ANCHOR ICE AND BOTTOM-FREEZING IN HIGH-LATITUDE MARINE SEDIMENTARY ENVIRONMENTS: OBSERVATIONS FROM THE ALASKAN BEAUFORT SEA

by

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INTRODUCTION

As early as 1705 sailors observed that rivers sometimes begin to freeze from the bottom (Barnes, 1928; Piotrovich, 1956). Anchor ice has been observed also in lakes and the sea (Zubov, 1945; Dayton, et al., 1969; Foulds and Wigle, 1977; Martin, 1981; Tsang, 1982). The growth of anchor ice implies interactions between ice and the substrate, and a marked change in the sedimentary environment. However, while the literature contains numerous observations that imply sediment transport, no studies have been conducted on the effects of anchor ice growth on sediment dynamics and bedforms.

Underwater ice is the general term for ice formed in the supercooled water column. It exists in 2 forms: frazil ice and anchor ice, also called ground ice or bottom ice. Frazil ice consists of disk-shaped crystals 1-4 mm in diameter and 1-100 microns thick, that form in turbulent, slightly supercooled water (Kivisild, 1970). Frazil crystals during periods of supercooling are sticky, adhering to each other and to foreign objects (Martin, 198 1). When turbulence carries frazil ice to submerged, supercooled objects or the bottom, the frazil therefore may adhere to the subtrate, forming anchor ice (Piotrovich, 1956; Benson and Osterkamp, 1974; Tsang, 1982; Osterkamp and Gosink, 1982). Once anchor ice is formed, it may grow rapidly by free growth in the supercooled water or by trapping other frazil crystals from the water column (Osterkamp and Gosink, 1982; Tsang, 1982). Since the substrate also has to be supercooled, an ice-bonded crust should form in sediment saturated with water of the same or lower salinity than the water column.

The most convincing observations on actual sediment transport by anchor ice have been made in the Niagara River. Here masses of "muddy colored ice filled with bed material" rise to the surface following a night of anchor ice growth. Also, traps placed on the bottom collect masses of spongy ice with sediment. Furthermore, settling basins near water intakes collect up to boulder-size material during winters, while almost no sediment is moved during summers (Tsang, 1982). Osterkamp and Gosink (1982), in a study of intenor Alaskan streams, describe how slabs of porous ice accreted on the bed are picked up with the attached sediment by turbulent flow and are moved slowly and intermittently downstream. The above observations imply that much more sediment is transported in a flow than that seen on the surface. Benson and Osterkamp (1974) discuss the possibility of sediment transport into northern seas by anchor ice.

The mechanism of anchor ice growth whereby "sticky" frazil crystals attach to bed material implies grain transport on a microscopic scale as well. The bed material has to be coarse, with individual clasts heavy enough to counteract the buoyancy or drag resistance of the ice buildup (Arden and Wigle, 1972). Particles smaller than fine sand are lifted off the bottom under the buoyancy of ice floes or even individual frazil crystals, and thence are carried along with the flow. For this reason, anchor ice has only been observed on sand and coarser substrate, but almost never on fine sand, silt, or clay (Wigle, 1970; Arden and Wigle, 1972; Tsang, 1982). Frazil crystals have also been noted for their scavenging of algae from the water column (Weeks and Ackley, 1982). This action removes fine suspended matter from the water (Altberg, 1938), so that “the first run of frazil has a remarkable cleansing effect in the water” (Barnes, 1928). Coarse organic matter and organisms also attract the growth of anchor ice. In the 18th century, Elbe River fishermen noted that willow baskets used as eel traps were commonly covered by ice when raised to the surface (Barnes, 1928). Dayton, et al. (1969) report anchor ice in the Antarctic to 33 m depth. Above this depth benthic organisms are lifted off the bottom and incorporated into the surface ice canopy when the buoyancy of growing anchor ice exceeds their weight or strength. The result is a reduction of certain benthic organisms above the 33 m isobath.

These observations of Dayton, et al. (1969) represent the best documented case of anchor ice in the sea. Sadler and Serson (1981) describe the formation of a narrow fringe of anchor ice near beaches in the Canadian Arctic, and large masses of anchor ice have been reported growing off Newfoundland to depths of 20 m (Barnes, 1928). Besides these observations we only find statements that anchor ice is well known, widespread, and sailors see it rising to the surface.

In this report we describe miscellaneous observations relevant to anchor ice in the Alaskan Beaufort Sea, including direct diving observations. We also discuss the conditions, timing, and probable extent of the phenomenon, and lastly we emphasize anchor ice as a potentially important geologic agent.

REGIONAL SETTING

The open shelf of the Alaskan Beaufort Sea, in the area of our observations (fig. 1) is covered by a thin layer of muddy sand to mud, has very little relief, and is shallower than 20 m for 10 to 50 km from land. The coast is fringed by island chains up to 15 km from land, protecting 1- to 6-m deep lagoons from pack-ice intrusion. Total ice cover exists for more than 9 months of the year, and even during the navigation season heavy pack ice concentrations normally remain over much of the shelf. River discharge begins in early June by flowing out across the extensive fast ice. By mid September the North Slope drainage basins begin to freeze, effectively eliminating river-water discharge to the sea, and a new sea ice cover begins to grow by late September. At this time bottom-water temperatures below the 10 to 15-m thick mixed layer are only slightly above their freezing points (Hufford, 1974). In winter the water column typically is at its freezing point down to a depth of 40 m (Aagaard, 1984). Much of the inner shelf is underlain by ice-bonded sediments at shallow depths below the seafloor (Neave and Sellman, 1984; Morack and Rogers, 1984) Along a transect from Prudhoe Bay northeastward to 12 m depth seaward of Reindeer Island (fig. 1) winter seafloor temperatures are at the freezing point for the salinity of interstitial waters, and seasonal freezing is assumed to occur (Sellman and Chamberlain, 1979). Widespread overconsolidation of surface sediment could be attributed to seasonal freeze-thaw (Lee et al., 1985), but general ice-bonding of the seafloor is not documented. For more details of the regional setting the reader is referred to Norton and Sackinger (1981).

