[AMERICAN JOURNAL OF SCIENCE, VOL. 273, JANUARY, 1973, P. 48-64]

FORMATION, STRUCTURE, AND GEOMORPHIC INFLUENCE OF LAKE SUPERIOR ICEFOOTS

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ABSTRACT. The long, narrow, continuous ridges of grounded ice, separated by broad areas of low-relief ice, that parallel much of the shoreline of Lake Superior and other Great Lakes are called icefoots. They form from fields of small fragments of lake ice and ball ice which have been heaped up by storm waves and grounded. Once stabilized, an icefoot can continue to grow by wave overwash and wave spray. A sequence of up to four ridges of successively increasing size generally forms during the winter, and the outermost ridge is subject to wave action, both erosional and depositional, most of the time.

Seismic refraction and drilling show that the subsurface configuration of the ridges is often irregular, while that of the interridge ice is smoother. The total thickness of an ice ridge may exceed 11 m. Icefoots are best developed off sandy, equilibrium shorelines with high exposure to storm waves. They may form on sandbars, but sandbars are also common in interridge areas. In plan view, the lakeward edge of an icefoot is wavy, and the amplitude/wavelength ratio is greater for icefoots off bedrock cliff shores than for those off sandy, equilibrium shores. No simple relationship exists between wavelengths of the ridges and wavelengths of formative waves.

The icefoots are of geomorphic importance, as they protect the shoreline from the high-energy waves of winter and spring, thereby reducing rates of erosion which otherwise might be expected. The net sediment transport along the southeastern shore of Lake Superior is on the order of 11 times greater than that along selected segments of the Arctic shoreline where icefoot formation is known to occur but is much less than that along comparable ice-free coasts.

INTRODUCTION

During March 1969, January and March 1970, and January 1971, field investigations were undertaken to determine the formation, structure, and geomorphic influence of icefoots in the vicinity of Grand Marais, Mich., located on the south shore of Lake Superior, approximately 130 km east of Marquette, Mich. (fig. 1). The icefoots observed along Lake Superior are composed of large masses of grounded ice that form long, narrow, essentially continuous ridges parallel to the shoreline which are separated by relatively broad areas of low-relief ice (pl. 1). Along shorelines with gently sloping offshore relief, several ice ridges usually form at various distances from the shore during a single winter. The size of these features generally increases with distance from the shore, though their morphology remains fairly constant. The smallest ridge forms in the lower swash zone and is usually less than 1 m high, whereas the largest ridge may form as far as 200 m from shore and reach a height of 7 m above the lake. In plan view, the lakeward edge of these ridges is wavy. The wave length and amplitude generally in-

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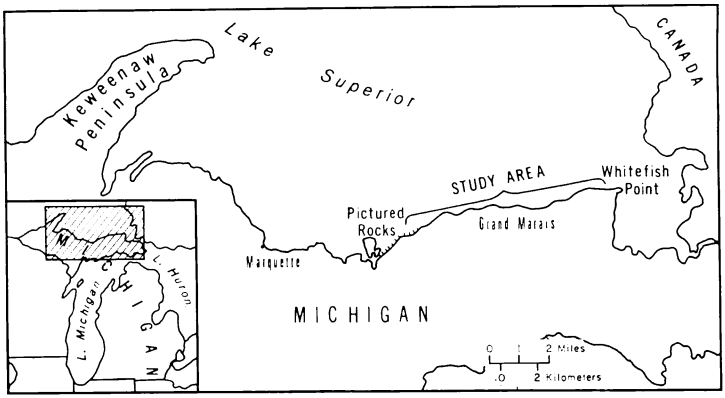


Fig. 1. The study area, southeastern Lake Superior. Most of the area is sandy, low-relief shoreline, backed locally by high backshore slopes.

crease with the size of the ridge. In transverse profile ice ridges have gentle landward slopes, usually less than 25°, and steep lakeward slopes, often reaching 90° or more. The composition of the ice ridges of Lake Superior icefoots is very heterogeneous, consisting of fragments of lake ice (fast ice), ball ice (brash and slush), glaze ice, fine granular ice (resembling névé), and sand and pebble-sized sediments. Although icefoots at many locations in the Great Lakes have been observed and described by several investigators (Zumberge and Wilson, 1953; Marshall, 1966; Bajorunas and Duane, 1967), only Bryan and Marcus (1972) have measured any of their specific parameters. Their study focused on the composition of ice ridges near Grand Marais, Mich.

