile. This deformation is too great to be dismissed as an error of measurement, especially inasmuch as the pattern of warping in this area is confirmed by the elevations of breaks in slope both in gravel and in bedrock.

The relationship between the two measures of the uplift of the raised beaches—(1) the preearthquake barnacle line to the break in slope of the raised beach and (2) the preearthquake storm beach to the raised storm beach—can be studied where the storm beaches abut against steep slopes at each end of Main Sandy Bay. Figure 26 shows the relationship at the northern end of the bay and figure 27 shows the relationship at the southern end. Both figures seem to show that, in the area of overlap, the difference in height between the two storm beaches tends to decrease and the difference in height between the break in slope and the barnacle line tends to increase.

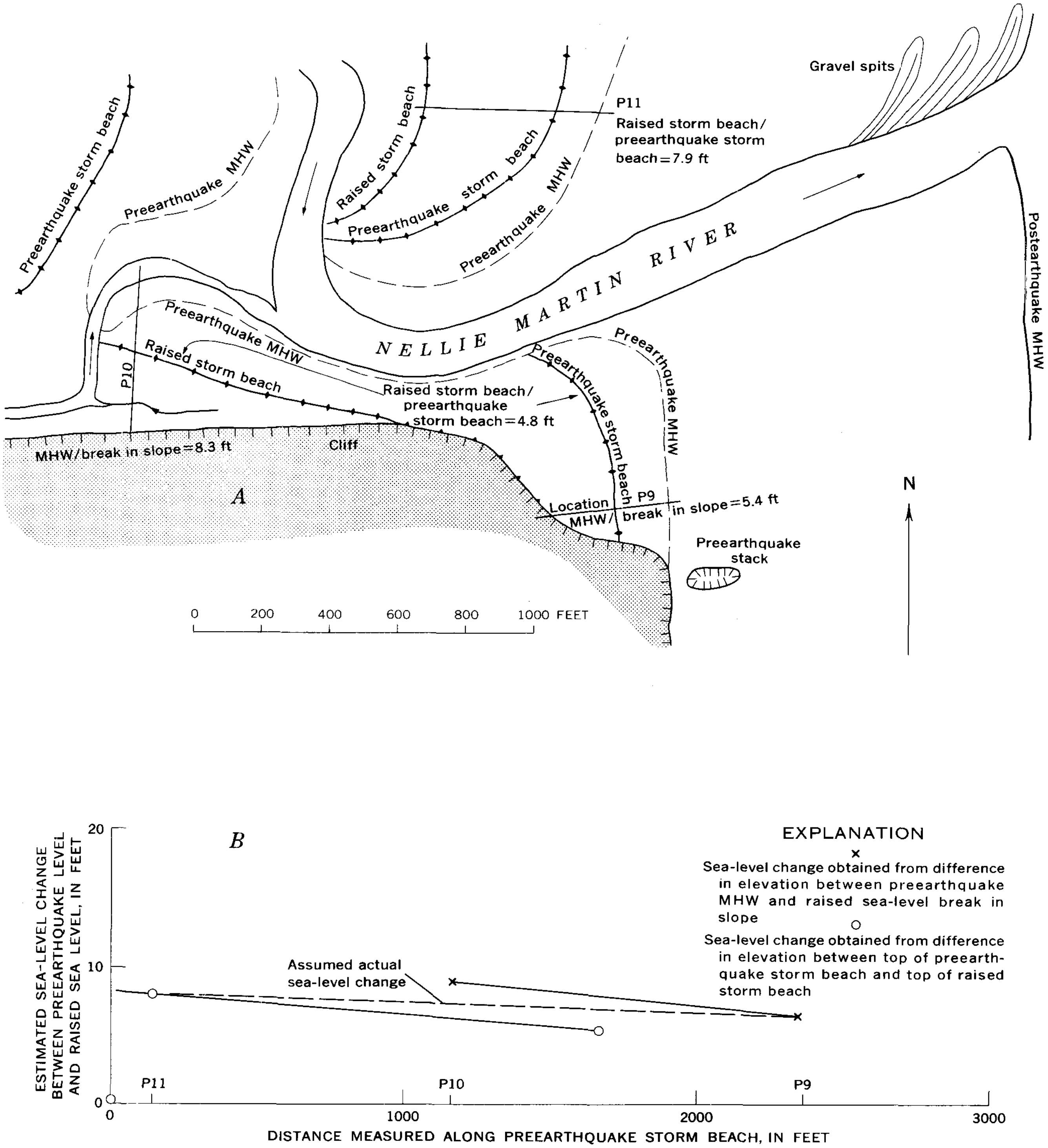
Inasmuch as the storm ridges are curved, the exposure to waves obviously will not be the same on each ridge at the points where a section line drawn at right angles to the ridge intersects them (figs. 26, 27). The landward, older ridge will have been closer to the hillside, and therefore less exposed to waves. As a result, the landward ridge is lower for a given sea level, and the difference in height between the ridges is smaller than the difference between the sea levels at which the ridges were formed. Elevations of breaks of slope, on the other hand, increase up the inlets (p. H27), so the differences in elevation between the break in slope on the raised platform and the preearthquake highwater mark will tend to become progressively larger than the differences between the two sea levels in the zone where the steep hillside is farther from the shore. It is concluded that the true change of sea level is greater than that shown by the storm beaches and less than that shown by the breaks of slope. An interpolated value of the sea-level change is shown by the broken lines in figures 25, 26*B*, and 27*B*.

