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Final Reports of Principal Investigators

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Outer Continental Shelf Environmental Assessment Program

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VOLUME 73 MAY 1991

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SUBSEA PERMAFROST:
PROBING, THERMAL REGIME, AND DATA ANALYSES, 1975–81

by

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I. SUMMARY, CONCLUSIONS, AND IMPLICATIONS FOR OCS DEVELOPMENT

The purpose of this research was to develop a greater understanding of subsea permafrost in order to evaluate problems it poses for offshore oil and gas lease development and production. Our strategy was to obtain reconnaissance-level information on subsea permafrost over a wide geographic area. Analysis and interpretation of these data and their use in relatively simple models have allowed us to make some general statements on the occurrence and distribution of subsea permafrost in Alaska's continental shelf.

Specific problems for petroleum development include differential thaw subsidence; difficulties with seismic data interpretation; geotechnical problems with structures, pipelines and other facilities in warm subsea permafrost; corrosion associated with concentrated brines in seabed sediments; and the dangers associated with decomposing gas hydrates.

Specialized methods and equipment were developed to investigate the properties of subsea permafrost and heat and salt transport processes that occur within it. These include methods and equipment for lightweight drilling, temperature measurements, in situ sampling of soil pore water, pore water pressure measurements, detection of the presence of ice in the sediments, and measurements of the thermal and hydraulic conductivities of the sediments. A general discussion of the results from our study areas follows.

Norton Sound.—Subsea permafrost does not occur at Nome or on the south side of Norton Sound, except possibly in nearshore areas where rapid coastal retreat occurs. A more detailed analysis taking into account cold, summer seabed temperatures and, possibly, greater-than-normal heat flow, will be required to assess the potential occurrence of subsea permafrost in the main portion of Norton Sound.

Northern Bering Sea and Bering Strait.—This area was not studied in the course of our investigations. Limited oceanographic measurements exist, but the data have not yet been used to estimate the potential for the presence of subsea permafrost in these areas.

Kotzebue Sound.—It appears likely that subsea permafrost occurs under Kotzebue Sound, particularly in shallow nearshore regions. The discharge of the Noatak and Kobuk rivers into the northern port of Kotzebue Sound may affect the subsea permafrost through warmer seabed temperatures and reduced under-ice water salinities. Fresher water found under the ice during winter at Kotzebue could be a source for producing potable water for OCS related development activities.

Chukchi Sea.—In the southeastern Chukchi Sea, subsea permafrost exists in the shallow, nearshore areas. Oceanographic data and thermal models suggest that subsea permafrost may be absent beneath deeper water although it may well occur in areas out to depths of 15–30 m. At Barrow, the seabed temperatures are slightly negative and it appears that ice-bearing subsea permafrost occurs there only very close to shore. Bedrock occurs close to the seabed between Peard Bay and Cape Thompson. This rock cannot be assumed to be a good foundation material since it may contain segregated ice. Trenching of the rock for subsea pipelines would be difficult. Gravel sources for OCS development appear to be relatively sparse and available in localized areas only.

Elson Lagoon, Lonely, and Harrison Bay.—The sediments in these areas are generally fine-grained. A thawed layer under the seabed of generally increasing thickness can usually be traced out from shore. However, the thawed layer is still relatively thin (10-15 m or less) out to water depths of about 15 m and distance from shore of about 20 km or more. The proximity of ice-bearing sediments to the seabed and the high potential for encountering segregated ice in them suggests that subsea permafrost problems for offshore development are likely to occur. Thaw subsidence problems for bottom-founded structures and pipelines in the seabed could be potentially serious in these areas.

Long Island.—The seabed sediments consist of a layer of fine-grained material, probably clay, which thins seaward and is underlain by thawed gravels. The clay at the seabed may be somewhat difficult to trench.

West Dock–Reindeer Island.—Most of our research was concentrated along this line. There is a thawed layer about 3–4 m thick out to about 400 m offshore. Part or all of this layer freezes seasonally and contains highly concentrated brines. The ice-bonded permafrost table deepens rapidly in a transition zone from about 400–440 m offshore. From about 440 m to about 3.4 km or more offshore, the ice-bonded permafrost table deepens gradually as the square root of distance (or submersion time) offshore. From there to Reindeer Island, seismic data suggest that the ice-bonded permafrost table is deeper than given by this functional relationship, possibly because of the presence of a paleo-river channel.

The sediments within a few kilometers offshore are sands and gravels containing some fines and are capped by a layer of fine-grained material up to about 5 m in thickness. Beyond Reindeer Island, the sediments are fine-grained with overconsolidated clays found at the seabed 15–20 km from the mainland. In this area, ice-bonded permafrost occurs typically from 5 to 15 m below the seabed.

The presence of ice-bearing, and usually ice-bonded, subsea permafrost close to the seabed near shore suggests that special methods will be required to bring offshore pipelines on shore. These pipelines will have to transect the area where the permafrost may be thawing at the rate of 0.3 m/yr and the potential for encountering segregated ice is high.

Beyond Reindeer Island the potential for encountering segregated ice in the sediments is also high. The very stiff clays near the seabed will also be difficult to trench and may hinder access to underlying gravel sources.

Point Brower–Jeanette Island.—The sediments immediately offshore of the Sagavanirktok River are fine-grained, with ice-bonded permafrost found from 5 to 25 m below the seabed. We found what appeared to be segregated ice about 16 m below the seabed about 1.6 km off Point Brower. A layer of massive ice, 0.6 m in thickness, was also found in this area by the U.S. Geological Survey (USGS) at a depth of 10 m below the seabed at a point about 5.5 km offshore. These observations show that offshore development in areas of fine-grained soils, where the ice-bearing permafrost is close to the seabed, must consider the potential problems associated with ice-rich permafrost, its thawing, and subsequent thaw settlement.

Flaxman Island.—The sediments near the seabed are fine-grained and sometimes hard. South of Flaxman Island the seabed sediments are underlain by gravel. Ice-bonded permafrost was found near the seabed north of the island. Sediment temperatures are relatively cold and suggest that the island may have retreated from these sites recently.

Barrier Islands.—Holes were drilled on Thetis, Cottle, Stump, Reindeer, Cross, and Flaxman islands. Ice-bonded permafrost was found under the islands but the state of ice-bondedness varies in a complex fashion laterally and with depth. Layers of unbonded material suggest that gravel islands may have planes of weakness. Gas of biogenic origin was found in one shallow hole on Flaxman Island and could be a drilling hazard.

Substantial progress has been made in our understanding of subsea permafrost during the past decade. However, each new field effort uncovers new and unsuspected facts, suggesting that the database is inadequate. Progress toward an understanding of the heat and salt transport processes responsible for the response of subsea permafrost to natural or man-made disturbances has been inadequate. The present program of offshore development and the results presented here and elsewhere show that subsea permafrost must be considered if offshore petroleum fields are to be brought into production. Sound engineering practice and environmental considerations will require an expanded database and an understanding of the factors controlling the evolution of subsea permafrost.

II. INTRODUCTION

This is the summary report of a program to study the distribution and properties of subsea permafrost in Alaskan waters. Our specific objectives were:

1) To obtain reconnaissance-level information on subsea permafrost conditions over a wide geographic area.
2) To investigate conditions at the seabed relevant to the evolution of the underlying permafrost regime.
3) To study the heat and salt transport processes in subsea permafrost with the eventual goal of predicting its response to natural or man-made conditions.

Field measurements included soil type, temperature, electrical conductivity of the soil pore water, thermal and hydraulic conductivity, in situ ice detection, and pore water pressure.

When this program began in 1975, very little was known about subsea permafrost in Alaska's offshore areas except near Barrow, where navy support had been available (Brewer 1958; Lewellen 1975; Rogers et al. 1975), and for limited proprietary industrial data. This study, related OCSEAP studies beginning the following year, and a USGS (Conservation Division) study in 1979 (USGS 1979; Miller and Bruggers 1980) provided the first offshore data from many regions, including Prudhoe Bay. In the last few years, subsea permafrost data have been obtained by the petroleum industry as part of geotechnical investigations for offshore pipelines and structures. Most of this information is proprietary, and the data obtained by the OCSEAP and USGS studies still compose the bulk of public domain information.

In keeping with the objective of obtaining wide geographic coverage, most of the data from our program were obtained with lightweight equipment, and, except in 1975, few soil samples were obtained. A great deal of information was nevertheless obtained from relatively shallow holes, typically 10–30 m but up to 50 m deep, using the sea ice as a working platform. Our data and detailed descriptions of our methods and results are described in our eight annual OCSEAP reports and in our other published papers and reports listed in Appendix A, covering field observations from 1975 to 1981. They are available from World Data Center A; Glaciology, Cooperative Institute for Research in Environmental Sciences, University of Colorado, Boulder, CO 80309. OCSEAP reports up to 1980 are available from the National Technical Information Service (NTIS), Springfield, VA. Corrections to these reports are contained in Appendix B. This summary is merely an overview; the OCSEAP reports must be referred to for most of the details and data. The overview is restricted to our results; some reference is made to related OCSEAP studies and the USGS program, but a full synthesis was not attempted. A synthesis of conditions in the Alaskan Beaufort Sea has been given by Sellmann and Hopkins (1984). The most relevant of the OCSEAP studies were a CRREL–USGS drilling and probing program in 1976 and 1977 (Hopkins and Hartz 1978a,b; Blouin et al. 1979; Chamberlain 1979; Sellmann and Chamberlain 1979); a seismic program carried out from a small boat (Rogers and Morack 1982; Morack and Rogers 1984); interpretation of industry seismic data (Sellmann and Neave 1982; Neave and Sellmann 1984); mapping of onshore conditions (Hopkins and Hartz 1978a,b; Smith et al. 1980; Smith and Hopkins 1982); and a study of shoreline erosion rates (Lewellen 1977). Additional information on OCSEAP research in the Beaufort Sea is given in Barnes et al. (1984).

This report contains, in addition to the permafrost information, a summary of an ad hoc study of sediments in sea ice (Appendix C).

III. RELEVANCE TO PROBLEMS OF PETROLEUM DEVELOPMENT

A list of some of the information gaps on subsea permafrost and problems posed to offshore development was compiled, with industry input, by the National Academy of Sciences (NAS) (1976). Another version was produced by OCSEAP at the Beaufort Sea Synthesis Meeting in 1977 (Weller et al. 1978). Although there have been some changes with increasing experience, these reports are still relevant. The NAS list is reproduced here with minor changes:

1. Lack of information on the horizontal and vertical distribution and properties of subsea permafrost.
2. Differential thaw subsidence and reduced bearing strength due to thawing of ice-rich permafrost.

a. Thaw subsidence around well bores causing high down-drag loads on the well casing. This problem is aggravated offshore by the need for drilling several holes from a single island and by the warm permafrost temperatures and salty pore water.

b. Differential settlement associated with hot pipelines, silos in the seabed, and pile and gravity structures.

c. Differential strain across the phase boundary between bonded and unbonded permafrost.

3. Frost heaving in very shallow water.

   a. Well bore casing collapse due to freeze-back.

   b. Pipelines—differential movement.

   c. Gravity structures including gravel islands—local heaving causing foundation instability.

   d. Pile structures—differential stress in pile-founded structures.

4. Seismic data in areas of subsea permafrost, particularly when the latter has a variable thickness and variable properties, can be misinterpreted and can lead to improper design of offshore production and distribution facilities. Special care needs to be taken in the interpretation of seismic exploration data.

5. Excavation—dredging, tunneling, trenching.

   a. Increased strength of material associated with bonded sediment.

   b. Over-consolidated sediment can influence excavation rates and approach.

   c. Thaw can be induced in deeper sediment by removal of material at the seabed.

   d. Highly concentrated and mobile brines can be found in the sediment.

   e. Insufficient data on engineering properties for design of excavation equipment and facilities.

6. Gas hydrates.

   a. Blowouts can result from gas hydrate decomposition during drilling operations.

   b. Fire danger.

   c. Misinterpretation of seismic data and other geophysical data.

7. Corrosion: Fluids (brines) with concentrations several times higher than is normal in seawater are common, particularly in water less than 2 m in depth.

The original version of this list concludes with a number of precautions and possible consequences if they are not heeded. It notes that detailed site specific information will be required for any development, especially in view of the highly variable nature of subsea permafrost, and that down-hole information at the exploratory drilling stage should be used as much as possible. To date, the latter recommendation has not been followed. From the beginning, a basic concern has been that methods developed for petroleum production and transportation onshore at Prudhoe Bay may not be adequate, because subsea permafrost is warmer, saltier, and more easily disturbed, and because the lithology may be different.

To the original list of special problems associated with subsea permafrost, several more may be added on the basis of our more recent experience. These include the presence of gases in shallow sediments, premature pile failure in salty sediments (Nixon and Lem 1984), and, perhaps most important, the design of insulated pipelines, the optimum design of which requires more information about salt transport mechanisms than is presently available.

IV. METHODS

The measurements made during our studies included soil type, temperature, electrical conductivity of the soil pore water, hydraulic and thermal conductivities, in situ ice detection, and pore water pressure. Although a standard rotary wash boring rig was used in 1975, in other years most of the equipment and many of the techniques were unconventional and had to be developed as part of the program. Most of the details are given in our OCSEAP reports, and by Osterkamp and Harrison (1981b) and Harrison and Osterkamp (1981a,b).

Our strategy was to use the sea ice as a drilling platform, sometimes using snow walls or tents as windbreaks. Transportation to the sites was by fixed wing aircraft, helicopter, snow machine, or hand-drawn sled. Usually the sites were located by distance and orientation with respect to some fixed and prominent onshore feature. Distance was measured by pacing, string measure, tape measure, or by electronic distance measuring equipment. Usually, orientation was obtained by Brunton compass, but in some cases, a 1-second theodolite was used. Some holes, usually those farther offshore, were located using the Global Navigation System of the NOAA helicopters.

Data on snow depth, ice conditions, ice thickness, freeboard, water depth, and bottom conditions were usually taken but have not been consistently reported in our OCS reports. These supplemental data are available from the authors.

The reconnaissance measurements emphasized by this program required methods for gaining access to subsea sediments with lightweight equipment. Two techniques were used: driving and rotary-jetting. These approaches do not permit the acquisition of the detailed soils data that can be obtained with a drill rig, but they are actually better suited than drilling for certain in situ measurements (such as temperature, pore water pressure, and thermal and hydraulic properties), and for sampling of the pore water in coarse sediments. Shallow soil samples can be obtained, and, with experience, lithologic information can be inferred. The equipment evolved over several years of field experience.

In the driving technique, a portable motorized cathead and tripod and 74-kg drop hammer are used to drive drill rods with attached probes into the seabed. The depth capability depends upon soil type and is optimum for the relatively coarse sediments typical of Prudhoe Bay, where the maximum depth reached was 26 m (determined by our available supply of drill rod). Good results were also obtained with a portable gasoline-powered driver similar to those used for breaking concrete.

In the rotary-jetting method, a 3/4-inch steel water pipe drill string with a bit and a one-way valve at the bottom of the string is rotated and jetted into the seabed using a small gasoline-powered pump and drill. The depth capability depends on soil type and is optimum for fine-grained soil conditions, where depths in excess of 50 m have been reached. When the sidewalls of a hole are stable, the pipe string can be "washed" down without rotation. The types of data that can be obtained with the driving and rotary-jetting techniques are briefly described in the following paragraphs.

Soil Conditions

Information about subsea soil conditions can be obtained from both the rotary-jetting and driving techniques, even when soil samples are not taken. The experience of the driller, as in any type of drilling, is extremely important, and drilling in known material was a factor in building up our experience. In rotary-jetting, the "feel" and sound of the drill are different in clay, silt, sand, and gravel. Compact clays feel like hard rubber when the string is dropped onto the bottom of the hole; bonded materials feel more like concrete and usually drill very slowly. Unbonded silt and silty sands usually drill rapidly, as long as the hole walls do not cave, which may happen if too much sand is present. The grains of sand and gravel can be detected when raising and lowering the string off the bottom, and by the roughness of the drilling. Gravels are extremely hard to drill by rotary-jetting, due to loss of circulation, caving, and the difficulties of flushing larger soil particles from the hole. Direct identification of soil type is possible when soil cuttings are washed to the surface, as occurs on land or where the sea ice is frozen to the seabed. Wood, shells, and other items may be recovered in this way, although one must realize that they may not always come from the bottom of the hole but may be washed out of the walls.

The driving techniques work best in the soils that are the most difficult to jet, sands and gravels, which can often be identified by the rapid progress of driving. On some occasions clay can be identified when it comes up stuck on the drill rod. By and large, soil identification is less positive than with the rotary-jetting technique. Both techniques can detect the presence of firmly ice bonded sediments, which can be penetrated by the jetting technique but not to any significant depth by the driving technique.

Temperature

Both techniques are well suited to provide access holes for temperature measurements. In the rotary-jetting technique the pipe is left in the hole. To prevent freezing, the pipe is normally installed with a check valve at the bottom, and upon hole completion a non-freezing fluid is pumped through it. In the driving technique, temperature can be measured inside the drill rod, but this is inconvenient, since the rod has to be left in place until the temperature measurements are completed. In 1978, a new technique was developed, wherein a continuous length of 12.7 mm O.D. tubing is run inside the drill rod, and attached to a suitable driving point at the bottom. When the hole is completed the drill rod is removed and the point and tubing remain behind, providing access for temperature measurements. This frees the drill rod, and the equipment needed to remove it, for immediate use elsewhere while temperatures are equilibrating. The undisturbed or "equilibrium" temperatures can usually be accurately estimated after a few days, depending upon the time spent in drilling and other factors, particularly any loss of drilling fluid. Temperature is logged inside the pipe or tubing with a single thermistor on a cable. Information about the approach to temperature equilibrium, accuracy, and thermistor calibration are given by Osterkamp and Harrison (1976a, 1982b). Other aspects of temperature measurements are discussed in a general way by Osterkamp (1985).

Soil Pore Water Electrical Conductivity

The electrical conductivity of the soil pore water is important in determining the presence or absence of ice, and in giving clues to the past history of the soil. Therefore, considerable effort was devoted to developing techniques for determining the pore water conductivity by using the driving equipment. NSF support was used for this phase of the work. The water is admitted into plastic tubing run inside the drill rod through a porous metal filter at the bottom, and sampled inside the tubing with a special bailer. The tubing can be cleared and a valve closed at its lower end before the rod is driven to the new sampling depth. It is not necessary to pull out the drill pipe between taking samples.

The method has an interesting limitation if the freezing temperature of the pore water is higher than the in situ temperature. Liquid will still be collected, but it will not be representative of the soil bulk H_2O conductivity, because the solid phase fraction of the H_2O in the soil is not collected. Therefore, when the measured freezing temperature of the collected liquid is equal to the in situ temperature, one can only conclude that the soil bulk H_2O freezing temperature is higher than or equal to the in situ temperature, and that ice may be present in the soil under these conditions.

Hydraulic Conductivity

The saturated hydraulic conductivity (essentially the permeability) of the soil is calculated from the rate at which water from the soil enters the filter, which is determined with a water level sensor. The exact calculation is extremely difficult for some of the geometrically complicated shielded filters that we used, but we were able to derive a simple and reasonably accurate approximate method (Harrison and Osterkamp 1982). However, it was noticed that the incoming water rarely obeyed Darcy's law.

Thermal Properties and the Detection of Ice

Past experience in the interpretation of temperature profiles in fine-grained subsea permafrost and in warm, saturated, fine-grained subaerial permafrost showed that it can be difficult or impossible to detect the position or presence of ice-bearing permafrost soils from temperature profiles alone. The difficulties appear to be associated with the presence of unfrozen pore fluids which lead to a variable ice content or ice bonding at temperatures near, but less than, the equilibrium freezing point of the pore fluids. Our drilling methods in subsea permafrost had the same difficulties since they depend on a certain degree of ice bonding to detect the presence of the ice. In addition, since we did not take soil samples at depth, it was desirable to have a technique to determine soil thermal properties in situ. These considerations led us to design a borehole heating experiment for the purpose of detecting ice-bearing permafrost and to determine the thermal conductivity, K, of subsea permafrost soils in situ.

In practice, the borehole heating experiment is similar to the probe method for measuring the thermal conductivity of materials (Bullard 1954; Lachenbruch 1957; Carslaw and Jaeger 1959). The "probe" in our borehole heating experiment was either 1/2" O.D. plastic tubing (3/32" wall) or 3/4" schedule 40 black iron water pipe installed in our boreholes for temperature logging purposes. The tubing was filled with antifreeze and the pipe with a mixture of antifreeze and seawater. A 5-kW generator was used to heat a length of 2-conductor cable which was placed inside the tubing or pipe in the borehole. Current and voltage were monitored periodically during the time of heating, which varied from 15 to 120 minutes. The boreholes were logged just prior to the time of heating and at selected times after heating, usually over a time period of a few days or less. Prior to heating, the boreholes were generally within a few hundredths C of equilibrium. The heat dissipated in the boreholes was relatively small. Where ice-rich permafrost was present, it was probably sufficient to melt the ice in the sidewall of the borehole to a depth of about 0.01 m.

It is usually possible to determine the depth distribution of ice-bearing permafrost in the borehole by comparing temperature profiles measured before and just after heating. With ice present in the soil, the latent heat required to melt it tends to buffer the temperature against changes. Comparison to an adjacent depth not containing ice often shows a strong temperature difference across the ice-bearing boundary.

The in situ determination of the thermal conductivity of unbonded (non–ice-bearing) subsea permafrost follows the same procedure as used in standard probe measurements. For a continuous linear heat source of constant strength in an infinite homogeneous medium the temperature disturbance at a time t is (Lachenbruch 1957)

$$\Delta T = (P/4\pi K) \ln (t/t{-}s)$$

where s is the period of heating and P is the power dissipated per unit length, which can be calculated from the current and voltage measurements and the length of the heating wire. This equation neglects the finite size of the borehole (pipe, water, tubing, fluid, and surrounding disturbed volume of soil), and effects associated with the proximity of the depth of measurement to the seabed or hole bottom, or with layers of varying thermal properties. It is possible to evaluate the associated errors and/or to modify the equation to take them into account (Lachenbruch 1957; Carslaw and Jaeger 1959). The usual approach is to graph ΔT vs. ln t/t–s and to select the straight line portion of the curve, which has a slope of $P/4\pi K$ and thus determines K. In our measurements, an error analysis showed that the total error in these determinations of K appears to be about 0.3–0.5 W $(C\ m)^{-1}$, and was dominated by the error in power determination.

The addition of heat to ice-bearing permafrost usually results in some melting, basically because the material is often warm and contains salt. The associated latent heat effect may make it difficult or impossible to determine K by this method. Ideally, the temperature rise to determine K by heating the borehole should be held to a very small value to minimize the latent heat effect.

The analysis of these borehole heating data appears to be relatively straightforward in some of the holes and very complex in others. Detection of the presence of ice-bearing permafrost seems possible provided the boreholes are logged immediately after heating. Determination of K appears to be a more complex matter, especially in ice-bearing permafrost. Unfortunately, these two type of measurements are also somewhat at odds since substantial heating seems to be desirable to detect the presence of ice-bearing permafrost, while very small heating is desirable to minimize latent heat effects when trying to determine K.

Pore Water Pressure Measurements

Pore water pressure measurements were made with a pneumatic pressure transducer (Sinco model 51481 with Model 51411A readout). Several of these piezometers were calibrated by us and were found to meet the manufacturers' specifications (typically 0.03 m [water head] sensitivity, 0.1% linearity, and 0.2 m offset) after some factory calibration data reduction errors were corrected. However, we had unsatisfactory results when trying to use the system in very cold weather. The transducer is mounted inside a piece of A-rod, and connected to the probe assembly of the pore water sampling equipment (Harrison and Osterkamp 1981) with a short piece of rubber tubing. The probe assembly contains a filter through which the transducer couples to the soil pore water. The tubing and probe assembly are completely filled with antifreeze, and the A-rod driven into the seabed sediments. Pressure is read sequentially at increasing depths down to the ice-bonded phase boundary. Equilibration times may vary from a few hours to a few seconds. Tide is determined concurrently by measuring the water level in a borehole through the sea ice with respect to a mark on a string of A-rod driven into the seabed to provide a stable reference.

V. REGIONAL RESULTS AND DISCUSSION

The primary purpose of our OCSEAP project was to obtain reconnaissance-level field information on subsea permafrost conditions over a wide area on the continental shelves of arctic Alaska. From this point of view, the data described below speak for themselves. However, considering the incredible variability of the subsea permafrost regime, it seems worthwhile first to gain some perspective by a brief discussion of some of the processes responsible for its formation and evolution.

Subsea permafrost is a product of changing sea levels over geologic time and of past cold climates. It forms in exposed continental shelves at times of lower sea level, under low air temperatures, by growth from the exposed surface downward. Subsequent rises in sea level and shoreline erosion at times of static sea level cover the permafrost with seawater, thus replacing the cold surface boundary condition with a relatively warm and salty one. Melting from the surface (seabed) downwards is controlled by salt and heat transfer to the ice-bearing subsea permafrost table. Melting at the base of the subsea permafrost, by geothermal heat, begins once the thermal disturbance (of submergence) has propagated to the base. The time scales for permafrost growth are such that several hundreds of meters of permafrost can be formed in the continental shelves while they are exposed. The time scales for thawing at the permafrost table and base are such that several tens of thousands of years may be required to completely thaw subsea permafrost a few hundred meters thick. Subsea permafrost is essentially a transient phenomenon, and it must have formed and thawed repeatedly over geological time. At present, sea levels are very high.

Melting at the subsea permafrost table, which occurs even in the presence of negative mean annual seabed temperatures, is determined by lithology and local oceanographic conditions. This melting is extremely sensitive to the efficiency of salt transport, which is apparently controlled by soil type. Other soil properties that play a role in the melting process include ice content, thermal properties, and possibly hydraulic conductivity. Oceanographic conditions that play a role in this melting process include seabed temperature and salinity, which are in turn influenced by sea ice cover, bathymetry, presence of nearby rivers, and currents. Sea level and shoreline history are of particular importance since they determine the time of emergence and submergence. The initial conditions, at the time of submergence, are set by the local meteorological and permafrost conditions.

This section summarizes the results of investigations of the permafrost regime in shallow drill holes (usually < 30 m depth) generally in relatively shallow, nearshore waters. Most of the drill holes were within a few kilometers of the coast, although some were up to 20 km offshore. A location map is provided for each area along with a table of information on each hole drilled.

Norton Sound

Five holes were drilled in Norton Sound during March 1980 at the locations shown in Figure 1; the drilling data are given in Table 1.

There is considerable uncertainty in our knowledge of the parameters and conditions needed to assess the presence and distribution of subsea permafrost in Norton Sound, not only over the long time scales involved, but also the spatial variations. In addition, subsea permafrost in the general area of the Bering Sea appears to be a near-borderline case so that an assessment of its potential presence is especially difficult and depends strongly on the assumptions made in that assessment.

The sediments in Norton Sound are primarily silts deposited by the Yukon River (Nelson and Hopkins 1972). Geothermal heat flow in the Norton Sound area may be anomalous. Our four drill holes on the south side of Norton Sound are near the Kaltag fault. An unvegetated lava flow was observed east of St. Michael and several lakes along the coast south of St. Michael were partially free of ice at the end of March 1980. A number of sidehill icings were observed east of Nome and there are several hot springs on the Seward Peninsula. These observations suggest that the geothermal heat flow may be greater than normal in Norton Sound. Biswas and Aki (personal communication) consider the heat flow in the northern part of the Bering Sea, including Norton Sound, to be about 3 HFU or about

Table 1.—Norton Sound drilling data.

Hole designation	Location or distance offshore	Water depth (m)	Sea ice thickness (m)	Drilling method	Date of drilling	Approximate depth below seabed (m)
Nome	N208°E from benchmark on Anvil Mt., 326 m from shore, along this line, 299 m offshore	3.70	0.82	driving	3/25/80	6.8
St. Michael River	About 8.7 km SW of mouth of St. Michael River	2.00	1.10	jetting	3/28/80	25
Charley Green Creek Hole 630 (Hole 1)	Near mouth of Charley Green Creek, about 630 m from shore	2.44	1.16	jetting	3/28/80	6
Hole 630 (Hole 2)	Same, but about 610 m from shore	1.69	1.16	jetting	3/30/80	7
Yukon Delta	63°27.2′N; 163°36.7′W, 28 km from shore	10.30	1.05	jetting	3/31/80	18

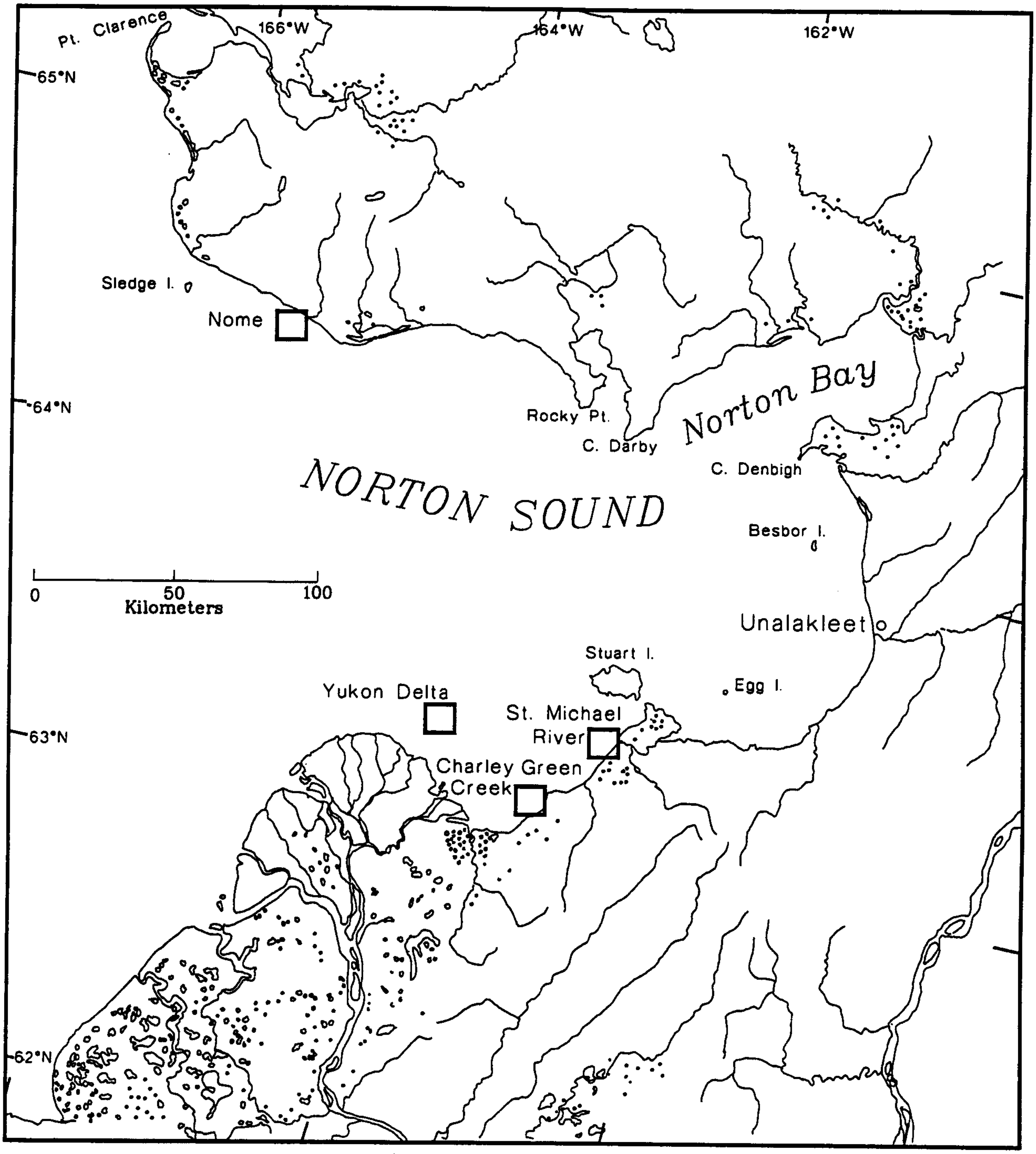

Figure 1.—Norton Sound hole location map.

twice the worldwide average. Therefore, basal melting of the permafrost after submergence may be expected to be greater than normal, perhaps several centimeters per year.

Muench and Coachman (1980) found that the eastern portion of Norton Sound contained a two-layered water structure during the summers of 1976, 1977, and 1978 with the bottom layer being colder, more saline, and denser than the upper layer. Bottom

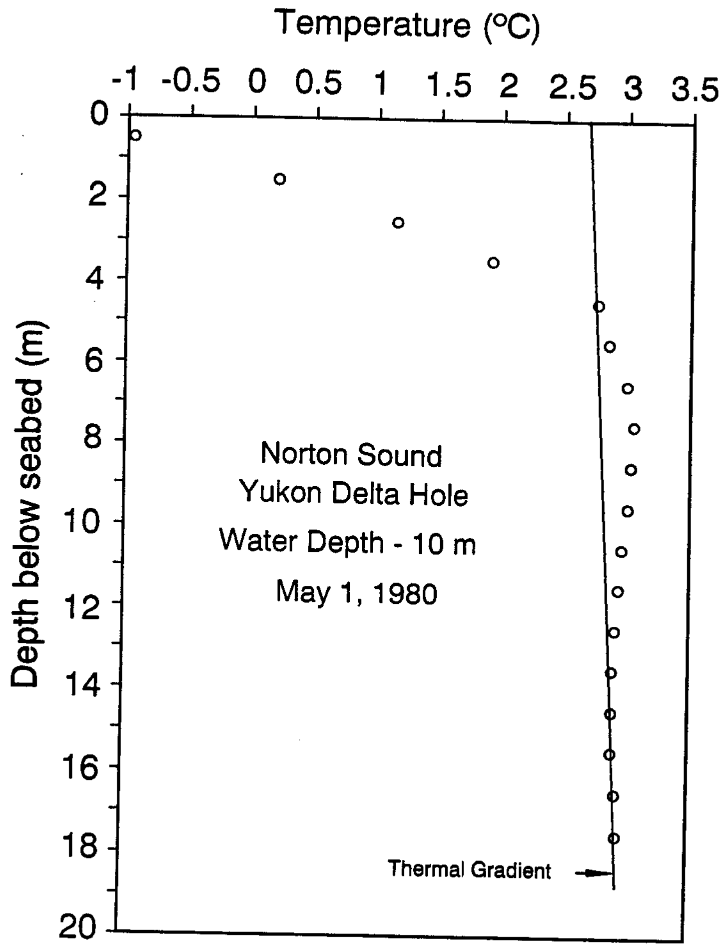

Figure 2.—Temperature profile in the Norton Sound Yukon Delta hole showing the warm seabed sediments and positive temperature gradient near the bottom of the hole.

temperatures on the north side of Norton Sound were colder than on the south side. During July 1977, near Cape Darby, bottom temperatures < 0°C were observed, which warmed to 2–3°C by late August. This suggests that the mean annual seabed temperature (MAST) could be < 0°C in this area. Bottom temperatures on the south side of Norton Sound were several degrees or more warmer. The persistence of this layering over time scales of decades to millennia is unknown.

The temperature data at Nome and on the extreme south side of Norton Sound show that the sediments there are very warm and that the temperature gradients appear to be positive. The temperature profile shown in Figure 2 illustrates the warm seabed temperatures in the sediments on the south side of Norton Sound.

We conclude, on the basis of very sparse data and a preliminary analysis, that subsea permafrost does not occur at Nome nor on the south side of Norton Sound. A more detailed analysis would be desirable. A possible exception for the occurrence of subsea permafrost would be in nearshore areas of rapid coastal retreat as suggested by Hopkins (1980). In the Cape Darby area, we believe that ice-bonded subsea permafrost is also absent except in nearshore areas of rapid coastal retreat. These statements are based on assumptions of the maximum expected permafrost thickness of about 125 m at the time of submergence about 10,000 years ago, a geothermal heat flow capable of melting 2–3 cm per year at the permafrost base, and the assumption that there will be some salt in the sediment pore water.

Northern Bering Sea

We did not take any measurements in the northern Bering Sea from Nome to the Bering Strait. Oceanographic measurements (Coachman et al. 1975) show that a relatively cold bottom water area exists to the southwest of St. Lawrence Island (about –1.5°C during midsummer). The Bering Strait has a mean annual water temperature near the seabed of about 0°C. The sediments along the coast are usually sandy and gravelly and a submerged belt exists, as much as several kilometers wide, where rocks are exposed at the seabed or lie below a few meters of the coarse-grained sediments (Nelson and Hopkins 1972; McManus et al. 1977). Surficial sediments on the northern Bering Sea floor are predominantly poorly sorted, silty, very fine sand, although gravel and coarser sand are common. The mean annual air temperature (MAAT) at Wales is near –6°C, which is several degrees colder than at Nome (about –3.4°C). Therefore, it is expected that, other factors being equal, the maximum permafrost thickness at Wales would be correspondingly greater, perhaps 200–250 m. However, the water is deeper (about 50 m) and this implies a different history of submergence compared to Norton Sound. The above noted factors have not yet been applied to estimate the potential for the presence of ice-bearing subsea permafrost in the northern Bering Sea.

Kotzebue Sound

Our database in Kotzebue Sound is very limited (Figure 3 and Table 2). It consists of five seabed temperature measurements and poorly defined temperature profiles from two shallow holes close to shore; the data from one hole are shown in Figure 4. These holes were drilled during April 1977. The temperature gradients in the holes are both negative, suggesting the presence of ice-bearing permafrost at depth. Seabed temperatures in the main portion of Kotzebue Sound were near normal for winter conditions (about –1.8°C).

Therefore, it seems likely that ice-bearing permafrost exists under Kotzebue Sound, particularly in shallow nearshore regions. The discharge of the Noatak and Kobuk rivers into the north side of Kotzebue Sound produces warmer seabed temperatures and reduced under-ice water salinities. These factors may have a strong impact on subsea permafrost on the north side of Kotzebue Sound. Fresher water under the ice during winter could be a source for producing potable water for OCS development activities.

Table 2.—Kotzebue Sound–Southern Chukchi Sea drilling data.

Hole designation	Location or distance offshore	Water depth (m)	Sea ice thickness (m)	Drilling method	Date of drilling	Approximate depth below seabed (m)
Rabbit Creek	75 m	4.0	1.2	driving	4/26/77	18
Kotzebue	310 m	1.8	≈1.2	jetting	4/28/77	25
Cape Blossom	≈300 m	1.37	1.07	jetting	4/27/77	10

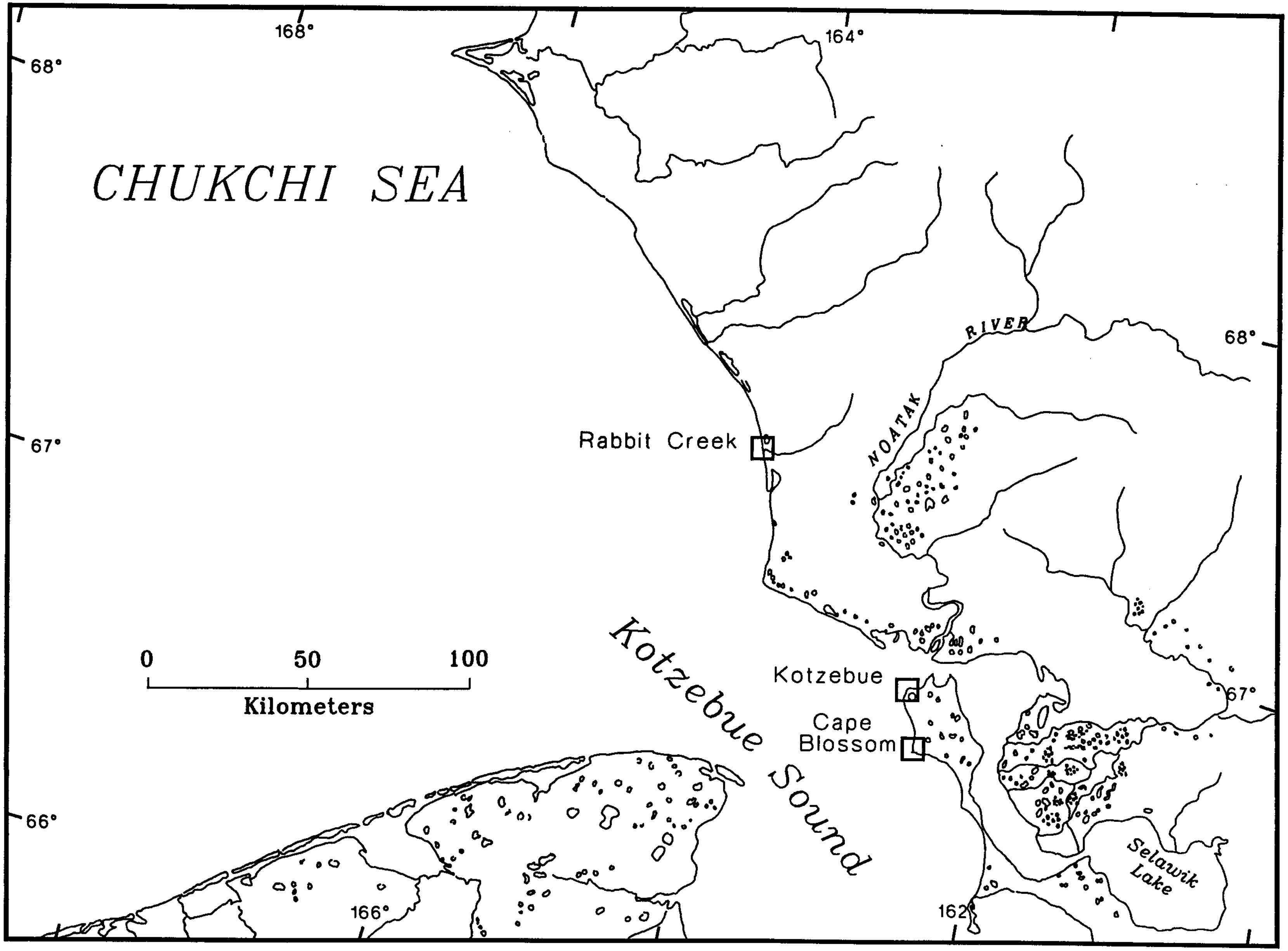

Figure 3.—Kotzebue Sound hole location map.

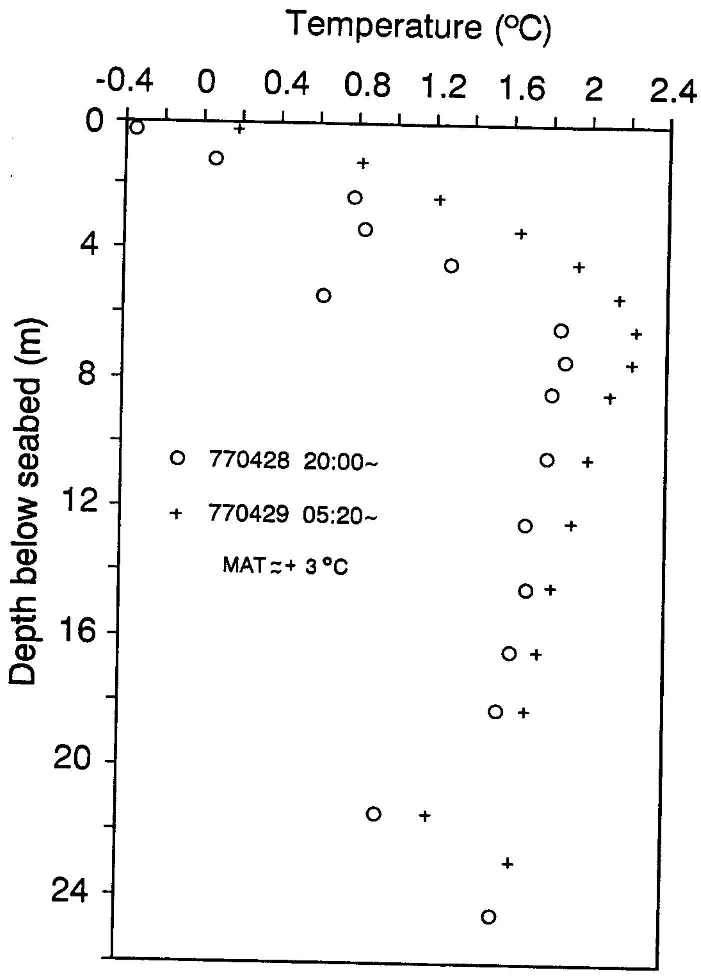

Figure 4.—Temperature profile in an offshore hole off the end of the airport at Kotzebue. The sediments were thermally disturbed but the profile shows a negative temperature gradient suggesting the possible presence of ice-bearing subsea permafrost at depth.

Chukchi Sea

Five holes were drilled in the Chukchi Sea at the locations noted in Figure 5 and Table 3 during the 1977, 1978, and 1979 spring field seasons. Attempts were also made to drill holes at the additional sites shown in Figure 5 but were unsuccessful because rock was encountered at or near the seabed. Rock, at or near the seabed, appears to be common between Peard Bay and Cape Thompson. It was not possible to penetrate the seabed at these sites. The holes at the Naval Arctic Research Laboratory (NARL) (Figure 6) show thermal evidence for a long-stable shoreline, and probably any ice-bearing permafrost there exists only very close to shore and is due to the proximity of the cold emergent land. The situation is quite different in the southeastern Chukchi Sea, where the holes show decreasing

Table 3.—Northern Chukchi Sea drilling data.

Hole designation	Location or distance offshore	Water depth (m)	Sea ice thickness (m)	Drilling method	Date of drilling	Approximate depth below seabed (m)
NARL						
Hole 27 (Hole A)	On a line bearing N330°E from sea ice radar mast at NARL. 27 m from shore	4.12	1.80	rotary-jet	4/18/80	6
Hole 78 (Hole B)	Same, but about 78 m from shore	5.0	1.75	rotary-jet	4/19/80	29
Hole 690 (Hole C)	Same, but 690 ± 100 m from shore	6.5	1.70	rotary-jet	4/21/80	22
Hole 705	705 m offshore	6.57	1.63	jetting	5/15/77	16
Peard Bay	In line with runway near Tachinisok Inlet. About 500 m from shore	6.5	rafted?	rotary-jet	4/23/80	2.0
Wainwright						
Hole 265	70°49.7′N; 159°31.2′W	4.75	rafted?	rotary-jet	4/24/80	17
Cape Lisburne						
Hole 1	075° magnetic from Cape Lisburne radome at ≅25 miles	6.10	1.50	driving	3/27/81	No penetration, rock
Hole 2	048° magnetic from Cape Lisburne radome at ≅32 miles	17.16	1.51	jetting	3/28/81	No penetration, rock
Hole 3	065° magnetic from Cape Lisburne radome at ≅37 miles	17.16	1.51	jetting	3/28/81	No penetration, rock
Hole 4	357° magnetic from Cape Lisburne radome at ≅22 miles	30.48	0.40	jetting	3/29/81	No penetration, rock

Table 3.—(continued).

Hole designation	Location or distance offshore	Water depth (m)	Sea ice thickness (m)	Drilling method	Date of drilling	Approximate depth below seabed (m)
Cape Lisburne						
Hole 5	005° magnetic from Cape Lisburne radome at ≅10 miles	21.03	0.72	jetting	3/29/81	No penetration, rock
Hole 6	050° magnetic from Cape Lisburne radome at ≅55 miles	14.42	0.76	jetting	3/30/81	No penetration, rock
Ogotoruk Creek	Hole to Crowbill Pt., N303°E; Hole to highest peak of Crowbill Pt., N313°30′E. About 75 m from shore	6.40	0.81	Not attempted, bottom found to be hard by hand probing	3/31/81	No penetration, rock

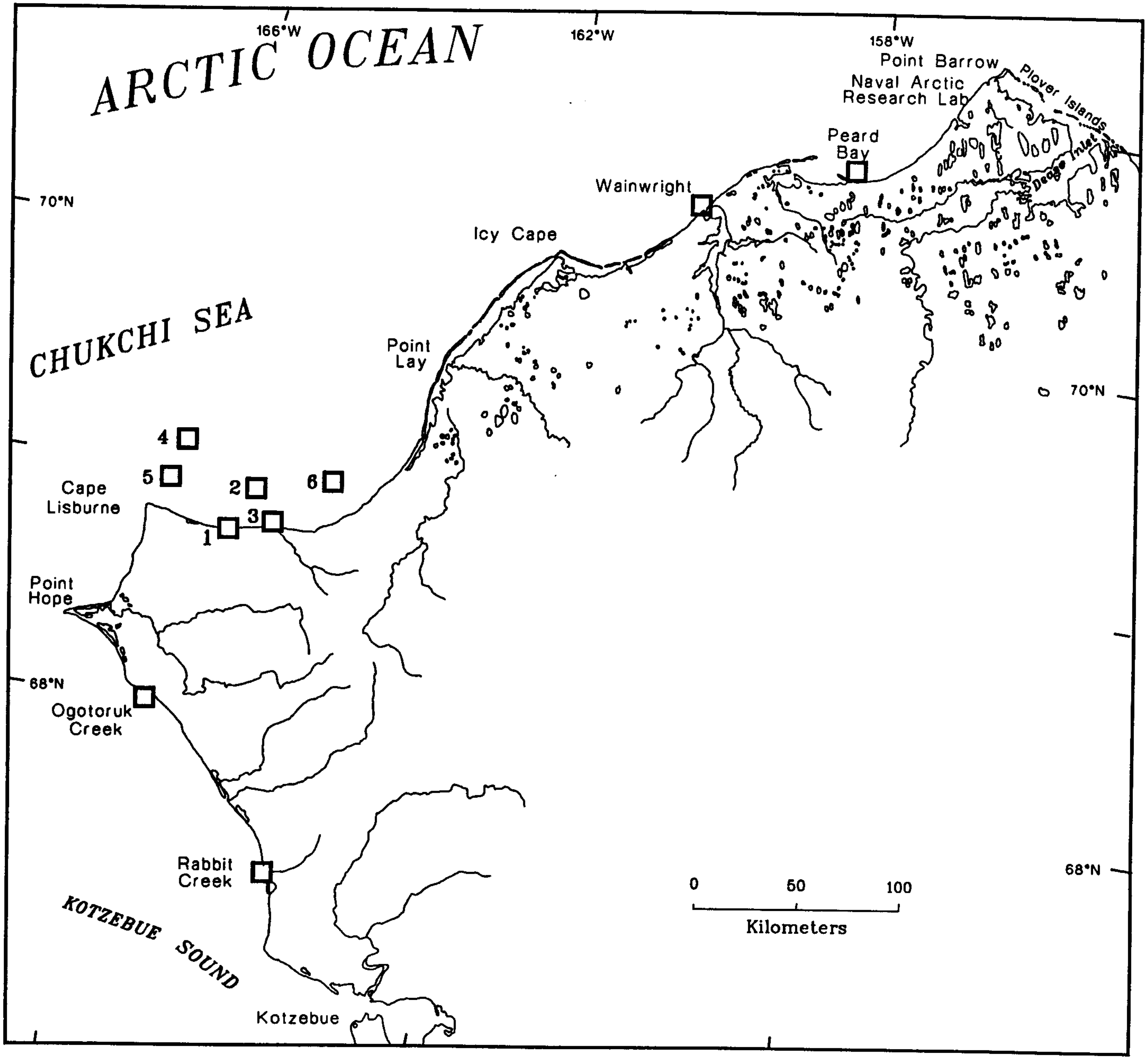

Figure 5.—Chukchi Sea hole location map. The six sites at Cape Lisburne and the site at Ogotoruk Creek encountered rock within 1 m of the seabed. Rock was also encountered in the Wainwright and Peard Bay holes.

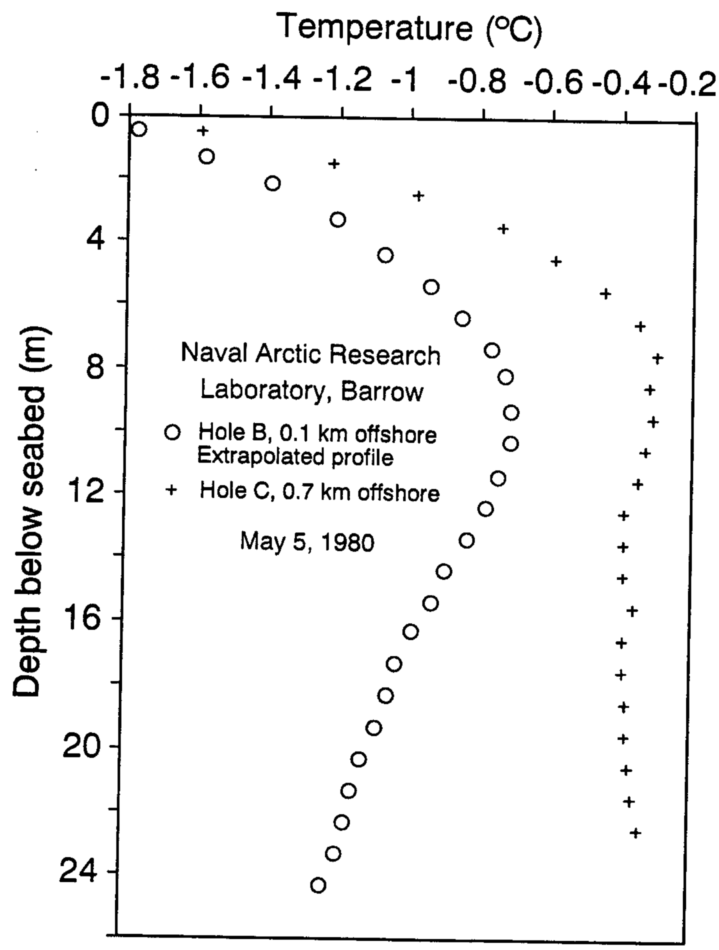

Figure 6.—Temperature profiles at the Naval Arctic Research Laboratory showing colder nearshore sediments with a negative temperature gradient and warmer offshore sediments with a positive temperature gradient.

temperature with depth (Figure 7), the effect of shoreline recession, and indirect evidence for ice at depth. All holes so far are confined to within 1 km of shore.

It seems entirely possible that permafrost might be widespread beneath the Chukchi Sea, because the entire area has been emergent and exposed to cold temperatures several times, and most recently only about 15,000 years ago. We have tried to address the question of the large scale permafrost distribution, using sea level history, seismic data, onshore boreholes, geologic and oceanographic data, and simple theory. This indirect approach suggests that, in the southeastern Chukchi Sea, permafrost may now be absent beneath deeper waters, although it may well survive in areas shallower than 15–30 m. The borehole data indicate that it probably exists in many nearshore areas. Elsewhere in the Chukchi Sea, except at Barrow, the situation is even less clear.

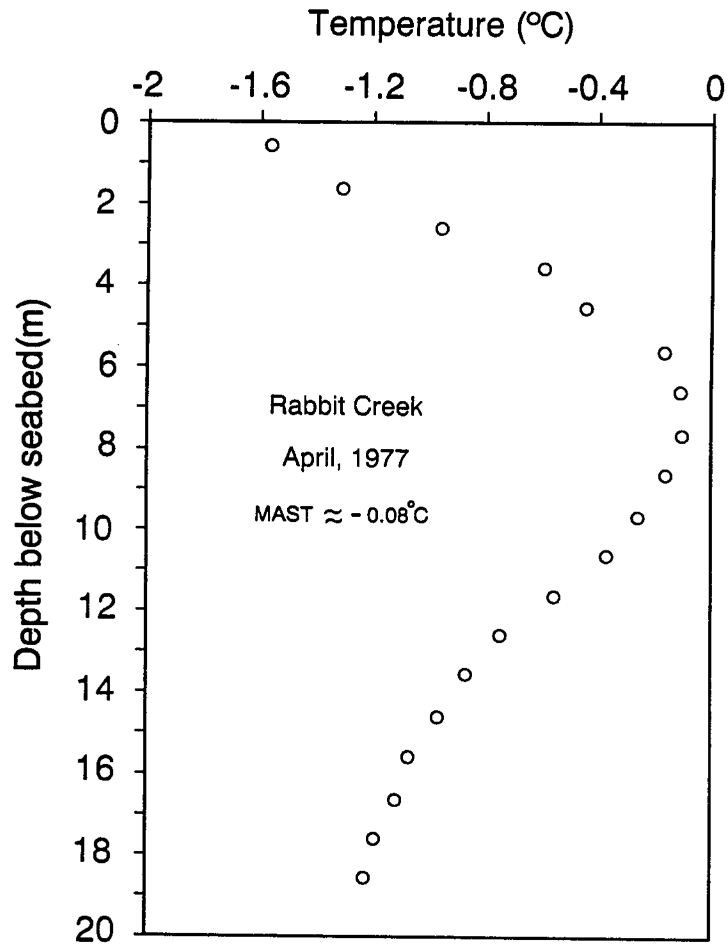

Figure 7.—Temperature profile at Rabbit Creek showing the negative temperature gradient found there.

The presence of rock at or near the seabed poses problems for OCS development. First, it cannot be assumed that the rock does not contain segregated ice. Both temperature profiles and borehole heating data will be required to determine the presence or absence of ice in the rock. The rock cannot be assumed to be a good foundation material for structures unless it can be shown that it does not contain segregated ice. Second, laying pipelines in this rock could be extremely difficult. Third, if the rock is as widespread as our sparse data suggest, then it may be very difficult to obtain gravel for construction of docks, causeways, and artificial islands. A brief discussion of sources of gravel in the Chukchi Sea area is given in Appendix D.

Elson Lagoon

A number of holes were drilled during the spring 1977 and 1978 field seasons along a line from Tekegakrok Point offshore at the location shown in Figure 8 and noted in Table 4. These consisted of eight jetted holes, one of which was started by driving, and three driven

Table 4.—Tekegakrok Point drilling data.

Hole designation	Location or distance offshore	Water depth (m)	Sea ice thickness (m)	Drilling method	Date of drilling	Approximate depth below seabed (m)
Hole 660 (1)	660 m	2.00	1.40	jetting	4/25/78	15.0
Hole 660 (2)	660 m	2.00	1.40	driving	4/21–29/78	18.0
Hole 880	880 m	2.32	1.45	jetting	4/26/78	28.0
Hole 1189 (1)	1,189 m	2.52	1.50	jetting	4/27–28/78	34.5
Hole 1189 (2)	1,189 m	2.52	≈1.40	driving	4/29–5/3/78	≈9.0
Hole 575	575 m	1.25	1.25	jetting	5/15/77	19
Hole 611	611 m	1.7	1.68	jetting	5/13/77	21
Hole 798	798 m	2.22	1.65	driving and jetting	5/13–16/77	14.6
Hole 1036	1,036 m	2.5	≈1.7	jetting	5/11/77	17
Hole 1466	1,466 m	2.80	1.8	jetting	5/12/77	20

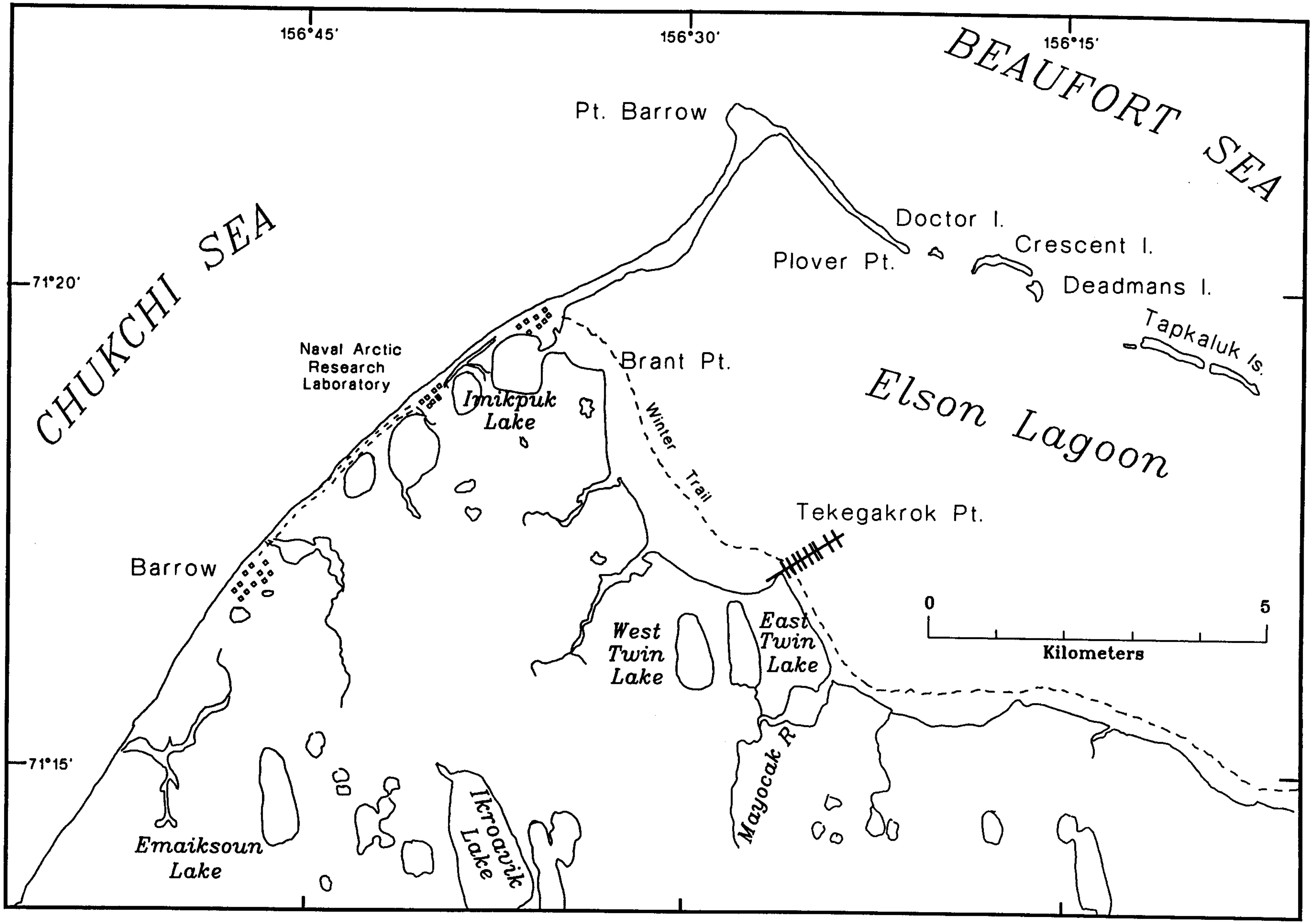

Figure 8.—Tekegakrok Point hole location map.

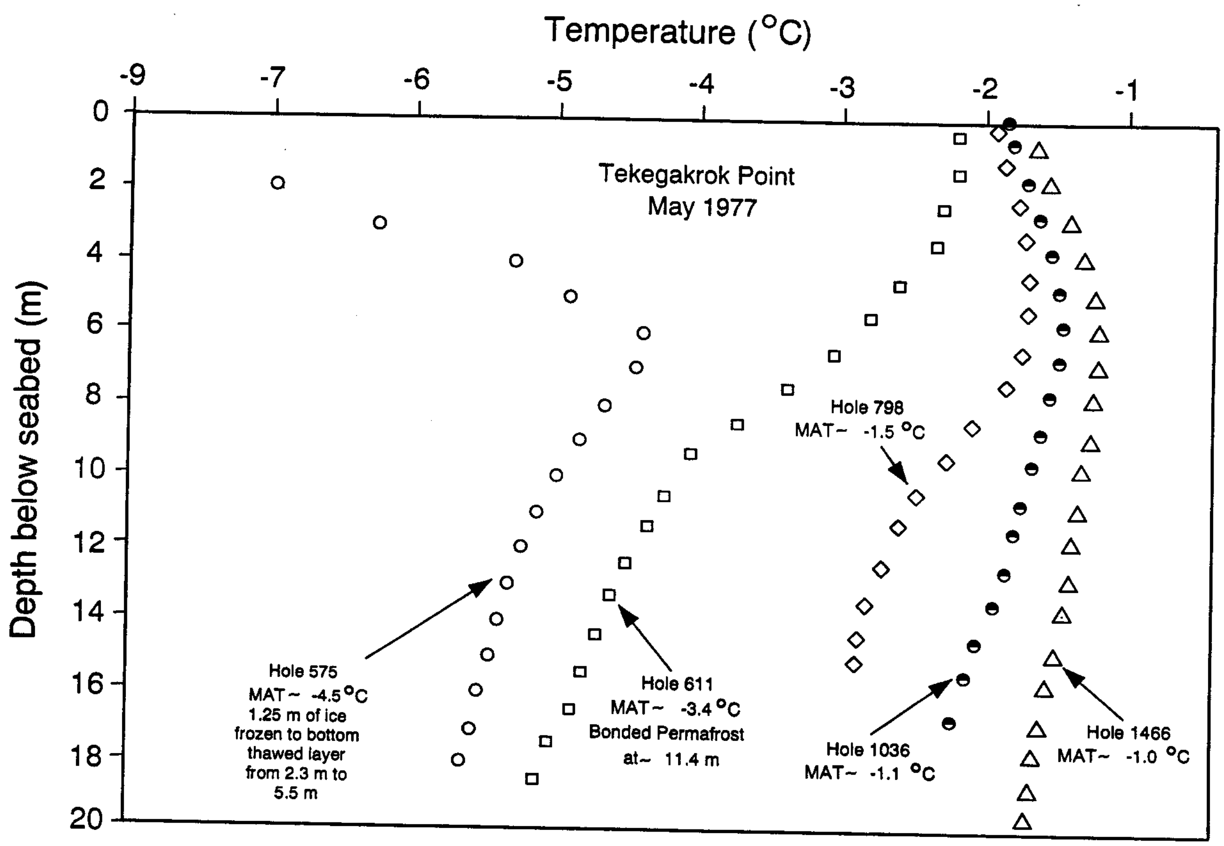

Figure 9.—Five temperature profiles at Tekegakrok Point showing the thermal evolution of the subsea permafrost with distance offshore.

holes. The shoreline retreat rate at these sites was estimated to be about 2.4 m/yr. These holes span the transition from cold nearshore conditions where sea ice freezes to the seabed, to the warmer conditions in deeper water maintained by the presence of seawater of normal salinity under the ice (Figure 9). The sediments were fine-grained.

Temperature, soil pore water conductivity, and hydraulic conductivity were measured. A thawed layer of generally increasing thickness can be traced out from shore under the seabed, although the phase boundary may not always be sharp. Simultaneous temperature and pore water conductivity measurements indicate that ice may be present at depths that are not well bonded. Within roughly 600 m of shore, where the sea ice can seasonally freeze to the seabed, pore water salinities in the thawed layer tend to be at least twice that of normal seawater. This high salinity persists to sediment depths of 5–10 m to at least 798 m from shore, but for an unknown reason it is absent at a site 1,198 m from shore, where the salinities to 9 m are only about 40% higher than normal seawater. The measured hydraulic conductivities are extremely low, and are typical of fine-grained silts or clays. This may have an effect on salt transport processes, and therefore on subsea permafrost evolution. A striking exception to this was found 1,189 m from shore, where there appears to be an extremely permeable layer 9 m below the seabed. At the same site a sharp boundary, evidently a lithologic one between the Pleistocene and Cretaceous (Black 1964; Lewellen 1976), was found at the 30-m depth from the temperature and drilling data.

The observations in hole 575 suggest that docks, causeways, or islands constructed from local fill material would likely become ice-bonded to 3 or 4 m during the first winter, and therefore resistant to ice forces, although some thawing would take place again the following summer. However, the high brine concentrations under the ice-bonded layer, which are probably concentrated during the freezing process, will severely reduce the degree of bonding unless some provision is made for their removal.

Lonely

Five holes were rotary-jet drilled during spring 1980 along a line bearing N328°E offshore from the DEW site at the location noted in Figure 10 and Table 5. Distances from shore were paced and checked by helicopter dead reckoning, but they may contain significant errors. This area was of considerable interest because there were no previous offshore hole data between Elson Lagoon and Harrison Bay. In addition, extremely high shoreline retreat rates, up to 10–15 m/yr, have been measured immediately to the west and to the east (Lewellen 1977). The onshore surficial deposits are mapped as interglacial nearshore and lagoon sand, silty fine sand, and pebbly sand (Hopkins and Hartz 1978b). Drilling data suggest that the offshore sediments are fine-grained to the maximum depth reached (31.3 m). The ice-bearing permafrost onshore at Cape Simpson, about 55 km to the west, is about 300 m thick and this depth contour passes near Lonely (Osterkamp and Payne 1981).

The temperature data (Figure 11) show evidence of rapid shoreline retreat. A preliminary analysis of the data suggests a retreat rate of several meters per year characteristic of the past 100 years or so. It was difficult to assess the presence of ice from drilling data alone at this site, as is often the case when fine-grained soils are present. Therefore, all the boreholes were heated after they had approached equilibrium, and the temperature response was used to determine the presence of ice. Ice-bearing permafrost seems to exist at 6–8 m below the seabed all the way out to the most seaward hole about 7.8 km from shore. Ice bonding exists somewhat deeper, characteristically at about 15 m below the seabed.

Because of the fine-grained soils and presence of ice-bearing sediments, possibly containing segregated ice, near the seabed, subsea permafrost problems for offshore development are likely to be very serious in this area. Thaw subsidence problems for bottom-founded structures and pipelines could potentially occur.

Harrison Bay

Holes drilled in Harrison Bay and seaward and west of Harrison Bay include two holes on the east side of the bay during spring 1977, and three holes near Esook to the west, one hole near Atigaru Point, and two holes north of Thetis Island during spring 1981. The hole locations and drilling information are given in Figure 12 and Table 6.

Table 5.—Lonely drilling data.

Hole designation	Location or distance offshore	Water depth (m)	Sea ice thickness (m)	Drilling method	Date of drilling	Approximate depth below seabed (m)
Hole 88 (Hole A)	Line intersects shore at a point N294°E from DEW line radar dome. From this point hole line bears N328°E. About 88 m offshore	1.98	1.45	rotary-jet	5/8/80	23
Hole 950 (Hole B)	Same, but about 950 m offshore	3.12	1.58	rotary-jet	5/9/80	31
Hole 2560 (Hole C)	Same, but about 2,560 m offshore	4.80	1.47	rotary-jet	5/10/80	21
Hole 4360 (Hole D)	Same, but about 4,360 m offshore	6.50	2.1 (rafted?)	rotary-jet	5/11/80	17
Hole 7770 (Hole E)	Same, but about 7,770 m offshore	7.70	1.39 (?)	rotary-jet	5/13/80	27

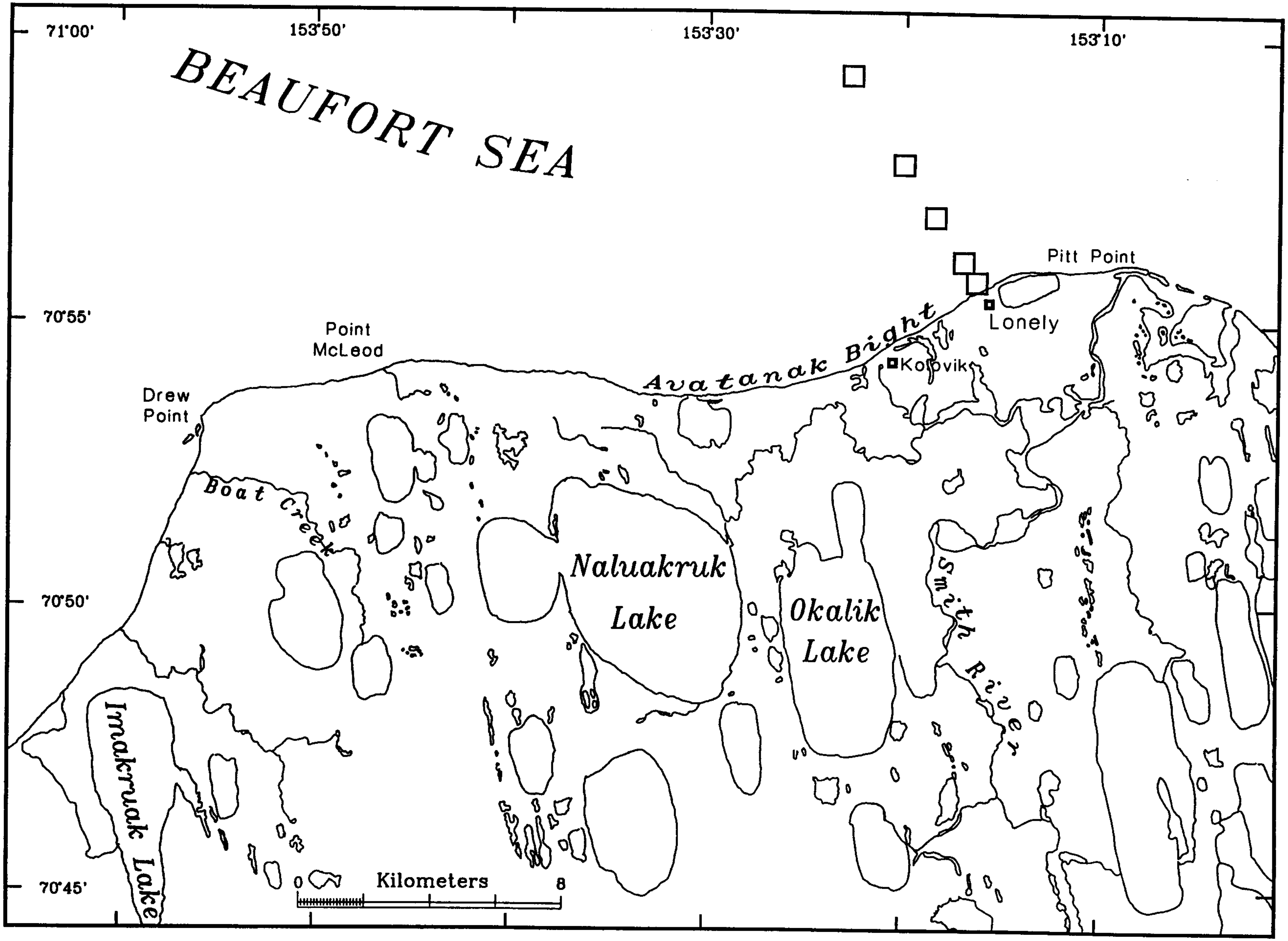

Figure 10.—Lonely hole location map.

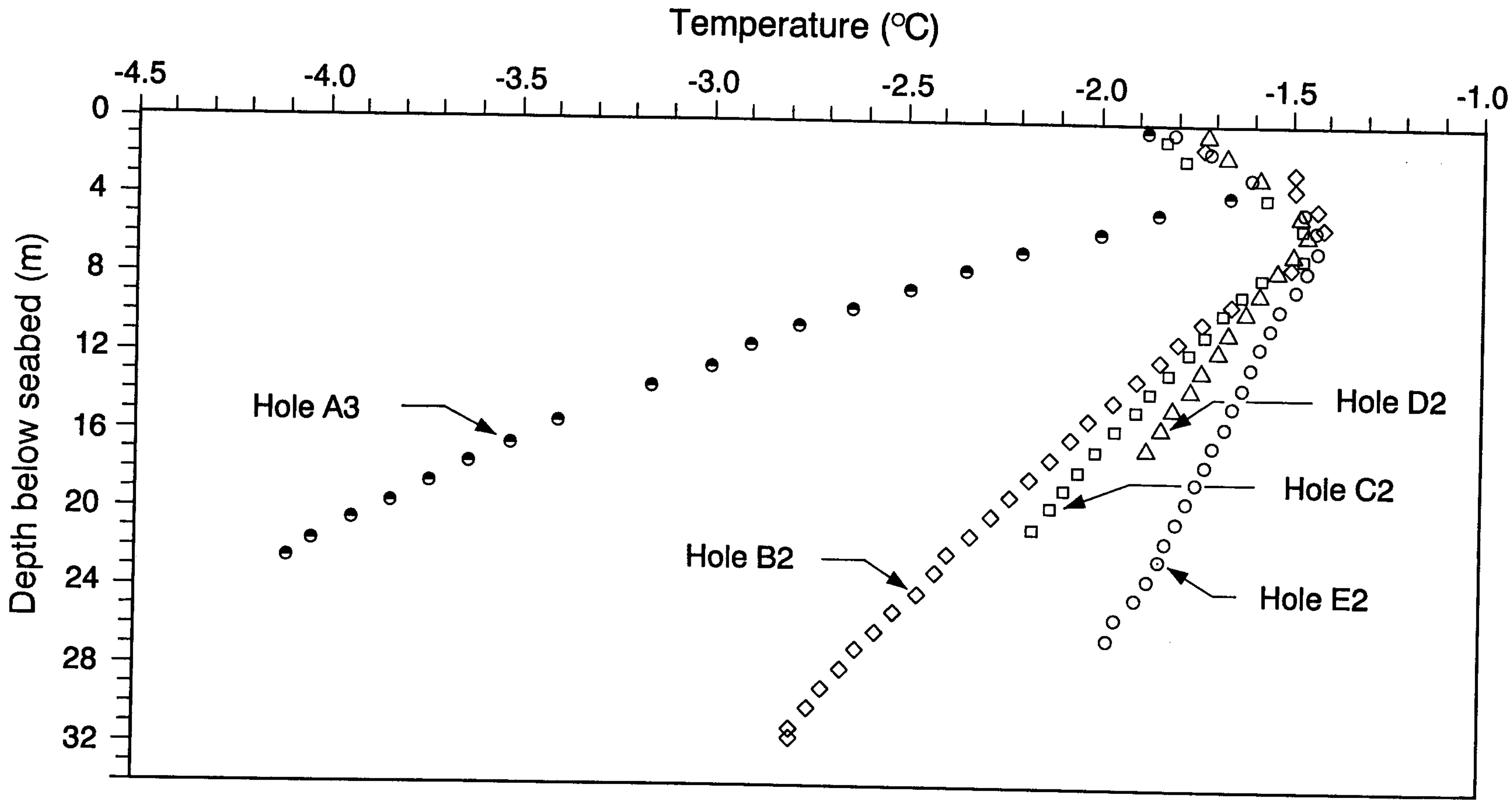

Figure 11.—Five temperature profiles at Lonely showing the thermal evolution of the subsea permafrost with distance offshore.

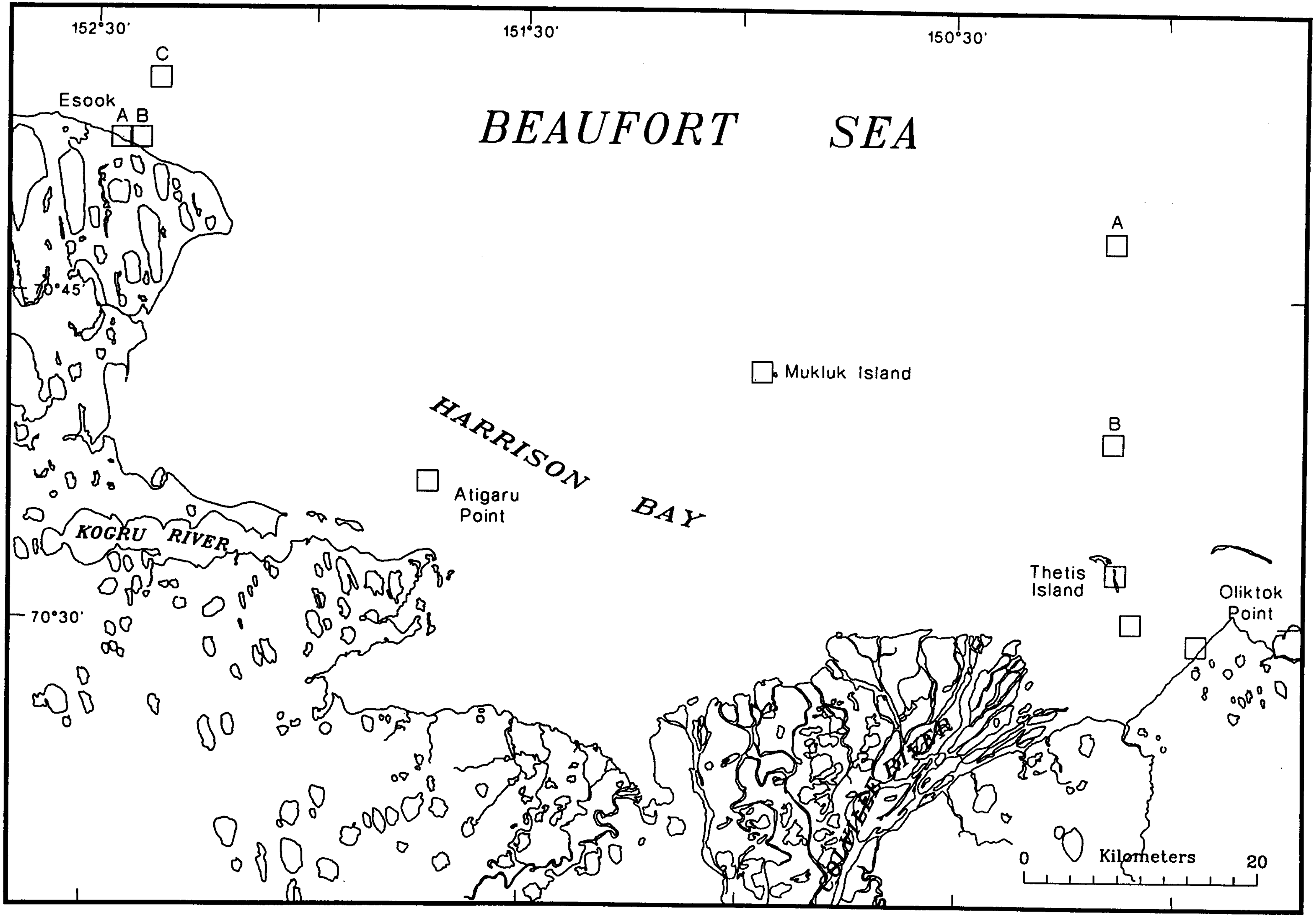

Figure 12.—Harrison Bay hole location map.

Table 6.—Esook–Harrison Bay drilling data.

Hole designation	Location or distance offshore	Water depth (m)	Sea ice thickness (m)	Drilling method	Date of drilling	Approximate depth below seabed (m)
Esook A	70°52.2′N, 152°25.3′W	2.72	1.98	rotary-jet	5/1/81	15
Esook B	70°52.5′N, 152°21.1′W	5.45	1.68	rotary-jet	5/7/81	19
Thetis A	330° magnetic from Thetis Island hole at 23 nautical miles	25.92	2.82	jetting	4/30/81	4
Thetis B	70°39.2′N, 150°08.9′W N148.5°E hole to Oliktok radome	14.78	1.81	jetting	4/30/81	16
Atigaru Pt.	70°36.4′N, 151°30.5′W	6.83	1.86	rotary-jet	5/1/81	17
Harrison Bay–south of Thetis Island	5.7 km	2.95	2.0	driving	4/30–5/1/77	15
Harrison Bay–Oliktok	≈400 m	2.4	2.1	jetting	5/2/77	8

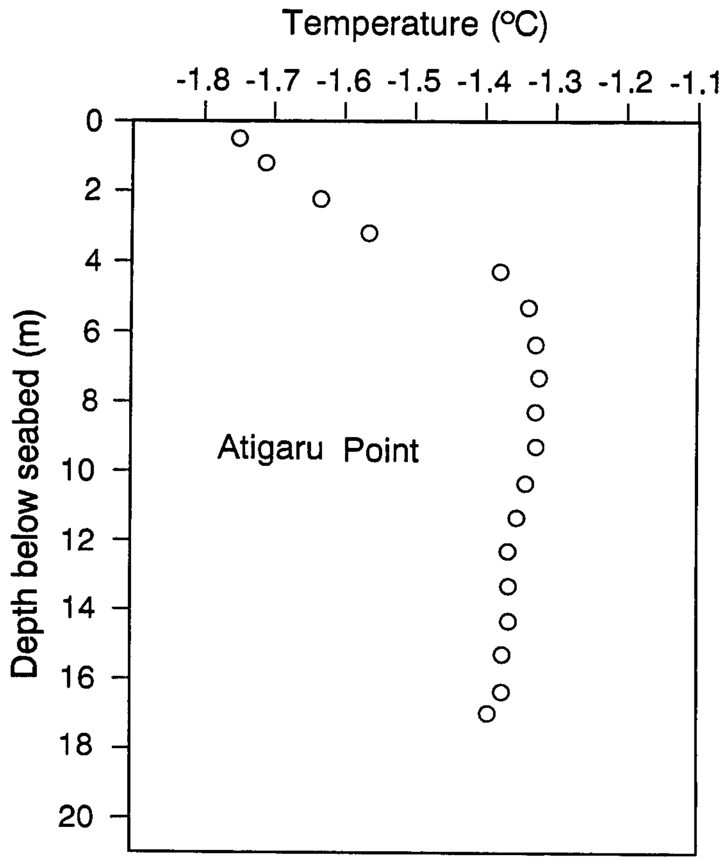

Figure 13.—Temperature profile in the Atigaru Point hole.

The holes drilled on the east side of Harrison Bay appear to straddle the transition zone between the fine-grained sediments to the west and the coarser sediments characteristic of Prudhoe Bay. It appears that this transition occurs somewhere between Oliktok Point and Thetis Island. A typical temperature profile for this area is shown in Figure 13.

All holes drilled in the fine-grained sediments encountered ice-bonded subsea permafrost within 15 m of the seabed or less even for a water depth of 15 m. These results are similar to those at Lonely and suggest that subsea permafrost problems for offshore development are likely to be serious in the Harrison Bay area. Gas was encountered in the 1977 hole south of Thetis Island and 8.5 km west of Oliktok DEW station.

A hole was drilled in 14.8 m of water north of Thetis Island and to the east of the Mukluk Island exploration hole. Ice-bearing fine-grained sediments were thought to exist below the 12-m depth in this hole which was about 20 km offshore from Oliktok Point.

Long Island

Four holes were drilled on a line offshore from Long Island during spring 1979 at the locations noted in Figure 14 and Table 7. This line is several kilometers to the west of the

Table 7.—Long Island drilling data.

Hole designation	Location or distance offshore	Water depth (m)	Sea ice thickness (m)	Drilling method	Date of drilling	Approximate depth below seabed (m)
Hole A	70°28.9′N, 148°50.5′W 0.315 km N25° of VABM on Long Island	3.6	2.30	rotary-jet	5/17/79	13.3
Hole B	70°29.0′N, 148°51.0′W 0.600 km N25°E of VABM on Long Island	5.8	2.30	rotary-jet	5/18/79	12.8
Hole C	70°29.5′N, 148°50.4′W ≅1.8 km N25°E of VABM on Long Island	9.7	2.00	rotary-jet	5/19/79	11
Hole D	70°30.7′N, 148°45.5′W ≅3.2 km N25°E of VABM on Long Island	14.0	1.40	rotary-jet	5/20/79	7

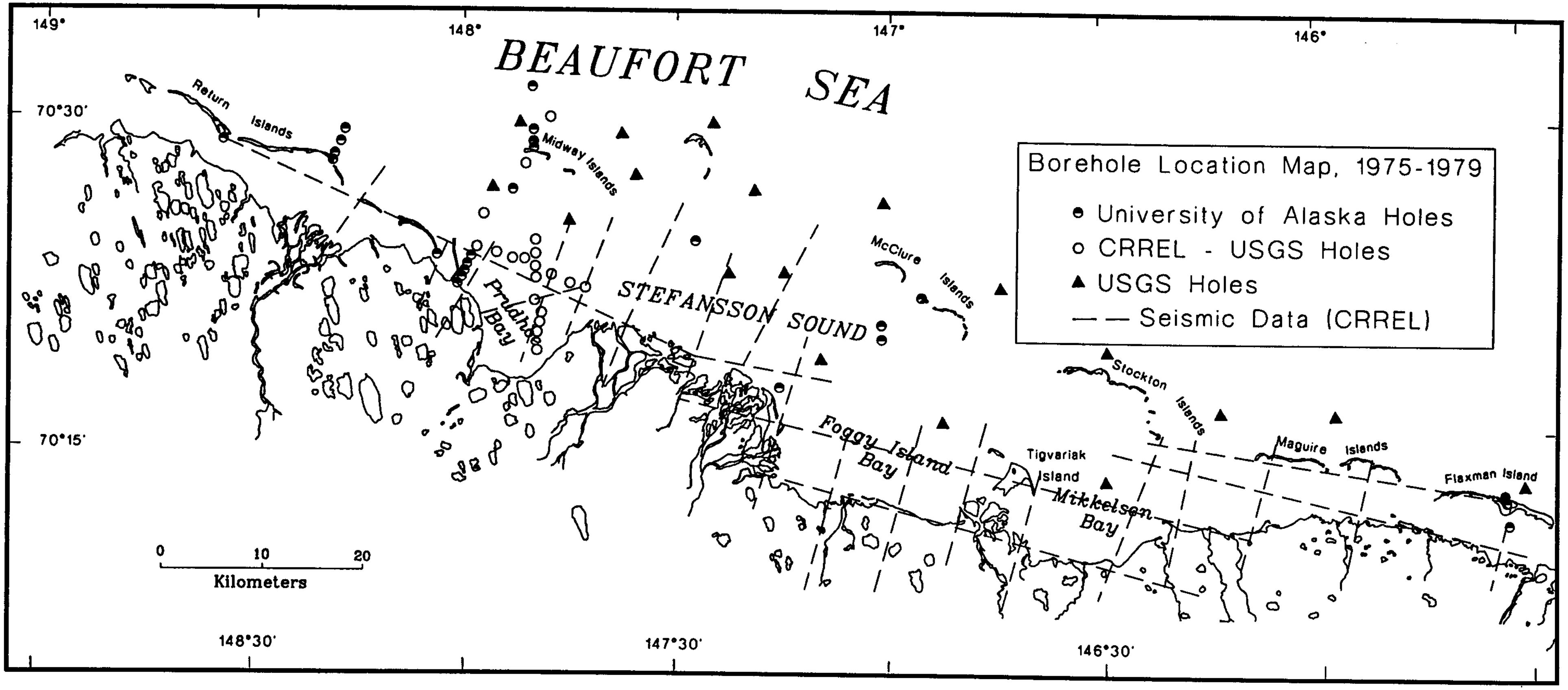

Figure 14.—Prudhoe Bay area hole location map showing USGS, CRREL–USGS, and University of Alaska holes. The dashed lines represent seismic lines investigated by CRREL.