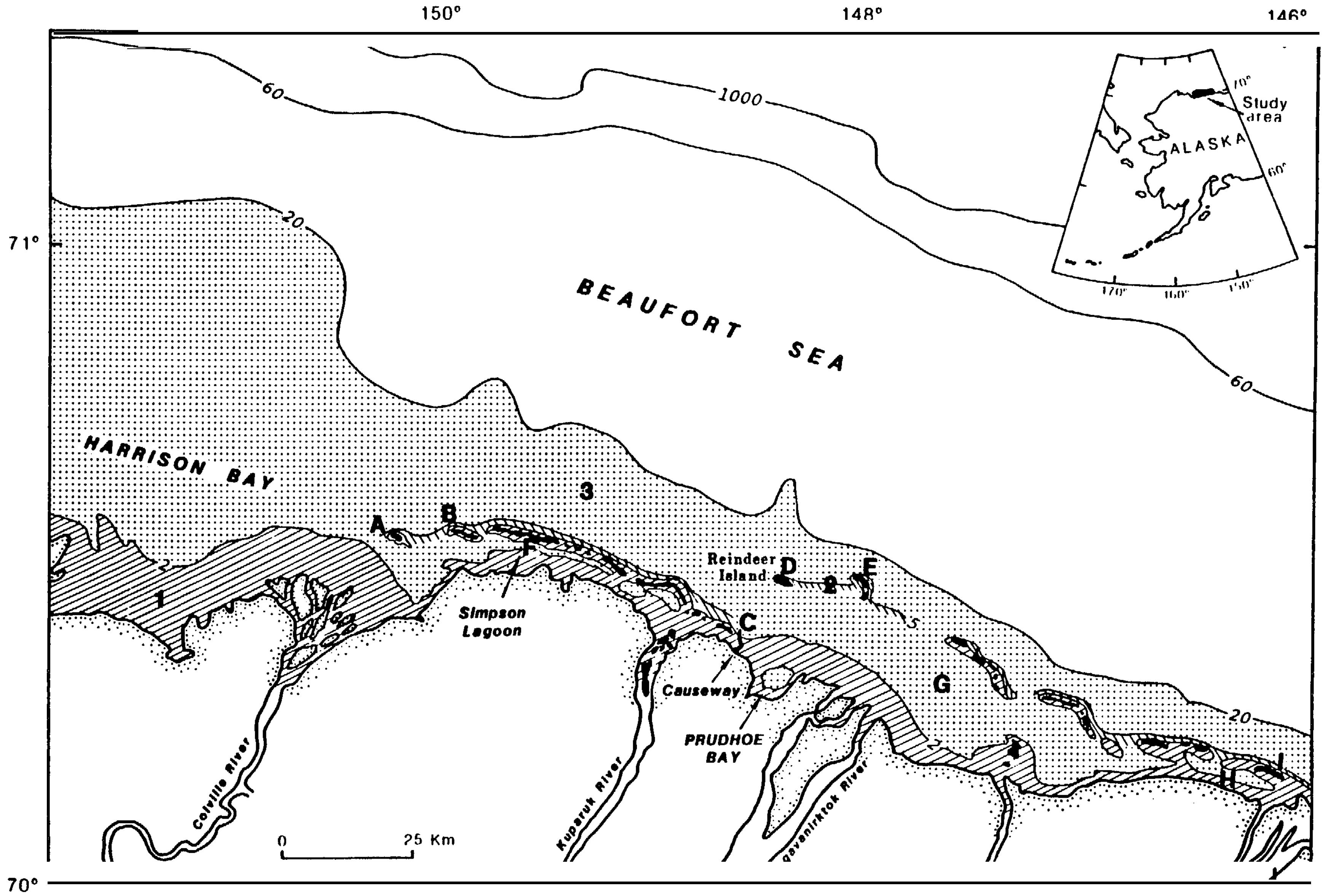

Figure Map of central Beaufort Sea shelf with lettered sites where evidence of anchor ice or bottom freezing was observed: (A) Thetis Island, (B) Spy Island, (C) West Dock, (D) Reindeer Island, (E) Cross Island, (F) Simpson Lagoon, (G) Stefansson Lagoon, (H) Point Thomson, and (I) Flaxman Island. Three regions with different likelihoods of seasonal bottom freezing are delineated: (1) certain at lower than 2 m, (2) very likely on exposed shorefaces to 5-m depth, and (3) possible to 20-m depth.

Based on our observations over the last 15 years, the onset of freeze-up and the formation of a new ice canopy are commonly initiated by one or more storms with 10 m/s or stronger easterly winds and freezing air temperatures. During this period night temperatures may drop to - 10°C, and increasing amounts of new ice are seen daily in sheltered areas between ice floes and on brackish-water plumes in extensive shallows near rivers. Storms lasting several days cause a number of important changes. We have observed that a cold 3 to 4 day storm generates inner-shelf currents of 50 to 100 cm/s, mixes waters laterally and vertically to the thermocline near 15 to 20 m depth on the mid-shelf, replaces brackish waters in bays and lagoons with higher salinity shelf waters, generates brash ice from the impacts and grinding action in fields of multi-year ice, and extensively rearranges the sea-ice distribution. Most important for the formation of underwater ice is the fact that such a cold storm cools the ocean, and results in the formation of large amounts of frazil ice aligned on the sea surface in streaks parallel to the wind, a sure sign that the surface layer of the ocean now is supercooled. Cooling of surface waters is enhanced by wave- and ice-induced turbulence associated with the storm. In the calm following storms this frazil ice is found as a surface layer of grease ice with reported thicknesses of up to 4 m (Collinson, 1889), depending on how the coast, island chains, and grounded ice-fields are oriented with respect the driving storm winds. Swell can travel through the grease-ice layer for a few km.