South of the Nellie Martin River there are cliffs along the coast; no certain evidence of a raised beach was found, but with elevation differences as little as 5 feet above the preearthquake level, evidence would be very difficult to interpret in the absence of a continuous feature. Possible evidence exists in the second storm-beach ridge at Bay Mouth Bar and at isolated locations at the mouths of streams where the cliff is absent. However, the primary purpose of this investigation was not to locate possible raised beaches, but to use the evidence provided by a raised beach to support observations of changes to the beaches and platforms uplifted by the earthquake.

Comparison of the radiocarbon dates for all samples from Patton Bay beaches indicates that observed data can be explained most simply by two relative sea-level changes before the 1964 earthquake. Each change has left abandoned shoreline features at various elevations (fig. 25). Data from elsewhere in Prince William Sound suggest that the 1,000 years before the earthquake was characterized by slow submergence of the land (Plafker and Rubin, 1967), and the data from Montague Island are consistent with this



26.—**Relationship between measures of uplift of raised beaches at the north end of Main Sandy Bay.** ***A*****, Differences in elevation of raised break in slope, raised storm beach, preearthquake storm beach, and preearthquake mean high water.** ***B*****, Measurements of sea-level change.**



27.—Relationship between measures of uplift of raised beaches at the south end of Main Sandy Bay. *A*, Sketch showing difference in elevation of raised break in slope, raised storm beach, preearthquake storm beach, and preearthquake mean high water. *B*, Measurements of sea-level change.

Date	Sea level relative to preearthquake level (feet)	C-14 sample	Location of C-14 sample (fig. 17)	Basis for estimating mean high water
Sometime before 2070 B.P.	+8	W-1770	P23	Raised beach peat.
By 2070 B.P.	≤−6			
600 B.P.	≤−25	W-1776	P11	Nellie Martin peat.
Preearthquake A.D. 1964.	0	--------	P11	Barnacle line.
Postearthquake A.D. 1965.	−11	--------	P11	Estimate from U.S. Coast and Geodetic Survey tide tables.

interpretation. If the central value for each radiocarbon date is used, then the sequence of events near the mouth of the Nellie Martin River seems to have been as shown in the table.

These dates are consistent with relative sea-level changes at the same time and in the same direction on both sides of the island, although the amount of vertical movement varies considerably from place to place. No clear evidence of other sea-level changes was found in the area studied.

OBLITERATION OF BREAK IN SLOPE AT TOP OF BEACH

One immediate effect of the 1964 earthquake was to initiate many landslides along the coastline, but much mass wasting of the former cliffs has taken place since the earthquake. The break in slope at the top of the former beach is being rapidly obliterated by several processes—talus accumulation, landsliding, gullying, and vegetation growth. The uplift has prevented the sea from removing mass waste from the base of the cliffs, so all the material that has been eroded from the cliffs since the earthquake can be measured. From these measurements the rates at which the different slope processes were operating were determined.

OBLITERATION BY TALUS ACCUMULATION

At profile P3 (fig. 17), talus accumulations have already obscured 85 percent of the break of slope at the foot of a vertical cliff. The distribution of fallen debris is shown in figure 28. Some of this material no doubt fell at the time of the earthquake, but the major part probably has accumulated since the earthquake; the sound of falling stones was still frequently heard in 1965.

The volume of talus material at profile P3 is estimated to be about 210,000 cu ft. The cliff from which it is derived is about 110 feet high and 580 feet long. The talus accumulation therefore represents a mean cliff retreat of 2.4 feet in 15 months (to July 1965), assuming a 30-percent porosity in the talus. At this rate, if the total accumulation postdates the earthquake and if the observed maximum angle of rest of 22° is correct (fig. 28), the cliff will be completely regraded in 71 years. If any of the present accumulation occurred at the time of the earthquake, then the time required for the cliff at profile P3 to become completely covered by its own talus will be as follows:

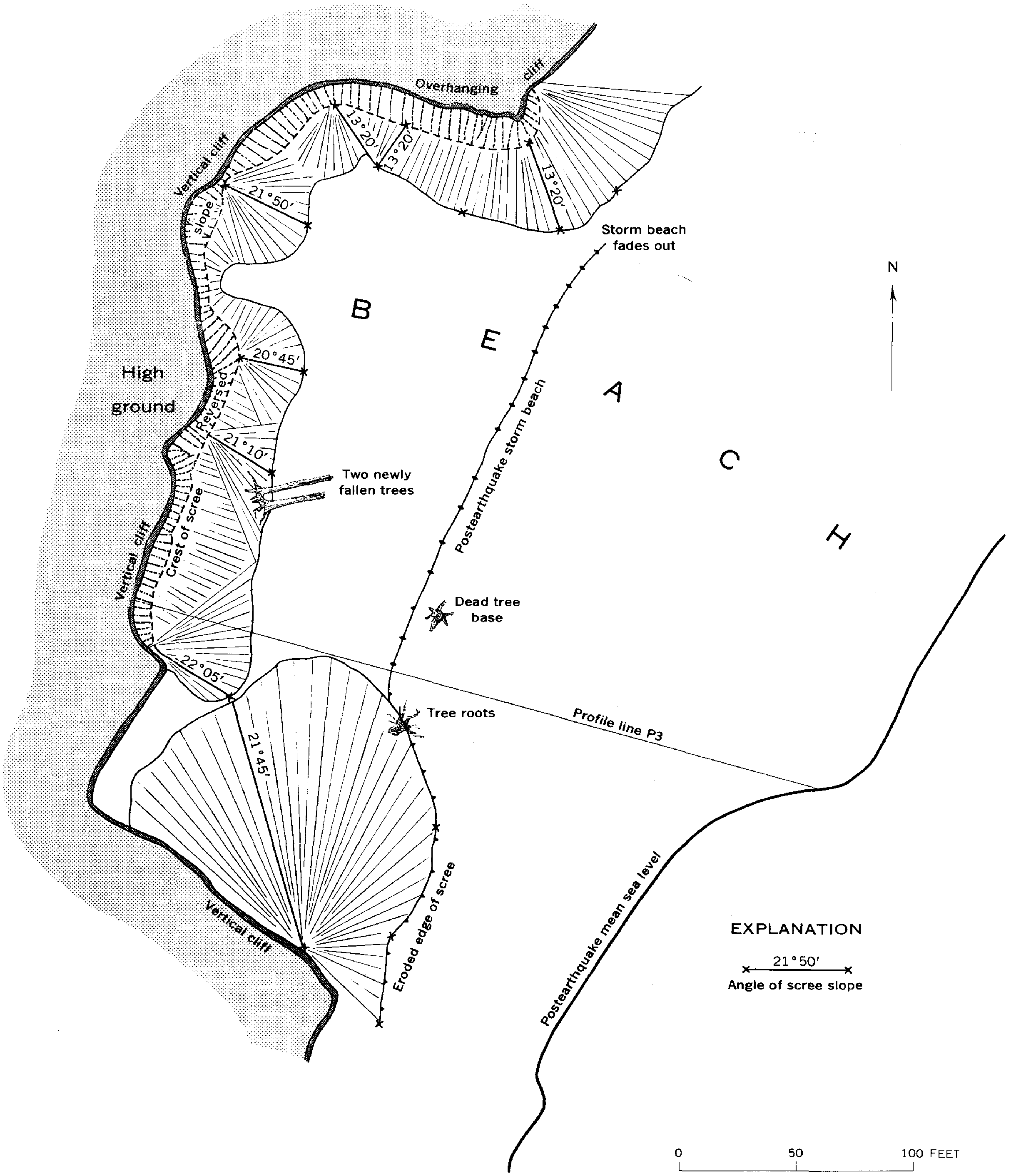
Accumulation at time of earthquake (percent)	*Time required for total degradation of cliff (years)*
0	71
40	117
60	176
80	350