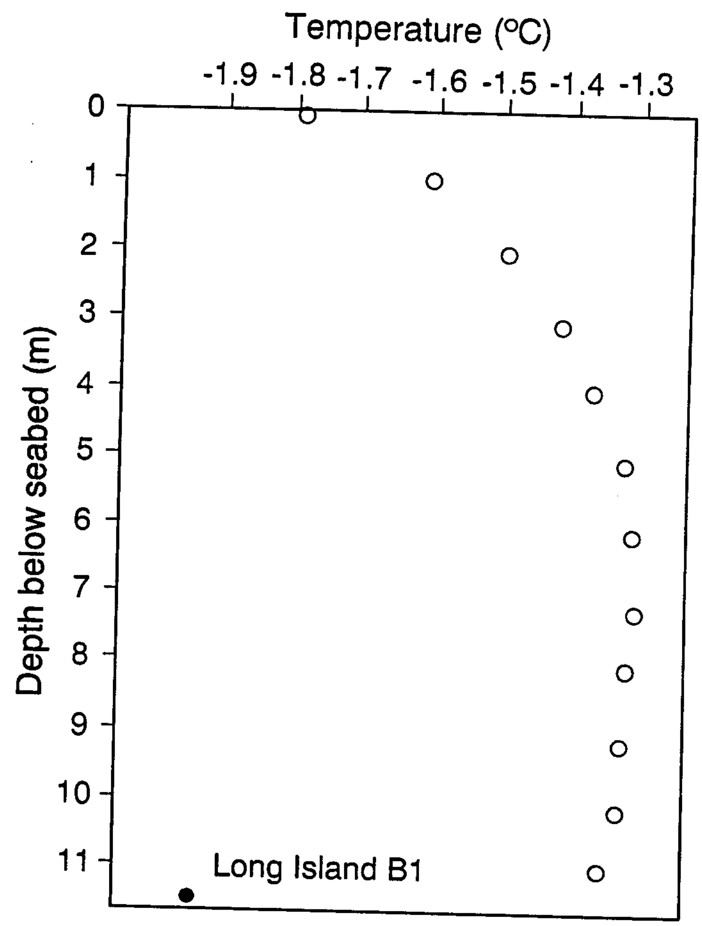

Figure 15.—Temperature profile in Long Island hole B.

Shell artificial Seal Island. The lithology determined by rotary jet drilling consists of a layer of fine-grained sediments, probably clay, which thin seaward from about 7 m thick at water depths of 3–5 m to about 2 m at a water depth of 14 m. These sediments were underlain by gravel-bearing material which could be penetrated about 6 or 7 m. The clay at the seabed may be somewhat difficult to trench. Figure 15 shows a temperature profile from a hole where the water depth was about 6 m.

West Dock

Prudhoe Bay, particularly its west side, has been the area most intensely studied by the OCSEAP projects (Figures 14 and 16 and Table 8). The West Dock line extends from North Prudhoe Bay State No. 1 Well (70°22'36"N, 148°31'28"W) onshore along a line bearing about N31.5°E to Reindeer Island, and then approximately due north. Shoreline retreat is roughly 1 m/yr.

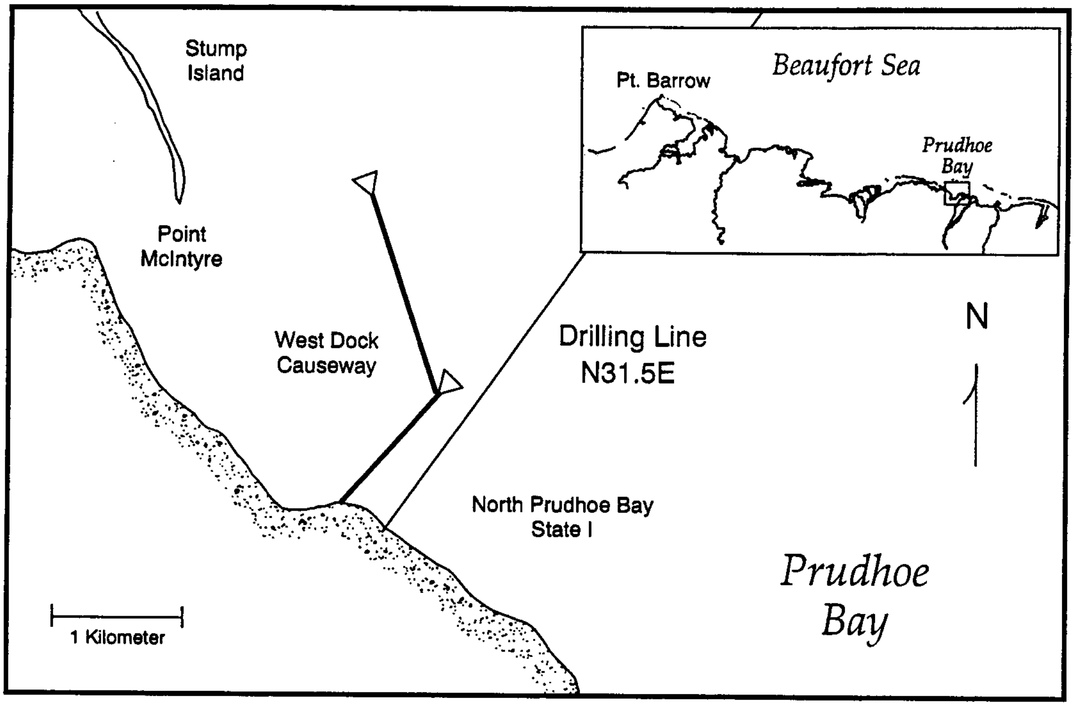

Figure 16.—West Dock line. The line continues to Reindeer Island, about 14 km from shore, and then turns due north. West Dock is also called the Arco Causeway.

The shallow permafrost regime along the West Dock line can be subdivided into five zones:

1. Nearshore zone, 0–400 m from shore.—In this zone, the permafrost regime is analogous to what is found onshore. There is a seasonably active layer which is ice-unbonded to a depth of 3–4 m by fall. Beneath it, judging from conditions onshore at North Prudhoe Bay State No. 1 Well, the permafrost is both ice-bearing and ice-bonded to a depth of about 550 m (Osterkamp and Payne 1981). The lack of deeper thaw in this region is probably due to the low seabed temperatures, which are a result of freezing of the sea ice to the seabed. By late winter, the active layer is mostly ice-bonded, although it seems to contain considerable liquid brine.

2. Ramp zone, 400 to about 440 m from shore.—In this zone, the depth to the ice-bonded permafrost table increases rapidly, and rather linearly, from a few meters below the seabed to about 14 m below it. The ice-bonded and ice-bearing boundaries coincide to within a small fraction of a meter; we usually refer to the ice-bearing boundary as the "phase boundary." To the extent that shoreline retreat is constant (about 1 m/yr), distance offshore is proportional to submergence time. The slope of the phase boundary in this zone therefore implies a thaw rate

Table 8.—West Dock drilling data.

Hole designation	Location or distance offshore	Water depth (m)	Sea ice thickness (m)	Drilling method	Date of drilling	Approximate depth below seabed (m)
Prudhoe Bay, West Dock	All holes on a line bearing about N31.5°E from NPBS #1 well to Reindeer Island					
Hole 398	398 m*	—	1.27	driving	5/27/81	2.74 (?)
Hole 418	418 m	—	1.51	driving	5/27/81	8.81
Hole 419	419 m	—	1.63	driving	5/31/81	8.95
Hole 433	433 m	—	1.58	driving	5/28/81	10.81 (?)
Hole 438	438 m	—	1.67	driving	5/27/81	13.98
Hole 439	439 m	—	1.54–1.67	driving	5/30/81	14.09
Hole 448	448 m	—	1.61	driving	5/28/81	15.30
Hole 300	300 m	1.37	1.37	auger	5/24/80	12.9
Hole 400	400 m	1.58	1.58	driving	5/23/80	3
Hole 438	438 m	1.53	1.53	driving	5/26/80	13.0
Hole 438 S	Same, but slightly displaced	1.53	1.53	driving	5/28/80	13.0
Hole 700	700 m	1.62	1.62	driving	5/23/80	24.8
Hole 701	701 m	1.62	1.62	driving	5/29/80	24.9
Dock Hole 700 (A)	700 m	1.8	1.82	driving	5/14/79	20.87 (?)
Dock Hole 689 (B)	689 m	1.8	1.82	driving	5/26/79	24.6
Dock Hole 701 (C)	701 m	1.8	1.75	driving	5/15/79	24.5

Table 8.—(continued).

Hole designation	Location or distance offshore	Water depth (m)	Sea ice thickness (m)	Drilling method	Date of drilling	Approximate depth below seabed (m)
Dock Hole 687 (D)	687 m	1.8	1.80	driving	5°23–24/79	24.6
Dock Hole 837 (E)	837 m	1.8	1.76	driving	5/26/79	shallow
Dock Hole 700 (F)	700 m	1.8	1.85	driving	5/29/79	10.4
Dock Hole 700 (G)	700 m	1.8	1.85	driving	5/29/79	7.3
Dock Hole 700 (H)	700 m	1.8	1.85	driving	5/29/79	5.8
Dock Hole 700 (I)	700 m	1.8	1.85	driving	5/29/79	8.8
Hole 9500	9,500 m	6.9	1.60	driving	5/23–25/78	25.5
Hole 1252	1,252 m	1.97	1.87	driving	5/4–5/77	15
Hole 2114	2,114 m	1.85	1.9	driving	5/8/77	26
Hole –226	226 m onshore	—	—	auger	5/6–7/75	12.2
Hole –225	225 m onshore	—	—	auger	4/30/75	2.9
Hole –69	69 m onshore	—	—	auger	5/8/75	12.2
Hole 0	0 m	—	—	auger	4/30/75	3.5
Hole 190	190 m	(ice frozen to bottom)	1.1	rotary wash boring	5/3–5/75	54.7
Hole 195	195 m	(ice frozen to bottom)	1.1	auger	5/1–2/75	8.8
Hole 196	196 m	(ice frozen to bottom)	1.1	penetration test	5/15/75	2.3
Hole 203	203 m	(ice frozen to bottom)	1.1	auger	5/1–2/75	2.3

Table 8.—(continued).

Hole designation	Loocation or distance offshore	Water depth (m)	Sea ice thickness (m)	Drilling method	Date of drilling	Approximate depth below seabed (m)
Hole 334	334 m	(ice frozen to bottom)	1.6	penetration test	5/15/75	1.4
Hole 403	403 m	(ice frozen to bottom)	1.6	penetration test	5/15/75	3.1
Hole 481	481 m	<0.1 (water under ice)	1.8	rotary wash boring	5/16–18/75	26.6
Hole 486	486 m	<0.1 (water under ice)	1.8	penetration test	5/16/75	19.5
Hole 493	493 m	<0.1 (water under ice)	1.8	penetration test	5/15/75	14.9
Hole 964	964 m	0.2 (water under ice)	1.8	penetration test	5/15/75	14.8
Hole 3370	3,370 m	0.8 (water under ice)	2.0	rotary wash boring	5/7–14/75	42.9

* The West Dock hole location designation used for the 1981 holes is not entirely consistent with that used the previous year, when "distance from shore" really was from a fixed marker which was about 2 m from the tundra edge in 1981. If 2 m is added to the 1981 distance designations, they are consistent, in a fixed reference system (not one moving with the shoreline), with the 1980 designations.

of about 0.3 m/yr. Although it is virtually certain that this thawing occurs in response to increasing mean seabed temperature seaward of the nearshore zone, it is difficult to understand in terms of any reasonably simple seabed temperature model.

3. Parabolic zone, from about 440 m to 3.4 km or greater from shore.—The phase boundary undergoes a break in slope between this zone and the ramp zone. In the parabolic zone, the phase boundary shape can be considered parabolic and described by

$$Y = a\sqrt{x - x_0}$$

where Y is the depth (m) of the boundary beneath the seabed, x = distance (m) from shore, $a = 1.147\ m^{1/2}$ and $x_0 = 276$ m. This shape seems understandable in terms of a $(time)^{1/2}$ dependence of thaw layer thickness, which is typical of many simple phase boundary propagation problems. The phase and mechanical ice-bonded boundaries correspond as they do in the ramp zone; this boundary is extremely sharp. The seabed temperature is warmer than its average value in the ramp zone, but the phase boundary temperature is the same, close to –2.40°C, over a distance of at least several kilometers. The shape of the phase boundary in these three zones is indicated in Figure 17.

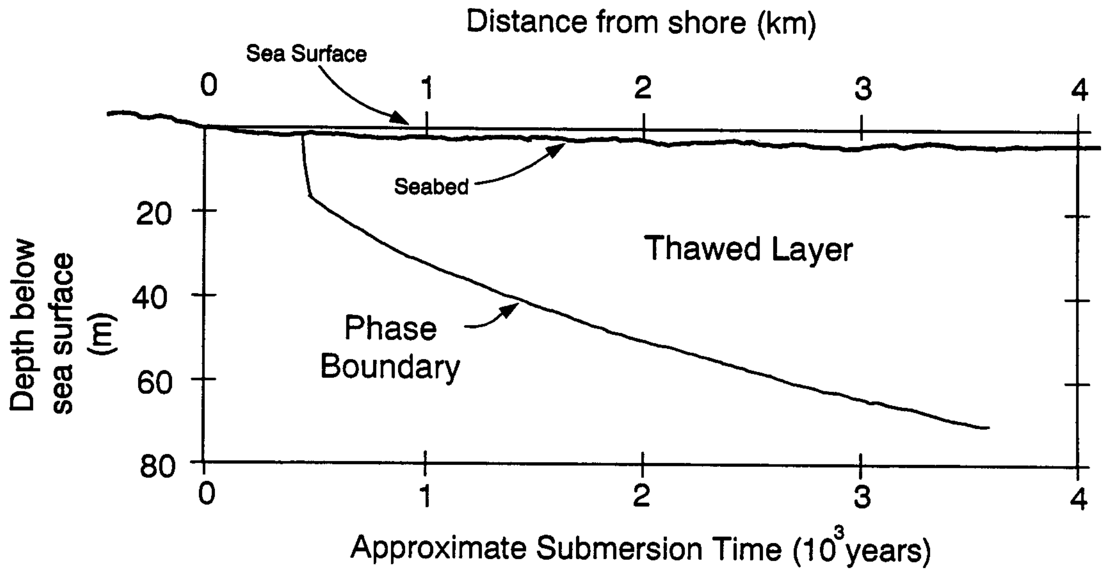

Figure 17.—Profile of the ice-bonded permafrost table along the West Dock line where it coincides with the phase boundary. The nearshore, ramp, and parabolic zones are located from 0 to 400 m, 400 to about 440 m, and about 440 to > 3,400 m from shore, respectively.

4. Intermediate zone, beyond 3.4 km offshore to Reindeer Island.—Fewer data are available from this zone, but seismic evidence suggests that somewhere beyond 3.4 km from shore the ice-bonded boundary becomes deeper than indicated by the above equation. A suggestion that this deep thaw may be due to the presence of a paleo-river channel was offered by Smith and Hopkins (1982).

5. Seaward zone, outside Reindeer Island.—Ice-bonded permafrost occurs typically 5–15 m beneath the seabed in this zone. Limited drilling data suggest that the phase and ice-bonded boundaries may coincide approximately, and that the phase boundary temperature is roughly –1.8°C, significantly different from the –2.40°C of the ramp and parabolic zones. This zone is also referred to as the Reindeer Island line and is described in more detail in the next subsection.

There are several general observations on the West Dock line that are of interest. The first of these is that seasonal freezing often occurs in the shallow subsea sediments, even beyond the nearshore zone. Beyond the ramp zone, however, there may not be significant ice bonding. Seasonal ice may be found to depths of several meters.

Another observation concerns the nature of permafrost below the phase boundary. A comparison of pore water bulk salinity and in situ temperature of the two existing samples from the nearshore and parabolic zones indicates that 30–50% of the water is in the liquid phase in situ. Evidence for deep briny layers both onshore and offshore also exists (Osterkamp and Payne 1981). Shallow permafrost samples onshore also contain salt, but insufficient to allow complete melting of ice at the phase boundary temperature of –2.40°C. This seems to indicate that a key part of the thaw process is transport of salt from the seabed through the thawed layer to the phase boundary. Other evidence for this important conclusion is discussed by Harrison and Osterkamp (1982b).

Some lithologic information, compiled from both USGS–CRREL and University of Alaska data, is shown in Figure 18. This lithological profile is similar to that reported by Hopkins and Hartz (1978), with minor differences at hole 3370 and hole –226 (onshore).

Temperature data exist from most of the holes on the West Dock line. The most obvious feature in the ramp and parabolic zones, discussed previously, is the uniformity of the phase boundary temperature at close to –2.40°C. Below approximately the 10-m depth of seasonal variations, all temperature profiles show negative gradients and are approximately linear in the thawed layer, although some curvature, suggesting pore water motion or yearly variation, is apparent. Figure 19 shows a typical example. Extrapolation of the linear portion of these profiles to the seabed yields estimates for mean annual seabed temperatures, although there are interpretive problems when the thawed layer is thin, or seasonal freezing at the seabed is significant. The results are summarized in Figure 20. Some data on the amplitude of seasonal temperature variations at the seabed were also obtained.

West Dock, which is really a gravel causeway (Figure 16), and whose leg parallel to the West Dock line is about 1.2 km long, probably has had an important effect on circulation near the West Dock line since its construction in 1974. It is probably influencing the seabed

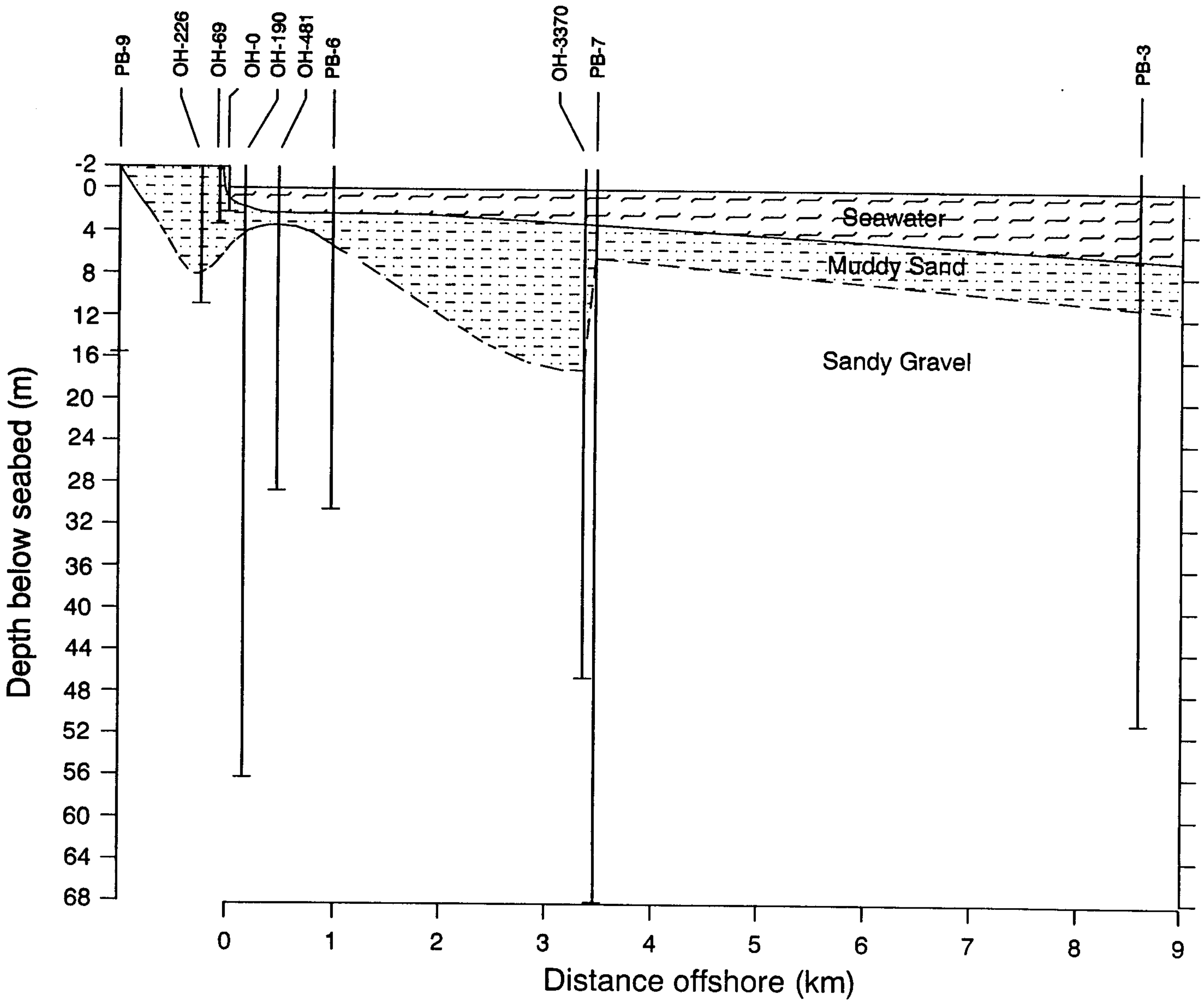

Figure 18.—Lithology along the West Dock–Reindeer Island line showing the contact between the finer-grained near-surface sediments and the underlying coarser sandy, silty gravel.

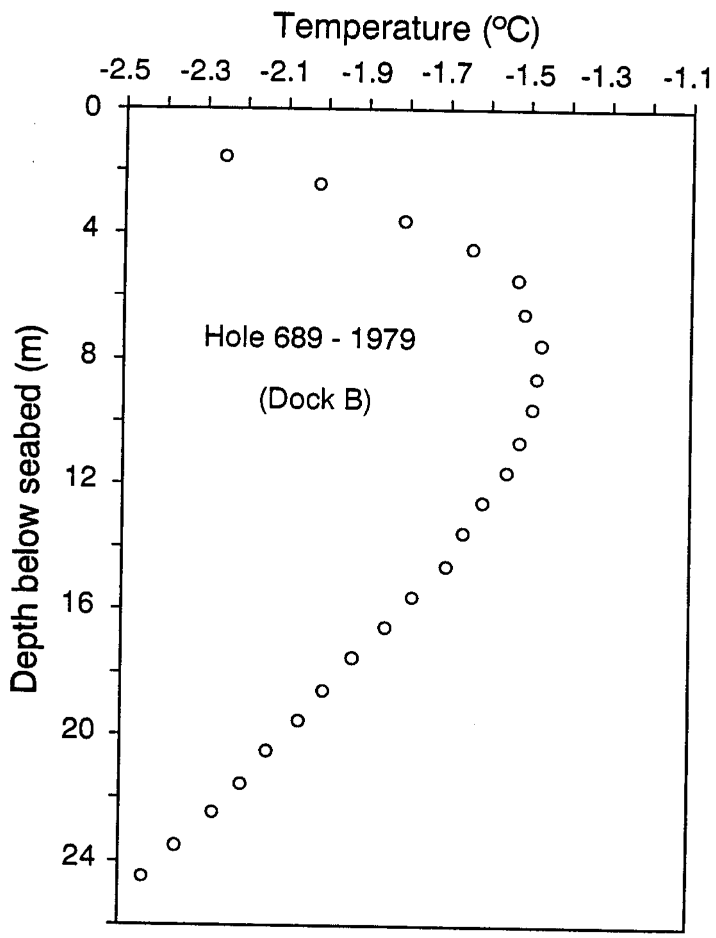

Figure 19.—Typical spring temperature profile through the thawed layer along the West Dock line.

temperature and salinity also, and consequently the evolution of the subsea permafrost regime.

Several special measurements, in addition to soil type and temperature, were made in the thawed layer on the West Dock line. In situ sampling of the pore water, and subsequent laboratory measurements of electrical conductivity, permitted estimates of pore water salinity, under the assumption that its chemical composition is rather similar to that of normal seawater. This assumption is supported by results of the CRREL–USGS program (Page and Iskander 1978). Sample data are in Figure 21. Salinity varies but little with depth and distance from shore. A characteristic value is 43 parts per thousand, which is about 25% higher than normal seawater. Some evidence exists for a thin, weak boundary layer just above the phase boundary, through which the salinity decreases slightly as the phase boundary is approached.

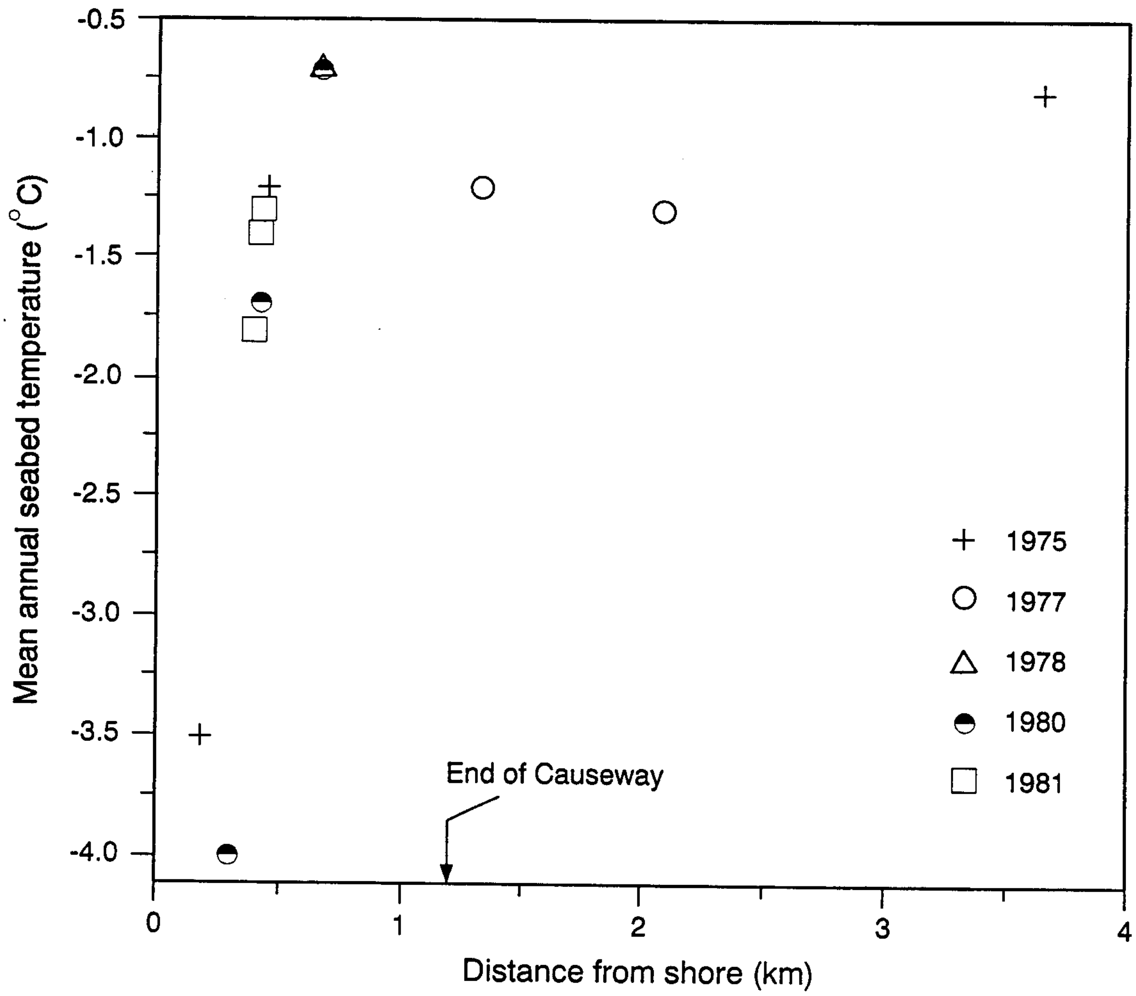

Figure 20.—Estimated mean annual seabed temperature along the West Dock line. The years in which the estimates were made are indicated.

Both laboratory measurements and the rate of inflow of pore water samples during collection were used to estimate hydraulic conductivity. There is a variation of about a factor of 3 (or 10 in a few cases) around the characteristic value of 3 m/yr. Darcy's law, upon which the results are based, is not obeyed well. Two laboratory measurements on samples recovered from hole 3370 gave an average value of 13 m/yr.

A pore water pressure profile through the thawed layer was obtained. The pressure was slightly less than hydrostatic, except at shallow depths probably influenced by seasonal freezing (Figure 22).

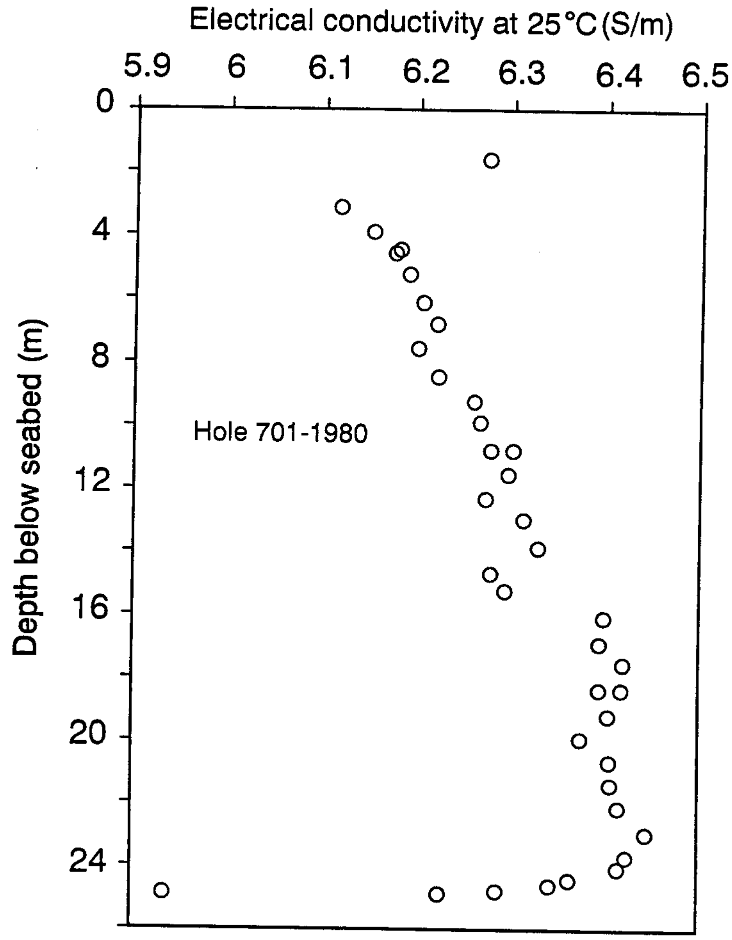

Figure 21.—Typical profile of pore water electrical conductivity, measured at 25°C, in hole 701 along the West Dock line.

Reindeer Island

Six holes were drilled on a line due north of Reindeer Island; the farthest offshore hole was 8 km from the island, where the water depth was 21 m. Figure 14 and Table 9 give the locations and drilling information. A temperature profile from the 5-m water depth is shown in Figure 23. The lithology determined during rotary jet drilling indicates that a layer of fine-grained sediments overlies somewhat coarser sediments. Depth to the ice-bearing/bonded permafrost first increases and then decreases with distance seaward while the sediments become finer-grained. In the three outermost holes the sediments at the seabed were clay, which appeared to be overconsolidated, and were covered by a thin layer of soft sediments; this suggests that the sedimentation rate is very low. It is difficult to understand this low sedimentation rate in the presence of shoreline erosion rates of 1 m/yr or more.

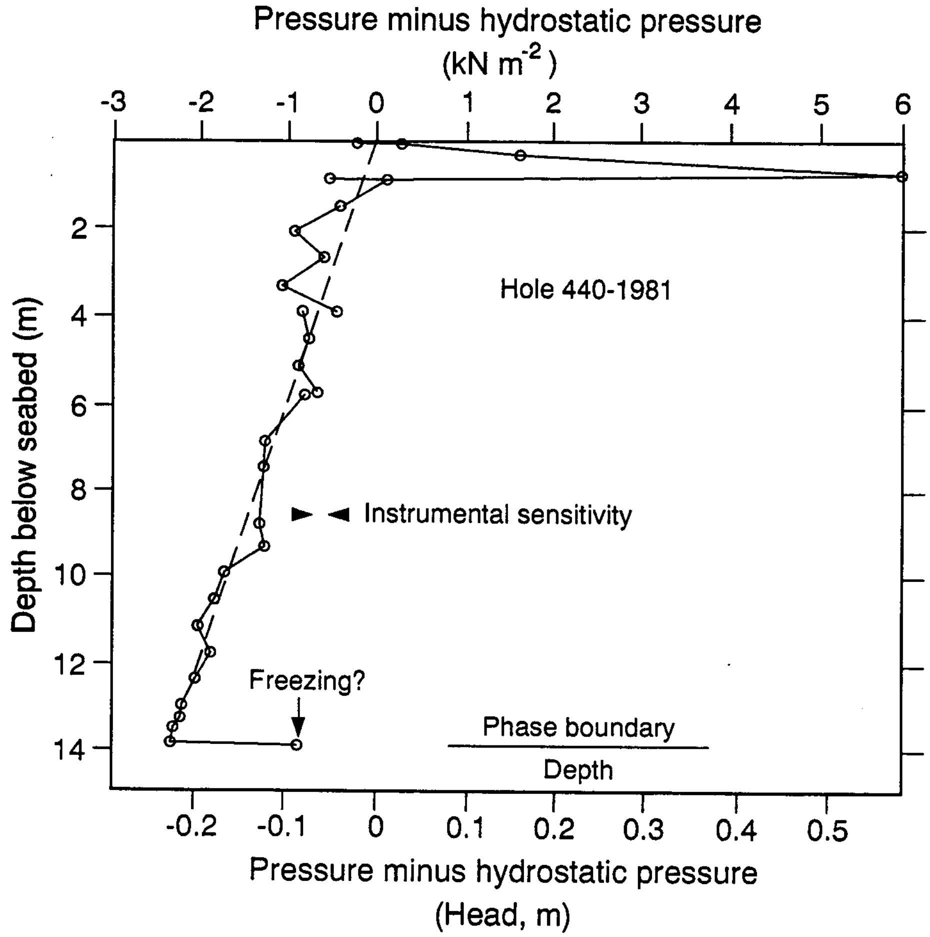

Figure 22.—Pore water pressure profile in hole 440 along the West Dock line.

The depth to ice-bearing permafrost under the clay sediments was less than 11 m. These clays will be very difficult to trench for laying pipelines in the seabed. Since the ice-bonded permafrost is shallow, the danger of permafrost containing segregated ice and subsequent problems of creep and thaw settlement must also be faced for bottom-founded structures and pipelines.

Sagavanirktok Delta

A hole was driven at a site off the Sagavanirktok Delta at the location given in Figure 14 and Table 10. The sediments were fine-grained, with clay near the bottom of the hole. A hard boundary which was probably ice-bonded permafrost was found at 25 m below the seabed.

Table 9.—Reindeer Island drilling data.

Hole designation	Location or distance offshore	Water depth (m)	Sea ice thickness (m)	Drilling method	Date of drilling	Approximate depth below seabed (m)
Hole A	0.352 km N9°E from the USGS Tower on Reindeer Island	3.8	1.93	rotary-jet	5/5/79	23.5
Hole B	70°29.0′N, 148°21.1′W 0.744 km N10°E from the USGS Tower on Reindeer Island	5.4	1.90	rotary-jet	5/5/79	23
Hole C	70°29.7′N, 148°20.8′W ≅2.6 km N9°E from the USGS Tower on Reindeer Island	11.5	1.91	rotary-jet	5/4/79	17
8 km	8 km true North of Reindeer Island	21	≈1.60	jetting	4/21/78	3.5
RDEE (1)	5.9 km true North of Reindeer Island	17	≈1.60	jetting	4/20/78	10.0
RDEE (2)	5.9 km true North of Reindeer Island (30 m separation)	17	≈1.60	rotary-jet	5/23/78	12

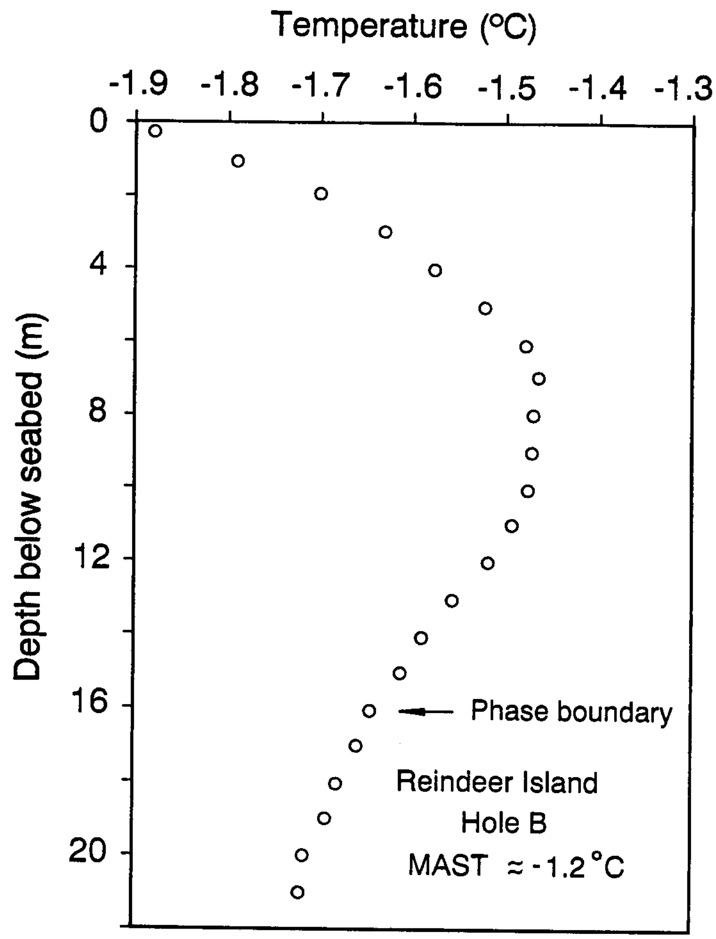

Figure 23.—Temperature profile on Reindeer Island hole B which was located at the 5-m water depth due north of Reindeer Island.

Jeanette Island

Three holes were rotary-jet drilled on a line from Point Brower through Jeanette Island as shown in Figure 14 and Table 10. The probable existence of ice-bonded permafrost in fine-grained soils at a depth of 5.5 m in the hole approximately 1.6 km offshore from Point Brower contrasts with what has been found in a different environment at West Dock where the ice-bonded permafrost is typically 5 or 6 times as deep for the same distance offshore. If the very hard layer in this hole at 16.2 m was ice, as we believe, then this is the first observation we have of such a thick ice layer in the sediments. This observation is reinforced by that of the U.S. Geological Survey (1979) in their boring number 13, which is located about 5.5 km offshore and almost on our drill line. They found a layer of massive ice containing clay soils about 0.6 m thick at the 10.0–10.6-m depth. This ice was just under the ice-bonded permafrost boundary at 9.8 m and will produce a thaw settlement of about 0.5 m when it

Table 10.—Sagavanirktok Delta and Jeanette Island drilling data.

Hole designation	Location or distance offshore	Water depth (m)	Sea ice thickness (m)	Drilling method	Date of drilling	Approximate depth below seabed (m)
Jeanette Island						
Hole A	≈1 km NE of Jeanette Island	7.5	1.72	rotary-jet	5/1/79	30.3
Hole B	≈1 km true North of dive site 11 which has coordinates of 70°19.2′N, 147°35.4′W	6.8	1.83	rotary-jet	5/2/79	15.4
Point Brower Hole C	70°17.3′N, 147°44.3′W ≈1.6 km NE of Pt. Brower	3.4	1.85	rotary-jet	5/3/79	21.9
Sagavanirktok	11.5 km from Sag. Delta #1 well	7.6	1.71	driving	5/25–26/78	18.3

thaws. Unfortunately, no temperature data are available from this hole since ice movement destroyed the pipe before it could be logged. We did not encounter ice-bonded permafrost in the hole about 11.5 km west of USGS boring 20. The ice-bonded permafrost table in the nearshore hole is more than twice as deep as that in boring 20. However, the temperatures there are nearly identical (about –1.6°C).

These observations suggest that offshore development in areas of fine-grained soils must consider the potential problems associated with ice-rich subsea permafrost containing segregated ice, its thawing and subsequent thaw settlement.

Flaxman Island

Four holes were drilled on a north–south line extending across the west end of Flaxman Island at the locations noted in Figure 14 and Table 11. South of the island, there was 9.4 m of clay underlain by gravel. Temperatures were relatively cold in these offshore holes, suggesting that the island may have recently retreated from these sites (Figure 24).

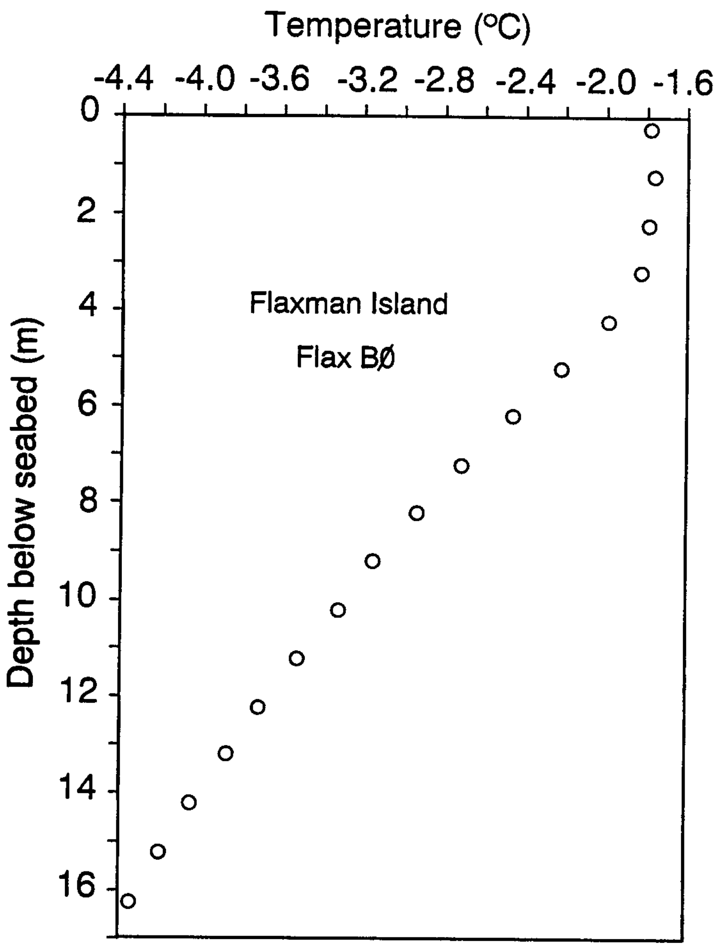

Figure 24.—Temperature profile from hole B which was located at the 3-m water depth on the Flaxman Island line.

Table 11.—Flaxman Island drilling data.

Hole designation	Location or distance offshore	Water depth (m)	Sea ice thickness (m)	Drilling method	Date of drilling	Approximate depth below seabed (m)
Hole A	0.893 km south of Leffingwell's cabin	2.2	1.75	rotary-jet	5/11/79	13.4
Hole B	0.165 km offshore from spit, bearing N2°E from Leffingwell's cabin	3.2	2.10	rotary-jet	5/13/79	17
Hole C	0.684 km offshore from spit, bearing N2°E from Leffingwell's cabin	6.1	2.01	rotary-jet	5/12/79	28.3

A layer of very hard sediments was found near the seabed in the offshore holes. This sediment appeared to be silt or clay with sand, sandy clay, or sandy silt. The USGS (1979) found similar hard sediments at 1–5-m in their bore number 18. They determined that this layer was hard, saturated gray silt, with numerous seams and pockets of gray, black, dense, fine-grained silty sand with small micaceous particles, about 2 m thick lying over a very stiff, saturated, gray silty clay which became less stiff below 5 m. USGS boring number 18 was located < 2 km ENE from our hole C in 11.3 m of water. This suggests that very hard sediments near the seabed may be a general feature of the Flaxman Island area.

Barrier Islands

Holes were drilled on the barrier islands along Alaska's northern coast of the Beaufort Sea from 1978 to 1981. The islands included Reindeer Island (five holes), Cottle Island (one hole), Stump Island (two holes), Thetis Island (one hole), Cross Island (two holes), and Flaxman Island (two holes). The drilling data are given in Table 12 and a temperature profile from Reindeer Island is shown in Figure 25.

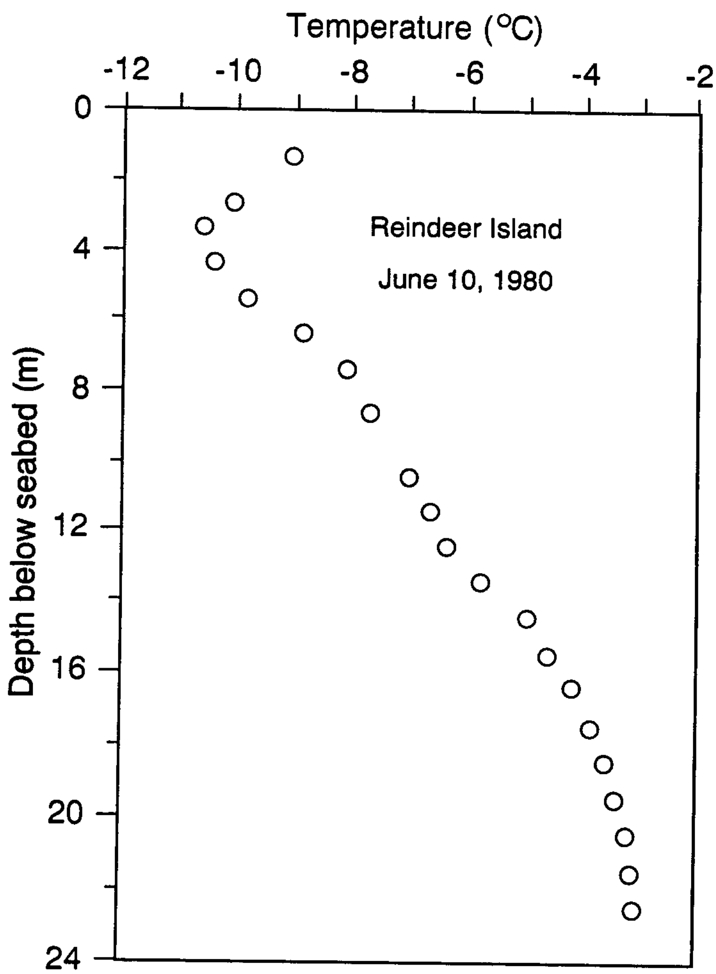

Figure 25.—Temperature profile on Reindeer Island.

Table 12.—Barrier Island drilling data.
The precise location of each hole is given in our previous OCSEAP reports.

Hole designation	Location or distance offshore	Water depth (m)	Sea ice thickness (m)	Drilling method	Date of drilling	Approximate depth below seabed (m)
Cross Island						
Hole 1	N320°E and 276 m from USCG navigation tower	—	—	rotary-jet	7/23/80	4.5
Hole 2	N294°E from USCG tower, and 103 m from south shoreline	—	—	rotary-jet	7/23/80	13.6
Flaxman Island	N118°E from Flaxman Island exploratory well. N90°E from Leffingwell's cabin. 213 m due N of shoreline	—	—	rotary-jet	7/28/80	9.9
Reindeer Island						
Hole D	N250°E and ≅300 m from the USGS tower on Reindeer Island, ≅4 m from the shoreline	—	1.91	rotary-jet	8/31/79	20
REIS 0	West end of Reindeer Island	—	—	jetting	5/24–25/78	12.2
REIS 1	West end of Reindeer Island	—	—	jetting/rotated by hand	8/19/78	6
REIS 2	West end of Reindeer Island	—	—	jetting/rotated by hand	8/19/78	3.5
REIS 3	West end of Reindeer Island	—	—	jetting	8/20/78	27
REIS 4	West end of Reindeer Island	—	—	jetting	8/25/78	16
Cottle Island						
COTT	—	—	—	jetting	8/24/78	≈6
Stump Island						
Stump I	—	—	—	jetting	9/21–22/78	10.5
Stump II	—	—	—	jetting	8/21–22/78	7.5

Ice-bonded permafrost was found under all the islands. The state of ice-bondedness in a particular hole varies with depth, sometimes in a complex fashion, and also with lateral position. The temperature profiles from the holes on Thetis and Reindeer Islands suggest that these islands moved over the present hole sites sometime during the past century.

A hole on Flaxman Island had a significant flow of gas which we collected for analysis by the USGS. The isotopic composition suggests that the gas was biologically generated.

VI. HEAT AND SALT TRANSPORT MECHANISMS AND MODELS

Transport Mechanisms

Thaw rates beneath the seabed vary by orders of magnitude. On the West Dock line seaward of the barrier islands, for example, thaw has been on the order of 10 m or even less after 10,000 years, while in the ramp and parabolic zones it is on the order of centimeters to tens of centimeters per year. This incredible variability must be due mainly to the fact that the mean annual seabed temperatures in Alaska's Beaufort Sea are negative, typically –0.7 to –1.5°C. (These are to be compared with mean surface temperatures onshore of about –10°C.) The seabed temperatures are low enough that salt must be present for significant thawing to occur. It was noted earlier that although some salt is present in the permafrost before submersion, it is probably insufficient to account for the thaw that occurs subsequently. Therefore, transport of salt from the ocean through the thawed layer beneath seems to be a key feature of subsea permafrost evolution. This idea can account for the variability of the thaw rates, since the efficiency of salt transport apparently depends sensitively on soil type. Evidence along the West Dock line indicates that this is reasonable because the slow rates occur in over-consolidated clay, while the rapid rates occur in material that is typically silty, sandy gravel. To understand the thaw rates quantitatively, then, it is evidently necessary to understand the salt transport mechanisms.

Early in this project it was realized that a potentially important salt transport mechanism was gravity-driven convection of the pore water in the thawed layer, the buoyancy of the relatively fresh water generated by thawing providing the driving mechanism (Harrison and Osterkamp 1978; Harrison 1982). This process would indeed be important if the pore water motion were primarily governed by Darcy's law, and if the molecular diffusivity of salt in the pore water were the same order of magnitude as in free solution. However, subsequent interpretation, based on detailed calculation and comparison with data from the West Dock line, indicates that this is not the case (Swift and Harrison 1983; Swift et al. 1984). Another candidate for a significant salt transport mechanism, resulting from the driving of pore water motion by surface wave action, also seems to be unlikely (Harrison et al. 1983).

The mechanism of transport of heat, unlike that of salt, is thought to be primarily conductive. This is indicated by the near linearity of the temperature profiles at all locations below the depth of seasonal variations. In a few cases, a slight curvature is observed, one of the many possible explanations for which could be slow pore water motion.

Until more is known about the salt transport mechanisms, it is doubtful that answers can be given to some of the obvious questions arising from the field observations, such as:

1) Why is the phase boundary temperature so uniform in the ramp and parabolic zones of the West Dock line, in the presence of known large variations in seabed temperature and salinity, and in pre-submergence surface morphology? What determines its particular value, –2.40°C, which indicates a pore water salinity at the phase boundary about 25% higher than that of normal seawater?

2) What is the mechanism of rapid salt transport implied by the rapid decrease of pore water salt concentration near shore at Tekegakrok Point, Elson Lagoon, with increasing distance from shore?

3) Why is the ice-bonded boundary sharp in some cases but not in others, or, perhaps equivalently, why do the ice-bearing and ice-bonded boundaries coincide in some cases but not in others?

One thing does seem to be certain: rapid salt transport occurs in some locations but not in others. There is another key question not clearly answered by either theory or field data: Does salt transport occur through the phase boundary and into the ice-bearing material below? This process, or the presence of salt initially, would influence the thermal properties of the material and greatly complicate heat conduction calculations. As noted, the limited field evidence indicates that this could be an important factor on the West Dock line, and in other locations it may be behind the failure of the ice-bearing and ice-bonded boundaries to coincide.

Although our interest was focused on salt transport mechanisms in the natural thawing of subsea permafrost, these mechanisms are equally important in the thermal design of insulated hot oil pipelines in subsea permafrost. For example, Heuer et al. (1983) (see also Walker et al. 1983) have shown that the optimum design is sensitive to the temperature at the contact between the developing thaw bulb and the permafrost that is still ice-bearing. This is the phase boundary temperature, and, as stated above, it is controlled by salt transport mechanisms so poorly understood that it cannot be predicted at present, or even any lower limit set on it. The transport of salt beyond the phase boundary may also be important in pipeline design.

Predictive Thermal Models

Even when heat transport is known to be conductive, there is coupling with salt transport processes, because pore water salt concentration controls temperature where brine and ice are in equilibrium. For this reason the lack of understanding of the salt transport mechanisms has restricted the development of thermal models. Thermal models can still be used to investigate the response of subsea permafrost in limiting cases, or at least in well-defined hypothetical examples. This approach has been used to infer shoreline stability in the Chukchi Sea by Lachenbruch (1957), Lachenbruch et al. (1966), and Osterkamp and Harrison (1978). It has also been used to infer the absence of permafrost in much of the southeastern Chukchi Sea as discussed earlier, and to infer the thaw rate at the base of the subsea permafrost along the West Dock line (Lachenbruch et al. 1982).

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APPENDIX A
LIST OF PUBLICATIONS

Published Articles

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APPENDIX B
ERRATA FOR ANNUAL REPORTS

1. Spring 1976 Report (Report on 1975 Field Work, Osterkamp and Harrison, 1976a)

The spring 1976 report was edited and re-issued as a joint Geophysical Institute (UAG R-249) and Sea Grant (76-5) Report (Osterkamp and Harrison, 1976b). The following errata are based on that report, rather than the original report for OCSEAP.

Page	Line	
iv	22	Change -0.7°C to -0.8°C.
v	11	Insert "in May" after "cover".
21	4 & 5	Replace sentence with "The age was found to be > 22,300 years."
34	Table 34	These values for the major ions are suspect because of lack of balance between cations and anions. See also page 35.
63	5	Change "-0.7°C" to "-0.8°C".

2. Spring 1977 Report (Report on 1976 field work, Harrison and Osterkamp, 1977)

Page	Line	
16	Subsection: Application of Simple Thermal Models	The result, Figure 8, is not in accord with more recent field observations.
Appendix II, page 5	Eq. 1	Subsequent analysis suggests that Equation (1) would better be interpreted as giving a reasonable lower limit to $\bar{T}$.
Appendix		Add the following references to the list:
II, pages 18 and 19		Brewer, M. C., 1958. Some results of geothermal investigations of permafrost in Northern Alaska. Transactions American Geophysical Union, 39, 1, p. 19-26.

Wilmovsky, N. J., 1953. Inshore temperature and salinity data during open water periods, Point Barrow, Alaska, 1951-1953, 14 pp. Was in Naval Arctic Research Laboratory, Barrow.

Wilmovsky, N. J., 1954. Inshore temperature and salinity data during open water period, Point Barrow Alaska, 1954, 5 pp. Was in Naval Arctic Research Laboratory, Barrow.

3. Spring 1978 Report (Report on 1977 field work, Osterkamp and Harrison, 1978)

Page	Line	
7	20	Replace "hold" by "hole".
7	22	Replace "Lackenbruch" by "Lachenbruch".
8	3	Insert comma after "temperatures".
18	34	Insert "steady state" after "corresponding".
24	3	Replace "Kruzenstorm" by "Kruzenstern, April, 1977".
24	9 (entry (d))	Place asterisk beside "2".
30	1	Insert "119 m offshore" after "Brewer (1958)".
30	19	Replace "exposed" by "emergent".
31 (Table 5)	2	Insert (1977) after "BEAUFORT SEA HOLES".
31 (Table 5)		Replace in column 1 "Harrison Bay - Thetis Island" by "Harrison Bay - south of Thetis Island".
33	7 (from bottom)	Delete "only".
38	2	Replace "k" by "K".
38	4	Replace "k" by "K".
39	2	Replace "k" by "K".
40	3	Replace "(ohm m^{-1})" by "(ohm $m)^{-1}$".
40	5	" " " "
42	9	Insert "probably" after "The sediments were".

42	11 (from bottom)	Insert "probably" after "this is".
46	1 (from bottom)	Insert "(Figure 25)" after "west dock".
52	Figure Caption 2 (from bottom)	Insert "Rogers and Morack (1978)" after "investigators".
55	Figure 26	Replace "MAT ≈ 1.2°C" by "MAT ≈ -1.2°C".
57	Figure 28	Replace "1.34°C by "-1.34°C".
Appendices I and II	II-1	This page is out of order.
II-2	1	Replace "Hole 78" by "Hole 75".
II-6	2	Replace entire line by "Extrapolation of May 18 and 28 data".
II-15	Fourth Reference	Replace "Bering Strait - Sky Cape" by "Bering Strait - Icy Cape".
II-15	Fifth Reference	Replace "Proceedings of the research international conference in permafrost" by "Proceedings of the second international conference on permafrost".
II-15	Sixth Reference	Replace "W. D. Harrison and W. D. Harrison" by "W. D. Harrison and T. E. Osterkamp".
	Twelth Reference	Replace "Reinhoed" by "Reinhold".
II-16		Add the following reference to the list:
		Rogers, J. C. and J. L. Morack, 1978. Beaufort seacoast permafrost studies. In: Environmental Assessment of the Alaskan Continental Shelf, Annual Reports, Vol. 11, p. 651-688.

4. Spring 1979 Report (Report on 1978, Harrison and Osterkamp, 1979)

Page	Line	
29	1 (from bottom)	Entry in first column is "5.2".

5. Spring 1980 Report (Report on 1979 field work, Osterkamp and Harrison, 1980)

Page	Line	
20	6	Replace "B and C" by "C and D".
25	18	Replace "uplife" by "uplift".
20	21	Add at end of paragraph, "The results are in Appendix D".
64	Entries 3 to 6 in column 7	Replace "- 27.45, 27.20, 18.81" by "24.6, 24.5, 24.6, shallow" respectively.
68	4 (Table 4)	Replace "Hole 700" by "Hole 701"
Appendix D-9	Entry 13 in column 1	Replace "70°" by "71°".

6. Spring 1981 Report (Report on 1980 field data, Osterkamp and Harrison, 1981)

Page	Line	
3	10 (from bottom)	Replace line by "A hole was driven near Nome, at a site about 326 m from shore along a line".
3	8 (from bottom)	Insert after "(Table 1 and Figure 1)", "The shortest distance to shore was 299 m."
4	8	Replace "driven" by "jetted".
8	8	Replace "increased" by "decreased".
9	6	Replace "29" by "27".
10	15	Replace "80-90" by "70-90".
21		Add to the reference list:
		Lewellen, R. I., 1977. A study of Beaufort Sea coastal erosion, northern Alaska. In: Environmental Assessment of the Alaskan Continental Shelf, Annual Reports, Vol. 15, p. 491-528.
Figure 2		Replace "326" by "299" on figure.
Figure 31		Replace this figure by Figure 21 of the present report.

56 (Table 1)	Entries 3, 4 and 7 in column 7	Replace "(7.28(?), ?, 28.8" by "17.28, 13.1, 24.9 respectively.
57	Entry 20 in column 3	Replace "5.586?" by "6.274?".
57	Entry 21 in column 3	Replace "6.291" by "6.303?".
Appendix B B-4	1	Replace "St. Michael Creek" by St. Michael River".
B-5	1	" " " " " " " "
B-6	1	" " " " " " " "
B-7	1	" " " " " " " "
B-8	1	" " " " " " " "
B-9	1	Line one should read "Charley Green Creek"

7. <u>1982 Report (Report on 1981 field work, Osterkamp and Harrison, 1982)</u>

In this report the West Dock hole location designation is not entirely consistent with that used in the previous report. In that report "distance from shore" really was distance from a fixed marker which was about 2 m from the tundra edge in 1981. If 2 m is added to the 1981 (1982 report) distance designations, they are consistent, in a fixed reference system (not one moving with the shoreline), with the 1980 designations.

Page	Line	
9	10-14	Delete sentence beginning "There is also...".
11	10-15	Delete paragraph beginning "The nature...".
15	3 (from bottom)	Replace "1.0350" by "1.0350 x 10^3".
16	3 (from bottom)	Delete sentence beginning "Its high value...".

18	5	Insert "If the interstitial water motion is governed by Darcy's law," at the beginning of the paragraph.
20	1 (from bottom)	Delete sentence beginning "These profiles suggest...".
21	2 (from bottom)	Delete paragraph beginning "The salt concentration...".
25	12	Delete paragraph beginning "Temperature profiles along...".
Table I	Hole 398	Note that the depth in column 7 is significantly greater than the greatest depth reached in temperature measurement (Appendix A), suggesting that the thermistor did not go down the hole all the way, or that a depth error has been made.
	Hole 433	"
	Insert under "Hole 433"	"Hole 438, same but 438 m offshore, 1.67, driven, May 27-30, 1981, 13.98, pressure".

APPENDIX C
SEDIMENTS IN SEA ICE

Summary

Sea ice in nearshore areas of Alaska's coasts has been found to contain concentrations of fine-grained sediment which are up to several orders of magnitude higher than the concentrations normally found in seawater. Measurements of the reduction in light intensity through the sea ice cover show that relatively small but visible sediment concentrations can drastically reduce the light intensity incident on the water column under the sea ice cover. Examinations of thin sections and ice samples from sea ice cores show that the sediments are clay and silt-sized particles with occasional fine sand and organic debris. The sediment incorporated into the sea ice was flocculated with floc sized a few tenths of a millimeter in maximum dimensions. Sediment flocs were found at grain boundaries, intersections of three or more grains, and in association with air bubbles, and occasionally they appeared to be interior to an ice crystal. High sediment concentrations were found to be restricted to the frazil ice in the sea ice cover, suggesting that the entrainment processes involve turbulence and frazil ice production during the freeze-up period. An evaluation of potential sediment entrainment processes suggests that those involving turbulence, frazil ice formation, sediment scavenging by individual frazil crystals and filtration by derived forms of frazil ice (e.g., grease ice, shuga, flocs, pans, floes) are the most viable.

A tentative sediment scavenging model for sediment entrainment in sea ice including the above components is proposed which is a combination and elaboration of the models of Naidu (1980) and Osterkamp and Gosink (1980). The proposed model assumes sufficiently windy conditions during freeze-up to produce the turbulence required for resuspension of fine-grained sediments and frazil ice formation and their entrainment in the water column.

A tentative filtration model for sediment entrainment in sea ice shows that the filtration process is also capable of concentrating sediment in the sea ice in the observed concentrations. This process will be most effective when the frazil ice has evolved into flocs, shuga, or small pans with dimensions up to a few tenths of a meter. The model suggests that as these forms evolve into large forms of frazil ice, the interstitial water and sediment load becomes trapped within the matrix.

When the derived forms of frazil ice congeal into an ice cover, it is proposed, sediment particles are flocculated by surface effects and by rejection of sediment particles from the growing frazil ice crystals which forces them into flocs in the interstices between crystals and into association with air bubbles.

While the above tentative models of sediment entrainment in sea ice are crude and somewhat speculative, they provide working hypotheses which can be tested as additional experimental data are obtained.

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Osterkamp, T. E., and J. P. Gosink. 1984. Observations and analyses of sediment-laden sea ice. *In* P. W. Barnes, E. Reimnitz, and D. M. Schell (eds.), The Alaskan Beaufort Sea: ecosystems and environment. Academic Press, New York.

APPENDIX D

GRAVEL SOURCES IN THE CHUKCHI SEA COASTAL AREA

Summary

Gravel appears to be scarce in the Chukchi Sea coastal area. Our program has provided some information about sources offshore, but mainly in a negative sense in that rock was encountered close to the seabed at one site off Ogotoruk Creek, at six sites between Cape Lisburne and Point Lay, and at one site in Peard Bay. Farther north, off Wrainwright and Barrow, reasonably good hole depths were achieved by rotary jetting, which suggests that gravel is not present in significant quantities or else is fine. Some soft rock was penetrated near Wainwright. Details are in our 1981 and 1982 reports. Some indications of gravel in the southern Chukchi Sea were found in our holes off Cape Blossom (18 km south of Kotzebue), Kotzebue Sound (where gravel, if present, is at a depth > 25 m), and off Rabbit Creek (36 km north of Cape Krusenstern). Details are in our 1978 report.

We are aware of several other sources of information, which are listed at the end of this appendix. AEIDC (1975) summarizes knowledge of gravel resources prior to 1975; of interest is the work of Creager and McManus (1966), who mapped several areas of gravel at the seabed in the southern Chukchi Sea. More recently, three engineering–geologic maps of northern Alaska (Williams 1983a,b; Williams and Carter 1984) cover the Meade River, Wainwright, and Barrow quadrangles, including the coastal area onshore from Tolageak, just south of Point Hope, to Barrow and beyond to Cape Simpson in the Beaufort Sea. Studies of sediments offshore in the northeast Chukchi Sea are currently being conducted by the OCSEAP project of Phillips and others. Three reports now available cover the nearshore region from Wainwright to Barrow (Phillips et al. 1982; Phillips and Reis 1984, 1985).

Some other information is also available from the coastal or near coastal areas of the southern Chukchi Sea, usually in connection with airport construction or improvements by the State of Alaska, Department of Transportation and Public Facilities (DOTPF), 2301 Peger Road, Fairbanks, AK 99701. The sites investigated by DOTPF include Shishmaref, Noorvik, Kivalina, Point Hope, Wales, Barrow, and possibly others. The offshore sites investigated were in Shishmaref and Kivalina lagoons. Permafrost was found under the latter. Only the reports that we have read are included in the list below. Gravel has been dredged from Kotzebue Lagoon.

Studies for proposed developments on land such as the Wainwright terminal, the Red Dog mine, and the Kotzebue to Chicago Creek highway have yielded, or will yield, other information about gravel sources that may be relevant to offshore development. Of special interest is the Draft Environmental Impact Statement for the Red Dog mine project (Environmental Protection Agency and Dep. Interior 1984).

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A GEOGRAPHIC BASED INFORMATION MANAGEMENT SYSTEM FOR PERMAFROST PREDICTION IN THE BEAUFORT AND CHUKCHI SEAS

PART I.
SUBMARINE PERMAFROST ON THE ALASKAN SHELF

by

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Final Report
Outer Continental Shelf Environmental Assessment Program
Research Unit 516

September 1978

ACKNOWLEDGMENTS

This is the first work done by this Russian-born author, on American soil, in the American scientific environment; and for a part of America that, in a geological sense, could be considered to be a part of the Asian plate and physiographically indistinguishable from Siberia.

For their help from beginning to end we would like to thank Professors Jack D. Ives, R. Berry, our colleagues in INSTAAR and the World Data Center in Glaciology, and Mr. J. Fletcher (NOAA). For having provided the data we would like to thank Dr. M. Loughridge, H. Hittelman, K. Potter (NGSDC, NOAA) and World Data Center-A, Oceanography.

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The author may only hope that the experience from the Eurasiatic Arctic, that was used in this work, will be as useful for the study of the American Arctic as the friendship, support, and new knowledge he has received here.

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LIST OF MAPS (Continued)

NOTE: It is difficult to use the reduced copies of the computerized environmental maps which are included in the text of Part I for BLM needs. Therefore, an atlas of these maps has been prepared and submitted separately from this document. The Atlas includes the 31 maps, which were made on a base map especially prepared by Science Applications, Inc., in a scale directly appropriate for practical use by BLM.

INTRODUCTION

According to the Program Development Plan (NOAA, BLM, 1976), this work is a part of the investigations related to the identification and estimation of the potential hazards posed by the environment to petroleum exploration and development (p. 3–3). This study involves quantitative assessments of information that can be used to assess future alterations resulting from various phases of lease site development. This study uses a computerized approach and analytical design which allows conclusions to be stated in terms of probability. The main products are computerized source, derived, and composite maps of several physical fields—oceanographic, geologic, topographic. The maps delineate environmentally sensitive areas; i.e., those conducive to formation of submarine permafrost. These areas are potentially dangerous for exploration and development of oil and gas in the Beaufort and Chukchi seas as a whole, and specifically in the limits of the lease areas. The maps are in scales directly applicable to BLM needs for prediction, assessment, and regulation.

TASK OBJECTIVES

This final report includes two parts which correspond to the two principal objectives of the work.

The *first objective* was to develop a computerized system to aid in the prediction of the distribution and characteristics of offshore permafrost. Development of this system involves (1) the gathering and study of all source data about direct and indirect indicators of permafrost in the given area (depth, temperature and salinity of water, topography, bottom deposits, ice, etc.); and (2) the generation of source and derived maps and construction of a candidate area map for submarine permafrost in the Beaufort and Chukchi seas. The decisive factor in this work was the close relationship with the NOAA Environmental Data and Information Service, especially the National Geophysical and Solar-Terrestrial Data Center (NGSDC) in Boulder, that also operates World Data Center-A for Solid Earth Geophysics and for Solar-Terrestrial Physics. We also used data from World Data Center-A, Oceanography, in Washington, D.C., and Glaciology in Boulder (INSTAAR). A description of the system and results is given in Part I of this report.

The *second objective* was to undertake a comprehensive review and analysis of past and current Soviet literature on subsea permafrost and related natural processes. The data and concept analysis on subsea permafrost of the Eurasiatic part of the Arctic in its relationship with the Arctic development in the Pleistocene, is given in Part II of this report as a monograph with bibliography related to the problem.

The work was done by M. Vigdorchik (Principal Investigator), B. Skholler (Computer Programmer), and J. Adams (Senior Graphic Artist) during the period of October 1976–September 1978.

SUMMARY OF RESULTS

1. An evaluation of existing environmental data on submarine permafrost of the Alaskan Shelf has been made. The reliability of the data and the gaps have been defined.
2. The paleoenvironmental aspects of the submarine permafrost of the Beaufort and Chukchi seas have been discussed. According to the specifics of the origin and development of the permafrost, the Beaufort Sea shelf has been divided into three parts: (a) the shelf area suitable for submarine relic permafrost–western part; (b) the area with low suitability for relic permafrost–central part; (c) the area without relic permafrost–eastern part. Estimations of thickness and other specifics of submarine permafrost in each area have also been made.
3. A computer system has been developed for the evaluation of all existing environmental data on recent subsea permafrost development. Thirty computerized maps of different oceanographic, geologic, glaciologic, and other parameters have been generated: source data maps, derived maps (three generations), and a composite map specifying the candidate area for submarine permafrost development. To develop the system a composite mapping algorithm for submarine permafrost prediction has been made.
4. The system has been checked in the Beaufort Sea areas with known permafrost. A comparison with Canadian data on submarine permafrost has shown positive results and high correlation.
5. The system can be readily updated according to new data and in this way the results can be enhanced.
6. The shelf maps showing suitability for submarine permafrost have been compared with the BLM lease nomination map. The extension of the suitable areas for permafrost at each nomination site has also been calculated.
7. The same work was done for the Chukchi Sea. It was found that there is a limited distribution of areas suitable for ice-bonded submarine permafrost. The areas were to the northwest from the Barrow Canyon and at some sites along the coast line.

GENERAL STRATEGY AND APPROACH

The study of submarine permafrost distribution and offshore ice content in the perennially frozen rocks is a part of the interdependent biological, physical, oceanographic, and geological investigations intended to minimize the environmental damage during future exploitation of oil and gas resources. The experience obtained in the terrestrial and shelf environments of northern Eurasia and America, along with the scenario worked out during the Barrow synthesis meetings (1977, 1978), clearly shows the danger of subsea permafrost, first for the production structures themselves, and then for biota and populations. Such peculiarities of the offshore permafrost as the high variability of its surface position in vertical and horizontal directions, sporadic distribution, and multi-layer character (two or more) could pose a threat to pipelines and the foundations of offshore structures because of possible differential subsidence through changes in the thermal regime. The buried thick ice lenses in the "ancient canyons" which cross the Arctic shelf usually in a meridional direction (see below and Part II) along with the thawing ice wedges in the thermokarst zones of the shelf areas formed in the case of their exposure during the last glaciation need to be taken into account.

We will discuss more of this problem later, but in the sense of the potential danger, it is useful to know that in the "ancient buried canyons" ice lenses 5–15 m thick were met in the Eurasiatic Arctic shelf in the first hundred meters from the sea floor in the narrow, usually meridionally oriented strips about several hundred meters in width. Ice wedges, in different stages of thawing under seawater influence, usually form the polygonal-tetragonal network straight from the seabed. These have an average size of about 100–150 m at a distance of 400–500 m one from another and may give the subsidence about 10 m. These figures might be kept in mind when assessing the potential difficulties for pipeline or foundations during the first phases of exploration and later when hot oil starts to flow from offshore wells.

The development and main features of the submarine permafrost on any part of the Arctic shelf depend on many factors and conditions:

1) Geologic structure of the shelf.
2) Hydrogeological structure of the shelf, specifically the formation, dynamics, and chemical composition of the underground water.
3) Morphostructural features of the shelf, the peculiarities of its relief.
4) History of the geological development of the shelf and adjacent coastal areas in Pleistocene and Holocene.
5) Recent tectonic activity.
6) Peculiarities of the modern marine basin, including bathymetry, salinity, and temperature regime of the seawater.
7) The specifics of the hydrological regime on the shelf close to the coast connected with the river activity.
8) Role of ice in the zone of ice/sea bottom development.

In a broad sense the specifics of subsea permafrost interaction, distribution, and local peculiarities are always related to the paleoenvironmental and modern environmental conditions. Knowledge of these eight factors is necessary to give a more or less comprehensive picture of the distribution and thickness of permafrost on the Alaskan shelf. Today, not all of these factors are represented by a sufficient data base for a comprehensive analysis. Pieces of information have been collected at various times and places and the coordinated programs to synthesize the existing data are now in a developmental stage.

The most comprehensive evaluation of current knowledge on Alaskan subsea permafrost has been done by T. Osterkamp and W. Harrison in their annual report of 1977–78. This report also summarized all existing direct data about submarine permafrost in the Beaufort and Chukchi seas, its probing, thermal regime models and data analysis. The study of Lachenbruch et al. (1962), Lachenbruch and Marshall (1977), Lewellen (1973, 1976), Osterkamp and Harrison (1976, 1977, 1978), Rogers et al. (1975), Rogers and Morack (1976), Chamberlain et al. (1977), and Sellman et al. (1976) took place only at two sites in the Chukchi Sea (near Kotzebue, Rabbit Creek, and near Barrow) and at two sites in the western Beaufort Sea (Elson Lagoon, Prudhoe Bay). In the Canadian Beaufort Sea the submarine permafrost was studied by Hunter et al. (1976) in the Mackenzie Delta.

Drilling, probing, seismic study and modelling were the major methods used at these sites. The results of these investigations are the following:

1) A negative temperature is typical for the shelf deposits at a depth of at least 80 m from the seabed.
2) The maximum distance from the shore where subsea permafrost was met by drilling was 17 km at Prudhoe Bay and about 25 km at the eastern Beaufort Sea (MacKay 1972).
3) The upper surface of ice-bonded permafrost may be quite variable with relief changes of several tens of meters over short distances and may be near the seabed at sites far offshore (according to the seismic studies).
4) Data on lower surfaces of permafrost are not sufficient at any site investigated.
5) Some data indicate the two bonded permafrost layers under some islands (Prudhoe Bay, Reindeer Island).
6) The grain size, pore water salt concentration, and temperature of the deposits play an important role in the distribution of ice-bonded permafrost in the shelf deposits zone of negative temperature.

The site-specific study of submarine permafrost and modelling are of great value for a better understanding of the nature, peculiarities, and thermal regime. Perhaps the site-specific character of the investigations limits their application to definition of areas of submarine extension because of the highly variable environmental conditions of the Alaskan shelf. It is understandable that site-specific information on offshore permafrost from the seismic or drilling methods cannot be obtained for all locations on the continental shelf. In order to meet the need for predictive information on the potential distribution and characteristics of offshore permafrost, a different kind of modelling approach has to be used, drawing on all existing data. This work is an attempt to use all available environmental data to produce the series of computerized maps of different "generations" to predict areas suitable for subsea permafrost. In addition to providing a tool for storing and retrieving geographically based data, the system is used to produce derived maps showing the important aspects of the conditions for the extension of submarine permafrost and its possible character.

Use of the data management system provides a comprehensive framework for recording, storing, manipulating, and displaying mappable information used in preparing planning studies. This program entails the use of electronic data processing and computer graphics to organize and present a variety of complex data in an orderly and systematic manner. Data are stored on magnetic discs allowing retrieval, analysis, and display of the data in the form of computer-generated maps. The program gives a dynamic base that can be readily updated, and it allows the evaluation of many alternatives. The system can automatically generate a great deal of secondary data, saving time and money during the collection phase of the project. During the data analysis phase, it was possible to aggregate a number of subjective judgments into an integrated set of evaluations. These evaluations identify the most suitable offshore candidate areas for permafrost, based on a multiplicity of geomorphological, geological, cartometrical, geophysical, and oceanographic factors. The system provides a complete trace of the decision-making process as well as an up-to-date base which can be used for siting and routing and environmental studies of this territory. This computer-oriented approach allows the coordination of information for project analysis, to control the selection and format of the data used, and to establish their value.

In this study we tried to analyze both the paleoenvironmental and modern environmental data to trace the area suitable for submarine permafrost. Of course, our attempts are restricted by the distribution and quality of the existing materials. New data might help answer the remaining questions.

DISCUSSION OF PALEOENVIRONMENTAL ASPECTS OF ALASKAN SUBSEA PERMAFROST DEVELOPMENT

Usually the paleoenvironmental aspects of the submarine permafrost are the most talked-about. One basic concept for subsea permafrost origin is that of a shelf exposure, but this is only one idea. In Part II we describe the types of the submarine permafrost on the shelf with undersea deposits saturated in saline cold water, supplied from the continent, or the perennially frozen rocks of the shelf which were formed under coastal conditions, often beneath a layer of low-temperature saline water. The long term role of cooling effects of the thick (250–600 m) body of the continental permafrost has not yet been the subject of adequate investigation.

Following the popular "shelf exposure" concept that helps to assume the shelf submergence–emergence history, taken into consideration in the models of the thermal conditions in submarine permafrost and its configuration (Lachenbruch 1957; Osterkamp 1975; Harrison and Osterkamp 1976; Lachenbruch and Marshall 1977), we will discuss the different possibilities for such exposure. For this purpose we will use the glacioeustatic sea level history of Hopkins (1973, 1977) for the time of the Last Glaciation, direct and indirect information on differential vertical movements at Arctic coastal areas, and some form of a morphostructural analysis.

Indicators of Vertical Movement at Different Parts of the Beaufort Sea Coast

Let us first discuss the role of the coastal retreat data for evaluation of vertical movement in the coastal area. As noted in the "Interim Synthesis: Beaufort/Chukchi, August 1978" (pages 122–126), the rates of coastal retreat differ depending on variations in the coastal bluff composition (including content of ice), exposure, morphology of the coast, and adjoining sea bottom. This rate also changes from year to year. The special investigations in the Laptev and eastern Siberian seas made by F. Are (1967, 1972) and P. Sisco (1970) show that this exposure rate in the Arctic does not depend on coastal exposure or steepness, that the snow cover reduces this rate by 30–40% in comparison with the uncovered coastal slopes, and that solar radiation and the temperature of air (in the case of thermodenudation) and seawater (thermoabrasion) are the key factors influencing this rate. At the same time, the average rate of coastal retreat in the limits of the different geological and geomorphological structures usually reflects the neotectonic recent vertical movements of these morphostructures. Of course, the highest rates of coastal retreat in the Arctic are on promontories and points, but it is important that "many bays and estuaries have persistently caspate outlines showing that thermal erosion and thermokarst collapse tend to cause parallel retreat of the shoreline regardless of coastal orientation" (Interim Synthesis, 1978, p. 126). Therefore, to use the coastal retreat data for approximation of the recent movements, we must consider only average rates in the areas with the different morphostructural conditions. The average figures help to exclude local factors and to specify the role of the seawater as a major agent of thermoabrasion, keeping in mind that the thermoabrasion processes are more intensive and consequently the coastal retreat more active at the areas of submergence or the low rate of the positive vertical movements. In evaluating vertical movements on the Beaufort Sea coast we have made a comparative analysis of the rates of coastal retreat and vertical movement according to data from the Laptev and East Siberian seas. The vertical uplift data of Baskakov and Shpaikher (1970) are based on the graphic method for the seaboard of the Kara, Laptev, and East Siberian seas and the Bering Strait. The method assumes that the year-to-year sea level variations are the same order of magnitude in all regions of a given sea or adjacent seas (Mesheryakov 1964). The rate, V, of the present-day movements of the seaboard and the calculation errors were calculated by means of the following equations:

$$V = \frac{\Sigma Ht}{\Sigma t^2} \qquad \delta_V = \pm \frac{\delta H}{\Sigma t^2} \qquad \delta_H = \pm \sqrt{\frac{\Sigma H^2 - V\Sigma Ht - b\Sigma H}{n - 1}}$$

where:

V = the rate of present-day vertical movements of the seaboard;

H = the mean annual level;

b = the mean long-term annual level ($\Sigma H/n$);

n = the number of years of observations (length of the series);

t = the serial number of the year counted from the mid-point of the observation period;

δ_V = the RMS error involved in the calculation of the rate of present-day coastal movement; and

δ_H = the RMS error involved in the calculation of the sum total of annual levels (ΣH).

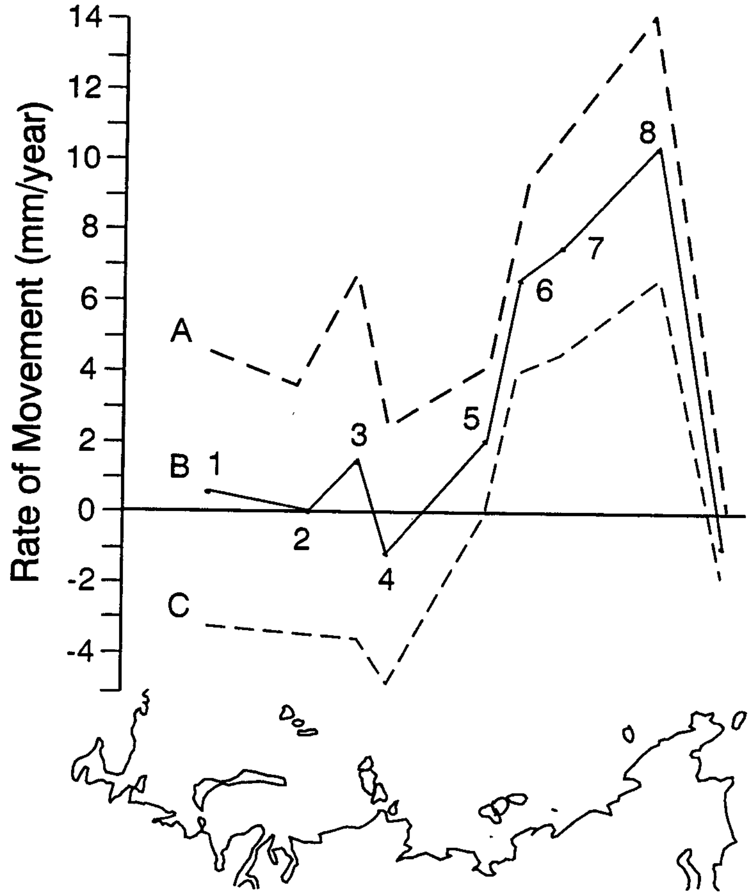
The rate of present-day coastal movements was calculated for nine polar stations, four of which are located in the Kara Sea, one in the Laptev Sea (Tiksi Bay), two in the East Siberian Sea (Cape Shalaurov and Ambarchik Bay), one in the Chukchi Sea (Cape Schmidt), and one in the Bering Strait (Providence Bay). The calculations were based on coastal observations of the sea level during the years 1951–65 (Table 1, Figure 1). The RMS errors make it possible to assess the present-day coastal movements more objectively by characterizing the possible variation range of the rate of these movements. We see that seaboards are currently rising, except for the regions of the Vilkitski and Bering straits (where the coasts are settling). The rate of uplift increases from west to east, from values not exceeding +1.5 mm/year in the Kara Sea to +10.5 mm/year in the Chukchi Sea. An especially high coastal uplift rate, ranging from +12.7 to +39.5 mm/year, was calculated by Pierce (1960) for the southeastern coast of Alaska (Table 2). Pierce's results were mostly confirmed by Bird and Barnett on the northern Canadian seaboard. From his sea level analysis at Churchill (1940–64), Barnett (1966) determined a coastal uplift rate of 7.3 mm/year. According to Bird (1954), coastal uplift of as much as 90 cm over a period of 100 years was registered in the region of Boothia Peninsula, and some 80 cm per 100 years between Chesterfield and Cape Eskimo. According to all this data on the Kara and Laptev sea coasts where the rate of uplift is very small and does not exceed the calculation error, the coast is relatively stable. The point of observation (Tiksi Bay,

Table 1.—Relative rate of present-day vertical coastal movements in Arctic seas in 1951–65, mm/year. After Baskakov and Shpaikher (1970).

Serial Number		Relative Rate of Movement	RMS Error of the Rate of Movement	Sea
1	Amderma	+0.7	±3.9	Kara
2	Dikson Island	+0.1	±3.6	Kara
3	Pravda Island	+1.5	±5.2	Kara
4	Cape Chelyuskin	−1.2	±3.8	Kara
5	Tiksi Bay (Bulunkan Inlet)	−2.2	±2.0	Laptev
6	Cape Shalaurov	+6.7	±2.6	East Siberian
7	Ambarchik Bay	+7.5	±3.0	East Siberian
8	Cape Schmidt	+10.5	±3.8	Chukchi
9	Providence Bay	−1.0	±1.0	Bering Strait

Table 2.—Variations of absolute elevations on the southeastern Alaskan seaboard. After Pierce (1960).

Observation Site	Observation Period	Change in Coastal Elevation over Observation Period (cm)	Rate of Vertical Coastal Movement (mm/year)
Skagway	1909–1959	89	+17.8
Haines	1922–1959	82	+22.2
Tepri Bay	1922–1959	82	+22.2
Muir Inlet	1940–1959	67	+35.3
Willoughby Inlet	1939–1959	53	+26.6
Bartlett Cove	1938–1959	83	+39.5
North Inian	1902–1959	135	+23.7
Lisianskii Bay	1917–1959	58	+13.8
Elfin Cove	1938–1959	40	+19.5
Hoopak	1932–1959	57	+15.8
Svenson Harbor	1901–1959	99	+17.0
Funter Bay	1922–1959	52	+14.0
Auke Bay	1937–1959	42	+19.3
Juneau	1911–1959	61	+12.7
Annexe Creek	1937–1959	30	+13.6
Greeley Point	1937–1959	30	+13.6



A = Rate of movement (mm/year); B, C = Rate of movement (mm/year) taking account of the RMS error; 1, 2, 3 . . . .9 = serial numbers of observation sites (See Table 1).

Figure 1.—Spatial variation in the rate of present-day vertical movements in Arctic seas. After Baskakov and Shpaikher (1970).

Bulunkan Inlet) has a relative rate of movement of +2.2 mm/year. This is part of the Laptev Sea submarine permafrost and thermoabrasion study polygon. According to Are (1972, 1973) and others, the average rate of the coastal retreat here is about 4–5 m/year. The seaboard of the eastern part of the Laptev and East Siberian seas with a higher rate of recent uplift has proportionally lower rates of average coastal retreat.

The data of Stovas (1965) and Gutenberg (1941) indicate a possible settling of the Bering Strait western coast and uplift of the areas east of the strait. Pierce (1960), Barnett (1966), and Bird (1954) also emphasize the very high uplift rate of the eastern part of the Bering Strait, especially east of Cook Inlet and on the northern Canadian seaboard.

Figure 2 shows the average rate of coastal retreat in the Beaufort Sea. Data from the Interim Synthesis (1978, p. 125) shows three areas with three different rates of retreat:

1) Pt. Barrow–Harrison Bay—4.4 mm/year
2) Colville River (148°?) to 142°—1.6 mm/year
3) 142° to Mackenzie River—1 mm/year (in the "Synthesis": Demarcation point–Mackenzie River delta).

Using the comparative data on coastal retreat and the vertical uplift equivalents from Siberian seas we consider that the recent uplift rate at the first area is about 2.2 ± 2.0 mm/yr, which identifies that territory as a relatively stable one (Figure 3).

The second area, with coastal retreat through time less than the first one, has a rate of vertical uplift of about 6.6 ± 2.6 mm/yr, and the third one, Canada, about 10.34 ± 3.7 mm/yr. This means that the first area is considerably late in the general uplift of the Arctic coasts, the second area is transitional, with a clear picture of the modest uplift, and the third one could be characterized as an area with a high rate of recent vertical uplift, comparable with the rates of southeastern Alaska, the territory situated to the east of Cook Inlet. It is possible to find the explanation of the differential vertical movements on the Beaufort Sea shelf in its geological and geomorphological peculiarities.

Morphostructural Peculiarities of the Coastal Areas Related to Vertical Movements

The differences in the rates of coastal retreat and the rate of vertical movement at the three named areas could be explained by their different morphostructural conditions. The first area (Point Barrow–Harrison Bay) is a low, flat plain with very high density of lakes generally oriented north. It is situated to the west from the major zone of seismic activity which has meridional orientation here. This zone is a continuation of the Aleutian Islands–Cook Inlet active zone. This gigantic transitional, tectonically active zone divides Alaska and its northern slope along 150°, 148°–142° and includes Prudhoe Bay, the seismic zone around

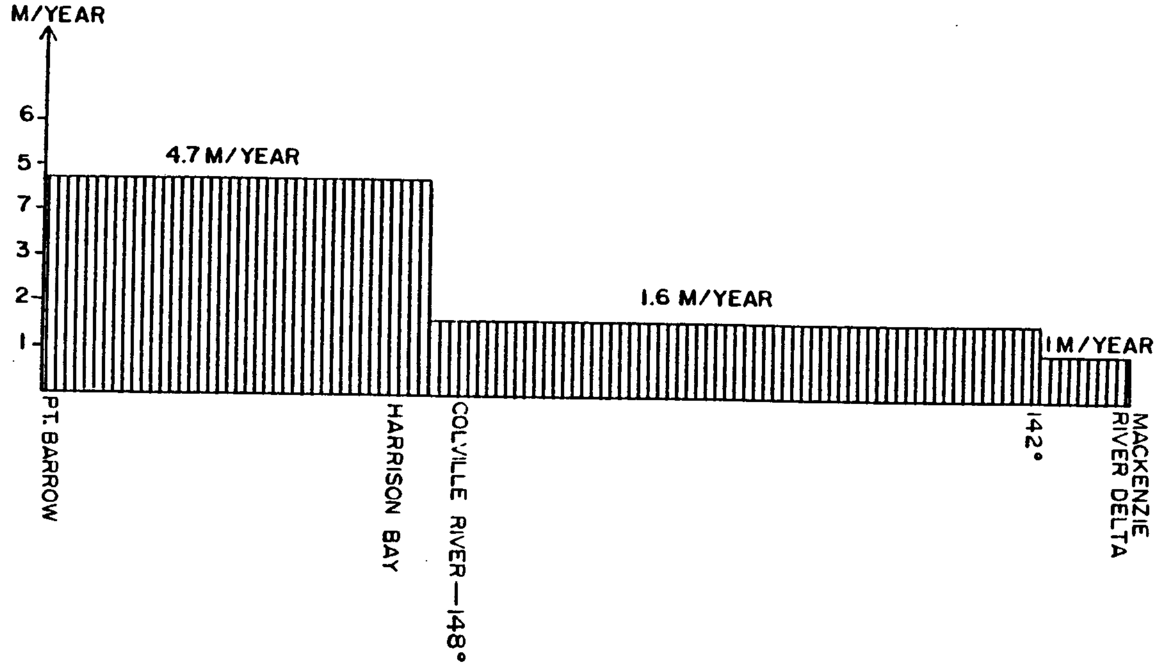


Figure 2.—Average rate of coastal retreat according to the Interim Synthesis, August 1978, Figure 3.13, pp. 125, 126.