ICE RIDGE FORMATION

Field observations and information from local inhabitants indicate that the large, offshore icefoots usually form in a two to three week period in January of each year. Several conditions are necessary for their formation: (1) sub-freezing air temperature, (2) open lake water, (3) storm waves, and (4) a supply of ice fragments. Observations of the processes of ice ridge formation show that during sub-freezing weather fields of small fragments of lake ice (usually less than 3 m in diameter) are forced by waves into shallow coastal water where they become compacted and grounded. Subsequent action by storm waves results in a heaping up of this ice along its lakeward edge into a ridge parallel to the shore. The ridge grows from (1) overwash by waves, which transports balls of slush ice onto it, and (2) wave spray. As build-up continues, the ridge develops a glazed, pimpled surface. Wave overwash was measured as far as 26 m down the landward slope of one 7 m high (above lake level) ridge. Ice balls as large as 0.75 m in diameter were deposited by waves near the crest of the ridge (pl. 1-B). A vertical growth in excess of a meter in one ridge occurred overnight.

PLATE 1

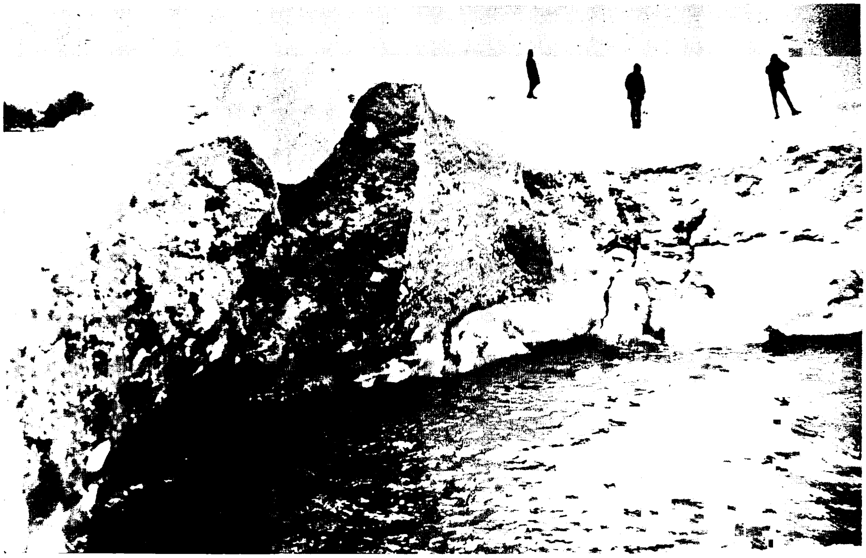
A.

B.

A. A newly-forced ice ridge, 5 to 6 m thick, about 150 m off shore. Landward beyond this ice ridge and the broad area of low-relief ice are two smaller ice ridges.

B. A large piece of ball ice deposited by waves 7 m above water level near the crest of an ice ridge.

PLATE 2



A.

B.

A. The lakeward face of an ice ridge after failure caused by wave erosion. Tonal differences in the ice are due to variations in sediment content. Note on the right the boundary between the high- and low-sediment portions of the ridge.

B. A view of the landward side of an active ice ridge showing the volcanic mountain-like morphology of the cones of deposition.

The first ice ridge of winter appears in the lower swash zone in December. The early growth of this ridge is characterized by interbedding of sand, pebbles, and ice. As the ridge grows—mainly from wave overwash and wave spray—the sediment content generally decreases. Rarely does this ridge exceed 1 m in height. Morphologically, it is a small replica of the larger ridges which form later off shore. The formation of the second ice ridge follows the accumulation and compaction of ball ice and lake ice fragments into a low relief mass of grounded ice which extends lakeward 15 to 35 m from the shore. Along the edge of this mass of ice the second ridge forms. Near Grand Marais in the past three years, this ridge formed in 1 to 2 m of water and reached heights of 1.5 to 3 m. The formation of the third, and usually the outermost, ridge follows the same sequence as that of the second ridge. The only major difference is one of scale; namely, the mass of grounded, low relief ice extends 75 to 100 m beyond the second ridge, and the third ridge reaches heights of 7 m above the water. Near Grand Marais, this ridge formed in 2.4 to 3.0 m of water.

Once formed, the outermost ridge may come into intermittent contact with open water during the winter. Wave action may modify it by deposition from overwash and spray or by erosion at the waterline. The distribution of wave erosion and deposition is controlled in part by the promontories and reentrants in the lakeward edge of the ridge. The initial locations of these features are probably related to natural concentrations of wave energy, but there is no simple relationship with the period of incoming waves. Observations show that promontories receive the severest erosion by waves. This is evidenced in several ways: (1) undercut profiles; (2) concentration of vertical cracks in the ice, some of which are as large as 0.75 m wide and 7 m deep; and (3) propensity to fail during periods of heavy wave action. The size of the blocks of ice that break off promontories increases with the size of the ridge. In January of 1971, a single failure from one large ridge removed 5000 m^3 of ice. The cavity created extended approximately 15 m into the ridge, forming a new reentrant (pl. 2-A).

Reentrants in the ice ridge are characterized by deposition of ice. This is evidenced in irregular and gently lakeward-sloping profiles covered with ice balls and in cone-shaped depositional features that form in the heads of reentrants. The cone-shaped features develop at the ends of "shoots" which extend 6 to 10 m up the ridge (fig. 2). Waves that converge on the reentrant heads are forced up shoots, carrying ball ice and producing overwash and spray, all of which build the cones to heights 1 to 4 m above the general ridge crest. The longitudinal profile of the ridge crest develops the aspect of a miniature volcanic mountain chain (pl. 2-B).