INDIRECT EVIDENCE FOR ANCHOR ICE IN THE BEAUFORT SEA

From the beginning of our Beaufort Sea studies in 1970 we were aware of the possible presence of anchor ice (Reimnitz and Barnes, 1974). While routine marine geological survey techniques may not reveal the phenomenon, our work has also included hundreds of research dives in diverse settings and seasons, as well as years of small boat and skiff operations in the nearshore environment during freezeup. Our knowledge also includes observations from increasing numbers of dives made under the ice canopy by consulting firm personnel and researchers.

Prior to 1982 we had no direct evidence for anchor ice in the Beaufort Sea, although indirect evidence suggested its presence. (Reimnitz and Dunton, 1979). One type of indirect evidence is the large amounts of fine sediment that are incorporated into the seasonal fast-ice canopy (Barnes, et al., 1982; Osterkamp and Gosink, 1984). The concentrations of fine sediments over extensive regions in this ice are at times an order of magnitude larger than concentrations of suspended matter seen in the open-water season (Barnes, et al., 1982). Barnes, et al. (1982) attribute these high sediment concentrations to scavenging of sediment particles from the water column by frazil ice. The sediment at the sea surface could also be attributed to the buoyant action of frazil or anchor ice adhering to sediment particles on the seafloor during fall storms (Reimnitz and Dunton, 1979), although Barnes et al. (1982) reasoned this is unlikely. Spotty occurrences of organic matter, coarse sediments, and clam shells, often with attached benthic organisms such as kelp (i.e. Reimnitz and Barnes, 1974), are commonly seen in the smooth ice canopy (fig. 2), or in the layer of soft ice below the solid' ice canopy. Such materials are never seen on the sea surface in open water, and therefore their isolated inclusion in the ice canopy requires suspension by anchor ice. For example,

Figure 2. Mass of coarse-grained sediment entrained in new, thin ice cover on seaward side of Flaxman Island (fig. 1-1). Masses such as these are probably carried to the water surface by the buoyant action of anchor ice acting on bottom sediments.

incidental collection from 500 m^2 of soft ice under a lagoonal ice canopy produced a handful of kelp and other benthic organisms, some attached to pebbles (fig. 3, from Dun ton, written communication, 1980). These materials match the unique local substrate.

Ice growth on objects in the water column, both below the ice canopy and in open water during fall storms, has been variously observed (i.e. Reimnitz and Dunton, 1979). For example, fish nets of certain materials are carried to the surface when the buoyancy of ice growing on webbing exceeds the weight of lead lines (Jim Helmericks, oral communication, 1979). Such ice growth has been seen on mooring lines and oceanographic equipment (fig. 4) in the quiet water column under an ice cover.

During or after fall storms there are various indicators that either the bottom is frozen or that anchor ice exists. An example are slabs of sand with vertical edges littering a gravel beach (fig. 5A and B). We commonly see 2- to 4-cm floes of sediment-laden ice rise to the clean sea-surface when anchoring a small boat in 2 m water depth. These floes apparently are released from the bottom by turbulent propwash when backing the vessel. In several instances, when anchored 150 m from a