Unless almost all of the accumulation occurred at the time of the earthquake—and the continued falling of material makes this seem improbable—then the cliff obviously will be completely obscured by its own talus in no more than a century or two.

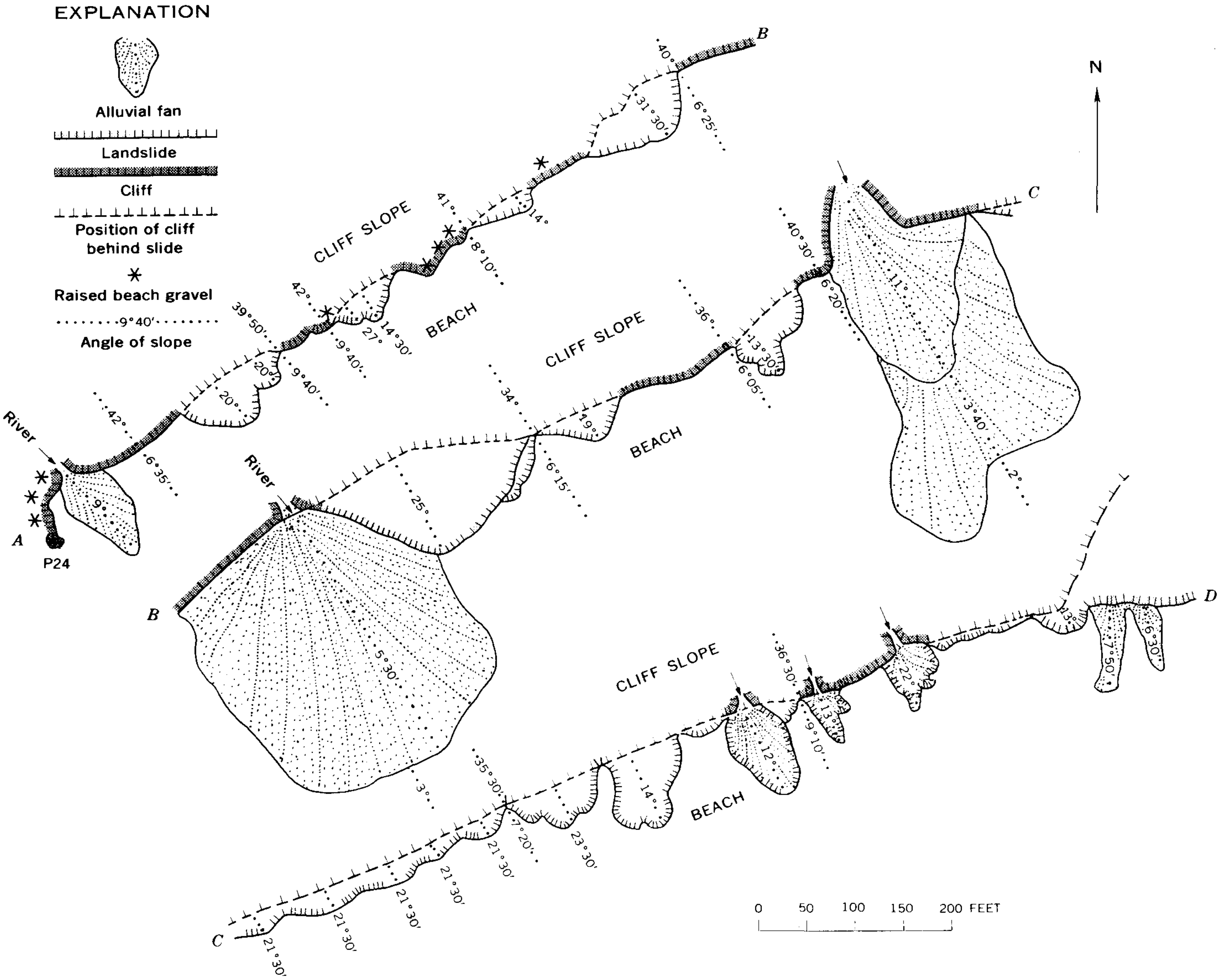
OBLITERATION BY LANDSLIDES AND GULLYING

In North Sandy Bay near profile P24 (fig. 17), landslides and gullying seem to have been more important than simple talus accumulation in obliterating the break in slope at the foot of the cliff. A horizontal distance of 2,940 feet of cliff was studied in detail, and the distribution of fallen material is shown in figure 29. In this total distance, 22 percent of the break in slope is virtually unobscured by debris, 17 percent of the break in slope is obscured by alluvial fan material, and 61 percent of the break in slope is obscured by landslide material. It was evident that much of the movement postdates the earthquake; the stream in one gully was very rapidly transporting debris while studies were being made in July 1965, and material in some of the slides was still very wet and sticky. The extent of postearthquake erosion is also shown by the very marked increase between May and August 1964 in fresh landslide scars that appeared on air photographs taken by the U.S. Coast and Geodetic Survey at those times. The estimated volumes of material transported are 670,000 cu ft in alluvial fans and 370,000 cu ft in landslides. The average height of the cliff is 400 feet; so the mean distance of cliff retreat since the earthquake has been 0.90 foot in 15 months.

Landslides are far more effective than gullying in degrading the cliff, because they act along its whole length. If landslides alone are taken into account, and if all the volume of landslide debris has accumulated since the



28.—Degrading of vertical cliffs and accumulation of talus at profile P3, Patton Bay.



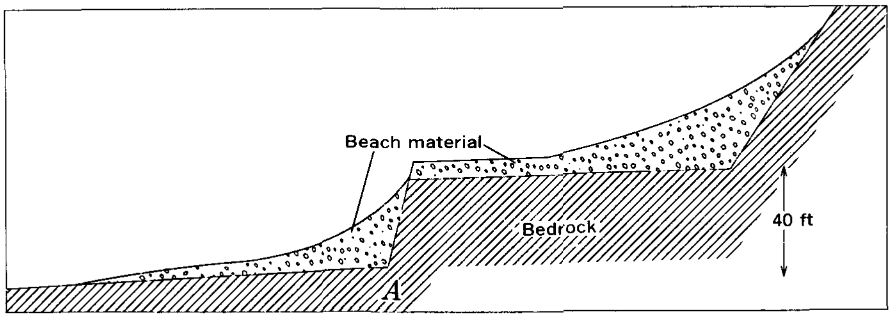
29.—Sketch showing degrading of cliff by landslides and gullying north of location P24 in North Sandy Bay.

earthquake, then the time required for total degradation of the cliff would be 2,300 years. However, the gradient of the vegetated steep slope behind the raised beach is rarely less than 38°, so it is unlikely that the process of degradation will proceed very far. At an early stage, vegetation will establish itself on the slope, and further degradation will be exceedingly slow, except perhaps locally along streams or gullies.

The striking difference in behavior between the resistant vertical cliff and the nonresistant 50°-slope may help to explain some apparent anomalies of raised-beach preservation in other areas. For example, rapid accumulation of talus may destroy topographic evidence of a raised beach within a short period, but it is this same talus cover which protects the raised beach from subsequent weathering and erosion and so over the long term preserves the beach in the stratigraphic section.

OBLITERATION BY VEGETATION

Where the upper limit of a raised beach is not a well-defined cliff, but a low step leading to another raised beach, the junction between the two beaches can be hidden by beach deposits and the growth of vegetation. It seems likely that this obliteration will occur extensively at the junction between the preearthquake beach and the raised beach, for example at the section shown in figure 19. Where the raised platform is well developed, the topographic step between the raised platform and the preearthquake beach is nearly always low, being nowhere more than a 10-foot step, even though