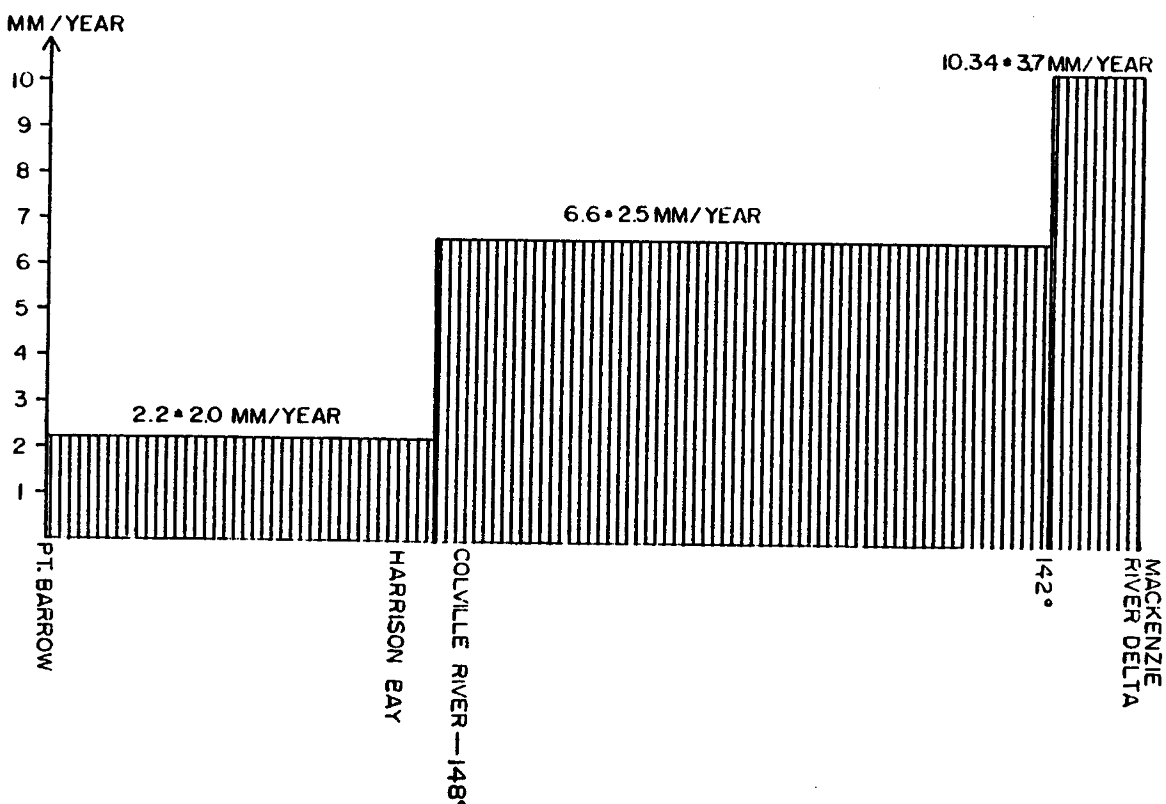


Figure 3.—Possible rate of the Beaufort Sea recent coastal uplift proportionally to the average coastal retreat.

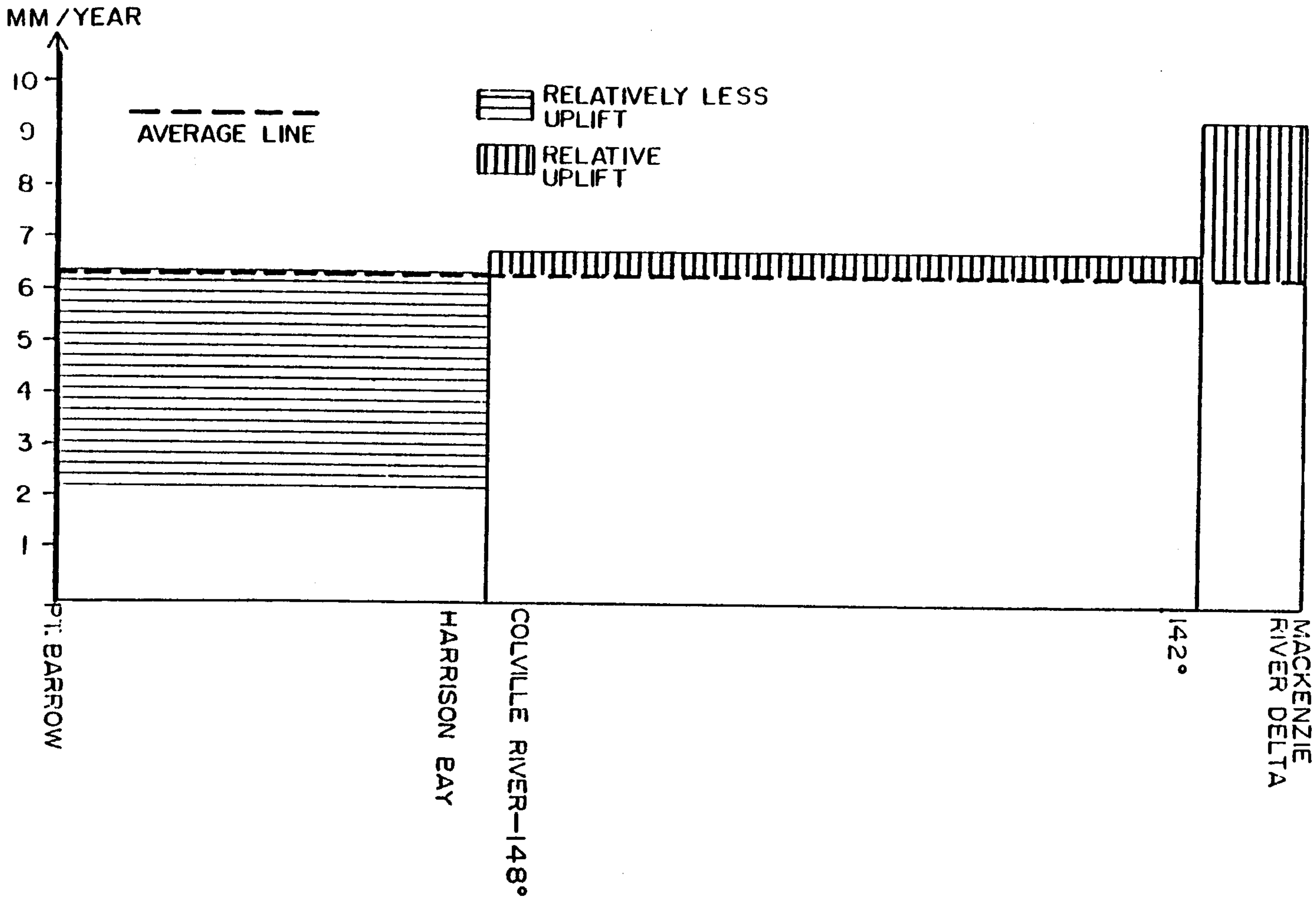


Figure 4.—Relative rate of the recent uplift of the three major morphostructures of the Beaufort coast.

Barrier Island and the point 30 km offshore (with the largest earthquake registered (M_L = 5.3) and a series of aftershocks) showing an ENE–WSW seismic trend along the axial traces of the offshore folded structures. The study made by Biswas (1977) traced this zone in northeast Alaska, as shown in Figure 5. Figure 6 shows the major transitional active zone of Alaska through distribution of the earthquakes. Both pictures are reproduced here from the "Interim Synthesis," (1978, pp. 103–104). If our first area is situated to the west of the tectonically active zone and it is not included in this zone, the second area is part of it. We might also suggest that the boundary between our first and second areas could be traced, not along the Colville River and its delta, but following the western limit of the named seismically active zone. At latitudes 70° 30′ –71°, the tectonic boundary corresponds more to 148° and this meridian looks more appropriate as a divide between the two morphostructures of our first and second areas. Thus, the second area fully corresponds to the tectonically active transitional zone (148°–142°). The third area is mostly situated to the east from Barter Island/142° and extends possibly towards the Mackenzie Delta, or east from the major seismic zone, and it is an area with low density of lakes, with domination of the higher absolute altitudes and more steep and narrow part of the Beaufort Sea shelf.

According to Bulard's suggestion (1971) the Anchorage–Prudhoe Bay transitional active zone could be considered as the boundary of the American and Asian plates. During the Paleozoic stage of the Alaskan Cordillera geosyncline development, this zone was one of the zones of relative uplift. The thickness of the Paleozoic system reached only 500 m. In the western and southeastern parts of Alaska this thickness reached several thousands of kilometers. During the Mesocenozoic stage this structure played the role of a barrier. Then and now the tectonic movements in this zone and to the east and west from it had, and have, different character and magnitudes. In the late Cenozoic the eastern part of the northern Alaska plain (White Hills Province) was under relative uplift. The fact that one can see dislocated tertiary rocks in the centers of the dome-structures here shows a result of the positive movement. On the other hand, the basin of Teshekpuk Lake in the west could be characterized by the great thickness of the unconsolidated Quaternary deposits (see Williams et al. 1977) on top of the Cretaceous Colville formation. The distribution of the surficial deposits of this part of the coastal plain is shown in Figure 7. This means that the western part of the Alaskan Arctic plain, or our "first area," was significantly late in the uplift of the coast and shelf in comparison with the more eastern parts of the plain, or our "second area," and "third area." Both the Tertiary and Quaternary periods at Alaska were the periods of relative positive movements of the eastern part of Alaska. At the same time the western plate was involved in relative submergence related to the development of the marine basins of the Arctic and North Pacific. Figure 8 represents the scheme of the major linear morphostructures of Alaska

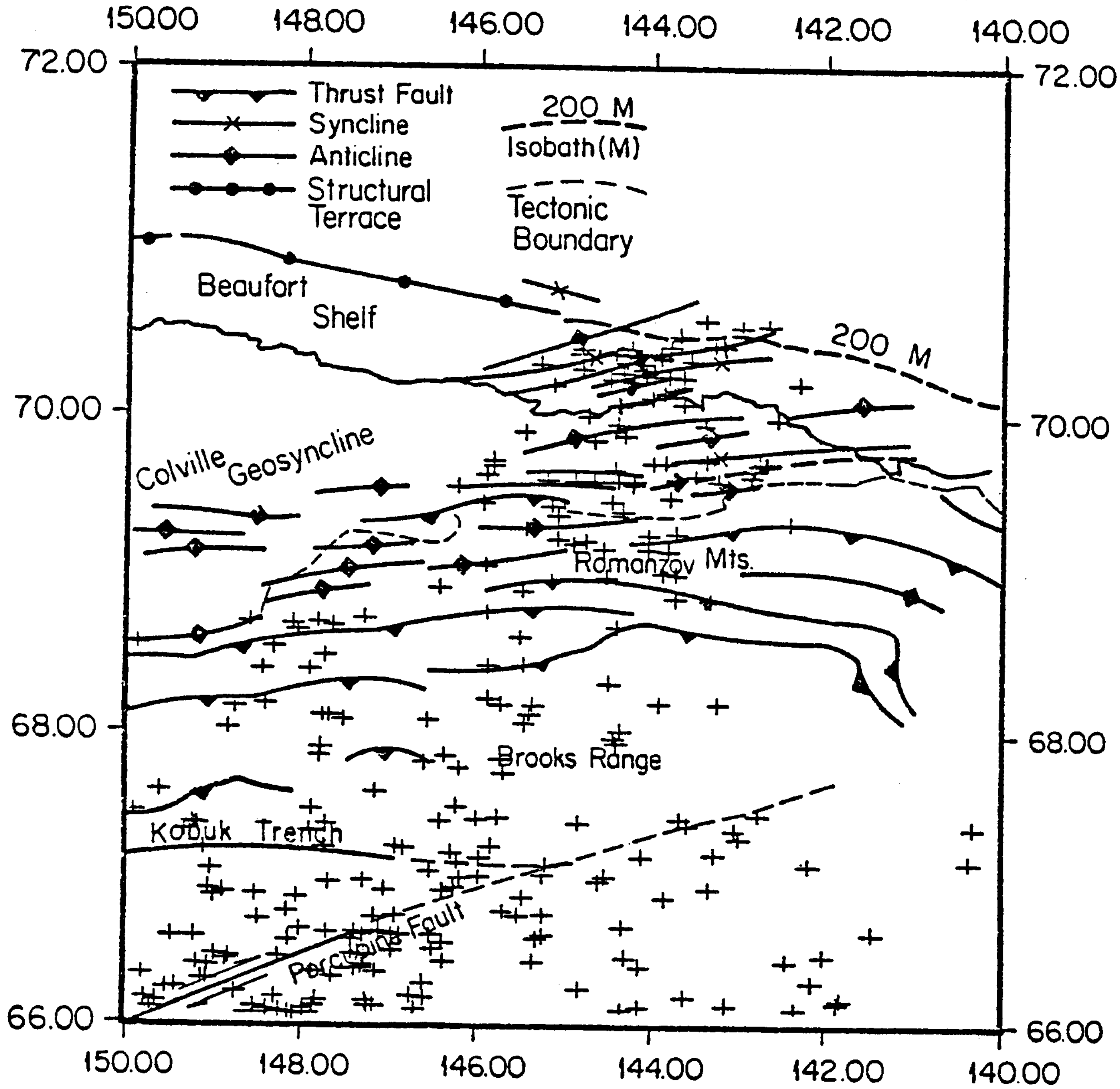


Figure 5.—Epicenters of earthquakes in northern Alaska according to Biswas (1977).

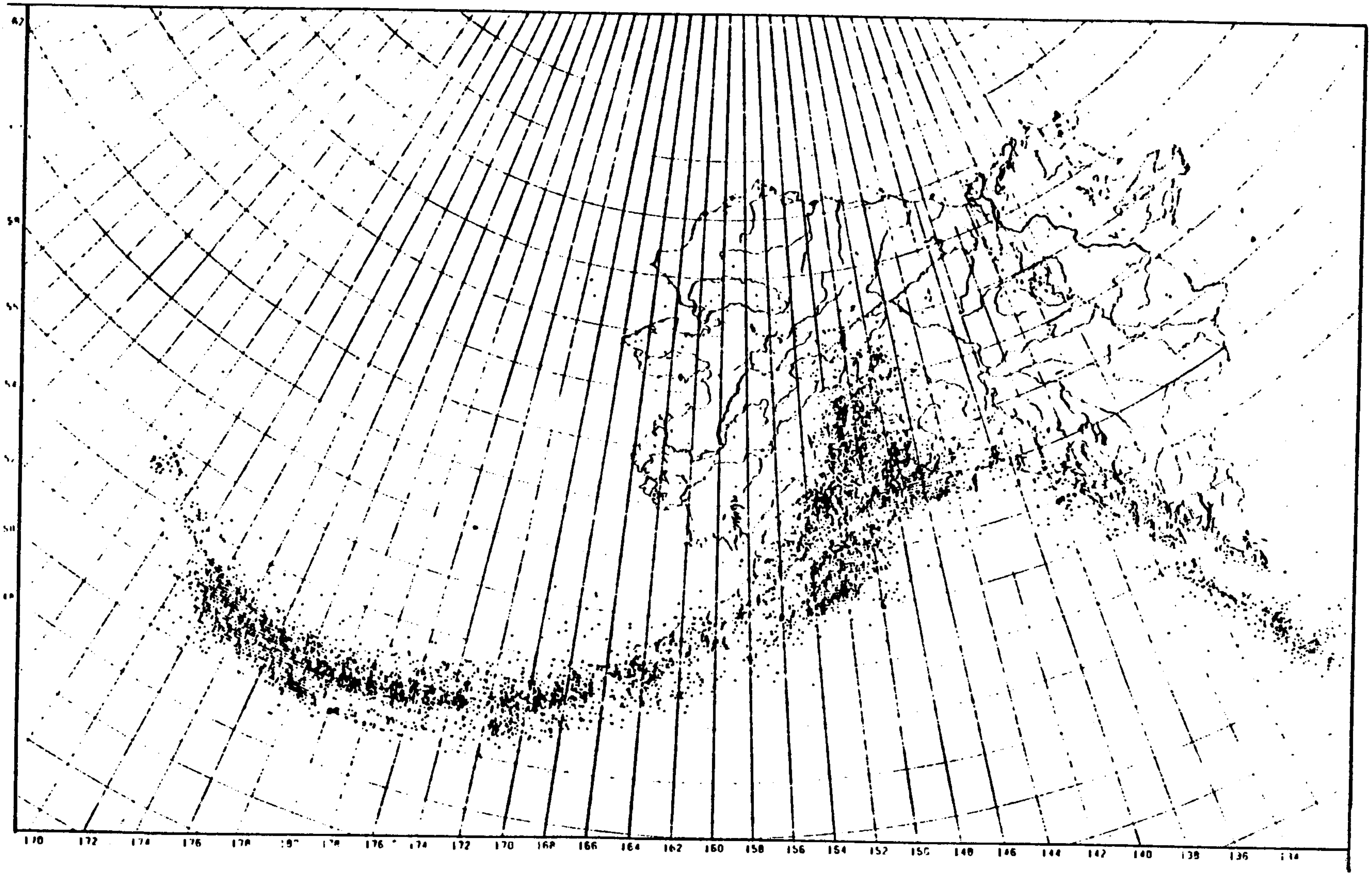


Figure 6.—Earthquakes in and near Alaska (Interim Synthesis 1978).

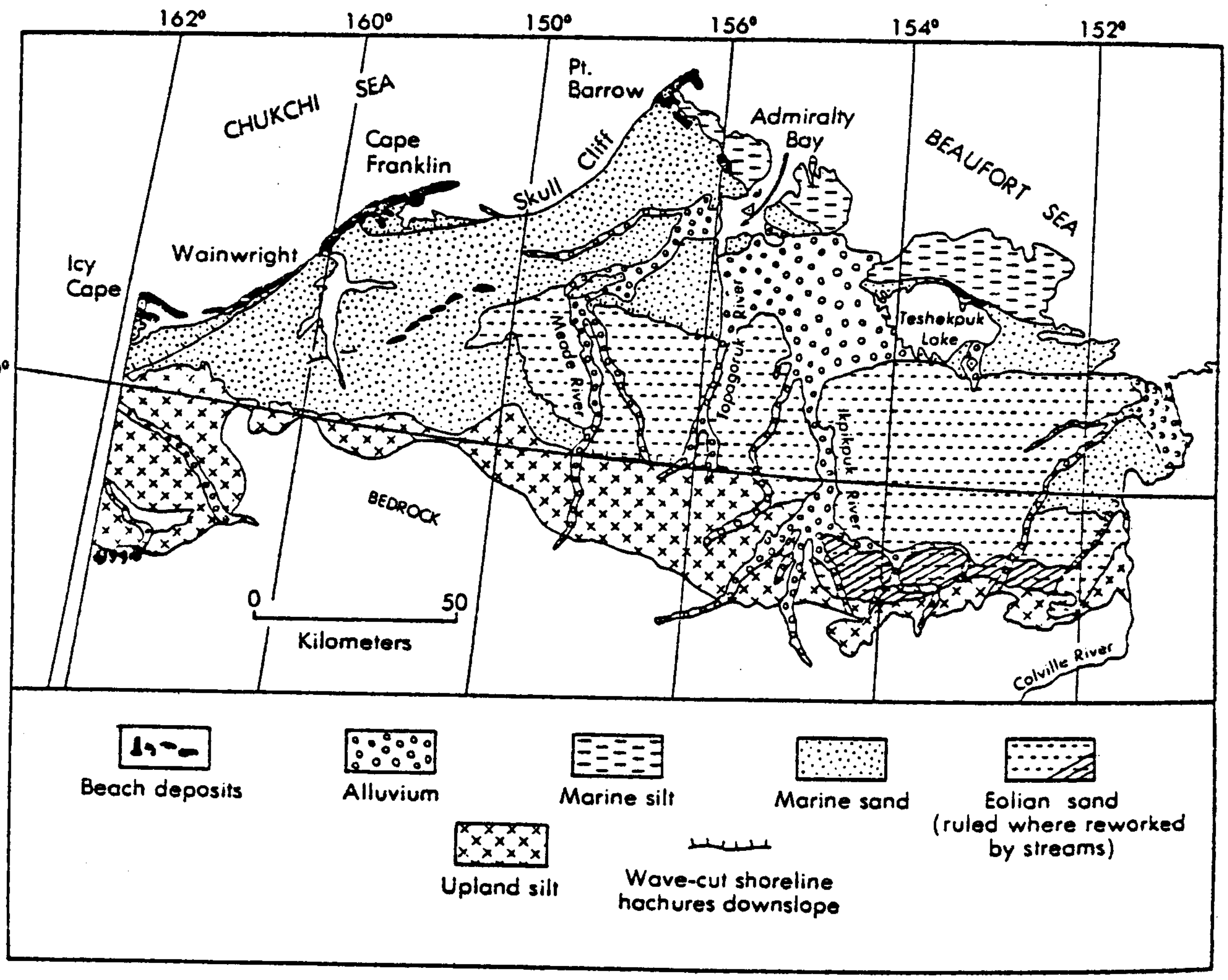
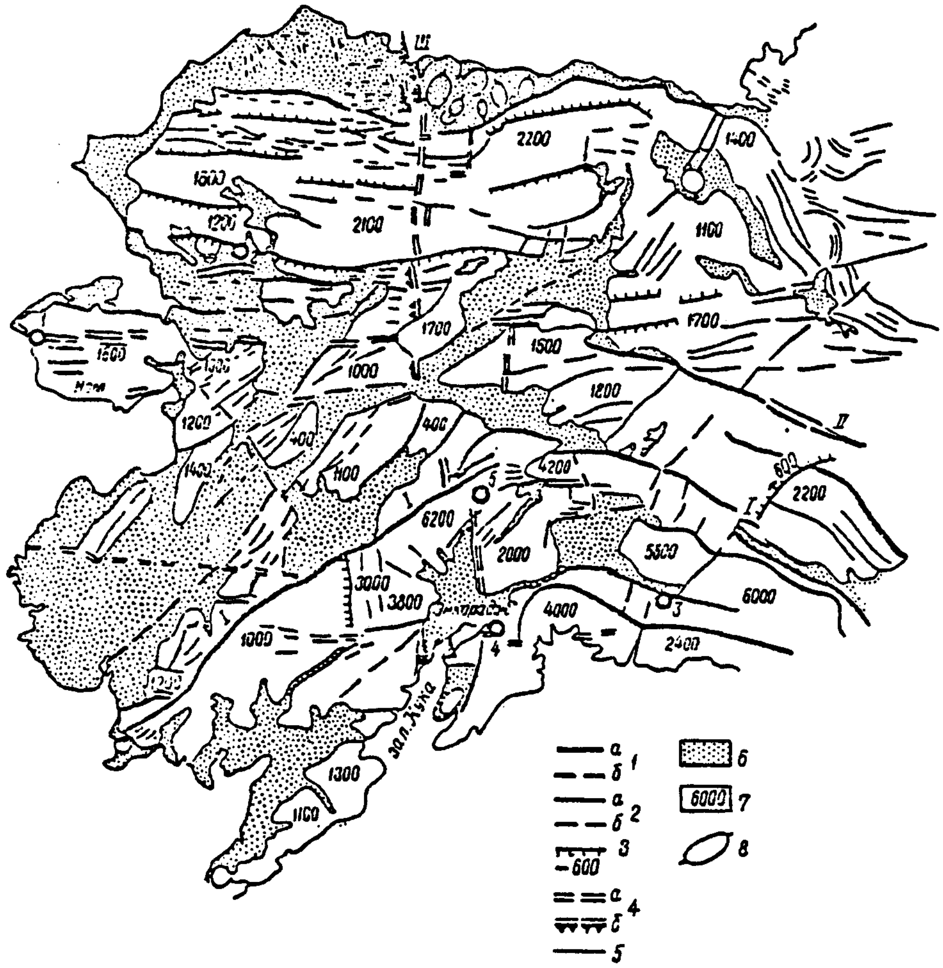


Figure 7.—Generalized surficial deposits map of Arctic coastal plain. After Williams et al. (1977).

made by I. Volochanskaya (1971) with use of satellite and topographical data, the scheme that could be used for better understanding of the nature of the differences in the recent vertical movement rate. It should be noted that these rates for northeastern and southeastern parts of Alaska are comparable (Table 2, Figure 3). Of course, the glaciostatic component of the eastern parts of Alaska close to the Laurentide Ice Sheet development increases the rates of vertical uplift because of the postglacial rebound. For the Mackenzie River basin such rebound looks appropriate.



1. Major lineaments: a) well seen, b) fragments (I–Denali system, II–Tintin fault)
2. Other lineaments: a) well seen, b) fragments
3. Topographical levels along the lineaments and their relative height
4. Transitional zones (a) and levels of regional basic heights (b) (III Anchorage–Prudhoe Bay Zone).
5. Boundaries of the morphostructures
6. Plains
7. Mountain relief and altitude of divides
8. Dome structures

Figure 8.—Linear elements in the Alaskan geomorphological structure. After Volochanskaya (1973).

We see that the major trends and differences in vertical movement rates in the western (Point Barrow–Harrison Bay), central (Colville River–142°), and eastern (142°–Mackenzie River Delta) parts of the Alaskan coastal plain (Figure 9) have a long term and genuine character explainable on the basis of plate tectonics, morphostructural specifics, and the role of glaciostatics.

Exposure of the Shelf During the Last Glaciation

If our considerations are right and the rates of vertical movement were constant during Late Glacial and Holocene times, we may evaluate the possible position of the Beaufort Sea shoreline in the past (Table 3). Then following Hopkins' curve (Interim Synthesis, 1978, p. 105) we will try to specify the areas of exposure suitable for permafrost development and then for the thermal influence of the seawater. In Figures 10 and 11, we can see that only the western part of the Beaufort shelf could have been exposed and only from 25,000 to 10,000 years ago, and isobath 60 ± 20 m is the best candidate to limit the exposed part of the shelf. Evidence of exposure, such as thermokarst and polygonal forms of relief, could be of great help to prove the exposure version.

Table 3.—Possible position of the Beaufort Sea shoreline in the past.

Thousands of Years Ago	Point Barrow–Harrison Bay (m)	Colville River–142° (m)	142°–Mackenzie Delta (m)
0	0	0	0
5	−11	−33	−51
10	−22	−66	−103
15	−33	−79	−154
20	−44	−111	−205
25	−55		

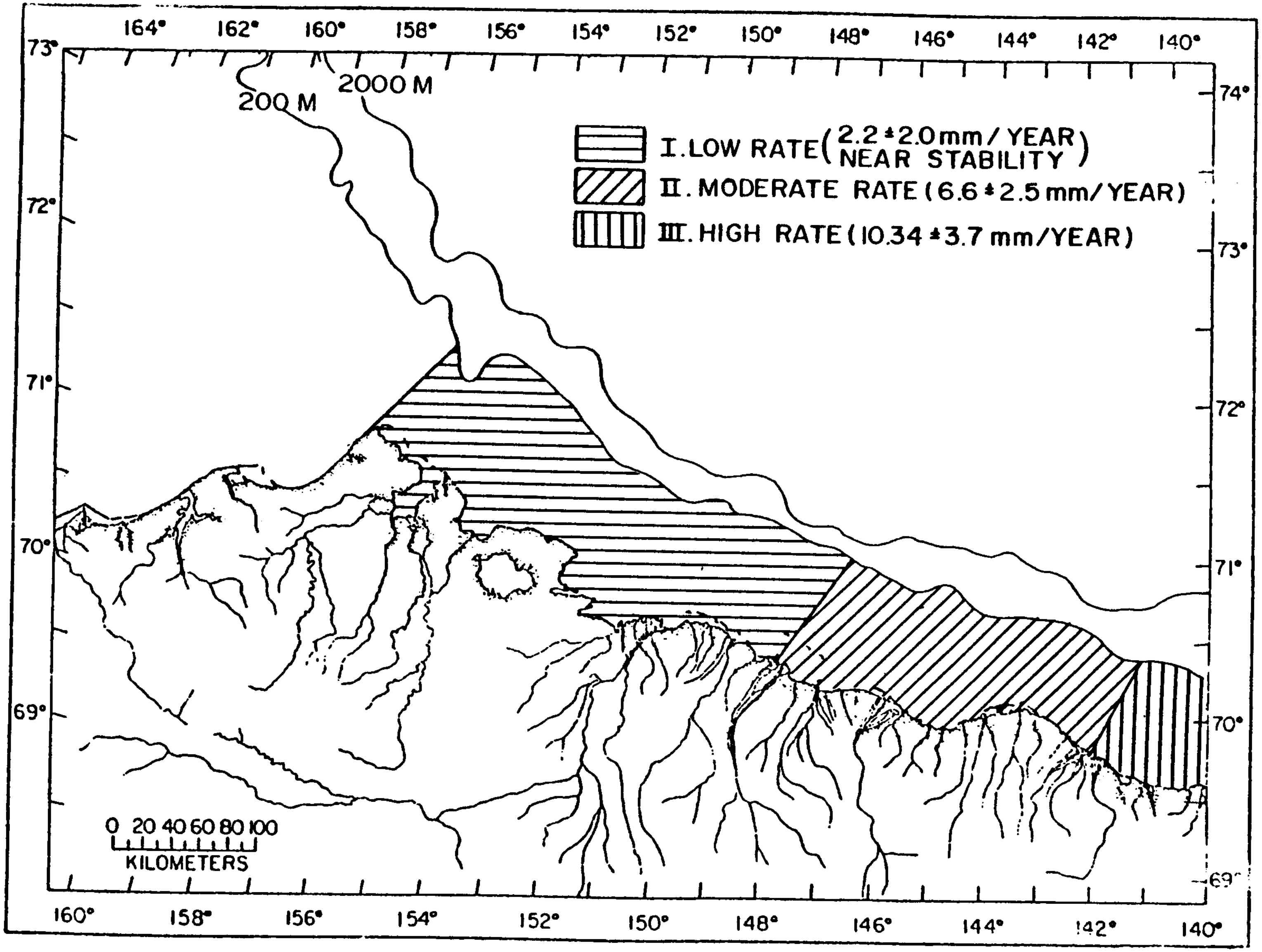
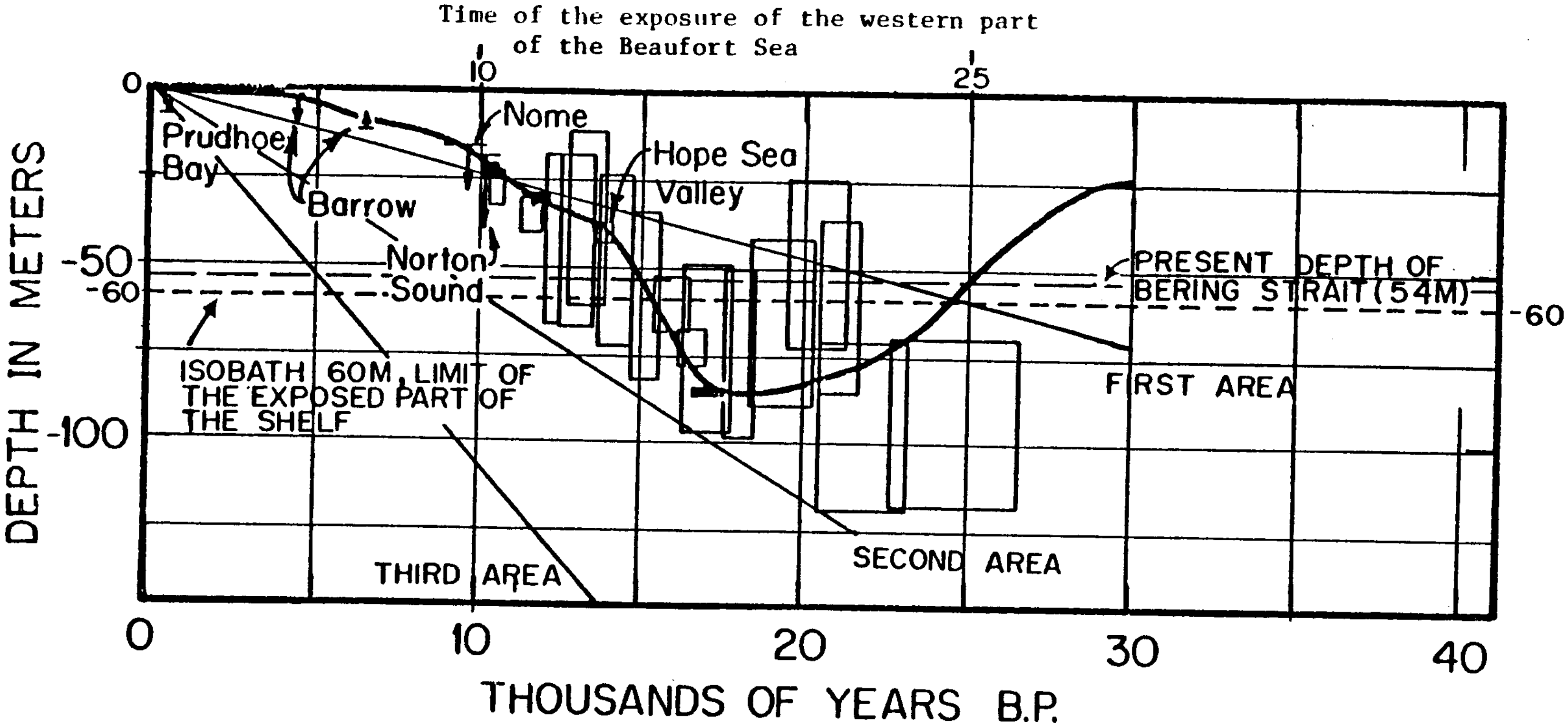


Figure 9.—Areas of the Beaufort Sea coastal plain and shelf with different rates of vertical movement.



First area: Barrow to Harrison Bay
Second area: Colville River (148°) to 142°
Third area: Mackenzie River to 142°

Figure 10a.—Reconstruction of sea level history on the continental shelves of western and northern Alaska. Possible position of the Beaufort Sea shoreline in the past, according to the Hopkins (1977) curve of the sea level reconstruction with our additions on differential recent tectonics.

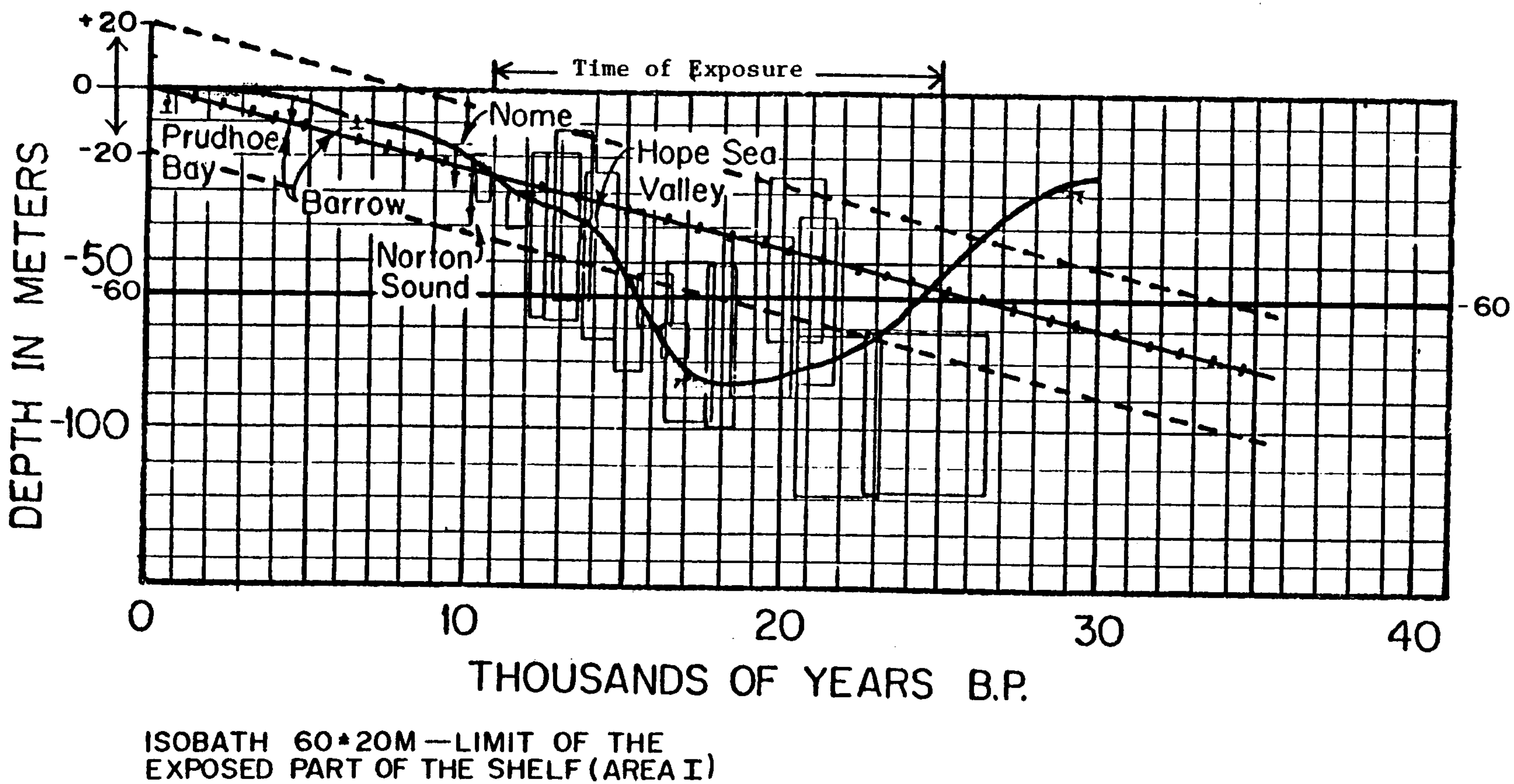


Figure 10b.—Reconstruction of sea level history on the continental shelves of western and northern Alaska. After Hopkins (1977) with our additions related to the differential recent tectonics. Possible position of the Beaufort sea shoreline in the past at the northwestern part of the shelf (first area) with correction on RMS error of the rate movement.

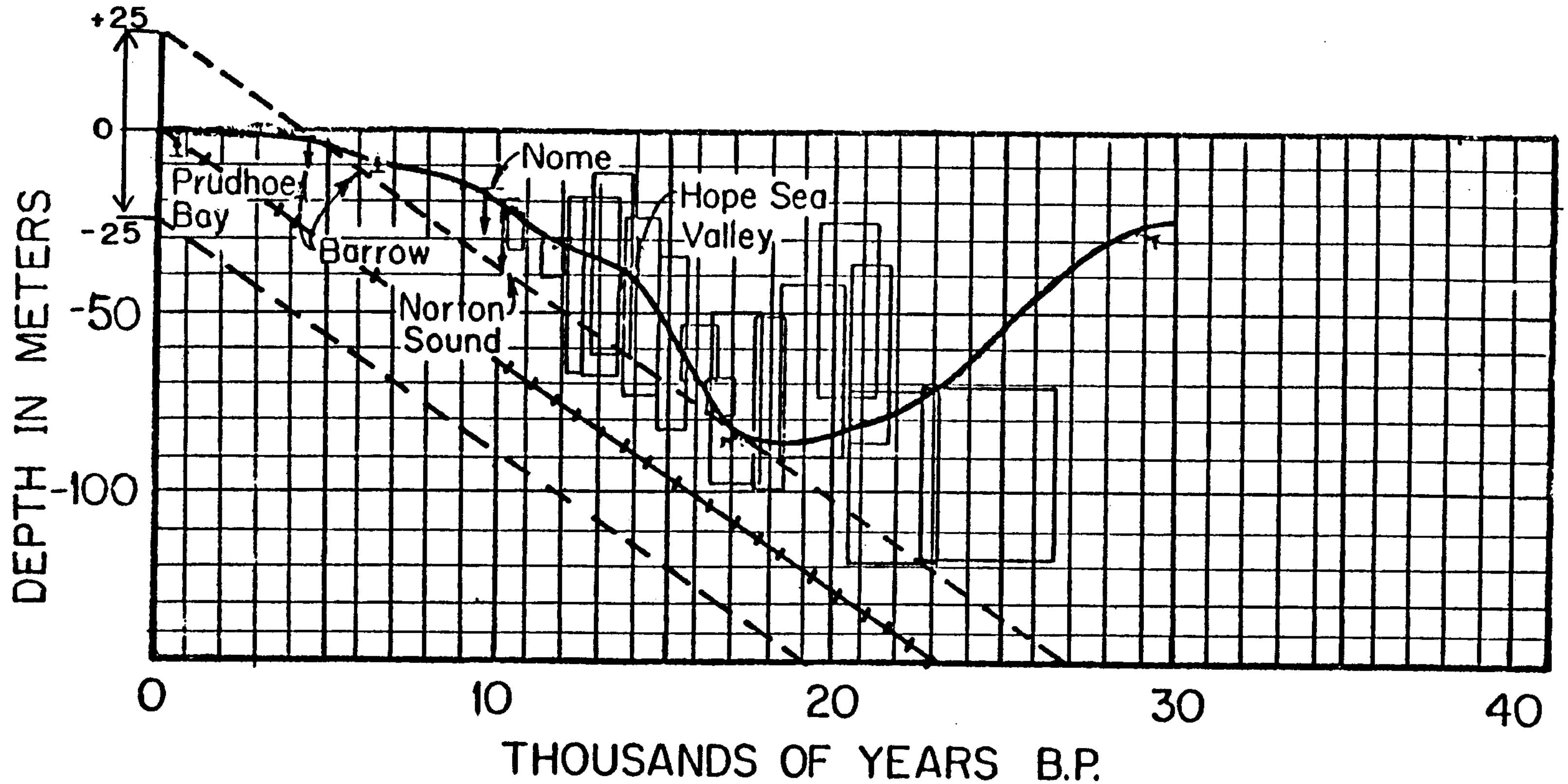


Figure 10c.—Reconstruction of sea level history on the continental shelves of western and northern Alaska. After Hopkins (1977) with our additions related to the differential recent tectonics. Possible position of the Beaufort Sea shoreline in the past at the central part of the American Beaufort Sea shelf (second area) with correction on RMS error of the rate movement.

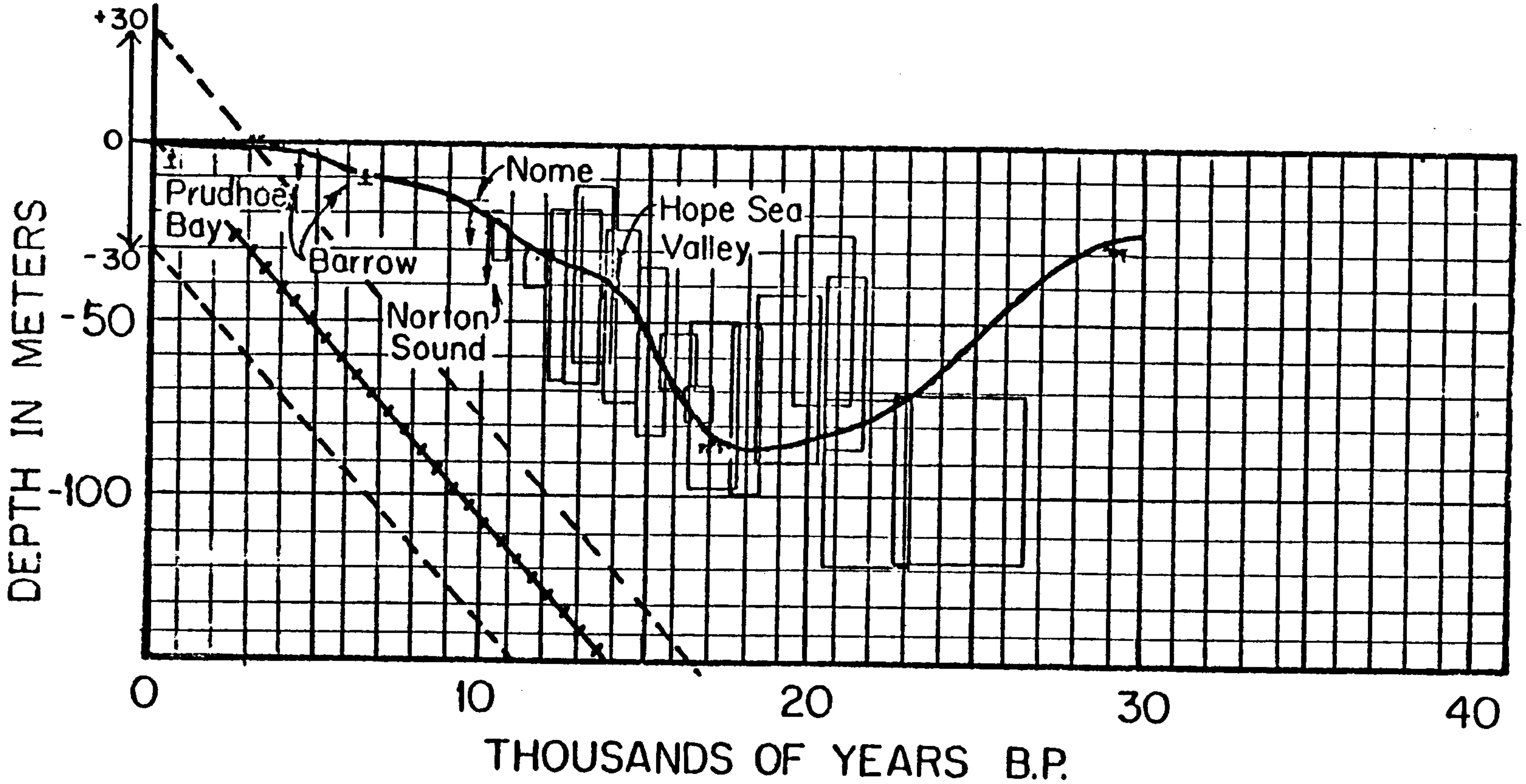


Figure 10d.—Reconstruction of sea level history on the continental shelves of western and northern Alaska. After Hopkins (1977) with our additions related to the differential recent tectonics. Possible position of the Beaufort Sea shoreline in the past at the eastern part of the American Beaufort Sea shelf (third area) with correction on RMS error of the rate movement.

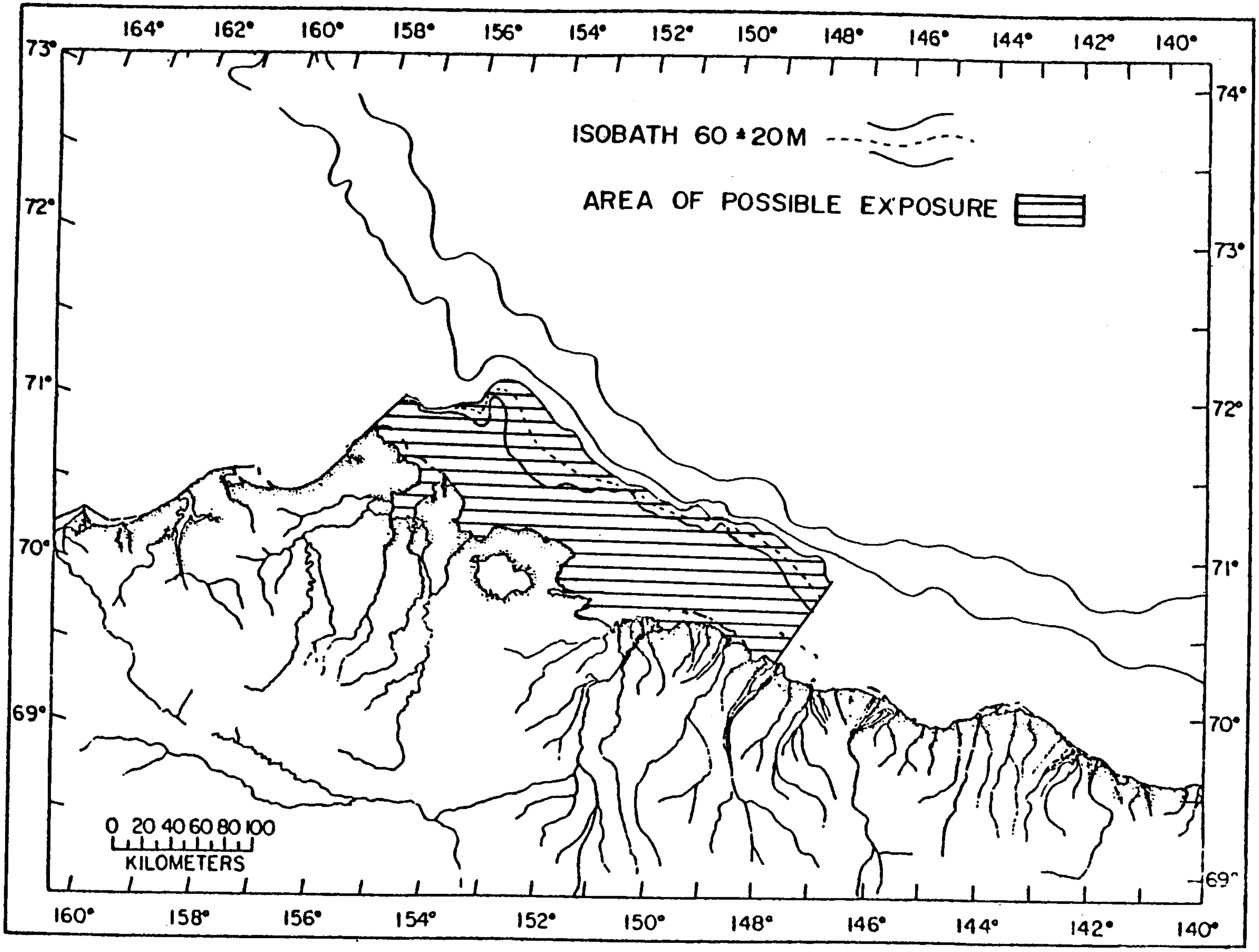


Figure 11.—Possible exposure of the shelf during the last glaciation.

Lack of Data on Thermokarst and Polygonal Forms at the Shelf

Hopkins' curve and the data on coastal retreat and morphostructural specifics today are the only base for the Beaufort Sea shelf exposure evaluation during the last 25,000 years. At the Eurasiatic part of the Arctic shelf the conclusions about the exposure of one or another area of the shelf are usually supported by such evidences as thermokarst or polygonal–tetragonal forms. Hydroacoustic investigations, for example, have been used by Klyev (1966) in the Laptev Sea for the detection of submarine thermokarst relief. Figure 12 shows the polygonal profiles and the sizes of the forms according to the acoustical diagrams. The author described wedge-like forms in the different stages of the ice thawing. The silty mud typical for this part of the Laptev Sea shelf has not filled in these forms yet. The later stages of thermokarst form development are shown in Figure 13. The space left by the thawed ice wedges are here in the process of being filled in. These forms are most typical for depths of 10–15 m where gravel and coarse sand dominate. The deeper part of the shelf is characterized by smoother microrelief of such forms. Some of the negative forms of relief, considered by Klyev as the thermokarst, are oriented along the isobaths and have a length of about 3–4 km (sometimes 7 km), an average width of about 135 m, and depth of 4–6 m. In this sense they look very similar to the ice gouging features described by Reimnitz and Barnes (1976, 1977). But such forms in the Laptev Sea are usually connected with the same kind of depressions forming polygons, tetragons, or "chess board" patterns. At some places these patterns of submarine thermokarst have a very clear expression. As a rule the subsea thermokarst forms are comparable with their analogs in the adjacent coastal areas. At some areas of the Eurasiatic shelf exposed during the late Pleistocene the thermokarst forms were found at a depth of 35–40 m. Often these formations have been found near areas of river-water influence. This shows the continuation of the submarine permafrost and its forms degradation in present time. It is possible that some of the depressions described on the Eurasiatic shelf are not thermokarst but are of ice gouging origin. As for the Beaufort Sea shelf, all the depressions of similar size and position are described here as the result of ice gouging processes (Reimnitz and Barnes 1976, 1977). The possible future finding of thermokarst and polygonal–tetragonal forms in the western part of the shelf (to isobath 60 ± 20 m) could give direct evidence for a shelf exposure. The sizes and distribution of these forms have to correspond to the parameters of their analogs on the western apart of the Beaufort coastal plain. If such direct evidence were found, it would be a great support for the shelf exposure concept related to the origin of the submarine permafrost now considered as a relic.

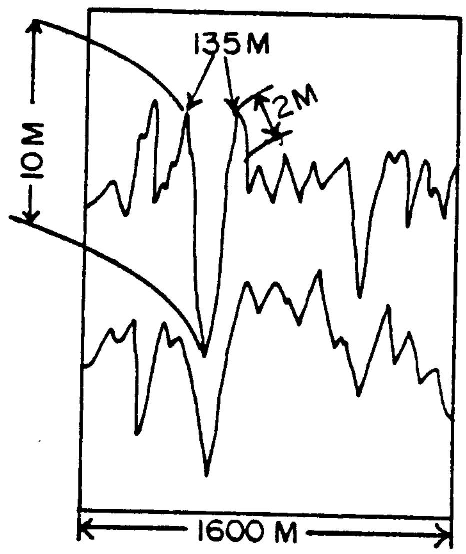


Figure 12.—Polygonal profile with the ice wedge in the process of thawing at the Arctic shelf, Laptev Sea. After Klyev (1964).

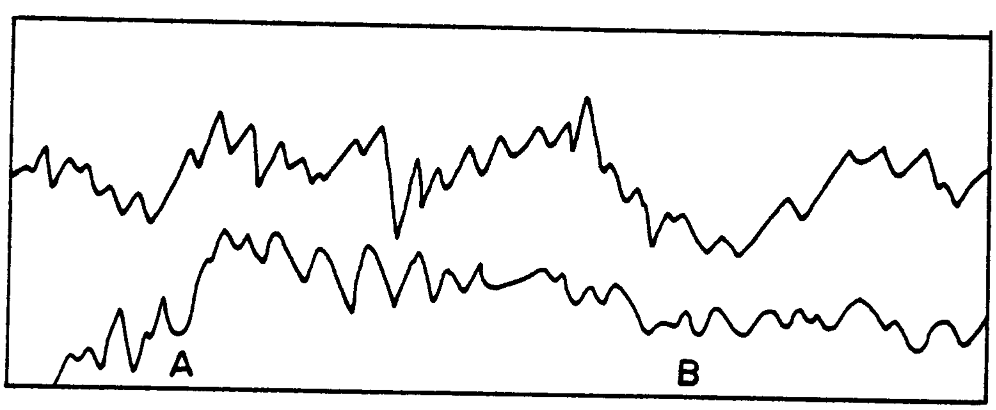
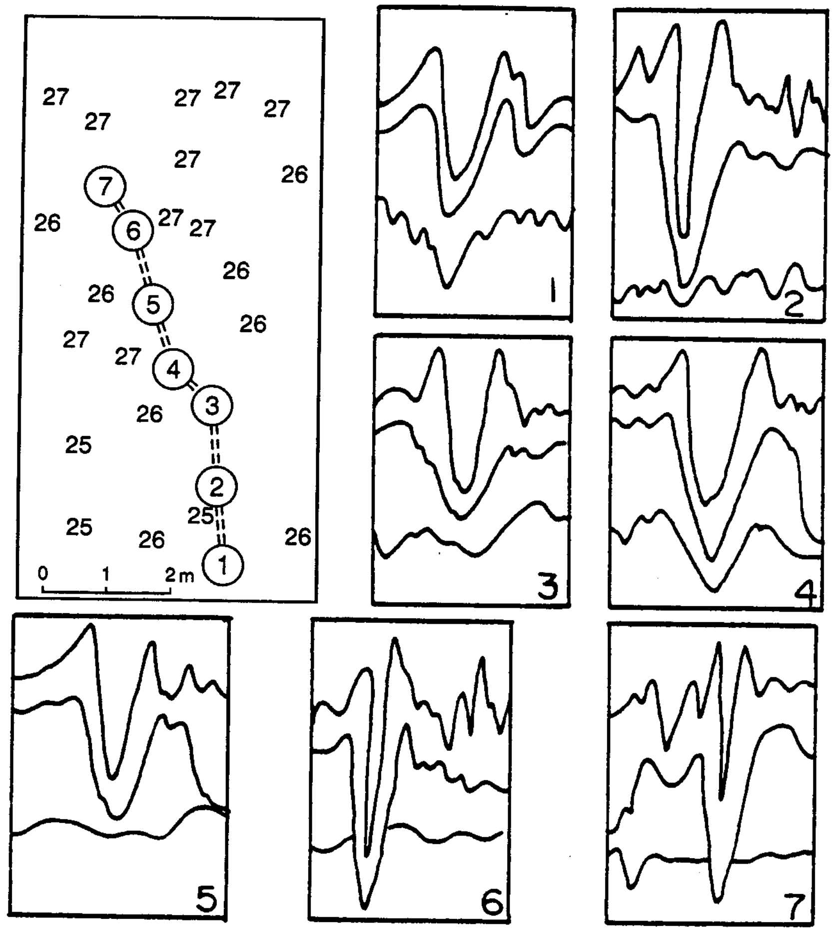


Figure 13.—Profiles of the thermokarst forms on the shelf after thawing, according to Klyev (1964).



Vertical scale for 1,3,5 - 1mm = 1m
for 2,6,7 - 5mm = 1m

Horizontal scale for 1,3,5 - 1mm = 12m
for 2,6,7 - 1mm = 23m

Figure 14.—Submarine thermokarst profiles on the shelf. After Klyev (1964).

Role of the Meridional Lineaments of the North Alaskan Coastal Plains and Shelf

Short and Wright (1974) have identified two sets of lineaments on the Alaskan North Slope by inspection of map contour patterns and surficial geomorphic trends. They have noted that primary and secondary lineaments, with 40° and 300° azimuths, have an effect on the coastal configuration and other physiographic features. Two systems of the lineaments of Short and Wright (Figure 15) fully correspond to the tangential stress orientation along the major lineament systems of the earth (Figures 16 and 17). These lineaments are a part of the global system and they do not specify some important regional morphostructural peculiarities of the Alaska Peninsula, including the northern slope. Usually, a comparison of maps which portray deep-seated faults with topographical maps of the corresponding areas gives an indication that the directions of the dominating faults are reflected in the general pattern of the drainage system (Kudryavtsev 1963; Chebanenko 1963). Close relationships between tectonic lines and the drainage patterns were noted on the Siberian platform (Vakat et al. 1958). The trend of the major faults is shown in the bends of the Lena River. Such Siberian rivers as Vilyui, Augara, and Aldan generally flow along the trend, which coincides with the primary structural pattern of the corresponding region. A rose diagram of lineaments from various areas in the different parts of the earth (Figure 17) gives some kind of summary in lineaments distribution and orientation made by Voronov et al. (1970). The same authors give the general picture of drainage patterns and direction of major lineaments in the Bering Strait area (Figure 18). This picture shows not only "diagonal" lineaments but also meridional structural features, the major of which is the transitional highly seismic zone, Cook Inlet–Prudhoe Bay. We tried to specify these meridional lineaments according to the direction of several rivers on the Beaufort Sea coastal plain (Figure 19). It seems that these lineaments are the continuation and reflection of the same features that were discovered by Eittreim and Grantz (1977) as the "sea valleys" at the upper slope of the Beaufort–Chukchi shelf (Figure 20). In Part II of this work we have described the phenomena of the "ancient valleys." Their development is connected with a sharp ocean level drop at the time of Brunhes-Mathuyama paleomagnetic changes. They look usually as the canyons in the sedimentary rocks of the coastal plains and shelves, filled in mostly by silty sand and gravel, frozen in Arctic basin, and could represent a big danger in the case of the disturbance of the thermal regime during any operational activity on the shelf. At the same time these valleys could be traced easily by the different geophysical methods. The meridional lineaments of the Beaufort Sea shelf as the special combination of the tectonic structures with the "ancient" eronsional forms, filled in with Quaternary deposits, or partly reflected in the relief (half buried), or at the sites of crossing with the "diagonal lineaments" could be responsible for anomalies in the parameters of the bottom deposits and seawater (temperature, salinity, etc.). These sites are convenient for discharge of the deep ground water

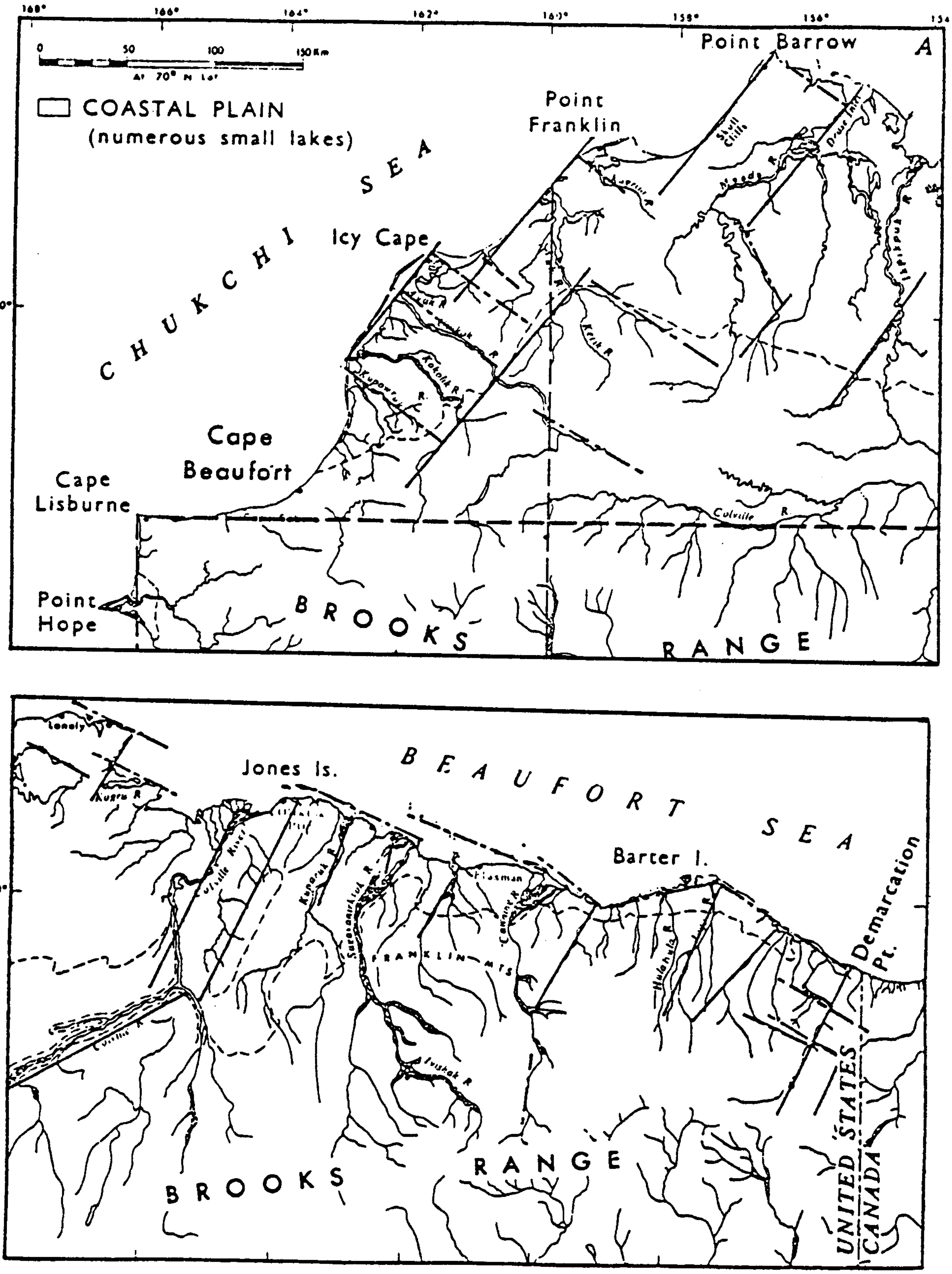


Figure 15.—Location of some of the most prominent lineaments in the Alaskan Arctic. After Short and Wright (1974).

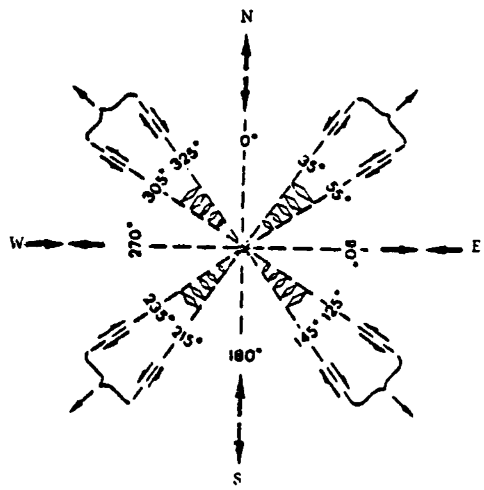


Figure 16.—Scheme of the tangential stress orientation along the major lineament system of the earth. After Voronov et al. (1970).

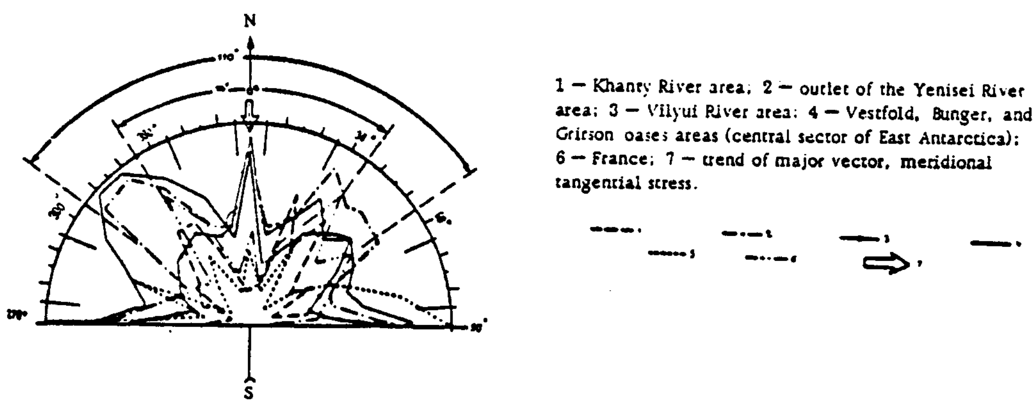


Figure 17.—Rose diagram of lineaments from various areas. After Voronov et al. (1970).

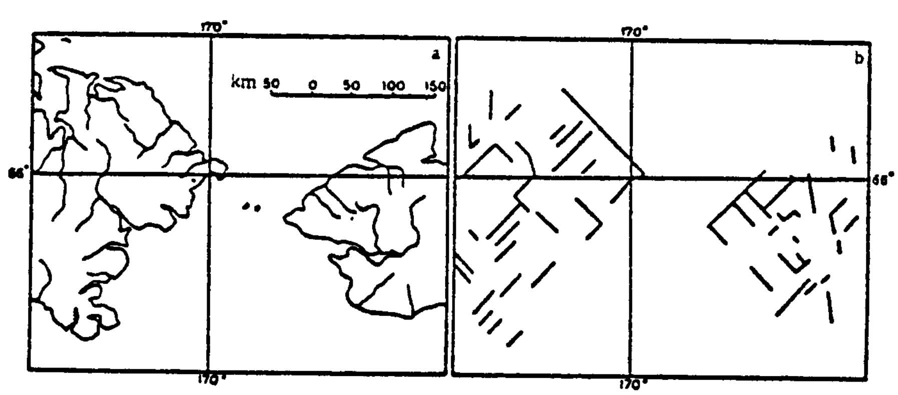


Figure 18.—Direction of drainage patterns (a) and major lineaments (b) in the Bering Strait area. After Voronov et al. (1970).

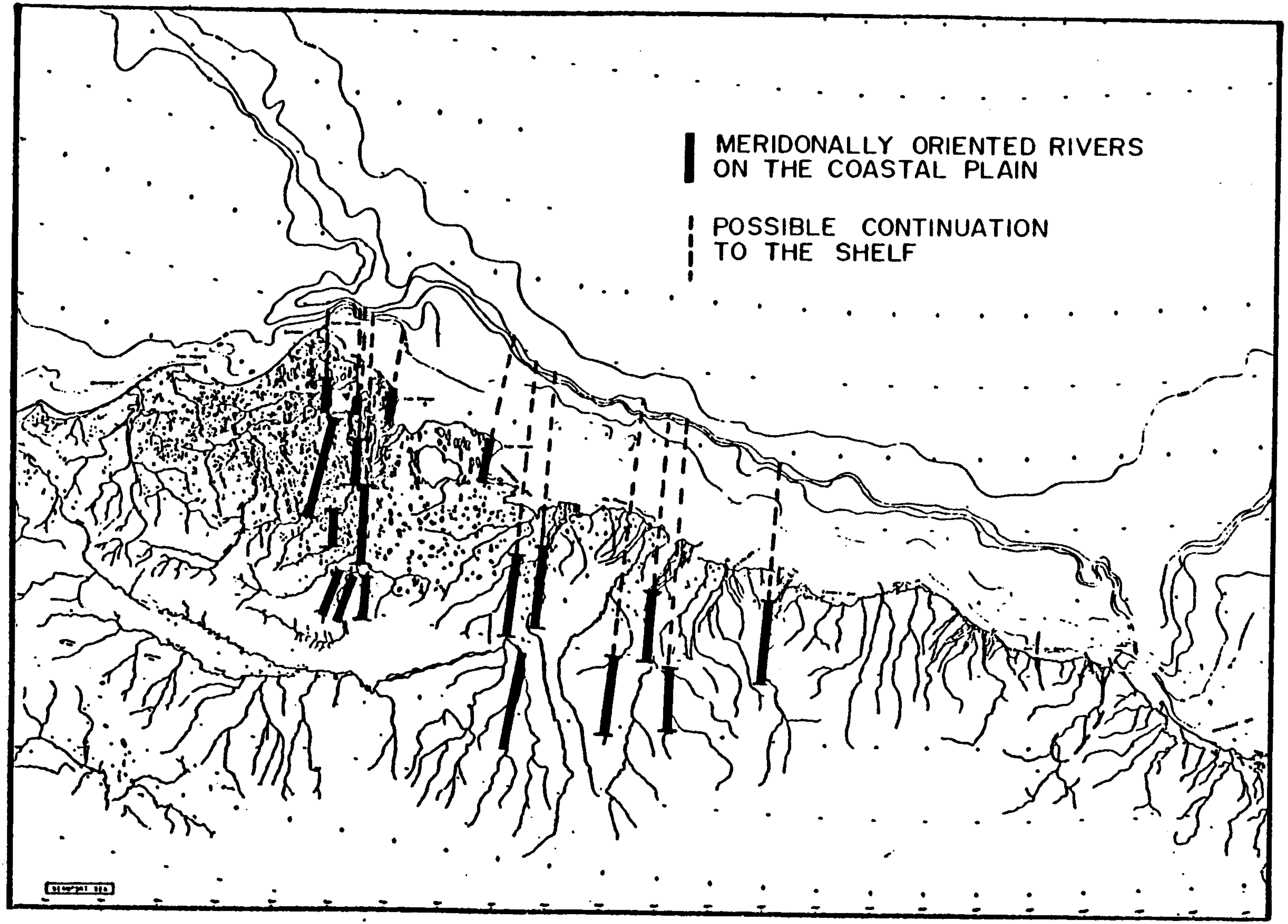
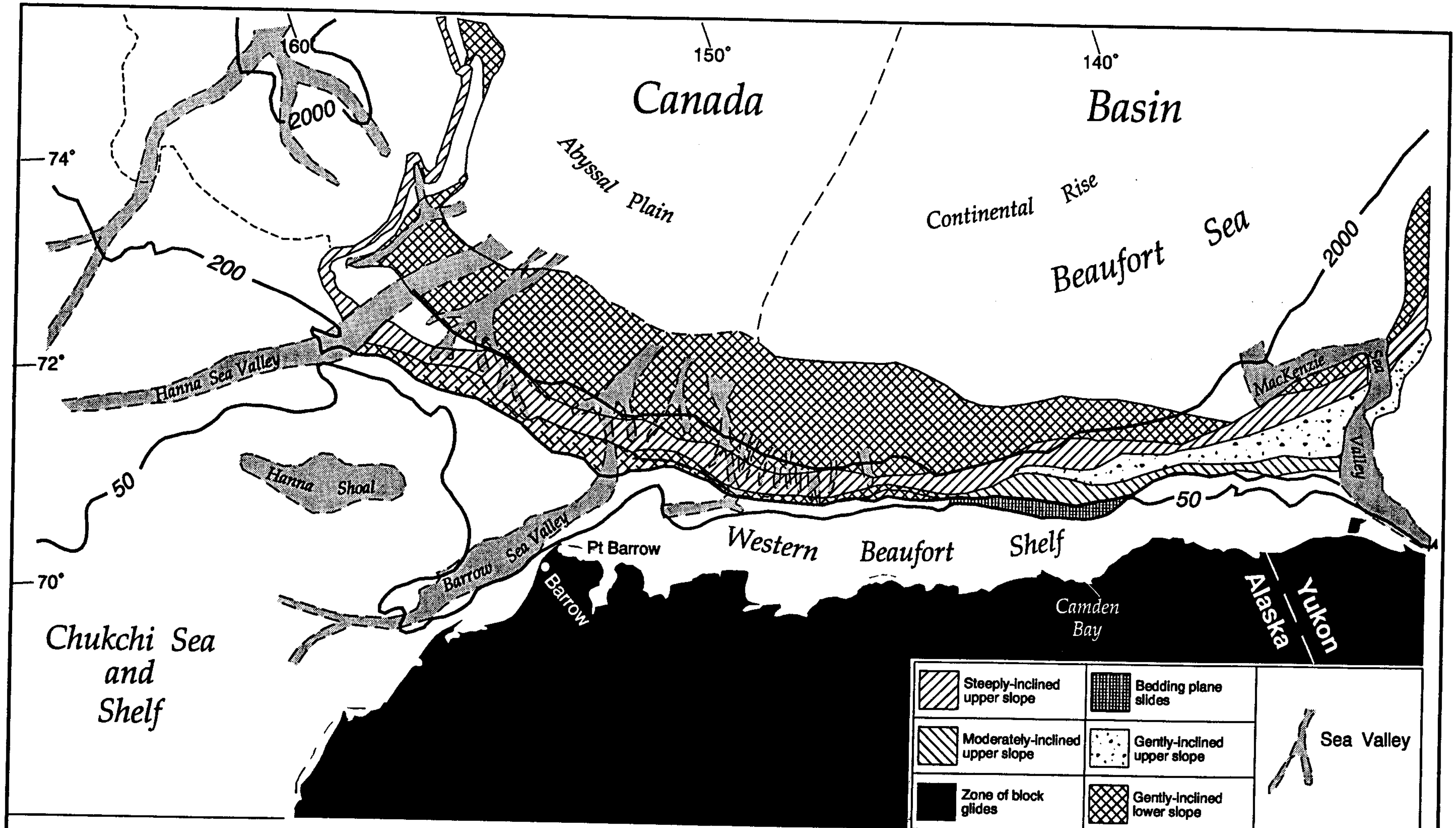


Figure 19.—Major meridional lineaments of the Beaufort Sea coastal plain (rivers) and their possible continuation on the shelf.



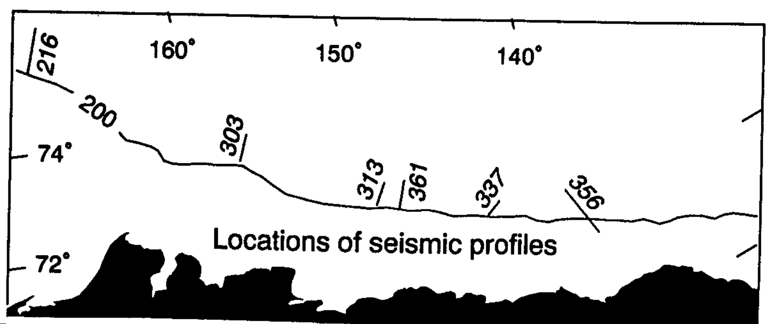


Figure 20.—Physiographic provinces and location of submarine terrains susceptible to sliding. Western Beaufort Sea after Eittreim and Grantz (1977).

with different ranges of temperature from warm to the "Criopegi" type (saline, cold, supercooled water; see Part II).

Possible Thickness of the Relic Submarine Permafrost

On the north coast of Alaska, the geothermal heat flux is equal to 0.050 Kcal/(hr m^2) (Gold, Lachenbruch 1973). Consideration of modern concepts about geological structure and the developmental history of the western part of the Beaufort Sea shelf shows that the value of the flux of geothermal heat within limits of the shelf is probably similar to the value on the coast; i.e., if is approximately 0.050 Kcal/(hr m^2) with possible deviations of at least ±14% (Are 1978). We saw that the development of the western part of the Beaufort shelf in marine conditions took place during the last 10,000 years (in the limits of isobath 60 ± 20 m). This means that thawing of the shelf deposits from below because the thermal heat flux continued here about 10,000 years. Following the simple formula used by Chekovsky (1972) for the calculation of the submarine permafrost thickness in the Kara Sea, we tried to use this approach for the western Beaufort Sea subsea permafrost. The formula:

$$G_t \lambda_t = \frac{Q_p H}{T}$$, where

G_t – Geothermal gradient of the thawing zone, in our case 0.030°C/m

λ – Coefficient of the thermoconductivity in the thawing zone, 1.1 (Kcal/M hour°C)

Q_p – Phase change heat, 24,000 (Kcal/M^3)

T – Thawing time, 10,000yr

H – The value of thawing zone (from below) thickness (m)

The solution is: $$H = \frac{G_t \lambda + T}{Q_p}$$

The figures of the thawing zone from below thickness depend on the changes in the value of the geothermal gradient here. They could change in the limits of 130–175 m. Now according to these data, we can decrease the figures of the present coastal permafrost thickness at the northern Chukchi and western part of the Beaufort seas that are approximately 400 m (near Barrow), 300 m (near Cape Simpson), 350 m (near Cape Thompson). It means that the average figure of the western Beaufort Sea subsea permafrost would be: 350 m – 150 m = 200 m (roughly), but these figures characterize the possible submarine permafrost thickness only with the decrease related to thawing from below. We need also to decrease the last figure because of thawing from above. In the works of Osterkamp and Harrison (1976) and Lachenbruch and Marshall (1977) the several possibilities have been considered. Two of them are the most important: first, when the mean seabed temperature is greater than melting at the top of the permafrost; and second, when the mean seabed temperature is less than melting at the top of the permafrost. Today we cannot answer which part of the Beaufort or Chukchi seas could be related to the first or second cases, or for how long. During sea level changes in Late and Post-glacial times, seawater currents, their direction and thermal regime in the lower layers, as the factors influenced the temperature of the seabed could be changed often and drastically, especially in the sense of Bering and Chukchi water exchange. In our opinion, the parameters of the Bering Strait—its exclusively shallow depth and small width, high recent tectonic activity, and potential to be dammed by ice (Péwé 1976)—today give no clue for reconstructions that could help in calculating the decrease of the subsea permafrost from above.

According to some very approximate data from the different sites of study in the Eurasiatic and American shelves of the Arctic the decrease of the subsea permafrost thickness from above during 10,000 years might reach about 80–100 m. In our also approximate calculations the thickness of the relic permafrost in the western Beaufort Sea consequently could be about 100–120 m and permafrost might be met at the limits of the first 50–100 m from the seabed (in the limit of the water depth—60 ± 20 m).

At the second, or the central, area of the Beaufort Sea a body of deep permafrost could be only a continuation of a thick (about 600 m) coastal cryogenic zone and limited to 18–22 km offshore. General thickness would be about 150–200 m; seaward this is less, 50–30 m. In the third, or eastern, part of the sea the continuation of coastal permafrost could not be extended farther than 2–5 km offshore. The depth of the permafrost here could be 50–100 m. There is not enough data on coastal permafrost thickness here to give an idea of the thickness of the permafrost offshore. According to comparable data from the Eurasiatic shelf it could be 2–3 times less than coastal permafrost here. Seasonally frozen layers might be met on any part of the Beaufort Sea and Chukchi Sea coasts within the limits of the sea ice–sea bottom interaction (2-m isobath). In the Chukchi Sea the permafrost could exist only extremely close to the coast (usually less than 1.5–2.0 km) and might be connected with the coastal permafrost body or disconnected (lenses). In both cases the thickness of the submarine permafrost could reach nearly 20–30 m.

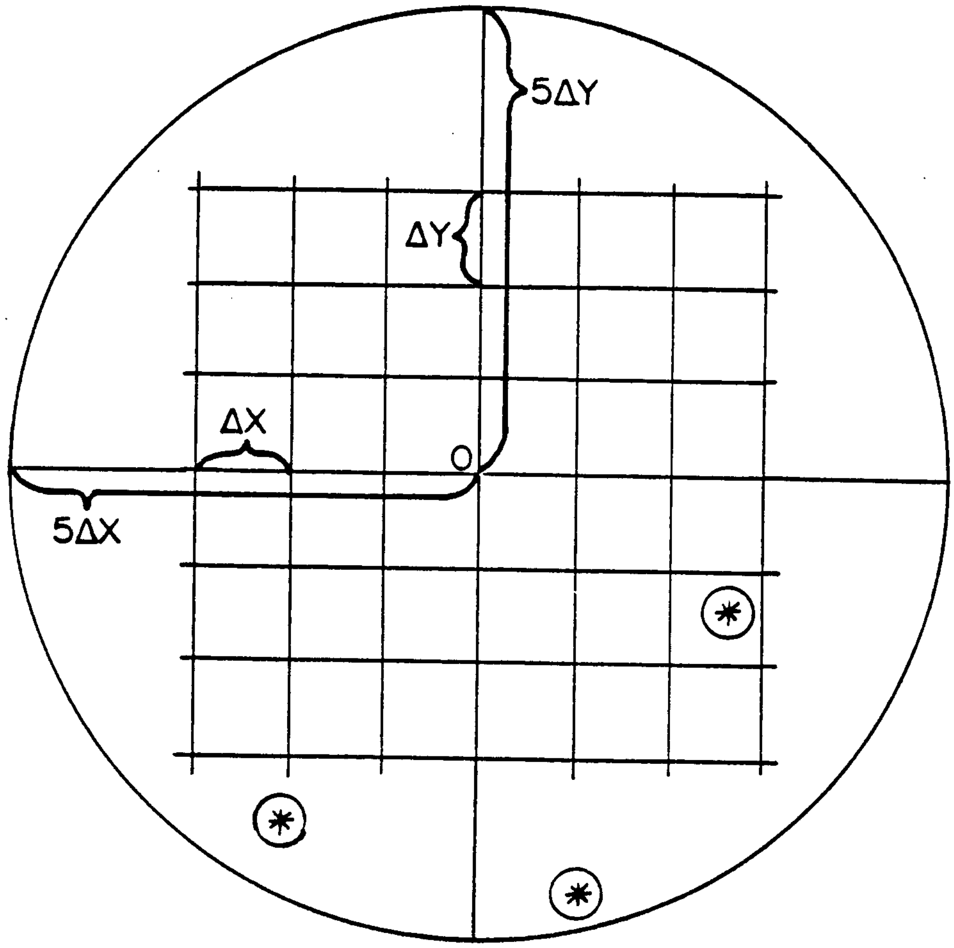
ENVIRONMENTAL DATA ON ALASKAN SUBMARINE PERMAFROST DISTRIBUTION

The available geologic, oceanographic, topographic and glaciologic materials have been included in our system in an attempt to find the most suitable combination of interdisciplinary data for submarine permafrost prediction. The evaluation of all these data has been done in terms of probability.

Data and Their Reliability; Defining the Gaps

As was mentioned above, the direct data on submarine permafrost (drilling) in the Beaufort and Chukchi seas are restricted to a very few sites and in variable environmental conditions of the shelf they cannot characterize any significant part of it. The seismic methods for defining the upper surface of the subsea permafrost are exclusively prospective, but according to the latest publications, some of the methods do not give the necessary answers. According to Reimnitz et al., "High resolution seismic reflection records in the Arctic today have given no clue on the depth to ice-bonded sediments" and "careful analysis of the seismic records provides no clues on the distribution of ice-bonded sediments" (Miscellaneous hydrologic and geologic observations in the inner Beaufort Sea shelf, Alaska, 1977, p. G–10). The only geological information available to use for our purposes is the bottom sediment grain size data. The topographical picture of the shelf is sufficient. The bathymetrical data provide knowledge of the bottom depth for any area of the Beaufort and Chukchi seas out to several hundreds of kilometers. The glaciological data, gathered mostly during the OCS program research, also sufficiently describe the processes of the sea ice–seabed interaction close to the coast and the barrier islands. The role of the sea ice as a cooling agent of the sea bottom in the stamukhi zone (Barnes and Reimnitz 1976, 1977) is not clear. Oceanographic data on temperature and salinity of the seawater close to the bottom in the shelf limits had been gathered during different years, different periods of the summer, and at different times of the day. This explains the necessity of designation of statistical reliability on source and derived products. This point was emphasized in the letter of D. A. Wolfe of July 17, 1978. The importance of the documentation of the statistical reliability of the data presented is obvious. Interval contouring of source and derived maps is based on multiple data points non-uniformly distributed over a broad geographic area. Because of this the confidence intervals vary in different parts of the field. The measure of the reliability or confidence level needs to be associated with the probabilities predicted and to be incorporated directly onto the product.

The reliability of contouring is connected with the distance between the grid points, or the size of the grid-cell chosen. The size of the cells is determined by several factors: the overall goals of the study, the character of the data, the size of the study area, the scale of mapping, and of course, the computer efficiency resources. Since the main objective of the study is to specify the lease areas with the different probability of subsea permafrost, special consideration is given to the elementary amount of land required for leasing. A grid-cell size of about 4 square miles, which approximated the size of an individual lease, was finally selected. At latitude 71° north the distance 1.8 miles corresponds to 5′. The Beaufort Sea area which is under consideration lies between 141°–157° west longitude and 69°30′–73°30′ north latitude. We need enough space in the computer memory for at least two different fields. For example,



O - Point for which we want to find the reliability

ΔY–5′ - Vertical distance between two nearest grid points

ΔX - Horizontal distance between two points

(*) - Points of observation

Figure 21.—Illustration for the reliability of data evaluation.

to compute the supercooling of the seawater close to the bottom, two parameters are used at the same time: seawater temperature and salinity. Coordinates of the shoreline are included in the computation too. The smallest step in longitudinal and latitudinal directions, which still allows only use of the main memory, is the same 4 square miles. This means a total of 9,457 words of computer memory.

Reliability of our maps at grid point is

$$R = 100\,(1 - \zeta/8) \text{ in percents}$$

where $0 \leq \zeta \leq \delta$ is the distance between grid point and the nearest observational point and $\delta = 5 \max(\Delta x, \Delta y)$. Δx is horizontal distance between 2 adjacent grid points and Δy is vertical distance between grid points. We chose Δx and Δy to equal approximately 2 miles. This was done by the following procedure:

$$\Delta y = a\Delta\varphi,\ \Delta x - a\Delta\lambda \cos\varphi, \text{ where}$$

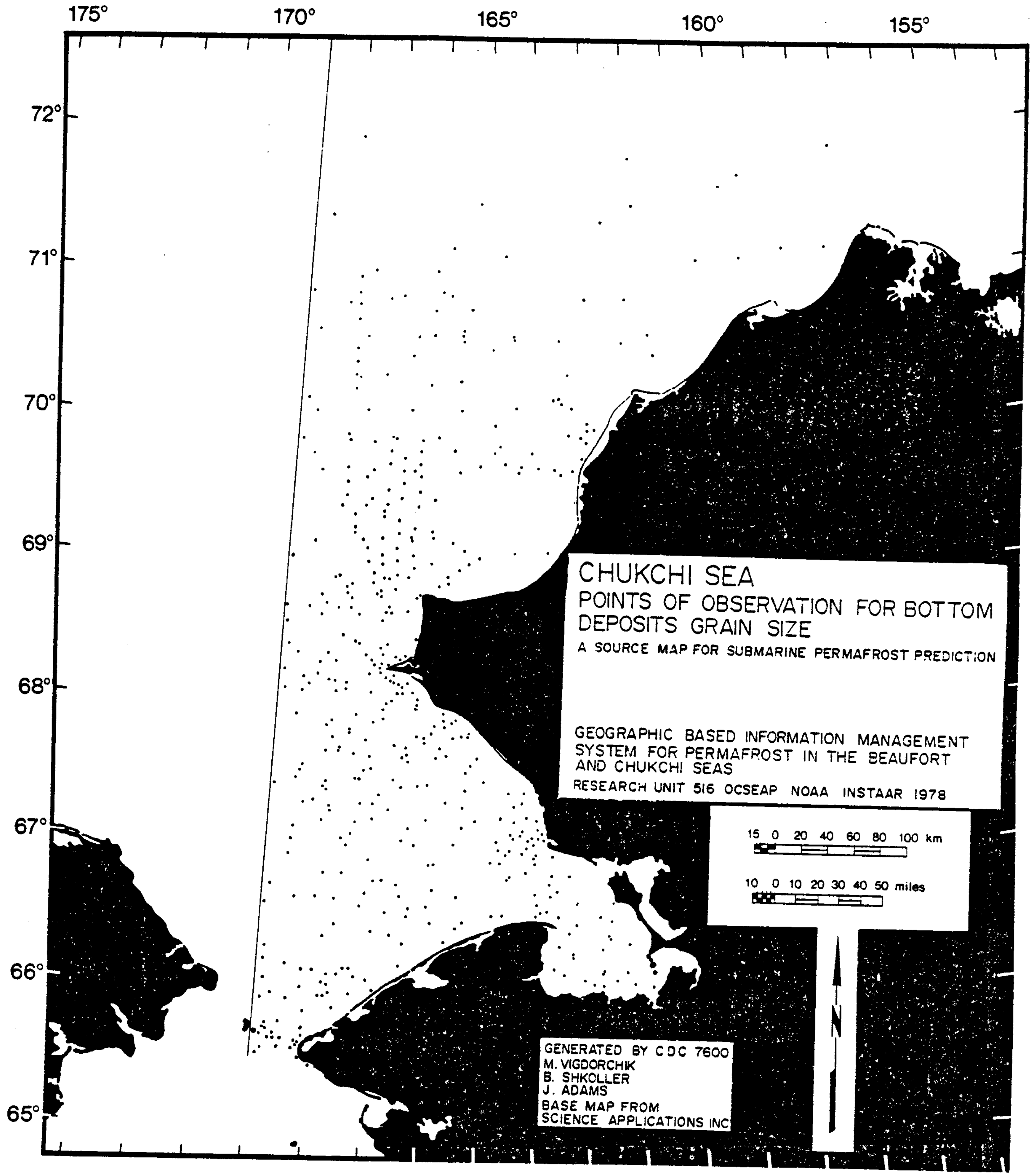
a (radius of earth) = 6,371 km

$\Delta\varphi$ (latitudinal distance between 2 grid points) = $(1/35)^\circ$

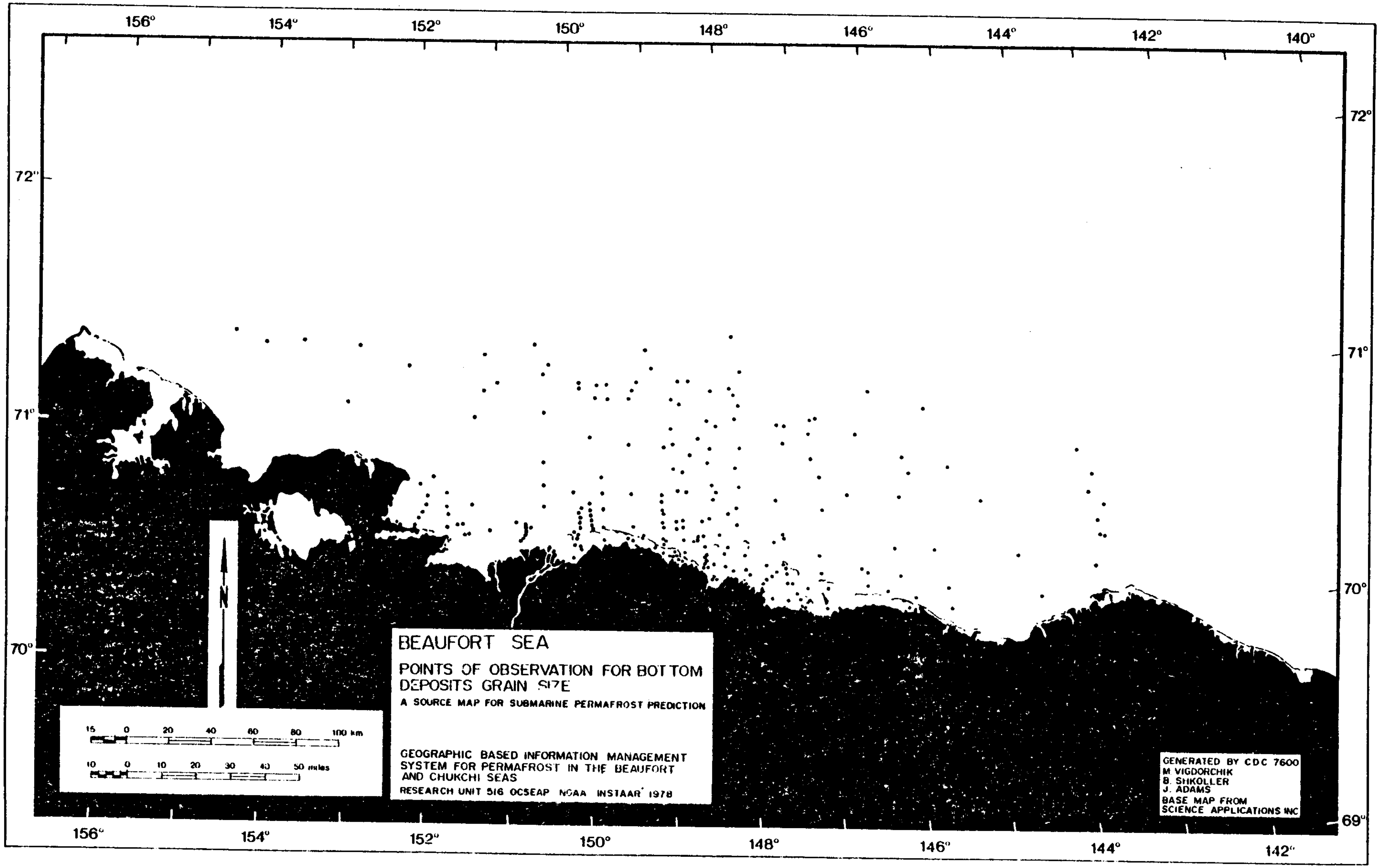
$\Delta\lambda$ (longitudinal distance between 2 grid points) = $1/12)^\circ$

Then $\Delta y = [a \cdot (1/35) \cdot \Pi/180 \cdot 1/16]$ miles $= 1.99 \approx 2$ miles, and Δx (at latitude 70°) = a $(1/12) \cdot \Pi/180 \cdot \cos(70^\circ) \cdot 1/16 = 1.98 \approx 2$ miles ($\Delta x = 1.92$ at latitude 71°).