The mass of interridge ice between the second and third ridges is frequently rougher than that between the first and second ridges. Pieces of ball ice are generally larger—the largest measuring 0.5 m in diameter

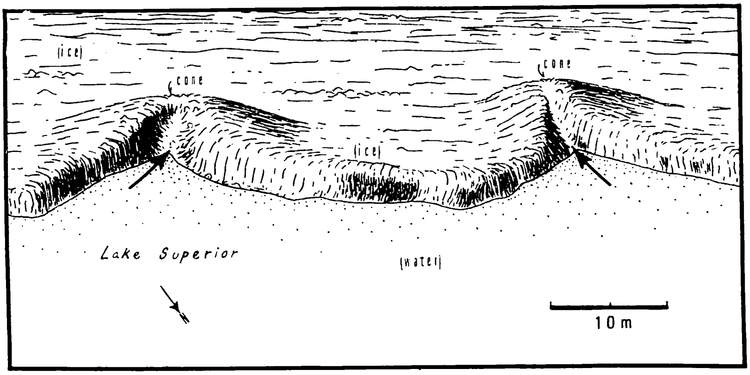


Fig. 2. An illustration of an offshore ice ridge, showing promontories, reentrants, and cones of deposition. The arrows indicate the paths of wave movement up the "shoots."

—and locally are arranged in several short ridges, less than 1 m high, resembling pressure ridges. These ridges apparently form during the accumulation and compaction of the ice mass prior to the formation of the outer ridge. The composition of the interridge ice appears to be very similar to that of the ice ridges, except it lacks glaze ice and associated concentrations of sediment. This suggests that interridge ice accumulates rapidly and therefore receives very little spray and overwash from wave action. Indeed, observations indicate that interridge ice may grow lakeward more than 35 m overnight. Based on their observations of interridge ice on Lake Michigan, Zumberge and Wilson (1953, fig. 2, p. 203) indicated that this ice was not grounded. However, sample drillings at several locations along southeastern Lake Superior indicate that not only is interridge ice usually grounded, but by late winter the ice for several tens of meters beyond the outer ridge is also often grounded.

STRUCTURE

To determine the subsurface structure of icefoots, two techniques were employed: (1) seismic profiling and drilling in the areas of ridge ice, and (2) drilling in the areas of interridge ice.

In March 1969, a portable seismic refraction unit was used to profile the outermost ice ridge near Grand Marais. Twelve geophones were laid out along the crest of the ridge. Blasting caps placed on the surface of the ice served as an adequate energy source. The ice on this portion of the ridge contained a number of cracks and voids, which in turn often made proper energy transmission impossible, and consequently only two areas were found where the refraction method could be successfully employed. For each of these places the times of the first arrivals were recovered from the photographic time records and were plotted in the conventional time versus distance fashion, that is, distance of the geophones from the blasting cap on one axis and arrival times of the

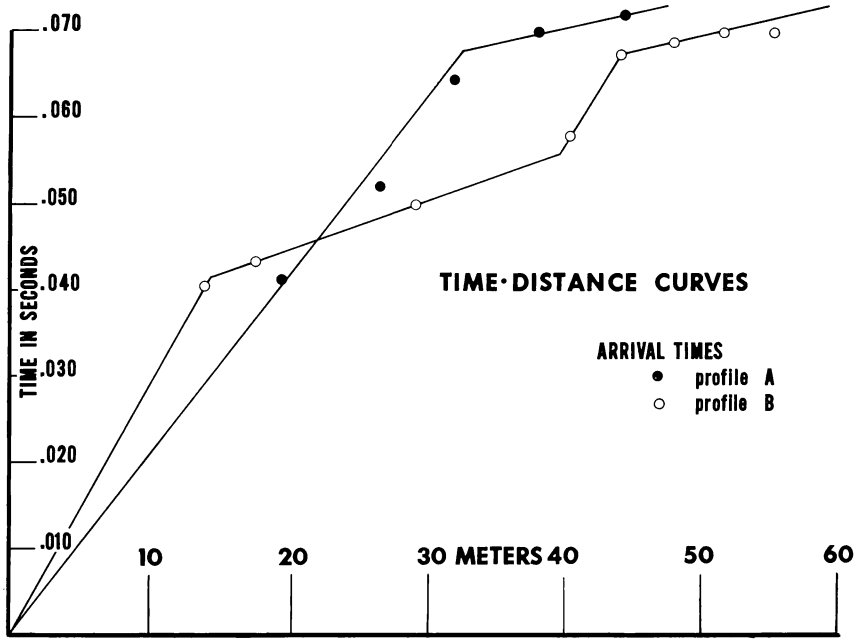


Fig. 3. The time-distance curves from each of the seismic profiles through a large ice ridge.