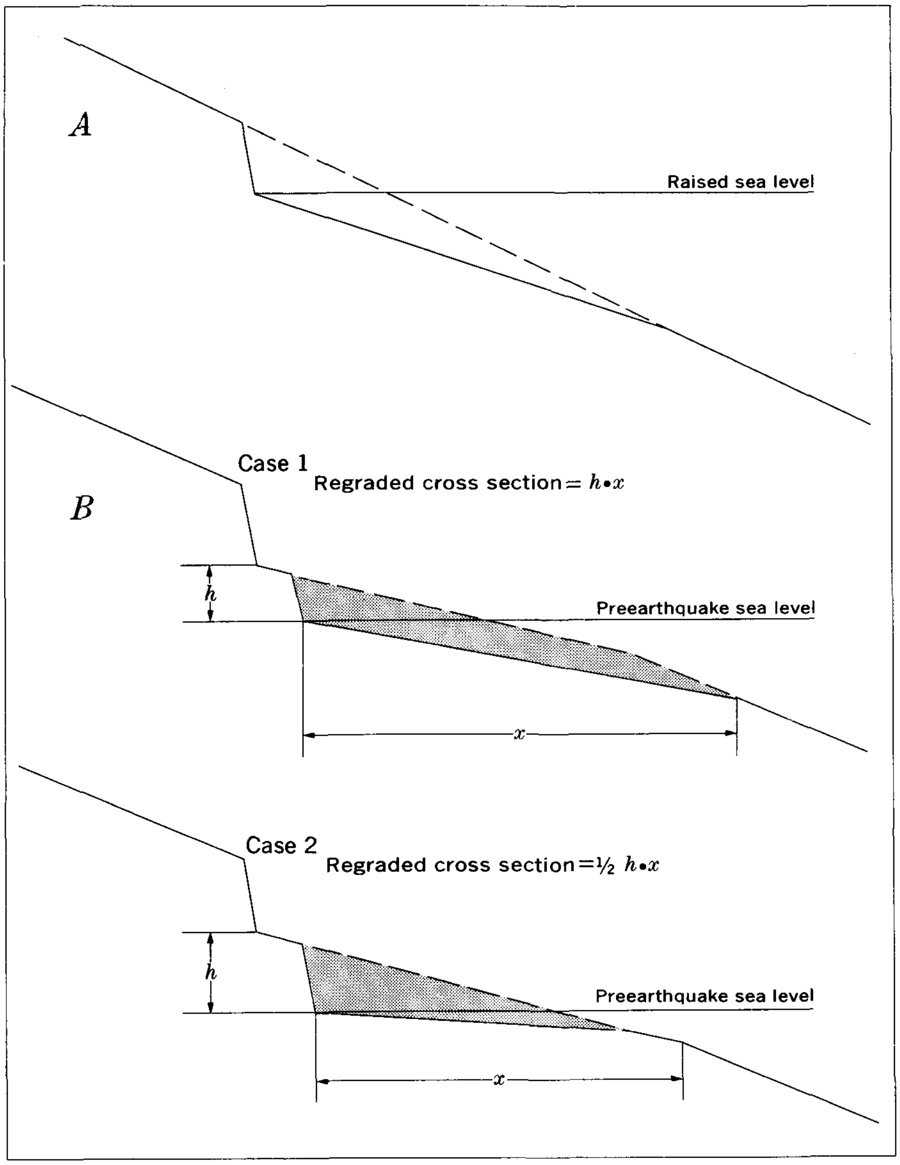
30.—Idealized profile through two raised beaches, with soil and vegetation omitted. Step between two rock platforms indicated by *A*.

The development of the preearthquake platform at the expense of the raised platform has been in progress only during the period that relative sea level was rising to preearthquake level, that is, since 600±200 B.P. or earlier. The probable sequence of sea levels and platform cutting is illustrated schematically in figure 31.

Figure 31*B* shows two of the possible extreme cases if the gradient the difference between the two sea levels may have been 40 feet or more. Figure 30 shows such an idealized profile, with soil and vegetation layers omitted. Most of the topographic step between the two rock platforms, as visualized in figure 29, is obscured by beach deposits, so the surface step is very low; a vegetation cover will soon make it extremely difficult to distinguish between the two surfaces by topographic form alone. Two valid criteria may still be used: (1) there should be a marked subsurface step in the bedrock at *A* of figure 31; and (2) if datable soil layers (for example peat) continue to form on the abandoned beaches of Montague Island as they have in the past, then an age discontinuity should occur in the lowest layers at *A*.

REGRADING OF ROCK PLATFORMS

The rock platforms accordant with the preearthquake sea level appear to have been formed by the regrading of a preexisting platform accordant with the raised-beach sea level. Regrading is well marked near location P21 (fig. 17) where many bedrock knobs on the upper part of the preearthquake platform stand as outliers, clearly accordant with the raised-beach platform, and in places surrounded by preearthquake beach deposits.



31.—Probable sequence of platform cutting in Patton Bay. *A*, Cutting of raised platform. *B*, Cutting of main platform during a time of rising sea level: case 1, main platform at same gradient as raised platform; case 2, main platform at lower gradient than raised platform.

dients of the raised platform, main platform, and initial land slope are allowed to vary; it may be seen that the cross-sectional areas removed in forming the main platform range from a minimum of $\frac{1}{2}hx$ to a maximum of hx, where h is the difference between the elevations of the raised beach notch and the preearthquake notch and x is the width of the main platform, measured at right angles to the shoreline.

TABLE 2.—*Relationship between change of sea level* (h), *width of platform regraded* (x), *and cross-sectional area of material removed* ($A \approx h \cdot x$)

Site (fig. 17)	h, in feet (raised level to preearthquake level)	x, in feet (base of raised beach to postearthquake mean high water)	$A \approx h \cdot x$), in square feet
P24	43	400	17, 200
P23	40	420	16, 800
P22	45	400	18, 000
P21	34	350	11, 900
P20	21	320	6, 500
P19	23	500	11, 500
	17	550	9, 200
P17	10	650	6, 500
	13	450	5, 800
P16	12	550	6, 600
P9	6	750	4, 500

In theory, the rate of regrading would depend on the resistance of the rock and on the thickness of rock to be removed and would decrease as either factor increases. Rock resistance is an unknown variable, but the fall of relative sea level from the raised level to the preearthquake level is a measure of the thickness of the material to be removed; it is compared with platform width in table 2.

Figure 32 shows the relationship of the cross-sectional area removed by regrading (A), to the depth through which regrading has had to work. It can be expressed by the best-fit relationship (fig. 32)

$$A=2{,}000+360h.$$

Dividing this expression through by h, the approximate width of the rock platform is given by the relationship