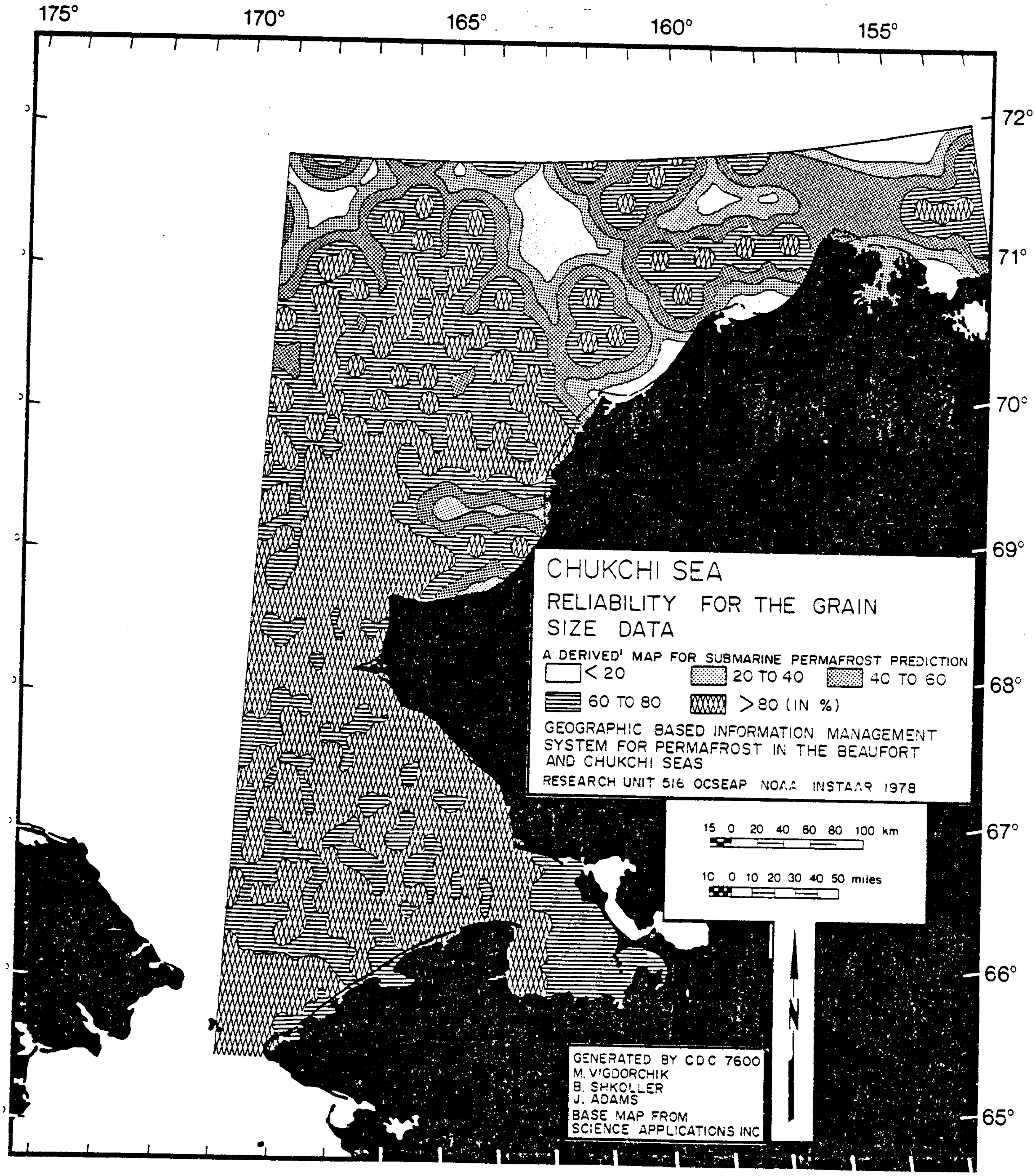
We distinguish two levels of data reliability, one for each parameter separately, another to characterize the relative density of all data used in the different parts of the area. Maps of the "Points of observation for bottom deposits grain size" in the Chukchi (Map 1) and Beaufort (Map 2) seas and maps of "reliability for the grain size data for these seas (Map 3 and 4) clearly show that the data distribution is irregular and sparse; some of the areas have no observations at all. This means that any statistical probability of the submarine permafrost prediction needs to be reduced according to the relative density of the basic data. It means also that to make contours with the computer process we have to use interpolation. The areas with a high reliability for grain size data concentration are situated in the southern and central parts of the Chukchi Sea and along the coast of the Beaufort Sea (70°–71° N, 144°–155° W). There are gaps in the observations to the north from 71° N in both seas and to the east from 145° W. Some of the areas with reliability of data less than 20% are situated along the Chukchi Sea coast between Cape Lisburne and Point Barrow and in the Beaufort Sea between this point and Smith Bay. The level of reliability for temperature and salinity according to the density of the observation network (Maps 5 and 6) is also variable but mostly higher than 60%, both in the Chukchi Sea and in the Beaufort Sea in the area close to Point Barrow and between Harrison Bay–Camden Bay at the south, and 71°–71°30′ at the north (Maps 7 and 8). "The space distribution" aspect of the data reliability for sea temperature and salinity is not the



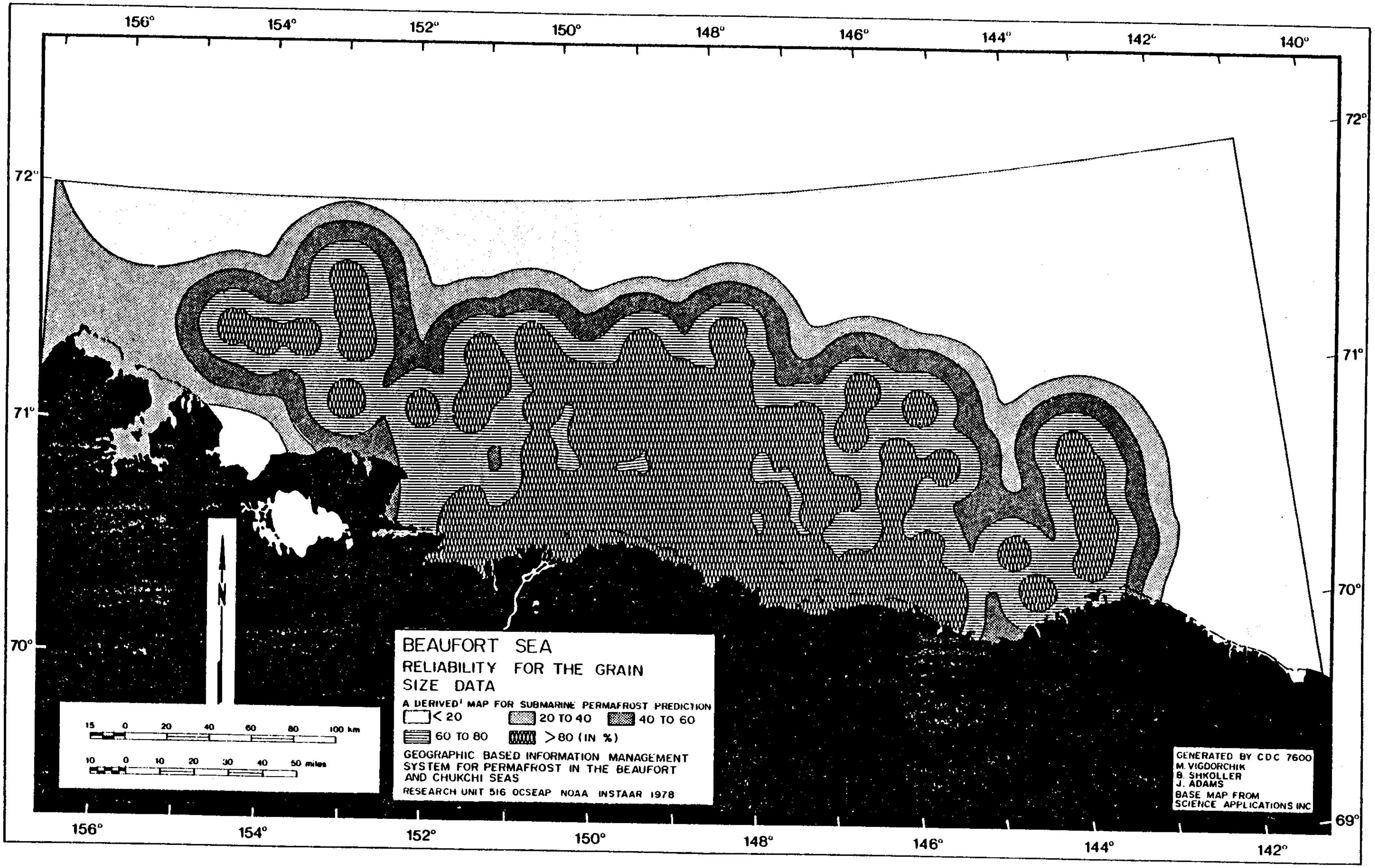
Map 1.—Points of observation for bottom deposit grain size, Chukchi Sea.



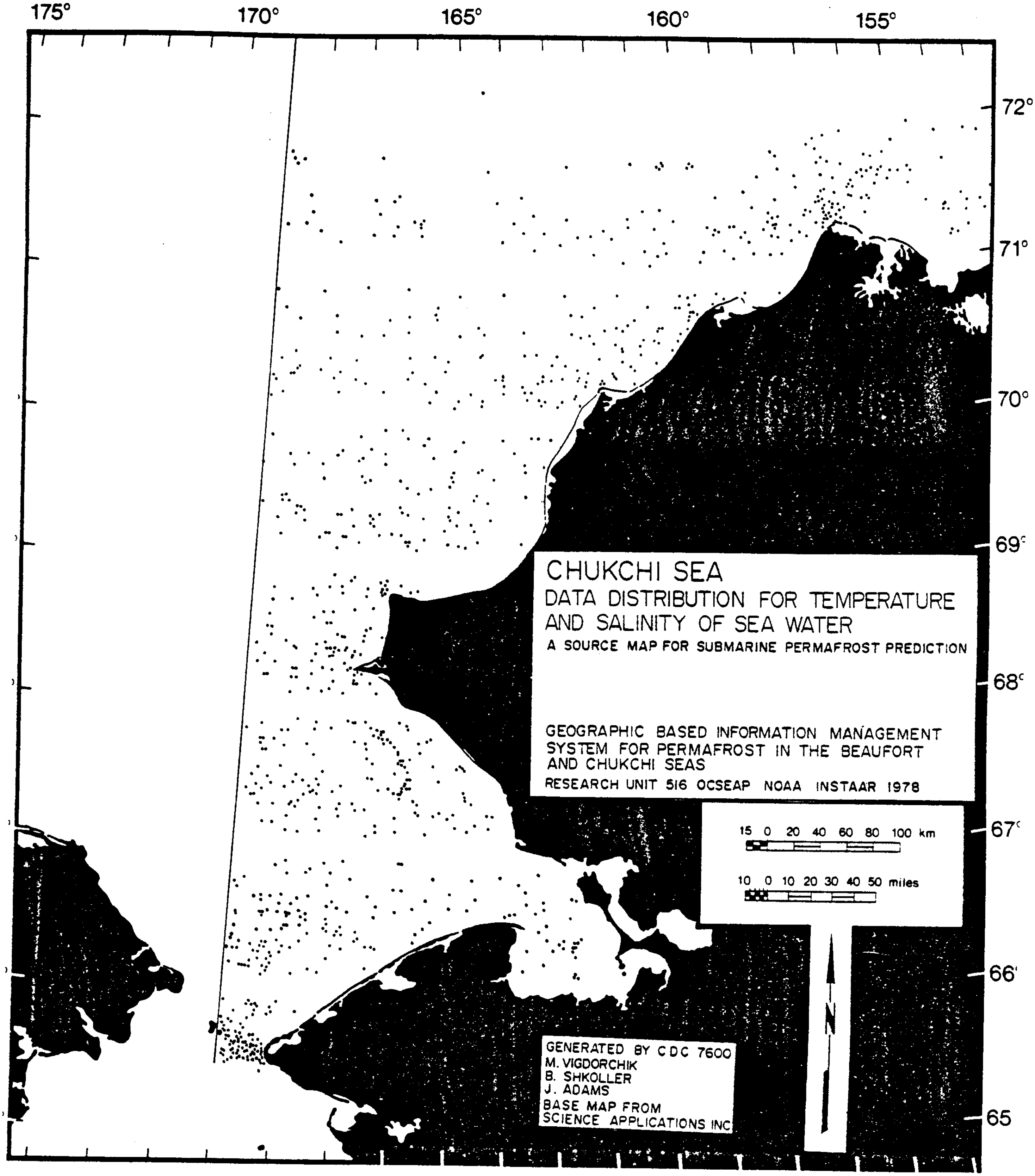
Map 2.—Points of observation for bottom deposit grain size, Beaufort Sea.



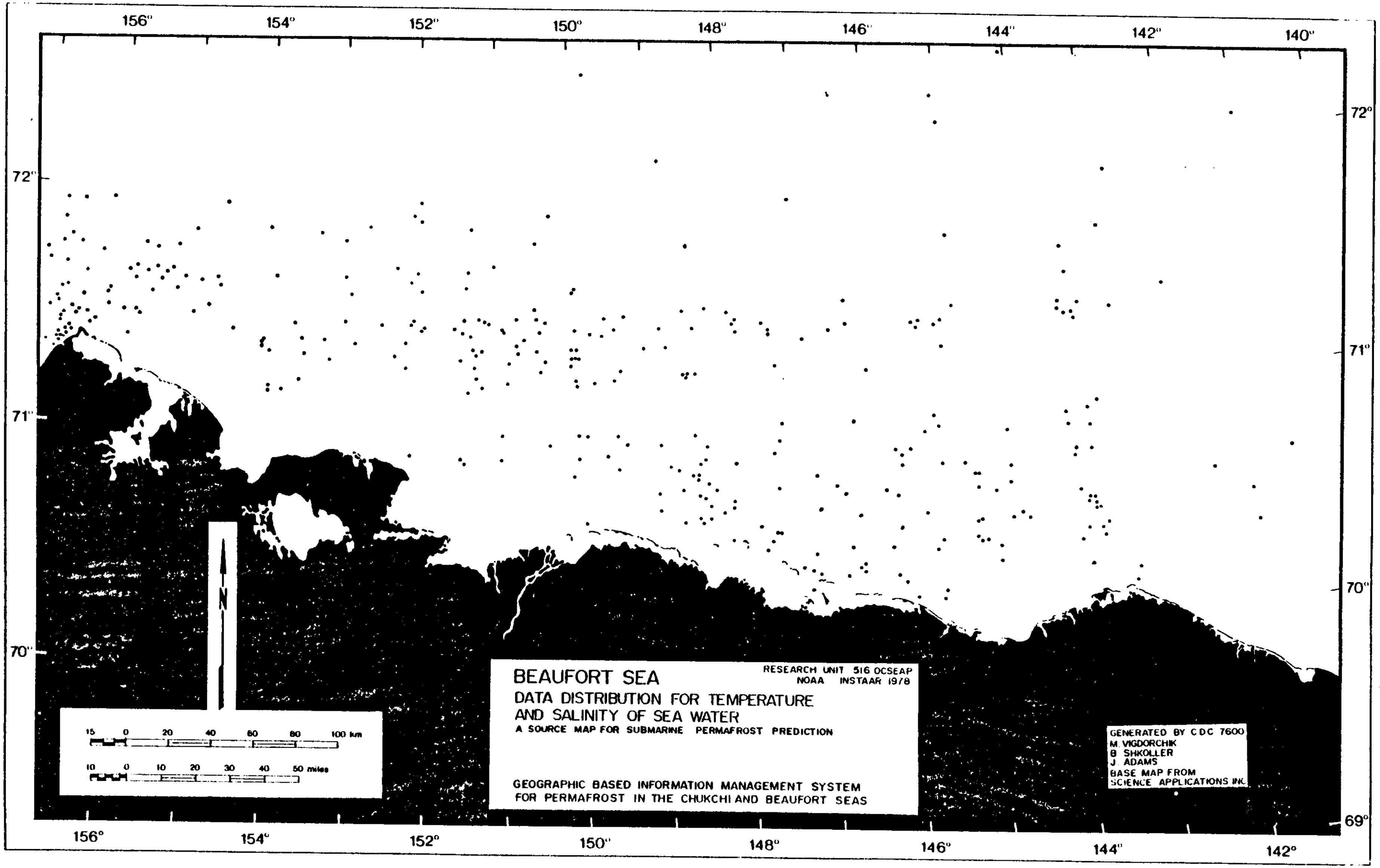
Map 3.—Reliability of the grain size data, Chukchi Sea.



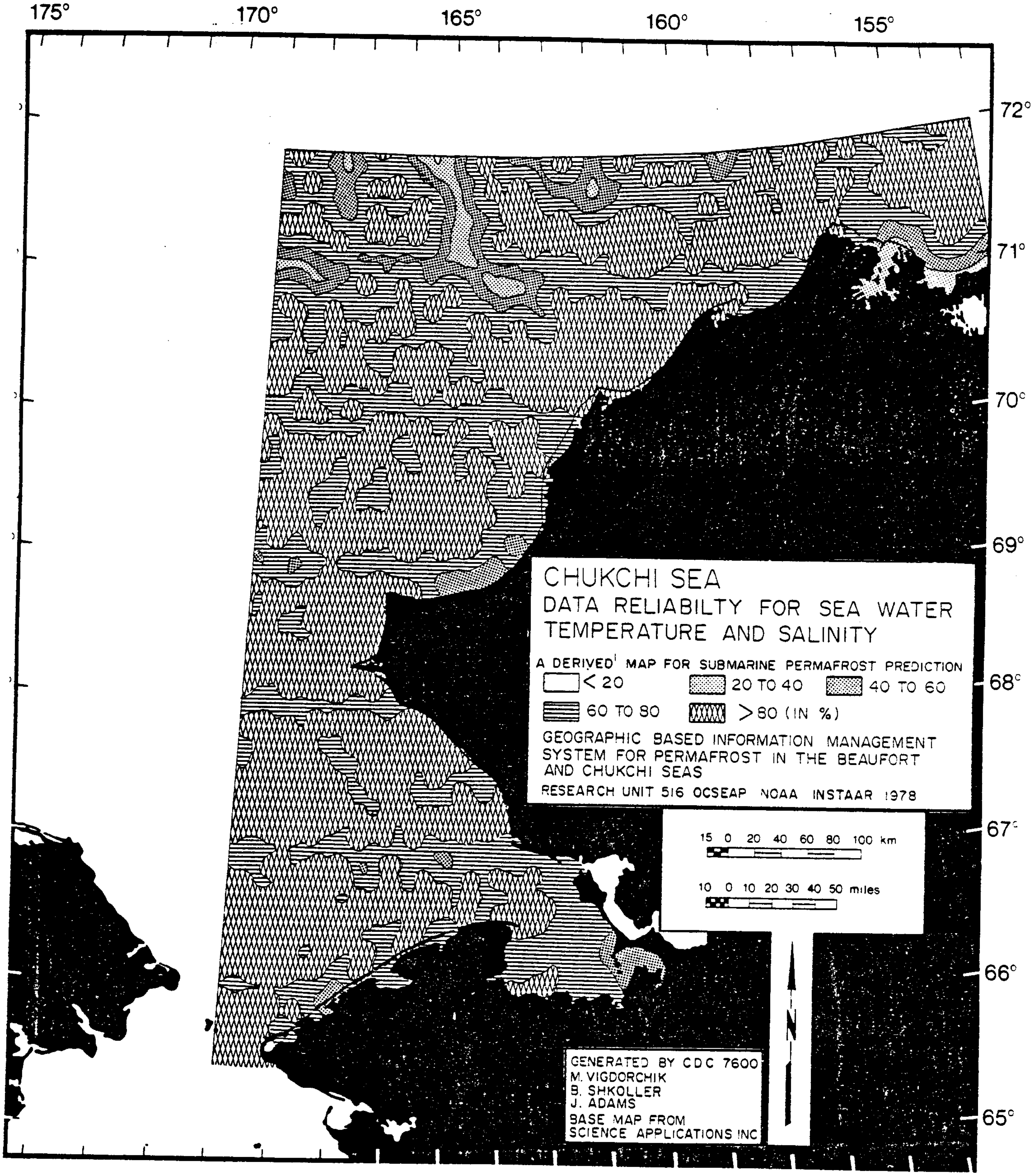
Map 4.—Reliability of the grain size data, Beaufort Sea.



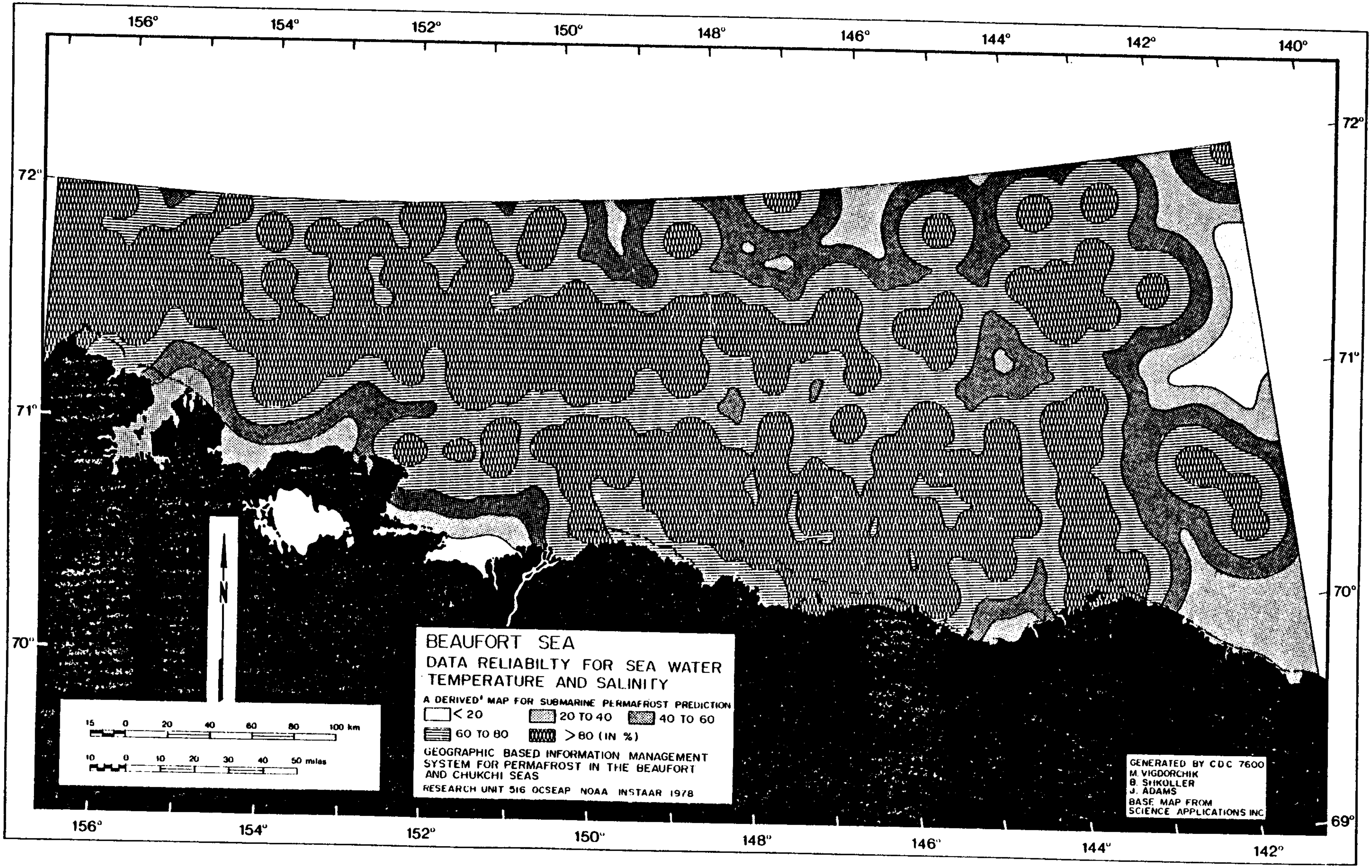
Map 5.—Data distribution for temperature and salinity of seawater at the maximal sampling depth, Chukchi Sea.



Map 6.—Data distribution for temperature and salinity of seawater at the maximal sampling depth, Beaufort Sea.



Map 7.—Data reliability for temperature and salinity of seawater at the maximal sampling depth, Chukchi Sea.



Map 8.—Data reliability for temperature and salinity of seawater at the maximal sampling depth, Beaufort Sea.

only one. The capability of the seawater to change these parameters yearly, monthly (Tables 4 and 5), and daily is shown very well by Barnes and Reimnitz (1977) for the surface layers at the Colville River delta–Oliktok Point part of the Beaufort sea and it poses another problem. Winter freezing coupled with the exclusion of solutes also affects the salinity and temperature regime of the seawater. But the bottom layers of the seawater, which area our concern, seem to be more stable, and we tried to choose the intervals of contouring according to possible limits of such temperature and salinity changes. Because the seawater sampling depth in the

Table 4.—Distribution of observations by month, Beaufort Sea.

Month	Number	Sampling Depth	Bottom	Temperature	Salinity
1	2	2	1	2	2
2	1	1	1	1	1
3	1	1	1	1	1
4	–	–	–	–	–
5	3	3	–	3	2
6	–	–	–	–	–
7	28	28	18	18	17
8	358	358	351	355	352
9	85	85	81	83	84
10	13	13	13	8	12
11	1	1	1	1	1
12	2	2	2	2	2
Total	491	491	469	472	472

Table 5.—Distribution of observations by month, Chukchi Sea.

Month	Number	Sampling Depth	Bottom	Temperature	Salinity
1	2	2	1	1	2
2	1	1	1	1	1
3	1	1	1	1	1
4	12	12	12	12	12
5	41	41	3	3	14
6	28	28	28	28	26
7	215	215	171	215	206
8	1,957	1,957	1,913	1,957	1,947
9	1,230	1,213	1,196	1,228	1,227
10	213	213	213	209	208
11	26	26	26	18	22
12	2	2	2	2	2
Total	3,718	3,709	3,764	3,764	3,668

Beaufort and Chukchi seas is not always close to the bottom, we need to consider also the relative distance between the deepest interval of sampling and the bottom. The increase of the sampling depth proximity to the bottom (Maps 9 and 10) gives more reliable data on these parameters close to the seabed. The maps of the general reliability (Maps 11 and 12) summarize the picture of the used data reliability (grain size, seawater temperature and salinity) for given areas of the Beaufort and Chukchi seas.

Interpolation Scheme

A 2-dimensional second order polynomial interpolation seems to give the best results. The numerical scheme follows. If we wish to calculate the value of some physical or geological feature at Point M (Xo, Yo), we first limit our considerations to the domain p′ (Xi, Yi), which satisfies the condition:

$$\sqrt{(Xo-Xi)^2 + (Yo-Yi)^2} \leq R$$

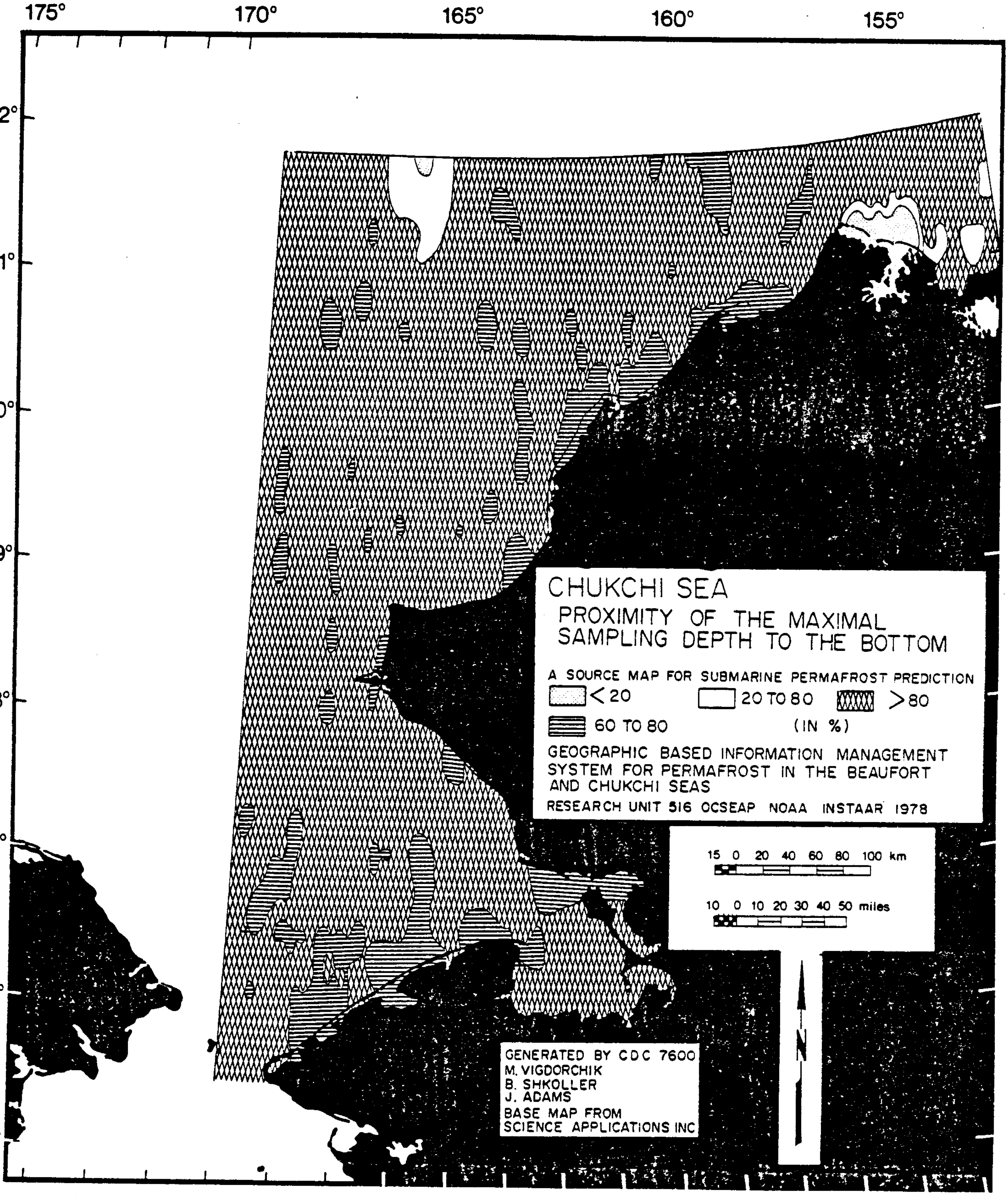
In other words, the point M is inside the circle of the radius R. Each point (Xi, Yi) has a special weight Pi, which increases when (Xi, Yi) is near (Xo, Yo) and decreases elsewhere. At the point M (Xo, Yo), the value is equal to 1. We take point M as an origin or coordinates. The value of the function in each point inside our domain can be approximated by a first or second degree polynomial.

$$Q_2(X,Y) = C_0 + C_1X + C_2Y + C_3X^2 + C_4XY + C_5Y^2$$

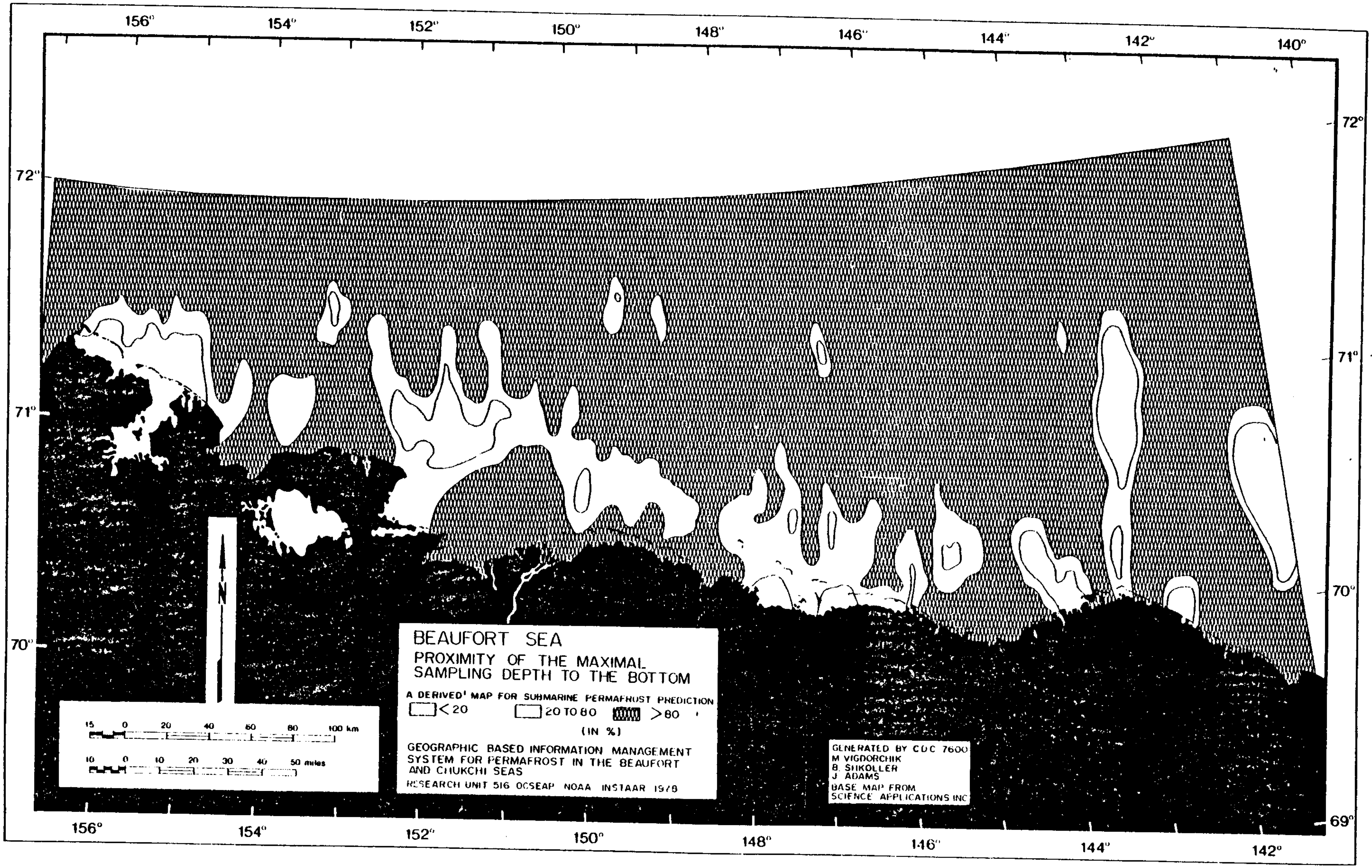
We will now show how to calculate the unknown coefficients C_0, C_1, C_2, C_3, C_4, C_5 in the case of the second order polynomial. This is done by using the least squares numerical method. We find such coefficients that will give the minimum to the sum

$$S = \sum_{i=1}^{N} P\,(C_0 + C_1X_i + C_2Y_i + C_3X^2_i + C_4X_iY_i + C_5Y_i - Q_i)^2$$

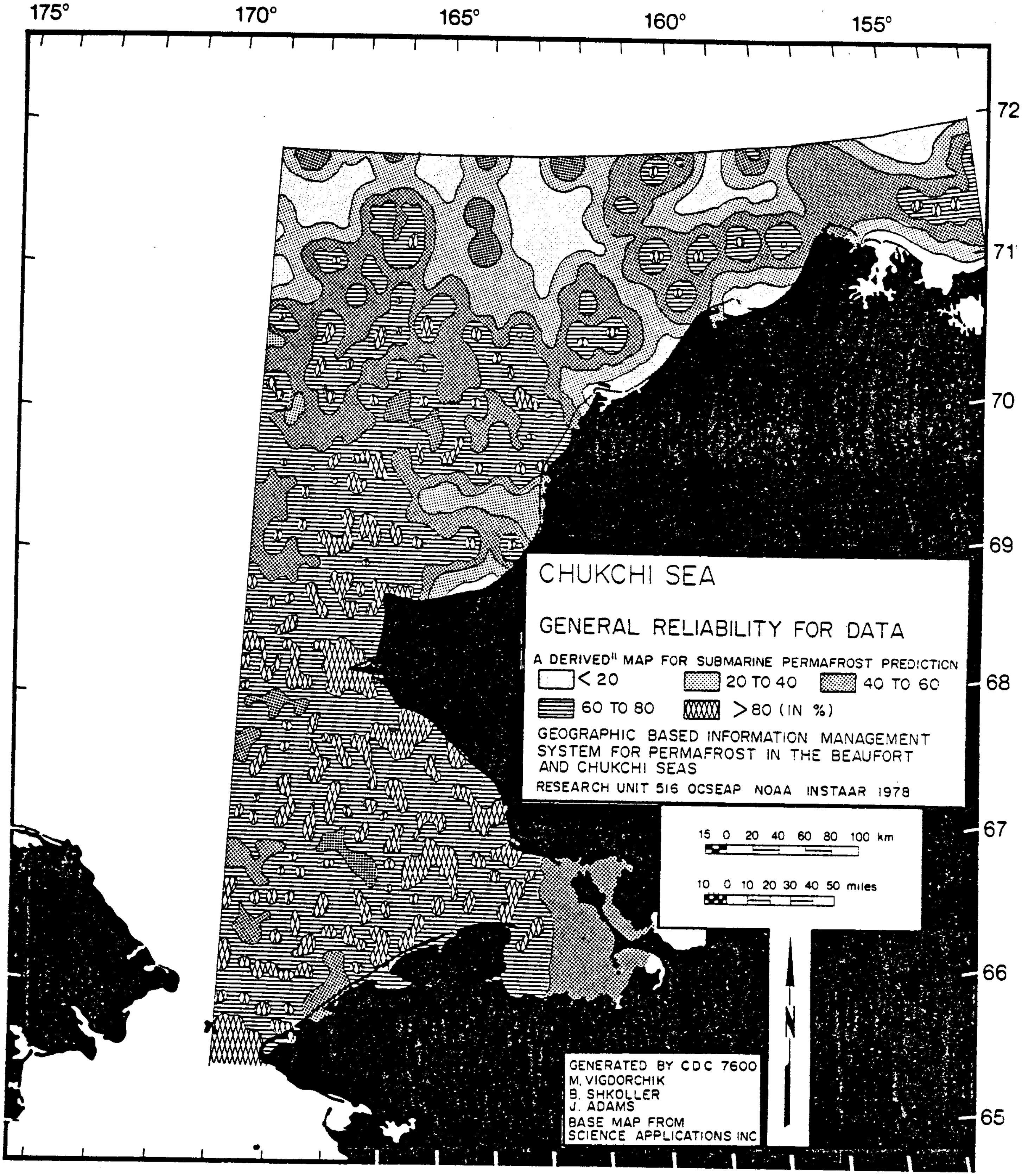
where N is number of observations inside the circle of radius R; X_i, Y_i are coordinates of the given observations inside the circle; and Q_i are the values of the function (salinity, temperature, depth) at these points. In order to obtain a minimum of S we must take the derivatives of this expression with regard to C_0, C_1, C_2, C_3, C_4, and C_5 and obtain six linear equations, which are called normal equations:



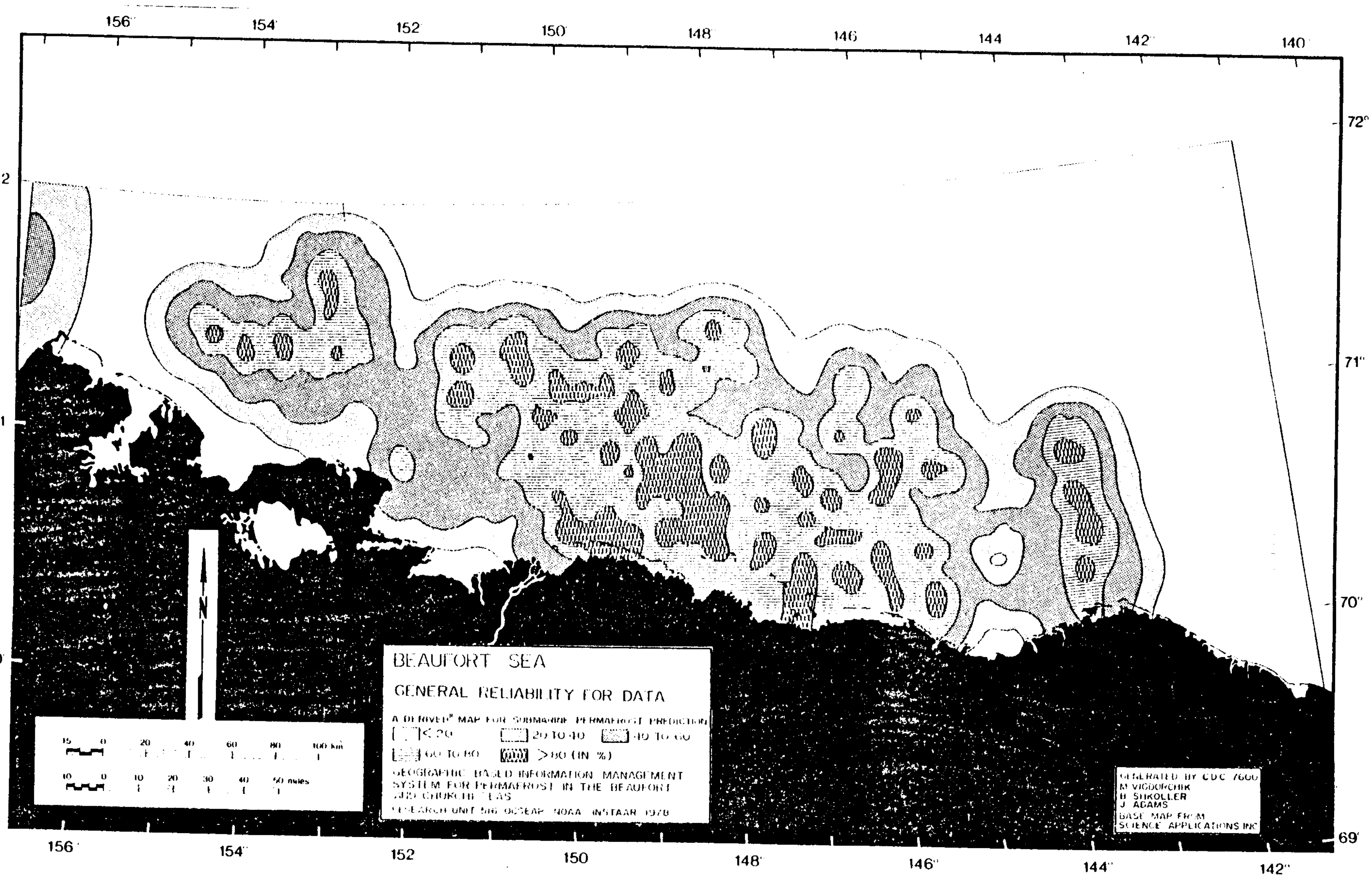
Map 9.—Proximity of the maximal sampling depth to the bottom, Chukchi Sea.



Map 10.—Proximity of the maximal sampling depth to the bottom, Beaufort Sea.



Map 11.—General reliability of the data, Chukchi Sea.



Map 12.—General reliability of the data, Beaufort Sea.

$\overline{X}_i = X_i - X_o$; $\overline{Y}_i = Y_i - Y_o$

$$C_o\tilde{\Sigma}P_i + C_1\tilde{\Sigma}P_i\overline{X}_i + C_2\tilde{\Sigma}P_i\overline{Y}_i + C_3\tilde{\Sigma}P_i\overline{X}_i^{\,2} + C_4\tilde{\Sigma}P_i\overline{X}_i\overline{Y}_i + C_5\tilde{\Sigma}P_i\overline{Y}_i^{\,2} = \tilde{\Sigma}P_iQ_i$$

$$C_o\tilde{\Sigma}P_i\overline{X}_i + C_1\tilde{\Sigma}P_i\overline{X}_i^{\,2} + C_2\tilde{\Sigma}P_i\overline{X}_i\overline{Y}_i + C_3\tilde{\Sigma}P_i\overline{X}_i^{\,3} + C_4\tilde{\Sigma}P_i\overline{X}_i^{\,2}\overline{Y}_i + C_5\tilde{\Sigma}P_i\overline{X}_i\overline{Y}_i^{\,2} = \tilde{\Sigma}P_i\overline{X}_iQ_i$$

$$C_o\tilde{\Sigma}P_i\overline{Y}_i + C_1\tilde{\Sigma}P_i\overline{X}_i\overline{Y}_i + C_2\tilde{\Sigma}P_i\overline{Y}_i^{\,2} + C_3\tilde{\Sigma}P_i\overline{X}_i^{\,2}\overline{Y}_i + C_4\tilde{\Sigma}P_i\overline{X}_iY_i2 + C_5\tilde{\Sigma}P_i\overline{Y}_i^{\,3} = \tilde{\Sigma}P_i\overline{Y}_iQ_i$$

$$C_o\tilde{\Sigma}P_i\overline{X}_i^{\,2} + C_1\tilde{\Sigma}P_i\overline{X}_i^{\,3} + C_2\tilde{\Sigma}P_i\overline{X}_i^{\,2}\overline{Y}_i + C_3\tilde{\Sigma}P_i\overline{X}_i^{\,4} + C_4\tilde{\Sigma}P_i\overline{X}_i^{\,3}\overline{Y}_i + C_5\tilde{\Sigma}P_i\overline{X}_i^{\,2}Y_i^{\,2} = \Sigma P_i\overline{X}_i^{\,2}Q_i$$

$$C_o\tilde{\Sigma}P_i\overline{X}_i\overline{Y}_i + C_1\tilde{\Sigma}P_i\overline{X}_i^{\,2}\overline{Y}_i + C_2\tilde{\Sigma}P_i\overline{X}_i\overline{Y}_i^{\,2} + C_3\tilde{\Sigma}P_i\overline{X}_i^{\,3}\overline{Y}_i + C_4\tilde{\Sigma}P_i\overline{X}_i^{\,2}\overline{Y}_i^{\,2} + C_5\tilde{\Sigma}P_i\overline{X}_i\overline{Y}_i^{\,3} = \tilde{\Sigma}P_i\overline{X}_i\overline{Y}_iQ_i$$

$$C_o\tilde{\Sigma}P_i\overline{Y}_i^{\,2} + C_1\tilde{\Sigma}P_i\overline{X}_i\overline{Y}_i^{\,2} + C_2\tilde{\Sigma}P_i\overline{Y}_i^{\,3} + C_3\tilde{\Sigma}P_i\overline{X}_i^{\,2}\overline{Y}_i^{\,2} + C_4\tilde{\Sigma}P_i\overline{X}_i\overline{Y}_i^{\,3} + C_5\tilde{\Sigma}P_i\overline{Y}_i^{\,4} = \tilde{\Sigma}P_iQ_i\overline{Y}_i^{\,2}$$

Remembering that M (Xo, Yo) is the origin of the coordinates $\overline{X}_i = 0$, $\overline{Y}_i = 0$, we have only to find C_0; then $Q(m) = C_0$. When calculating the value of the needed function at point M, we have only to move to the next point and repeat the above calculations. The value of Pi is a function of the distance

$$\zeta_i = \sqrt{(Xi - Xo)^2 + (Yi - Yo)^2}$$

and must equal zero when di=R. In our calculations we have chosen the expression

$$P_i = ((R^2 - \zeta_i^2)/\zeta_i^2)^2$$

The radius R was chosen as $R = 25\Delta X$; however, it is important to be sure that at least six observations are inside the circle when using a second order interpolation. Generally all calculations were made with second order polynomials. As a test, we also used the first order, and usually the results were almost the same. However, the second order as a rule gives the smoother contours.

Data Structuring

Data Structure Diagram (Figure 22) represents and organizes data requirements, the stages of mapping, and the production of information resulting from the study. This diagram differs from the Data Structure Diagram of the first reports in that some changes were made after studying the available source data in this area, including such parameters as: sea bottom temperatures and salinities, grain size of the Holocene sediments, and bathymetry.

This Data Structure Diagram illustrates the relationships among the data information used in the study, and it can be viewed as describing the flow of mappable information. The layout and content of the diagram are developed in response to the relevant issues. The diagram is organized both horizontally and vertically, with the vertical organization arranged by the type of map analysis. The *source data* column contains "nonvalue-oriented" data from maps. The next few columns, *derived data* maps, contain the results of the two or three stages of the usage of the data management. Then, the composite map displays information developed from source data and derived maps. This map is defined by interdisciplinary knowledge and the relationships between source data topics. All these maps serve as a basis for further subjective analysis.

We have used computer methods as the mechanisms for identifying and organizing the multiplicity of the values of the data into a form useful in the composite analysis stages. Composite mapping records and illustrates our opinion about the major problem—the tracing of areas with potential for the existence of permafrost in the Beaufort and Chukchi seas offshore.

Four basic techniques are used for interpreting and analyzing the data: (1) the Translation technique for converting a single source data map into a secondary data map; (2) the Comparison technique for comparing two or more maps in order to produce a third derived map showing the results of the comparison; (3) the Overlay technique for combining two or more maps in order to produce a composite map showing the results of the overlay process; and (4) the Distance technique which is used for calculating the distance of all geographical areas from a given point, line, or area.

We have six blocks in our system: seawater, bathymetry, geology, sea ice, direct data on permafrost, and reliability of data. Three generations of derived maps and one of composite map have been produced.

COMPUTERIZED MAPPING

The computerized mapping involves the following phases:

1.) a) Data digitation and preparation for computer use.

b) Preparation of *source* maps (one map—one characteristic).

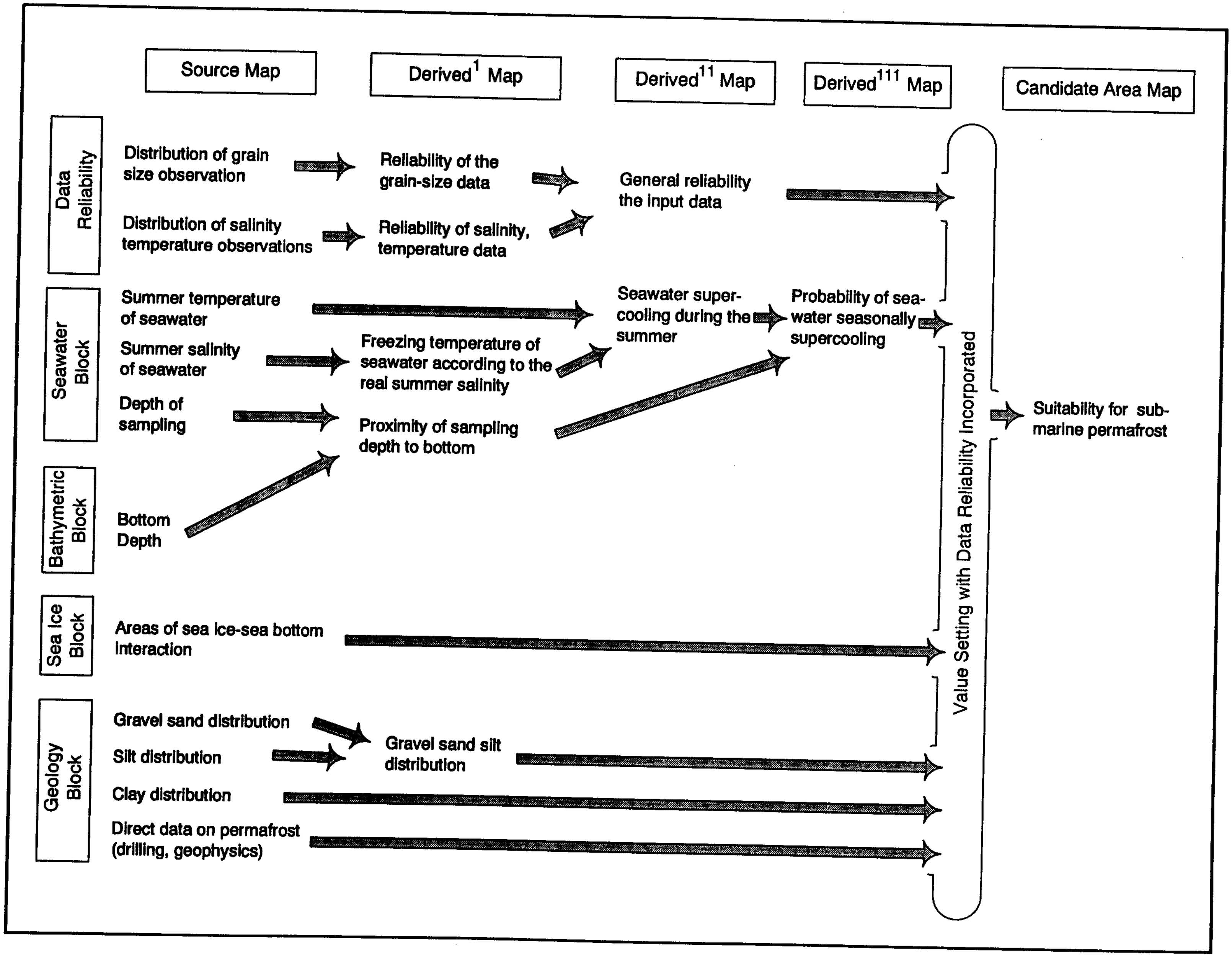


Figure 22.—Structure diagram of the data management system for submarine permafrost prediction in the Beaufort and Chukchi seas.

2.) Generation of *derived* maps, based on source maps and/or on the physical and statistical relations between different characteristics, belonging to one discipline (first generation of the derived maps) or several disciplines (second or third generations).

3.) Compilation of *composite* maps based on the assessment value setting obtained from all previous maps. A composite map is produced as a result of the overlay process for combining several maps in order to reach the ultimate goal of the work. In terms of probability it has to specify the candidate areas suitable for submarine permafrost in the Beaufort and Chukchi seas according to the paleoenvironmental and environmental data. To compile the composite map we used the source data on bathymetry, grain size, sea ice, and seawater.

Source Maps

The grain size source maps are based on the grade scale most commonly used for sediments—Wentworth Size Classes (Wentworth 1922). The materials of Barnes and Reimnitz (1976), Naidu and Mowat (1974), and Creager and McManus (1967) have been used for computerized source mapping of the distribution percent of clay, silt, silt plus sand plus gravel at the Beaufort and Chukchi seas. It is known that the grain size of the sediments is an important factor in ice segregation during the freezing process. According to the empirical data, ice segregation is very active in the silty deposits. The impermeable clay is not suitable for ice segregation. In pure sands, ice segregation is possible in the course of water freezing, usually if there is a piezometric head. In Figure 23 and in Table 6 (after Tsythovich 1973) we see that the finer (more clayey) the soil, the larger the amount of unfrozen water it contains at a given subzero temperature. This is understandable when we realize that finer soils have larger specific surface mineral-particle areas and, consequently, have a greater capacity for the binding of pore water. Another factor in ice segregation is the heterogenic structure of the sediments, with changes of grain size mostly in the vertical direction. That is why the areas with silt and silty sands (sometimes with gravel) could be considered as more suitable for ice bonding (if the temperature and other conditions are favorable) and subsea permafrost development. Areas of clay domination usually have low potential for ice-bonded permafrost. The aerial distribution of these grain size classes is shown in source maps 13–18.

Temperatures of the seawater (Aagard 1977, 1978; Aagard and Haygen 1978; Coachman and Aagard 1974; Hufford 1973; Hufford, Fortier et al. 1974; Hugget et al. 1977) at the maximal sampling depth during the summer (Maps 19 and 20) in the southern Chukchi Sea and close to Point Barrow are above 0°C. The main boundary of this "warm" deep water and the cold deep water (0°C to −3°C) could be traced along latitudes 70°40′–71°20′. In the Beaufort Sea deep water layers, negative temperatures dominate but there are several spots of "warm" water

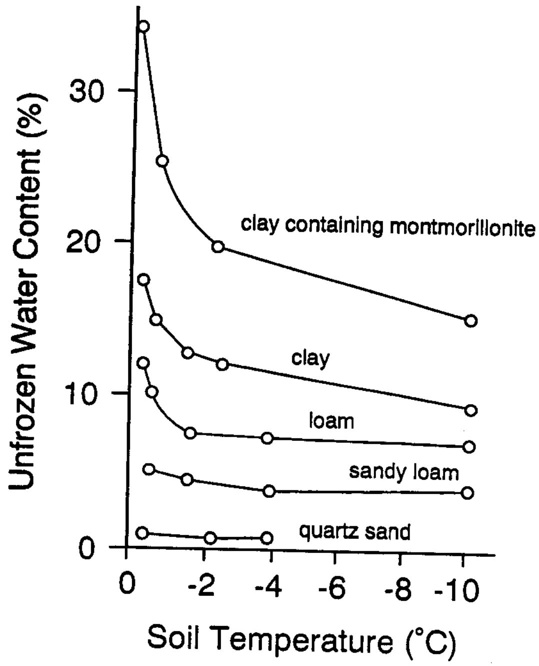


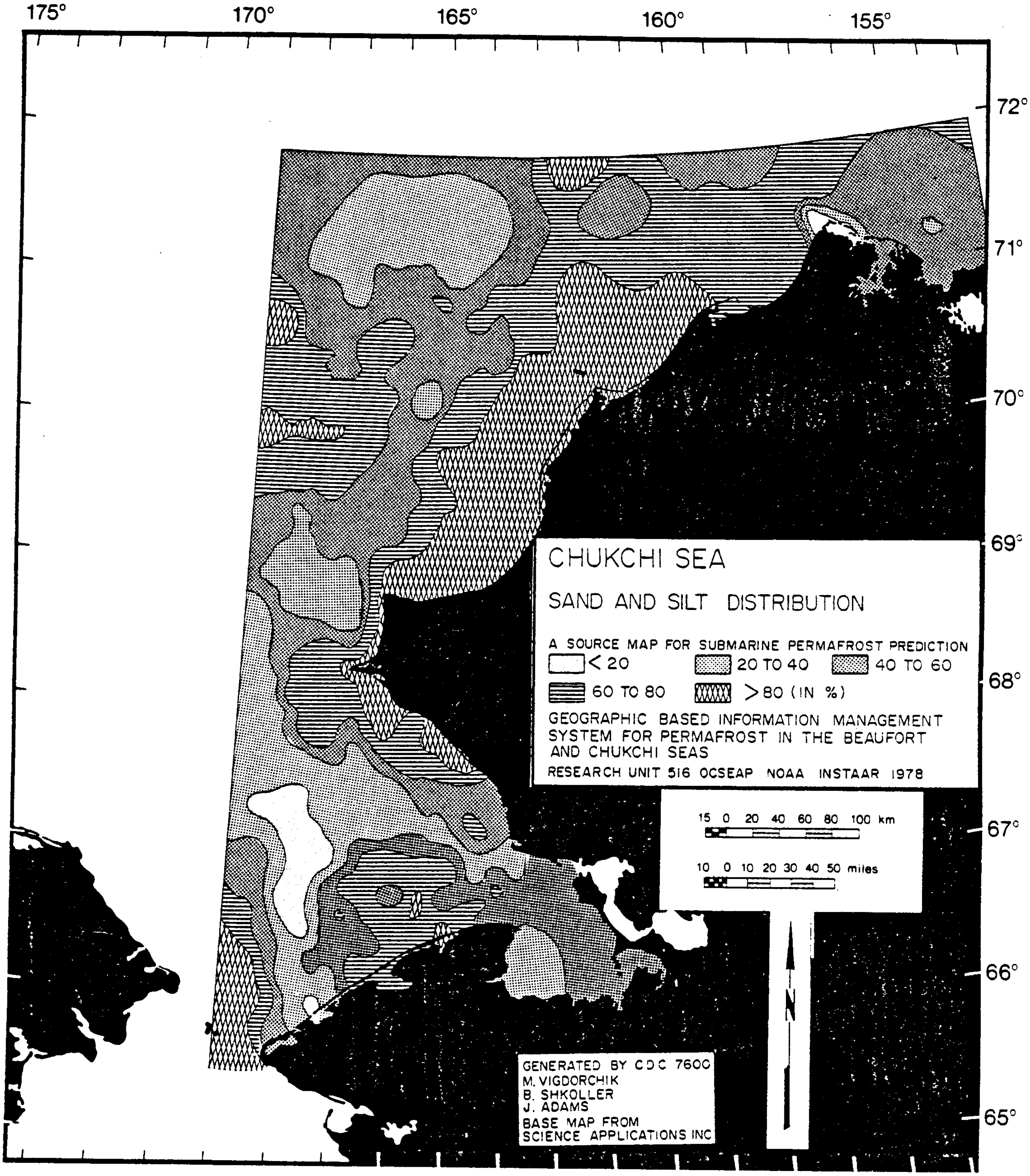
Figure 23.—Graph of unfrozen water content in frozen soils. After Tsytovich (1973).

close to the coast and river deltas, which is normal. But the specific of some of these spots seemed to be their meridional shapes oriented to the north (seen in the case of the very detailed intervals of mapping). There is a possibility that some of them could be related to the meridional geomorphological and structural lineaments of the shelf. We can distinguish these meridional lineaments in the river systems (see above), lakes, and topography of the coastal area, the former sea bottom continuing toward the shelf. The "meridional type" of these morphostructural elements could be easily reflected in the distribution of other natural conditions, including the temperature of the deep water layers. On the Eurasiatic shelf of the Arctic basin the alternation of the relatively cool and warm belts, sometimes considered in connection with the hydrogeological peculiarities of the areas and the discharge of the underground and artesian water through the fractures (see Part II). Salinity of the deep seawater layers during the summer (Maps 21 and 22) in the Chukchi Sea is usually between the 27–31 ppt along the coast and 31–35 ppt for the rest of the sea. In the Beaufort Sea close to Point Barrow, Elson Lagoon, Cape Halkett, the Colville Delta, and Prudhoe Bay, the salinity is usually low, less than 27 ppt, sometimes even less. The more saline deep water extends to the northern parts of the shelf (31–35 ppt and greater).

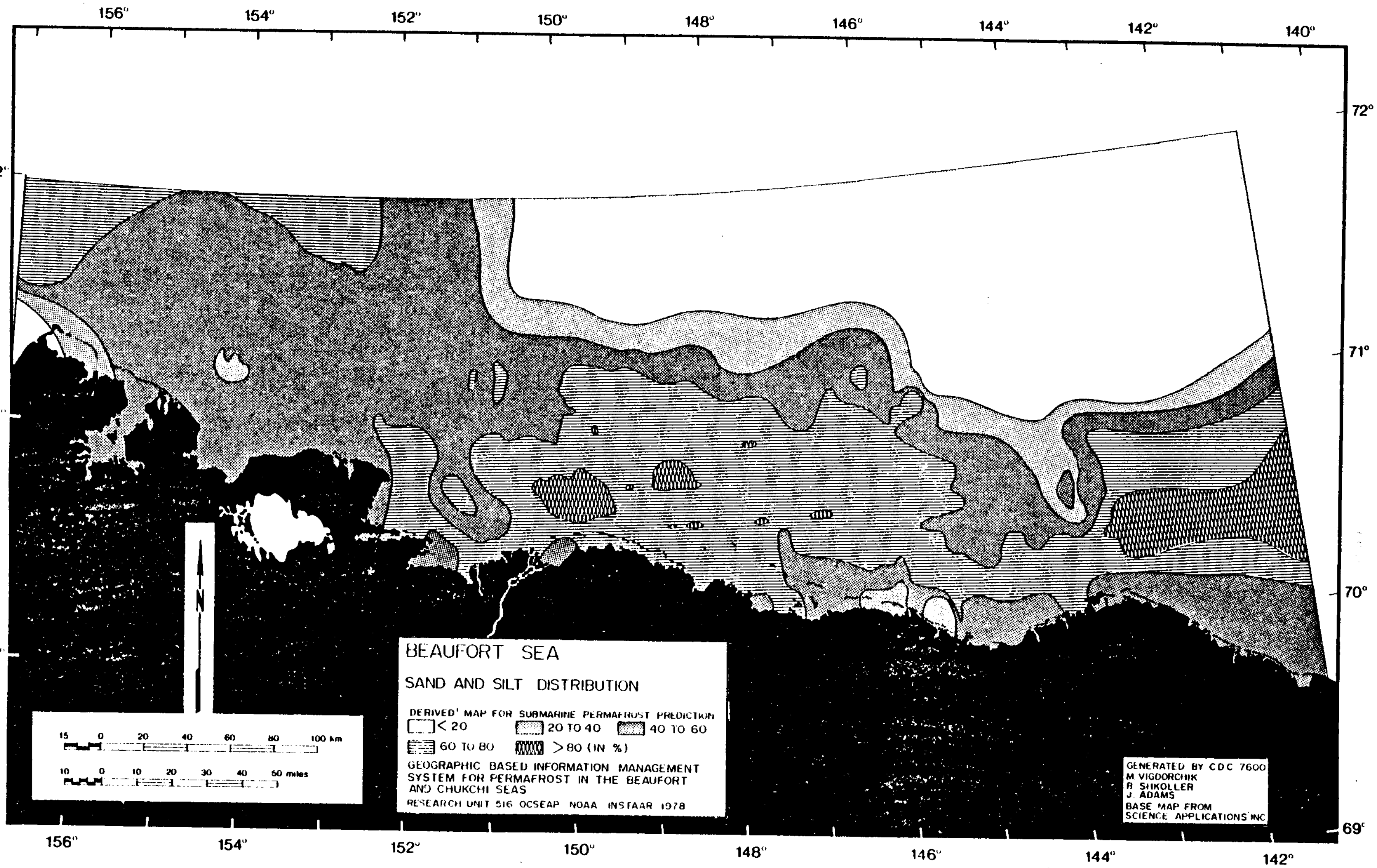
Table 6.—Unfrozen-water contents of salt free soils vs. degrees of negative temperature. After Tsytovich (1973).

Designation of Soil	Amount of unfrozen water, in percent of weight of dry soil, as a function of temperature in °C					
	−0.2 to −0.5	−0.5 to −0.5	−1.0 to −1.5	−2.0 to −2.5	−4.0 to −4.5	−10.0 to −11.0
Sand	0.2	0.2	–	–	0.0	0.0
Sandy loam	–	5.0	4.5	–	4.0	3.5
Loam	12.0	10.0	7.8	–	7.0	6.5
Clay	17.5	15.0	13.0	12.5	–	9.3
Clay containing montmorillonite	34.3	25.9	–	19.8	–	15.3

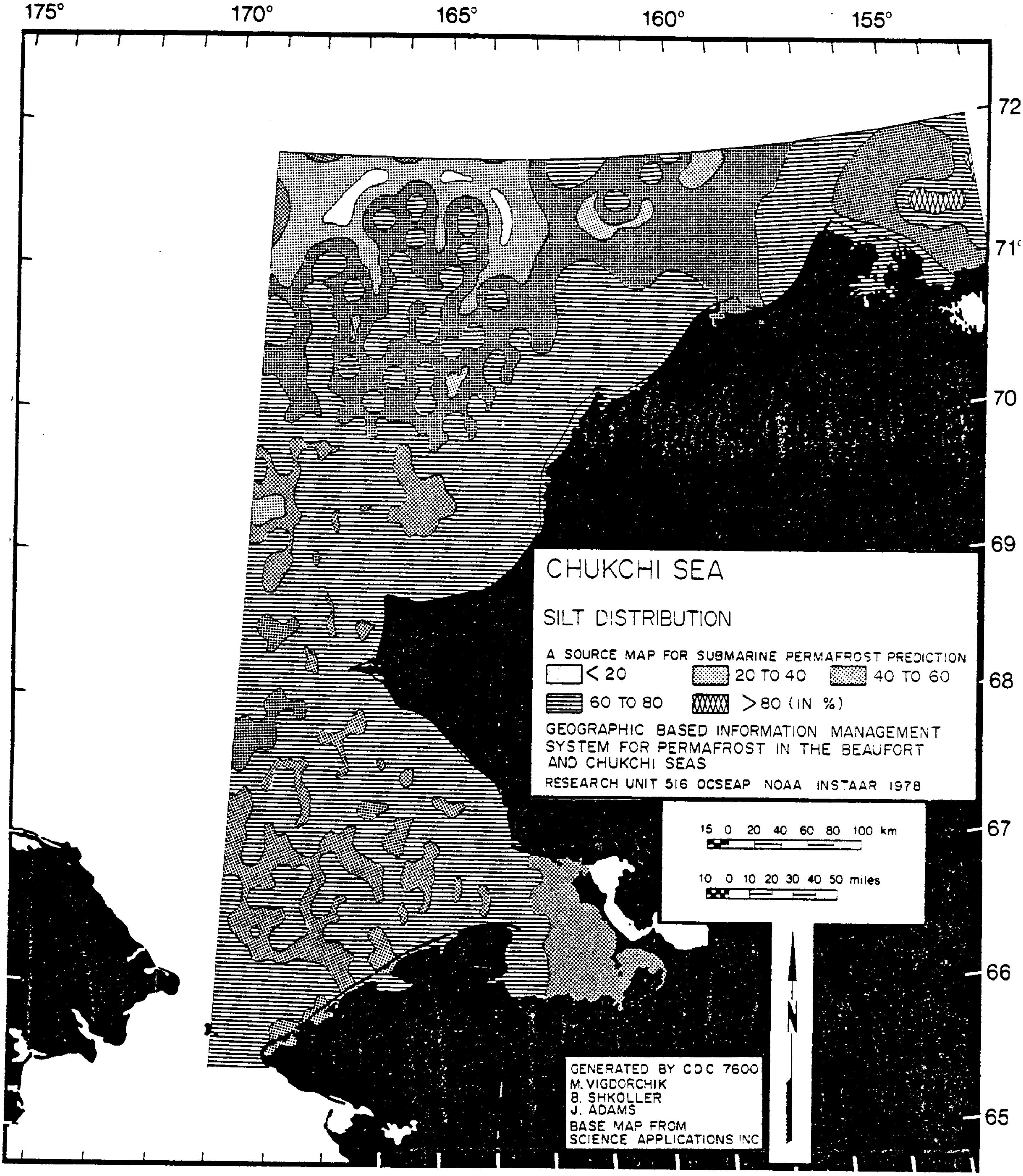
* (continuing montmorillonite)



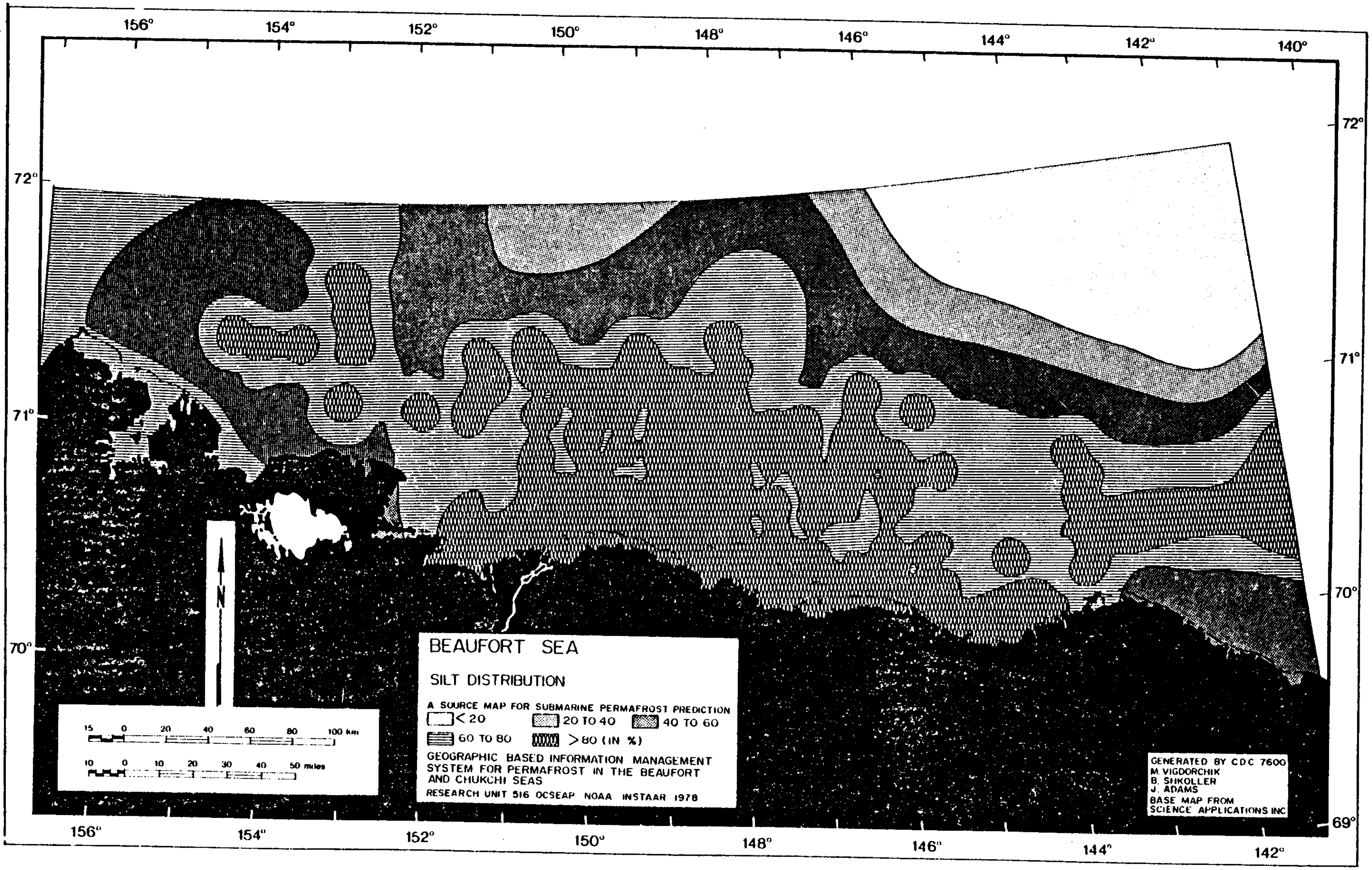
Map 13.—Sand and silt distribution, Chukchi Sea.



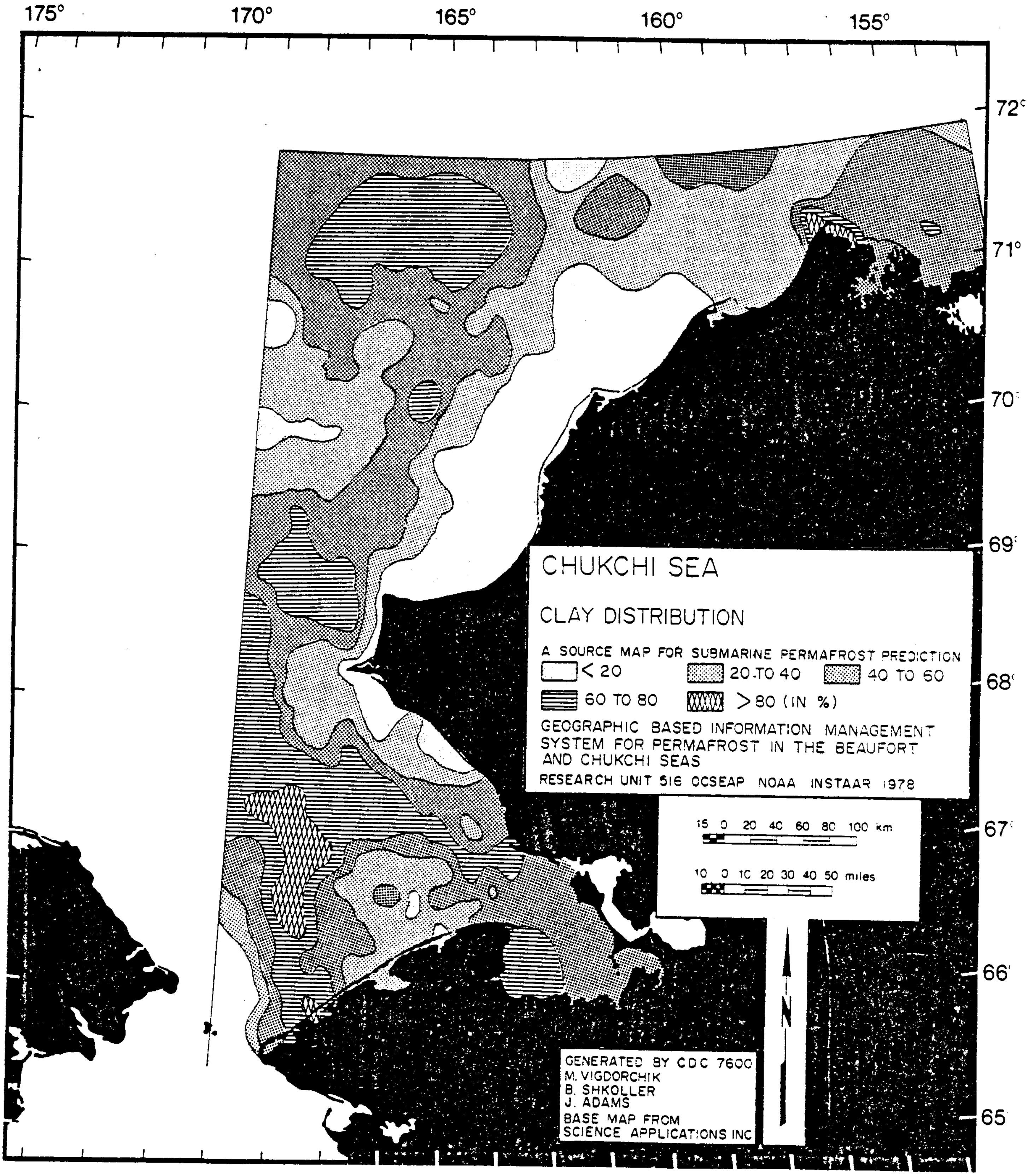
Map 14.—Sand and silt distribution, Beaufort Sea.



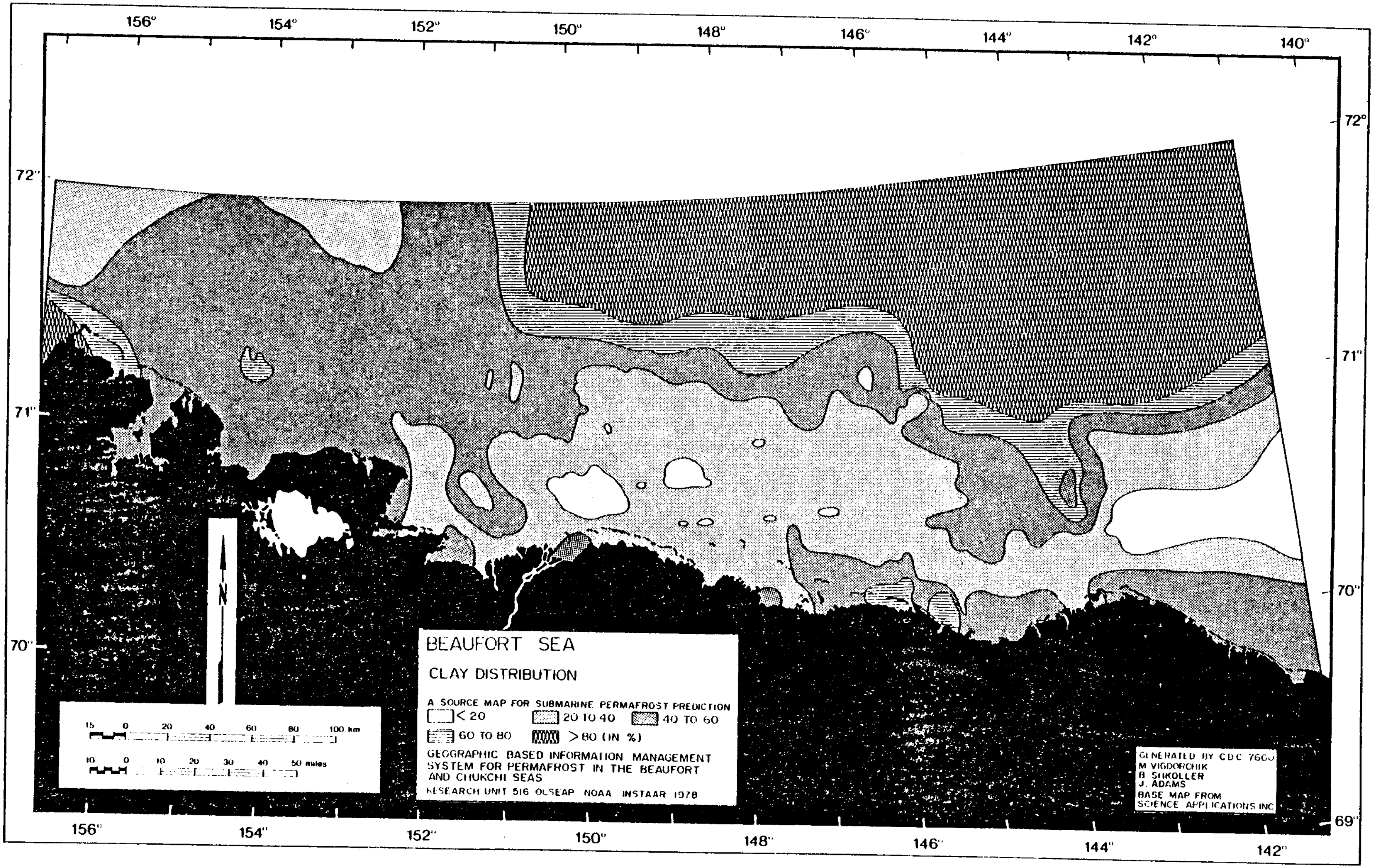
Map 15.—Silt distribution, Chukchi Sea.



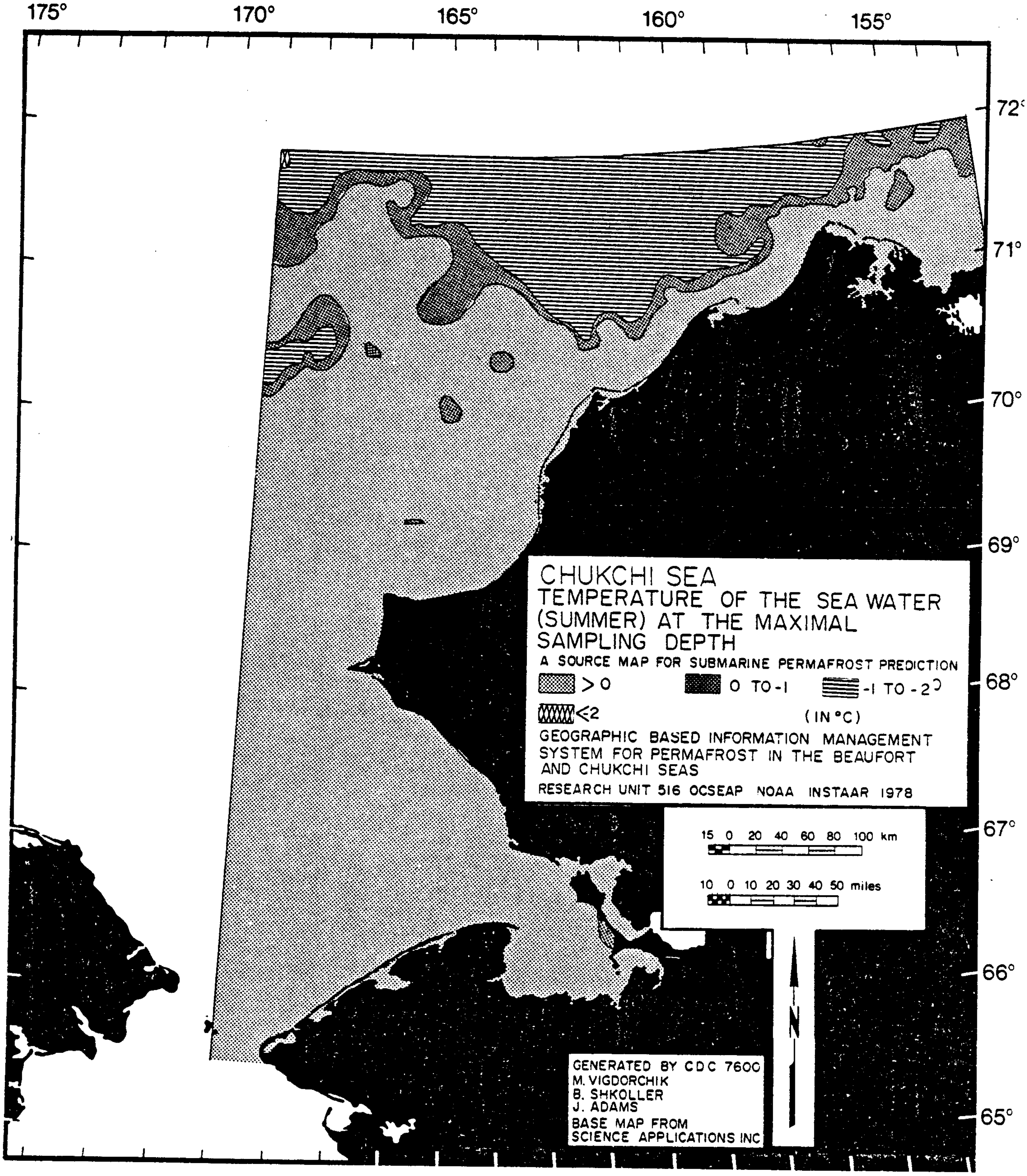
Map 16.—Silt distribution, Beaufort Sea.



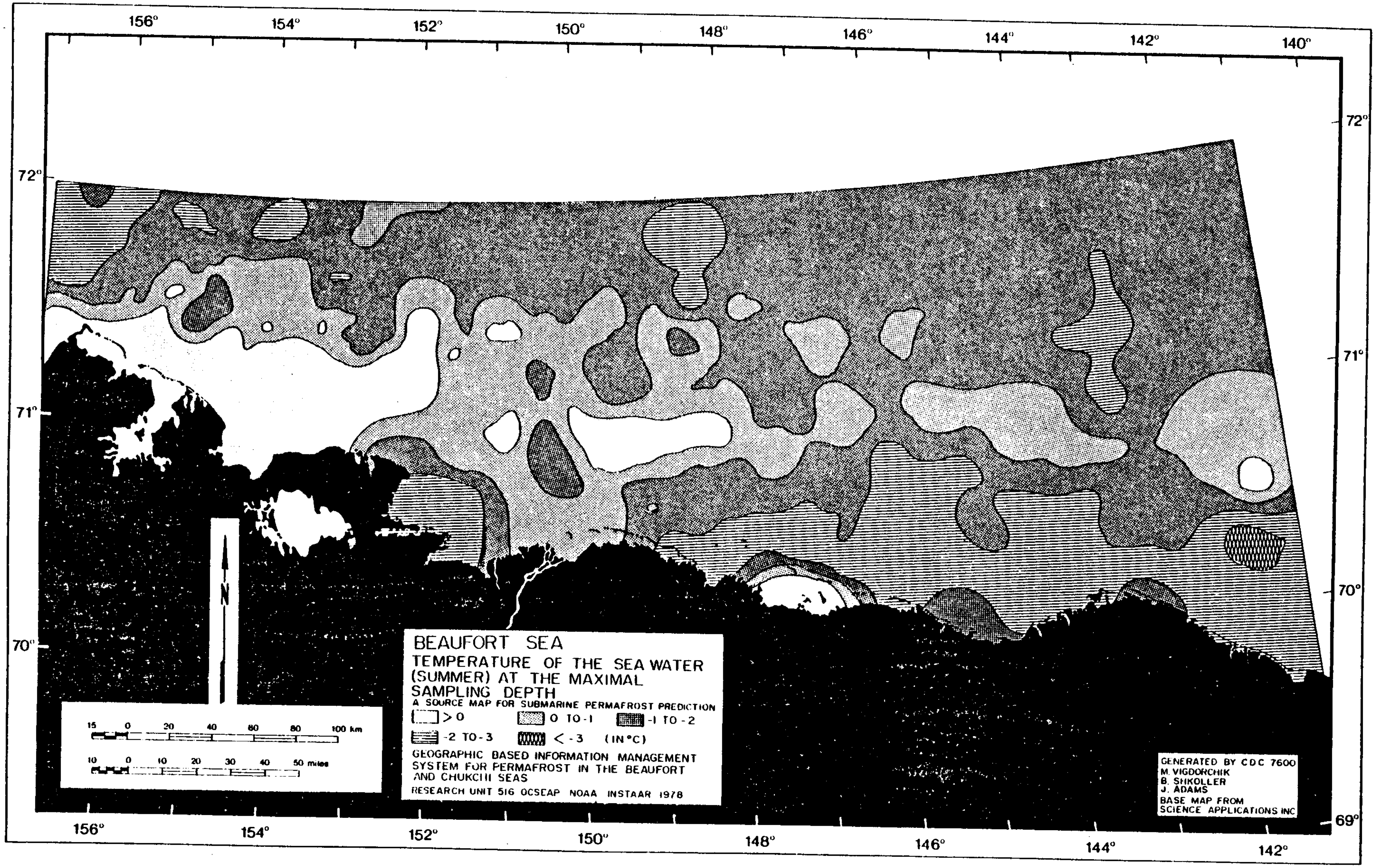
Map 17.—Clay distribution, Chukchi Sea.