seismic energy on the other (fig. 3). Several different curves can be drawn through the data which will give various interpretations of levels of stratigraphic changes. However, through careful study of lead-line water depth measurements, all but one of the possible interpretations can be eliminated. Water depths 50 m lakeward of the ridge were on the order of 3.5 m and decreased slowly toward the ridge. The depth of water at the immediate outer edge of the ridge, though unobtainable at the time, was found to be a minimum of 3 m. By combining this minimum depth and the height of the geophone spread above the water level, it was calculated that the ridge thickness must be a minimum of about 7 m. Therefore, only those interpretations of the seismic data consistent with this boundary condition need to be considered. The data also yield information on the configuration of the lower boundary of the ice ridge.

From the time-distance graphs, the velocities and intercept times were obtained for the two layers involved. Because of warm weather immediately preceding and during the time of the study and the general inhomogeneous composition and density of the ridge (documented by Bryan and Marcus, 1972), very low velocities were characteristic of these areas. Using the elementary seismic interpretational theory of Dobrin (1960), the following data were extracted from the time-distance graphs:

	Area A	Area B
Velocity of ice layer	0.477 km/sec	0.345 km/sec
Velocity of sand layer	2.49 km/sec	1.83 and 2.49 km/sec
Intercept time	0.0544 sec	0.0337 and 0.050 sec

The velocity variation for the sand layer in Area B could have been due in part to the pressure added locally from the increase in ice thickness, which may have compacted the sand and thereby increased the velocity of energy transmission through it. From the above data, the depths of the various interfaces were calculated using the relationship:

$$Z = \frac{T_i}{2} \frac{V_o \cdot V_1}{\sqrt{V_1^2 - V_o^2}}$$

where Z = depth
T_i = intercept line
V_o = velocity through ice
V_1 = velocity through sand

Using an average velocity for the ice layer (0.412 km/sec) the thickness of the ridge, from the geophones downward, was found to be:

Area	Ice thickness, m
A	11.20
B	10.30 and 7.42

Area B showed a thickness variation of almost 3 m over a distance of 50 m.

During the following winter at approximately the same location, we were presented an opportunity to profile the outer ridge using an ice drill. Other investigators from the University of Michigan had dug a trench through the ridge in order to examine its stratigraphic nature (Bryan and Marcus, 1972). From the floor of their trench we were able to drill to the lake bed. A detailed transverse profile was determined which generally confirmed the seismic findings. It revealed relief variations in the bottom of the ice ridge as great as 0.75 m in a distance of 1.5 m (fig. 4).

The subsurface configuration of the interridge ice is smoother than that of ridge ice. Profiles, determined by drilling, across the interridge ice between the second and third ridges indicated that small scale relief variations on the order of those found under the ridges are lacking here. However, gentle subsurface relief variations, corresponding to sand bars and troughs, generally less than 1 m over distances of several tens of meters, are common under interridge ice.

DISTRIBUTION

The locations offshore of ice ridge formation appear to be controlled by several factors including: (1) the size of the lake ice fragment-

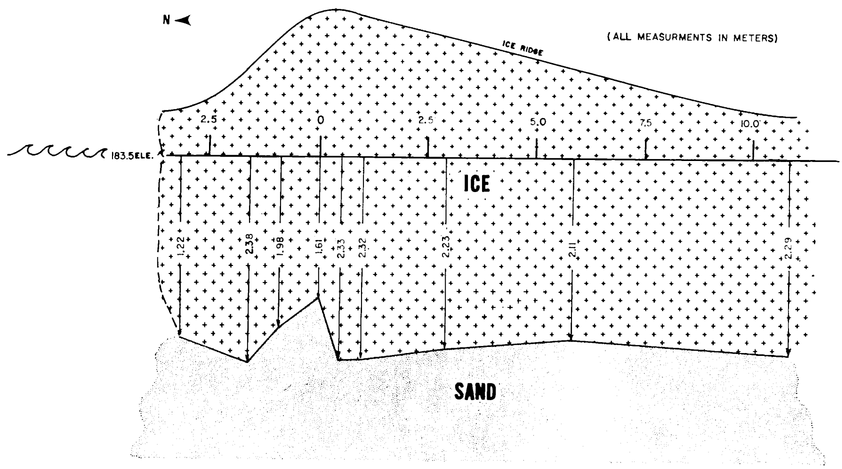


Fig. 4. An ice ridge cross section determined by drilling.

ball ice fields along the shore, (2) offshore bottom relief including the location of sand bars and the depth of water, (3) the size and duration of onshore waves. Although their spacing suggests that they may form on sand bars, a bottom profile showed as many sand bars without ice ridges as with ice ridges within 150 m of the shore. The distances from shore at which ice ridges form appear to be fairly consistent from winter to winter. Resurveys of a transect across the icefoot at one location near Grand Marais revealed that the location of the second and third ice ridges varied a maximum of only 16 m over three consecutive winters.