$$x=360+\frac{2{,}000}{h}.$$

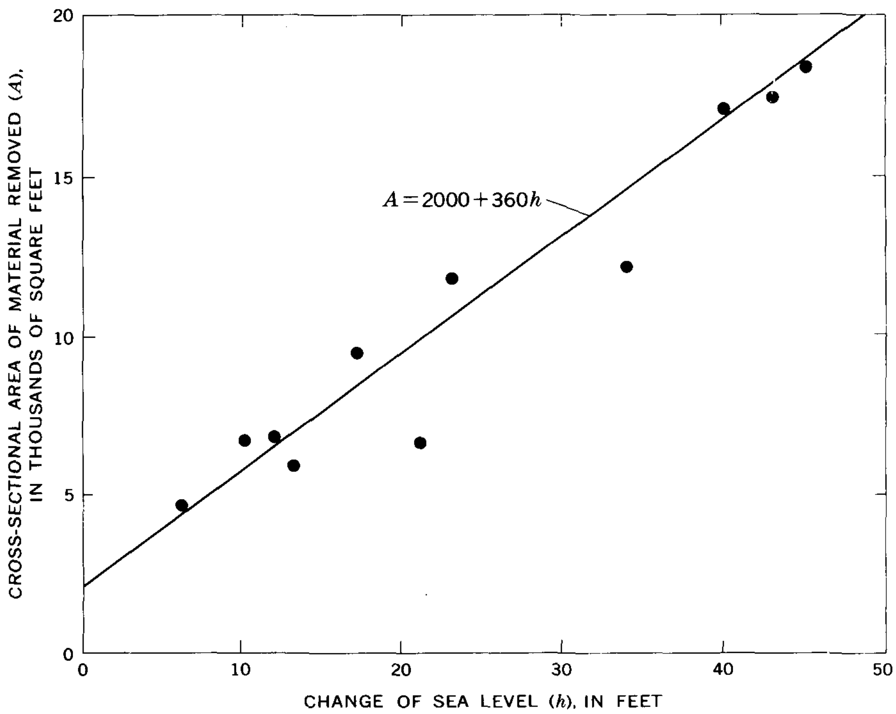
The mean rate of regrading can be found by dividing these two equations by the period during which relative sea level rose to the preearthquake level. Relative sea level was already rising at the time of formation of the low sea-level peat, dated at 600±200 B.P. (p. H33). The rise of relative sea level had probably begun before the deposition of sample W–1768 in MacLeod Harbor at 800±200 B.P. The period of platform cutting has therefore been taken as 1,000 years (T), and the estimated mean rate of removal of material has been

$$A/T=2.0+0.36h \text{ sq ft per year } \pm 30 \text{ percent.}$$

The mean rate of widening of the platform has similarly been

$$x/T=0.36+2.0/h \text{ ft per year } \pm 30 \text{ percent.}$$

These empirical equations can be compared with two limiting ideal situations, the first a constant rate of removal of material, and the second a constant rate of lateral cutting. The first situation typically would exist where the cliff is soft and is easily eroded, or the cliff is very high; the process is



32.—Relationship between change of sea level and cross-sectional area of material eroded during regrading of raised bedrock platform, Patton Bay. Points are plotted from data in table 2.

limited by the capacity of the sea to carry away the eroded material. In the notation of the equations above, this situation gives:

$$A/T = \text{constant},$$
$$x/T \propto 1/h.$$