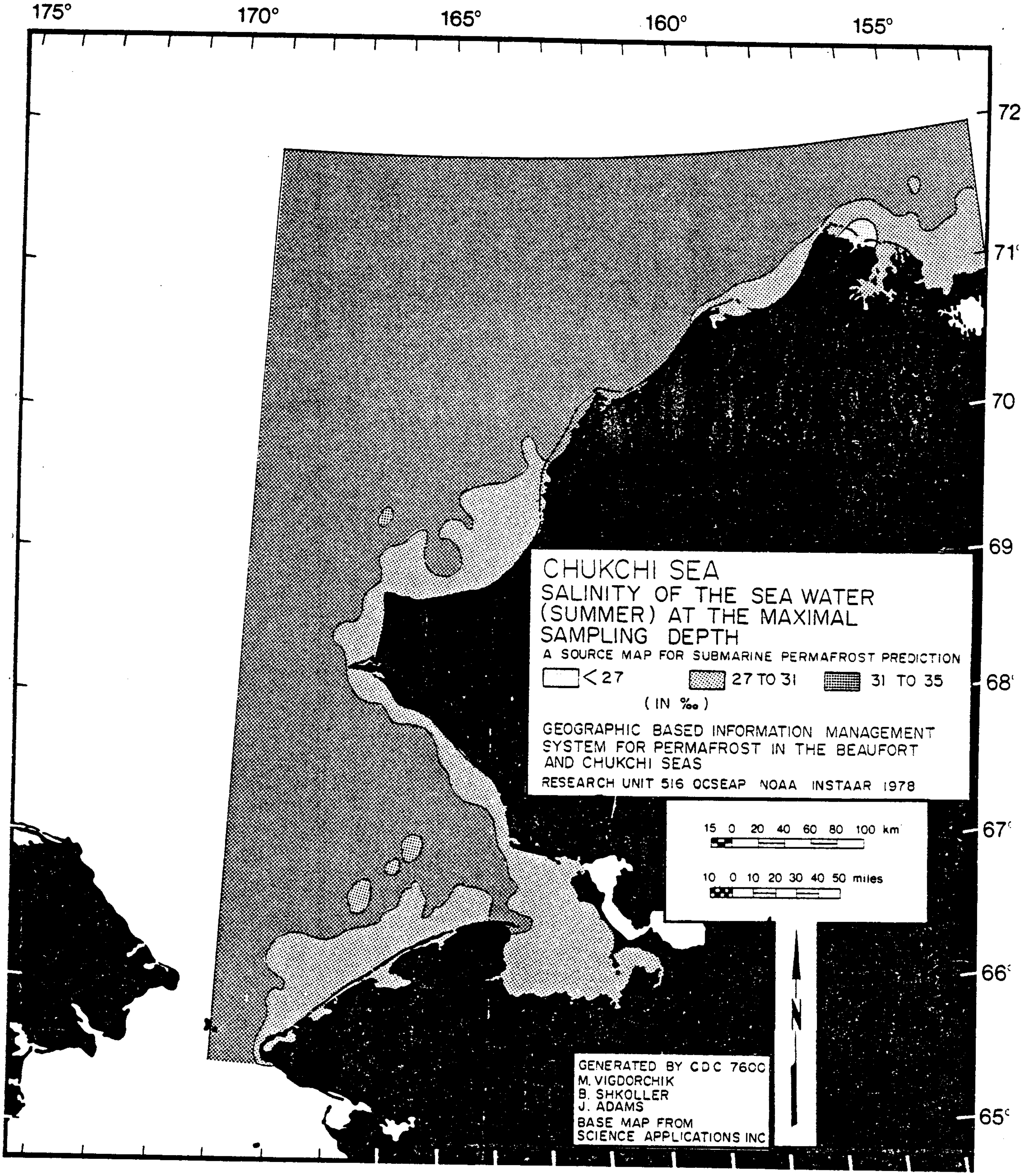
Map 18.—Clay distribution, Beaufort Sea.



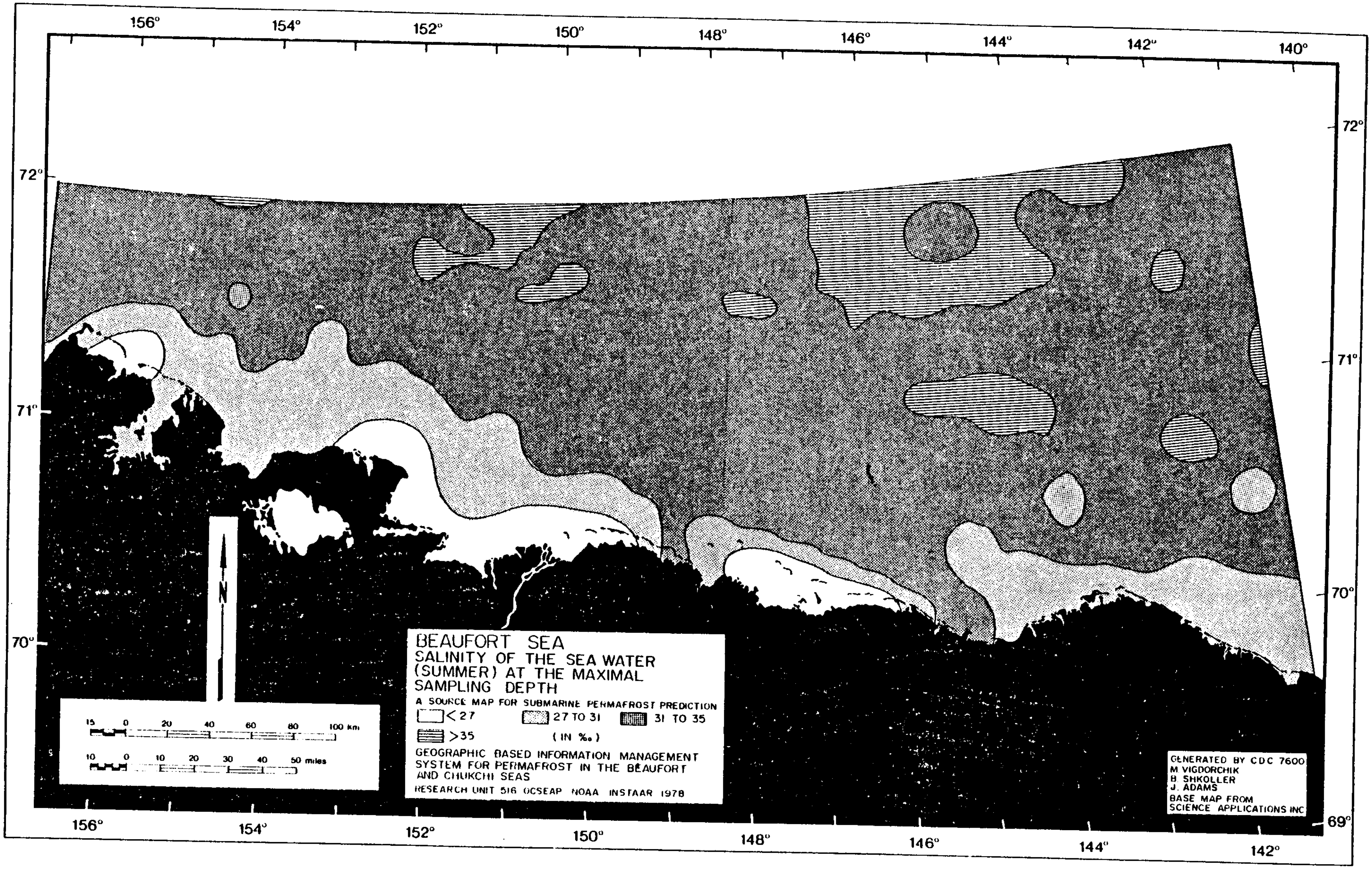
Map 19.—Temperature of seawater in summer at the maximal sampling depth, Chukchi Sea.



Map 20.—Temperature of seawater in summer at the maximal sampling depth, Beaufort Sea.



Map 21.—Salinity of seawater in summer at the maximal sampling depth, Chukchi Sea.



Map 22.—Salinity of seawater in summer at the maximal sampling depth, Beaufort Sea.

Derived Map of Freezing Temperature of Seawater according to Real Summer Salinity (Fij)

This can be produced using Savel'ev's formula (1963)*

$$Fij = 2.6 \cdot 10^{-3} - 5.265 \cdot 10^{-2}S - 2.89 \cdot 10^{-6}S^{2} - 3.6 \cdot 10^{-7}S^{3} - 1.2 \cdot 10^{-9}S^{3},$$

or according to Krummel's formula:

$$Fij = 0.003 - 0.0527S - 0.4 \cdot 10^{-4}S^{2} - 0.4 \cdot 10^{-6}S^{3}$$

To generate the map of this characteristic (Maps 23 and 24) we have used the second equation according to the real summer salinity (Sij).

Seawater Supercooling in Summer (Cij)

This map (Maps 25 and 26) gives the difference between real summer water temperature (Tij) and freezing temperature (Fij).

$$Cij = Tij - Fij$$

Probability of Seawater Seasonal Supercooling (Maps 27 and 28)

Coachman (1966)** used the observations from more than 300 oceanographic stations in the Arctic basin to determine the deviations of water temperature in the upper 50 m layer from the freezing temperature for the given salinity (Figure 24). The results of the calculations for depth levels of 5 m and 25 m were grouped according to months. They show that the supercooling of water in the Arctic Ocean is quite well-defined throughout the whole year and that it is most pronounced from October to April. The greatest supercooling (0.13°) was observed at the end of February at a depth of 60 m.

In the Beaufort Sea 0.07° supercooling of the water was recorded during the drift of ice island T-3 (Coachman 1966). Coachman also processed bathymetric data for a 10-m water layer, obtained during the drift of the *Maud* along the land side of the eastern Siberian Sea from April to June 1924. He established that the greatest supercoolings (close to 0.1°) were recorded at the surface of the sea from October to December; that is, from the time when low air temperatures began to be observed.

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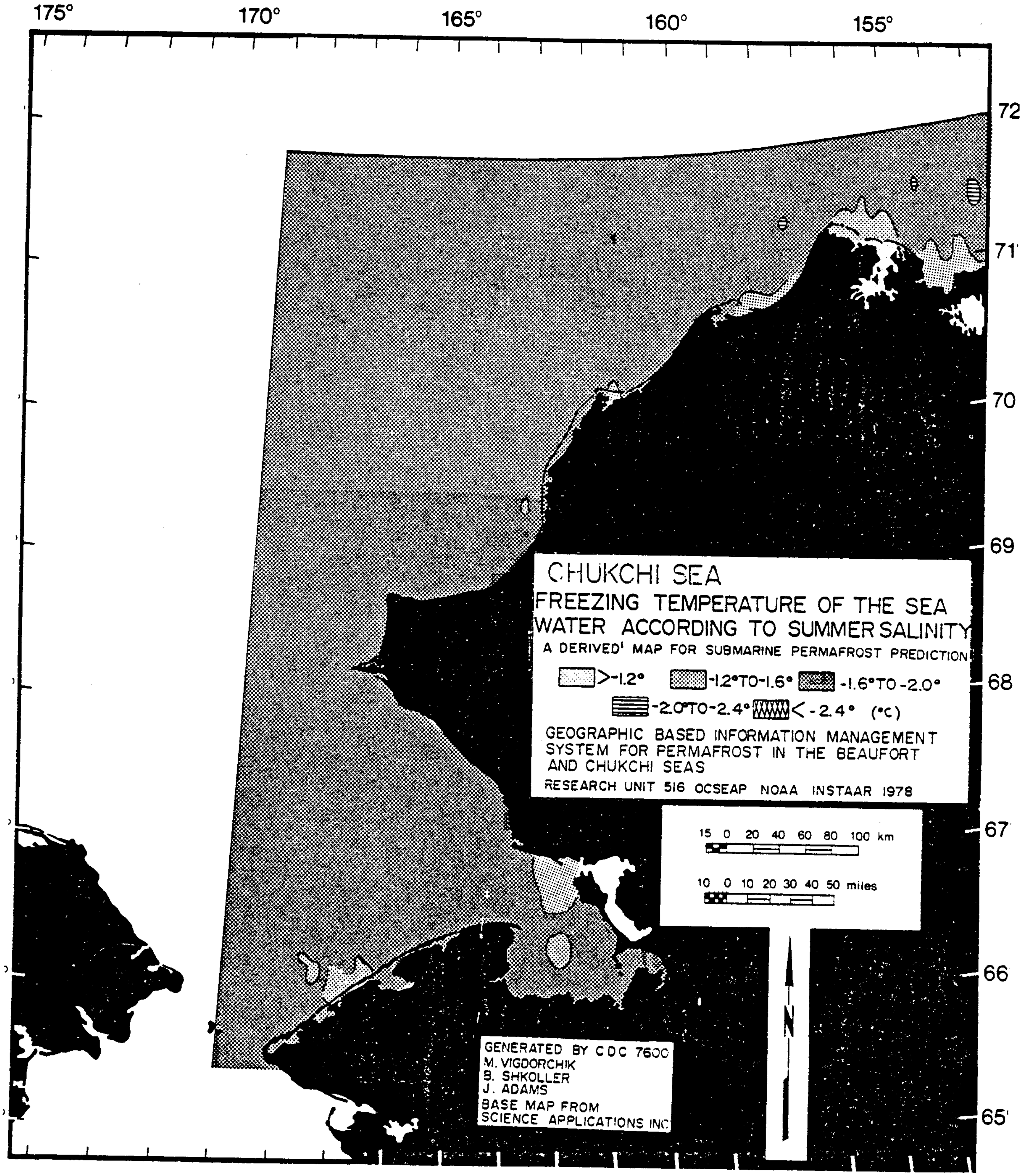
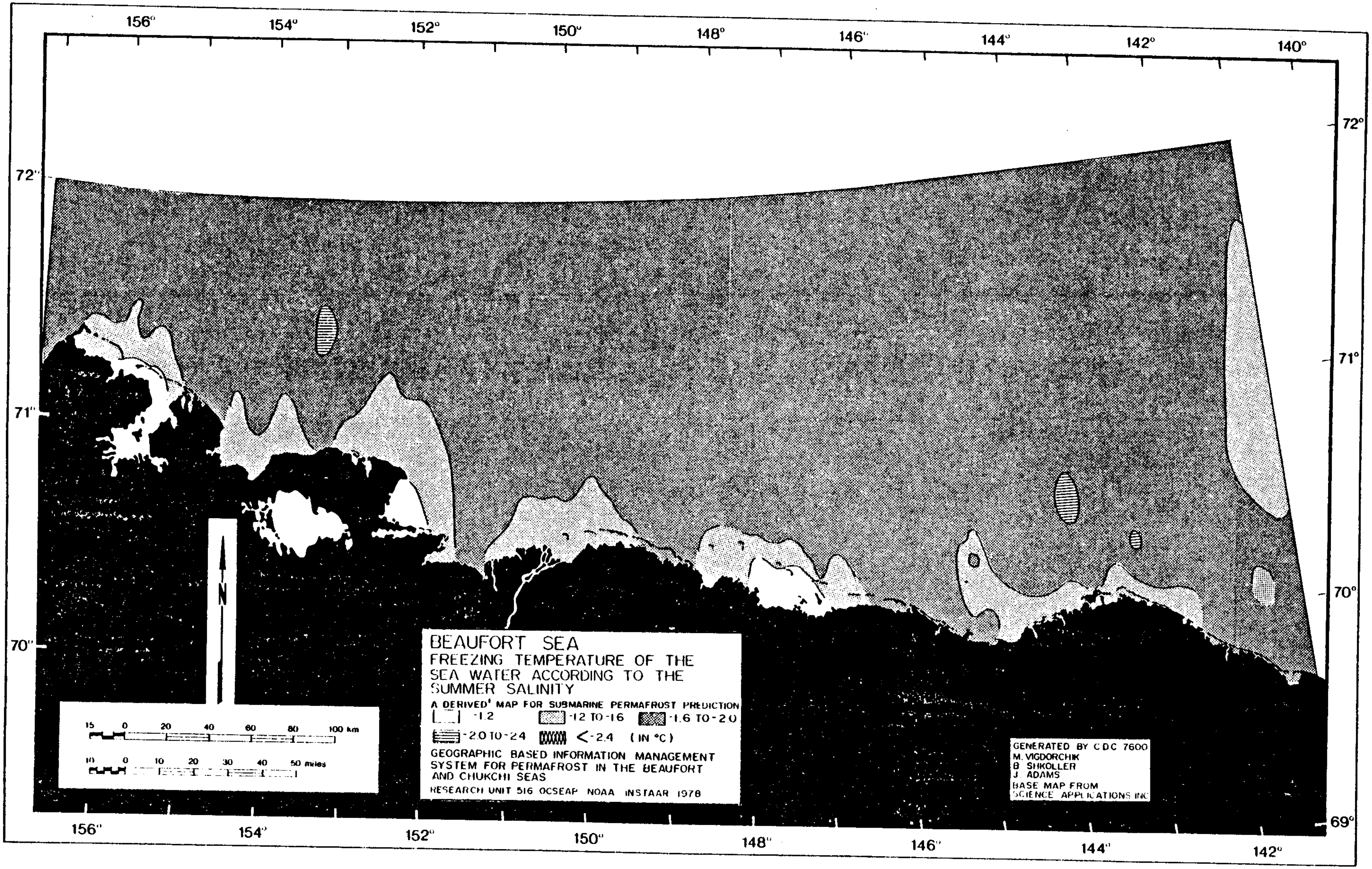
* Savel'ev, B. A. Stroenie, sostav i svoistva ledyanogo pokrova morshikh i presnykh vodoemov (The Structure, Composition, and Properties of Ice Covers of Marine and Freshwater Bodies). Moskva, Izadatel'stvo Mosk. Gos. Universiteta. 1963.

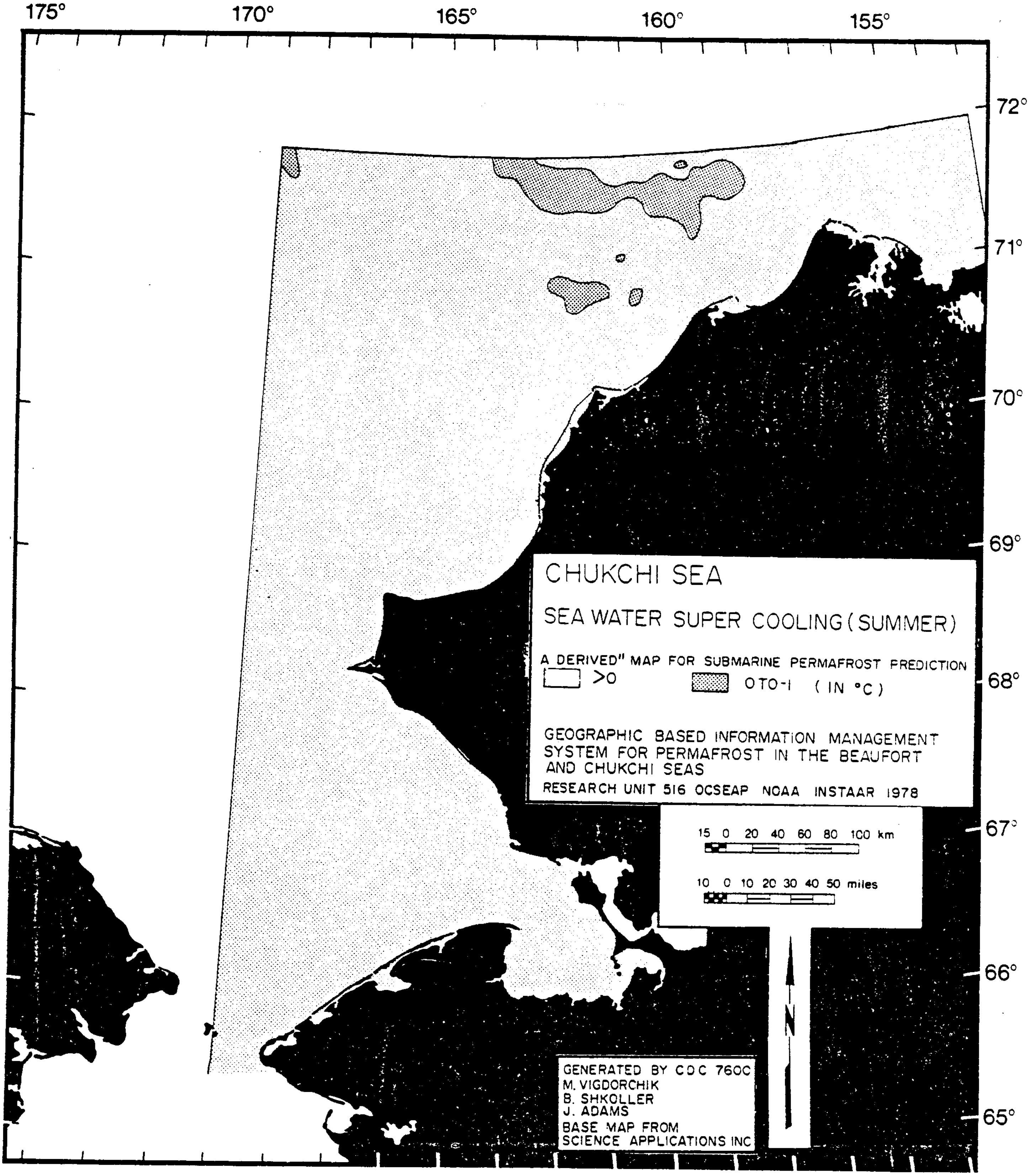
** L. K. Coachman. Production of Supercooled Water during Sea Ice Formation. Contribution number 383, Dept. Oceanography, Univ. Washington, 1966.



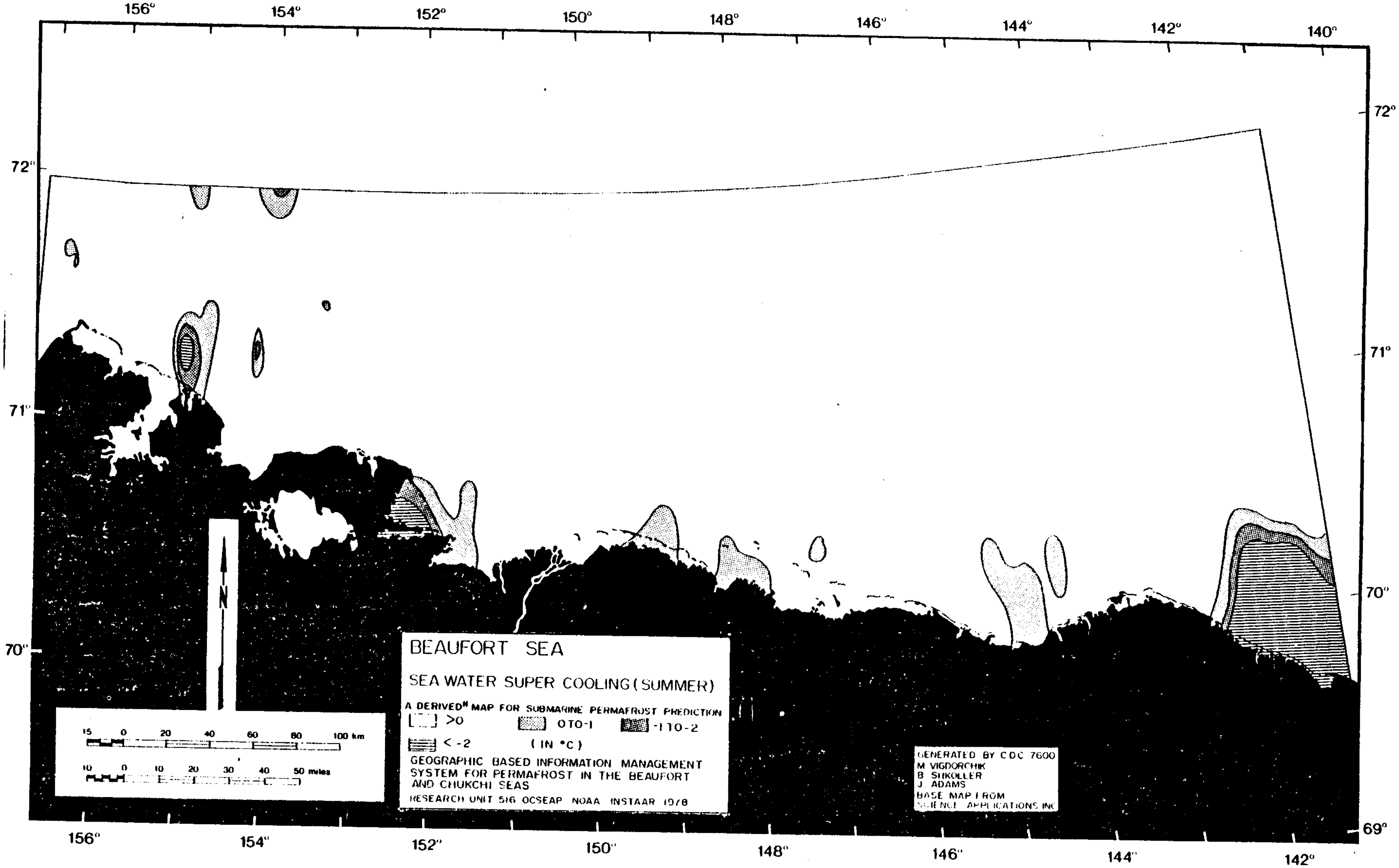
Map 23.—Freezing temperatures of seawater at summer salinities, Chukchi Sea.



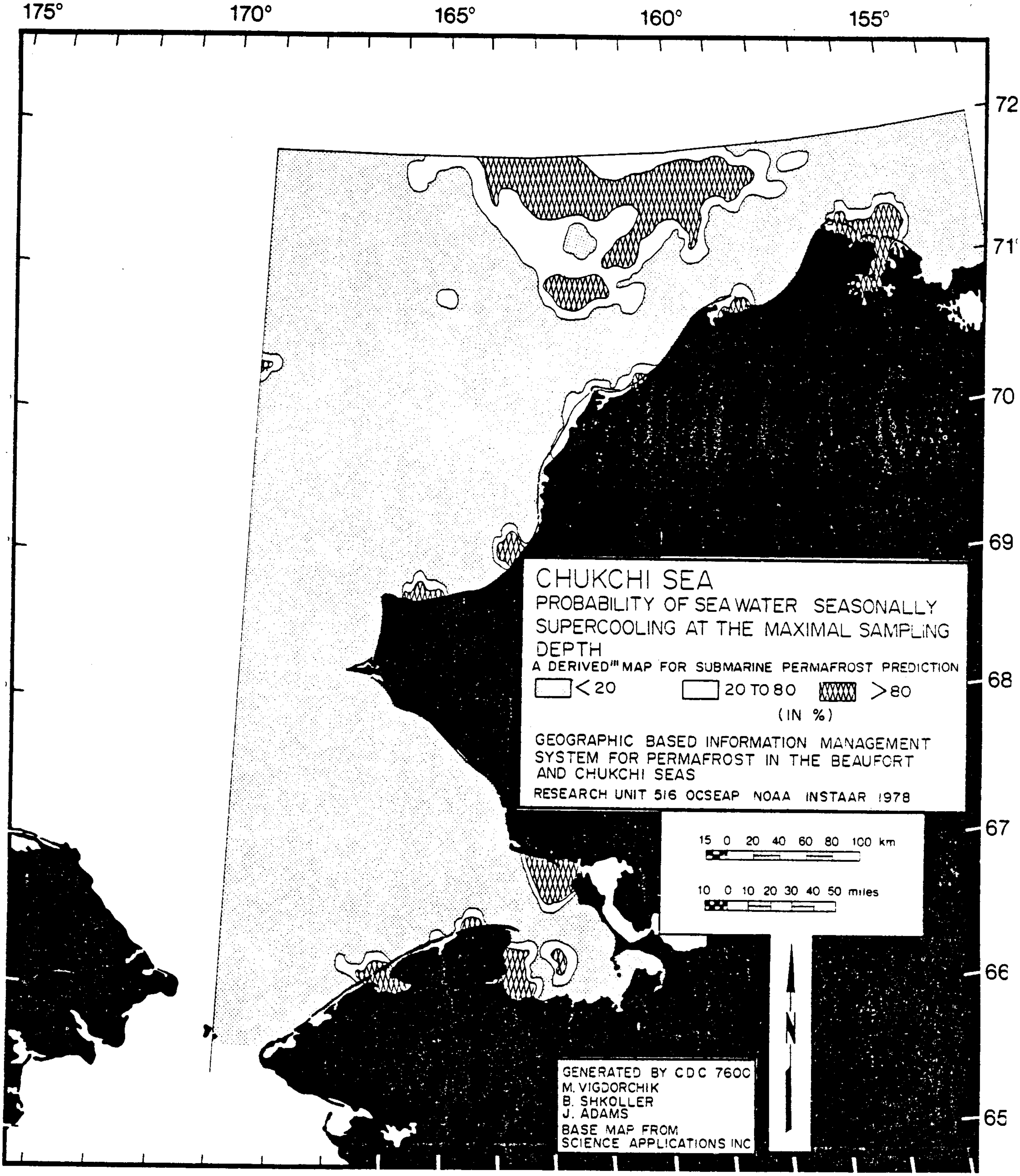
Map 24.—Freezing temperature of seawater at summer salinities, Beaufort Sea.



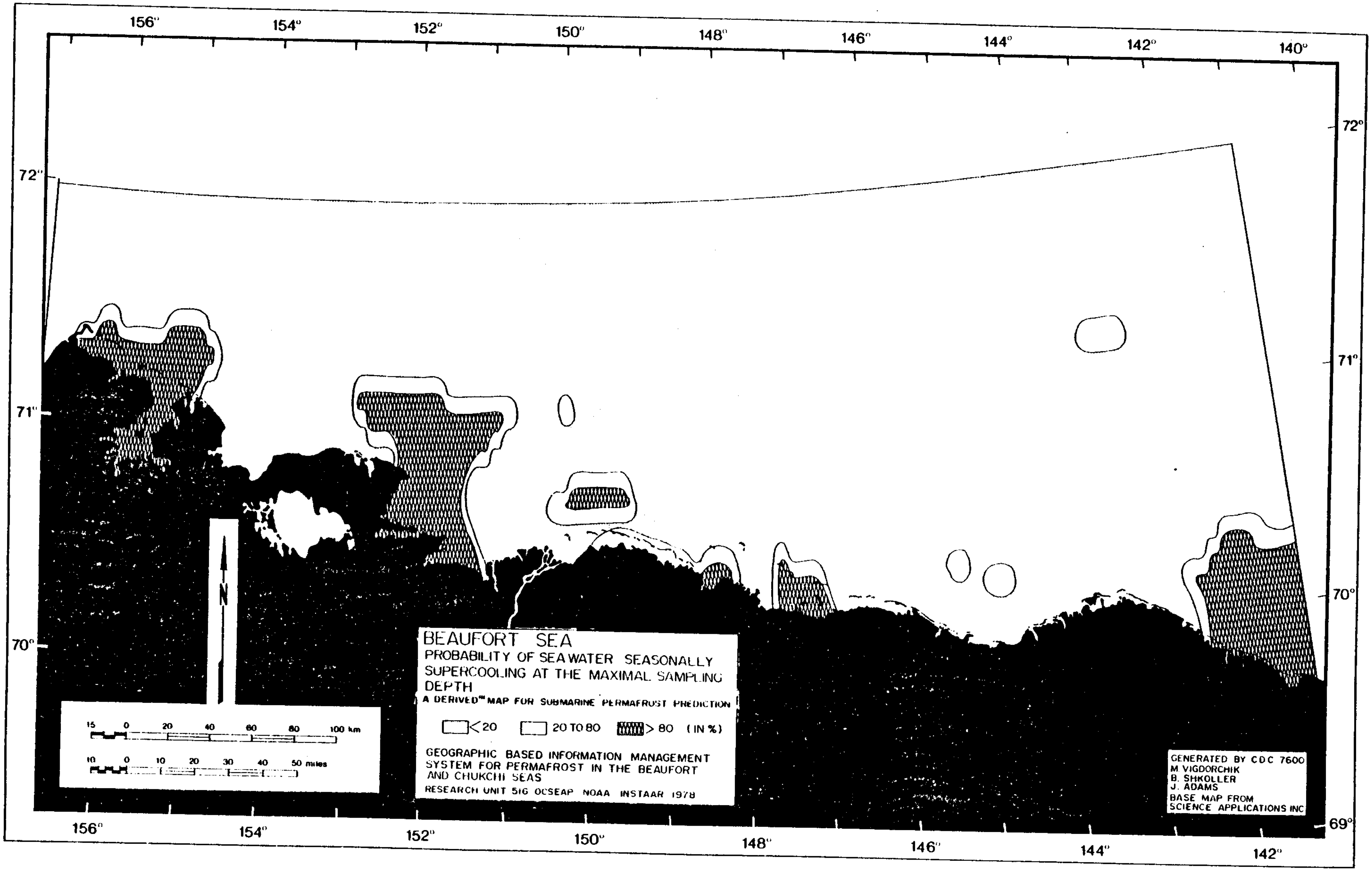
Map 25.—Supercooling of seawater in summer, Chukchi Sea.



Map 26.—Supercooling of seawater in summer, Beaufort Sea.



Map 27.—Probability of seasonal supercooling of seawater, Chukchi Sea.



Map 28.—Probability of seasonal supercooling of seawater, Beaufort Sea.

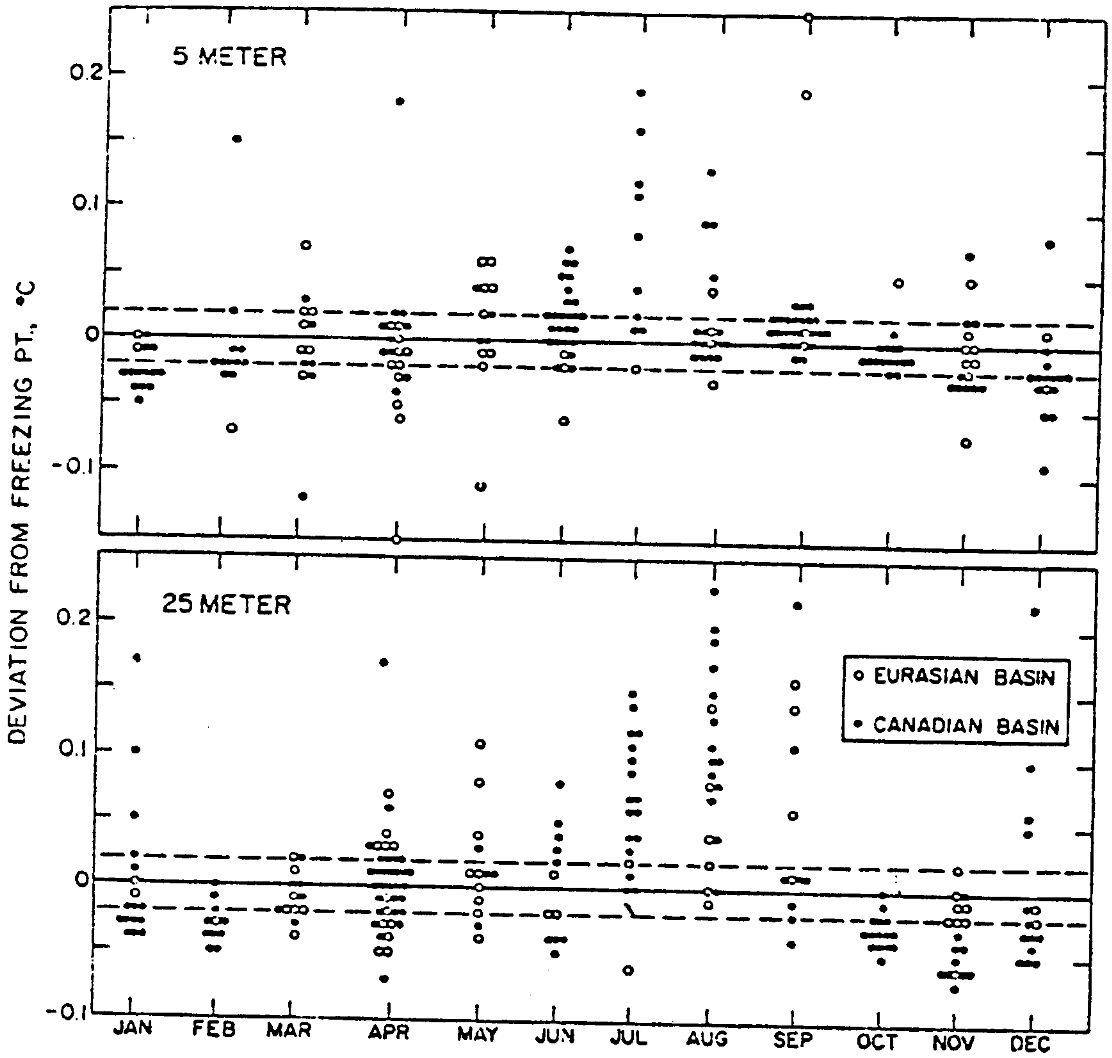


Figure 24.—Temperature deviation from the freezing point at 5–25 m, central Arctic Ocean, grouped by month. See Table 1 for data sources. After Coachman (1966).

In the Soviet Union Chikovskii (1970)*, studying the supercooling of seawater under natural and laboratory conditions, made the following important conclusions:

1. Under natural conditions, the supercooling of seawater depends on the distance between the observation point and the source of the supercooled water (on the open water stretches and the velocity of the currents dispersing the supercooled water).

* S. S. Chikovskii. Studies on ice physics and ice engineerng. Arkticheskii i Antrarkticheskii Instit., Trudy, Vol. 300, 1970, p. 132.

2. In the Arctic Ocean, mainly because of the warm Atlantic waters, supercooling is observed primarily in the upper 50 m and only in exceptional cases at some deeper layers.
3. The supercooling of a water mass located in the watery depths of the sea, the temperatures of which are close to the freezing point, can apparently endure for a long time. This is because of the lack of any direct contact with an ice surface or with a bedrock surface, the presence of which would either sharply curtail or completely eliminate the supercooling of the water.
4. Under natural conditions, seawater contains considerably smaller amounts of different impurities in comparison with fresh water, and thus it would be supercooled to a greater degree. This result is verified by observations in nature.

However, in the calculations and diagrams, a temperature correction for the water pressure is, as a rule, not introduced, which distorts the results somewhat. But for the shelf of the shallow Beaufort and Chukchi seas this correction is not of great importance.

According to these data and ideas we have tried to generate the map of seawater seasonal supercooling probability in the following way. We first define a rectangular network with N points in WEST/EAST direction and M points in SOUTH/NORTH direction, so:

$i = 1, 2, \ldots\ldots\ldots N$ (indexes of horizontal coordinates

$j = 1, 2, \ldots\ldots\ldots M$ (indexes of vertical coordinates)

At each point, we define

Dij – sampling depth

Bij – bottom depth

Tij – temperature at depth Dij

SiJ – salinity at depth Dij

Fij – freezing temperature, computed by use of Sij

Cij – supercooling, which is Cij = Tij – Fij

Pij – probability of seasonal supercooling, which is

$$Pij = 100\% \text{ if } Dij \leq 2 \text{ m}, 0\% \text{ if } Dij \geq 60 \text{ m}, 100\% \text{ if } Cij \leq 0°C, 0\% \text{ if } Cij \geq 0.2°C$$

Otherwise: $100 - 100\ Cij/0.2 = 100 - 500 \cdot Cij$

Following Coachman (1966) and Chikovskii (1970), we have found that usually Cij ≤ 0.2°C and becomes negative if supercooled. Therefore, we simply perform linear interpolation between maximum positive value of 0°C, for if it is negative, then it is already supercooled. We are seeking the conditions favorable for cooling of the bottom floor and bottom deposits. That is why we are interested more in the supercooling conditions at the bottom level, rather than the maximal sampling depth. Usually extensive particles near the bottom increase the possibility of supercooling. The increase of the *sampling depth proximity to the bottom* also increase this possibility. Thus, taking this characteristic into account, we may do the following:

if $Eij = Dij/Bij$ (proximity in %), then $Pij^* = Pij \cdot Eij$, in %

where Pij^* is the probability decreased according to the proximity data.

Composite Mapping Algorithm

The Composite Mapping Algorithm is used for combining a number of individual maps in order to produce a single map showing the combined influence of each of the individual maps relative to a particular evaluation. In our case it will be the candidate area map showing the suitability of the environmental data for the ice-bonded permafrost in the upper layers of the seabed (in %). It is understandable that the environmental source data today provide us only with the information related to the upper layers of the sea bottom. In addition to the notations mentioned above, the following are used:

- R – General reliability of data, in %
- R_1 – Reliability of the grain size data, in %
- R_2 – Reliability of the seawater data, in %
- r_1 – Distance between grid point and the nearest observational point (grain size data), in miles
- r_2 – Distance between grid point and the nearest observational point (seawater data), in miles
- Δx – Horizontal distance (W–E) between adjacent grid points changing from latitude to latitude, in miles
- Δy – Vertical distance (S–N) between two adjacent grid points, in miles
- G – Percentage of silt-sand classes appropriate for active ice segregation
- A – Probability of ice-bonded permafrost distribution, in %
- W – Suitability for ice-bonded permafrost in the upper layers of the seabed

The algorithm to find W is described by

$$W = A \cdot R,$$

$$A = Pij^* \cdot G \quad \text{and}$$

$$R = R_1 R_2$$

$$R_1 = 100\% \, (1 - (r_1/5\Delta y))$$

$$R_2 = 100\% \, (1 - (r_2/\Delta y))$$

It means: $W - Pij^* \cdot G R_1 R_2$, or $W = (Pij \cdot Dij \cdot G \cdot R_1 R_2)/Bij$ (in %)

Candidate Areas for Submarine Permafrost Related to BLM Lease Nomination

The candidate areas for ice-bonded submarine permafrost in the upper layers of the seabed in the Chukchi and Beaufort seas (Maps 29 and 30) might be characterized the following way:

On the Beaufort Sea shelf the probability of ice-bonded permafrost is not very high (20%). At the western part of the shelf there are three major areas of higher probability. Area 1, about 60 square miles, is situated in the northern part of Cape Simpson. The probability here is 20–40%. Area 2, about 2,100 square miles, is situated at the northeastern part of Harrison Bay and at the open sea to 50 miles from the shore. The probability here is 20–40% and 40–60%. Area 3, about 350 square miles, is situated to the N-NE from the Colville River delta, close to the northern edges of the barrier islands and far to the north. The probability here reaches 60–80%. The small area with probability of 20–40% could be defined to the north from Area 3.

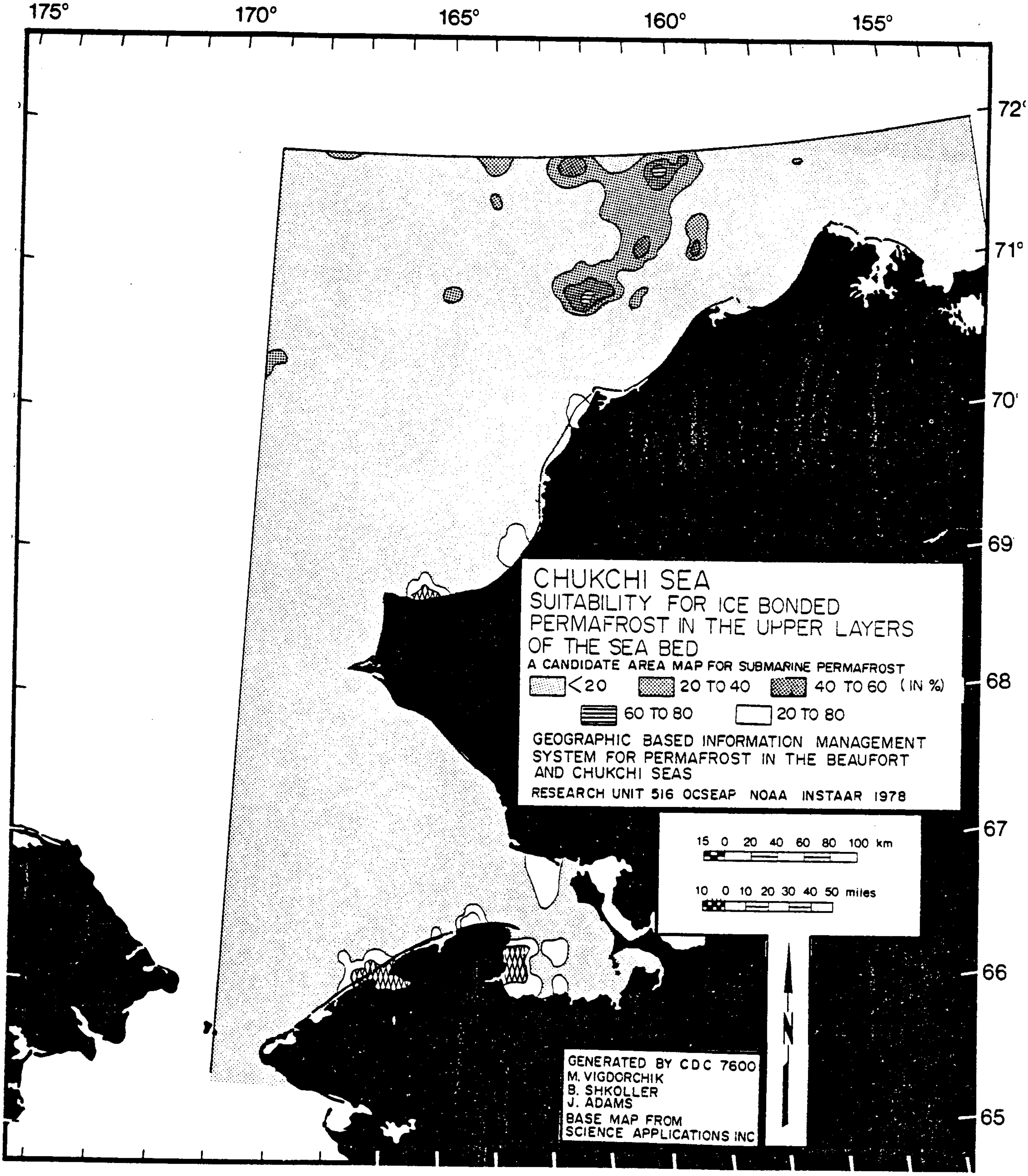
In the central part of the Beaufort Sea shelf the prospective permafrost areas are situated much closer to the shore mostly between the shore and barrier islands, or at the shallow areas a little north from the islands. Area 4 is such an area, about 780 square miles, with the probability in its inner part (closer to the shoreline) about 60–80%. Two very small spots with probability more than 20% are defined at Camden Bay (Area 5). Table 7 characterizes the level of probability for submarine ice-bonded permafrost in the upper layers of the seabed (limited to 5 m) for each unit of the lease areas.

We also need to specify the suitability for submarine relic permafrost within the boundaries of the same three areas of the Beaufort Sea (Map 31).

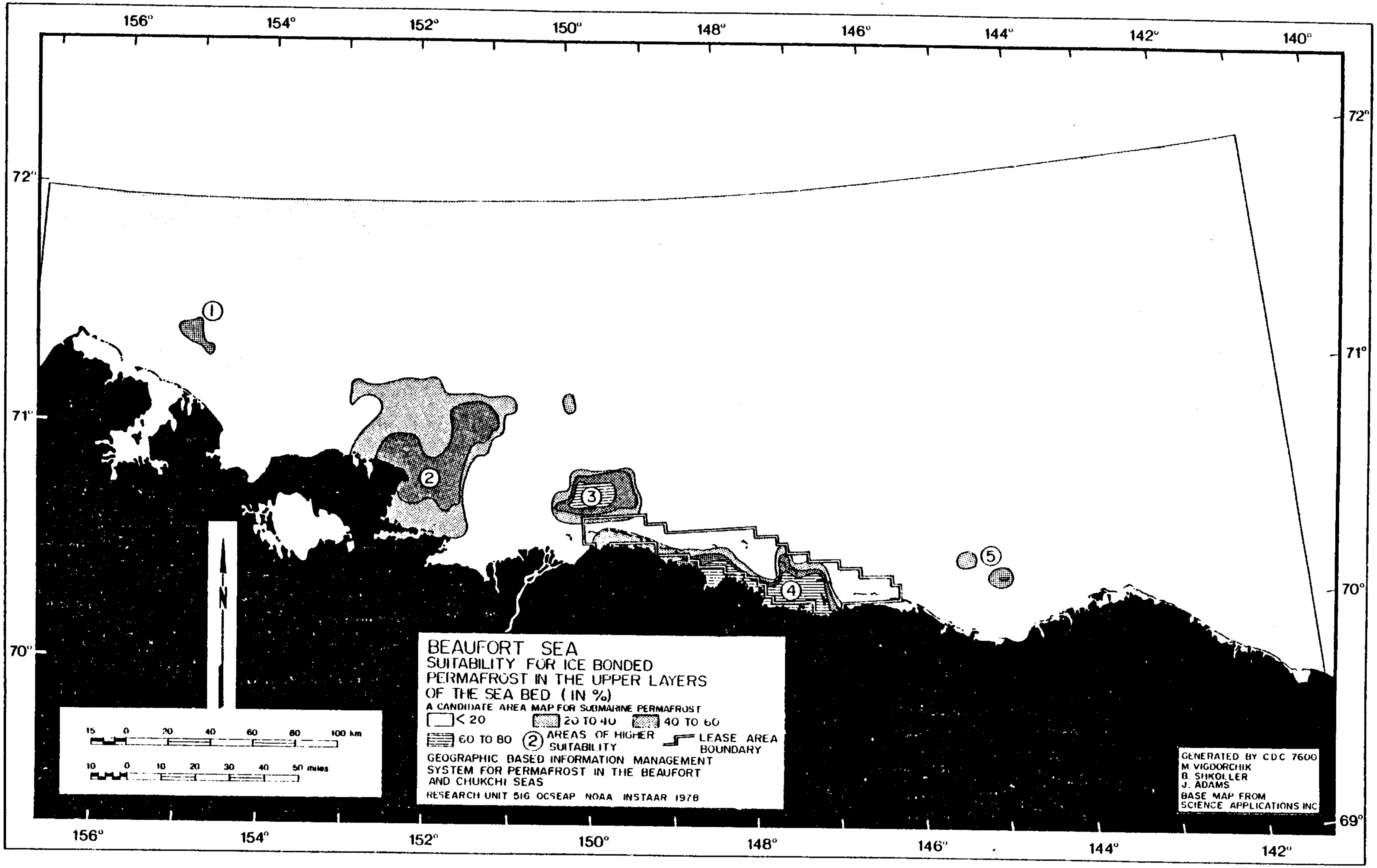
The western part of the Beaufort Sea can be considered as suitable for relic permafrost. Average thickness could reach 100–120 m. Permafrost might be met at the first 50–100 m from the seabed where the seawater depth is 0–60 ± 20 m. "Taliks" are typical for such kinds of relic permafrost extension according to studies of the Eurasiatic shelf. The relic permafrost body could be separated from the upper ice-bonded permafrost spots on the seabed, by taliks, and this connection in each case could be clear only from site specific investigations.

The second, or central, area of the Beaufort Sea can be characterized by low potential for relic permafrost, possible only at the boundary with the western area. A body of deep permafrost could be a continuation of a thick (about 600 m) coastal cryogenic zone and limited to 18–22 km offshore. General thickness would be about 150–200 m; seaward this is less—30–50 m. Taliks are also typical for such submarine permafrost and sometimes give the impression of a multilayered structure of permafrost.

No relic permafrost can be considered to exist in the third, or eastern, part of the sea. The coastal permafrost could not extend farther than 2–5 km offshore. Permafrost depth here could be 50–100 m. Lack of data on coastal permafrost thickness prohibits speculation on



Map 29.—Suitability for ice-bonded permafrost at the upper layers of the seabed, Chukchi Sea.



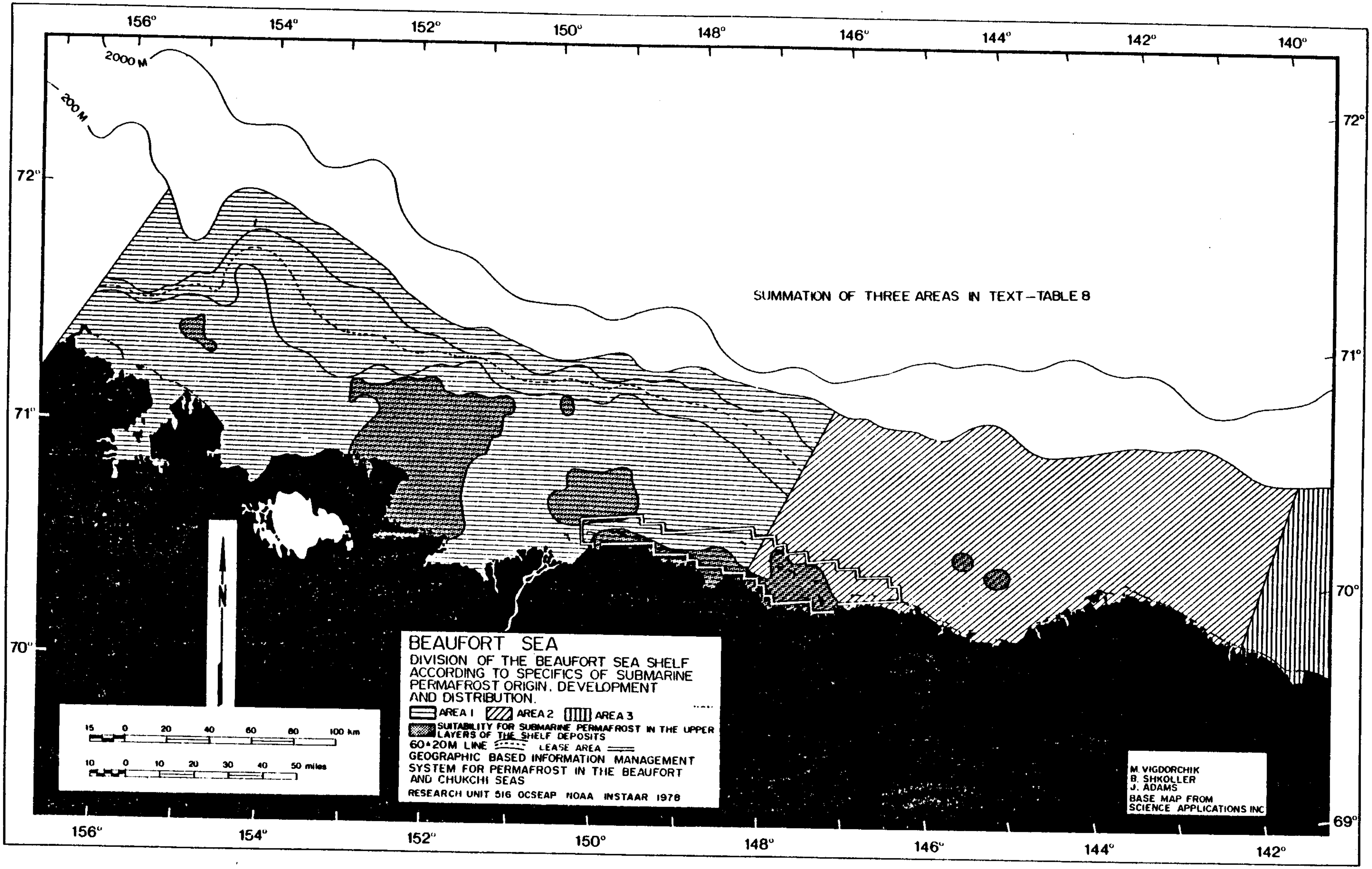
Map 30.—Suitability for ice-bonded permafrost at the upper layers of the seabed, Beaufort Sea.

Table 7.—Probability of submarine permafrost in the lease areas (by lease unit numbers).

<20%	20–40%	40–60%	60–80%
1–22	23–29	95–101	57
30	32–40	110	58
31	59–61	114–116	111–113
41–56	77–90	129–131	126–128
62–76	102	135	139
91–94	103	140	158
104–108	109	153	171–175
119–121	117	157	182–189
125	118	145–148	201–208
132	122–124	159–162	220–221
133	130	168–170	233
143	131	176	234
151	134	190	
152	136–138	209	
163–167	141	222	
178–181	142	235	
192–200	144		
211–219	149		
224–232	150		
	154–156		
	177		
	191		
	210		
	223		
	236		

the thickness of the permafrost offshore. Depending on the coastal permafrost thickness, it must be 2–3 times less. Seasonally frozen layers might be met on any part of the Beaufort Sea coast within the limits of the sea ice–sea bottom interaction (isobath ~2 m).

In the Chukchi Sea the areas with relatively high potential for the development of ice-bonded permafrost in the upper layers of the shelf deposits are situated to the west-northwest from the Barrow Canyon and at the east-southeastern slope of Hanna Shoal. Some relatively small spots could be traced along the coast close to Icy Cape, Cape Beaufort, Cape Lisburne, between Cape Thompson and Kivalina at Kotzebue Sound and Shishmaref (Map 30). There are no data available on the relic ice-bonded permafrost in these areas and the relationship between the prospective areas for ice-bonded permafrost at the surface of the seabed and any deeper permafrost body if it exists. Generally speaking, the influence of the relatively warm water of the Bering Sea, lack of thickness of the permafrost in the coastal zone (<300 m), its discontinuous character (Péwé 1976), and possibly a higher thermal flux



Map 31.—Division of the Beaufort Sea shelf according to the specifics of submarine permafrost origin and distribution.

in the zone of convective heat transfer in the southern part of the Chukchi Sea—all these factors are negative for preservation of relic ice-bonded permafrost, if it ever existed. The deep submarine permafrost could exist only extremely close to the coast (usually less than 1.5–2.0 km) and might be connected with the coastal permafrost body or disconnected (lenses). In both cases the thickness of the submarine permafrost could reach nearly 20–30 m. The spots suitable for ice-bonded permafrost in the upper layers of the shelf deposits might represent the area where the intrusion of the coastal permafrost body into the shelf takes place. To answer the question special investigations need to be done.

APPLICATION OF THE SYSTEM TO THE BEAUFORT SEA AREAS WITH KNOWN SUBMARINE PERMAFROST (COMPARISON OF RESULTS)

An opportunity to check the efficiency of our system for permafrost prediction occurred when we compared our maps (computerized variations) with maps made for the Canadian shelf. We will therefore show here the following maps of the Mackenzie Bay and Delta and adjacent areas (after Hunter et al. 1976): bathymetry (Figure 25); summer and winter bottom water temperature in the southern Beaufort Sea (Figure 26); summer and winter bottom water salinities in the southern Beaufort Sea (Figure 27); and an interpretation of the occurrence of subsea bottom ice-bonded permafrost from industry seismic records (Figure 28). Then we compare them with the following maps of our system ("seawater block") made for the Canadian part of the Beaufort Sea according to existing data at the same scale and projection: summer temperature of the seawater at the maximal sampling depth (Figure 29); summer salinity of the seawater at the maximal sampling depth (Figure 30); freezing temperature of seawater according to the summer salinity (Figure 31); seawater supercooling during the summer at the maximal sampling depth (Figure 32); probability of the seawater seasonally supercooling at the maximum sampling depth (Figure 33).

In general, the summer seawater temperature distribution on the Canadian map and our map look similar. The main areas with positive or negative seawater temperatures are the same. The differences may be observed in the size of the northeastern positive spot, which is oriented more to the west on the Canadian map, and also in the discontinuous character of the positive spots in the middle and upper parts of the territory. The temperature contours on the Canadian map are discontinuous also in the western areas. Following the same trend, the summer temperature contours on the computerized map show considerable extension of negative temperature zones during the summer too. Both maps of salinity are also very close but the computerized map shows much more detail, especially for the northern part of the shelf and to the north from the Mackenzie River delta. According to the freezing temperature of the seawater (calculated by using the summer salinity), the areas of seawater

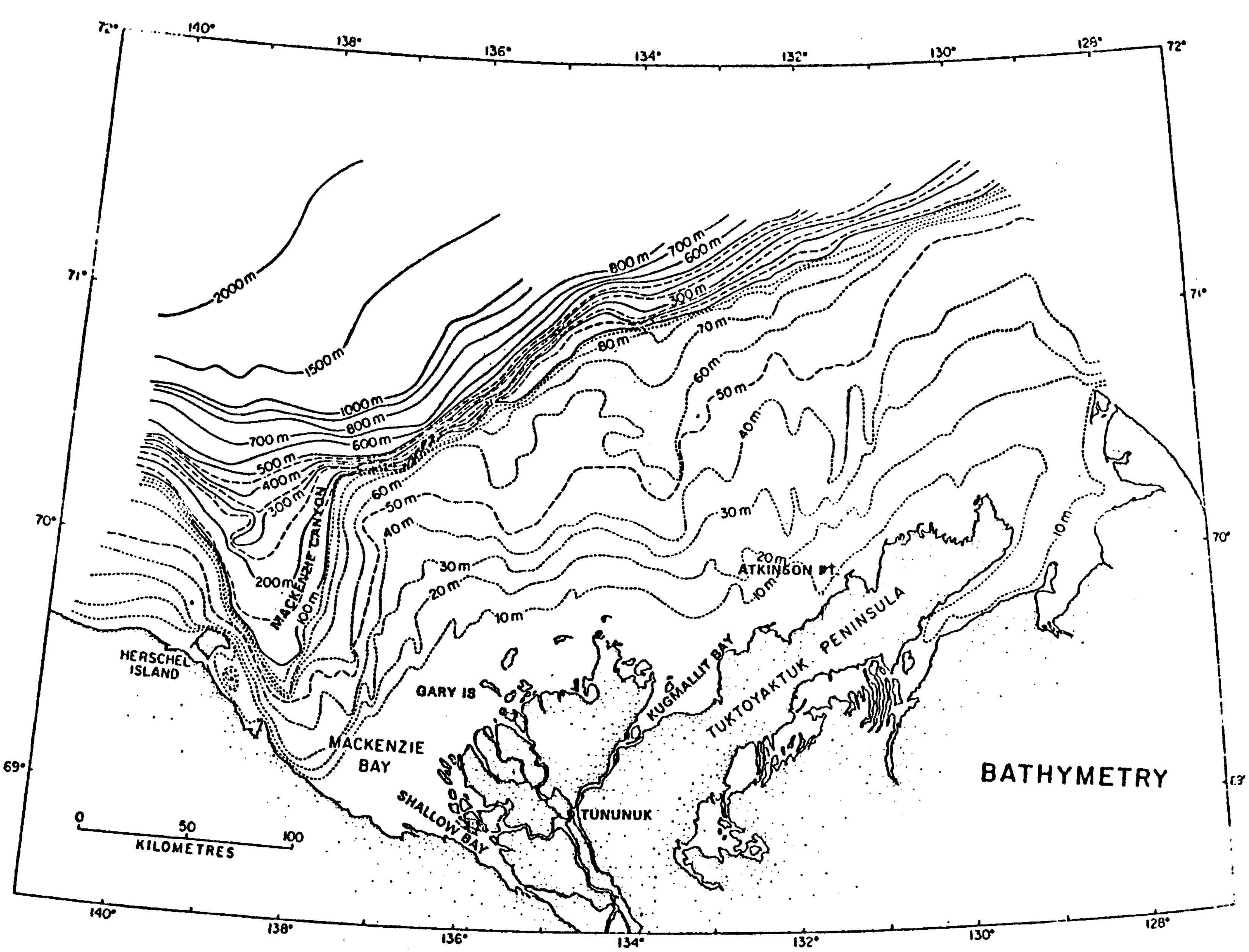


Figure 25.—Bathymetry of the southern Beaufort Sea. After Hunter et al. (1976).

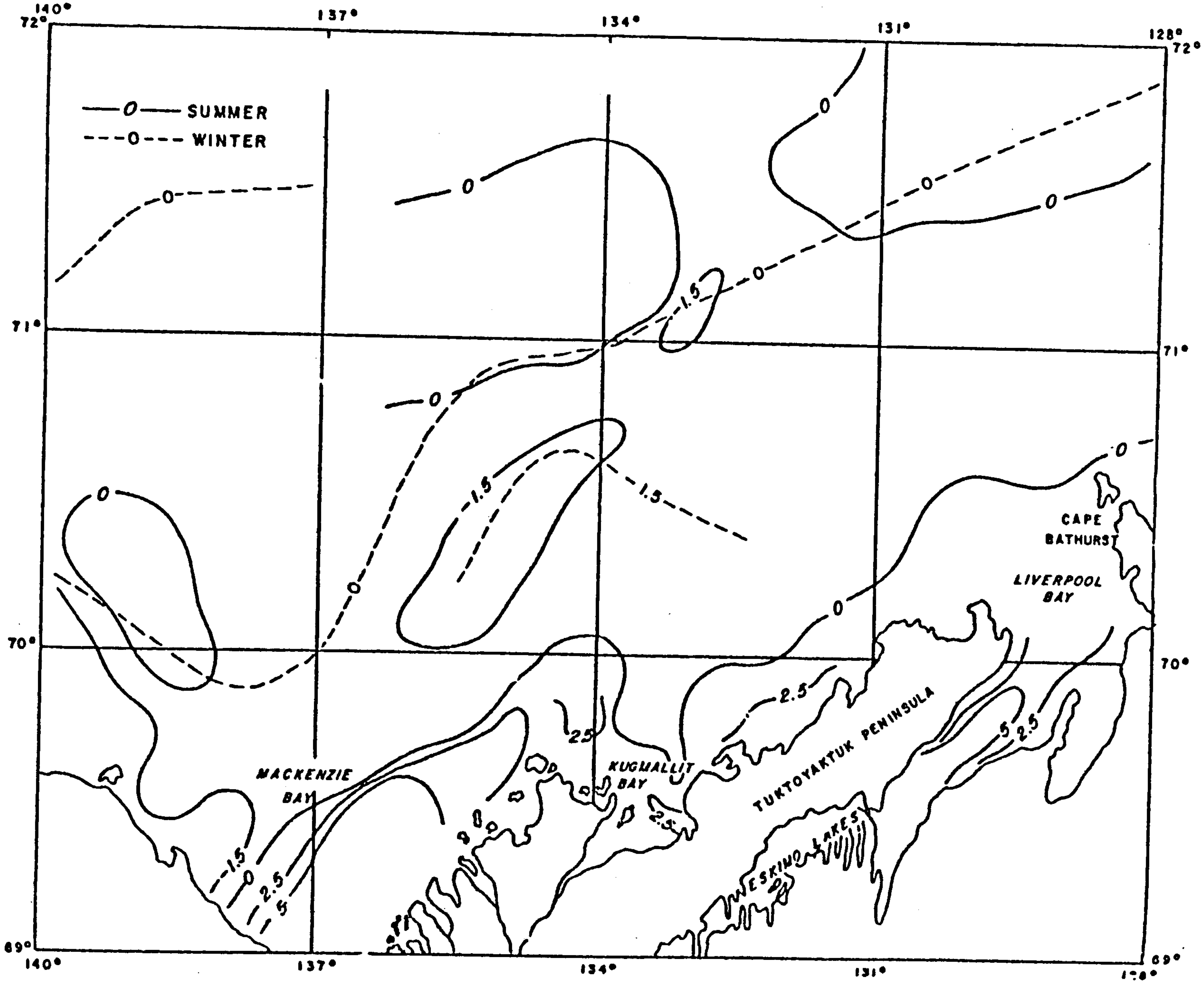


Figure 26.—Summer and winter bottom water temperature in the southern Chukchi Sea. After Hunter et al. (1976).

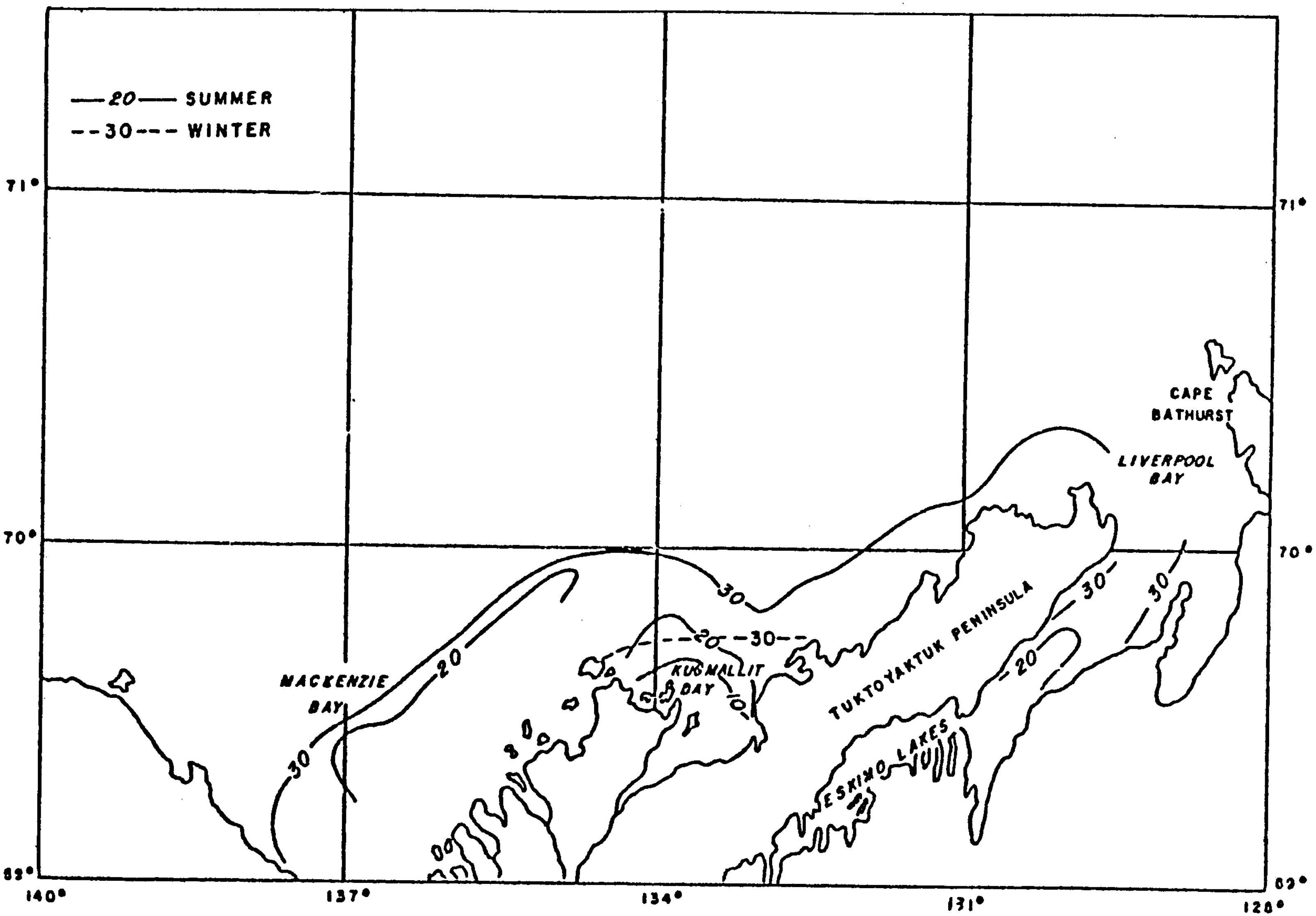


Figure 27.—Summer and winter bottom water salinities in the southern Beaufort Sea. After Hunter et al. (1976).

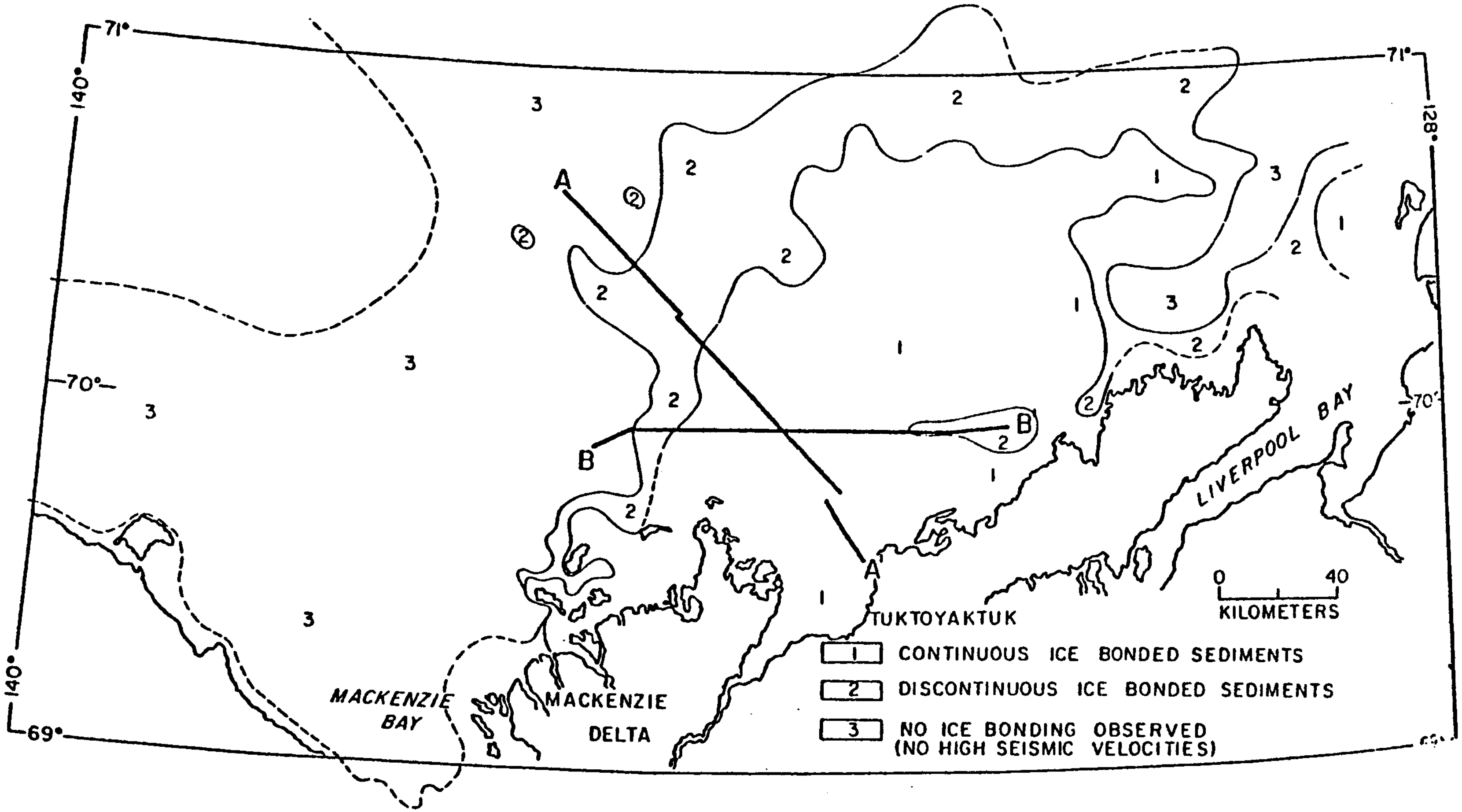


Figure 28.—An interpolation of the occurrence of subsea bottom ice-bonded permafrost. After Hunter et al. (1976).

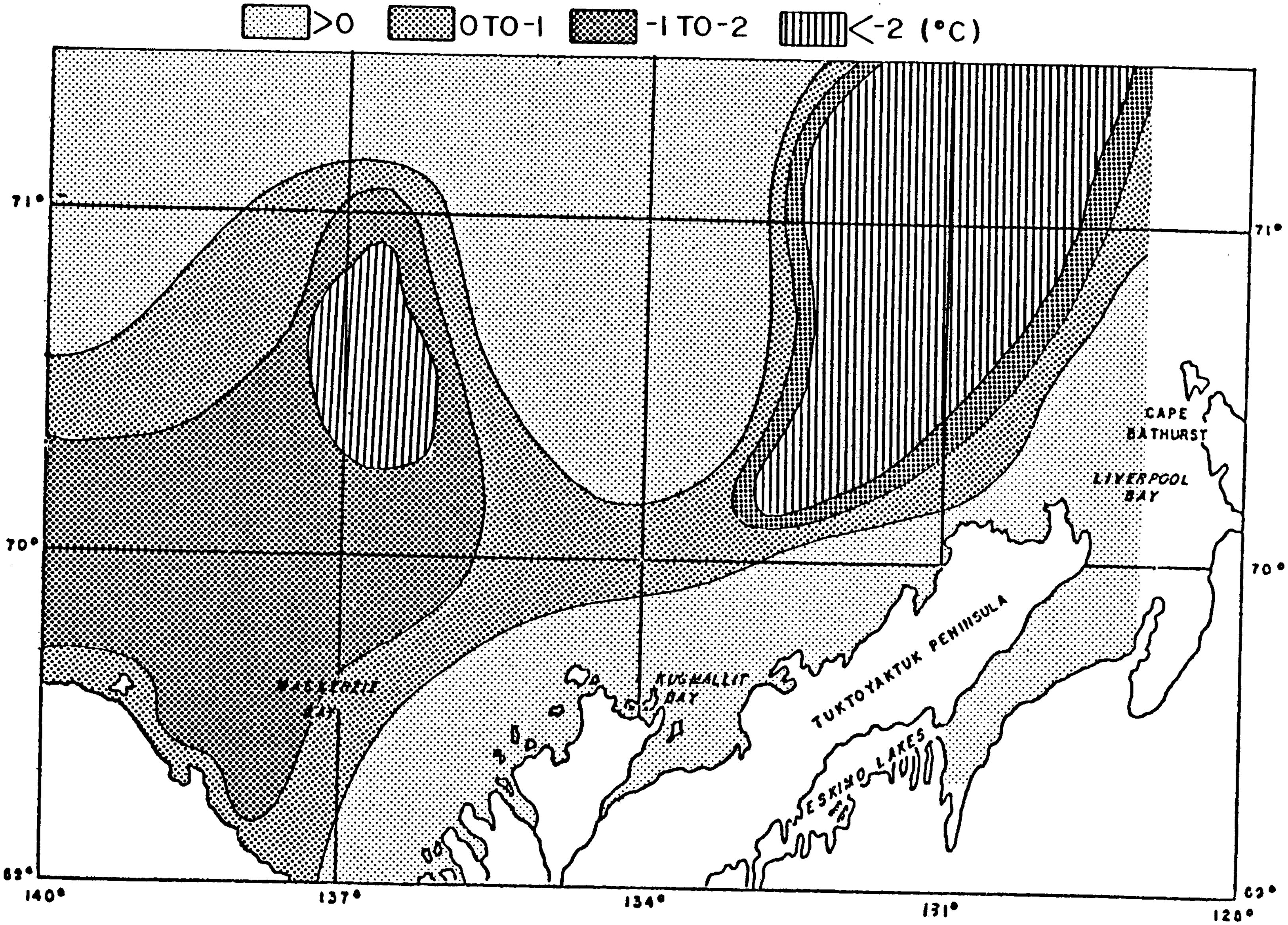


Figure 29.—Summer temperature of seawater at the maximal sampling depth in the southern Beaufort Sea.

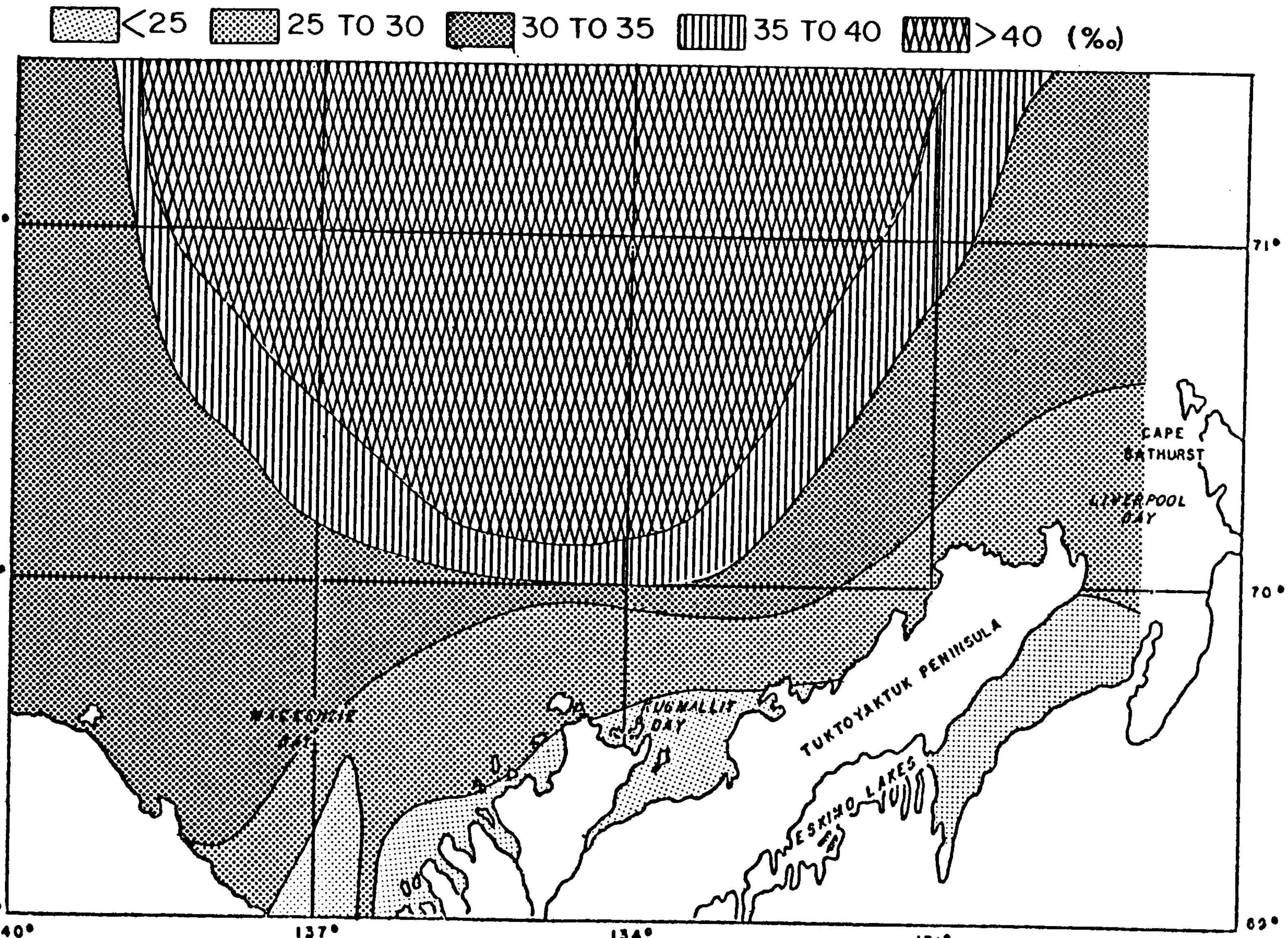


Figure 30.—Summer salinity of seawater at the maximal sampling depth in the southern Beaufort Sea.

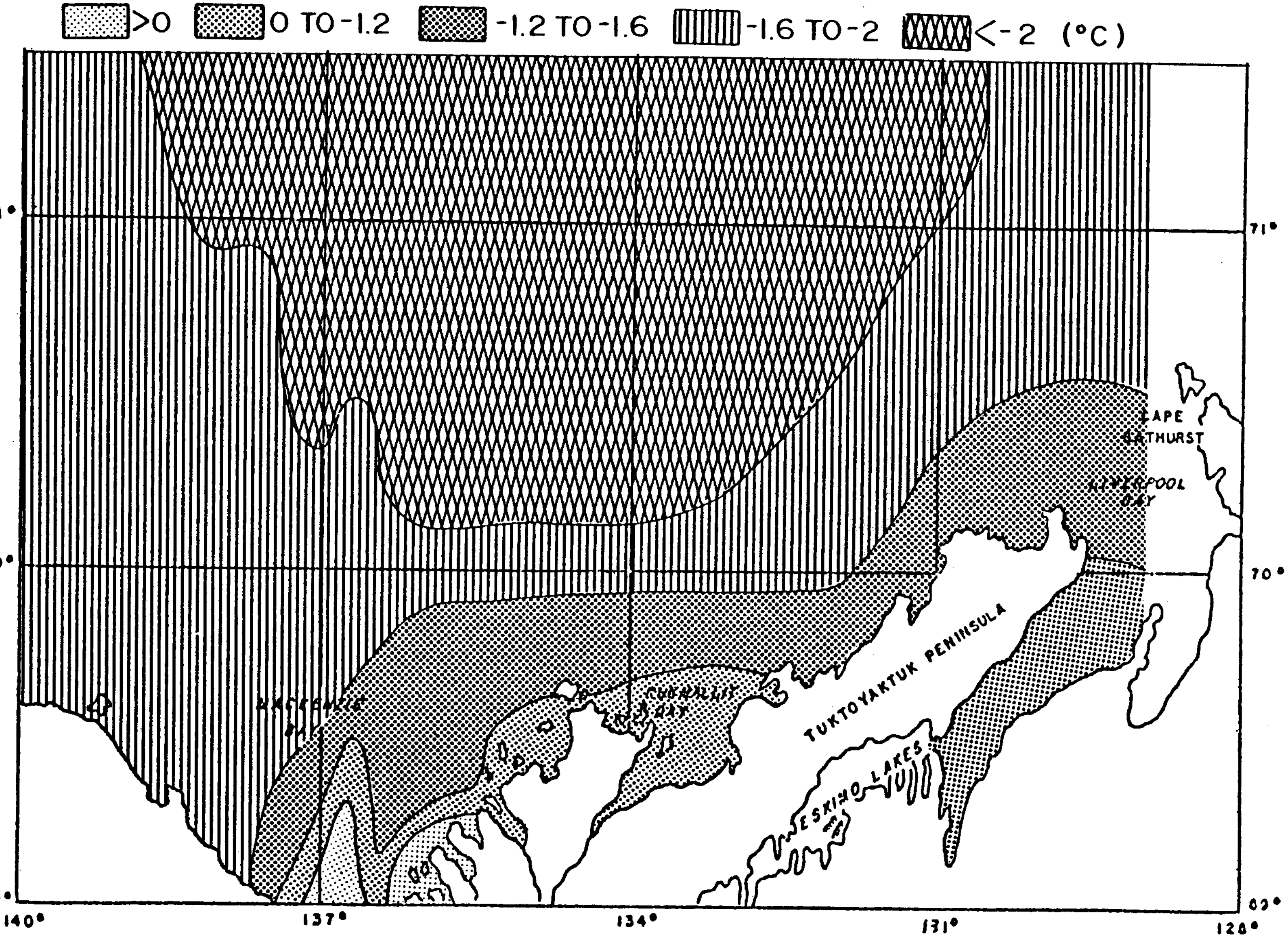


Figure 31.—Freezing temperatures of seawater according to summer salinity in the southern Beaufort Sea.

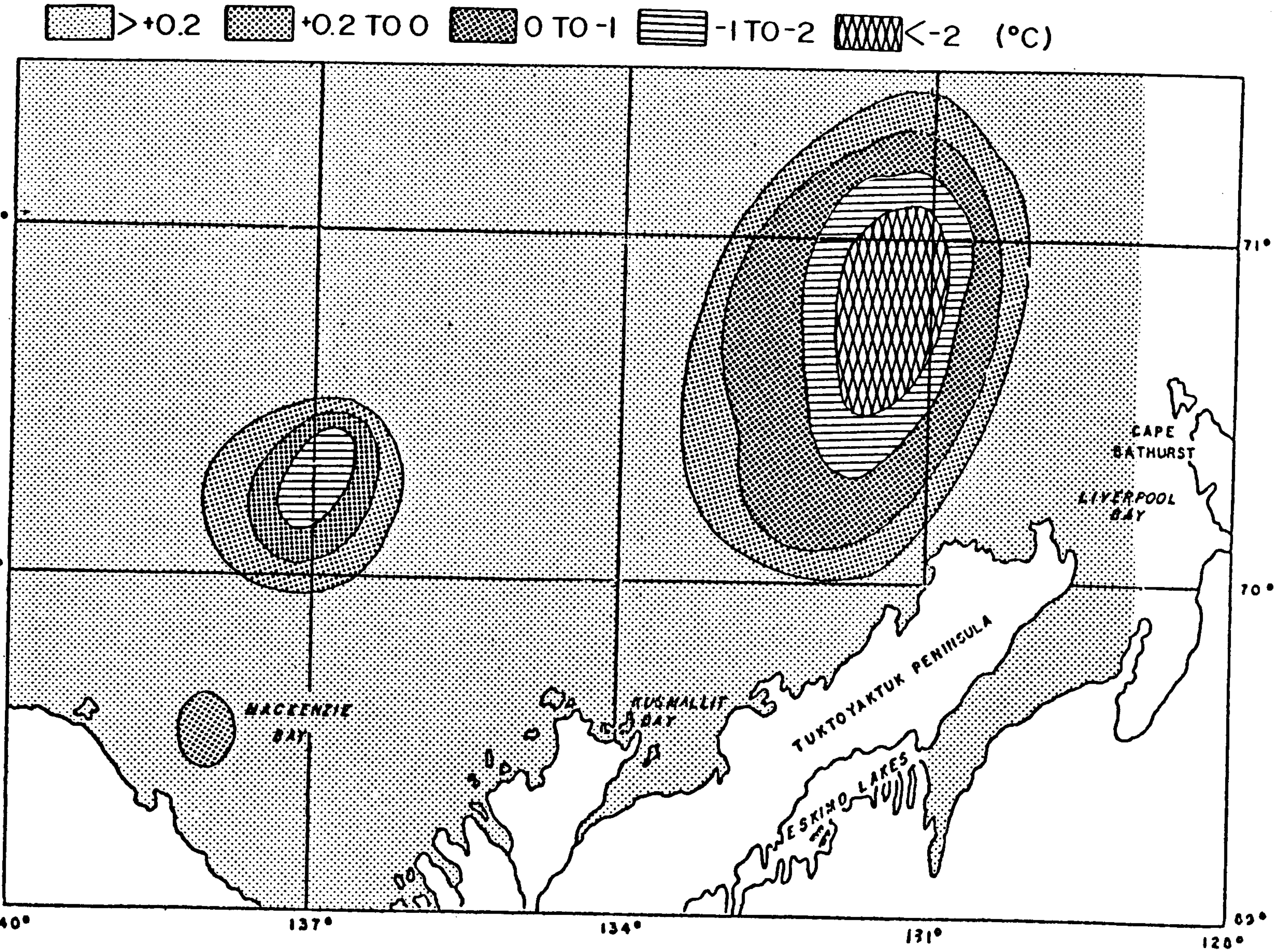


Figure 32.—Seawater supercooling during the summer at the maximal sampling depth in the southern Beaufort Sea.