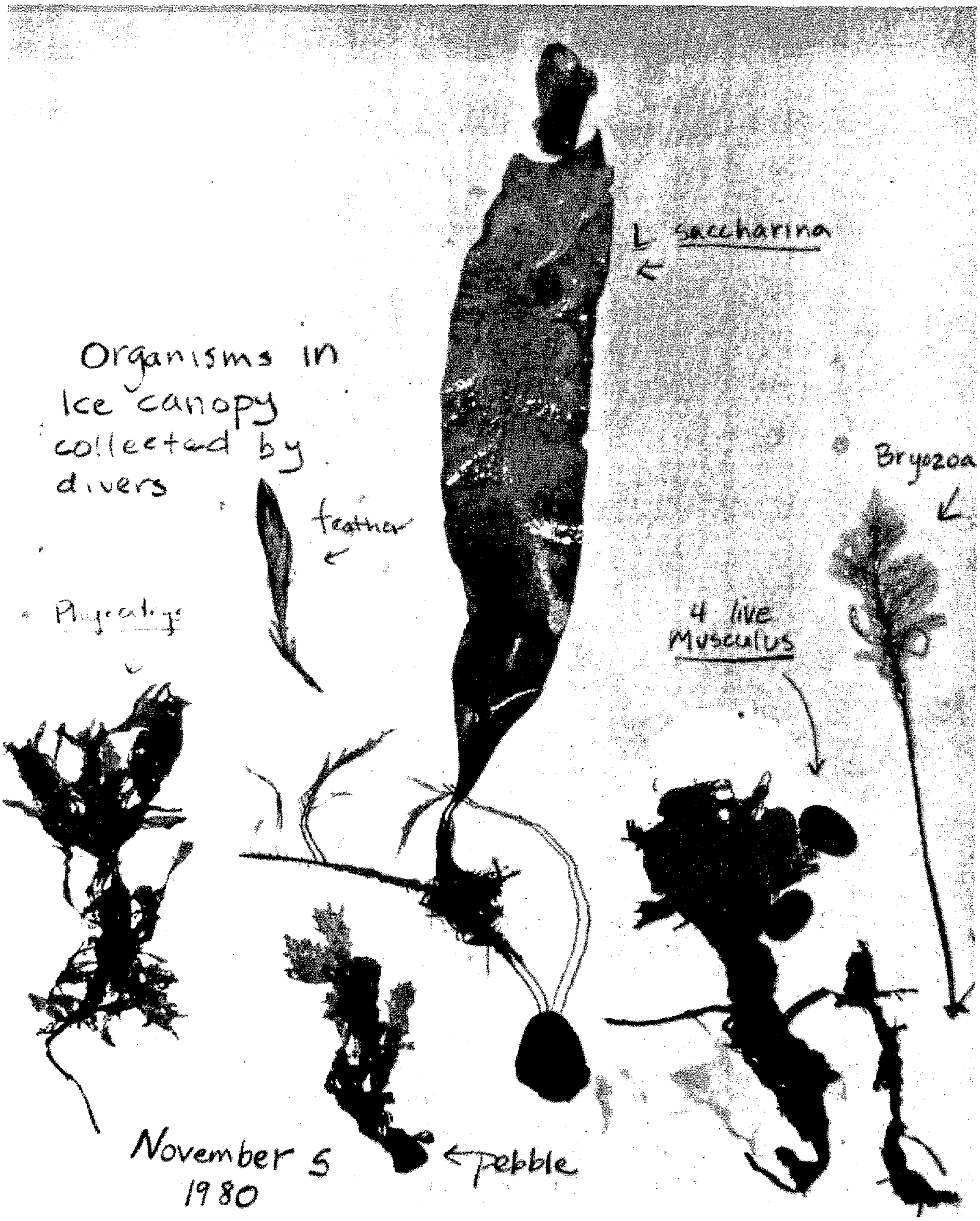

Figure 3. Example of different organic materials found in the soft ice under the solid ice canopy in Stefansson Sound These materials are never found floating on the water surface in the summer, and are believed to be incorporated into the ice canopy by the action of anchor ice (Photograph by K.Ii. Dunton).

Figure 4. Anchor ice growth below ice canopy on instrument mooring ice in Stefansson Sound (K.H. Dunton, Harding and Lawson Assoc.).

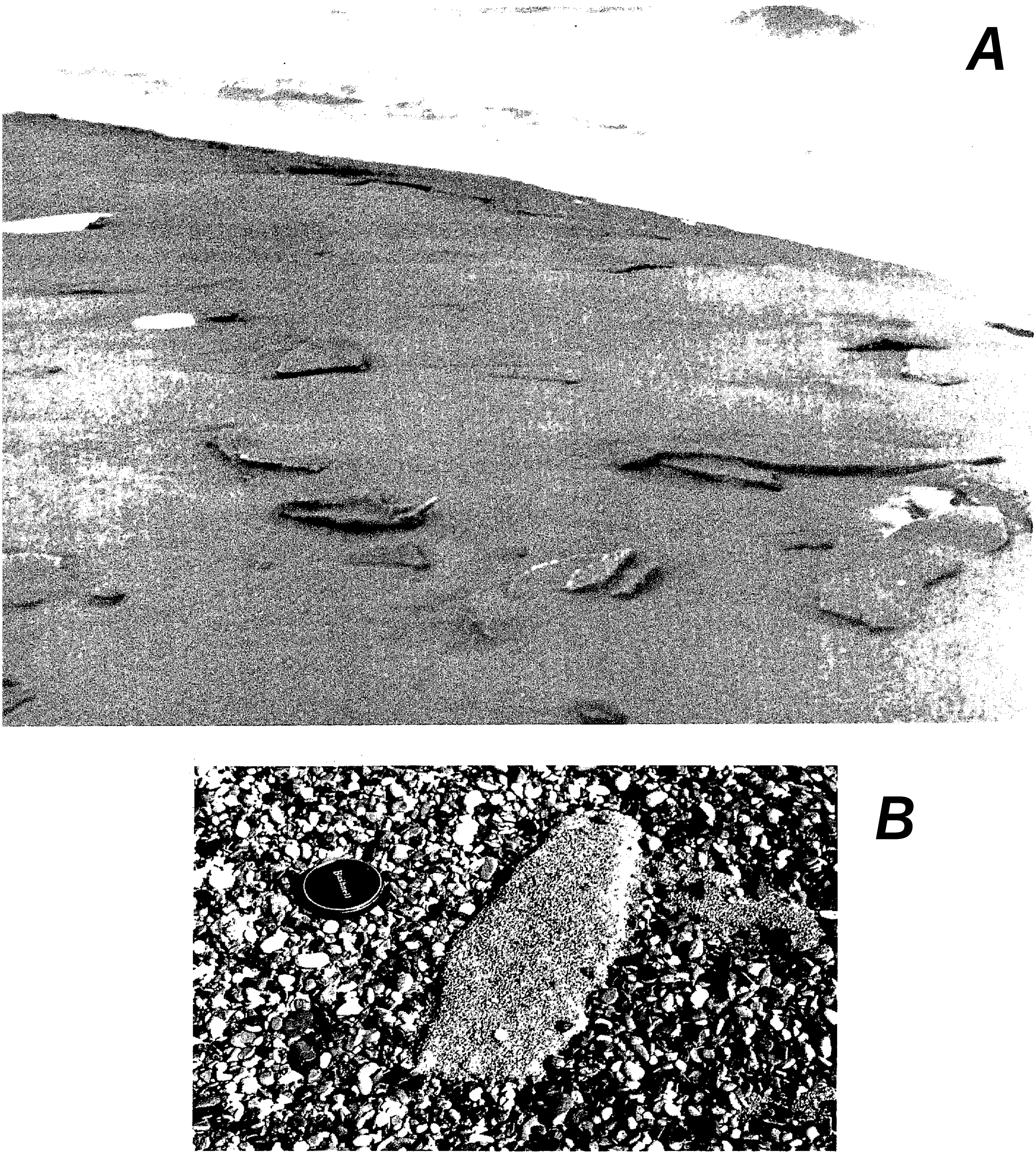

Figure 5. Plaques of sand, tossed as a frozen slab onto beach during fall storm. (A) Plaques on beach face **during** a fall storm (B) Sand plaque on gravel beach **during** the summer, after the **ice** holding the plaque together has melted. Such evidence for seafloor freezing **is** common on **seaward facing beaches.**

protecting barrier island at water depths of 1.5 to 2 m during freezing storms, we observed such sediment-laden spongy ice drift past the vessel. At Thetis Island (fig. l-A) on 9-21-1980, wind-parallel streaks of slush extended windward to within 10 m of the lee shore. These streaks contained many ice floes carrying sand and coarser material, which most likely had an anchor ice origin, By jabbing the seabed we determined that it was ice-bonded. At this same time, near the beach at a water depth of 0.8 m we used a shovel to retrieve a slab of ice-bonded sand. On the same day in Simpson Lagoon (fig. 1-F) we repeatedly encountered hard bottom at a water depth of 1.5 m, where on] y unconsolidated sediment is normally found. This hard bottom is evidence for ice-bonding of sediments. A Beaufort Sea tugboat operator related that during a fall storm two anchors bounced along the seafloor without digging in at a normally good anchor site (Jim Adams, oral communication, 1984). He believes this may have been due to anchor ice and frozen seafloor.