The second situation is applicable to resistant cliffs where the limiting factor is the power of the waves to undercut the cliffs, and the transporting capacity of the water is more than adequate:

$$A/T \propto h$$
$$x/T = \text{constant}.$$

Figure 32 shows that the data for Patton Bay more closely fits the second situation. The intercept of the area for zero net change of elevation is smaller than the h-dependent term and may be interpreted as the amount of material removed from a coast where the sea returned to its original level and merely formed a steeper offshore platform. A sufficiently great relative lowering of sea level presumably would overload the transporting capacity of the system and thus would approximate the first ideal situation, but so great a change was not observed in the field.

In Patton Bay the resistance of the rocks seems to have been the main control of the erosion and regrading of the rock platform to its preearthquake level, and the transporting capacity of the waves has been more than sufficient to remove the debris produced.

REGRADING OF SAND AND GRAVEL

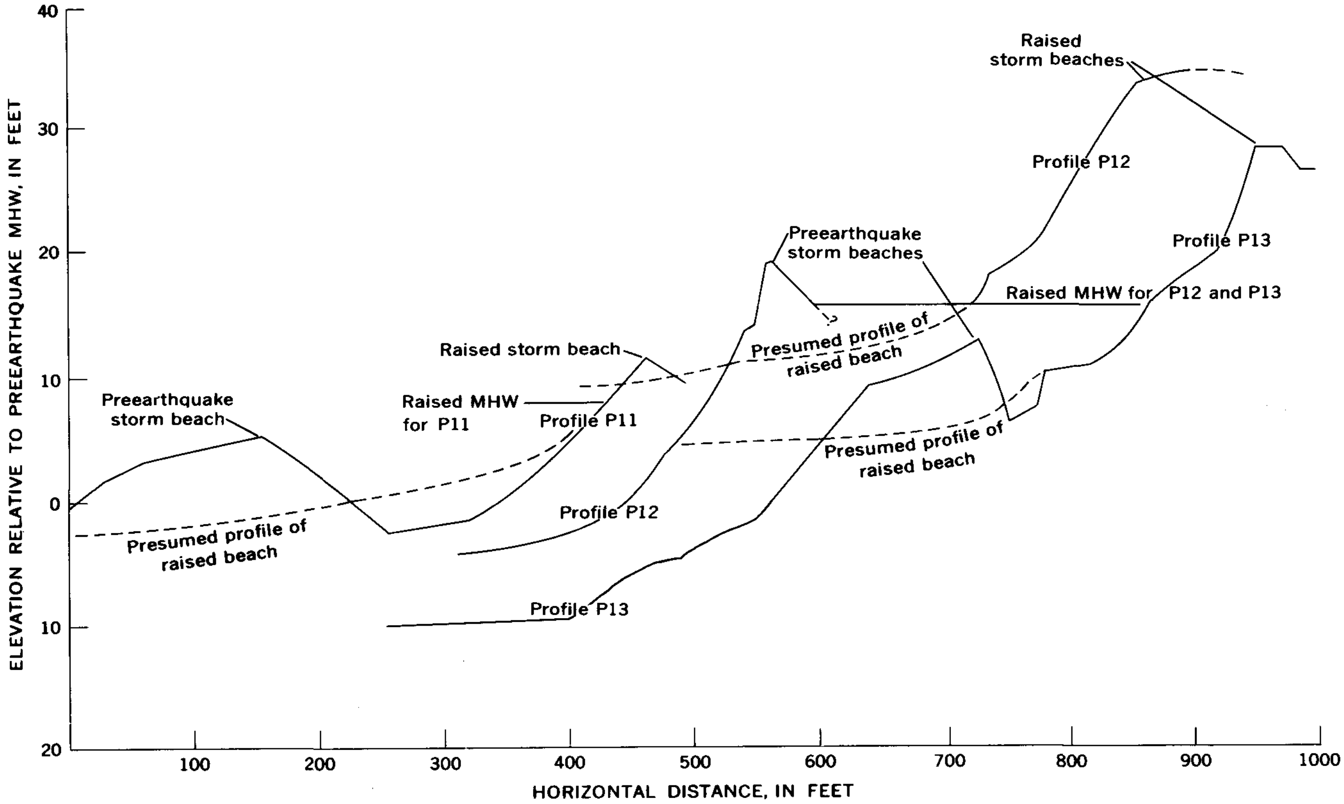
Where the beach is in unconsolidated materials, the whole area of the beach below the crest of the newly forming storm beach has apparently been regraded by marine processes. It seems probable that the vertical range of this regrading will increase as storms of greater magnitude build a higher storm-beach ridge.