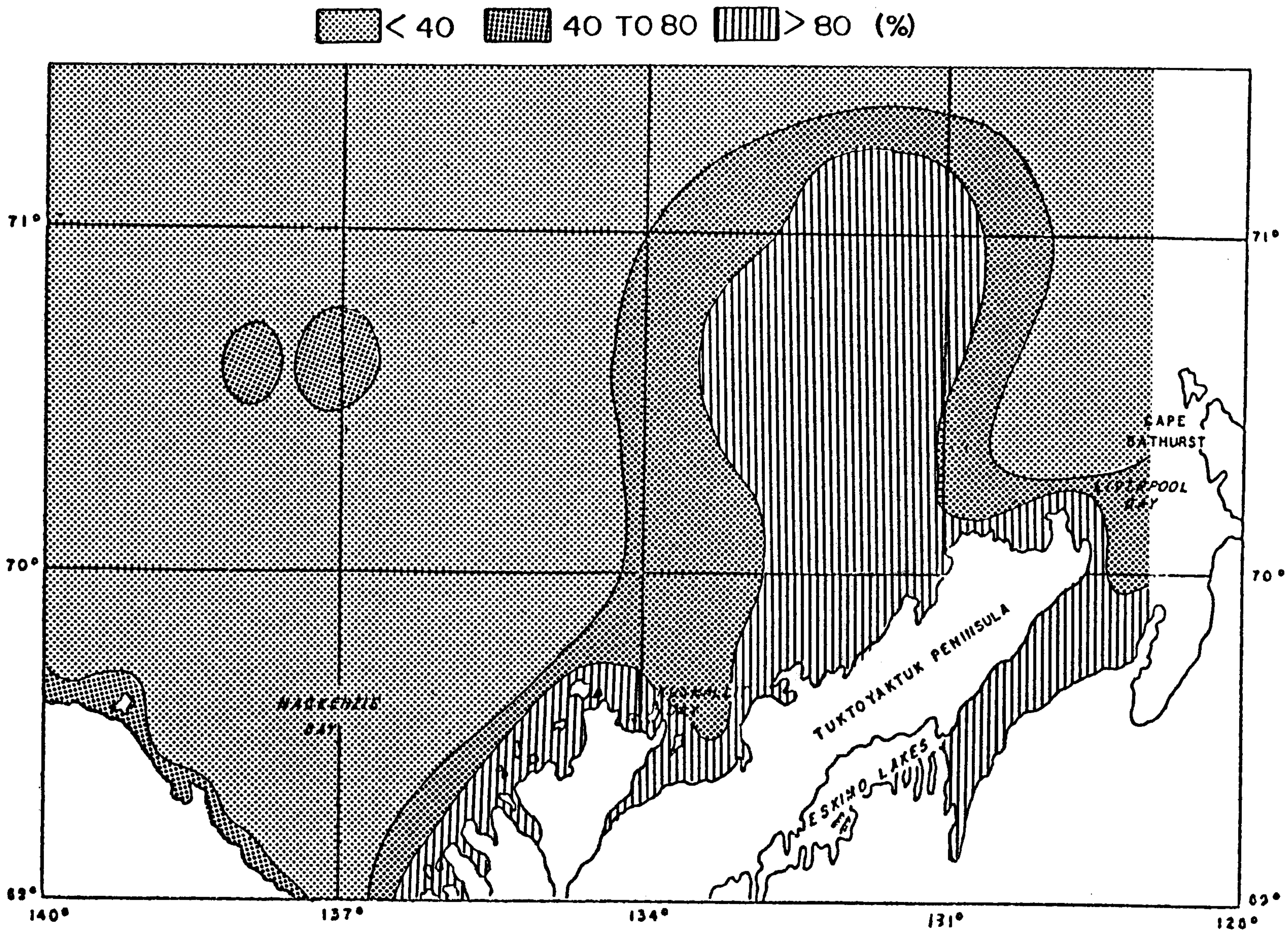


Figure 33.—Probability of seasonal supercooling at the maximal sampling depth in the southern Chukchi Sea.

supercooling during the summer look like three isolated spots on the computerized map. The biggest one is between 70°–71°15′N and 130°30′–133°W. A couple of other spots can be seen in the western part of the territory.

Because of the sparse data the seawater winter temperatures on the Canadian map are shown only by two contours following the general direction of the bathymetry contours.

Most interesting are the results of the comparison of the final map from the Canadian publication and the derived map produced by our system. Hunter et al. (1978) have compiled a map of the subsea bottom ice-bonded permafrost as an interpretation from industry seismic records. The Canadian geologists show continuous ice-bonded permafrost in the area that looks in a bathymetrical and geomorphological sense, like the submarine continuation of the Mackenzie Delta, between 69°30′–71°20′N and 130–135°W. The area of continuous ice-bonded sediments is surrounded by a relatively narrow belt of discontinuous ice-bonded permafrost. The probability map for the seawater seasonally supercooling at the maximal sampling depth gives essentially the most probable areas (80–100%) inside or very close to the zones of continuous ice-bonded sediments at the submarine part of the Mackenzie Delta and 40–80% probability in the limits of discontinuous zones of ice-bonded sediments there. The shape of the territory without ice-bonded permafrost also looks similar in both maps. A small part of the peripheral discontinuous submarine permafrost zone in the center of the territory is indented in an area of low probability for seasonal supercooling (<40%), but along the coast this probability is higher (40–80%).

Our impression is that according to the comparison the areas of seawater seasonally supercooling are good indicators of submarine permafrost, lying not only right under the sea bottom, but to a considerable depth below it and at the areas with bathymetry ranging from 7 to 100 m. The absence of ice-bonded permafrost below the sea bottom when it exists at greater depths could be explained by higher concentrations of salt in the pore water in the sediments forming the upper layers of the seabed.

We see that the results of the comparison are positive. This means that the use of the described system could help in tracing the areas with possible submarine permafrost extension. It is also important that the shelf areas with the seawater under summer and seasonal supercooling have an obvious correlation with the subsea permafrost. Perhaps these spots of supercooling water could be the result of cooling effects of the perennially frozen shelf deposits. In such a case, the spots themselves might be indicators of submarine permafrost.

CONCLUSION

The results of the work are summarized in Table 8, in which the division of the Beaufort Sea shelf, according to the specifics of the submarine permafrost's origin and distribution,

Table 8.—Division of the Beaufort Sea shelf according to specifics of submarine permafrost origin and distribution.

Area	Seawater Depth	Geological Structure	Geomorphological Structure	Extension of the Area Seaward
1	2	3	4	5
Western part of the shelf: Point Barrow–148°	0–60 ± 20 m	Colville Syncline	Continuation of the north-Alaskan flat plain. Probability of the "ancient valleys" development buried & half-buried & thermokarst polygonal–tetragonal forms in the different stages of thawing	70–80 km
Central part of the shelf: 148°–142°	0–200 m	Tectonically active transitional zone, possible boundary of Asian & American plates	Gentle plain on the complicated system of the submeridional, meridional, and latitudinal faults, blocks & dome structures	100 km
Eastern part of the shelf: 142°–Mackenzie River	0–200 m	Anticline	Relatively steep slope of the mountain range	85 km

Table 8.—Continued.

Average Rate of Coastal Retreat	Possible Rate of the Recent Vertical Movement	Possibility of Exposure during Last Glaciation and How Long	Time of Submergence
6	7	8	9
(W) 4.7 m/year	2.2 ± 2.0 mm/year	During last glaciation 25–10,000 years ago	10,000 to present
(C) 1.6 m/year	6.6 ± 2.5 mm/year	Low probability of exposure & only for small blocks at the boundary with the western part of the Beaufort shelf for a short time	>25,000 years
(E) 1 m/year	10.34 ± 3.7 mm/year	No possibility	>25,000 years

Table 8.—Continued.

Main Agents of the Permafrost Development	Suitability for Submarine Relic Permafrost and its General Characteristics	Suitability for Submarine Ice-bonded Permafrost at the Upper Layers of the Seabed
10	11	12
(W) Thickness of the permafrost at the coast is relatively moderate (average 350 m). Exposure of shelf was the main case of the submarine permafrost development. Ice at depth of ≤2 m.	Suitable. Average thickness 100–120 m. Permafrost could be met at the limits of depth first 50–100 m from the seabed. Taliks. Seawater depth ≤60 ± 20 m.	There are several spots with higher probability for ice-bonded submarine permafrost at the upper layers of the seabed (by 5 m) at seawater depth of ≤35 m. The spots are possible indicators of the more deep-lying body of relic permafrost. Taliks.
(C) Thickness of the permafrost at the coast is high—about 600 m. The cooling effect of this factor is the major one. Ice at depth of ≤2 m.	Low suitability. Only at the boundary zone with the western area. Body of permafrost is the continuation of the thick coastal cryogenic zone limited by first 18–22 km. Taliks. General thickness about 150–200 m. Seaward less, 50–30 m. Seawater depth ≤12 m.	The areas suitable for submarine ice-bonded permafrost at the upper layers of the seabed are restricted by the first 18–22 km from the shore, including barrier islands (seawater depth ≤10 m). Taliks.
(E) Ice at depth of ≤2 m. Influence of the coastal permafrost body.	No relic permafrost. Probably permafrost limited by continuation of the coastal permafrost body, at a distance no more than 2–8 km from the shore. The depth to permafrost could be 50–100 m. Thickness depends on coastal permafrost. Usually reduced 2–3 times. Lack of site specific data.	No areas suitable for submarine permafrost at the upper layers of the seabed. Only at the limits of the 2 m of seawater depth (including barrier islands).

has been described. This division has been made according to the evaluation of the paleoenvironmental and environmental aspects of the subsea permafrost.

According to the specifics of the origin and development of the permafrost, the Beaufort Sea shelf has been divided into three parts: (1) the shelf area suitable for submarine relic permafrost (western part of the shelf); (2) area with low suitability for relic permafrost (central part of the shelf); and (3) area without relic permafrost (eastern part). Estimations of thickness and specifics of submarine permafrost in these areas have also been made.

To evaluate all existing environmental data on recent subsea permafrost development, a computerized system has been developed. Thirty computerized maps of different oceanographic, geologic, glaciologic, and other parameters have been used to generate source data maps, derived maps (three generations), and a composite map specifying the candidate area for submarine permafrost development. In order to develop the system a special composite mapping algorithm for submarine permafrost prediction has been made.

The system has been checked in the Beaufort Sea areas with known permafrost. Comparison with Canadian data on submarine permafrost has shown positive results and high correlation.

The shelf maps showing suitability for submarine permafrost have been compared with the BLM Lease Nomination map. The extension of the suitable areas for permafrost at each nomination site has also been calculated.

It would seem to be helpful, in the future, to pay more attention to the western part of the Beaufort shelf. This is an area that could have been exposed in the Pleistocene and the probability of thermokarst forms in a thawing state and the buried ancient valleys (with possible lenses of ice and ice-bonded permafrost) could represent additional environmental hazards. The relationship between the submarine permafrost at the upper layers of the shelf deposits and the deeper relic permafrost, or between the submarine permafrost and the coastal "body" of permafrost, also need to be objects of special investigations.

The same work was done for the Chukchi Sea. It was found that there are limited areas suitable for ice-bonded submarine permafrost, situated to the northwest from the Barrow Canyon and at some sites along the coastal line.

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A GEOGRAPHIC BASED INFORMATION MANAGEMENT SYSTEM FOR PERMAFROST PREDICTION IN THE BEAUFORT AND CHUKCHI SEAS

PART II.
SUBMARINE PERMAFROST ON THE ARCTIC SHELF OF EURASIA AND THE DEVELOPMENT OF THE ARCTIC IN THE PLEISTOCENE

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INTRODUCTION

Since permafrost is defined upon the basis of temperature, its submarine variation may exist wherever mean annual sea bottom temperatures are below zero. The areas of its development, or the "subsea cryolithic zone" is the designation for a zone of the Arctic shelf having a negative temperature. In this zone the frozen rocks are rocks which contain ice; the cold (frost) rocks are rocks whose temperature is below zero but which do not contain ice. The permafrost is extensively developed below the bottom of the Arctic Ocean and its fringe seas. However, the geocryological study of the sea bottom is extremely deficient. The information about the distribution, properties and development of rocks of the sea bottom cryolithic zone has already long been necessary to support navigation through the northern maritime routes and for the construction of ports. Their requirement has sharply increased in connection with the initiated conquest of natural resources of the Arctic shelf. The practical significance, poorly studied nature and poor access of the subsea cryolithic zone render its investigation one of the most interesting problems of modern cold research.

In view of the prospective geological exploration and commercial exploitation of oil and other useful mineral deposits, as well as the construction of hydraulic engineering structures in the shelf zone of the Arctic seas, and particularly the shallow-water coastal area, the comprehensive analysis of the submarine permafrost data on the Soviet Arctic seas shelf is acquiring more and more importance.

The development of research on submarine permafrost of the Alaska shelf has much to gain from Eurasiatic Arctic shelf investigations. The early and steady northerly settlement of the USSR Arctic coast, development of the "Northern Marine Road," and shelf mineral resources research created a growing interest in the properties of subsea permafrost and their adverse effect on construction. Expeditions to study subsea permafrost occurrences, made as early as the beginning of this century, had provided a steady, and centralized accumulation of experience and data. In 1930 the USSR, then promulgating decrees for the further settlement of the sparsely populated north and east, established the first formal agency to assemble data and further permafrost investigation, the Commission for the Study of Permafrost of the Academy of Sciences of the USSR.

In 1939 the commission was reorganized and became the V. A. Obruchev Institute of Frost Studies (later changed to Geocryology). As early as 1940, the Institute maintained special permafrost laboratories at Moscow and Leningrad, as well as four field stations in the North. After several more reorganizations, most permafrost work was assigned a few years ago to two institutes of the State Construction Board (the Gosstroy): (1) the Scientific Research Institute of Foundation Soils and Underground Structures (NIIOSP), and (2) the Operations and Scientific Research Institute for Engineering Site Investigations (PNIIS). A field station at Yakutsk then became a full-fledged institute for permafrost studies under the jurisdiction of the Siberian Division of the Academy of Sciences. Most other field stations are now administered from Yakutsk or by the Gosstroy. There are other institutes, such as the Building Research Institute of the Russian Soviet Federated Socialist Republic at Krasnoyarsk, and an institute at Magadan. These institutes and others like them have their own field stations.

Permafrost investigations are also being carried out very actively at some universities, particularly at the Moscow State University. There is a Department of Permafrost and a Department of Polar Regions and Geocryology in the Faculty of Geology. In addition to instructing students, the staffs of these departments study permafrost at various Arctic coast locations in Siberia and their work ranks with the investigations being carried out by the State institutes. The total result is that probably about three hundred people are engaged in subsea permafrost investigations in the Soviet Union.

The list of Soviet organizations taking part in the Submarine Permafrost Study in the Laptev Sea, Kara Sea, East Siberian Sea, and Chukchi Sea is as follows:

1. Permafrost (Geocryology) Science Institute, Novosibirck, Yakutsk, Siberian Division of the USSR Academy of Sciences.
2. Hydrographic Administrations of the USSR Navy Ministry (Different points of the Soviet Arctic Coast).
3. Scientific Research Institute of Arctic Geology of the USSR Geology Ministry, Leningrad.
4. Moscow University, Permafrost Study Department, Department of Polar Regions and Geocryology, Scientific Research Laboratory for Problems in the Mastery of the North.
5. Geological Survey of the USSR Geology Ministry (Different expeditions), Moscow, Leningrad, Norilsk, Vorkuta, Magadan, and others.
6. Leningrad Mining Institute, Department of Marine Geology, Department of Hydrogeology and Engineering Geology.
7. Leningrad University, Geological Department.
8. State Construction Board (the Gosstroy). Scientific Research Institute of Foundation Soils and Underground Structures (NIIOSP) and the Operations and Scientific Research Institute for Engineering Site Investigations (PNIIS). Moscow, Leningrad, Vorkuta, Magadan, and others.
9. Northeastern Regional Scientific-Research Institute (*SVKNII*), Magadan.
10. Far Eastern Scientific Center (*DVNTS*), AS USSR, Magadan, Vladivostok.

HISTORY OF INVESTIGATIONS

The report of A. Ye Nordenshel'd is among the first information about the presence of frozen rocks on the bottom of the Arctic seas (1880). The report notes that the sandy bottom in one of the bays of the eastern part of the Chukotsk Sea shoreline is cemented by ice.

The monograph written by V. Yu Vize et al. (1946) states that on the bottom of the Dmitriy Laptev Strait ice has been detected which is not subject to thawing due to the sub-zero water temperature.

The factual data about the permafrost below the bottom of the Arctic seas, until recently, were available solely for the coastal shallows. These data were chiefly obtained at the Permafrost Institute of the Siberian Department of the Academy of Sciences of the USSR by N. F. Grigor'yev (1966), who made a great contribution to studying the subsea permafrost. Beginning in 1953, under his supervision, a large number of boreholes were drilled and these boreholes encountered permafrost or cold rocks at the receding shores of the open sea and in the shallows of sea gulfs to a distance of several hundred meters from the coast, as well as in the mouth debouchment areas of the Yana and Indigirka rivers to a distance of up to 25 km from the outer edge of the delta. In all cases, the water depth was less than 2 m. V. A. Usov (1965) and M. S. Ivanov (1969) participated in these investigations. In 1962–72, Ye. N. Molochushkin (1973) studied the permafrost and cold rocks near the outcropping shores of Mostakh Island in Burokhaya Gulf and in the Van'kin Gulf of the Laptev Sea, as well as in the near-mouth salt water part of the Lena River in the region of the Oleneksk tributary. The first reports of the presence of subsea permafrost outside the limits of the 2 m water level were obtained by V. M. Ponomarev (1960, 1961) for Kozhevnikov Bay and Khatangsk Gulf, where many boreholes revealed permafrost and cold rocks. At one of the drilling points a distance of 3 km from the shore at a water depth of 3 m, the thickness of the permafrost was in excess of 66 m. V. M. Ponomarev also points out that V. F. Zyukov and N. I. Saltykov encountered permafrost at a water depth of less than 2 m in the coastal zone of the Ob Gulf. In 1970, Ye. N. Molochushkin (1973) detected permafrost during sampling of bottom deposits of the Ebelyakhsky Gulf and the Dmitriy Laptev Bay in several locations at distances of up to 30 km from the coast in the lower parts of extracted cores during the use of a vibropiston tube. According to L. A. Zhigarev and I. R. Plakht (1974), heavily ice-bonded rocks were discovered in the submarine part of Van'kin Gulf of the Laptev Sea 10 km from the shore.

Based on general concepts and extremely limited factual material, I. Ya. Baranov (1958, 1960) identified two zones of subsea permafrost layers. The first of these zones extends over the entire Arctic shelf of Asia to an isobath of 100 m. The second lies in the coastal zone of the Karsk, Laptev, and Eastern Siberian seas to an isobath of 20 m. These strata were formed in the Pleistocene and partly in the Holocene, when the shelf was dry. The first zone was converted to the subsea position as a consequence of marine transgression. The second underwent this process as the result of abrasion at a constant sea level. Today, the subsea permafrost strata are degrading, in the opinion of I. Ya. Baranov. His ideas were further developed in the works of N. F. Grigor'yev (1966); Ye. N. Molochushkin (1970), F. E. Are and D. N. Tolstyakov (1970), and S. V. Tomirdiaro (1974) also extended their scope.

The possibility and necessity of a prolonged existence of the submarine cryolithic zone are due to the constant negative temperature of the near-bottom layers of seawater over a large part of the territory of the Arctic seas and the Arctic Ocean (AANII 1954; Nordenshel'd 1880 b; Timofeyev 1957; Sovetskaya Arktika 1970; Neizvestnov et al. 1972; Chekhovskiy 1972; Brewer 1958; Collin 1960; Lachenbruch, Marshall 1968; Lewellen 1973, and others). Thermal aspects of the Arctic waters have been studied in quite detailed fashion. Therefore, today, it is quite possible to construct a diagram of distribution of the submarine cryolithic zone, as, for example, was done by A. L. Chekovskiy (1972) for the Kara Sea shelf. But it is much more complicated to determine the thickness of rock with a negative temperature, and still more to explain the morphology of the permafrost strata. Drilling deep boreholes on the bottom of the Arctic seas is only beginning, and making geothermal measurements within them presently remains a matter for the future. Geophysical methods are inadequately developed for the examined conditions and cannot provide reliable data without confirmation by drilling (F. Frolov 1961; Artikaev 1964; F. Frolov, Zykov 1971).

Today the concept of the morphology and nature of the submarine permafrost of the Arctic shelf can only be devised on the basis of approximate calculations. The chief factor that must be taken into account in this case is the contemporary temperature of the near-bottom water layers (the sea bottom). But one should bear in mind that under continental conditions the surface temperature correlates poorly with the thickness of the cryolithic zone (Balobayev 1973). Obviously, this is also valid for the submarine permafrost of the shelf, where the surface temperature of the bottom in the recent geological past changed many times, rapidly and extremely significantly as a consequence of the alternation of transgressions and regressions of the sea. In this connection, a consideration of the developmental history of the Arctic shelf has primary significance.

No less important are reports about the paleoclimate, under whose direct influence the cryolithic zone of dry land formed and upon which the temperature of the seawater depended. An important factor affecting the thickness of the cryolithic zone is the stream of intraground heat. It is entirely necessary to take into account the degree of mineralization of the seawater and subterranean waters of the shelf, since the phase state of water in rock pores depends not only on its temperature, but also its mineralization. It is known that rocks are widespread in the coastal zone of the Arctic seas which are saturated with mineralized water and are frozen

(Vittenburg 1940; Ponomarev 1950, 1960; Grigor'yev 1966; Govorucka 1968; Molochushkin, Gavril'yev 1970; Neizvestnov, Semenov 1973; Pinneker 1973). In order to estimate the possible thickness of the submarine cryolithic zone, one must know the geological structure of the sea bottom in order to use it as a basis for an approximate determination of the thermophysical properties and moisture content of the rocks. It is also necessary to take into account the abrasion process and erosion washing of the bottom by sea currents which can reduce thickness or cause destruction of the equilibrial state of the cryolithic zone. The accumulation of bottom sediments can lead to limestone formation in the sea and to the formation of a new or increased thickness of an earlier existing cryolithic zone. According to F. Are (1976), among the most important factors which affect the formation of the submarine cryolithic zone one should include the following: relative sea level and dry land level; climate and temperature of the sea bottom; the intraground heat flux; mineralization of the near-bottom layers of water and subterranean waters of the sea bottom; properties of rocks which underlie the bottom; and processes of abrasion and accumulation of bottom sediments.

The Russian literature on the research and theory of submarine permafrost and the practical applications of the results of the study of the properties of the frozen deposits comprises about four hundred publications. All these publications demonstrated the complex nature of the cryogenic shelf formations and studied several regional features of their formation, structure, and distribution. However, many aspects of this theoretically and practically important problem have either not yet been solved or are still being discussed. For example, the concepts of the thermal state and structure of submarine permafrost are not sufficiently clear, the thermodynamic aspects of the conditions for the formation, preservation and destruction of cryogenic formations on the shelf have not been explained, seasonal underwater cryogenic formations and their dynamics have been studied very little, there is little information on the physical and mechanical properties of these deposits and changes in them depending on temperature, humidity, and ice conditions, the terminology has not been put in order, and methodological principles for investigating and mapping these strata have not been worked out. All the literature may be divided into five parts reflecting the main aspects of submarine permafrost study in the Soviet Union: (1) regional distribution and characteristics, composition and structure; (2) thermal state and regime. Environmental data on submarine permafrost. Regional coastal dynamics, arctic shoreline processes, the ice processes in the coastal zone, influence of the river flow, thermal and chemical characteristics of the seawater; (3) Dynamics of the Arctic shelf and coastal development; (4) The problems of the exposed shelf glacio-isostasy, block-tectonics related to permafrost development (Figures 1, 2, 3, 4); and (5) The types of the Arctic shelf permafrost. These parts correspond to the following chapters.

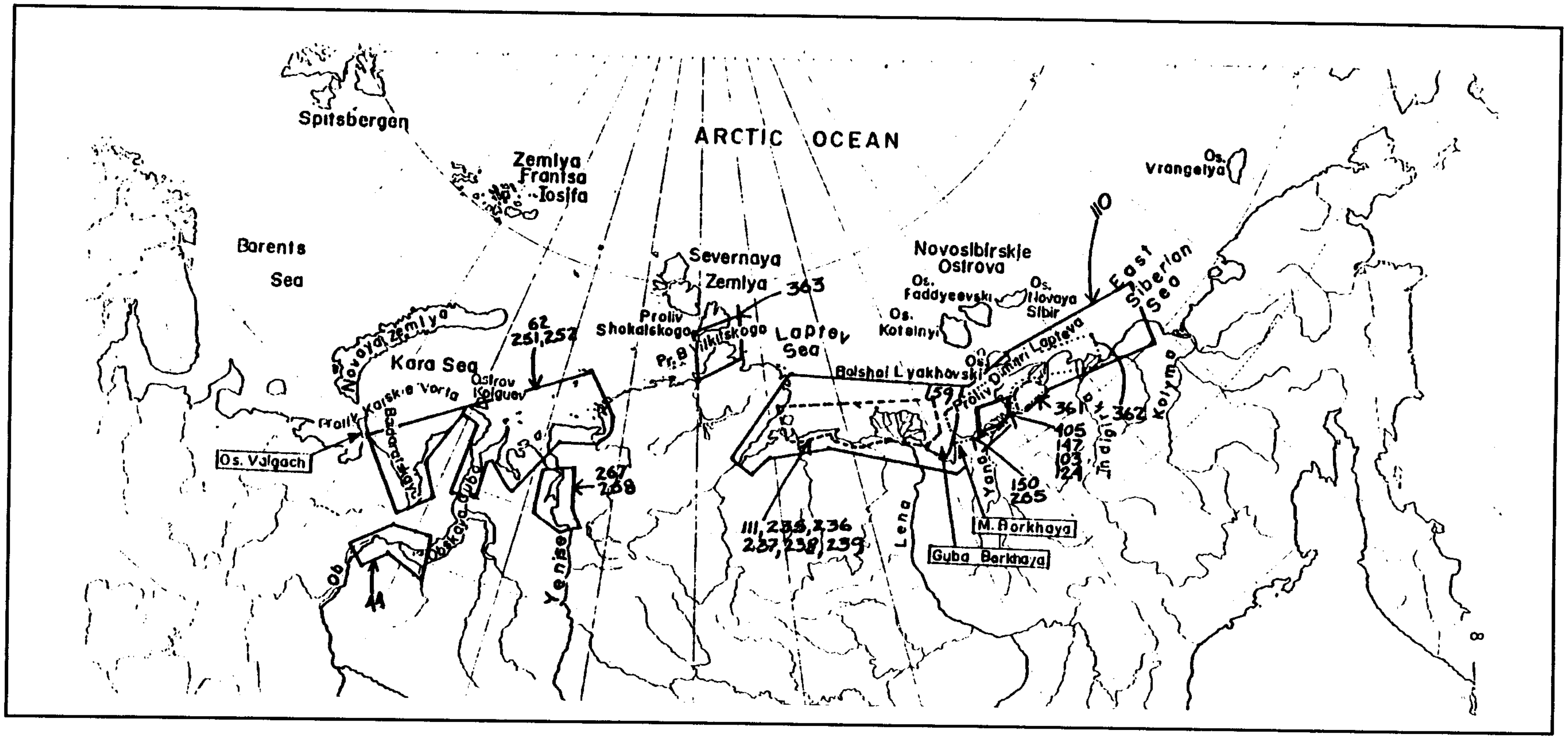


Figure 1.—Areas of investigation connected with submarine permafrost regional distribution, characteristics, composition, and structure.

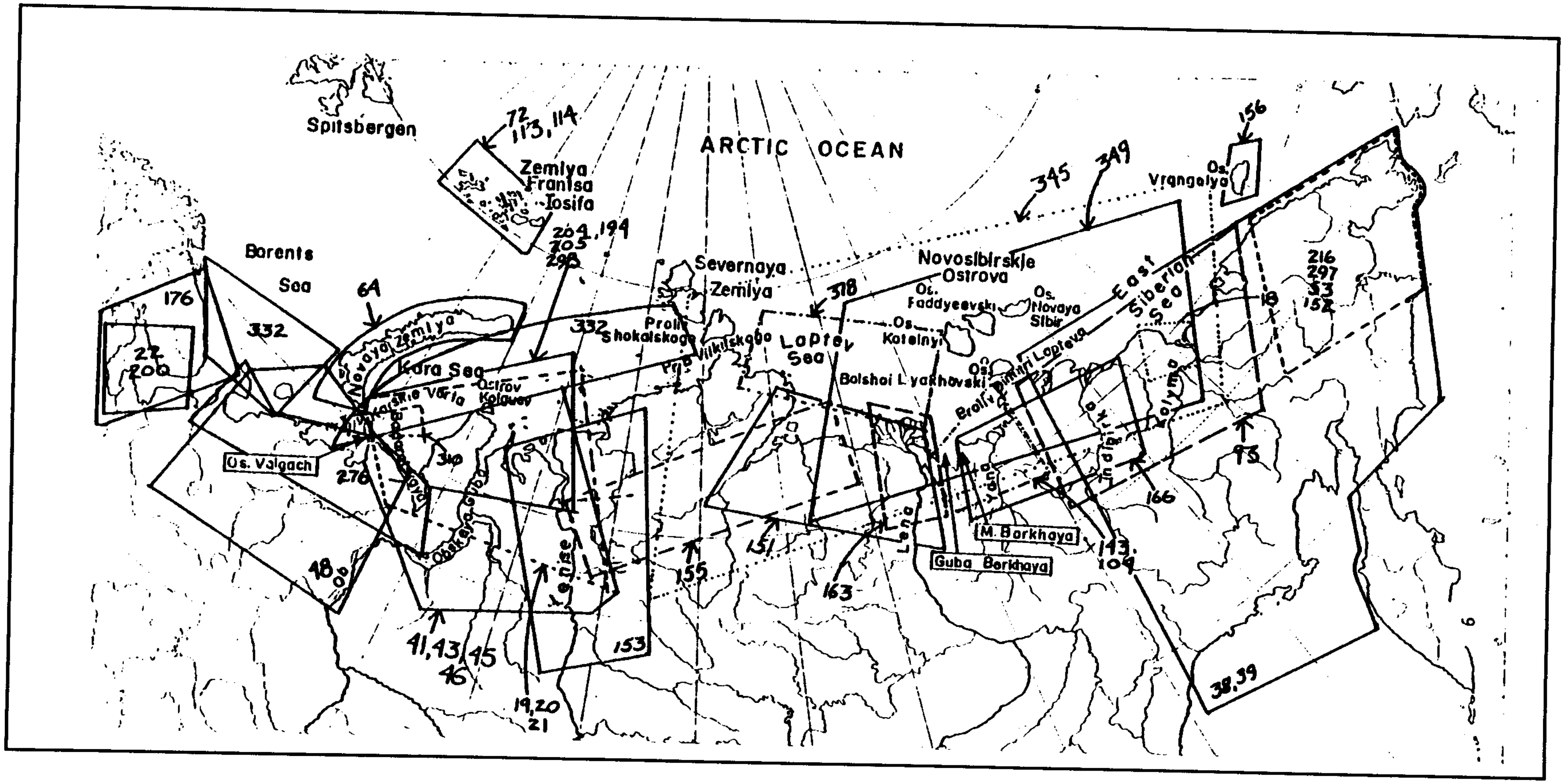


Figure 2.—Areas of investigation connected with submarine permafrost, genesis history, paleogeographical conditions (changing of the sea level, regressions and transgressions, Pleistocene and recent tectonics, paleoclimatic data). Numbers correspond to numbers in bibliography.

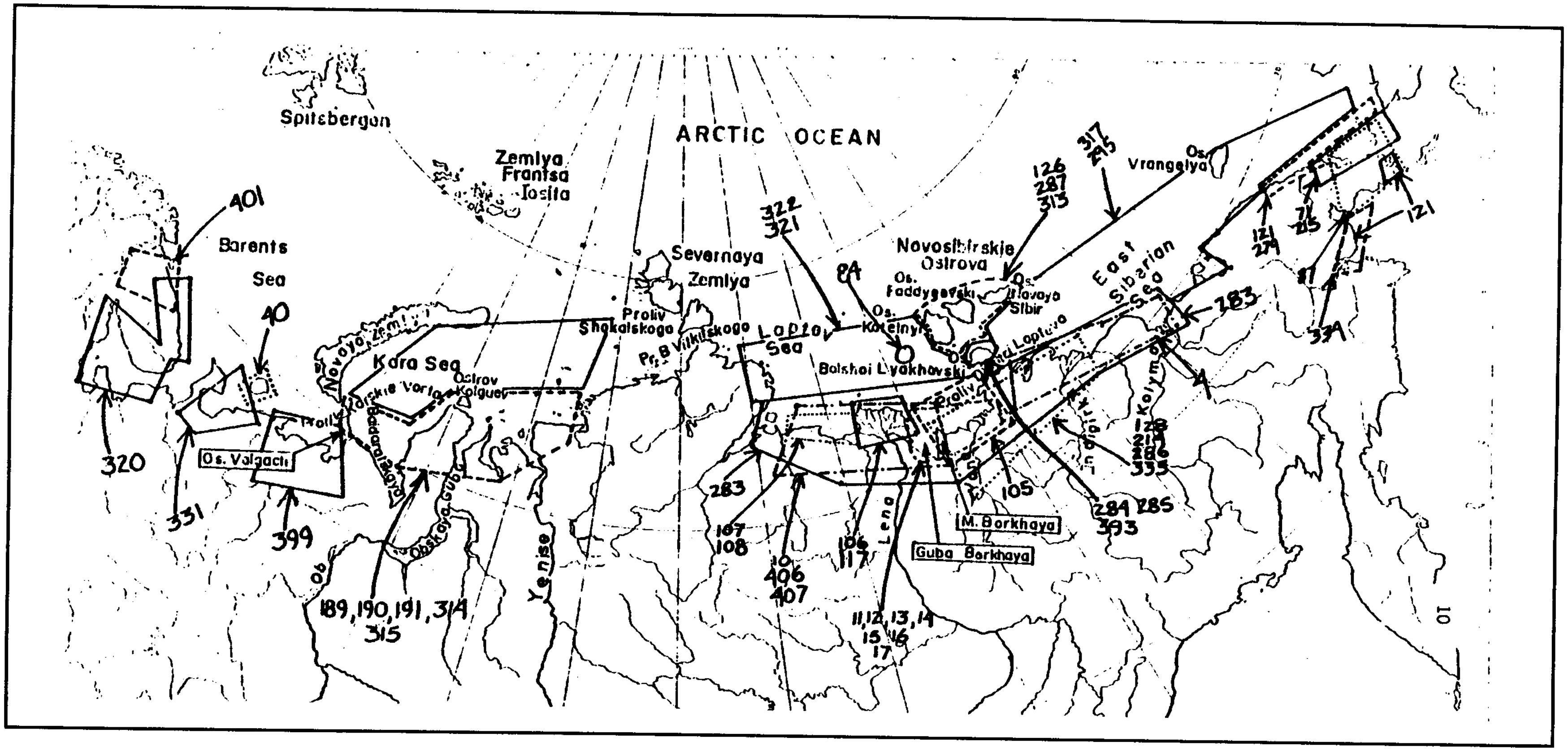
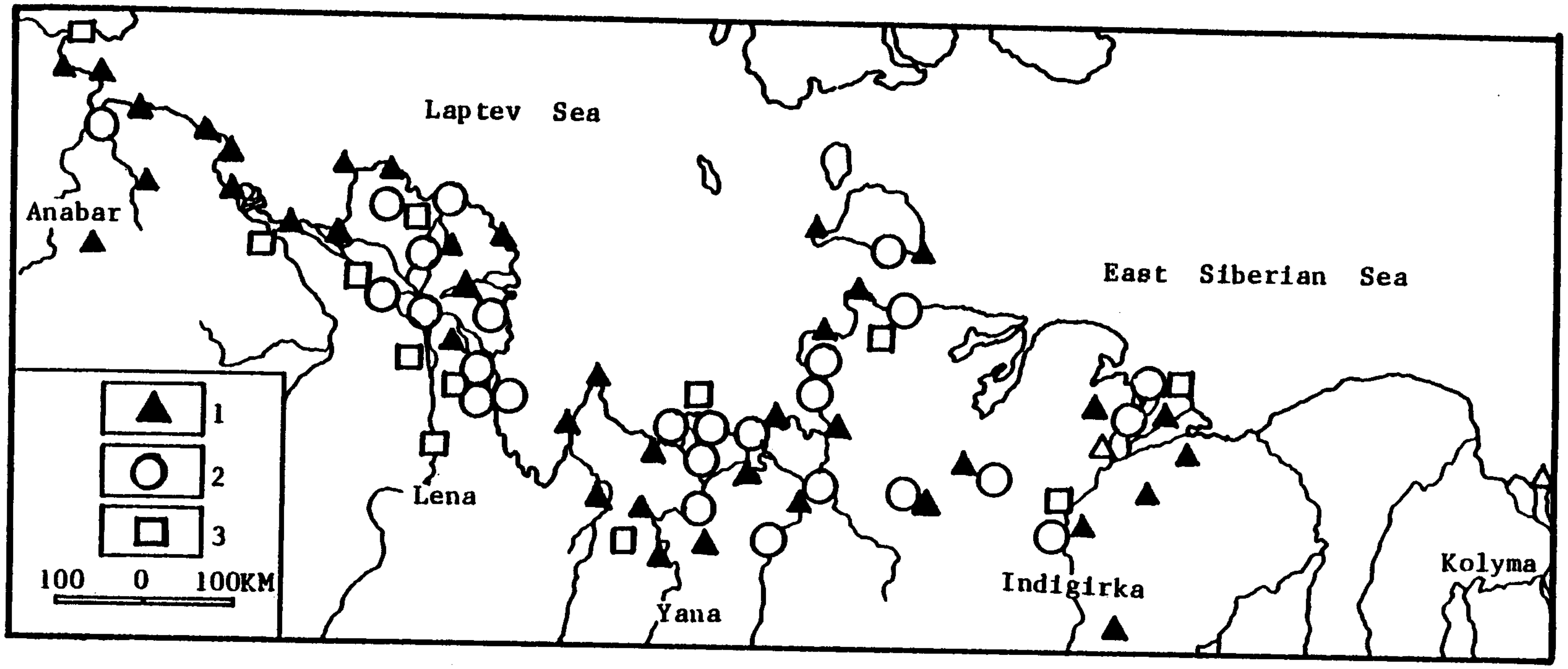


Figure 3.—Areas of investigations connected with submarine permafrost, geological and geomorphological environments, thermal erosion, coastal dynamics, arctic shoreline processes, shelf bottom relief and deposits, ice processes, hydrological peculiarities. Numbers correspond to numbers in bibliography.



1. Regions where there have been reconnaissance permafrost-geological investigations by the Permafrost Science Institute, USSR Academy of Science.
2. Regions where drilling and thermometric investigations have been made by the Permafrost Science Institute.
3. Regions where permafrost investigations have been made by other organizatons incidentally with geological prospecting work.

Figure 4.—Schematic map of permafrost studies of the arctic part of Yakutia (Grigor'yev 1966).

SUBMARINE PERMAFROST REGIONAL DISTRIBUTION, COMPOSITION, AND STRUCTURES

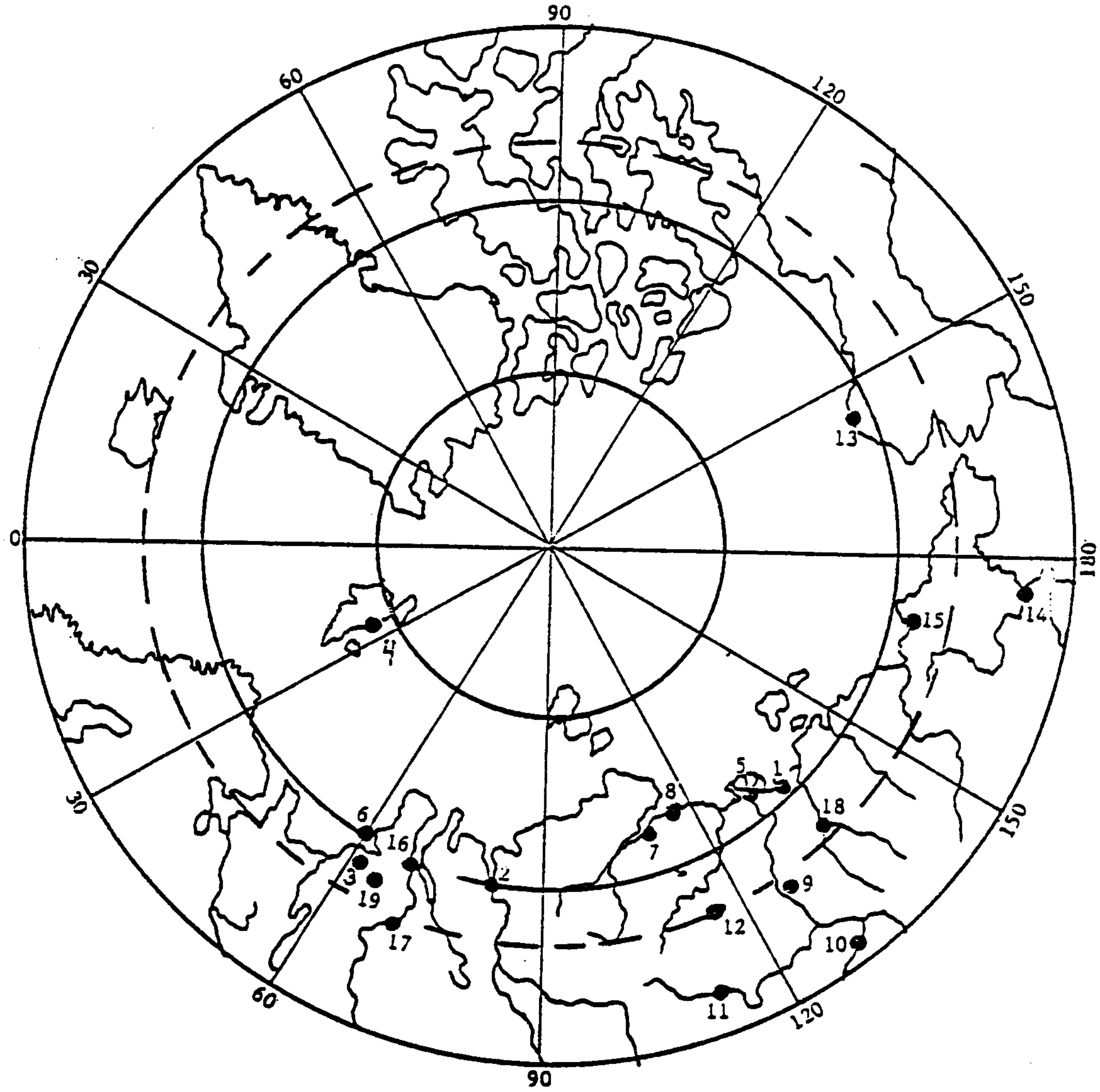
In this chapter we are considering the following four questions: (1) thickness of the rock zone with subzero temperature on the Eurasia Arctic coast; (2) data on submarine permafrost extension in Laptev and East Siberian Seas; (3) depth and thickness; and (4) composition and structure.

Figure 5 shows the points of geothermal observation in the Arctic. Tables 1, 2, and 3 give the temperature of permanently frozen ground in some of these points according to Grigor'yev (1966). Tables 4, 5, and 6 give the geothermal gradients based on temperature observations made in Arctic coastal zones according to Oberman and Kakunov (1973). Using the geothermal gradients of the 250–1,000 m interval and hydrogeological data (salinity and temperature), these authors made a conclusion that the zero isotherm in the Laptev Sea coastal zone has to be at 900–1,000 m (Kojevnikov Bay and Chay Tumus), in the Kara Sea at 500–900 m (Ust Port and Amderma). Newer boreholes have reached the bottom of rocks with subzero temperature in some of these coastal areas on corresponding depth.

Permanently frozen deposits were discovered in the Laptev Sea at a maximal distance about 26 km from the shore near the mouth of the Yana River (Grigor'yev 1966, Ivanov 1969) and in Dimitri Laptev Strait at a distance of 25–35 km (Molochushkin 1973). Figure 6, Table 7, Figures 7 and 8 show the results of drilling and observations in these areas.

In the East Siberian Sea perennially frozen ground was met at a maximal distance about 18 km from the shore in the region of beach area at the mouth of the Indigirka River (Figures 9, 10 and 11; Usov 1965, Grigor'yev 1966). In 1960 a general geocryological map of the USSR was published at a scale of 1:10,000,000. It was compiled by I. Ya. Baranov, and for the first time in the practice of compiling small-scale permafrost maps he defined a zone of permanently frozen ground under the bottom of Arctic seas. The northern boundary of this zone within the limits of the Laptev Sea was drawn approximately along the margin of the shelf. In the west of Siberia the permafrost zone on the sea floor was shown to the region of the Yamal Peninsula in the Kara Sea and on the east to the region of Cape Billings in the East Siberian Sea. Later, in 1974, Fotiev et al. also had shown the permafrost extension over all the shelf at 300–400 km from the shore (Figures 13 and 14). All these maps are based on general paleogeographical estimations and we would like to consider this later.

Regional details concerning the area, distribution, thickness, and depth of the upper boundary of the permafrost are known in the Arctic seas of Eurasia thanks to investigations of many geologists, especially V. M. Ponomarev (1940, 1960, 1961), N. F. Grigor'yev (1952, 1964, 1966), V. A. Usov (1967, 1970), Ye. N. Molochushkin (1972), U. A. Zhigarev and T. R. Plakht (1974) and others. The scientists usually specify the kind of submarine permafrost they



1. Tiksi Bay
2. Ust-Port
3. Bolshezemelskaia Tundra
4. Spitzbergen
5. Lena Mouth, Chay-Tumus
6. Amderma
7. Kojevnikov Bay
8. Nordvik
9. Bahinai
10. Namtci
11. Mirni
12. Marha
13. Alaska, Barrow
14. Anadir
15. Pevek
16. Ob Mouth
17. Ob Bay
18. Verhsyansk
19. Vorkuta

Figure 5.—Points of geothermal observation. After Oberman and Kakunov (1973).

Table 1.—Temperature (°C) of permanently frozen ground in Chay-Tumus region. After Grigor'yev (1966).

Depth (m)	Borehole				Depth (m)	Borehole			
	1	4	5	9		1	4	5	9
4	-6.6	-8.7	-8.2	-7.4	70	–	–	-10.3	–
6	-7.8	-9.8	-10.4	-9.0	74	–	-9.6	-10.3	–
8	-8.5	–	-11.3	–	80	–	-9.4	-10.2	–
10	-8.7	-10.4	-11.2	-10.0	90	-7.4	–	-9.8	–
12	–	-10.6	-11.2	-10.0	100	–	-9.1	-9.6	–
14	–	–	-11.2	-9.8	120	-6.8	-8.8	-8.8	–
16	-8.5	-10.4	-10.8	-9.8	140	–	-8.4	-7.8	–
18	-8.5	–	–	-9.7	160	-6.2	-8.4	-7.5	–
20	–	-10.4	-10.8	-9.6	180	-5.8	-7.8	-7.3	–
24	–	-10.4	-10.3	–	200	–	–	-7.0	–
30	-8.4	-10.2	-10.8	-9.6	220	–	–	-6.2	–
40	-8.4	–	-10.8	-9.6	230	–	–	-6.0	–
48	–	-10.0	-10.6	-9.5	240	–	–	-5.5	–
50	-8.3	-10.0	-10.6	–	250	–	–	-5.4	–
54	–	-10.0	-10.6	–	260	–	–	-5.3	–
56	–	-9.9	-10.6	–	280	–	–	-5.2	–
58	–	-9.9	-10.4	–	300	–	–	-5.0	–
60	-8.1	–	-10.4	-9.5	330	–	–	-4.9	–
64	–	-9.8	-10.4	–					

Table 2.—Temperature of permanently frozen ground in neighborhood of Tiksi Bay. After Grigor'yev (1966).

Depth (m)	Temperature (°C)	Depth (m)	Temperature (°C)
20	-11.1	340	-6.0
100	-10.5	360	-5.6
140	-9.8	380	-5.2
200	-8.6	400	-4.8
220	-8.2	420	-4.4
240	-7.8	440	-4.1
260	-7.5	480	-3.7
280	-7.1	480	-3.3
300	-6.8	500	-2.9
320	-6.4		

Table 3.—Temperature of permanently frozen ground in the neighborhood of Cape Val'kumey. After Grigor'yev (1966).

Depth (m)	Temperature (°C)	Depth (m)	Temperature (°C)
50	-5.1	230	0.0
90	-4.0	255	+0.7
150	-2.5	290	+2.0
200	-1.0	348	+4.2
220	-0.1		

Table 4.—Ground temperature (°C) in bar of channel in middle delta of the Indigirka River. After Oberman and Kakunov (1973).

Depth (m)	Borehole and Observation Date						
	1 22–26/7	2 20–21/7	3 23–24/7	5 30/7	6 2/8	7 3/8	8 12/8
0	–	7.2	10.0	11.8	–	7.9	4.2
0.5	−0.8	–	0.6	2.0	−0.6	3.2	4.6
1.0	−3.2	−3.6	−0.9	−0.4	–	−1.5	4.8
1.5	−5.0	–	−1.7	−1.3	−0.5	−3.9	3.8
2.0	−6.5	−7.3	−2.8	−1.6	−0.5	−5.0	2.5
2.5	−8.0	−8.9	−3.1	−1.6	−0.3	–	1.6
3.0	−9.2	−10.5	−3.4	−1.5	−0.2	−8.5	1.0
3.5	−10.1	−11.3	−3.4	−1.2	–	−9.1	–
4.0	−10.7	−11.6	−3.5	−1.2	–	−9.3	–
4.5	−11.3	–	−3.5	–	–	−9.6	–
5.0	−11.7	–	−3.5	–	–	−9.6	–
5.5	−12.1	–	–	–	–	–	–
6.0	−12.2	–	−3.2	–	–	–	–
6.5	−12.3	–	−3.2	–	–	–	–
7.0	−12.3	–	−3.1	–	–	–	–
7.5	−12.2	–	−2.9	–	–	–	–
8.0	−12.1	–	−2.8	–	–	–	–
8.5	−12.0	–	–	8	–	–	–
9.0	−11.8	–	–	–	–	–	–

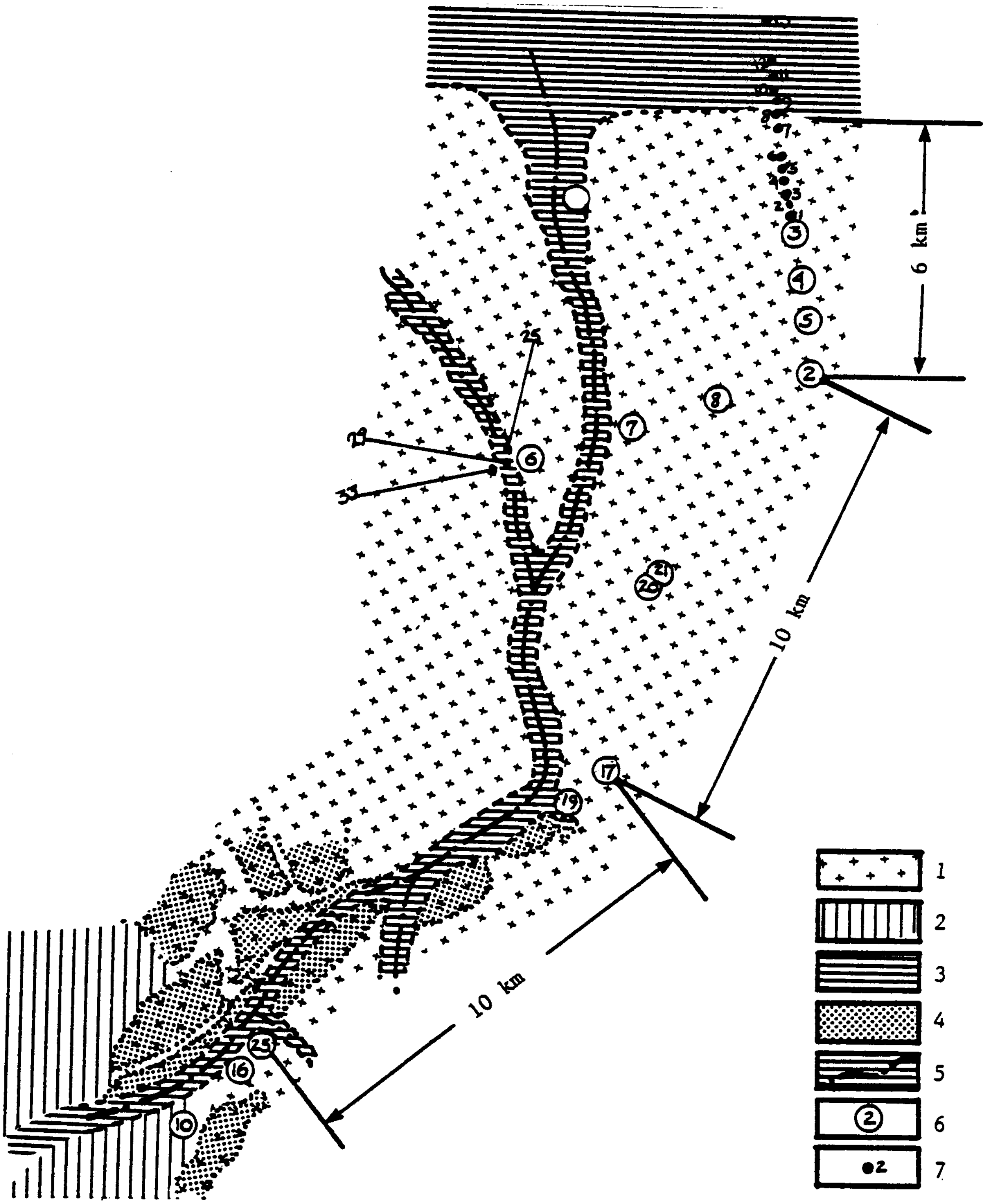
Table 5.—Geothermal step and geothermal gradient based on temperature observations made in 1964 in Chay-Tumus region. After Oberman and Kakunov (1973).

Depth Interval	Geothermal Step m/Degree	Geothermal Gradient °/100 m
	Borehole 1	
60–120	47	2.1
120–180	62	1.6
	Borehole 5	
20–120	42	2.4
120–220	41	2.2
220–320	61	1.5

describe as simple sediment or rock at a year round temperature below 0°C; ice-bonded, brine soaked sediment; or cold, dry rock at a negative temperature. They also specify the permafrost as that: which in equilibrium with the modern temperature regime and the relict permafrost, which was formed when the climate was colder than now or, in the case of submerged shelf areas, which was formed before submergence. Sometimes the authors try to use the terminology

Table 6.—Rock temperature at the bottom of the ground heat storage layer and geothermal gradients. After Oberman and Kakunov (1973).

Points of Observation	Permafrost Thickness Borehole Data (m)	Depth of Measurement (m)	Depth Interval 30–250		Depth Interval 250–1,000	Depth Interval 30–1,000	Rock Temperature at the Depth of 30 m	Mineralization (g.l.)
			Geothermal Gradient, Degree 100 m					
			Negative	Positive	Positive	Positive		
TO THE EAST FROM TAYMYR								
Nordvik B.8	600(800)	320		2.40	1.31	2.10	−11.6	300
Kojenvinkov Bay B.6	600	503		4.09	0.40	2.13	−11.5	165
B.5	600	380		4,00	0.38	2.58	−11.0	
Tiksi Bay	640	500		1.50	1.88	1.71	−11.1	2
Lena Mouth Chay-Tumus B.5	540–560	330		2.45	0.63	1.97	−10.8	
Anadir	>150	77		5.41		5.41	−5.8	27–70
Pevek	230	348		2.86		2.86	−5.2	>1
Average				3.2	0.9		−9.5	
TO THE WEST FROM TAYMYR								
Amderma B.75	400	274	−0.67	0.87	0.42	0.82	−4.7	63–133
Ust-Port B.32	>400	400		1.00	0.40	0.76	−3.2	45
Spitzbergen B.248	400	310	−0.94	1.20	0.75	1.04	−3.0	34–44
B.13	180	180	−1.00	2.00		2.00	−2.4	34–44
Ob Bay B.U-1	120	120	−1.50	1.25		1.25	−1.9	Salt water
Amderma B.AD-2	900	773	−0.62		0.44	0.44	−1.7	60–92
Ust-Port B.2GBG	310	850	−1.08	0.75	1.17	0.85	−0.6	
Bolshezemelskaia Tundra B.9	350	1060	−0.43		1.02	1.02	−0.5	10–20
Ob Bay B.59	250	245	−0.12				−0.3	1–18
Average			−0.8	1.2	0.7		−2.0	



1. Underwater layer of modern perennially frozen deposits
2. Continental Holocene-delta perennially frozen deposits
3. Talik zone near delta
4. Above-water, periodically inundated shoals—"sands"
5. Underwater channel and talik zone
6. Boreholes
7. Points of thermal soundings

Figure 6.—Schematic permafrost map of beach near mouth of Yana River, Laptev Sea (Grigor'yev 1966).

Table 7.—Underground water mineralization and geothermal gradients. After Oberman and Kakunov (1973).

Points of Observation		Permafrost Thickness Borehole Data (m)	Depth of Measurement (m)	Depth Interval 30–250 Geothermal Gradient, Degree 100 m Negative	Depth Interval 30–250 Geothermal Gradient, Degree 100 m Positive	Depth Interval 250–1,000 Positive	Depth Interval 30–1,000 Positive	Rock Temperature at the Depth of 30 m	Mineralization (g.l.)
Tiksi Bay		640	500		1.50	1.88	1.71	-11.1	2
Ust-Port	B.2GBG	310	850	-1.08	0.75	1.47	0.85	-0.6	9
Bolshezemelskaia Tundra	B.9	350	1060	-0.43		1.02	1.02	-0.5	10-20
	B.16	500	1000		0.85	1.16	1.05		10-20
Average				-0.8	1.0	1.3		-4.1	
Spitzbergen	B.248	>400	310	-0.94	1.20	0.75	1.04	-3.0	34-44
Lena Mouth Chay-Tumus	B.5	540-560	330		2.45	0.63	1.97	-10.8	
Ust-Port	B.32	400	400		1.00	0.40	0.76	-3.2	45
Amderma	B.AD-2	900	773	-0.62		0.44	0.44	-1.7	60-92
	B.75	400	274	-0.67	0.87	0.42	0.82	-4.7	63,133
Kojevnikov Bay	B.6	600	503		4.09	0.40	2.13	-11.5	165
	B.5	600	380		4.00	0.38	2.58	-11.0	165
Nordvik	B.8	600(800?)	320		2.40	1.3	2.10	-11.0	300
Average				-0.7	2.3	0.6		-7.2	30

Table 8.—Temperature (°C) of bottom deposits for beach near the mouth of the Yana River. After Ivanov (1969).

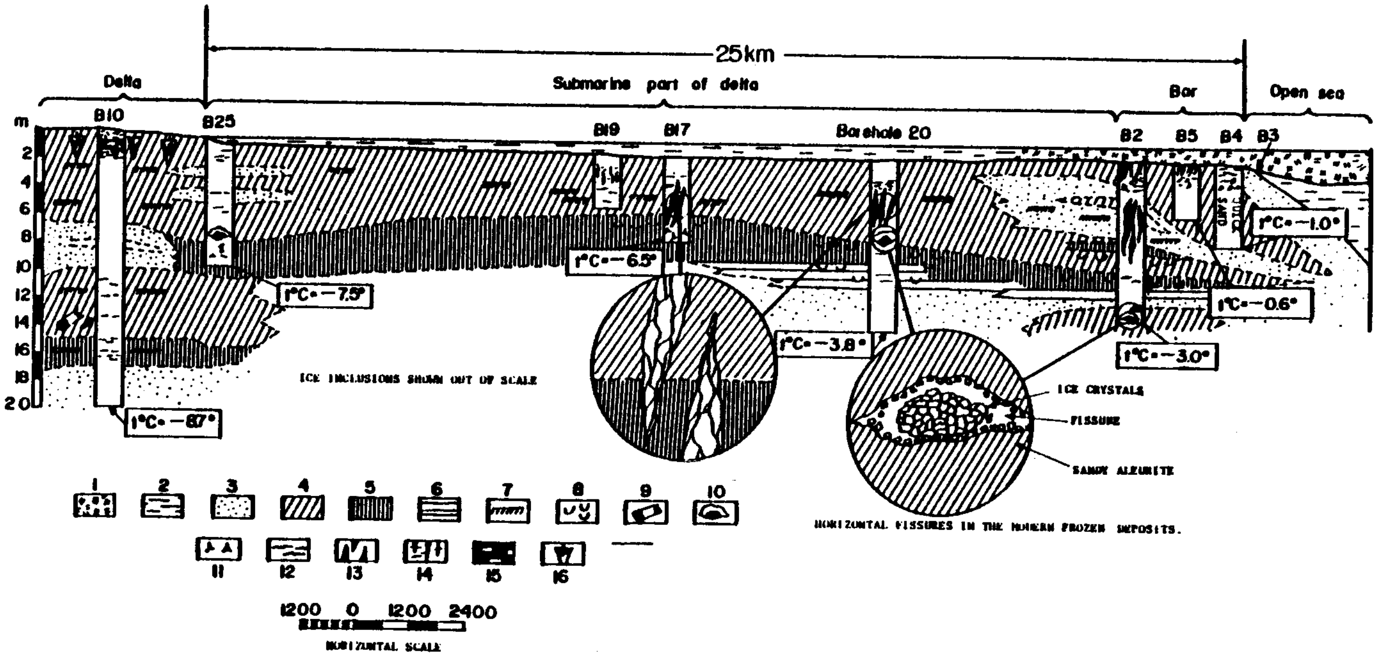
Depth (m)	Borehole and Observation Date					
	10 17–22/7	16 13/8	25 16/8	17 30/7	20 4/8	2 8/5
1	−3.4	−3.4	–	–	–	–
2	−7.0	−7.0	−5.6	−3.6	−3.1	–
3	−9.0	−9.4	−9.4	−4.1	−3.1	–
4	−9.6	−9.8	−10.2	−6.5	–	−4.0
5	−10.0	−10.2	−9.9	–	–	−4.0
6	−10.8	−10.5	−10.2	–	−2.0	−3.5
7	−9.7	−10.1	−9.4	–	−4.1	−3.2
8	−9.5	–	−8.3	–	−3.5	−3.2
9	−9.0	–	–	–	−3.7	−3.0
10	−8.4	–	–	–	−5.2	−3.0
11	−9.3	–	–	–	−4.6	−3.0
12	−8.9	–	–	–	−3.6	–
13	−9.0	–	–	–	–	–
14	−8.6	–	–	–	–	–
15	−8.6	–	–	–	–	–
16	−9.1	–	–	–	–	–
17	−8.6	–	–	–	–	–
19	−8.6	–	–	–	–	–
20	−8.7	–	–	–	–	–

Note: Borehole 10 was drilled on the surface of the low floodplain of Pridel'tovy Island, borehole 16 on the marine periodically inundated shore of this island; boreholes 25–17–20 under a water layer with a thickness of 0.8 m; borehole 2 was drilled early in May with sea ice at a thickness of 0.8 m.

"zone of the negative temperature," not "permafrost," emphasizing that the deposits are not ice bonded in spite of their very low temperature. Usually they explain this phemomena by hydrological, especially hydrochemical, conditions of the layers and water (Table 9).

One of the first attempts to show the complicated structure and position of the submarine permafrost in connection with its depth and the sea depth was made by Zyukov et al. in 1953 for the Ob Bay area in the Kara Sea (near Ust Port). In Figure 15 we can see some of the divisions of submarine permafrost. The seasonally freezing and cooled deposits lay directly on the bottom of the bay. Their thickness (about 1 m) decreases sharply, when the sea depth reaches 2 m. The authors divide permafrost into (1) seasonally freezing and cooled layers; (2) those separated from the seasonal layers; (3) those separated; and (4) "pereletok" (short-term permafrost).

The thickness of permafrost in the Kara Sea near Amderma and Vaigach Islands reaches 100 m at the water depth of 4 m, at 5 km offshore, and 60 m at the depth of 15 m, at 8 km offshore (Figure 16).



1. Sea Ice
2. Water
3. Sand
4. Sandy aleurite
5. Clay aleurite
6. Silt
7. Plant detritus
8. Shells
9. Wood
10. Siltlike voids with lengths of 10–30 mm
11. Vertical tubular voids with widths of 1–3 mm
12. Ice cement, ice nest
13. Ice intercalations and lenses with a thickness of 1 mm to 2 mm, vertical veins of granular ice
14. Vertical veins of ice with horizontal intercalations and lenses of ice
15. Basal texture
16. Secondary vein ice

Figure 7.—Permafrost-geological cross section in the region of the beach area at the mouth of the Yana River. After Ivanov (1969).

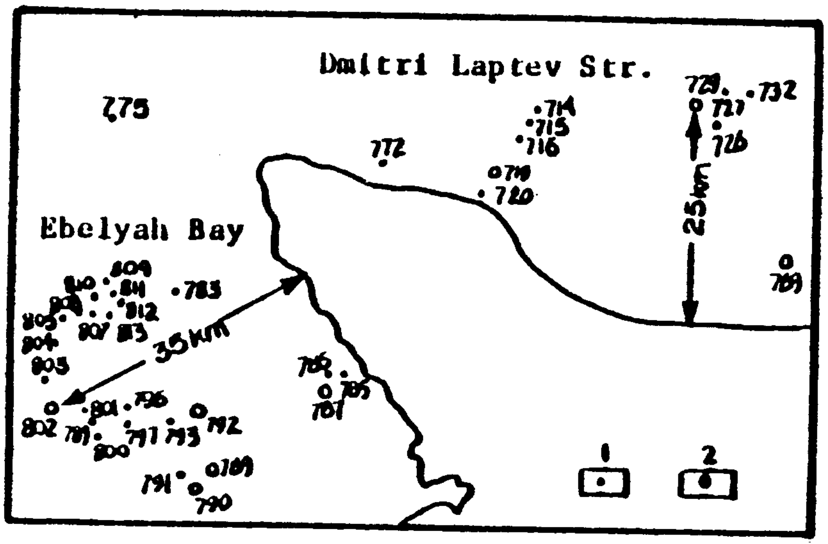
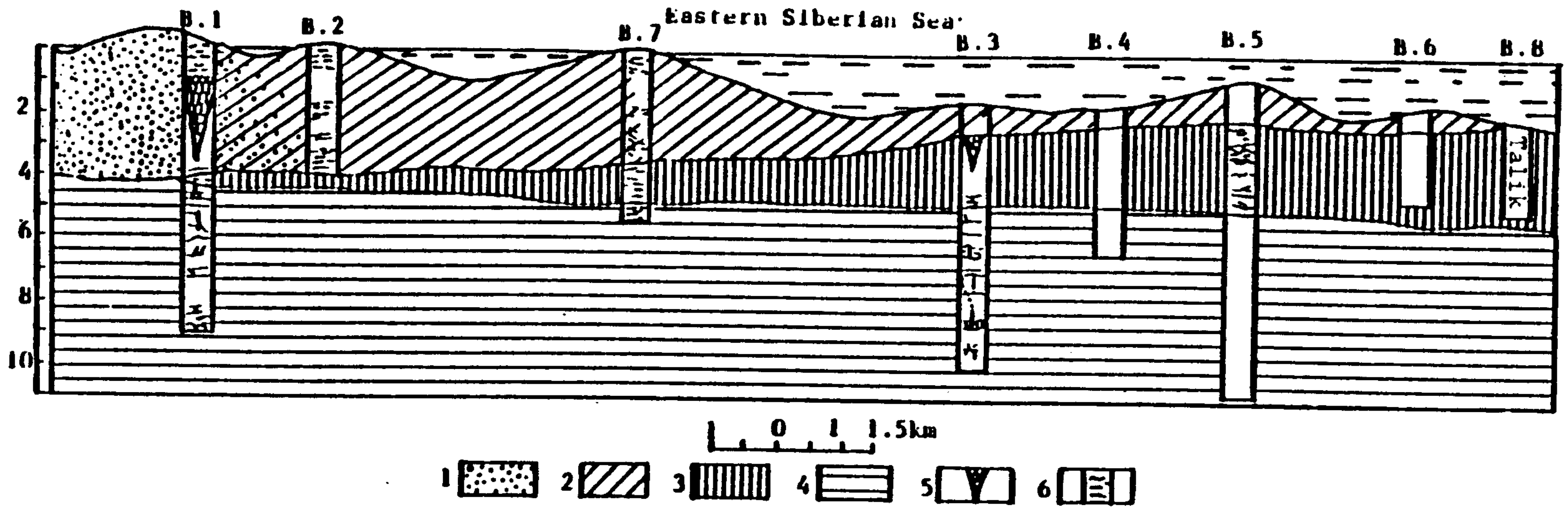


Figure 8.—Areas of the Laptev Sea shelf with unsalted deposits. After Molochushkin (1973).

In the Laptev Sea (Kojevnikov Bay) the thickness of the permafrost at the distance of 3 km offshore is more than 66 m and the permafrost body is separated by several layers of unfrozen rocks (Figure 17). In the eastern part of the sea the surface of the bonded permafrost becomes lower from 35 m to 2 km to 85 m at 10 km offshore (Figure 18). The borehole reached a big lense of buried ground ice at a depth of 86 m. The geologists did not give any data about submarine ice thickness in this area, but lenses of ground ice in submarine conditions in the Eurasiatic shelf of the Arctic Ocean are the usual phenomena. For example, in the Kara Sea near Amderma the ice lense was reached at a depth interval of 28.55 to 42 m in the "ancient valley" continuing from the continent to the shelf. Figure 19 shows the position of this lense; Figure 20, the temperature data for the same lense.

The almost universal presence of underground ice, which varies in form, dimensions and origin, is the most characteristic peculiarity in the structure of permanently frozen Quaternary deposits, developed on the seacoast. It has been established that this underground ice, of considerable thickness, is not buried or fossil remnants of valley glaciers or continental glaciation, but is an entirely independent formation having a water origin. According to the classification formulated by P. A. Shumskyi (1955), all the main ice formations present in the upper layers of the frozen stratum can be classified as the ice cement of frozen rocks, segregation ice and secondary vein ice. It is also possible to differentiate ice forming during the burial of snow "pereletoks" and drifts, during the freezing and burial of floodplain lakes and other variants of underground ice.



1. Sand 2. Sandy aleurite 3. Clay aleurite 4. Silt 5. Vein ice 6. Ice textures

Figure 9.—Schematic permafrost-geological cross section of the region of beach area at the mouth of the Indigirka River. After Usov (1965).

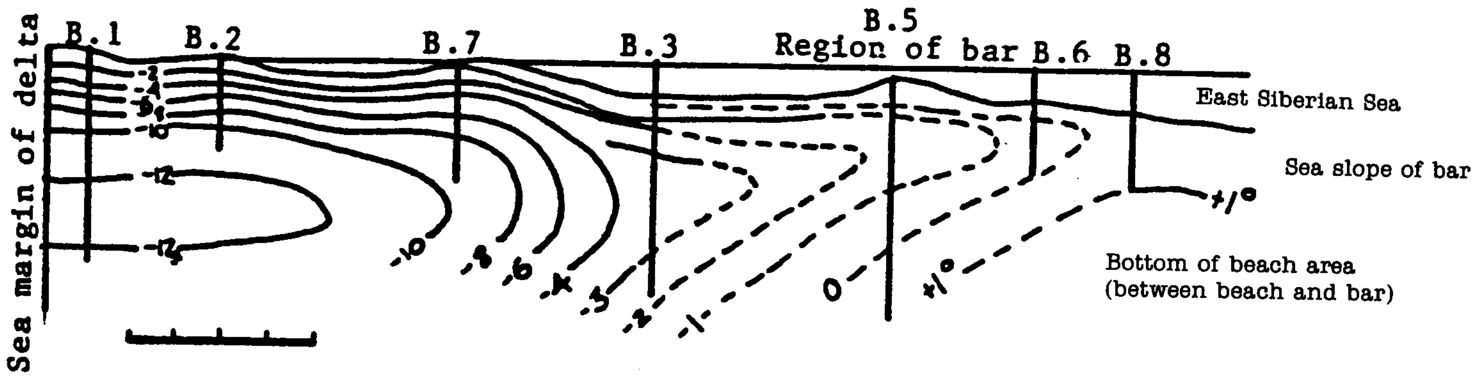
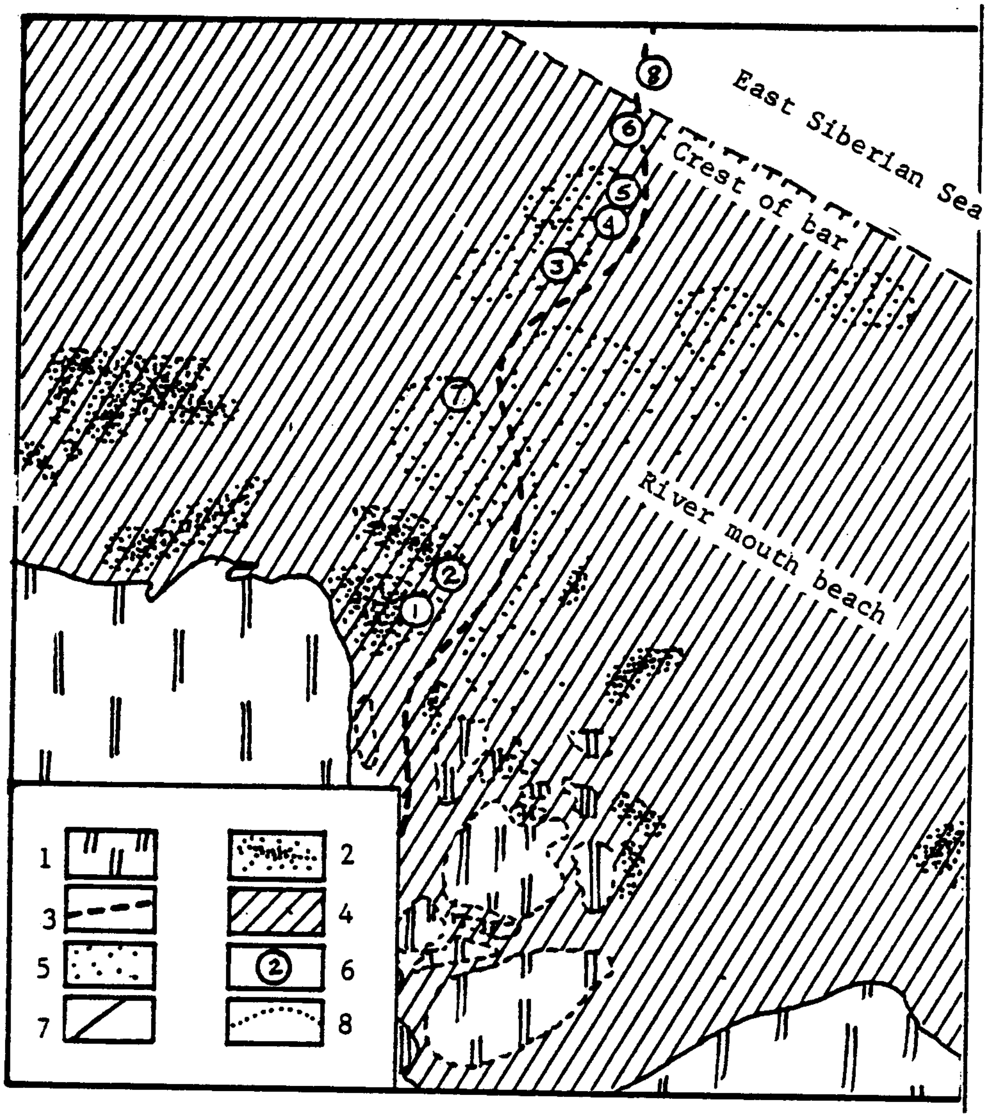
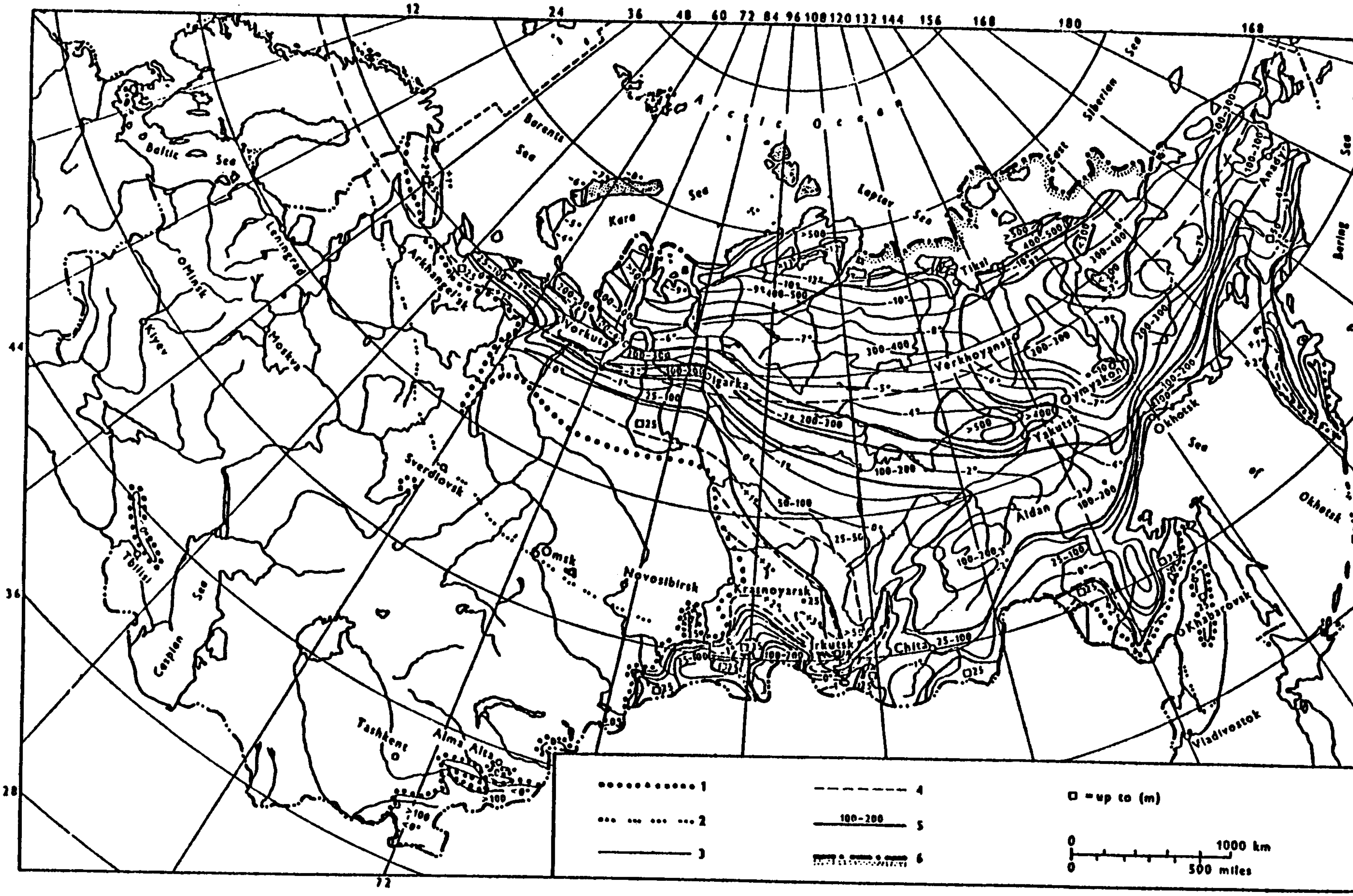


Figure 10.—Schematic temperature cross section of underwater layer of perennially frozen deposits on the bottom of the beach area near the mouth of the Indigirka River. After Grigor'yev (1966).



1. Holocene delta permanently frozen deposits
2. Above water, periodically inundated shoals—"sands"
3. Underwater channel and strip of talik beneath channel
4. Shelf delta-area underwater perennially frozen deposits
5. Underwater permanently water-covered shoal—"sands"
6. Boreholes
7. Shoreline of periodically flooded islands
8. Boundary of underwater shoals

Figure 11.—Schematic permafrost-geological map of the river mouth beaches of the shore of the East Siberian Sea (Grigor'yev 1966).



1. Boundary of permafrost area
2. Boundary of zone of frequent short-term permafrost
3. Minimum ground temperature at level of zero annual amplitude (in mountainous regions shown for valleys)
4. Soil isotherms at 1–2 m depth under natural conditions
5. Maximum thickness of permafrost (m)
6. Permafrost zone under Arctic Ocean

Figure 12.—Permafrost in the Soviet Union (Baranov 1960).

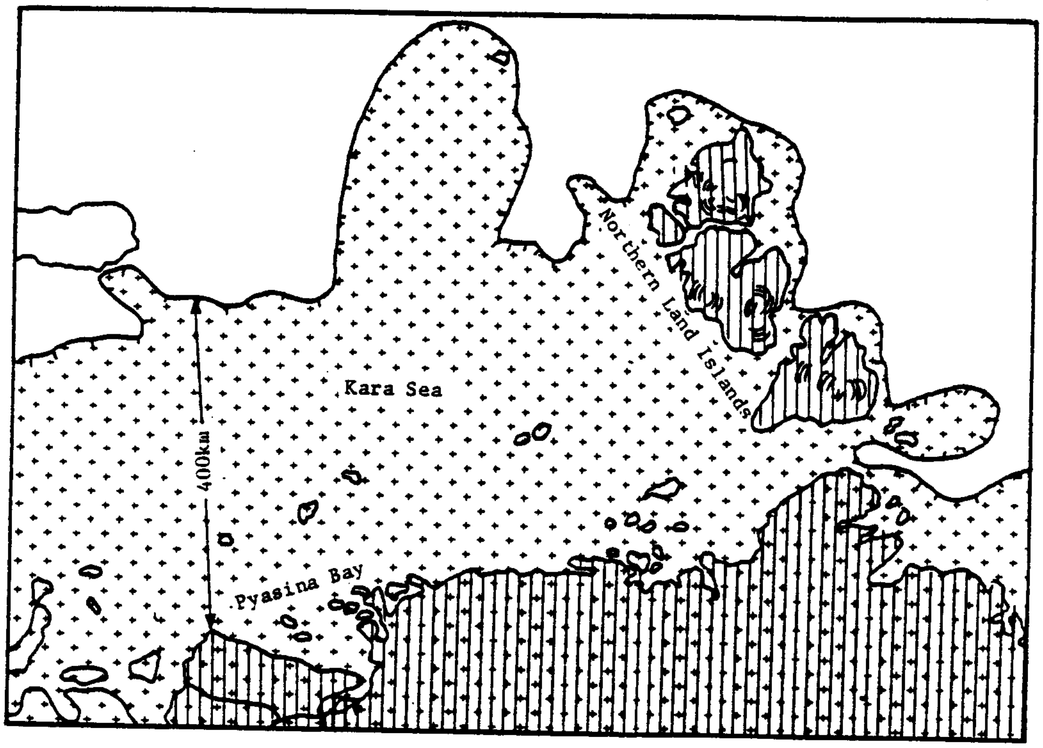


Figure 13.—Boundary of submarine permafrost in the Kara Sea.
After Fotiev (1974).

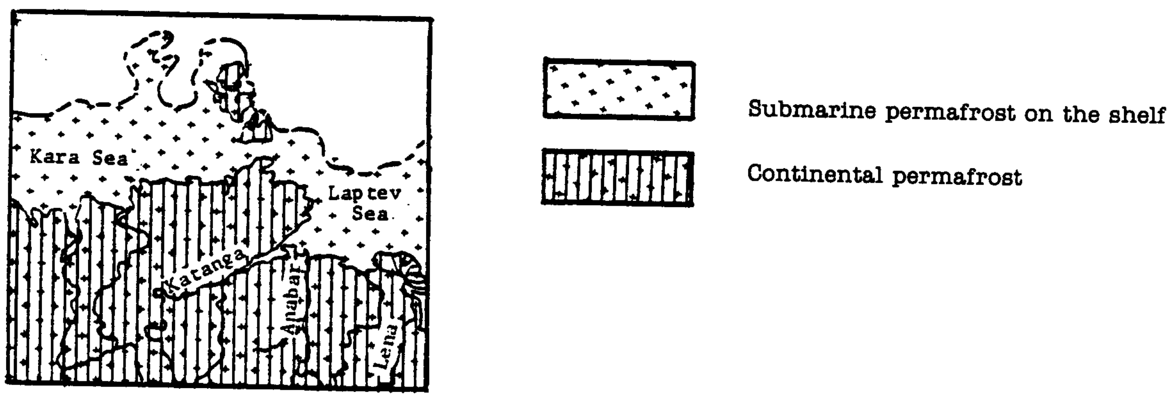


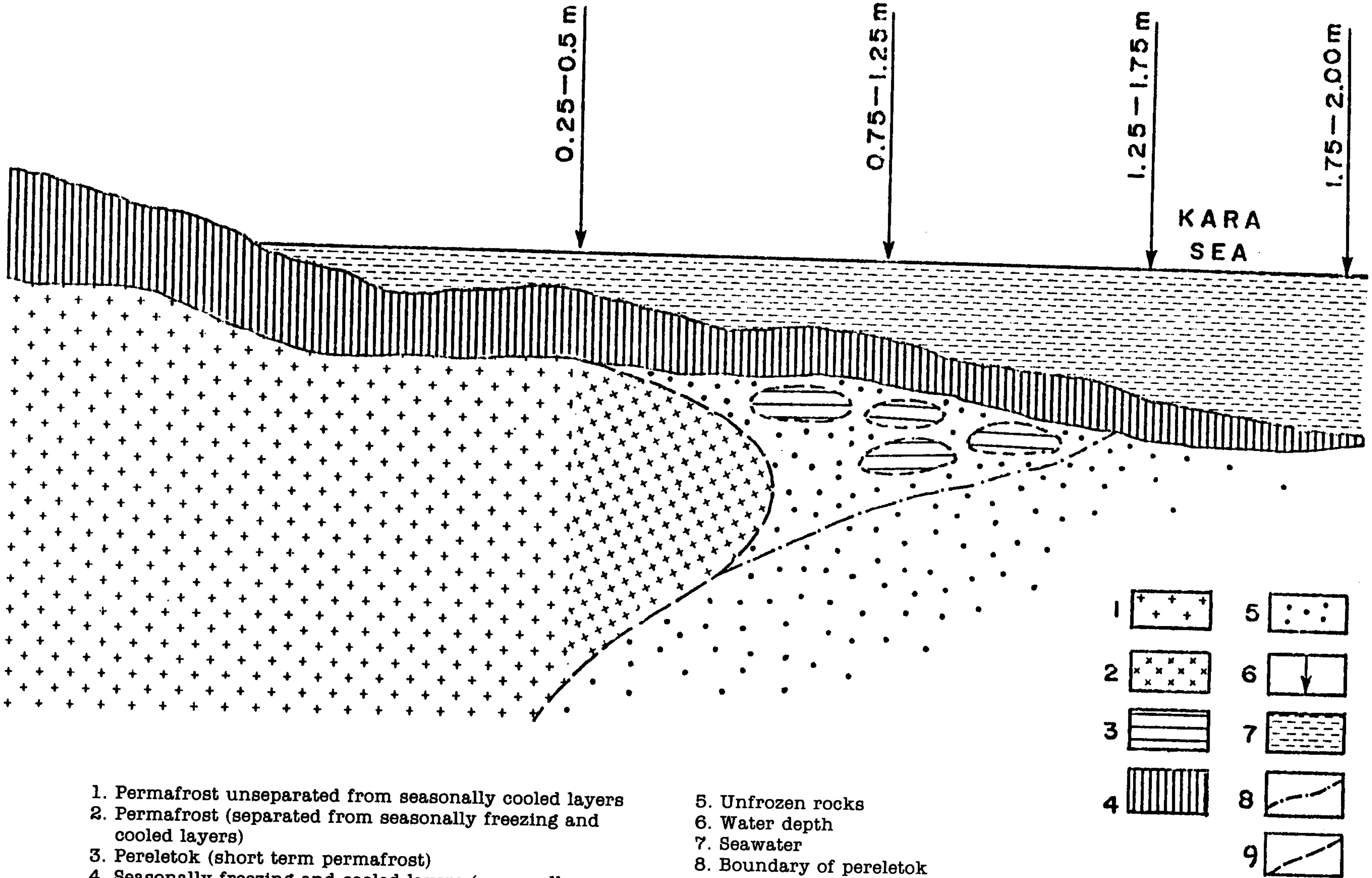
Figure 14.—Submarine permafrost extension in the Kara and Laptev seas.
After Fotiev (1974).

Table 9.—Freezing temperature of seawater of different salinity (N. N. Zubov, 1944).

Salinity (ppt)	Freezing Temperature of Seawater (°C)
0	0.0
10	−0.5
20	−1.1
30	−1.7
40	−2.2
50	−2.8
60	−3.4
70	−4.1
80	−4.8
90	−5.6
100	−6.4
110	−7.1
120	−8.0
130	−8.8
140	−9.8
150	−10.5

In the considered regions, those underground ice formations of greatest volume are the secondary vein ice formations. The general features of the mechanism for the formation of ice veins can be represented in the following form. In the case of great vertical temperature gradients in the upper horizons of the stratum of permanently frozen rocks frostlike tracks are formed. The annual repetition of cracking in the permanently frozen layers and the freezing of the water penetrating into the frost clefts, appearing in one and the same place, leads to the formation of the large frost vein. Horizontally, such ice veins usually form a polygonal lattice. As a rule, the cracking of the permafrost layer occurs at the depth greater than the thickness of the seasonally thawing layer and therefore the ice veins forming during winter do not thaw during the summer. Intensive frost cracking on the coast occurs primarily in highly icy peat and silty ground. The depth of the frost clefts can be different (from 1–4 m), depending on the thickness of these horizons and on the nature of the underlying rocks.

The most favorable conditions for the growth of secondary vein ice are the conditions of a floodplain regime. Here the surface horizons are always greatly moistened, peaty, and silted. In addition, under floodplain regime conditions, the growth of secondary vein ice can occur syngenetically; that is, simultaneously with the accumulation of precipitation, and thus ice veins can grow both in width and height. Secondary vein ice can grow both syngenetically and epigenetically; that is, after there has been an accumulation and freezing of the entire



1. Permafrost unseparated from seasonally cooled layers
2. Permafrost (separated from seasonally freezing and cooled layers)
3. Pereletok (short term permafrost)
4. Seasonally freezing and cooled layers (seasonally thawing in the coastal area above sea level)
5. Unfrozen rocks
6. Water depth
7. Seawater
8. Boundary of pereletok
9. Boundary of permafrost

Figure 15.—Schematic cross section of the permafrost in the Ob Bay (Kara Sea) coastal zone. After Zyukov (1953).