DIVER OBSERVATIONS OF ANCHOR ICE AND ICE-BONDED SEDIMENTS

Direct evidence for anchor ice was obtained from two dives made after fall storms. The dives were prompted after observing rising ice while wading in waist-deep, open water se-award of Cross Island (fig. l-E), where slabs of frozen sand had been thrown onto the beach face by waves. During the 5 days prior to the first dive (10-5- 1982), daily minimum temperatures of -8 to -10”C and maximum wind velocities of 21 m/s were recorded along the Beaufort Sea coast (U.S. Department of Commerce, NOAA, 1982). New ice covered large areas of the inner shelf during this time, indicating surface water temperatures at the freezing point.

The first dive, a shore-normal, 130-m-long traverse from the beach to 4.5 m depth, was made off Reindeer Island (fig. l-D). Turbid waters obscured the seafloor viewed from above even where the depth was only 0.5 m. Open water prevailed on the inner 50 m, and a 20-cm thick layer of weakly sintered slush ice covered the outer part of the traverse. About 90 m from shore the traverse crossed a 2-m-high, shore-parallel sand bar. Bottom sediments ranged from gravelly sand nearshore to sorted fine sand offshore. At the time of the traverse near-bottom salinity was measured at 36.8 ppt, and the temperature was -2°C.

From the beach to 2-m depth, about 30 m from shore, the seafloor was partially to totally ice-bonded. At less than 1-m depth, 5-10 cm high, ill-defined l-m or larger slabs of rippled gravelly sand, elongated parallel to the beach, gave the bottom an eroded appearance. At depths greater than 1 m irregular pillow-shaped masses of ice and ice-bonded sediments 10 to 40 cm across and 5 to 8 cm high rose above a rippled sand and gravel bottom (fig. 6). These pillows had an external crust of randomly oriented ice platelets up to 1 cm in diameter, with sediment grains trapped in the interstices between the ice platelets. With increasing depth below the outer crust the pillows were increasingly firm and dense, and their cores could be extracted from the surrounding sediment intact. Figure 7 shows several of these cores that were extracted from the seafloor and carried up to the surface. The anchor ice pillows were largest and best-developed at 3 to 4 m water depth on either side of the offshore bar, and were absent on the bar crest. Pillows occurred singly, or in patches up to four meters

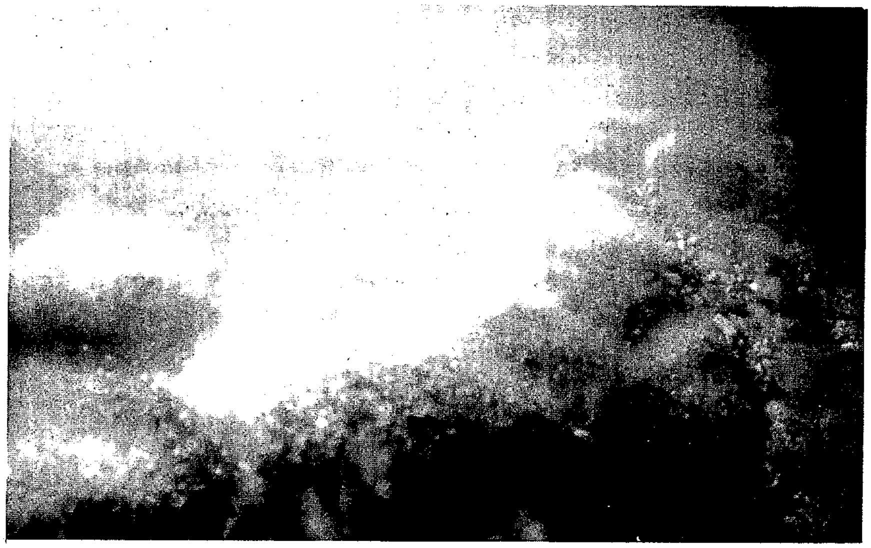

Figure 6. Fuzzy-looking anchor ice attached to gravel bottom at 1.5 m water depth off Reindeer **Island (fig. l-D.). Width** of scene is about 1 m.

Figure 7. **ice-bonded sediment cores of anchor-ice pillows collected off Reindeer island (fig.** 1-D). **Cleat is 30 cm long.**