Given high exposure to storm waves, the distribution of ice ridges along the southeastern coast of Lake Superior varies chiefly with the morphology of the shoreline. Shores approaching an equilibrium morphology, that is, those with gently sloping offshore relief across which wave energy is rather evenly distributed, are characterized by the ice ridge formation described earlier in this paper. Three ridges usually form, but it is not unusual to find stretches of equilibrium shorelines with only two ridges.

The continuity of ice ridges along the shore tends to increase with the size of the ridge. The innermost and second ridges locally may be highly fragmented. In most instances there is no apparent reason for the fragmentation, except at river mouths where the area between the shore and the outer ridge is often free of ice throughout the winter. Although the third, outer ridge commonly extends many kilometers without a break along an equilibrium shore, breeches and fragmentation of it have been observed at several localities. Breeches occur where storm waves erode through a weak point in the ridge. This often leads to an incursion of waves into the interridge area and fracturing of the ice there. If storm waves are subsequently accompanied by an additional supply of ice fragments from the lake, the breech may be filled with this ice and the ridge rebuilt. On the other hand, if a supply of ice fragments is not available, erosion of the ridge may continue until large segments of it are consumed.

The distribution of ice ridges along shorelines comprised of bedrock cliffs is markedly different from that described above. Observations at the Pictured Rocks, located 25 km west of Grand Marais (fig. 1), indicate that along these cliffs, ice ridge formation is (1) very localized, (2) limited usually to a single large ridge, (3) restricted to a zone roughly within 50 m of shore. Where the depth of water adjacent to the cliff exceeds 5 to 7 m, ice ridges appear rarely to form. However, where water depths at the cliff foot approach 3 m or less, for example, along underwater talus slopes, ice ridges often form. Ice ridge formation is conspicuously absent from shores lacking exposure to storm waves. The shortest fetch known to produce ice ridges over 1 m high is 35 km.

PLANIMETRIC SHAPE

One very salient feature shared by all ice ridges examined is a wavy lakeward edge. Both the amplitude and the wave length generally increase with the size of the ridge. The wave lengths of inner ridges are on the order of 10 to 20 m, whereas those of outer ridges are on the order of 35 to 75 m. Wave lengths along a single ridge are highly variable over distances of several hundred meters.

In order to investigate the possibility of a relationship between size and direction of onshore storm waves and the planimetric configuration of the outer ice ridge, Fourier (or harmonic) analyses were undertaken from oblique aerial photographs of ice ridges offshore from both sandy, equilibrium shorelines and bedrock, low-cliff shorelines. A Fourier series for a length along the ridge is a sine-function approximation of the actual planimetric configuration, namely:

$$y = \sum_{n=1}^{K} \left[a_n \cos\left(\frac{2\pi nx}{L}\right) + b_n \sin\left(\frac{2\pi nx}{L}\right)\right]$$

where x = distance along ridge from arbitrary starting point,
L = total length of ridge section,
y = distance of outer edge of ridge from an axis parallel to the ridge.

Any set of data arranged so that only one y-value is associated with every x-value can be reproduced exactly, if K is equal to half the number of data points. Usually the data can be approximated closely for much smaller values of K. In the actual situation, the approximation surpassed the 99 percent level when K was set to 20.

The coefficients a_n, b_n are found by:

$$a_n = \frac{2}{N}\left[\frac{y_o + y_N}{2} + \sum_{i=1}^{N-1} y_i \cos\left(\frac{n \pi x_i}{L}\right)\right] \quad n = 0, 1, 2, \ldots, N/2$$

$$b_n = \frac{2}{N} \sum_{i=1}^{N-1} y_i \sin\left(\frac{n \pi x_i}{L}\right) \quad n = 1, 2, \ldots, N/2$$

From these the square root of the power spectrum estimate for each frequency can be calculated:

$$A_n = \sqrt{a_n^2 + b_n^2}$$

This value is a measure of the importance of each frequency, and strong periodicities in the data will be expressed by a large value of A_n, where n is the appropriate frequency. There is a large body of literature dealing with Fourier analysis and its applications; we have found Harbaugh and Merriam (1968, chap. 6) a useful synthesis.

Discrete frequency spectra for a sandy, equilibrium shoreline and a bedrock, low-cliff shoreline and the actual tracing for each are shown in figure 5. Obviously the amplitude/wavelength ratio is greater for the bedrock, low-cliff shoreline, and both the spectra and the planimetric tracings show this. The spectra also show that a single, dominant wavelength does not exist. While undoubtedly a relationship does exist between the period and approach angle of formative, onshore waves and the planimetric configuration of the ice foot, the relationship is exceedingly complex, as the form of the ice ridge is the result of waves of different events with different periods and approach angles.