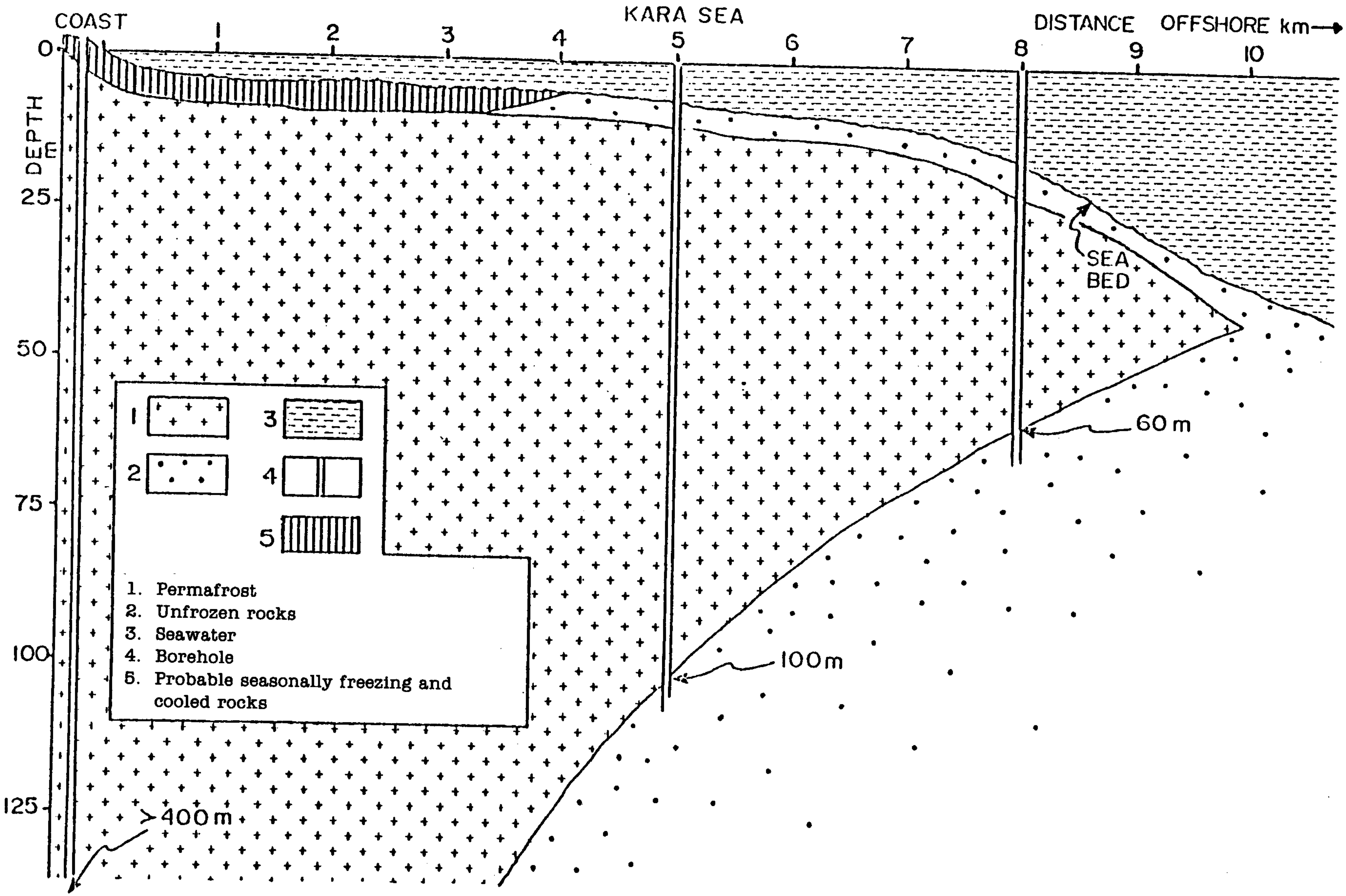


Figure 16.—Schematic cross section of bonded permafrost position and thickness in the areas of the Vaigach Islands and Amderma. After Vittenburg (1940) and Neizvestnov (1973).

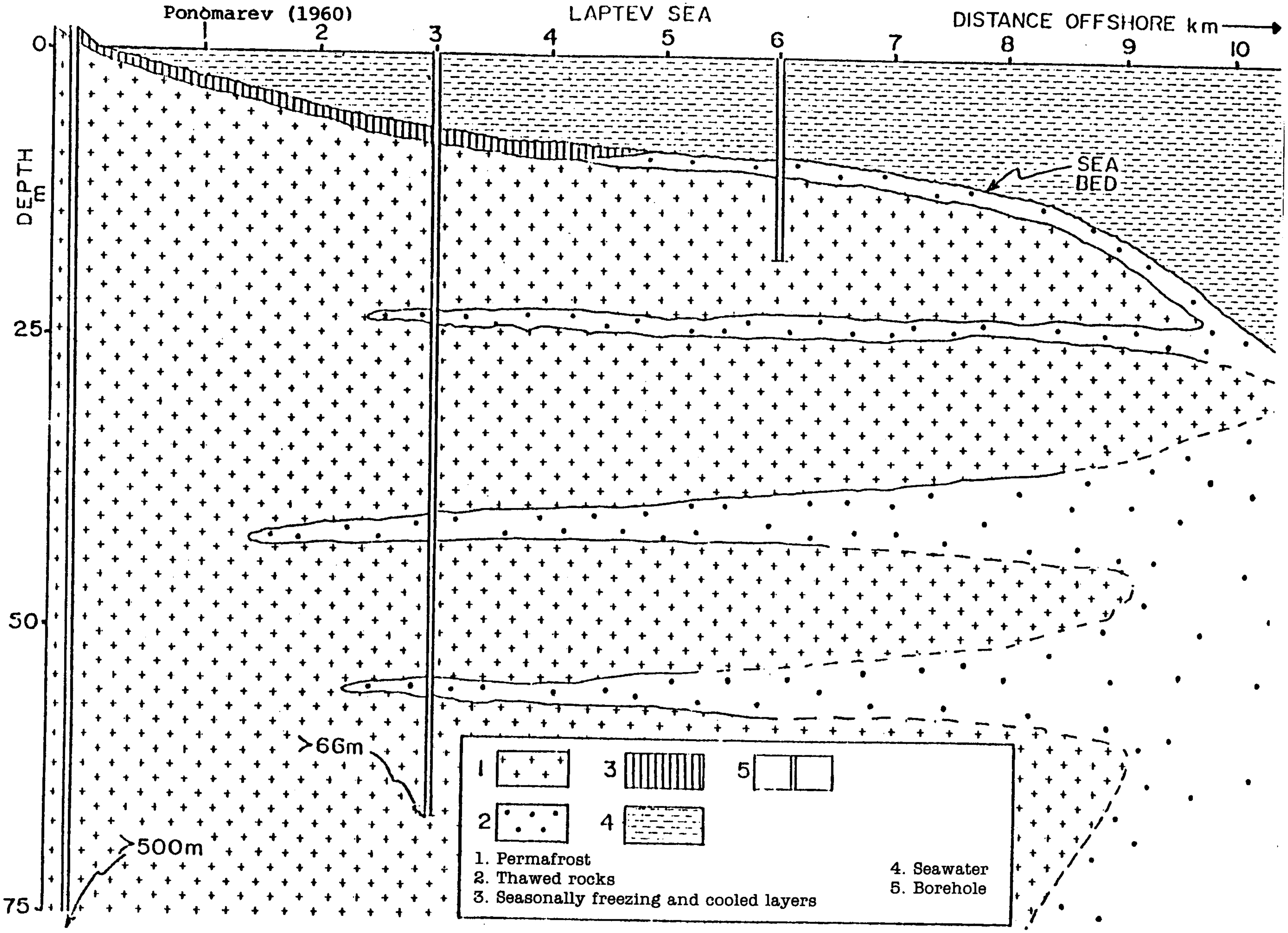


Figure 17.—Schematic cross section of submarine bonded permafrost position and thickness in Kojevnikov Bay. After Ponomarev (1960).

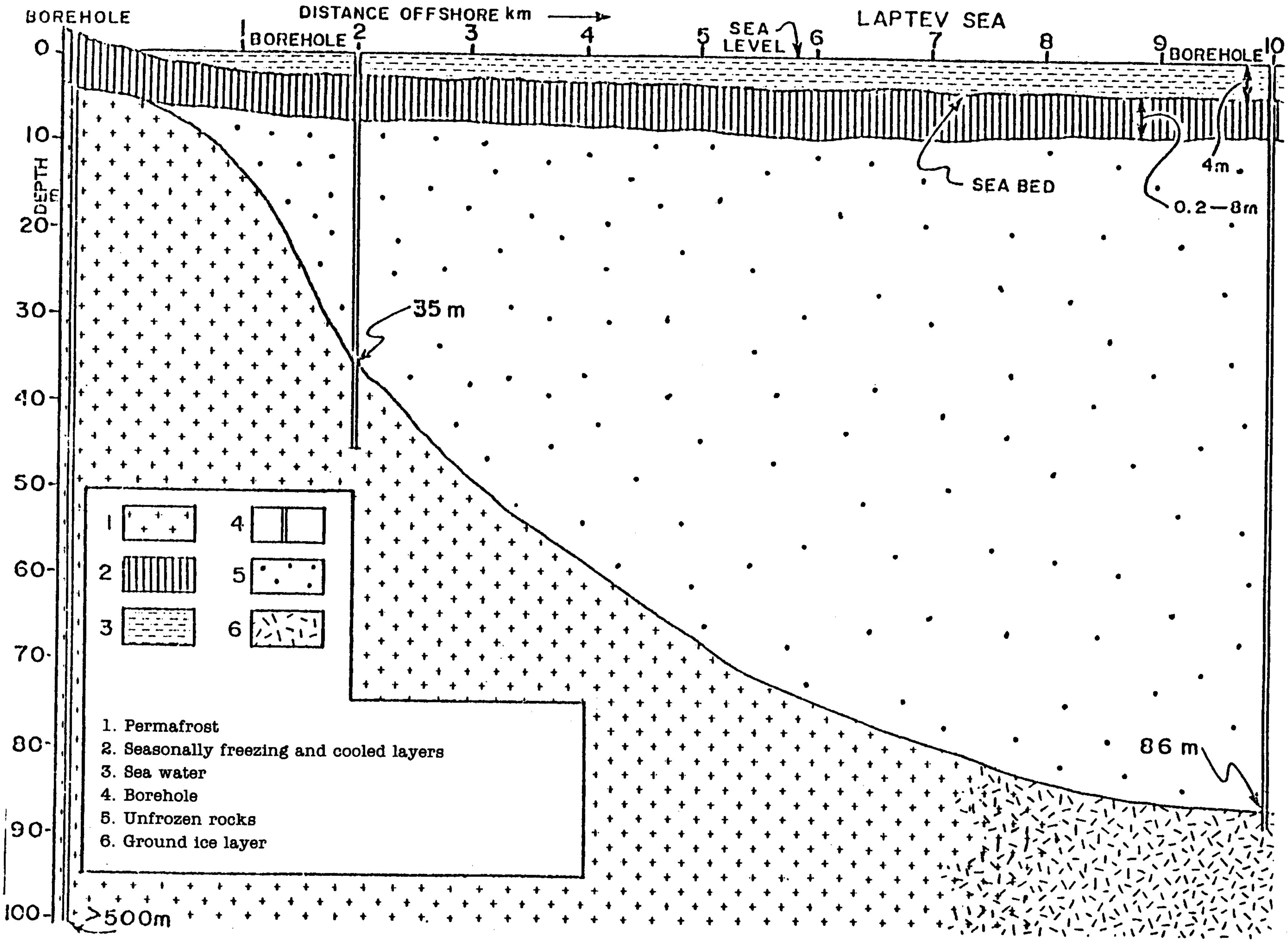
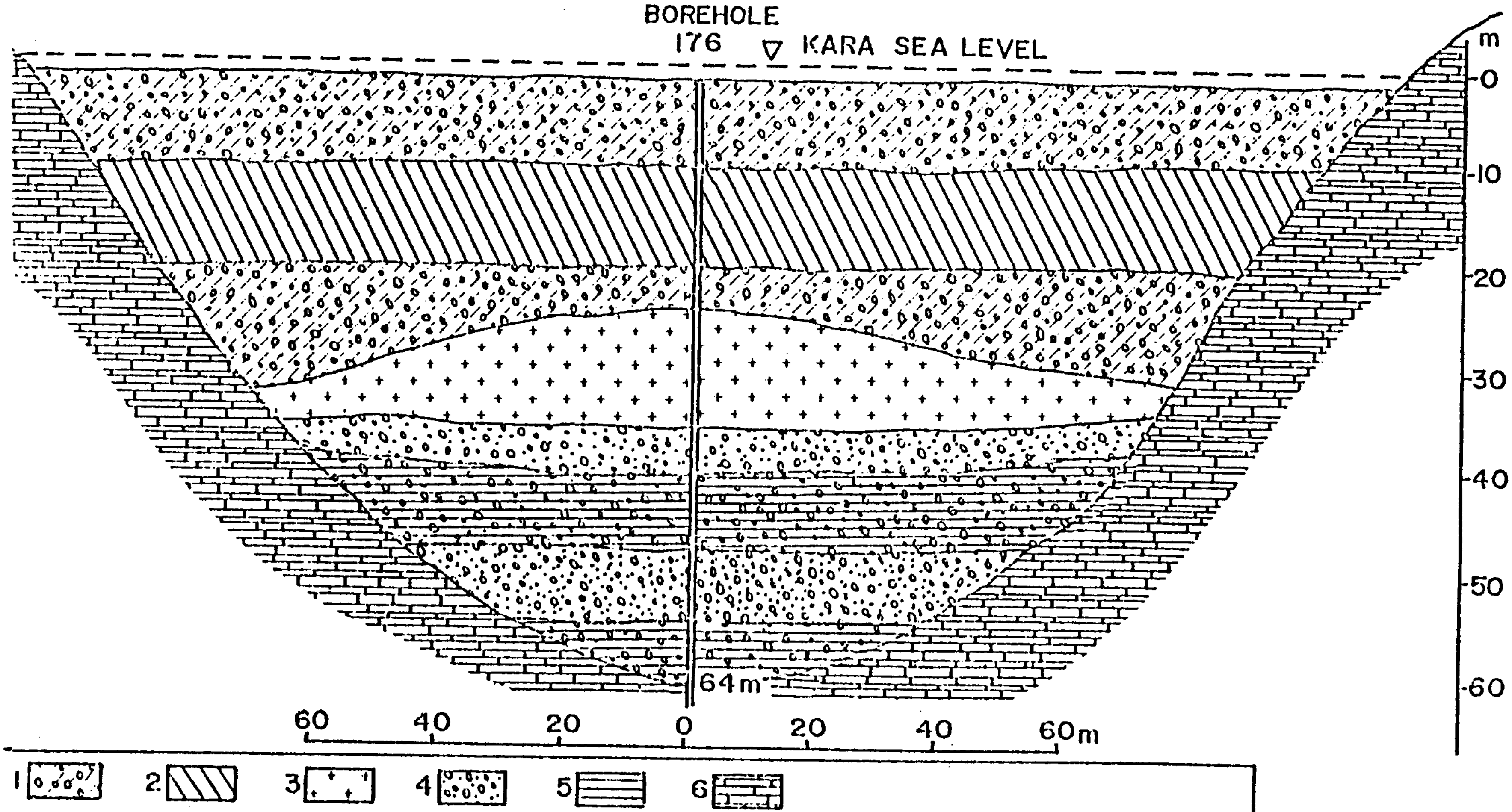
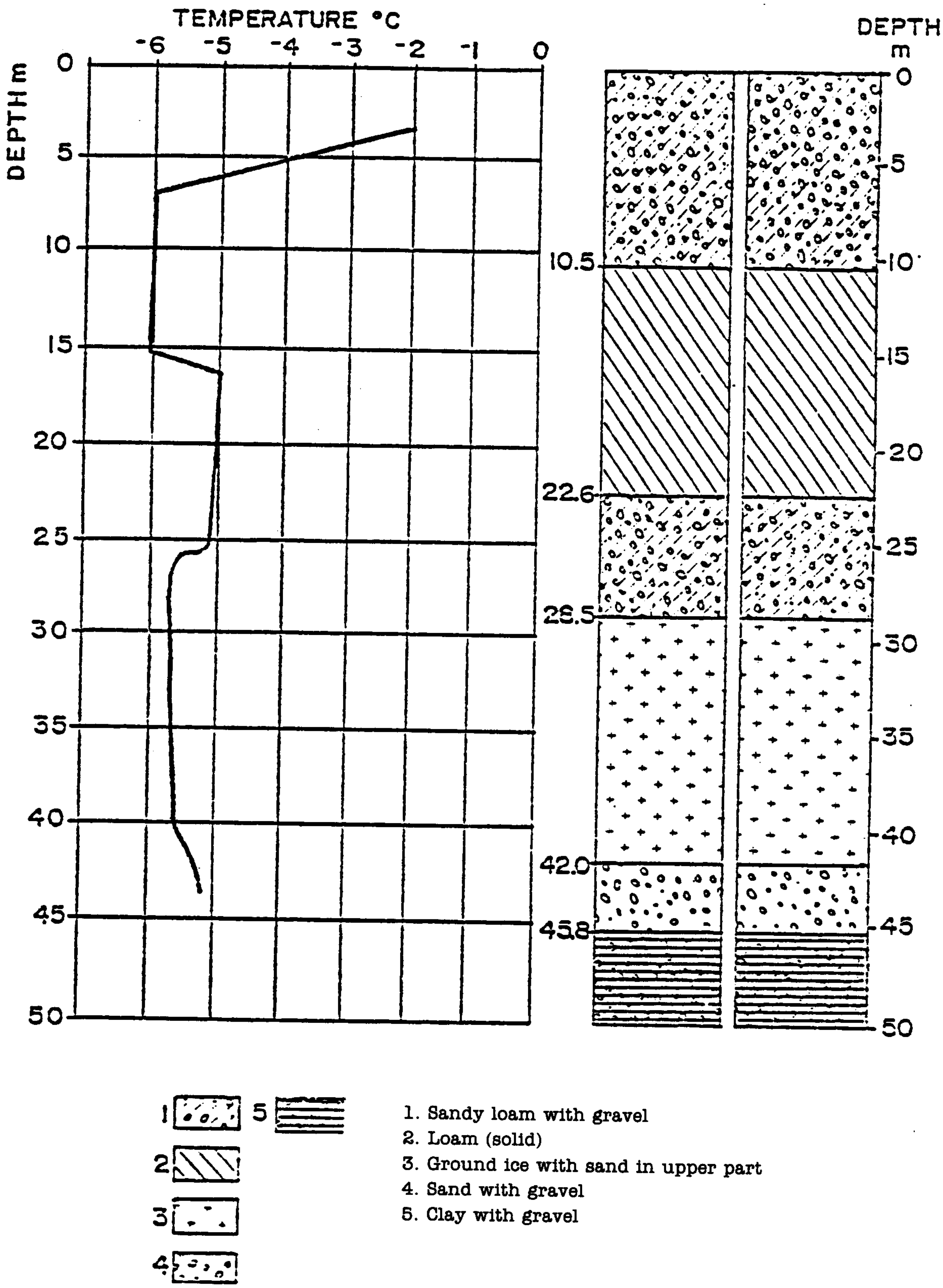


Figure 18.—Bonded permafrost in the eastern part of the Laptev Sea. After Zhigarev and Plakht (1974).



1. Sandy loam with gravel
2. Loam (solid)
3. Ground ice with sand in upper part
4. Sand with gravel
5. Clay with gravel
6. Limestone

Figure 19.—Geological cross section of the "ancient valley" inundated by seawater in Amderma (Kara Sea). After Ponomarev (1960).



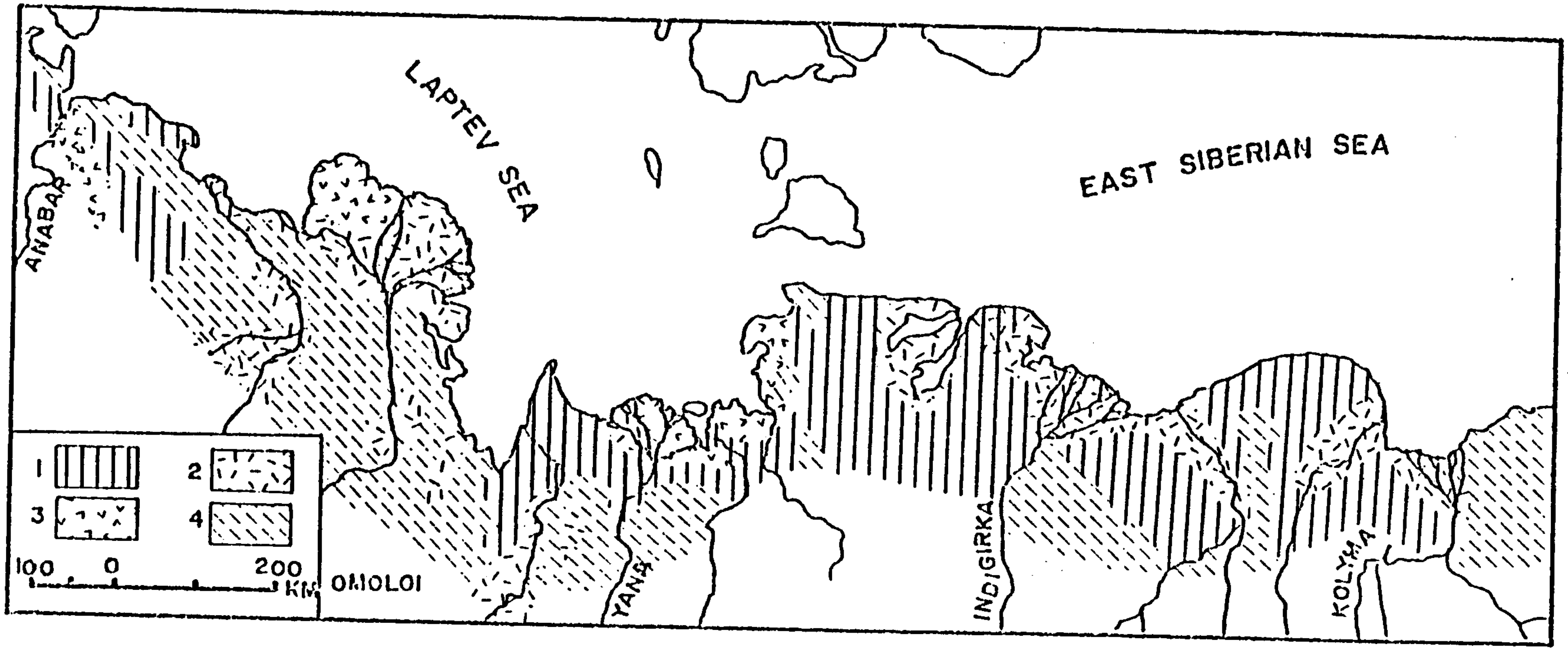
1. Sandy loam with gravel
2. Loam (solid)
3. Ground ice with sand in upper part
4. Sand with gravel
5. Clay with gravel

Figure 20.—Temperature of the ground ice in the Kara Sea. After Ponomarev (1960).

stratum of deposits. The most distinguishing characteristic of secondary vein ice is vertical stratification caused by the inclusion of mineral and plant admixtures entering the frost clefts with the water. Another of the distinguishing characteristics of the structure of secondary vein ice is the fine air bubbles dispersed in the ice mass. In secondary vein ice the air bubbles are rounded, elongated, or slightly dendritic in form, in contrast to a true cylindrical or acicular shape in river and lake ice. The size of the air bubbles in the ice varies from 0.5 to 3 mm. In the coastal zone of the alluvial-lacustrine plains, N. G. Grigor'yev (1966) discriminated the following regions with different ice saturation of the disperse deposits (Figure 21):

1. Regions with primary development of ancient thick secondary vein ice (the ice is associated with remnants of ancient alluvial plains). The dimensions of the ice veins are up to 10 m in width to 40–50 m in height.
2. A region with primary distribution of an intermediate density of the network of ice veins (the ice is associated with alassy depressions and floodplains). The dimensions of the ice veins are up to 2–3 m in width and up to 10–12 m in height.
3. A region with primary distribution of a thin network of thin secondary ice veins. The dimensions of the ice veins are up to 1–2 m in width and up to 3–5 m in height.
4. A region of primary distribution of bedrock outcrops in which there are no large formations of underground ice.

The thickest and oldest ice veins are encountered in deposits of an ancient alluvial plain. This ice is not developed at the present time. The study of secondary vein ice in shore scarps of the coastal plain revealed that the ice veins have considerable thickness. In width they attain 10 m and are locally traced to a depth of more than 30 m. In many cases the ice veins occupy about 60–70% of the volume of the entire ground mass. In its coastal form the ice veins frequently are dependent on the angle at which these veins are cut by the shore. If in their strike the ice veins are directed perpendicular to the shore, they usually have a rather regular form of narrow wedges or vertical columns. In those cases where the ice veins (especially intersecting ones) are cut by the shore at an angle, their form is more complex. Thus, in natural shows, as a result of intensive melting and destruction of the shore scarp, the form of the cross sections of the ice veins can change very rapidly, even in the course of a single summer season. The depth of the upper parts of the secondary vein ice is not everywhere the same. This depth varies from 0.7 to 3 m and the depth of the upper parts of the ice veins is usually greater than the depth of summer thawing. In general, the upper parts of the ice veins have a mostly even surface as if they had been melted. The lower parts of the ice veins, in rare cases, are clearly visible in the shows; more frequently they are covered by land slips or extend deeper than the bottom of the shows.



1. Regions with primary development of ancient thick secondary vein ice associated with remnants of ancient alluvial plains (dimensions of ice veins up to 6–8 m in width and up to 40–50 m in height.
2. Regions with primary distribution of intermediate density of network of ice veins associated with floodplains and alassy (dimensions of ice veins up to 2–3 m in width and up to 10–12 m in height.
3. Regions with primary distribution of thin networks of thin secondary vein ice associated with remnants of sandy coastal marine plain (dimensions of ice veins up to 1–2 m in width and up to 3–5 m in height.
4. Regions with primary distribution of bedrock outcrops.

Figure 21.—Schematic map of distribution of disperse deposits with different ice saturation in the coastal zone of the Yakutia. After Grigor'yev (1966).

Ice veins are also encountered on the bottom of alassy basins forming as a result of the melting of thick masses of ancient secondary vein ice. On the even surface of the alassy with thick peat bogs and a dense mossy cover extremely favorable conditions exist for frost cleft formation and the formation of secondary vein ice. Some of these ice veins were formed at the time of draining of the lake and others are forming during the present time. The ice veins in the alassy are usually thin and have a regular wedgelike form. The depth of the upper surface of the ice veins generally does not exceed the depth of seasonal melting of the ground, which is from 20 to 40–50 cm. In the sandy deposits of the coastal marine plain, the secondary vein ice is developed and is characterized by a thinness. The upper parts of the ice veins are usually found at a depth of 0.5–0.6 m; their width is 0.5–1.0 m. They are traced most frequently to a depth of 1.0–1.5 m. In the sandy deposits of the coastal marine plain secondary vein ice is extensively developed only in places which have peaty, silted and glazed horizons. Secondary vein ice is developed very extensively in deposits of the present-day floodplains. The upper parts of the ice veins are usually encountered at a depth of 0.5 m. The width of the ice veins is 1.2 m to, in rare cases, 4 m. The ice veins extend vertically 1–3 m and, in rare cases, 8 m.

On the coastal zone of Yakutia the development of the ice in deluvial deposits can frequently be judged from the forms of clumplike microrelief which are extensively developed on the slopes of the bedrock masses.

Among the structural peculiarities of the layers of frozen disperse deposits, the geologists include the appearance of "ice tectonics." In the numerous scarps of the ancient alluvial plain and the scarps of the alassy and the floodplains, it is common to observe singular forms of folds and layered frozen ground along side contacts with the ice vein.

The clear separation of the layers of frozen ground, which is observed in the polygonal blocks between the ice veins, is caused in most cases by ice inclusions in the form of horizonal intercalations with a thickness of 1–3 cm, less frequently 10 and even 20 cm. This is the so-called "spurious stratification" caused by the inclusion of ice intercalations forming at the boundary with the seasonally thawing layer and not having anything in common with the primary stratification of the frozen ground. As a result of this clear stratification of the frozen ground, the flexure of the layers along the side contacts with the ice veins is clearly visible. A smooth flexure of the layers occurs as a result of the gradual growth of secondary vein ice. Despite the monolithic nature and density of the frozen ground, and as a result of the enormous lateral pressure, which increases in the width of the ice vein, the ice can become denser and can be bent into folds. The flexure of the layers thus cannot be regarded only as a result of nonuniform thawing within the polygons. The development of the polygons is caused primarily by growth of the ice veins, and the ridges forming on the surface of the polygons are a result of pressure exerted on the ground by the growing ice veins.

The slight smooth concavity of the ice intercalations in the polygonal blocks probably corresponds to the lower isothermal surface of the once-existing seasonally melted layer. However, deep "coffer" folds outlined by ice intercalations scarcely coincide with the position of the surface of the seasonally melted layer.

In Grigor'yev's opinion (1966), the position of the isothermal surface is caused by pressure from the growing ice vein (the direction of the ice intercalations along the lateral contacts with a vein is almost vertical). As a result of such pressure the smoothly warping ice intercalations are deformed, sometimes folded and, in many cases, bent back almost at right angles. The nature of the contact between the ice veins and the ground mass is evidence that the main masses of secondary vein ice in ancient and modern alluvial deposits of the coastal lowlands increased simultaneously with the accumulation of sediments in the floodplain regime. At the present time, intensive growth of syngenetic secondary vein ice is observed in extensive sectors of floodplains (this is graphically indicated by the presence of growing ridges on the surface of the polygonal floodplains). However, on the ancient alluvial plain the growth of ice veins is not noted at this time. In the coastal scarps of the ancient coastal plains, one also finds vertical thin veins of ice from 10 to 30 cm in thickness which have been traced to a depth of 3–4 m. The nature of ice stratification and the bedding of the surrounding ground indicated that these veins have an epigenetic origin. This kind of thin epigenetic ice vein is also formed in thick ancient syngenetic ice veins, but the growth there is evidently difficult. The age of the deposits on the coastal plain is indicated by the ancient age of the syngenetic secondary vein ice. If the onset of formation of the ancient coastal alluvial plain is related to the beginning of the Middle Pleistocene, the onset of freezing of the unconsolidated Quaternary deposits and the formation of thick underground ice is related to the same time. However, the main mass of epigenetic vein ice has a Holocene age. The broad development of secondary vein ice in all elements of Quaternary deposits is evidence that the conditions for the development and growth of ice were favorable during the course of the entire second half of the Quaternary period.

Simultaneously with the secondary vein ice of syngenetic and epigenetic origin, constitutional ice, scattered through the entire mass of frozen ground, developed extensively in the layers of permanently frozen ground in the coastal and shelf parts of the Arctic on the north of Eurasia. This ice is represented, for the most part, by ice cement and segregation ice. The first was formed as a result of the rapid freezing of moisture present in the ground pores and the second as a result of the segregation and freezing of moisture migrating in the ground pores toward the freezing front. This increased ice content constitutes the most characteristic peculiarities of the permanently frozen ground layers developed in the coastal shelf zone.

The thickness of the permafrost on the Arctic islands sometimes reaches 220 m, but in the submarine conditions around the islands, it usually decreases by 5–35 m at distances offshore of 10–15 km.

In general, the picture of the main features of submarine permafrost distribution in the Arctic seas looks as it is shown in Figure 22. Of course, this figure shows only the scheme of the relict Pleistocene and Holocene permafrost of the coastal plains, shelf and islands. The characteristics and distribution of recent permafrost, including submarine permafrost characterized by seasonal and partly multi-year cryogenic stages, will be considered later and will be shown in greater scale. In the illustration (Figure 22), we can see that the permafrost is ususlly not monolithic but often a discontinuous body, half of it being separated by unfrozen deposits. We can also see the trend in permafrost depth and thickness decreasing in the offshore direction, and the limits of these characteristics are typical for the Arctic shelf of Eurasia.

Cryogenic phenomena found in Pleistocene formations on the shelf provide evidence of the existence of permafrost, although the mere fact of their presence or absence in the cross section affords no basis for paleogeographic conclusions. In order to reconstruct the history of permafrost formation, to determine the time of freezing of the sediments whose age has been ascertained on paleontological and chronological data or by means of other methods, one must find convincing proof that a given cryogenic phenomenon developed together with accumulation of deposits, i.e., syngenetically. This is the prime condition. Furthermore, cryogenic phenomena do not occur everywhere, even under the most severe climatic conditions. They are characteristics of only definite sediments and particular facies. Sometimes horizons which lack any "traces of frozen ground" may therefore be erroneously attributed to a warmer period.

It seems very difficult to solve the problems relating to the syngenesis of cryogenic phenomena and their association with those or other formations, even if the physical substance of all these phenomena could be ascertained. The geologic–cryogenic regularities controlling the development of these phenomena can only be explained with the aid of adequate research methods. One of those methods is the frozen-ground facial analysis of Quaternary formations. It is justified by the following facts. In cross sections, properties of the sediments are studied as may provide distinguishing evidence of their genesis (mineral composition, stratification pattern, faunal and floral remains, etc.) and facies. At the same time such syngenetic cryogenic phenomena are identified that are already known to bear traces of permafrost action. In subaqueous sediments, they appear in two forms: either oblique or vertical cryogenic. In subaerial formations, the leading features are a stratified or striated cryogenic texture, as well as ice veins with irregular border contrasts. The thus ascertained syngenetic freezing of the sediments, occurring in the cross section and in a definite area, must be correlated

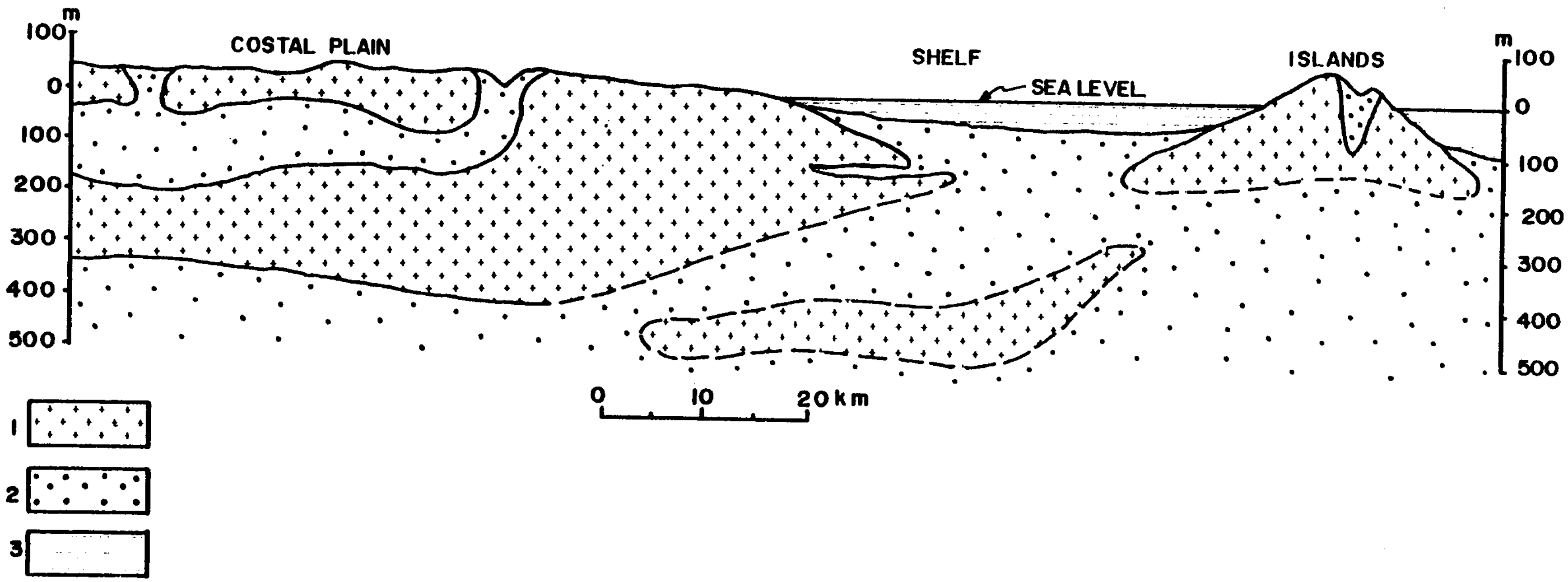


Figure 22.—Schematic cross section of submarine relict Pleistocene and Holocene permafrost distribution on the Eurasiatic part of the Arctic shelf.

with the facies from which cryogenic phenomena are absent or in which they are recorded by forms occurring in the active layer of permafrost.

Since the early fifties, some places of the Eurasiatic northern seas and (in particular) shallow-water coastal areas, were chosen as the "proving ground" for the comprehensive investigations of the layers of shelf deposits that had been transformed by cryogenic processes. In 1971 S. M. Fotiev* had proposed the term "subaquatic cryogenic stratum," of *SKT*, for such kind of layers; *SFL* for seasonally frozen layer; and *STL* for seasonally thawed layer. Some works were carried out by the expedition of the USSR Academy of Sciences and Moscow State University in the eastern part of the Vankina Gulf (Laptev Sea). In 1972 Ye. Katusonov and G. Pudov published the results of the cryolithological investigations of this area. In Figure 23 we see that permafrost is developed in most of the area under the sea ice. The depth of the permafrost here is more than 50 m. The area under unfrozen seawater is "talik." There are three lithological sorts of deposits under Vankina Gulf: silty sands, silty aleurites, and sandy aleurites. The authors divide all these deposits into two series. The upper one is wet and ice-saturated (45–70%). The lower one looks "dry" and very dense. Sands and aleurites of the lower series have the fissures filled by ice crystals. The material of the upper series is less dense, viscose after melting, and often flowing. There are many "broken" lenses of ice about 1.5 cm thick. They create a cross-bedded cryogenic structure and the small fissures are half filled by crystals of ice. The upper series is high in salts (chlorides and sulphates of magnesium, calcium, and sodium). The lower one has the fissures completely filled by ice crystals. This series is unsalted. Table 10 shows the moisture and salinity of the deposits for the different depths in this area. Because of the definite regularity of the ice content, the cryogenic structure distribution, and the properties of the deposits, the geologists concluded that differences in the two series could be explained with a cryolithological approach. They supposed that the more time the deposits had been in an unfrozen state and subjected to diagenesis and consolidation, the weaker and more monotonous the ice formation process must have been. On the contrary, the transition of the deposits from an unfrozen to a frozen state was fast, then these deposits were not subjected to serious changes and the cryogenic structures were developing much more intensively. The authors distinguished three types of permafrost in the sedimentary deposits: (a) *syngenetic type*, if the deposits were not changed by the final moment of the complete freezing. Their formation took place near the surface due to the influence of the permafrost basement. Typical cryogenic structures are usually

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* The article "Role of the Chemical Composition and Mineralization of Subterranean Waters in the Freezing Process . . ." *TR-PNIISA*, Vol. 11, 1971.

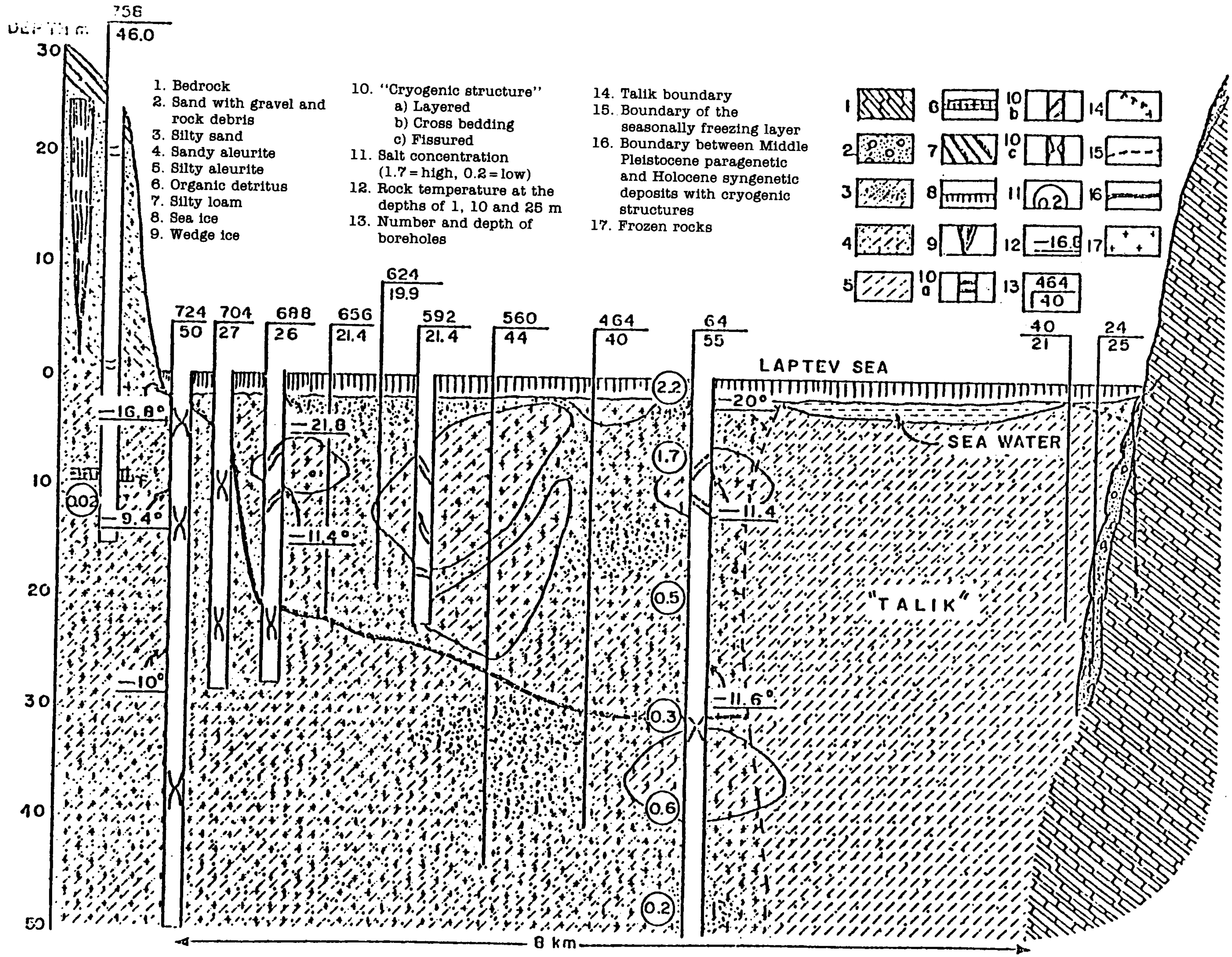


Figure 23.—Permafrost geological cross section of Vankina Gulf (eastern part of the Laptev Sea) (Katasonov and Pudov 1972).

Table 10.—Moisture and salt content in shelf deposits, April–May 1969, according to Katasonov and Pudov (1972).

Depth (m)	Moisture in % of Dry Weight	pH	Salt Content (g in 100 g)	Salinity
		Borehole 64		
1	22.0	5.5	2.245	High
3	25.6	7.2	0.757	Moderate
11	32.0	6.0	1.273	High
15	17.0	7.9	0.451	Weak
20	26.0	7.0	0.541	Moderate
25	28.0	7.26	0.320	Weak
35	29.6	6.0	0.606	Moderate
40	23.5	6.72	0.256	Non-saline
50	22.2	6.0	0.292	Non-saline
		Borehole 96		
1	19.4	6.38	1.421	High
8	36.8	7.22	1.160	High
25	23.6	7.94	0.411	Weak

layered for subaerial and cross-bedded for subaqueous deposits; (b) *parasyngenetic type*, if the deposits were formed in permafrost conditions but were in an unfrozen state for a long time. They had been consolidated and fissured; which is why the fissure cryogenic structures are typical for them; (c) *epigenetic type*, if the deposits had been formed before the permafrost conditions. These cryogenic structures are connected with tectonic dislocations and lithogenic cracks forming the large blocks.

Katsonov and Pudov suppose that the cryogenetic structures of the upper series of deposits (Figure 23) were formed syngenetically in the relatively salted Holocene basin. The lower one was formed in the Middle Pleistocene; it has the parasyngenetic cryogenic structures formed in the freshwater basin. We can see that in spite of the fact that both series are very similar lithologically, they are different in the cryolithological sense.

M. Ivanov's investigations (1969) of the cryogenic structure of coastal-delta perennially frozen deposits show the following:

(1) In the structure of most of the sectors of the beach areas of river mouths situated on the coast of the Laptev Sea and the East Siberian Sea, the principal components of this structure are coastal-delta deposits represented primarily by sands, sandy and clay aleurites, and also silts and clays. The structure of the upper layer of perennially frozen sandy and clay aleurites is characterized by siltlike voids with a width up to 1 cm and a length of 3–4 cm, whose formation is evidently associated with temperature stresses in the bottom sediments during their freezing. In these

same deposits one also finds obliquely and vertically arranged tubular voids with a width to 3 mm and a length up to 40 cm, associated with the vital functioning of mud-eaters (worms) moving about in the layer of thawed ground.

2. The moisture-ice content of coastal-delta and marine deposits varies in a rather wide range (Table 11). In determining the moisture content of ground taken from boreholes drilled in the bottom of the beach areas at the mouths of the Indigirka and Yana rivers, the geologists usually noted a decrease in the moisture-ice content of bottom deposits with depth, and only at individual horizons, to which the accumulation of segregation ice or ice cement is associated, is this tendency impaired.

Table 11.—Moisture content of coastal delta deposits (for typical boreholes) on the beach near the mouth of the Yana River (in % of dry weight. After Grigor'yev (1966).

Depth (m)	B10	B11	B16	B17	B19	B20	B25
0.5	–	56	39	–	28	49	73
1.0	–	229	61	–	41	58	50
1.5	–	172	52	74	45	52	89
2.0	–	32	76	25	77	144	48
2.5	–	29	55	44	52	53	26
3.0	–	35	56	31	42	35	33
3.5	–	33	33	28	56	52	20
4.0	–	27	58	30	55	42	26
4.5	–	37	49	28	–	35	40
5.0	–	142	46	29	–	58	43
5.5	–	40	43	33	–	45	30
6.0	–	27	48	50	–	30	47
6.5	–	26	31	–	–	34	27
7.0	–	24	47	–	–	57	26
7.5	–	47	–	–	–	31	40
8.0	–	23	–	–	–	33	39
8.5	27	24	–	–	–	30	38
9.0	27	25	–	–	–	18	–
9.5	25	24	–	–	–	19	–
10.0	26	22	–	–	–	21	–
10.0	35	–	–	–	–	23	–
12.0	36	–	–	–	–	22	–
13.0	34	–	–	–	–	25	–
14.0	35	–	–	–	–	–	–
15.0	66	–	–	–	–	–	–
16.0	40	–	–	–	–	–	–
17.0	60	–	–	–	–	–	–
18.0	21	–	–	–	–	–	–
19.0	22	–	–	–	–	–	–
20.0	22	–	–	–	–	–	–

3. Foredelta deposits are also characterized by the broad development of massive cryogenic textures, the formation of fissured and radial textures, and also intercalations and lenses of ice with broken outlines. A peculiarity of this complex of foredelta deposits is that they contain secondary vein ice forming during the freezing of the bottom deposits under the layer of ice covering them (Figures 7 and 9).

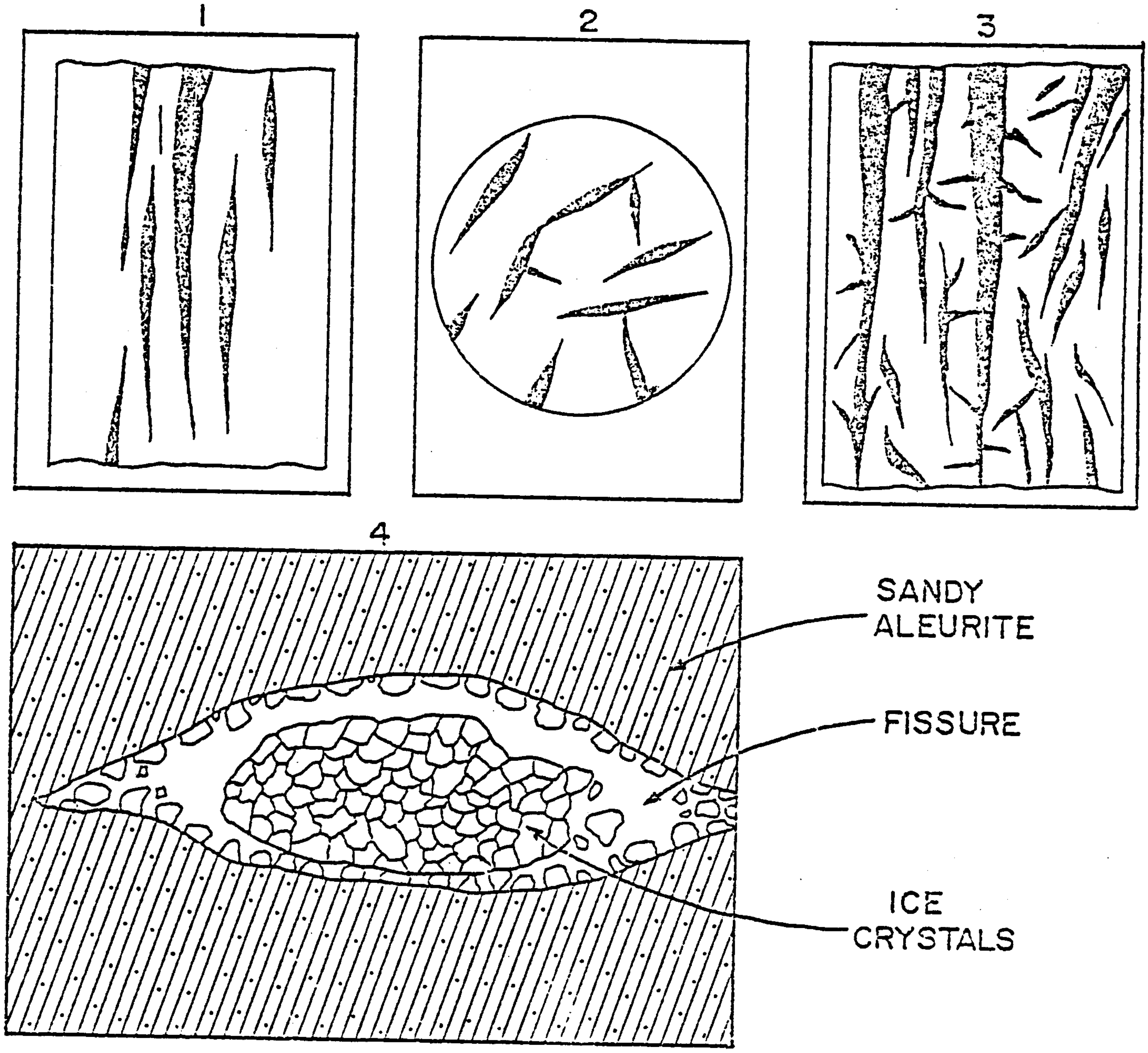
The silty deposits usually underlying the layers of sandy aleurites are characterized not only by ice cement, but also by individual schlieren and intercalations of ice with a granular structure. In many cases the horizontal ice intercalations have vertical offshoots which create an irregular rectangular grid.

The texture-forming ice consists of rounded grains similar to fish eggs measuring up to 1–2 mm, not firmly bound one to the other. The ice grains are usually covered by a dull whitish encrustation. In individual schlieren with voids at the center the granular ice in many cases makes up only the walls of the schlieren. The granular structure of the texture-forming ice which we encountered in the frozen silty deposits on the bottom of the beach area at the mouth of the Indigirka is evidently associated with the salinity of the bottom sediments. The tiniest particles of salt could serve as singular centers for the formation of individual grains of ice during the freezing of bottom sediments.

The broad propagation for texture-forming ice having a granular structure is also indicated by the fact that such ice is present in the frozen silty deposits lying on the floor of Siellyakh Bay in the Laptev Sea and on the bottom of the lagoon on Vil'kitskiy Island, situated in the Kara Sea (Figures 25 and 26). The cryogenic structure of the seasonally freezing layer in general is similar to the cryogenic structure of the upper layer of permanently frozen ground. In the case of syngenetic freezing, when with the accumulation of precipitation the upper surface of the permanently frozen stratum rises, the lower horizons of the seasonally freezing layer pass into a permanently frozen state. At the same time, the cryogenic textures forming in the seasonally melted layer are simultaneously preserved.

The formation of different types of cryogenic textures is dependent on the facies of the ground, the moisture content, and the nature of freezing. The following is the P. A. Shumskiy approach, N. F. Grigor'yev, 1966: (1) fused, characterized by the development in the ground pores of only very small formations of ice cement; (2) cellular or reticular, for which the formation of intersecting ice intercalations in the ground is typical; and (3) layered, clearly expressed intercalations of ice and ground.

Earlier, Ye. M. Katasonov (1960, 1962) had formulated a detailed classification of cryogenic textures for the principal genetic varieties of both seasonally and permanently frozen Quaternary deposits which can be used as well for the ground in the Eurasiatic coastal and shelf areas of the Arctic Ocean (Figure 27).



1. Vertical fissure ice in consolidated aleurite, longitudinal section.
2. Vertical fissure ice in consolidated aleurite, transverse section.
3. Cryogenic structure of radiating fissure ice in the silty sand and aleurite of the seasonally frozen bottom layer deposits.
4. Horizontal fissure in modern frozen deposits.

Figure 24.—Fissure ice. After Ivanov (1969).

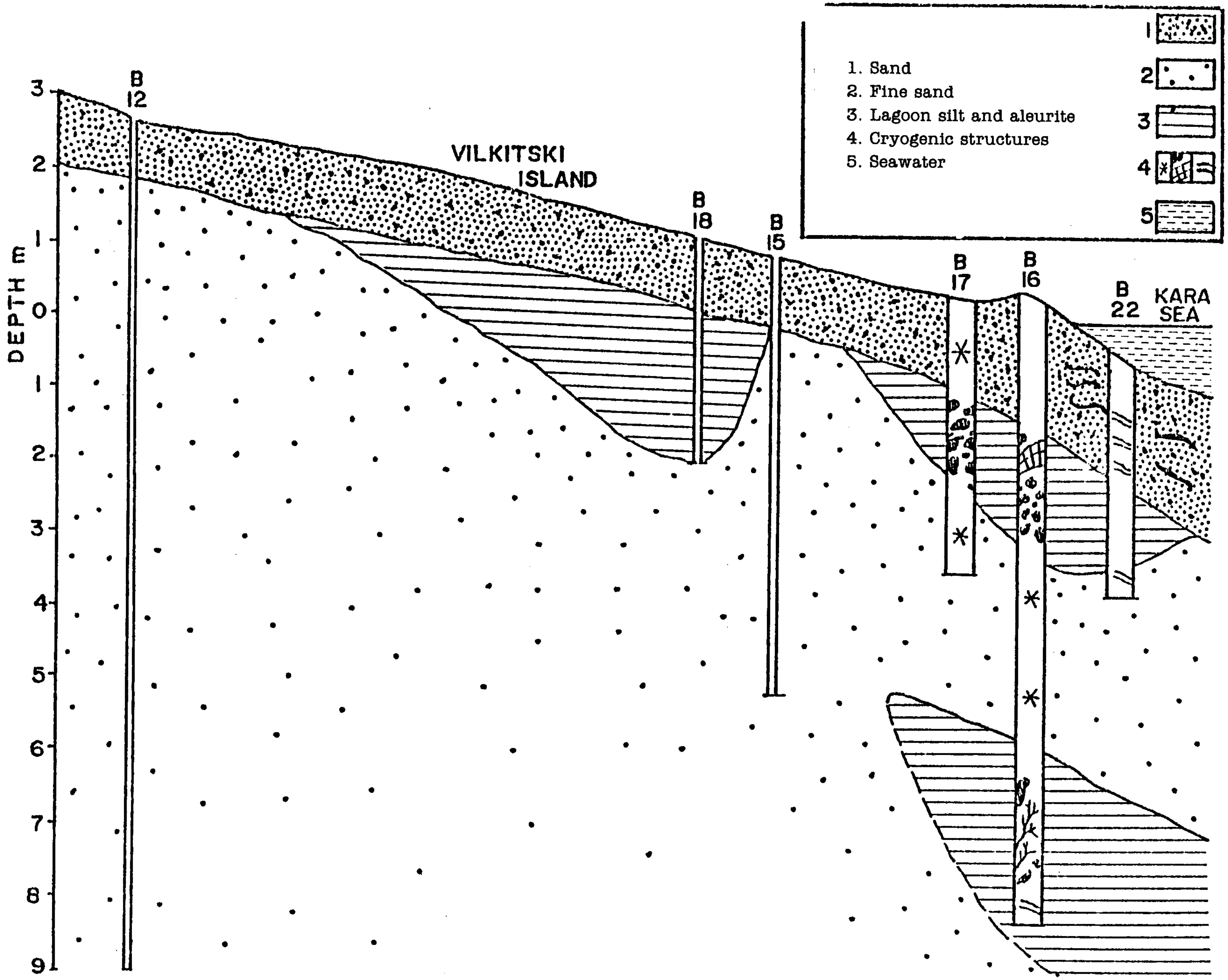
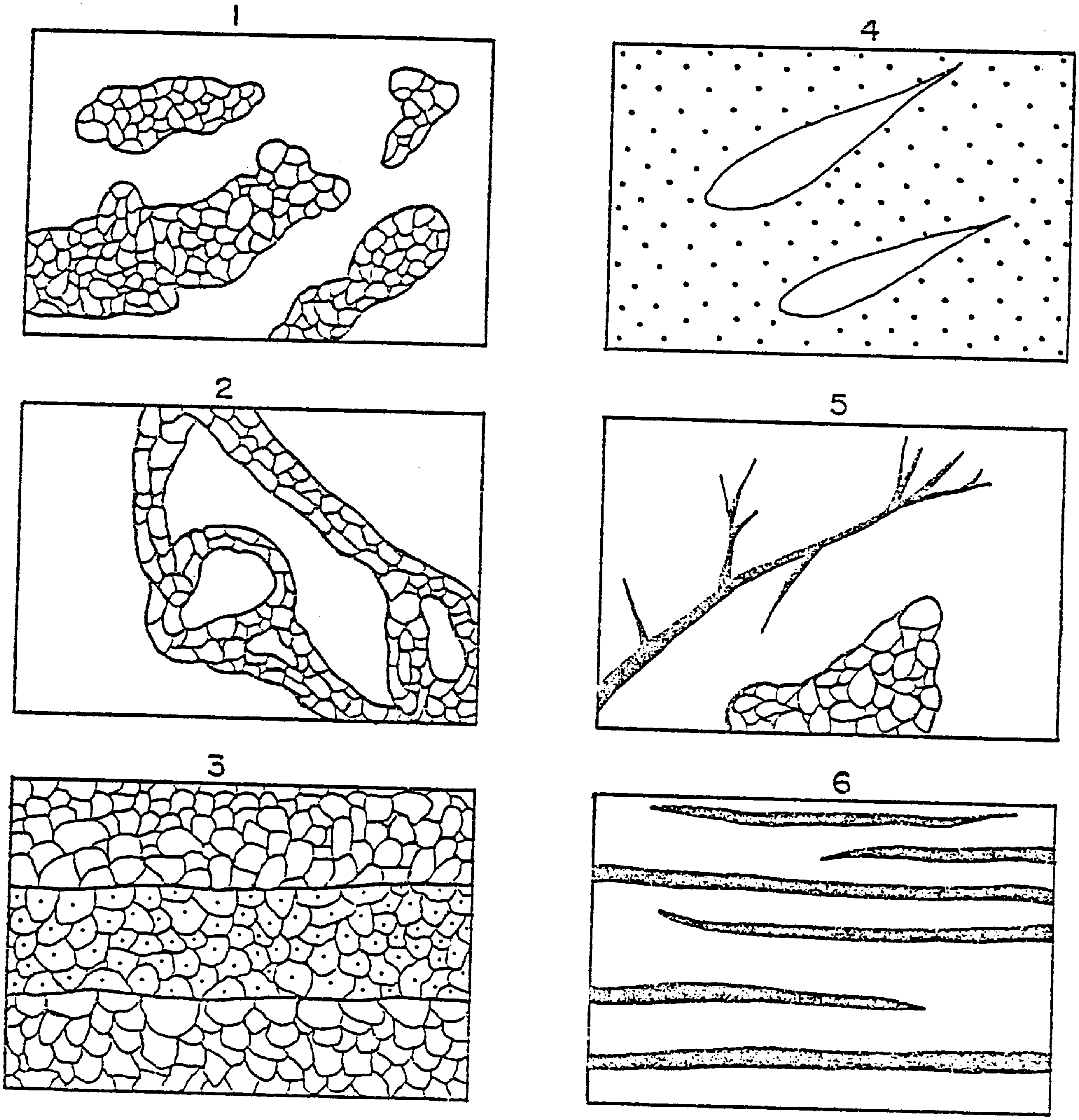
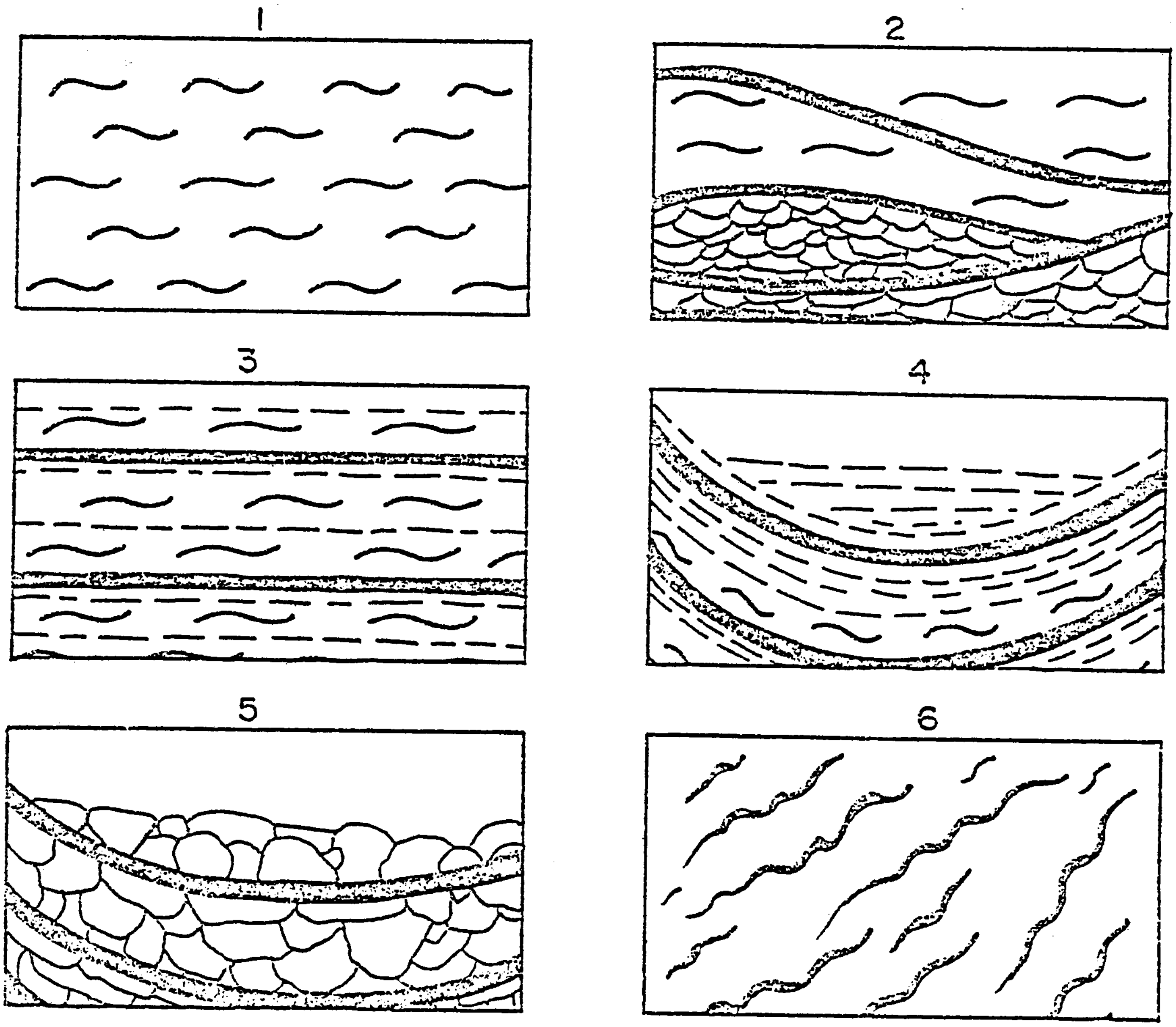


Figure 25.—Geological cross section of the north coast of Vilkitski Island (Kara Sea). After Usov (1967).



1. Ice inclusions in the upper lens of lagoon deposits.
2. (Same as number 1)
3. Injected layered coarse-crystalline ice.
4. Closed cavity ice on walls.
5. Injected ice inclusions in the lower lens of lagoon deposits.
6. Layered cryogenic texture of the submarine slope.

Figure 26.—Ice inclusion forms in the shallow water deposits of a sea embayment (enlarged four times) (Usov 1967).



1. Thin lenticular, characteristic of deposit (facies) of dry slopes.
2. Gently undulating lenticular or recticular, characteristic of deluvium (facies of gentle wet slopes).
3. Horizontal parallel bedding, characteristic of floodplain deposits.
4. Concave parallel bedding, characteristic of deposits of polygonal floodplain (with ice veins).
5. Concave, parallel bedding, recticular, characteristic of deposits in troughs and wet meadows.
6. Obliquely bedded, cryogenic, forming during freezing of bottom slopes.

Figure 27.—The most widely occurring cryogenic textures of the permanently frozen Quaternary deposits. After Katasonov (1962).

Later, the same author published the more detailed "Classification of Frost-Caused Phenomena" (1973), dividing them into two categories: (1) surface phenomena, including relief forms due to freeze and thaw, such as small and large polygons, frost fissures, frost-heaving mounds, ostioles, and mud strips; and (2) subsurface phenomena, among which ground and ice veins, streaks, and deformations of sediments occur. Cryogenic phenomena include, moreover, slope troughs (dells) produced by thermo-erosion, icings, and thermokarst depressions, as well as depressions formed as a result of the melting of glacier ice. His classification, presented below (Figure 28), refers to phenomena but not to the deposits in which they occur.

The author divides cryogenic phenomena into subterraneous and correlated surface structures, according to their genesis; i.e., to the conditions under which the deposits were accumulated and frozen. The choice of such a classificatory distinction is inspired by the necessity of correlating the phenomena under consideration with the properties of components (including also paleontologic remains) and the structure of Quaternary sediments. Such a classification is designed to serve geocryology (permafrost studies) and paleogeographic purposes. Depending on the accumulational environment, present-day cryogenic phenomena are being divided into two groups, namely, into subaqueous or subaerial ones.

Subaqueous cryogenic phenomena are due to freezing of the aqueous sediments, deposited in abandoned river channels (oxbow), lakes, and marine coastal zones. Characteristic of these formations are encrustations and agglomerations of ice. At one surface, the presence of ice is revealed by "bulgunnyakhs" rather than frost-heaving surfaces. The latter are not always well developed and are not therefore presented in the table.

Subaerial cryogenic phenomena are initiated during freezing of the sediments accumulated on floodplains, in deltas, and on slopes. They fall into two sub-groups: terrace-delta cryogenic structures which are characteristic of alluvial, deltaic as well as fluvioglacial formations. The sub-group of cryogenic slope phenomena comprises all those that occur in colluvial-solifluction, eluvial and colluvial formations. It further includes the phenomena occurring in the eolian sediments and peaty swamps deposited on the surface of supra-inundational terraces and watershed plateaus.

Subaerial cryogenic phenomena are associated with particular forms of accumulational surfaces, with slopes, whether steep or gentle, with various landscape types. Moisture of the active layer provides the most reliable index of such topographic, or rather frozen-ground facial conditions which are likely to either promote or inhibit the development of those or other cryogenic phenomena. Abundant moisture leads to formation of ice veins and small ice layers. In dry places with a deficiency of water in the active layer, ground veins are usually the result. These conditions have been marked in the classification table in which cryogenic

ORIGIN OF ROCKS				CRYOGENIC PHENOMENA			PROCESSES		
	LANDFORM ELEMENTS		ORIGIN AND FACIES OF DEPOSITS	SURFICIAL	UNDERGROUND: HUGE MASSES OF ICE, FISSURES, DISTURBANCES LAYERS.	UNDERGROUND: ICE INTRUSIONS	PHYSICAL	MECHANICAL	REMARKS
1	2	3	4	5	6	7	8	9	10
SUBAERIAL ENVIRONMENT	SLOPES INTERFLUVES	MOIST GROUNDS	1 ELUVIUM OF BEDROCK	1 FISSURES POLYGONS (10–20m)	1 ICE WEDGES			FROST CRACKING IN PERMAFROST	
			2 ELUVIUM OF PLEISTOCENE DEPOSITS	2 DEBRIS ISLANDS	2 CRYOTURBATIONS		SORTING	DISRUPTION AND SQUEEZING OF LAYERS	SOLIFLUXION
			3 COLLUVIUM	3a FISSURES POLYGONS (10–20m)	3a ICE WEDGES			FROST CRACKING IN PERMAFROST	
					3b PACKED LAYERS OF "GOLETZ" ICE		CRYSTALLIZATION OF WATER		SOLIFLUXION
			4 SOLIFLUXION DEPOSITS	4a SOLIFLUXION FORMS	4a CRYOTURBATIONS				
				4b DEBRIS ISLANDS ELONGATED DOWN THE SLOPE	4b CRYOTURBATIONS		SORTING	DISRUPTION AND SQUEEZING OF LAYERS	SOLIFLUXION
				4c FISSURES	4c BENT ICE WEDGES			FROST CRACKING IN PERMAFROST	SOLIFLUXION
						4d ICE INTERLAYERS DISRUPTED	CRYSTALLIZATION OF WATER		SOLIFLUXION
			5 DELUVIAL DEPOSITS		5 PACKETS OF LAYERED ICE	5 ICE INTERLAYERS WAVY	CRYSTALLIZATION OF WATER		
			6 PEAT—BOG DEPOSITS	6a POLYGONS (10–20m)	6a ICE WEDGES			FROST CRACKING IN PERMAFROST	
				6b PEAT THUFUR	6b LENSES OF ICE		INJECTION SORTING		
						6c BENT ICE INTERLAYERS	CRYSTALLIZATION OF WATER		

Figure 28.—Classification of frost-caused subaqueous and subaerial phenomena. After Katasonov (1973).

ORIGIN OF ROCKS				CRYOGENIC PHENOMENA			PROCESS		
	LANDFORM ELEMENTS		ORIGIN AND FACIES OF DEPOSITS	SURFICIAL	UNDERGROUND		PHYSICAL	MECHANICAL	REMARKS
					HUGE MASSES OF ICE FISSURES, DISTURBANCES LAYERS	ICE INTRUSIONS			
1	2	3	4	5	6	7	8	9	10
SUBAERIAL ENVIRONMENT	SLOPES—INTERFLUVES	SLIGHTLY MOIST AND DRY GROUNDS	7 ELUVIUM OF BEDROCK	7 STONE POLYGONS (0.5–4.0m)	7 PATTERNED GROUNDS			SORTING	
			8 ELUVIUM OF PLEISTOCENE DEPOSITS	8 MICRO-POLYGONS (0.5–3.0m)	8 FROST CRACKS WITH MINERAL AND/OR ORGANIC MATERIAL			FROST CRACKING IN ACTIVE LAYER	
			9 COLLUVIUM		?			?	
			10 DELUVIAL DEPOSITS	10 MICRO-POLYGONS (0.5–3.0m)	10 FROST CRACKS WITH MINERAL AND/OR ORGANIC MATERIAL			FROST CRACKING IN ACTIVE LAYER	
			11 EOLIAN SANDS	11 HIGH-CENTER POLYGONS (2–8m)	11 GROUND FISSURES (SAG VEINS)			FROST CRACKING IN ACTIVE LAYER	
	FLOOD PLAINS — DELTAS	MOIST GROUNDS	12 PEAT-BOG DEPOSITS	12a LOW-CENTER POLYGONS (10–20m)	12a ICE WEDGES			FROST CRACKING IN PERMAFROST	
						12b BENT AND HORIZONTAL ICE INTERLAYERS	CRYSTALLIZATION OF WATER		
			13 FLOOD-PLAIN AND DELTA DEPOSITS (SILT/SAND)	13 FISSURES, POLYGONS	13 ICE WEDGES			FROST CRACKING IN PERMAFROST	
			14 SEDIMENTS OF NEAR-BED SHOALS	14 FISSURES INDISTINCTLY VISIBLE	14 ICE-MINERAL WEDGES			FROST CRACKING IN PERMAFROST AND ACTIVE LAYER	
			15 GLACIOFLUVIAL PEBBLES	15a POLYGONS (10–40m) AND FISSURES	15a ICE WEDGES			FROST CRACKING IN ACTIVE LAYER	
					15b FROST FISSURES WITH SECONDARY INFILLING			FROST CRACKING IN ACTIVE LAYER	
		DRY GROUNDS	16 FLOOD-PLAIN DEPOSITS	16 FISSURES POLYGONS	16 FROST FISSURES WITH SECONDARY INFILLING			FROST CRACKING IN ACTIVE LAYER	
			17 SEDIMENTS OF NEAR-BED SHOALS	17 FISSURES INDISTINCTLY VISIBLE	17 GROUND FISSURES (SAG VEINS)			FROST CRACKING IN ACTIVE LAYER	

Figure 28.—(Continued)

ORIGIN OF ROCKS				CRYOGENIC PHENOMENA			PROCESSES		
	LAND-FORM ELEMENT		ORIGIN AND FACIES OF DEPOSITS	SURFICIAL	UNDERGROUND		PHYSICAL	MECHANICAL	REMARKS
					HUGE MASSES OF ICE, FISSURES, DISTURBANCES LAYERS.	ICE INTRUSIONS			
1	(2)	(3)	4	5	6	7	8	9	10
SUB AQUEOUS ENVIRONMENT	18		OXBOW AND LACUSTRINE DEPOSITS	18a PINGOS	18a LENSES OF ICE		INJECTION	BENDING OF LAYERS	
						18b VERTICAL AND DIAGONALLY ORIENTED SEGREGATION ICE	SORTING		
	19		MARINE LITTORAL DEPOSITS		19a FLOES		INJECTION SORTING		PERSISTING OF ICE PACK ( )
					19b BLOCKS OF ICE				PERSISTING OF ICE PACK
						19c HORIZONTALLY AND DIAGONALLY ORIENTED SEGREGATION ICE	SORTING		

Figure 28.—(Continued)

phenomena occurring in "swampy" and "dry" slope facies are distinguished from deltaic and terrace phenomena.

The frozen-ground facial conditions which determine the composition and moisture content (the amount of ice) of present-day sediments are responsible for the depth of thawing, the thermal conditions of rocks, and, consequently, for the intensity of cryogenic processes in any given area. Ye. M. Katasonov thinks, therefore, that cryogenic processes should be regarded as a complementary indication of frozen-ground facial conditions. In the classification table a distinction is made among: (1) physical processes such as migration of film water, injection and crystallization of water, etc.; (2) mechanical processes which produce disintegration, displacement and crushing of the material, its sorting, deformation of layers during frost heaving and frost cracking; and (3) specifically geologic processes which cause preservation of ice and solifluction. This division is somewhat simplified in that various processes may operate simultaneously, encroach upon one another, or one may give rise to another. In this classification, the morphogenetic criterion permits local conclusions certifying to the existence of a relationship between cryogenic phenomena and specific sediments, frozen-ground facial conditions of both their accumulation and their freezing.

Frozen-ground facial conditions can be most readily reconstructed on the basis of occurrence of ice which originates syngenetically under the influence of pre-existent permafrost. In subaqueous deposits, ice takes the form of either oblique or vertical lenses and schlieren, which repeat the shape of taliks. In subaerial formations, ice accumulations appear in the form of intervening layers at the border between the active zone and permafrost.

Tiny ground and humus veins, corresponding with micropolygons, are in fact widespread cryogenic phenomena occurring principally on slopes and developing within the active layer under conditions of denudation and instability of accumulation. Such small veins, together with the large ground veins that reflect the polygons of the active layer which are 10–30 m in size and have ice wedges due to cracking of the passive zone of permafrost, constitute a series of frost-fissure polygons.

Deposits with higher moisture content (accumulation of ice) that fill the marshy and peaty dells thaw to a depth of hardly 0.4–0.8 m and yet the mean annual temperature is low (down to −6°C); frost cracking occurs within the permafrost, thus inducing formation of ice veins but no ground veins. The sediments of the ridges on the coastal floodplains consist of fine sand and silt with a negligible content of ice, that thaw to a depth of 3 m and undergo mean annual temperature oscillates from −1.5° to −2.5°C. Frost fissures develop within the active zone, often extending down to the permafrost. The ground and ground-ice veins developed here are large, whereas the colluvual and alluvial silts covering the major part of the slopes surfaces exhibit predominantly ground and humus veins forming micropolygons. Since there are no deep cracks here, the active layer is intersected by a dense network of small fissures and constitutes a sort of "elastic core" in which the tensile stresses, called forth by winter thermal gradients, are released.

Some experts attribute the development of fissuring to differential cooling of deposits. N. N. Romanovskiy (1961), basing his inferences on the results obtained by measurements, believes that in deposits whose mean temperatures are −5 to −6°C and below, only ice veins are apt to form, whereas temperatures of −2 to −3°C and above give rise to ground veins.

These data testify to certain regularities due to thermal regime. However, two facts should be taken into account: First, ground veins are found in deposits whose mean annual temperature is −7 to −8°C (as described from the Anabara lowland, Lena delta); second, even under the most favorable geothermal conditions, ice veins fail to develop in slope sediments whose upper portions are usually dissected by tiny fissures.

The thermal and physical regularities controlling the development of ice and ground veins are of importance for the solution of many problems relating to geocryology. In the case in question those regularities are obviously associated with the genetic types and facies of the sediments in which cryogenic phenomena occur, since the mineral composition, moisture content, and thermal regime of these deposits are determined by one and the same cause, which is their genesis, the conditions of their formation.

Analysis of the facies distinguished, comparison of their specific features and present-day homogeneous formations of cryogenic phenomena can help to elucidate the paleogeographic conditions and history of development of submarine permafrost. The advantage of the frozen ground-facial method is that it does not permit study of cryogenic phenomena apart from the sediments within which they occur and sets the investigation of these phenomena upon firm geologic foundations.

In describing the present-day layers of perennially frozen deposits encountered on the bottoms of shallow-water rivers, lakes, and seawater bodies, scientists usually note some peculiarities of the cryogenic structure of the frozen strata forming under different physiographic conditions; the complex of cryogenic textures characteristic, for example, for the upper syngenetic layer, a combination of radial-fissured textures with oblique-broken intercalations and lenses of ice according to observations made by M. S. Ivanov in the beach area near the mouth of the Yana (Figure 7). The lower, epigenetic part of the frozen layer is characterized by a combination of massive cryogenic textures with horizontal ice intercalations. But in this layer it is most typical to observe texture-forming ice with a granular structure and also thin bandlike fissures and siltlike voids only partially filled with ice. It was also noted that there is a close correlation between the composition of the deposits and the cryogenic textures. For example, in the sandy and sandy-aleurite deposits there is a predominance of a massive cryogenic texture, but in the clayey aleurites, containing lenticular inclusions of plant remains, there is widespread development of horizontally oriented lenticular cryogenic textures.

It should be noted also that, depending on the conditions of the freezing of alluvium in the formation of underwater frozen strata in northern Yakutia, the factor some geologists underline as of greatest importance is the complex polygenetic freezing in which a relatively thin syngenetic layer of perennially frozen deposits usually covers a thick epigenetic or parasyngenetic layer of frozen rocks.

Some examples of this cryolithological approach can be found in the works of V. Usov published in 1965–69. This author studied the formation of permanently frozen deposits and cryogenic structures under lagoon conditions. Usov supposed the possibility of the relatively deep-water deposits freezing by the level of the middle sublittoral. The main condition for this is the presence of the same kind of freezing agent. Most often it is the ice body. In his opinion, freezing of subaquatic deposits in arctic lagoon conditions is the most probable source of permafrost development. He divides the marine accumulation area into three zones:

1) Subaquatic and subaerial deposits of the shallow coasts freezing syngenetically.
2) Relatively shallow water deposits, including bay deposits, freezing under water during small changes in reservoir parameters (syngenetic and diagenetic types of cryogenic structure formations).
3) Areas closed to middle littoral, formed by thawed, cold, and partly perennially frozen deposits.

V. Usov emphasizes the existence of relict permafrost in all three zones and its influence as an agent in stimulating the freezing of the younger layers. Figure 29 shows the scheme of the relationship between recent and relict permafrost. Permafrost forms in lagoons when sea ice interacts with the lagoon and bar bottom. The presence of the relict Holocene permafrost aids the development of recent permafrost from below, thus repeating the phenomenon in the Holocene, when the older Pleistocene relict permafrost acted as a freezing agent.

V. Usov considered the possibility of the determination of the different types of freezing processes and cryogenic structures in the first and second zones. On the basis of differences in the cryogenic structures, he delineates the formations of the lagoon, embayed, deltaic, and tidal-marsh types of the Arctic coast. He specifies the peculiarities of the way epigenetic freezing of bottom marine deposits (clay, in particular) is connected with duration of subaerial exposure. The freezing of such deposits usually is followed by intensive formation of ice layers and the cryogenic structures that appear to be syngenetic, but in reality have an epigenetic origin.

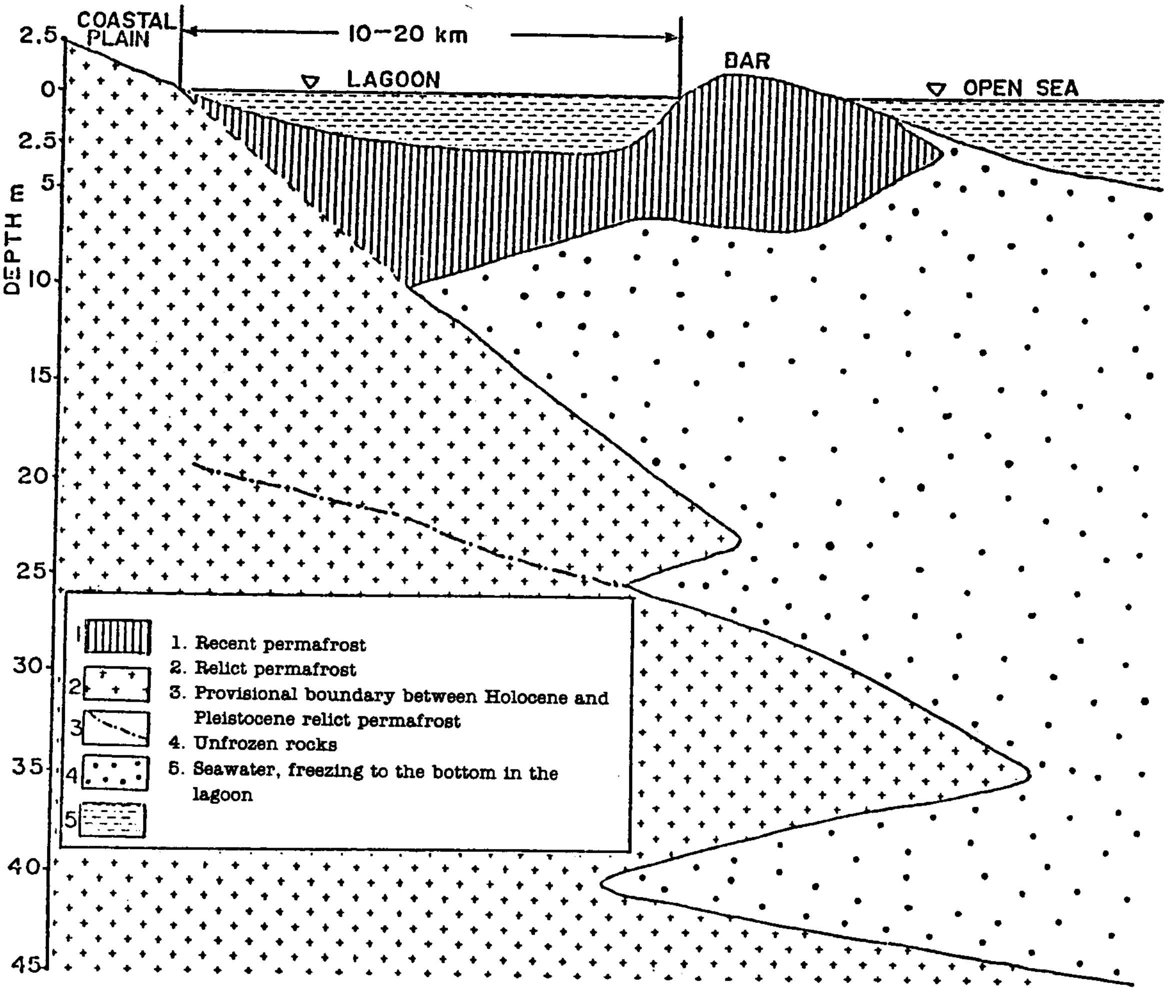


Figure 29.—Scheme of the relationship between recent and relict permafrost.

THERMAL STATE AND REGIME OF SUBMARINE PERMAFROST DEVELOPMENT

Information about the thermal state and regime of submarine permafrost in the Soviet Arctic has been gathered by Ye. N. Molochushkin (1965, 1970, 1973, 1975*), Molochushkin and Gavril'yev (1971), L. A. Zhigarev and I. R. Placht (1974), and I. Danilov and L. Zhigarev (1977)**. In the zone of the contemporary accumulation of sediment in the Laptev Sea, cold

* Recent publication "Merzlotno-geologiceskie issledovania moria Laptevih in "problemi geoligii shelfa." "Nauka," Moskva, 1975. (Cryogenic conditions investigations in Laptev Sea.)

** Recent publication "Merzlie porodi Arcticheskogo Shelfa" in "Merzlie porodi i Snegni pokrov." "Nauka," Moskva, 1977. (Perennially frozen rocks of the Arctic shelf.)

bottom deposits have been detected and studied which are not cemented by ice at a temperature up to −6°C. An interesting feature of such deposits is the fact that mineralization of their threshold solution is significantly above the mineralization of seawater (Marchenko 1966; Molochushkin and Gavril'yev 1970; Neizvestnov Semenov 1973).

Thermophysical Characteristics of the Shelf Deposits

Molochushkin and Gavril'yev's work was done in the eastern part of the Laptev Sea. The character of the deposits can be seen in the geological cross section (Figure 30). Changes in the deposits' salinity and moistness with depth for different boreholes are shown in Figure 31. We can see that the highest salinity is typical for deposits 2–2.5 m below the sea bottom. At depths of 4–15 m, the salinity is much less and the oscillations of this characteristic are small. The pore water in the deposits is mineralized more than the seawater during the winter salinity maximum. In borehole 5, the most distant point from the shore, the pore water is 1–1.5 times more mineralized than the seawater. In boreholes 2 and 3, closer to the shore, the pore water is 2–3 times more mineralized. The composition of the salt in the pore water is shown in Table 12. We can also see the relationship between salt concentration and moistness in Table 13. The relationship of bottom deposit freezing temperature with moistness and salinity according to Molochushkin is shown in Figure 32. Figure 33 gives the dependence between bottom deposit freezing temperature and salt concentration in the pore water, and Figure 34, the relative content of the unfrozen water in its connection with temperature and salinity of the deposits.

The temperature curve for the freezing points and their relationship with moistness and salinity can be shown by the following empirical formula (1):

$$(1) \qquad Tf = -28{,}4\ (e^{0{,}022\,\frac{S}{W}} - 1) \;-\; 5e^{-35}\ (W - 0.035)$$

Tf - Temperature of the freezing points

S - Salinity in $^{o}/o$

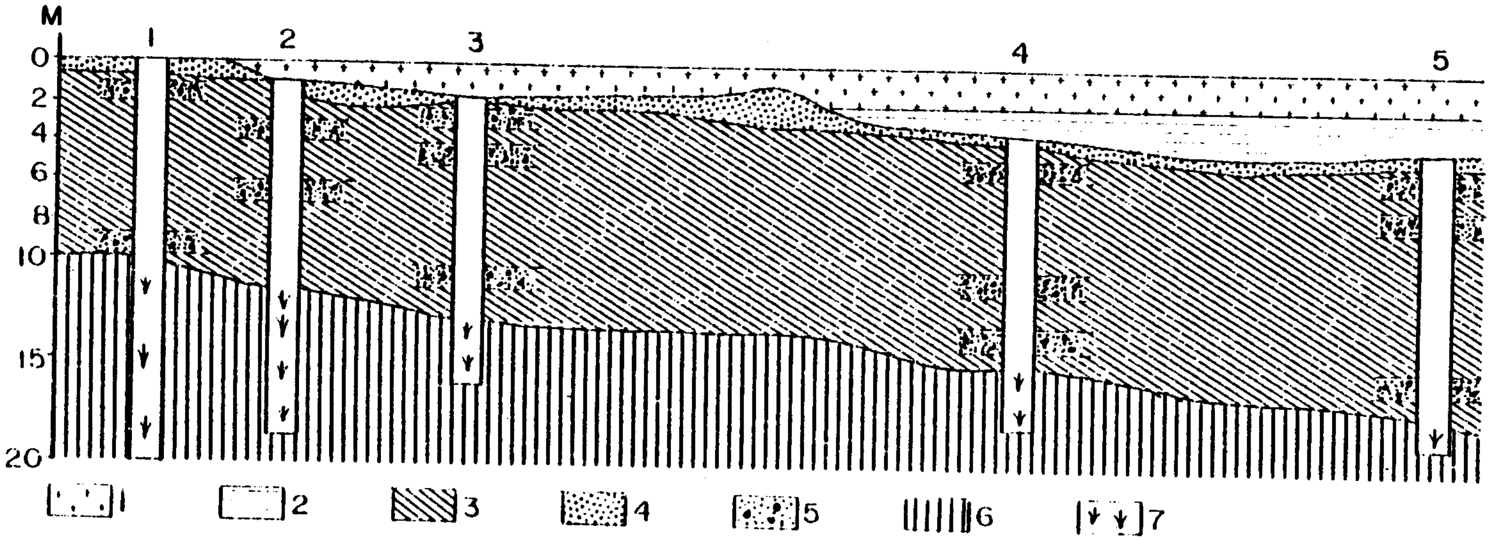
W - Moistness

e - Base of the natural logarithm

The amount of the unfrozen water (2):

$$(2) \qquad W\ \text{unfrozen} = \frac{0.022S}{\ln\left(-\frac{Tf}{28{,}4} + 1\right)}$$

$$\text{if } W \geq 20\%$$



1. Sea ice
2. Seawater
3. Gray silt, some places sandy
4. Sand
5. Silty sand with wood remnants
6. Silt
7. Peaty material

Figure 30.—Geological cross section, eastern part of the Laptev Sea. After Molochushkin (1970).