The upper limit of the future marine erosion and deposition of sand and gravel seems to be limited by the final storm-beach height, which should approximate the preearthquake storm-beach height, that is, 8 to 16 feet above high-water level. The horizontal extent of beach resorting up the bay is less clear. On a rock platform, the shape of the platform in cross section should determine the position of the storm beach. The beach profiles (fig. 24) indicate that the base of the storm ridge forms from 4 to 6 feet below high-water level, or at about mean sea level.

Main Sandy Bay is the best example of a beach of unconsolidated material, where waves could change the position of mean sea level in a horizontal direction by erosion or deposition. Figure 33 shows the presumed form of the raised beach before the preearthquake sea level and the position of the preearthquake storm beach for profiles P11, P12, and P13. Whether the effect of the sea-level change was erosional (P12 and P13) or depositional (P11), the position of the new storm beach is immediately at the foot of the old one, the horizontal distance between the two storm beaches being almost constant at 300 feet despite variations in the amount of net relative sea-level change. The postearthquake storm beaches in profiles P1 and P24 also are about 300 feet from the old storm ridges.

In North Sandy Bay an estimate was made of the percentage of the area regraded by fluvial action (even if later covered with alluvial fan deposits). This estimate was made for the same section of beach as the estimates of obliteration of the break in slope by landslides and gullying (p. H33). At the top of the beach, close to the cliff, 24 percent of the area had been regraded by fluvial action; just above the postearthquake storm beach 33 percent had been regraded. The extent of regrading depends on the position of the streams flowing off the hillsides, and the lateral extent of regrading will increase as the streams change course on the beach. The streams flowing into North Sandy Bay are small and drain very steep slopes. Regrading consists mainly of erosion of the beach materials, followed by deposition of hillside material as alluvial fans, often above the original beach level.

The overall course of regrading is illustrated by the beaches in MacLeod Harbor (profiles 3, 5, 8, fig. 13) and North Sandy Bay, and by the raised-beach section in Patton Bay (profiles P19, P23, fig. 17). After a raised beach or rock platform that is backed by a steep hillside has been elevated, zones of marine and fluvial action are divided by a storm-beach ridge. On a beach, the storm ridge is formed about 300 feet away from the old storm ridge. On a rock platform, the storm ridge is formed with its seaward base at about mean sea level. This ridge remains until it is demolished by wave erosion of the rock, after which there will only be a gravel beach banked against a bedrock step as at A in figure 31. Below the storm-beach ridge a new equilibrium beach or bedrock platform will be formed by marine action; the exact profile of this ridge cannot be predicted (fig. 24). Above the storm beach, infilling by alluvial fan or colluvial material alternates with the development of soil and peat until the streams are able to establish semipermanent courses and equilibrium gradients through the new storm beach.



33.—Position of preearthquake storm beaches in relation to the presumed original surface of the raised beach. Profiles of original surface were obtained by transposing present profiles to raised high-water marks.

CONCLUSIONS

Although most of the results of studies on Montague Island have only local relevance, some conclusions may be drawn which have a wider application. In MacLeod Harbor, the 33-foot uplift of the beach deposits and their subsequent erosion provided a rare opportunity to observe and measure the actual effect of a change in sea level on fluvial erosion. It was found that the streams flowing in sand and silt immediately cut a course of almost uniform gradient between the old and the new sea levels. This phase of the readjustment occurred within a matter of days for all streams flowing over unconsolidated materials and having low-flow discharges greater than about 1 cfs. Whether the long-term adjustment of the long profile would be aggradation or headward recession of a knickpoint could not be determined, but it is here suggested that the long-term adjustment may depend on whether the new course is less steep or steeper than the course above the former sea level. Thus, in mountainous regions, where the streams have relatively steep gradients, a fall in sea level might lead to either aggradation or erosion; there would not necessarily be a relation between changes of sea level and the presence of knickpoints. In lowland regions, however, where the lower courses of rivers have very low gradients, it is anticipated that knickpoints will develop and migrate upstream following a relative fall in sea level.