The greater amplitude/wavelength ratio of the lakeward edge of the icefoots off bedrock cliff shores appears to be related to irregular offshore topography. Deep reentrants in the icefoot front often correspond to zones of deep water (usually greater than 3-4 m) off cliffs which are due to relief variations in the bedrock and uneven distribution of underwater talus. These zones allow for deep landward penetration of large waves, thereby limiting icefoot formation to a narrow strip of shallow water at the cliff foot or, in the absence of shallow water, prohibiting icefoot formation altogether.

ICEFOOT DECAY

The decay of the icefoot along southeastern Lake Superior roughly follows its sequence of growth as the outer ice ridge is generally the last major feature to disappear. By late March the microrelief of most of the icefoot is reduced to a fairly smooth surface. The influence of sun angle is readily apparent as the ice balls on south-facing slopes of ridges ablate differentially, forming sun cups. Also on these slopes networks of rills form from melt water drainage. A thin (less than 2 cm), uneven

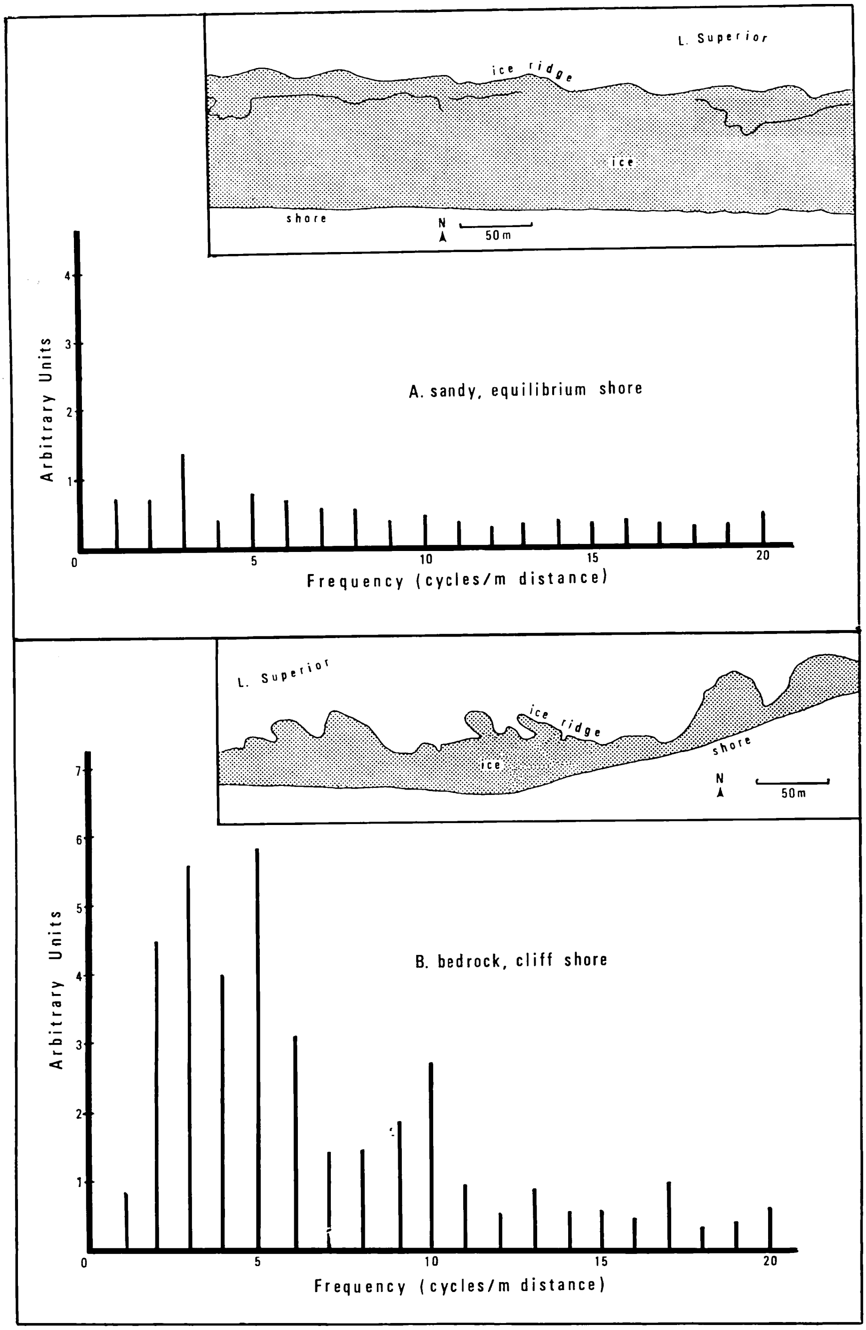


Fig. 5. The discrete frequency spectra and the planimetric configurations of the lakeward edge of (A) a sandy, equilibrium shore, and (B) a bedrock, cliff shore.

layer of sediment covers most surfaces. During melting days, pools of standing water form on the interridge ice in late afternoon. By mid-April, the entire interridge area is covered by water, and large segments of the lakeward edge of the outer ice ridge may be free of lake ice.