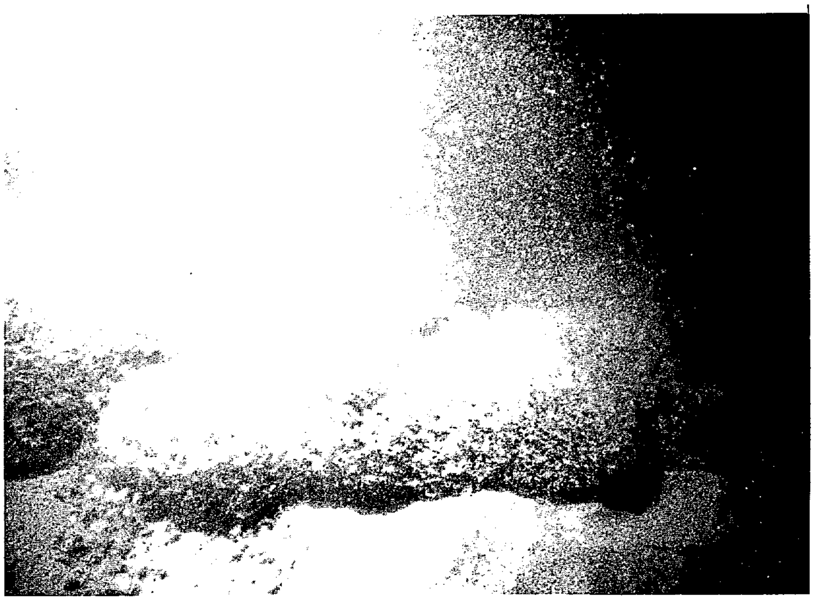

Figure 8. Anchor-ice pillows at 4-m depth, 130 m seaward from Reindeer Island. Individual pillows are 0.3 to 0.5 m in diameter, with a massive core of ice-bonded sand below a rind of delicate ice-crystals.

across, with individual pillows in close contact, but retaining well-defined boundaries (fig. 8). A rippled, thawed sand bottom surrounded the ice pillows in these deeper regions of the traverse. But the pillows were not only surficial features. To raise one off the bottom, we had to probe into the thawed sand to reach underneath an ice-bonded pillow-root, 5-10 cm below the seafloor. Such massive cores of ice pillows were difficult to break by hand, producing angular blocks of ice-bonded sand. Besides the large ice pillows we saw numerous 4-5 cm wide, 3-cm high balls of porous ice resting on the seafloor. When disturbed, these fragile balls of ice rose to the sea surface with some incorporated sediment. We did not detect any nuclei of ice-bonded sediment for these ice floes.

None of the large, massive ice pillows dislodged were sufficiently buoyant to ascend to the surface. The massive cores of ice rich sand carried to the surface (fig. 7) contained 20% of excess ice. A 100 kHz side-scan sonar traverse recorded along the diving traverse provided no clues of the existence of anchor ice.

On the next day, anchor ice and ice-bonded seafloor were observed during a dive made near a gravel causeway at water depths between 1 and 2 m (fig 1-C). The weather was calm and sheets of fresh ice covered the sea surface. The underwater visibility was 0.75 m. Along the 175-m traverse the bottom sediments ranged from solidly ice-bonded sandy gravel, to lightly bonded rippled muddy sand, to soft mud in 5- to 10-m wide depressions riddled with fragile ice platelets. The surfaces of mud-filled basins were marked by faint geometric patterns. Running a gloved hand through such soft mud-and-ice mixtures gave a distinct crunching sensation. Patches of ice-bonded muddy sand could be broken and disturbed under strong pressure by hand, but in the sandy gravel even an entrenching tool was useless to dislodge a single clast protruding above the bottom.

Also seen on this diving traverse were small, scattered accumulations of ice platelets with traces of entrapped sediment. These accumulations had nuclei of fibrous organic matter, kelp fragments, and gravel clasts, but wooden twigs apparently did not attract ice growth. The most remarkable growth of anchor ice was seen on trash, such as a rubber tire and a heap of steel banding. Accumulations of randomly oriented, thin ice platelets constituted a very open framework 40 cm across, appearing similar to an irregular bush with a structure of branches. The lightest water turbulence generated by our hands broke apart these large ice aggregates and sent the components rising to the surface.

DISCUSSION AND CONCLUSION

Diving observations suggest that anchor ice and bottom freezing are important for sediment dynamics. For future studies of how seafloor sediments are entrained into and rafted by ice, the generally accepted definition for anchor ice quoted initially is imprecise, as already pointed out by Tsurikov (1966) and Dunbar (1967). For a clear distinction from other forms of ice that may be stuck on the bottom and in the process may incorporate sediments, we suggest the following definition for anchor ice: "Ice accreted on a substrate submerged in either quiet or turbulent supercooled water, and remaining attached to the substrate." When dislodged, the ice generally transports some components of the substrate. The sediment-enriched base of anchor ice may not be distinguishable from underlying pre-anchor-ice sediments, which became ice bonded. However, the present definition excludes simply ice-bonded sediment from ice and sediment that accreted during an event, but future work may justify combining the two.

The different heat sinks for the formation of underwater ice may need to be distinguished for future studies of related sediment dynamics. Thus, the anchor ice forming in quiet conditions under an ice canopy in the Antarctic discussed by Dayton, et al. (1969) or shown in figure 4, and that forming where different water masses interact (i.e. Sadler and Serson, 1981), will have a different impact on the sedimentary regime than the anchor ice we find forming in the Beaufort Sea during turbulent open-water conditions, where the immediate heat sink is the atmosphere. Because of the hostile environment during the latter condition, this will be the most difficult to study.

Our sketchy observations from the shallow Beaufort Sea suggest ice-bonded sediments and anchor ice are widespread, if not ubiquitous, in years when the freeze-up is initiated by 12 m/s or stronger winds. Applying our understanding of why the bottom may become ice bonded and how anchor ice forms to the environment within the area shown in figure 1 allows regional extrapolation from our spotty observations. Wind-driven currents during fall storms, when freshwater input from rivers has terminated, homogenize the waters of the inner shelf. With these waters supercooled, subtle differences in salinities of interstitial waters in bottom sediments result in different seafloor freezing points and thereby influence the occurrence of ice-bonded sediments. Bottom sediments in Harrison Bay and Simpson Lagoon (fig. 1), infiltrated by river water during summer, have higher freezing points than bottom sediments seaward of the barrier islands, where sea-water salinities are higher. In general, ice bonding apparently extends to 2 m depth, but in embayments influenced by fresh water we expect bottom-freezing and perhaps anchor ice formation to begin sooner, and to extend seaward beyond the 2-m isobath (fig. 1). Conditions off exposed beaches, such as off Reindeer Island, are different. We observed patchy anchor ice and ice-bonded sand out to 150 m from shore, where the traverse terminated. But during the preceding storm we observed waves breaking as far seaward as 500 m from shore (4-6 m depth on shoals). Within this 500-m-wide surf zone, high turbulence and entrainment of cold air bubbles enhances the heat transfer from the water and seafloor to the atmosphere, We believe that anchor ice and ice-bonded sediments existed throughout this zone at least in patches, but possibly as a continuous sheet (fig. 1). The externally fragile crystal lattice of ice pillows obviously was a product of post-storm ice growth, as was the buildup of rippled, loose sand around the pillows. Flume studies show that such crystal lattice structures can grow by accretion of frazil ice onto an ice-bonded sediment core (Kempema, 1986).