In addition to being an example of a dissected uplifted surface, the head of MacLeod Harbor is also an example of an area where fluvial processes are occurring so fast that changes may be readily studied. A mean rate of net sediment removal of 1 cfs in a stream of low-flow discharge of 300 cfs may be unusually rapid, but the fact that the erosion rate varies in direct proportion to low-flow discharge may have wide application, inasmuch as it shows the effect of discharge on the rate of erosion and the transport of uniform materials.

The relationships between the preearthquake sea level and the beach features associated with it have been very clearly exposed by the uplift. Measurements over about 10 miles of coastline show that the height of storm beaches varies from 6 to 22 feet above high-water level, the mean being 12 feet. Where gravel was banked against bedrock cliffs at the top of a beach, the height of the break in slope between the gravel beach and the cliff varied from 2 to 15 feet above high-water level, the mean being 9 feet. Therefore, neither of these features can be depended upon to indicate the relative sea level within less than 5 to 10 feet unless the exposure and configuration of the shoreline is known in unusually great detail.

The area studied was not large enough to serve as the basis for general deductions about where raised beaches are best preserved, but some observations were made. Gently sloping shores in unconsolidated materials seem well suited to preservation of raised storm beaches. On rocky shores, such factors as rock resistance and wave exposure tend to influence the rates of formation and destruction of rock platforms in the same way, so there is no reason to assume that these factors influence preservation. In Patton Bay, however, the raised beach platform has in fact been best preserved where the rock seems to be most resistant (although not highly resistant), and the platform narrows dramatically and disappears where the rock becomes less resistant. Other factors which should theoretically aid beach preservation are (1) a low gradient on the earlier shoreline, which favors more rapid formation of a platform at the old sea level, and (2) a large uplift where the lowered relative sea level exposed a larger area of platform and thus delayed its total destruction. The influence of the slope of the former coast could not be clearly seen in the field, but broader platform remnants do seem to be associated with some of the areas of greater uplift. In Patton Bay, therefore, rock platforms seem to be best preserved in an area (1) slightly protected from wave attack, (2) where the rock is moderately resistant and (3) where relatively great uplift has followed the formation of the platform.

The 1964 deformation has provided information about the way in which a raised beach is modified after uplift. Rates of widening of a rock platform by erosion have been calculated and have been shown to be roughly constant at 0.49 feet per year for cliffs of moderate height. The rate of recession of 0.7–2.0 feet per year for cliffs on Montague Island should indicate the order of magnitude of recession in other areas. Storm beaches form on old platforms and their bases are roughly at mean sea level; they are converted to gravel banks against a step as the platform is eroded by the waves.

Beaches of sand and gravel (other than bay-head deposits) seem to be regarded by marine action within 1 year, except for the upper part of the beach, where a new storm beach forms about 300 feet seaward of the previous one, provided the sea level drops at least 5 feet. The new storm beach may take many years to attain its full height. Behind a storm beach, there is a rapid fluvial regrading of beach material near the mouths of streams and a general slow infilling by alternate layers of vegetation and alluvial and colluvial deposits. Larger streams cut through the storm beach and drain part of the area behind it. Within a period estimated to be about 5 years, areas of extensive deposits, such as the bay-head sediments of MacLeod Harbor, will be regraded mainly by fluvial action before a new storm beach develops.

The break in slope between the top of the beach and a cliff is almost completely obscured after a few years unless the cliffs are particularly resistant. Talus and landslides may degrade some or all of the cliff, unless the rocks weather to fine-grained material and are stabilized at an early stage by vegetation. Where the top of the beach is backed by another older beach, quite large (40 ft) differences in sea level were found to produce only low (5 ft) topographic steps. These steps could be so obscured by unconsolidated beach deposits and vegetation that a series of raised beach levels might appear on the surface to be one continuous slope.

Raised beaches formed at a known date and having their initial profiles almost perfectly preserved are very rare, but they did exist on Montague Island immediately after the 1964 earthquake. Thus an ideal opportunity was provided to study the rates and processes of destruction of raised beaches. Such studies of presently forming geomorphic features are an excellent basis for understanding and identifying similar features formed in the past.

REFERENCES CITED

Barrett, P. J., 1966, Effects of the 1964 Alaska earthquake on some shallow-water sediments in Prince William Sound, southeast Alaska: Jour. Sed. Petrology, v. 36, no. 4, p. 992–1006.

Plafker, George, 1965, Tectonic deformation associated with the 1964 Alaska earthquake: Science, v. 148, no. 3678, p. 1675–1687.

Plafker, George, 1967, Surface faults on Montague Island associated with the 1964 Alaska earthquake: U.S. Geol. Survey Prof. Paper 543–G, 42 p.

Plafker, George, and MacNeil, F. S., 1966, Stratigraphic significance of Tertiary fossils from the Orca Group in the Prince William Sound region, Alaska, *in* Geological Survey research 1966: U.S. Geol. Survey Prof. Paper 550-B, p. B62–B68.

Plafker, George, and Mayo, L. R., 1965, Tectonic deformation, subaqueous slides, and destructive waves associated with the Alaska March 27, 1964, earthquake—an interim geologic evaluation: U.S. Geol. Survey open file report, 21 p.

Plafker, George, and Rubin, Meyer, 1967, Vertical tectonic displacements in south-central Alaska during and prior to the great 1964 earthquake: Jour. Geosciences [Osaka City Univ.], v. 10, art. 1–7, p. 1–14.

Sparks, B. W. 1960, Geomorphology: London, England, Longmans, Green and Co., Ltd., 371 p.

Wolman, M. G., 1954, A method of sampling coarse river-bed material: Am. Geophys. Union Trans., v. 35, no. 6, p. 951–956.

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