In April 1969 a combination of events occurred that destroyed the icefoot near Grand Marais in two days. On April 20, the entire icefoot was intact, and water covered the interridge ice. At 5 p.m. of that day the first rainfall of the year began and was accompanied by strong (20-30 knots; 35-55 km/hr) southerly winds. The rain continued to fall throughout the following day, the icefoot began to break up, and the wind shifted to the north. Breeches formed in the outer ridge, and most of the interridge ice disappeared. By the following day, the entire icefoot was gone. Although this represents only one set of observations, it is likely that such a combination of events is necessary for the rapid decay of the icefoot. Barring such a combination, the icefoot appears to decay gradually, lasting well into May in some locations.

GEOMORPHIC INFLUENCE

Ice ridges influence shoreline geomorphology in several ways. The most important influence is that they function as natural seawalls, preventing waves from reaching the shore. As the icefoot grows in December and January, wave action is displaced progressively lakeward. The winter shore, potentially the highest energy shore of the year, is therefore isolated from waves for approximately five months of the year. Measurements show that backshore slope retreat due mainly to wave action approaches 0.75 m per year at several locations along southeastern Lake Superior. Undoubtedly this rate would be much greater if not for the barrier posed by the ice ridges to winter storm waves. Although waves are largely absent from the shore in winter, several other geomorphic processes are very active at this time. On barren, sandy backshore slopes onshore wind is the principal winter erosional agent. Measurements made over two winters indicate that 10 cm of sand per winter were eroded by wind from the upper parts of a 90 m high backshore slope near Grand Marais. Mass wasting on backshore slopes is very limited in winter due to the strength added to slopes by ground frost. However, as ground frost melts in late March and April, slumps, slides, and rock falls are released on active slopes. Observations and measurements indicate that these processes displace as much as 1000 m^3 of debris per kilometer from the 90 m high slope each spring. In addition, mudflows and runoff related to ground water seepage also contribute locally to the retreat of backshore slopes in late winter and spring. The debris from these processes accumulates on the shore and backshore during this period. In terms of the mass balance between the quantity of debris produced by mass wasting and erosion from backshore slope and the quantity of this debris removed by waves from the backshore and shore, winter and spring are clearly seasons of positive balance. Not until the ice ridges disappear in April or May is this debris susceptible to erosion by waves and incorporation into the beach system. By mid-December the icefoots have usually begun to reform.

Compared with Arctic beaches, the number of icefoot-free months is 4 to 5 times greater on Lake Superior. Studies by Hume and Schalk (1967, p. 91) indicate that net sediment transport by shoreline processes near Point Barrow, Alaska, is approximately 10,000 m^3 per year. Owens and McCann (1970, p. 413) suggest that this quantity may be even lower for the shore of southwest Devon Island, Northwest Territories, Canada. Based on measurements by Bajorunas and Duane (1967, p. 6199), the sediment transport along the southeastern shore of Lake Superior is estimated at 130,000 m^3 per year. Although a comparison is limited by differences in fetch, storm frequency, and other variables, a significant share of the difference between Lake Superior and Arctic net sediment movements appears to be attributable to the greater icefoot-free season on Lake Superior. Data on sediment transport along ice-free coasts indicate that the Lake Superior icefoots give the shore considerable protection. Along the Pacific Coast in southern California, where wave fetch is unusually long, net annual drift rates exceed 200,000 m^3 on several stretches of the coast, and along the Oxnard Plainshore net littoral drift is about 1,000,000 m^3 per year. Such an extreme value is not found along the Atlantic and Gulf of Mexico Coasts, but there are locations with net transport rates in excess of 300,000 m^3 per year (Johnson, 1956, p. 2218; Ingles, 1966, p. 162-163). Zenkovich (1967, p. 354) cites measurements of 150,000 m^3 per year at Sochi on the northeastern coast of the Black Sea. In terms of both net sediment movement and beach morphology related to ice processes, Lake Superior shores appear to be intermediate between Arctic shores and shores of essentially ice-free coast environments.

A comparison of the winter beach morphology of Lake Superior with that of the Arctic coast indicates that Lake Superior beaches lack the ice-push ridges of sediment common in the Arctic (Owens and McCann, 1970, p. 408-412). Another beach feature, hummocky microrelief due to differential ablation of sediment-covered ice, is found on both Lake Superior and Arctic beaches; however, the relief is much greater in the Arctic.

Ice ridges also influence the distribution of wave energy in the offshore zone. Wave energy is dissipated in a narrow zone along the ice ridge front rather than over the much broader zone toward shore. The turbulence created by waves as they crash into the ice ridge results in suspension of sediments which are often incorporated into the ice ridge by wave overwash and spray. Indeed, the weight added by sediments to low density ice (ball ice and granular ice) is sometimes great enough to sink pieces of this ice that have been eroded free of an ice ridge. Although it is evident that part of the wave energy is concentrated downward at the ice front, profiles reveal negligible net erosion of the bottom in this zone.