We have no observational basis for extending the occurrence of anchor ice seaward across the shelf beyond the 5-m isobath. However, large amounts of storm-produced anchor-ice form to at least 15 m depth at Molodeshnaya in the Antarctic (Cherepanov and Kozlovskiy, 1972). This anchor ice forms before an ice cover, when the water column is well mixed by fall winds. The conditions therefore seem similar to those in the Beaufort Sea, except that here waters shallower than 15 m are vastly more extensive. One may argue that increased pressure lowers the freezing point and therefore water depth will limit the seaward extent of storm-generated anchor ice. However, such ice forms under strong, cold offshore winds to 17-m depth in Lake Ontario (Foulds and Wigle, 1977). We suggest that in the Beaufort Sea the thickness of the mixed layer, all being supercooled to the same degree, should determine the seaward extent of anchor ice formation. This layer is about 15 m at the end of the summer, and should deepen during storms. We therefore map the probable extent of anchor ice to 20 m in figure 1.

Cherepanov and Kozlovskiy (1972) report that anchor is particularly common on banks and shoals. This observation is supported by the work of Untersteiner and Sommerfeld (1964), who speculate that downward protuberances in the ice canopy are flow obstacles which act as nucleating surfaces for anchor ice. We suspect that the upward protuberances on the middle Beaufort Sea shelf in the form of 5-10 m high shoals should also focus the growth of anchor ice.

The strong bond between anchor ice and substrate holds only during and shortly after formation (Tsang, 1982), as long as the water is supercooled. The growth of a surface ice sheet eliminates the atmosphere as heat sink, and results in a rise of water temperature from the heat released by further ice growth. The sub-bottom-heat conducted upward from the geothermal temperature gradient (fig. 9) probably weakens the

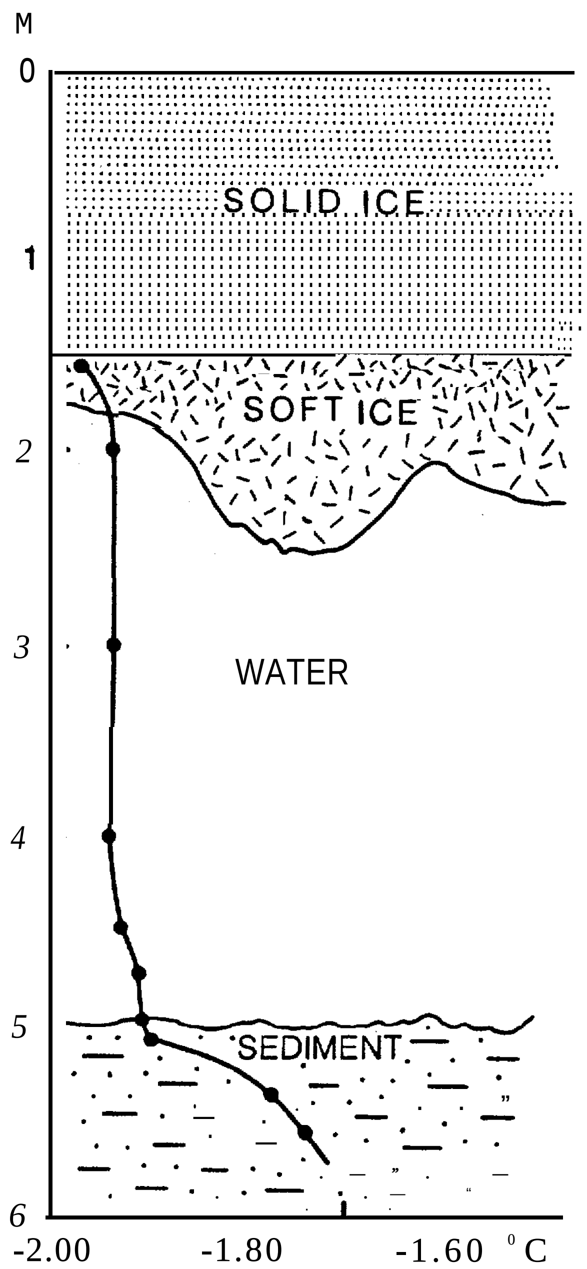

Figure 9. Depth/temperature curve measured in Stefansson Sound on March 6th, 1979. (From T. Osterkamp, written commun., 1979).

strong bond between substrate and anchor ice (Tsang, 1982), to where its buoyancy may carry it to the surface. This porous anchor ice containing some components of the substrate then becomes part of the seasonal ice canopy. Cherepanov and Kozlovskiy (1972) describe the resulting under-ice deposits, which to us seem similar to the billows of soft ice with included sediment observed in different years under the Beaufort Sea ice canopy (fig. 10) (Reimnitz and Dunton, 1979). The disruption of the bond from conduction of sub-bottom heat also explains why storm-generated anchor ice is not observed in winter diving.