Ground and aerial photographic surveys indicate that the sediment content of icefoots varies at three different scales. Within a single large ice ridge, the sediment content generally increases with elevation above lake level. Often the sediment-laden upper portion of an ice ridge appears as a single layer, separated by a sharp boundary from the cleaner, underlying ice (pl. 2-A). This results from the two-part sequence of ice ridge formation: (1) the lower portion is formed from the initial grounded mass of lake ice fragments and ball ice, whereas (2) the upper portion is formed from ice added by the turbid overwash and spray of waves. The broad interridge areas of the icefoot are formed from the same grounded mass of ice as the lower portion of the ridges and therefore also tend to have relatively low sediment contents. Locally along the foot of bedrock cliffs, several inches of sediment are deposited on the icefoot by mass wasting, but this sediment is usually not incorporated into the ice mass. At the largest scale, the sediment content of icefoots decreases from sandy to bedrock shorelines. This is attributable to the corresponding decrease in available unconsolidated offshore sediments and the resultant decrease in wave turbidity from sandy to bedrock shorelines.

A minor geomorphic influence of ice ridges is grooving of the lake bottom along sandy shorelines. During the final phases of icefoot decay, the fragments of the outer ridge become mobile and are moved by winds and waves in and out of the shallow offshore water. As the ice drags across the bottom, the protrusions at the base of the fragments produce grooves in the sand (Weber, 1958). Although very little is known about the dimensions and distribution of these features in the Great Lakes, networks of them have been observed from the air near Whitefish Point during the breakup of the ice there (Anthony Brazel, 1970, personal commun.).

CONCLUSIONS

Given subfreezing temperatures and storm waves, two conditions are particularly important for the formation of large, offshore ice ridges: (1) stabilized, grounded offshore ice, and (2) a supply of lake ice fragments and ball ice. A grounded agglomeration of ice, compact and frozen enough to resist displacement by waves, deflects much wave energy upward along the ice front. This provides a sufficient source of energy to lift ice fragments and ball ice onto the surface of the stationary ice and to produce overwash and wave spray, thereby building a ridge. Lacking a supply of ice fragments and ball ice, storm waves produce essentially the same effects as along a small sea cliff, namely, overwash, spray, and erosion. Growth of the ridge from overwash and spray alone is generally not great enough to offset erosion by wave undercutting, and therefore a ridge does not form.

The configuration of the sand-ice interface at the bottom of the icefoot is smooth under interridge ice and rough under ridge ice. The roughness of the lower boundary of ridge ice is likely due to characteristics inherited from the initial agglomeration of ice during ridge formaton. The anchoring effect of the protrusions at the base of this ice, in addition to the fact that the mass of the outer ridge is situated above the surface of the shallow water (3 m) in which it forms, surely contributes to its longevity in coastal water in spring.

The distribution of ice ridges along southeastern Lake Superior varies with exposure to storm waves and with offshore morphology. Along high exposure shores with offshore morphology approaching equilibrium conditions, a broad continuous icefoot with three ridges usually forms. Ice ridge formation is discontinuous along bedrock cliff shores and is limited to locations with shallow water immediately off shore. Usually only a single ice ridge forms at a distance of less than 50 m from shore. Although the relationship between formative, onshore waves and the planimetric configuration of the icefoot is not clear, the greater amplitude/wavelength ratio of the wavy lakeward edge of icefoots off bedrock cliff shores than that off sandy, equilibrium shores appears to be related to the irregular offshore topography along cliff shores.

The main geomorphic influences of ice ridges are the lakeward displacement of the active wave zone and the resultant protection of the shore from high energy winter waves. Mass wasting and erosion on active backshore slopes bring debris to the footslope where it accumulates during the winter and spring. Incorporation of this debris into the shoreline sediment system is limited to the icefoot-free season, generally from May to December. The estimated net sediment transport along the southeastern shore of Lake Superior is 11 times greater than that near Point Barrow, Alaska, where the ice-free season lasts only one or two months, but is much less than that along ice-free coasts.

ACKNOWLEDGMENTS

Thanks are due to many citizens of Grand Marais, Mich., who directly and indirectly contributed to this study. We are particularly grateful to Mr. Joseph Massey for logistical assistance and to Mr. David Becker for recording observations on icefoot decay. Dr. M. Leonard Bryan, Dr. J. C. Savage, and Dr. James Zumberge read an earlier draft of this manuscript and made suggestions for improvement. Dr. Bryan also brought some pertinent references to our attention. Part of the field work was supported by the Institute of Science and Technology, University of Michigan, research funds under the direction of Dr. Melvin G. Marcus, Department of Geography, University of Michigan, Ann Arbor. Part of the publication cost was paid by National Science Foundation Grant GA-11735, under the direction of Dr. Ian Carmichael, Department of Geology and Geophysics, University of California at Berkeley.

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