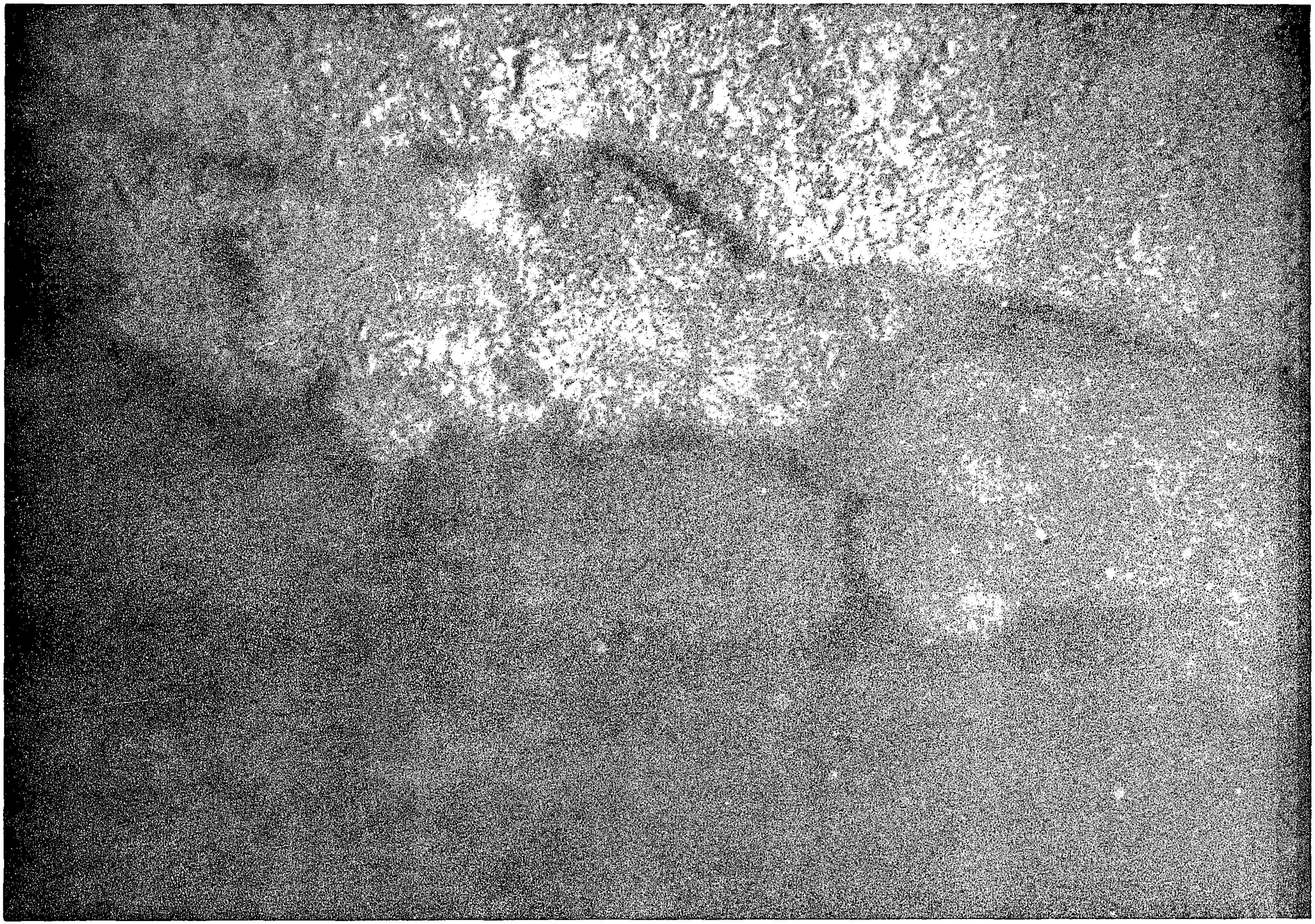

Figure 10. Photograph of highly irregular base of soft-ice laycr in Stcfansson Sound, where depth/tcmpcrmrc profile of figure 9 was measured. Both the relief and large icc crystals arc best explained by an anchor-ice origin. This icc offers littlc resistance to a diver's arm until the congelation ice base, 0.5 to 2 m above is felt, In the background the field of view is 2 to 3 m wide.

The short-lived nature of storm-generated anchor ice makes study of related sediment transport and bedform dynamics extremely difficult. The huge amounts of sediment held by the immobile winter ice canopy in some years (Barnes, et al., 1982) reflect only what rose to the surface once the storm died down. Sediment movement with anchor during the storm near the seabed is probably more important for overall sediment transport than that rafted on the sea surface. The frozen sand slabs we see cast by waves onto beaches probably are moving along the bottom during storms, similar to the ice/sediment masses observed moving along stream beds (Arden and Wigle, 1972; Osterkamp and Gosink, 1982). We commonly observe wind-driven surface currents of 100 ends in the shallow Beaufort Sea. When wave orbital motion is superimposed on such flow, bedload transport of sediment with ice may be significant. But there is a total lack information on the extent and strength of sea-bed ice bonding, which may resist transport during storms. There also is a total lack of information on the kinds of internal sedimental structures produced by anchor ice accretion and by ice bonding of surface sediments.

The ice pillows observed off Reindeer Island may result from post-storm ice nucleation on angular slabs of ice-bonded sand produced during the preceding storm, Transport of slabs from bar crests to the troughs would explain the observed scarcity of ice pillows on the shore-parallel bar and abundance in adjacent troughs. Lastly, our observations of wide-spread ice bonding in shallow regions supports the theory of Sellmann and Chamberlain (1979) that cyclic freeze-thaw may be a cause for dewatering and widespread overconsolidation of surface sediments in the Beaufort Sea.

We believe that the influence of anchor ice and ice bonding of surface sediments during fall storms may be one of the most important phenomena for the sedimentary regime of shallow arctic seas. Twenty five percent of the world's continental shelves are seasonally ice covered, and might be affected by the phenomenon. Anchor ice forms in Eurasia as far south as the Sea of Azov (46°N) (Zubov, 1945), and should be equally widespread in North American seas. The lack of knowledge about anchor ice and frozen seabed thus remains an important gap in our understanding of high-latitude sedimentary environments.

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