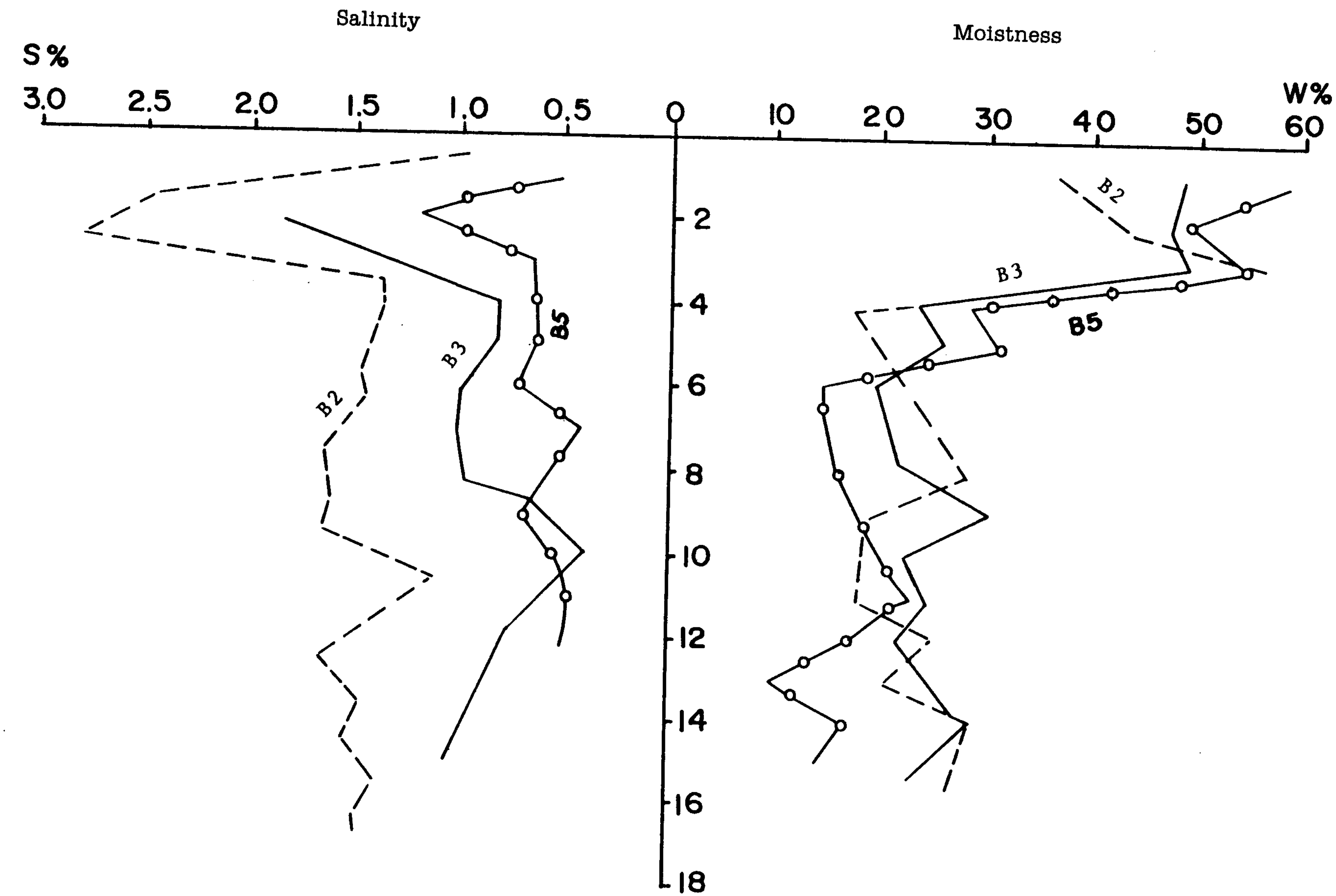


Figure 31.—Changes in bottom deposit salinity and moistness with depth. After Molochushkin (1970).

Table 12.—Composition of salts in pore water. After Molochushkin (1970).

Depth (m)	pH	Σ	Ca	Mg	Na	NH_4	HCO_3	SO_4	Cl
				Borehole 2					
0–1	7.0	995.7	9.4	23.2	339.0	2.5	16.5	65.2	550.5
1–2	7.2	2418.2	11.0	37.8	871.1	3.7	30.1	95.9	1387.3
2–3	7.2	2804.9	13.5	44.8	1010.3		48.6	94.6	1618.4
3–4	7.2	1364.3	6.3	21.6	493.1	6.6	30.1	35.8	792.5
4–5	7.2	1369.3	6.3	20.7	495.5	7.5	25.5	40.5	793.5
5–6	7.1	1484.4	11.0	26.3	527.0	10.4	15.0	35.3	877.2
6–7	7.0	1481.5	19.6	33.5	504.3	12.5	13.5	49.4	868.0
7–8	8.0	1641.2	23.1	41.4	550.2		12.0	57.0	963.5
8–9	7.0	1509.8	20.8	33.6	514.3	12.5	9.0	48.5	888.4
9–10	7.0	1541.7	21.9	56.4	515.4	12.0	18.1	56.1	882.6
10–11	7.1	1121.1	16.4	20.7	385.8	12.0	27.1	54.7	629.8
11–12	7.0	1392.7	20.8	32.2	468.3	20.0	19.5	69.3	792.4
12–13	7.0	1656.1	26.6	38.5	552.1	15.0	19.5	100.0	926.0
13–14	7.0	1453.1	21.9	28.9	497.9	17.5	19.5	88.7	816.0
14–15	7.0	1544.5	25.1	33.1	517.7	17.5	19.5	107.4	851.0
15–16	7.0	1389.7	22.5	31.4	460.2	17.2	17.8	96.1	761.6
16–17	7.0	1487.0	20.4	28.9	509.3	17.5	22.6	85.5	831.6
17–18	7.0	1482.6	17.6	26.0	515.6	17.0	19.5	84.1	829.6
18–19	7.0	1166.0	17.7	26.0	392.0	17.0	18.1	72.4	649.0
				Borehole 3					
2.0	7.2	1844.2	19.1	48.9	588.3		17.6	236.2	942.9
3.0	7.2	1241.8	8.2	18.0	436.3		14.6	150.7	621.2
4.0	7.2	817.6	5.7	10.5	292.3		24.3	71.8	425.1
5.0	7.1	824.5	6.7	11.5	289.5		30.4	68.7	433.1
6.0	7.1	998.6	13.4	21.0	365.0		16.7	42.9	607.9
7.0	7.1	1016.9	12.3	13.3	357.0		30.4	93.6	525.4
8.0	7.1	986.6	9.8	13.5	349.0		33.4	81.8	515.8
9.0	6.9	606.5	3.6	6.3	218.9		10.6	59.5	312.9
10.0	7.2	404.2	3.9	3.1	148.5		39.5	26.0	203.0
12.0	7.2	763.5	9.0	9.1	270.0		36.5	71.0	386.1
13.0	7.2	876.0	10.6	12.1	306.1		30.4	87.2	448.8
14.0	7.1	990.1	13.6	16.9	340.9		30.4	79.8	523.6
15.0	7.1	1056.6	13.1	17.0	364.9		30.4	102.2	543.6
				Borehole 5					
1.0	6.9	651.9	6.7	13.0	224.4	6.4	26.7	51.9	342.4
2.0	6.0	1027.6	11.0	26.1	400.5	11.2	29.7	221.9	533.4
3.0	7.0	685.5	4.3	10.8	241.6	5.5	66.8	84.0	311.4
5.0	6.9	641.0	4.3	7.4	220.7	10.0	13.4	152.9	249.2
6.0	6.8	702.1	3.1	6.1	253.7		19.3	97.9	331.6
7.0	6.4	436.8	3.8	2.5	161.8		23.7	25.0	231.9
8.0	6.4	588.9	2.8	6.8	212.9		11.9	51.7	308.6
9.0	6.4	680.1	3.8	6.8	246.0		13.4	68.9	347.9
10.0	6.4	543.3	2.4	4.5	200.1		34.2	48.0	271.2
11.0	6.7	486.0	3.3	2.8	179.3		57.9	50.1	218.5
12.0	6.7	496.8	2.8	2.3	187.0		69.8	34.0	232.9
13.0	7.0	603.2	3.1	2.7	266.9		68.9	37.7	295.1

Table 13.—Relationship between salt concentration and moistness in the bottom deposits. After Molochushkin (1970).

Borehole	Depth (m)	Moistness (%)	Salinity (%)	Concentration (%)
2	2–2.5	50	2.8	5.6
	4–18	28	1.5	7.5
3	2–2.5	50	1.8	3.6
	4–15	20	0.9	4.5
5	2–2.5	50	1.2	2.4
	4–13	20	0.6	3.0

The relative content of ice in boreholes 2, 3, and 5, its connection with the minimal temperatures and the freezing temperatures, is shown in Figure 35. Table 14 gives the thermophysical characteristics of the silty deposits in the boreholes. According to investigations of the Geothermal Laboratory of the Permafrost Institute of the Siberian Department of the Academy of Sciences of the USSR, the heat conductivity coefficient of the most prevalent rocks of Yakutia varies from 1.5 to 4.5 Kcal/(hr • m • °C); i.e., it can vary by a factor of 3. The equilibrial thickness of the cryolithic zone is directly proportional to the value of this coefficient. Consequently, depending upon the choice of the magnitude of the heat conductivity coefficient, the calculated thickness of the cryolithic zone can differ 3 times. Figures 36, 37, and 38 show the dependence of the thermophysical characteristics of the silty deposits from the moistness. This dependence is expressed by empirical formulas (3) and (4):

$$\lambda = \frac{W}{3{,}55\ W^2 + 1{,}4\ W - 0{,}25} + 0{,}5 \qquad (3)$$

λ – coefficient of the thermal conductivity of the silty sediments of $W \geq 18\%$

$$C_\gamma = \left(0{,}15 + W\right) \cdot \frac{2700}{2{,}70\ W + 1} \qquad (4)$$

C_γ – volumetric thermal capacity of the silty sediments.

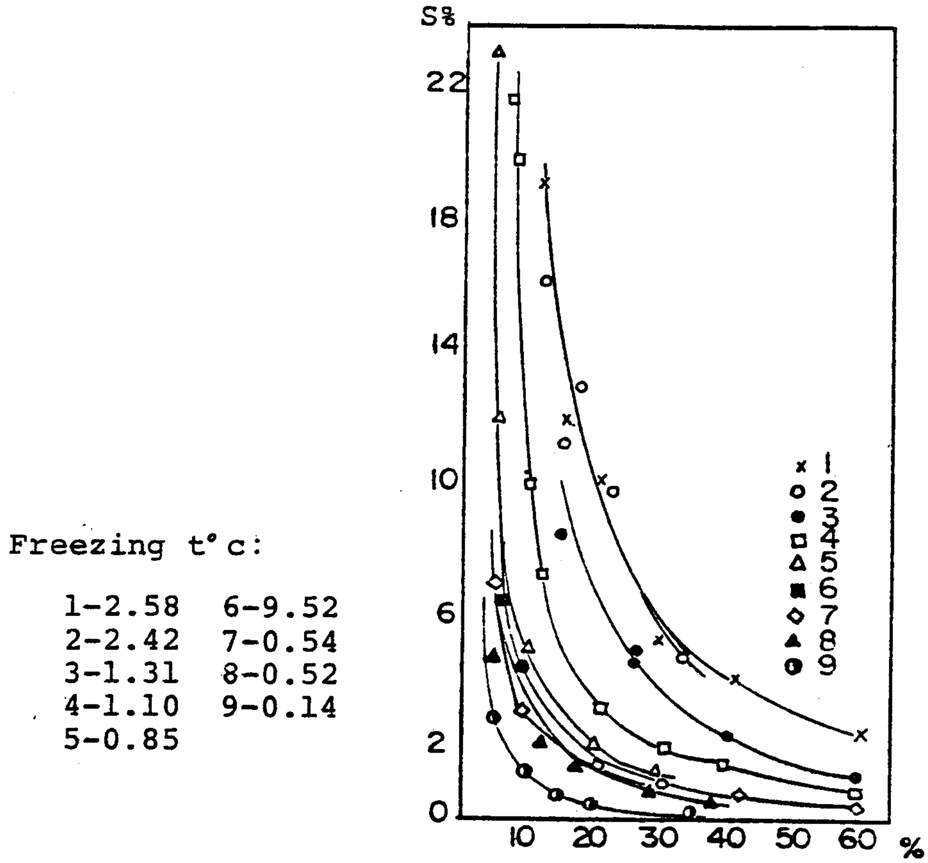


Figure 32.—Relationship of bottom deposit freezing temperature with moistness and salinity. After Molochushkin (1970).

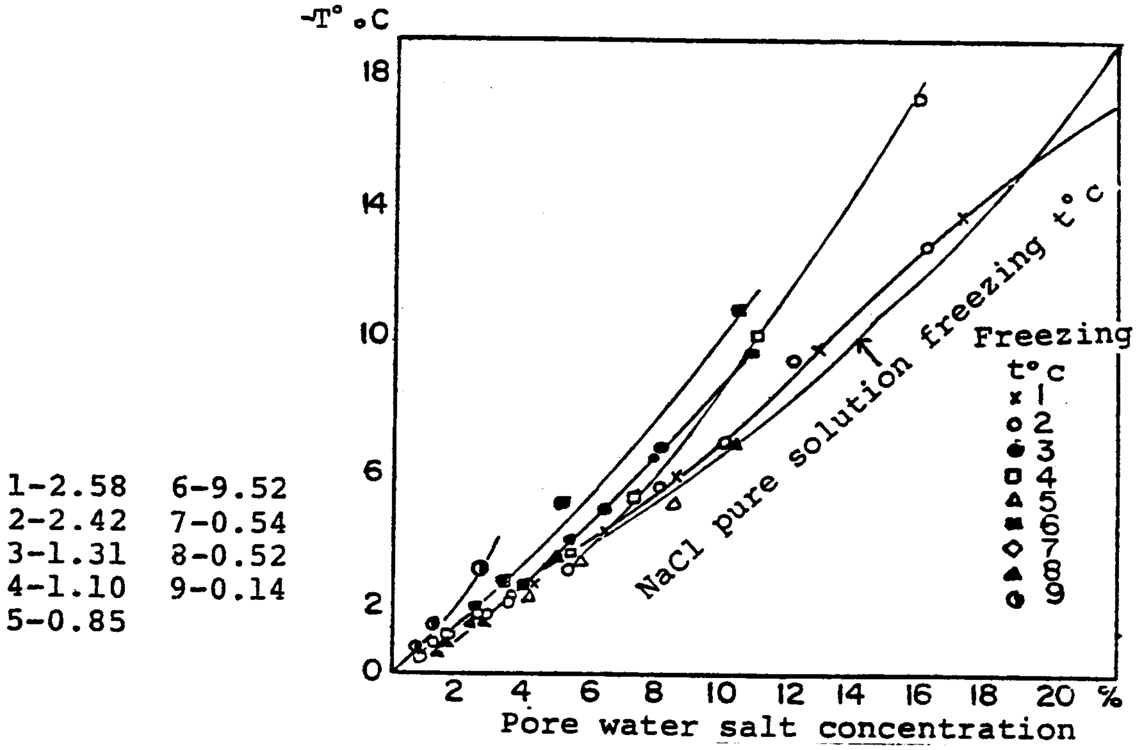


Figure 33.—Relationship between bottom deposit freezing temperature and the salt concentration in the pore water. After Molochushkin (1970).

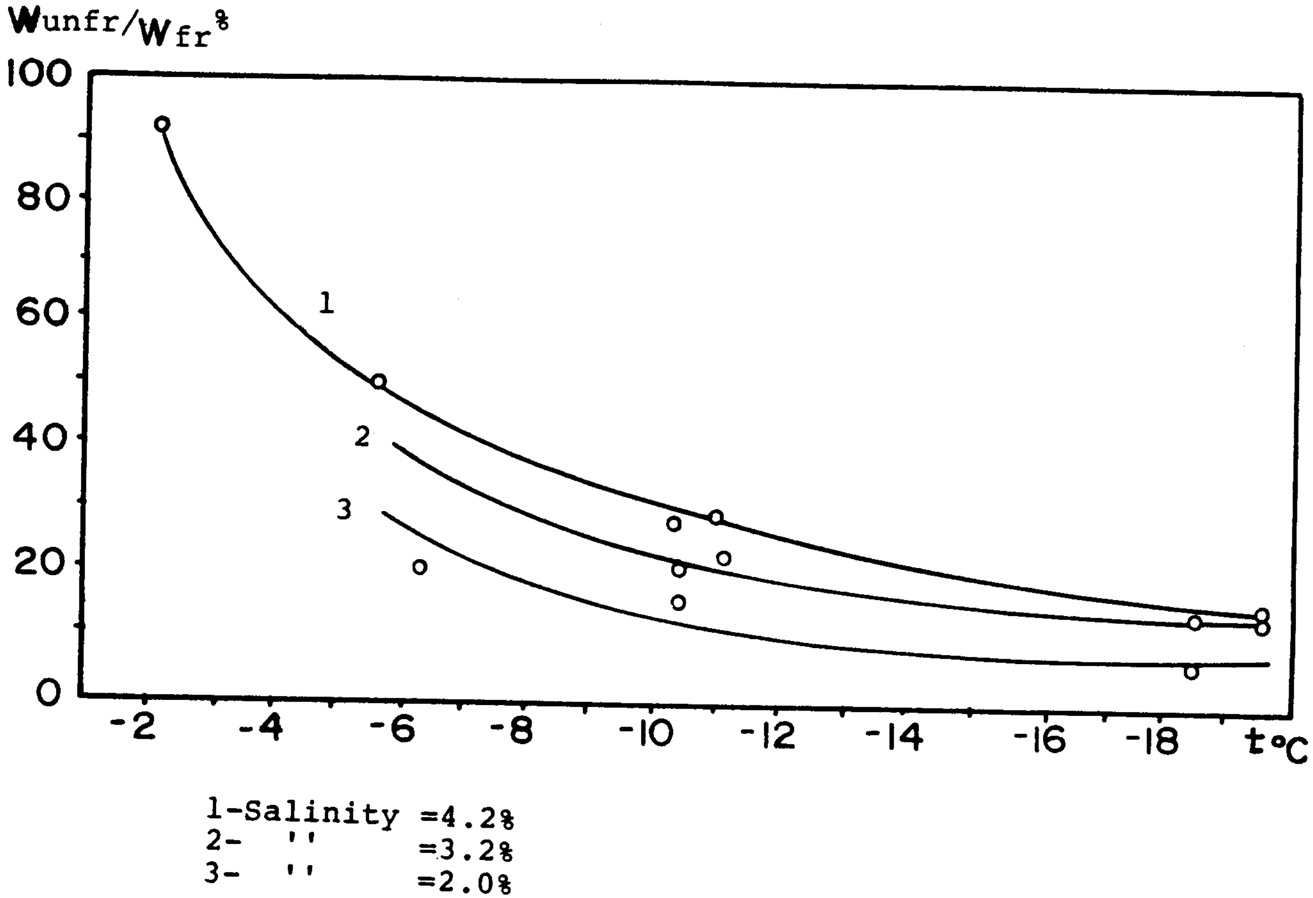
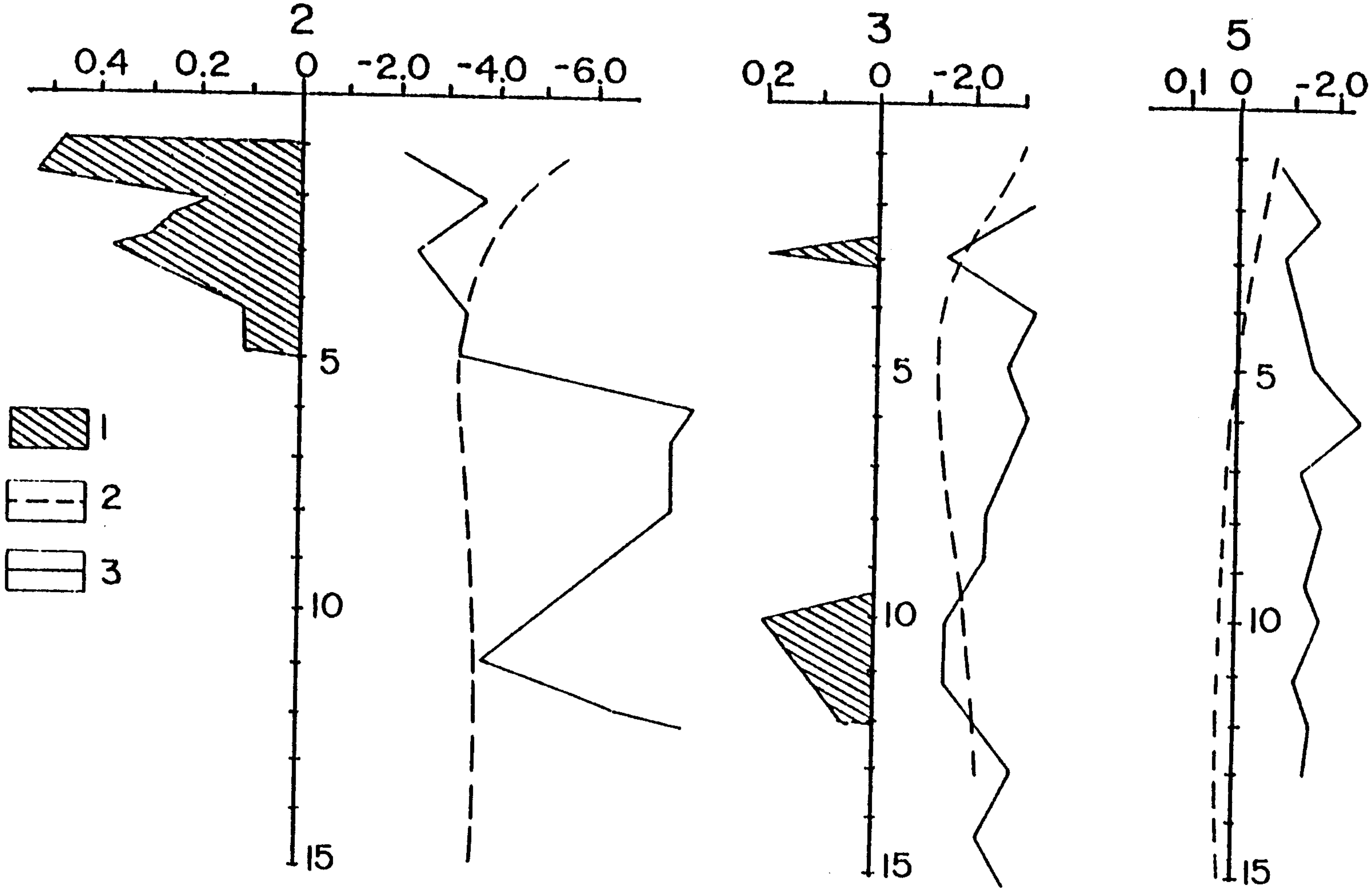


Figure 34.—Relative content of unfrozen water in connection with temperature and salinity of the deposits. After Molochushkin (1970).



1. Relative content of ice   2. Minimum t°C   3. Freezing t°C

Figure 35.—Relative content of ice in boreholes 2, 3, and 5. After Molochushkin (1970).

Table 14.—Thermophysical characteristics of silty deposits. After Molochushkin (1970).

Borehole	Depth m	w	γ	λ	a	$c\gamma$
2	0–1	21.1	1.61	1.56	0.00268	582
	1–2	40.5	1.21	1.32	0.00196	672
	2–3	41.7	1.20	1.05	0.00154	680
	2–3	36.6	1.25	0.95	0.00143	643
	4–5	20.4	1.45	0.86	0.00167	513
	5–6	30.2	1.35	0.97	0.00159	610
	6–7	32.5	1.25	0.84	0.00142	593
	7–8	27.0	1.44	1,25	0,00206	605
	8–9	30.0	1.34	1.11	0.00183	606
	9–10	38.1	1.18	0.86	0.00137	628
	10–11	19.2	1.43	1.00	0.00200	488
	11–12	14.6	1.60	1.27	0.00268	474
	12–13	22.5	1.55	1.39	0.00239	581
3	13–14	18.2	1.79	2.27	0.00382	594
	3	19.7	1.64	1.55	0.00270	572
	4	21.4	1.62	1.67	0.00282	592
	6	18.1	1.70	1.71	0.00303	563
	7	20.0	1.68	1.43	0.00243	588
	8	16.4	1.84	2.34	0.00403	579
	9	13.8	1.62	1.25	0.00266	468
	10	19.6	1.71	2.07	0.00343	593
	12	21.5	1.83	1.51	0.00253	597
	13	22.4	1.62	1.60	0.00262	612
	14	24.4	1.51	1.26	0.00212	593
5	15	28.3	1.49	1.30	0.00200	646
	1	17.8	1.45	1.14	0.00238	478
	2	311.7	1.36	1.26	0.00198	636
	3	35.7	1.34	1.43	0.00210	679
	4	32.1	1.25	0.93	0.00158	588
	6	14.6	1.78	1.75	0.00330	528
	7	16.9	1.78	1.95	0.00342	570
	8	20.0	1.64	1.49	0.00260	575
	10	19.1	1.82	1.91	0.00310	618
	11	34.5	1.27	0.88	0.00139	632
	12	23.3	1.66	1.70	0.00267	636
	13	17.0	1.90	2.27	0.00362	627
	14	19.2	1.81	2.03	0.00328	618
	15	18.2	1.79	2.27	0.00382	595

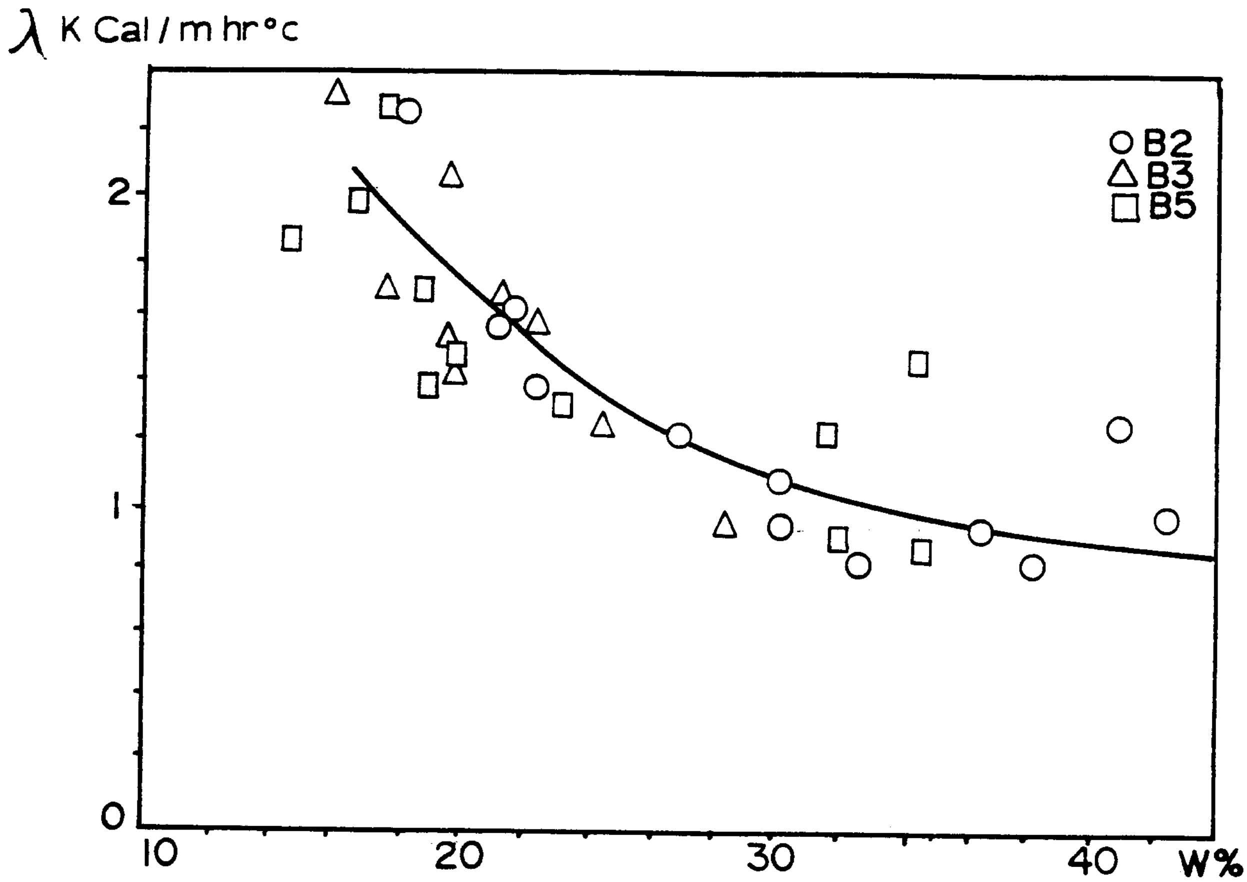


Figure 36.—Relationship of the thermophysical characteristics of silty deposits with moistness (1). After Molochushkin (1970).

According to Molochushkin, Gavril'yev, Zhigarev, Placht and Danilov, the bottom freezing temperature depends on salt concentrations in the pore water. If the moistness is 25%, the freezing temperature may change from –1 to –7.8°C, with different salt concentrations. If there is little salt, the freezing temperature of the deposits is equal to the freezing temperature of seawater with the same salinity. If the pore water salt concentration is 3–4%, the freezing temperature of the deposits is lower than for water with the same salt concentration. For instance, freezing begins at –5.7°C for sandy silt with pore water and a salt concentration of 5.2%. This figure is lower (2.9°C) than the seawater freezing temperature with the same salt concentration. This gives Molochushkin a base for his conclusion that subaqueous freezing of the deposits is a rather limited, not a widespread event. Grain-size composition also plays a role in the freezing process. For example, silty clay with 50% pore water and a salt concentration of 2% began to freeze when its temperature reached –4°C. Silty sand with the same amount of water and salt concentration stayed unfrozen at –5.3°C.

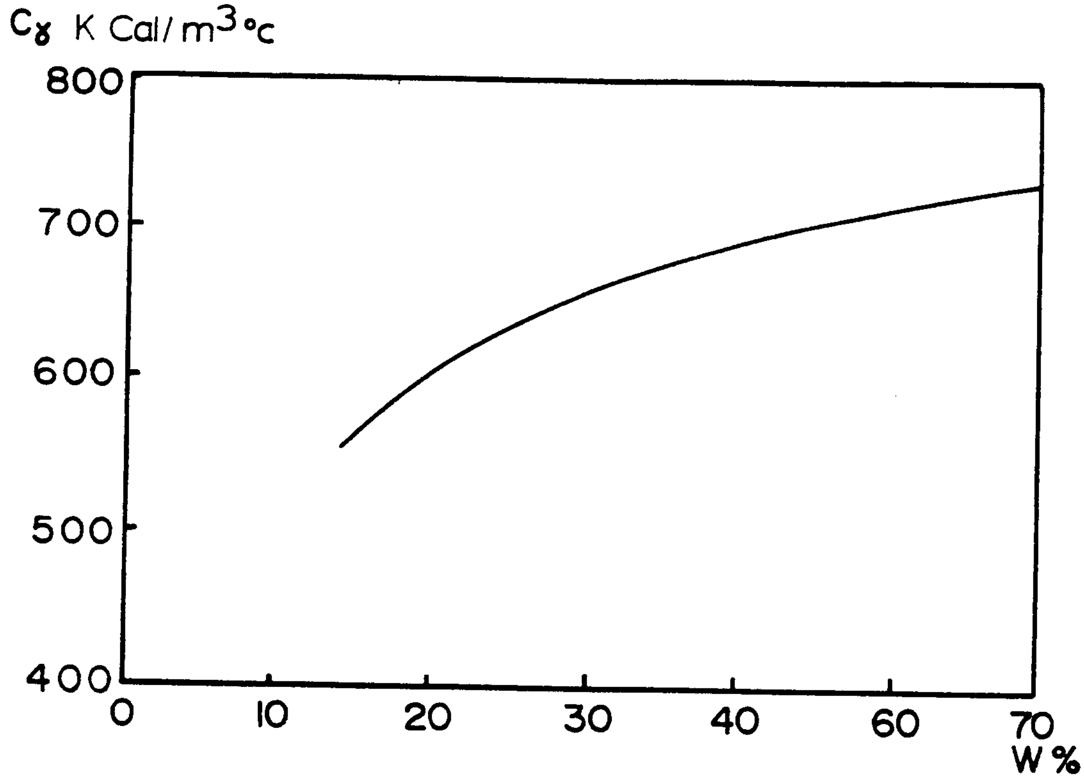


Figure 37.—Relationship of the thermophysical characteristics of silty deposits with moistness (2). After Molochushkin (1970).

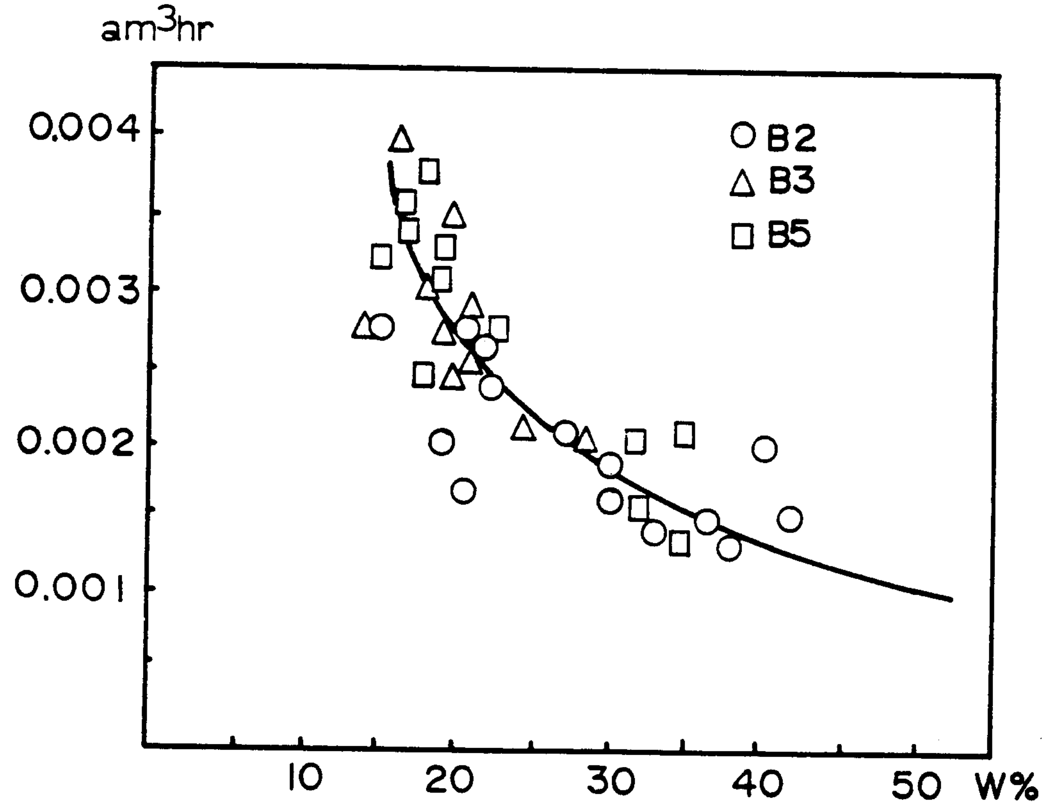


Figure 38.—Relationship of the thermophysical characteristics of silty deposits with moistness (3). After Molochushkin (1970).

Physico-mechanical properties of the deposits depend on the amount of water in frozen and unfrozen states: the higher the salt concentration, the greater the amount of unfrozen water. At –5.0°C in sandy silts with pore water salt concentration of 2.0, 3.2, and 4.2%, the amounts of unfrozen water are 29, 41, and 59%, respectively. During the rising of the deposit temperature from –10.0 to –1.5° at a salt concentration of 4.2% the amount of unfrozen water increased from 30 to 93%. This makes the drilling process much easier. The experiments show also that the coefficients of the thermal conductivity, temperature conductivity, and volumetric capacity are connected, primarily with moistness, and less from the salt concentration.

Dynamics of Temperature Changes

The depth of the temperature oscillations in the deposits (during 1 year observation) was 7–9 m. In the shallow part of the sea with ice cover on the bottom, the mean annual temperature of the deposits decreases with water depth. At about the 1–m isobath, these temperatures decrease from –1.5 to –3.5°C; at the 2–m isobath, from 0.2 to 2.1°C. With the direct ice contact the upper layer of the deposits cools more intensively, sometimes to –6°C (isobath 1 m) and –3°C (isobath 2 m). Because of the lower salt concentration at the deeper layers and the pattern of the mean annual temperature distribution, the possibility of freezing is greater in the deeper part of the rock mass. At seawater depths of 3–5 m, the deposit temperatures at the one depth from the sea bottom are equal and these temperatures are equal also to the mean annual temperature of the seawater there. This means that some areas close to the river delta are taliks with a positive mean annual temperature. According to Molochushkin, deposits in shelf areas with a mean annual seawater temperature below zero are usually in a frozen state. Modern permafrost develops on the shallow parts of the shelf surrounding the river deltas (the sedimentation of the unsalted deposits). These areas are divided by the narrow taliks.

The temperature of the upper layers of the perennially frozen deposits changes greatly under the thermal influence of the seawater if it is rising. For instance, a rise in temperature of 5–7°C at deposit depths of 10–15 m continues usually for 20–30 years. Figure 39 shows the temperature changes of continental frozen rocks under the thermal influence of seawater. For an area of shelf with mean annual seawater temperature of 0°C and ice-bonded deposit thickness of 15–45 m, the time of thawing could be about 6,000–12,000 years. If the deposits have a relatively high concentration of salt (Figures 40 and 41), this time is about 4,000 years. Areas of the Laptev Sea shelf with negative mean annual seawater temperatures are suitable for submarine permafrost.

Molochushkin gives some maps showing that in the areas of seawater with a mean annual temperature of –0.8°C, the thickness of the submarine permafrost is about 50–60 m; areas with a seawater temperature of –1.3°C are about 90–100 m thick. Figure 42 shows the scheme

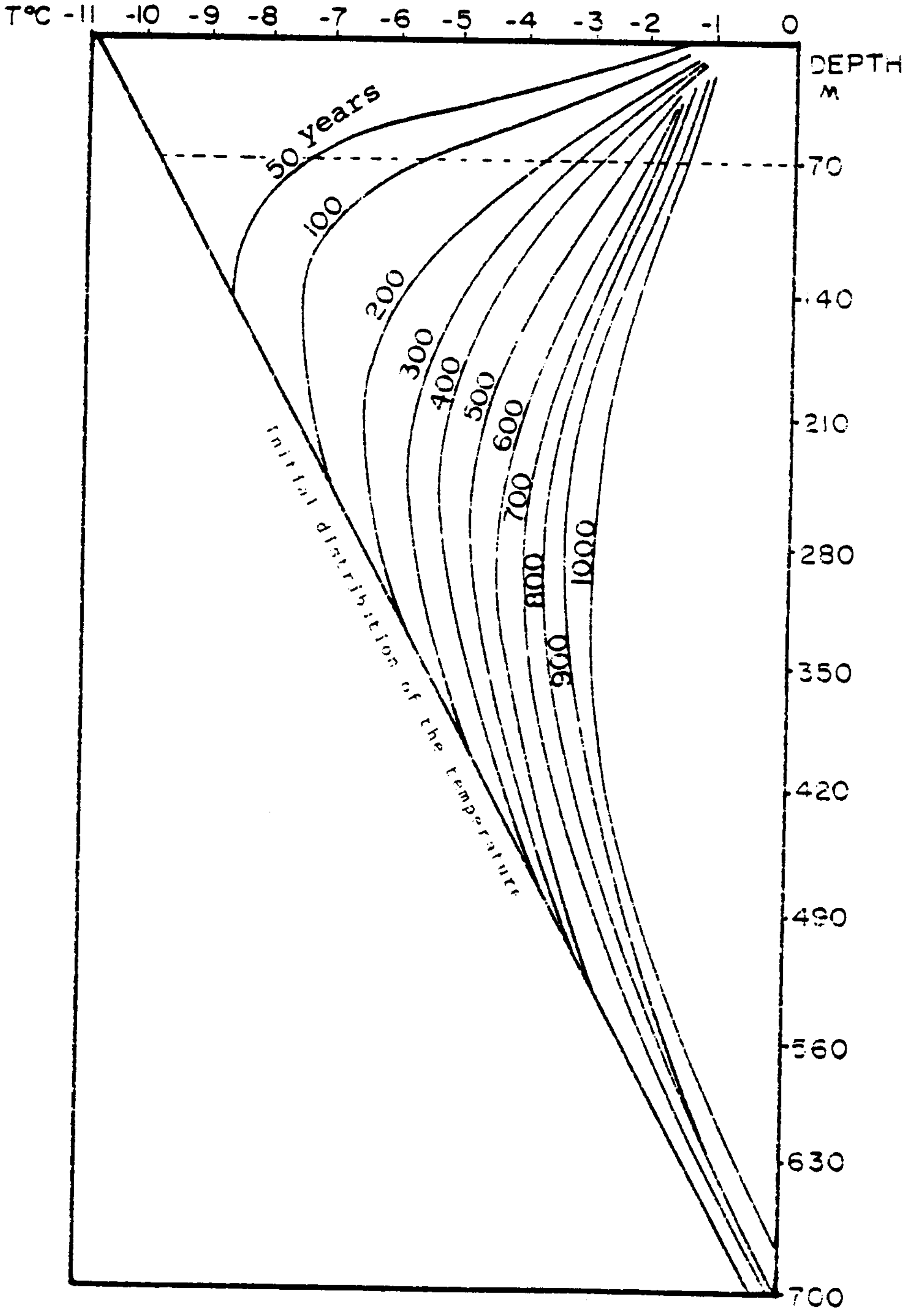


Figure 39.—Temperature changes of continental frozen rocks under thermal influence of seawater. After Molochushkin (1970).

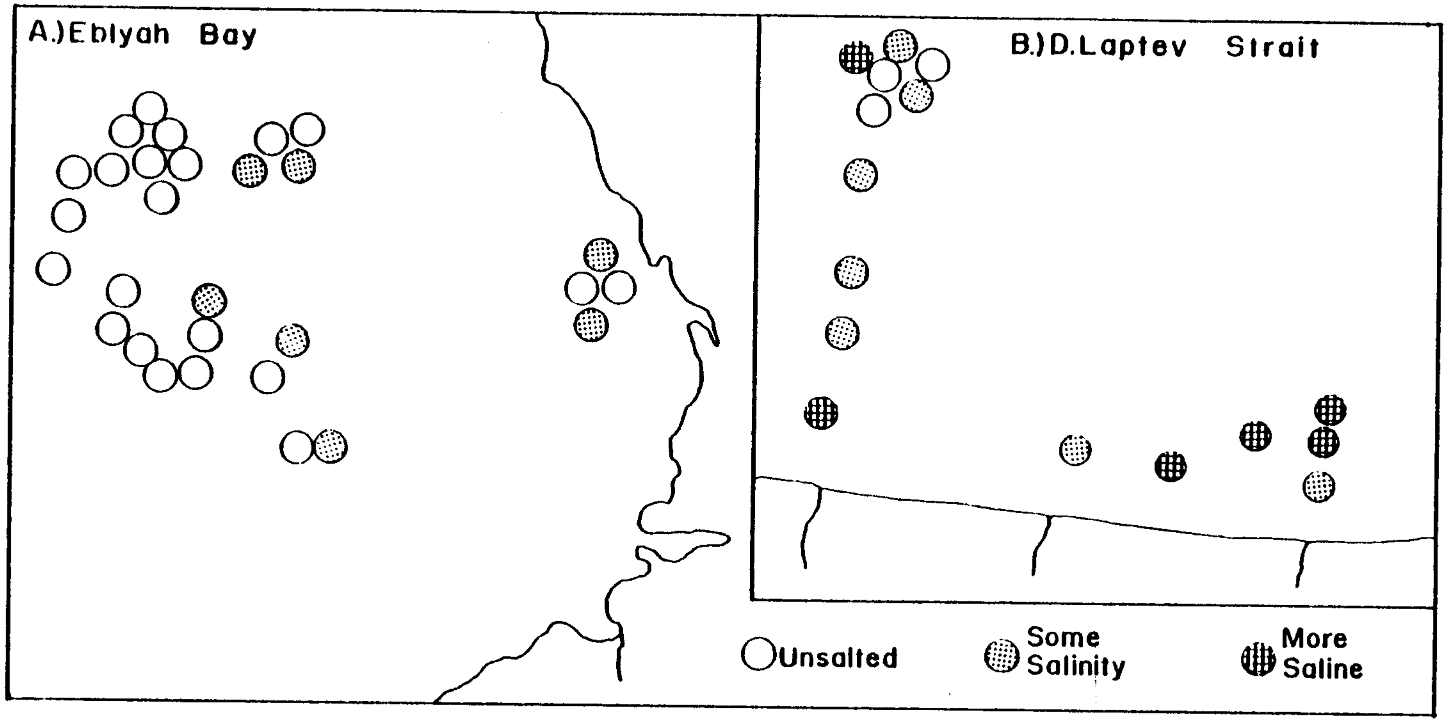


Figure 40.—Laptev Sea shelf with different concentrations of salt in the deposits. After Molochushkin (1975).

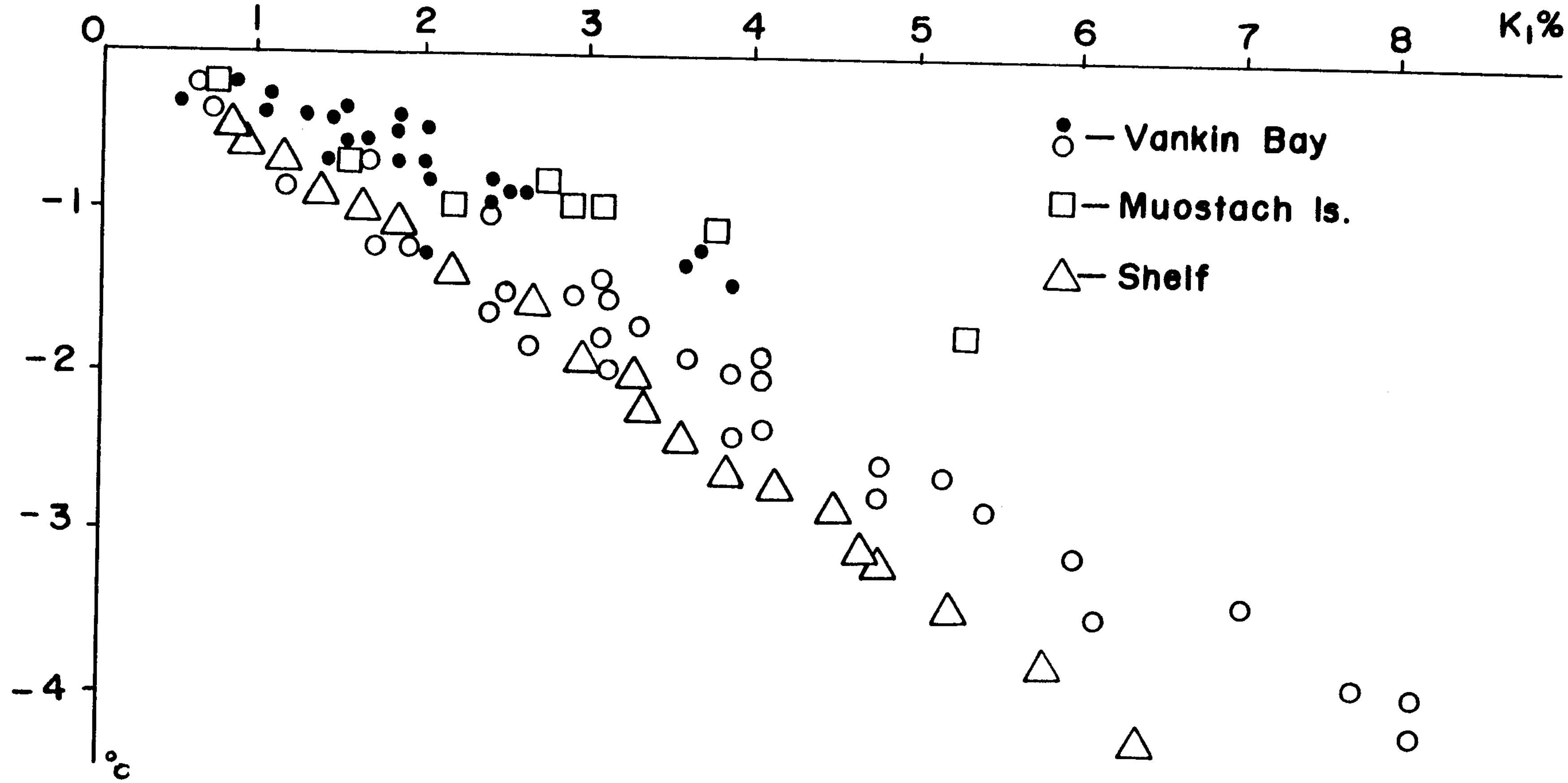
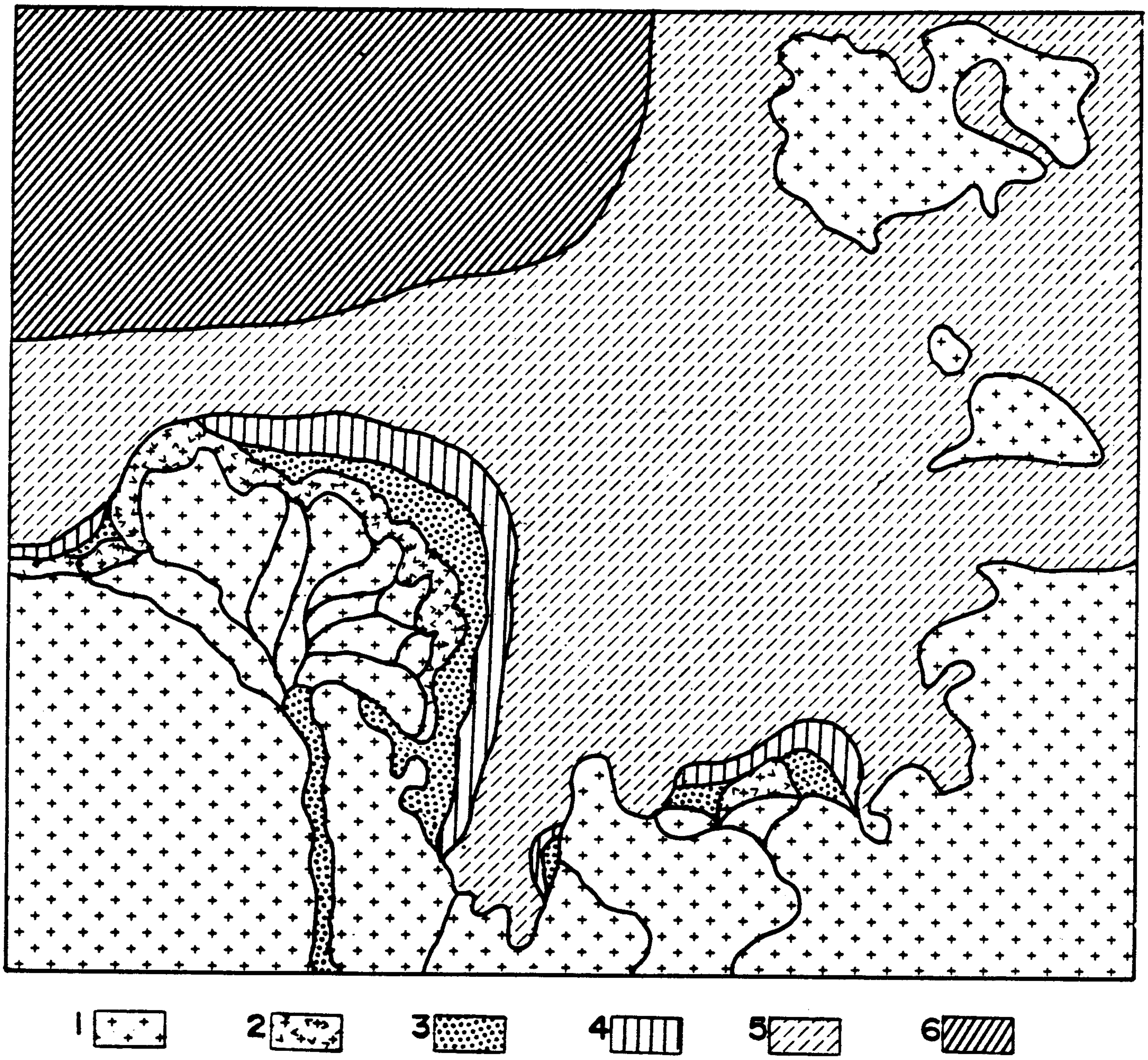


Figure 41.—Relationship of deposit freezing temperature with pore water salt concentration in different wells. After Molochushkin (1975).



1. Area of continental permafrost
2. Area of modern submarine permafrost
3. Thawed deposits (taliks)
4. Area of discontinuous "submerged" permafrost
5. Area of continuous "submerged" permafrost $h_{min} \leq 50$ m
6. Area of continuous "submerged" permafrost $h_{min} \leq 100$ m

(h = submarine permafrost thickness)

Figure 42.—Scheme of the different types of permafrost development in the Laptev Sea. After Molochuskin (1970).

of different types of permafrost development in the Laptev Sea according to Molochushkin's publications.

L. H. Zhigarev and I. R. Placht (1974) describe the "subaquatic cryogenic stratum" (SKT; term of S. M. Fotiev, see above). It is taken to mean the layer of shelf deposits that have been transformed by cryogenic processes. The SKT in the Laptev Sea (nearshore) lies on top of permanently frozen relict deposits inherited from the Pleistocene freezing epoch. This is confirmed to a certain extent by the nature of the distribution of salts and temperatures in the deposits, as determined with the help of shore and offshore test holes. The permanently frozen relict deposits apparently also lie under the positive-temperature rocks, but at a considerably greater depth.

The permanent cryogenic stage of the SKT in the Van'kina Gulf (Laptev Sea) is considered as a neoformation that has formed as the result of a Late Holocene retreat of the sea. Toward the Van'kina Gulf the vertical heat flow's effect increases, which is reflected in the gradient of the temperature curves. The temperature curve takes on a harmonic form only at a distance of about 0.5 m from the shore (Figure 43).

In the Laptev Sea this layer is characterized by seasonal and multiyear cryogenic stages. Within the seasonal cryogenic stage, it is possible to distinguish thawing, freezing, and cooled (negative-temperature rocks that do contain ice and that change into positive-temperature rocks, depending on the time of year) rock layers. Frozen and cooled rock layers are seen in the permanent cryogenic stage. Depending on their content of unfrozen water, the frozen rocks are subdivided into hard-frozen and semifrozen, with the relative amounts of unfrozen water being less than 22% and approximately 22–50%, respectively. A significant part of the Van'kina Gulf's water area is underlain by strata containing positive-temperature rocks. Zhigarev and Placht, in Figure 44, show the structure of the SKT in the offshore area of the Laptev Sea. On the surface in the beach zone there is a layer of seasonally thawing deposits 1.8–2.0 m thick that changes to permanently frozen deposits at that depth; toward the sea it changes to seasonally freezing deposits. This layer tapers off at a distance of 400–500 m from the shore. Under this layer there are seasonally cooled deposits ranging in thickness from 0.2 to 8 m. Near the shore, these deposits lie on permanently cooled rocks, while at a distance of more than 100 m from it, they are underlain by positive-temperature rocks. At a distance of up to 200 m from the shore, the positive-temperature rocks are characterized by average annual temperatures of 0–0.5 °C. In these rocks, the zero annual amplitude layer lies at depths of 9–10 m. Toward the shore, the positive-temperature rocks change into permanently cooled rocks that are delineated by the zero thermoisopleth. The slope of the permanently cooled rocks' roof (along the zero thermoisopleth) is about 60°. The depth of the zero annual amplitude layer in these rocks is 9 m, while the temperature at the bottom of this layer ranges from

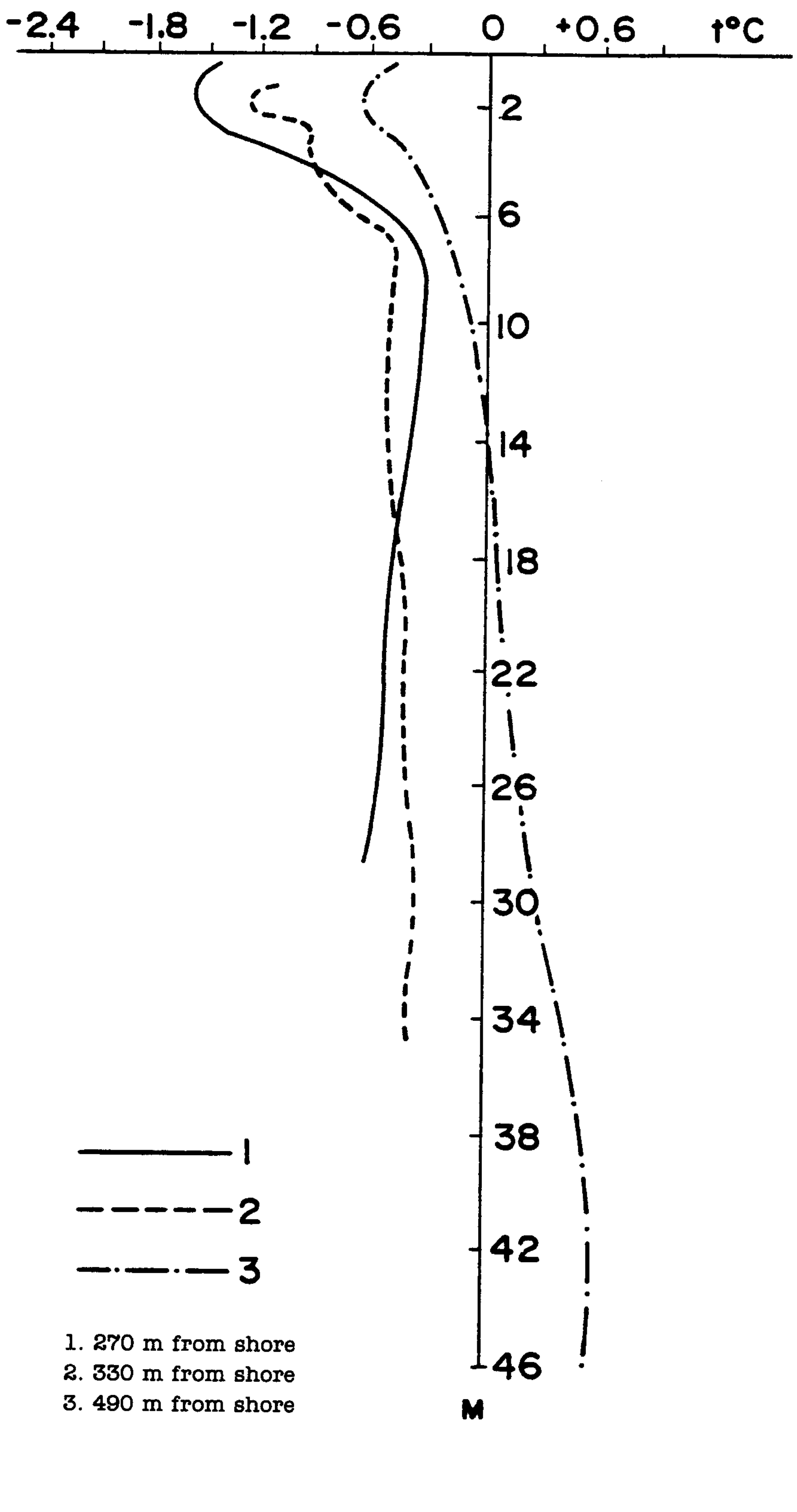
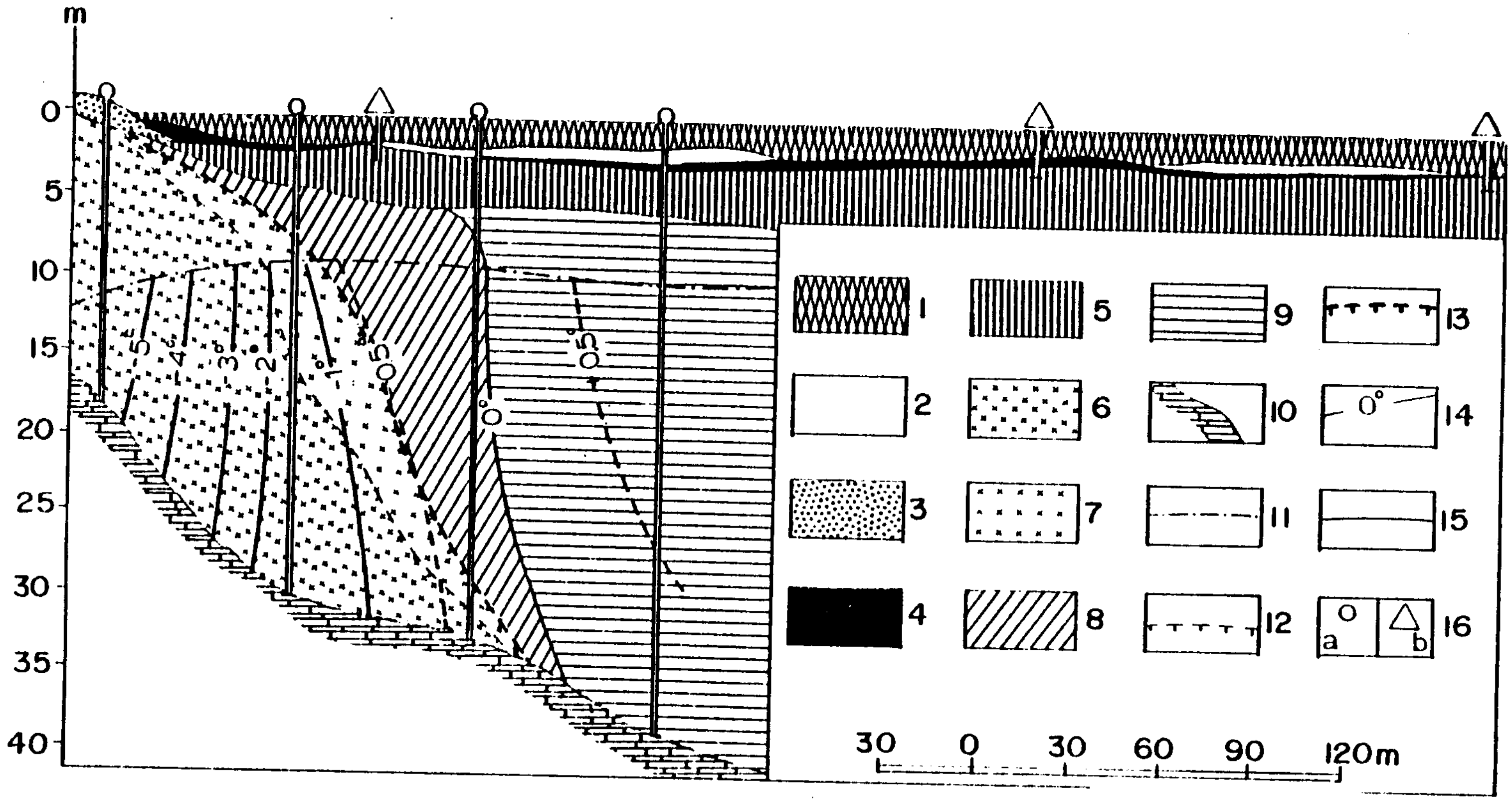


Figure 43.—Temperature distribution in the subaquatic cryogenic stratum. After Zhigarev and Placht (1974).



1. Ice cap
2. Water
3. Seasonally thawing layer
4. Seasonally freezing layer
5. Seasonally cooled layer
6. Permanently frozen rocks
7. Rocks in a semifrozen state
8. Permanently cooled rocks
9. Positive-temperature rocks
10. Bedrock
11. Boundary of layer of zero annual temperature changes
12. Boundary of permanently frozen rocks
13. Boundary of rocks in semifrozen state
14. Isotherms for 1.0 and 0.5°C
15. Zero isotherm
16. Test (a) and sounding (b) holes

Figure 44.—Structure of the subaqueatic cryogenic stratum in the Laptev Sea. After Zhigarev and Placht (1974).

0 to –0.8°C. In the coastal part of the shelf the permanently cooled deposits lie on top of permanently frozen rocks with a high (up to 22–50%) content of unfrozen water (semifrozen state). The slope of these rocks' roof reaches 30°. The semifrozen rocks' zero annual amplitude layer lies at a depth of 9 m, and the temperature at the bottom of this layer ranges from –0.8 to 1.8°C. Close to the shore, the semifrozen rocks change into hard-frozen rocks, which has been clearly determined by test holes. The slope of the hard-frozen rocks' roof is 25°. The depth of their zero annual amplitude layer drops from 9 to 12 m toward the shore, while their temperature ranges from –1.8 to –5.8°C. Describing SKT of the Laptev Sea, the authors write that the formation of the temperature field in the shelf deposits is determined primarily by the Sun's radiant energy, geochemical processes, and the Earth's internal heat. The effect of the Sun's radiant energy is much more significant and longer lasting during the winter than during the summer. It is sufficient to mention that in the eastern part of the Laptev Sea, the sum of the negative air temperatures is five 200–6,000 degree-days, while for the positive temperature it is only 110–600 degree-days. However, the snow and ice covers and the layer of seawater under the ice prevent substantial winter cooling of the shelf deposits. The greater the thickness of the snow and ice covers and the layers of seawater under the ice and the higher the salt concentration in the latter, the higher the total thermal resistance will be. As the salt concentration in the seawater and the bottom layer of sediment increases, not only is there a decrease in the freezing point temperature of the solution and the ground, but there is also an increase in their crystallization period. Therefore, during the entire time when the ice thickness is increasing, the cooling of the seawater layer is retarded to a great extent. According to Zhigarev and Placht, the salt content of the seawater in Van'kina Gulf is quite variable. During the summer period, in the absence of overtaking and colliding currents, it is 18–22%. After the formation of the ice cap in the fall and during its winter growth, the salt content gradually increases, reaching 50–72% in the spring (April–May). Later during ice melting, the seawater's mineralization decreases sharply to only 4–8%, but increases again after the ice returns. An increase in seawater mineralization can also take place when northerly currents are present.

During the summer, seawater salinity in the area of Muostakh Island (Molochushkin 1969) is 20–30 times less than it is in the Van'kina Gulf region during the same period, which is explained by the strong desalinization of the water caused by the discharge of the Lena River into Buorkhaya Gulf.

The thermal screening role of the snow, ice, and layer of seawater under the ice is shown for the eastern and southeastern parts of the Laptev Sea in Table 15. From the table, it is obvious that in the area of Muostakh Island at the end of December, when the ice thickness was increasing, the layer of seawater under the ice and the bottom sediment had rather high

Table 15.—Temperature distribution in bottom deposits in the Laptev Sea. After Molochushkin (1970), from data given by N. F. Grigor'yev (1964).

Time of Measurements	Distance from Shore (m)	Sea Depth (m)	Snow Thickness (m)	Ice Thickness (m)	Thickness of Seawater Layer (m)	Temperature (°C)		
						Air	Water	Sediment
				Van'kina Gulf				
April–May 1972	40	1.81	0.15	1.81	–	-8 to -12	–	-3.2
	80	1.96	0.14	0.85	0.11		-1.6	–
	280	1.88	0.10	1.82	0.06		-4.4	–
	480	1.89	0.21	1.80	0.09		-2.0	–
	680	5.48	0.22	1.87	3.61		-1.3	–
				Muostakh Island				
December 1963	60	1.50	0.00	1.25	0.25	-30	-0.5	-0.35
	220	2.00	0.00	1.20	0.80		–	-0.35
	400	3.00	0.00	1.15	1.85		-0.4	-0.4
	520	4.00	0.00	1.25	2.75		-0.3	–

temperatures under conditions of low air temperatures and the absence of snow cover on the surface of the ice. At the end of April and the beginning of May in the Van'kina Gulf region, the ice thickness reached its maximum value; that is, ice accretion was basically completed and, despite the comparatively high air temperature and the presence of a snow cover on top of the ice, the seawater layer under the ice and the bottom sediments were characterized by their lowest temperatures. In connection with this, we see that the temperatures depend on the thickness of the layer of seawater under the ice. In areas where the ice froze down to the bottom sediments, the latter had their lowest temperature.

The layer of seawater near the bottom has a negative temperature for 8.5–9 months (September–June). Based on an analysis of the temperatures and the nature of their temporal changes, as well as the seawater's salt content, it is possible to assume the sum of the negative temperatures to be 250–300 degree-days.

In contrast to the winter cooling, the summer warming of the shelf deposits is hindered only by the seawater layer; the greater the depth of the sea, the higher its total thermal resistance will be. In the Van'kina Gulf area, the depth of the sea varies from 1.5 to 10 m, averaging 2–3 m. During the summer, the temperature of this thin layer of seawater reaches 10°C at a depth of 6 m and 13–15°C at a depth of 2–3 m. Positive temperatures predominate in the bottom layer of the seawater in this region for 3–3.5 months (June–September), and total approximately 300–400 degree-days.

Consequently, the sum of the positive temperatures exceeds that of the negative temperatures by 50–150 degree-days. This gives an approximate average annual temperature for the bottom layer of seawater of 0.1–0.5°C, which is close to the average annual temperature of the rocks at a depth of 9.0 m. According to Ye. N. Molochushkin (1969), in the area of Muostakh Island the sum of the bottom layer of seawater's positive temperatures for a multiyear period exceeds that of the negative temperatures by 33–273 degree-days. This gives an average annual temperature for the water layer of 0.1–0.75°C, which is slightly higher than the average annual temperature in the Van'kina Gulf area; this is apparently explained by the warming effect of the water from the Lena River. At such temperatures for the bottom layer of seawater, the formation of permanently frozen and permanently cooled rocks because of vertical heat flows is eliminated. It is possible only because of a horizontal (from the shore) heat flow that intensifies as the sea retreats. A horizontal heat flow can lead to aggradation of the permanently frozen strata, while a vertical flow results in the opposite—its degradation.

The bedrock, and also the unconsolidated deposits (the thickness of which increases toward the water area of the Van'kina Gulf) are permanently frozen. For instance, in some sections of the inlet underwater bars, at a distance of 10 km from the shore, very icy permanently frozen relict deposits were found in the loose cover at a depth of 86 m.

As a result of the fact that negative temperatures with average values from −0.7 to −1.0°C predominate in the bottom layer of seawater for the greater part of the year, conditions for seasonal cooling and freezing are created in the upper part of the deposits. In this case also, however, the horizontal heat flow plays a substantial role, which is confirmed by the cryogenic structures of the seasonally frozen layer. The nature of the change in the seasonally frozen layer's thickness, from the shore toward the sea, is shown in Table 16. From this table it is obvious that in those cases where the ice freezes to the bottom deposits at the shore and in close proximity to it, the depth of the seasonal freezing is the greatest. As the distance from the shore and the thickness of the layer of seawater under the ice increase, the thickness of the seasonal freezing layer decreases. In those sections—even close to the shore—where a layer of seawater is preserved under the ice, the seasonal freezing layer is either very thin or generally absent.

Seasonal freezing begins on the shore during the fall, and gradually spreads toward the water. Judging by the temperatures of the bottom layer of seawater and the thickness of the frozen layer, seasonal freezing begins in the spring at a distance of 400–500 m from the shore, and is over at the beginning of the summer.

Close to the shore, the deposits in the seasonal freezing layer are dark gray, fine-grained, powdered sands with pebbles and rock debris. Toward the sea, the sands become even more powdery. The amount of pebbles and rock debris in them decreases, and they change to dark gray silt with a significant organic matter content. The moisture content of the sands at a distance of 270 m from the shore is 32–36% in the layer on the bottom, but drops to 26–27% at a depth of 0.4 m. The cryogenic structure of these deposits is not unidirectional, and is

Table 16.—Change in the seasonally frozen layer's thickness. After Molochushkin (1970).

Number of Measurement Range	Distance from Shore (m)	Thickness of Ice Cover (m)	Thickness of Seasonally Frozen Layer (m)
I	0	0.4–0.6	1.8–2.0
	160	1.55	1.15
	270	1.80	0.52
	330	1.80	0.30
	490	1.85	0.20
	680	1.87	0.00
II	60	1.81	0.74
	120	-	0.00
	280	1.82	0.18
	430	1.80	0.00

frequently thin (up to 1 mm) subvertical schlieren of ice up to 5–7 cm long that penetrate into the deposits to a depth of 20–25 cm. It is typical for highly moisturized, rapidly freezing sediments.

The seasonal thawing of the deposits usually begins during the first 10 days of June. Their average thawing rate at a distance of 500 m from the shore was the same at different points in the water area, and equaled 1.0 cm per day. However, since the thickness of the seasonally frozen layer decreases from shore to sea, thawing of the deposits was completed as follows: at a distance of 490 m from the shore, end of June; at 330 m, middle of July; at 270 m, end of July; at 160 m, end of August. The deposits thawed unevenly. The highest thawing rate was seen during the end of July, after the ice cap disappeared. The seasonal freezing and cooling layer is very dynamic, and its extent and thickness depend on the level of the sea at the moment when the ice cap forms and on the thickness of the ice. If the ice cap forms when the level of the sea is low, then the ice freezes to the bottom sediments and the seasonal freezing layer stretches a considerable distance out from the shore and is quite thick, while the freezing rate is high. If ice cap formation takes place several years in a row while the sea level is low, intergelisols (pereletoks) of frozen rock form and the seasonal freezing layer becomes a seasonal thawing layer. When the ice cap forms while the sea level is high and there is a layer of mineralized seawater between the ice and the sediments (which retards the seasonal freezing process), the seasonally frozen layer will form close to the shore and will not be very thick, while the freezing rate will be quite low. In separate years or for a number of years after the ice cap has formed when the sea level in the shore was high, it is possible for the seasonal thawing layer to break away and become a seasonal freezing layer. Consequently, the level of the sea at the moment of ice cap formation is the main factor in predicting seasonal thawing, freezing, and cooling.

There are no significant data on the value of the geothermal flux on the Eurasiatic shelf of the Arctic seas. This is due both to technical and methodological difficulties. The temperature gradient graphs which have been employed to determine the heat flux in deep water ocean basins are unsuited to shallow water shelves, where the temperature of near-bottom water layers is variable. Even at a water depth of about 1 km, the interpretation of data obtained with the aid of the temperature gradient graph can encounter great difficulties in individual cases (Lachenbruch and Marshall 1968). According to F. Are (1976), analysis of the presented data with consideration of modern concepts about geological structure and the developmental history of the Arctic shelf shows that the value of the flux of geothermal heat within the limits of the shelf is probably similar to the value on the coast; i.e., is approximately 0.05 kcal/(hr $\cdot$ m^2) with possible deviations of at least $\pm 14\%$. Much more significant anomalies are also possible, whose value is hard to predict; for example, in the tectonically active regions where the continuity of the permanently frozen relict deposits is disrupted along the faults and fracture zones. Mineralization of seawater and subterranean water has a great effect on the phase state of rocks which underlie the bottom of the shelf seas. We saw that the subterranean waters which are highly mineralized prevent freezing of the intervening rocks and form super-, inter- and subpermafrost water tables.

At the interface of the mineralized water table with the frozen layer at a temperature below zero, above the freezing point of the threshold solution, a fusion of ice which fills the pores of the rocks occurs. The intensity of fusion under natural conditions is unknown, but it is so slight that it is not identified by direct observations. An increase in mineralization of the threshold solution near the boundary of permafrost is necessary for development of the process of freshwater ice fusion and occurs through the medium of ion migration under the effect of the concentration gradient which diminishes proportional to the development of the process. Hence, in the absence of pressure filtration the process automatically slows. Therefore, the mineralized freshwater tables and permafrost layers can evidently coexist without significant movements of the freezing boundary over the course of geologically significant periods. All of the above also fully pertains to the interaction of sea water with permafrost which underlies the bottom of the sea. In the opinion of V. M. Ponomarev (1961), a solution of ice can only penetrate into the pores of rock under the effect of seawater at a depth of several meters. This is also indirectly confirmed by some laboratory experiments which showed that melting of ice-saturated dispersed rocks in contact with mineralized water at a temperature below zero slows with the increase in depth of thawing and the degree of dispersion of the rock, as well as with the decrease in mineralization of the water. Processes of erosion and accumulation in the coastal zone of the sea have a significant effect on the development of the permafrost on the shelf.

F. Are (1977) writes that as the result of thermal abrasion, the continental frozen layers of great thickness proved to be below the bottom of the sea. Their existence significantly depended upon the zone of prevalence of the warm and freshwater effect of rivers. It is known that river waters do not descend to a depth greater than 20 m on the shelf (Vize 1926; Vize et al. 1946; Gorbatskiy 1970; Sovetskaya Arktika 1970). Under the effect of river water, which has an above-freezing temperature, the continental-submerged permafrost layers are degradated. The approximate calculations of Ye. N. Molochushkin (1970) demonstrate that the frozen layer could either totally thaw or still be maintained at a certain depth from the surface of the sea bottom depending on distance from the shore and the rate of recession of the coast in such areas, by now. Outside the limits of the zone of effect of rivers, near the surface of the bottom, a freezing temperature constantly predominated which was significantly higher than on dry land. In the given case, a partial degradation of the frozen layer occurred from below with an increase in its temperature to an equilibrial state corresponding to the temperature of the near-bottom water. In the zone of accumulation of the contemporary bottom deposits, new permafrost layers can form. Their formation strongly depends on mineralization of the seawater. In the near-river mouth inlets, where the water is fresh or mildly saline, permafrost bottom freezing begins when water decreases to approximately 1.5 m.

Outside the limits of the zone of influence of the rivers, a still greater decrease in depth is necessary for the beginning of permafrost freezing. The precise shallow depth depends on salinity of the seawater and bottom deposits. In this case one should emphasize that the decrease in water depth with any mineralization of the seawater deposits leads to permafrost freezing of the bottom, since upon the shallows' reaching sea level the mean annual temperature of the surface drops to –10 to –12°C and the freezing point of seawater with a salinity of 30% is –1.7°C (N. N. Zubov 1944). In winter, in the closed coastal shallows, the salinity of the water increases proportional to the increase in thickness of the ice cover and reaches an anomalously high level; for example, about 74% (Molochushkin 1969). But even this water freezes at –4.4°C. Even if one takes into account that salinity of the threshold solution of the bottom deposits can be 1.5 times higher than salinity of the near-bottom layers of the water, it should be recognized that freezing of bottom deposits under ordinary marine conditions is unavoidable at temperatures of –10 to –12°C.

Arctic Shoreline Processes

Arctic shoreline processes, especially the reworking of shores in the permafrost zone, were considered by F. Are (1970, 1973), who shows that the processes of reworking of shores formed by permafrost have some peculiarities, connected with the rocks' frozen state and ice content. Particularly, the shores formed by Quaternary deposits with a large content of constitutional ice and ice-wedges (widespread in the Soviet Arctic area) have original development. If the main motive force of development of shores formed of the unfrozen rocks is the mechanical energy of waves, in reworking of shores formed by permafrost the thermal factor, which can cause the retreat of shores even in the absence of significant wave action, is the decisive factor. There are three main processes of reworking of shores formed by permafrost: (a) thermoabrasion—the destruction of shore zone under the action of mechanical and thermal energy of water; (b) thermodenudation—the destruction of cliffs under the action of air thermal energy and solar radiation; (c) thermokarst—the thawing of the basin's bottom under the action of thermal energy of water leading to subsidence of the bottom surface, conformably to reworking of shores. According to F. Are, the consequence of thermoabrasion is the shore's retreat: thermodenudation leads to the flattening of cliffs only; thermokarst, deepening the basin, promotes abrasion and, in certain conditions, interacting with thermodenudating, leads to shore retreat without thermoabrasion.

The specific of reworking of shores formed by permafrost depends on the ice content in rocks. Shores formed by ice only retreat very rapidly. Thus, in the permafrost zone there is a critical value of total ice content for each shore, above which the shore becomes unstable. After full thawing of rocks having critical ice content subsidence level may coincide with water level in the basin.

If the subsidence level is above the water level, the shore is stabilized and can retreat under the action of abrasion only. In this case the end profile of the submerged shore slope forms when the thermokarst process is over. If the subsidence level is lower than water level the permafrost thawing under the basin inevitably leads to its exposure and thawing in the cliff. In such a case the shore is unstable and will retreat even without the action of abrasion. F. Are shows also that the comparative roles of thermoabrasion, thermodenudation, and thermokarst in the processes of shore development depend on climatic conditions. So the regional peculiarities of shore reworking are defined by the climate. The climate of Arctic coastal lowlands is characterized by high winds and cold summers. The temperature of water in the basins is low. That is why thermoabrasion plays the main role in shore reworking in these regions. The rate of shore retreat has reached great values. The thawing of basin bottom is developing slowly. Closed taliks are widespread even under the bottom of large lakes. The climate of the eastern Siberia taiga zone is characterized by weak winds, hot summers, and large insolation. In these regions the role of thermoabrasion in shore reworking is unimportant. Basically, the shores retreat under the action of thermokarst and thermodenudation at a slower rate than in the northern coast, other things being equal.

According to the form of the profile, the cliffs may be divided into four main types: sloping, sloping with plumb lower part, plumb, and stepped. The cliff form is the approximate indicator of shore retreat rate. In this respect the stepped cliffs give the greatest possibilities (Are 1968). If the cliffs are plumb and have wave-cut notches, or there are remains of fallen blocks at their foot and there are no large accumulations of products of thermodenundational destruction, then it means that the shore is retreating rapidly, and the rate of retreat exceeds the intensity of thawing of permafrost exposed in the cliffs. If the cliffs are sloping or sloping with plumb lower part, with ice-wedges exposed on their surface and there are no large accumulations of thermodenudational destruction products at their foot, the rate of shore retreat is approximately equal to the rate of thermodenudation. If the cliffs of any form have large accumulations of thermodenudational destruction products at their foot, the rate of shore retreat is less than the rate of thermodenundation.

In the process of thermodenudation of cliffs formed in the ice complex, thawing of exposed enclosing rock material may be slower than the thawing of the ice wedges, but on the whole the rate of thermodenudation of such cliffs is defined by the intensity of ice-wedge thawing.

This value may be easily calculated by climatic characteristics and be used for approximate definition of the rate of retreat of shores formed in the ice complex on the basis of indications stated above. On the Yakut coast of the Laptev Sea the rate of thermodenudation is 4–5 m a year (Are and Molochushkin 1965), and in central Yukutia is 9–10 m a year, and along the shores of the Laptev Sea does not exceed 4–6 m a year. In Arctic coastal lowlands the rate of lakeshore retreat has reached 10 m a year under the active influence of thermoabrasion (lakes more than 1 km across), and under the weak influence of thermoabrasion 1–3 m a year (Tomirdiaro, Ryabchun, and Golodovkina 1960)

Seawater Salinity Changes during Ice Formation in Shallow Seas

An empirical method was suggested by L. O. Kuzenkova (1972) for calculating water salinity during the winter period. This method is based on taking into account the increase in salinity due to variation in the volume of water during ice formation. The calculation formula accounts for the relation of ice thickness with the ocean area depth at which the ice is formed. The necessity of such a method is connected with the lack of any winter measurements. In a shallow sea the ice is usually fresh rather than saline. This means that the salt concentration in the seawater beneath the ice increases with the process of ice development. The winter salinity could be defined according to this empirical formula:

$$(5) \qquad \frac{V_1}{V_2} \quad S_1 = KS_1 = S$$

V_1 - Volume of the water before the ice development

V_2 - Volume of the water after the ice development

S_1 - Water salinity before the ice development

S - Water salinity after the ice development (winter salinity)

$V_1 = Fh$; F is the surface and h is the depth of the sea area.

$$(6) \qquad V_2 = V_1 - V_3 = V_1 - 0{,}9\ V\ ice = Fh - 0{,}9\ F\ \bar{H} = F\ (h - 0{,}9\ \bar{H})$$

$\bar{H}$ - Average thickness of ice

V_3 - Volume of water for ice development.

(7)
$$K = \frac{V_1}{V_2} = \frac{Fh}{F\,(h - 0{,}9\,\bar{H})} = \frac{h}{h - 0{,}9\,\bar{H}}$$

(8)
$$S = \frac{h}{h - 0{,}9\,1\,\bar{H}} \quad S_1$$

Practically, it is possible to use the formula of the special graphs (Figure 45) of the salinity changes.

(9)
$$\Delta S = S - S_1 = S\,(K - 1) = S \frac{0{,}9\,\bar{H}}{h - 0{,}9\,\bar{H}}$$

We see that the parameter ΔS directly depends on changes of the water volume and initial salinity.

DYNAMICS OF THE ARCTIC SHELF AND COASTAL AREA DEVELOPMENT DURING THE PLEISTOCENE

Relative changes of sea and dry land level cause transgression and regression of the sea and, consequently, the filling or drying of the ground surface, which fundamentally alters the thermal regime of the rock and the hydrogeological conditions of rock development. A vast number of scientific investigations whose results in many cases are contradictory have been devoted to a history of the marine transgressions either directly or indirectly in the geological past of the earth.

The Main, though Contradictive, Concepts

An important and urgent scientific problem is the question of the Quaternary emergence–submergence history being the same for the Arctic Ocean as for the World Ocean, or during some stages did the Arctic Ocean develop separately. This question could seem to be strange for the geologists accustomed to considering the Arctic Ocean as a constant (in Pleistocene) and "unseparable" part of the World Ocean. But during the last decade or two most of the Soviet geologists studying Pleistocene deposits in northern Eurasia (including primarily western and northern Siberia and the northeastern part of Europe) believe it proved these deposits were formed mainly by repeated marine transgressions which were simultaneous to major glaciation stages in northern Europe and America. These scientists, so called "marinists," are represented by people of the All-Union Scientific Oil Research Institute in

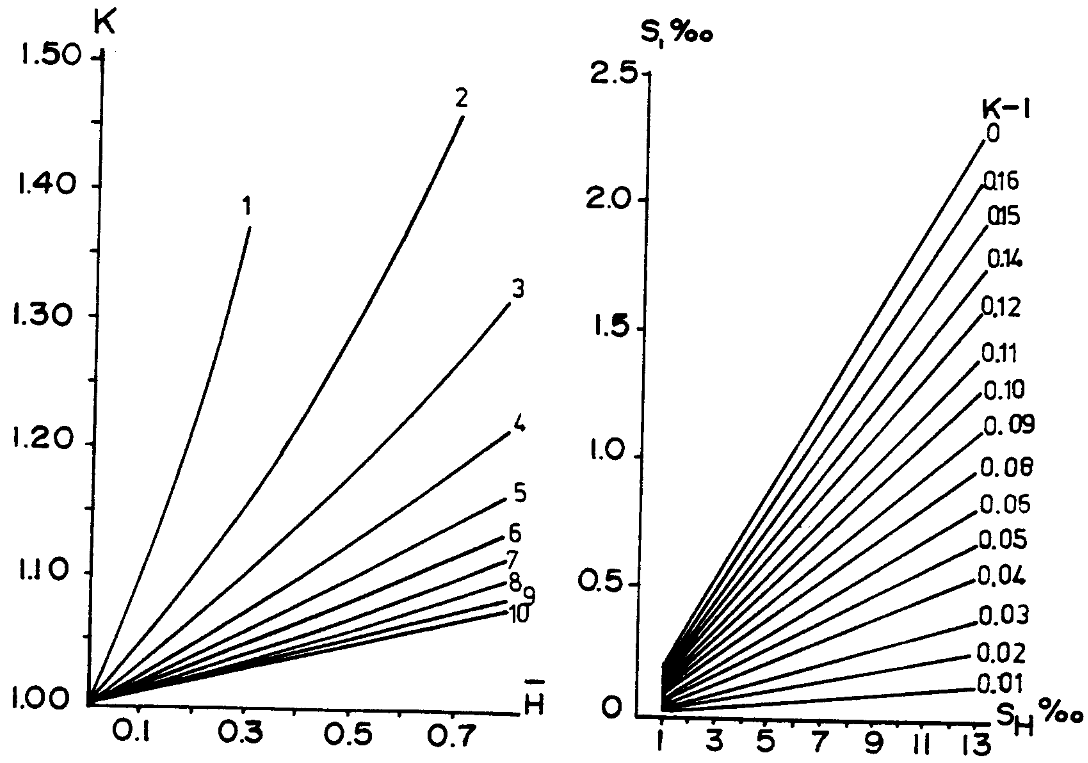


Figure 45.—Graphs for salinity calculations for sea areas under ice. After Kuzenkova (1972).

Leningrad, Scientific Research Institute of Arctic Geology, also in Leningrad, USSR Geologic Survey, Moscow University (Geographical Department), and in other institutes. The best known works connected with "marinistic" conception are written by G. Lazukov, A. Popov, I. Kusin, I. Zagorskava, O. Suzdalsky, V. Zubakov, and many others. Their opponents or "glaciologists," are represented mostly by the scientists of the Novosibirsk branch of the Academy of Sciences of USSR: S. Troitskyi, S. Strelkov, V. Saks, and partly, by S. Archipov.

Some of the marinists or "antiglacialists," following the extreme point of view, exclude the role of the glaciations in the development of the Siberian plains and consider only the marine and glacial marine activity. Some of the glaciologists exclude the role of the "cold" marine transgressions. A critical analysis of the marinistic concepts was published in 1975 by S. Troytskyi. The author himself summarizes the ideas and assumptions of his book the following way:

The book *Modern Antiglacialism, Critical Essay* is concerned with the first critical analysis in Russian geological works of the modern antiglacialism conception, i.e., the scientific trend that contradicts the theory of Quaternary glaciation of the plains in the moderate and subarctic belts. The author consistently examines general ideas, assumptions, and data of antiglacialists. He reveals the internal contradictions in their conceptions and comes to the conclusion about a very great significance of the sheet glaciers for lithogenesis and morphogenesis on the north Siberian plains.

Sometimes the discussion about the role of the glaciations and marine transgressions in the development of northern Eurasia during the Pleistocene was very sharp. But the results of these discussions are the new data and new ideas connected with the basic problem of the Arctic basin geological history of that time.

We have tried to summarize the data and ideas of the World Ocean level changes during the Pleistocene and also data and assumptions of the Arctic Ocean level changes during that time. As a result of this work we made a conclusion that it is possible to get agreement between "marinistic" and "glaciologistic" conceptions; of course, if we do not follow the extremes.

Trends in Ocean Level Changes

Three types of oceanic curves are "catastrophic," glacio-eustatic, and "reservoir." Figure 46 provides a summary of the World Ocean level changes during the Pleistocene (Vinkovetsky and Vigdorchik 1971). If the Russian geologists made a considerable input in the knowledge of the Arctic Ocean history, using their own data, the basic assumptions on the World Ocean development in the Pleistocene were made by them according to the data gathered by western scientists. Usually, studying the glacial period, we are primarily concerned with glacio-eustasy, realizing, however, that glacio-eustatic changes of sea level may have been superimposed on a speculative long-term fall in sea level that commenced in the Tertiary and earlier (Figures 47 and 48). In the Western geological literature J. T. Andrews (1975), for instance, noted that part of the fall in sea level was probably related to the glacierization of Antarctica, Greenland, and other parts of the world in the period of the early to mid-Tertiary; but this glacierization can account for only a sea-level drop of about 80 m. From our point of view the main trend in lowering of the ocean level is connected with regular periods of the Earth extension (P. Dirak theory). Anyway, the glaciation of Antarctica commenced long before the Pleistocene, perhaps even in the early Tertiary. Mercer (1968) suggests that the East Antarctic ice sheet developed first, followed by Greenland, and finally by another part of Antarctica. Depending on which value one accepts for the volume of the existing ice sheets, the glacio-eustatic sea level lowering led to a 100-m drop in ocean level; partial mitigation of this drop resulted from isostatic compensation of the ocean basins, yielding a net reduction of about 70 m. The most significant drop of the ocean levels took place about 700,000 years ago (Matuyama–Brunhes

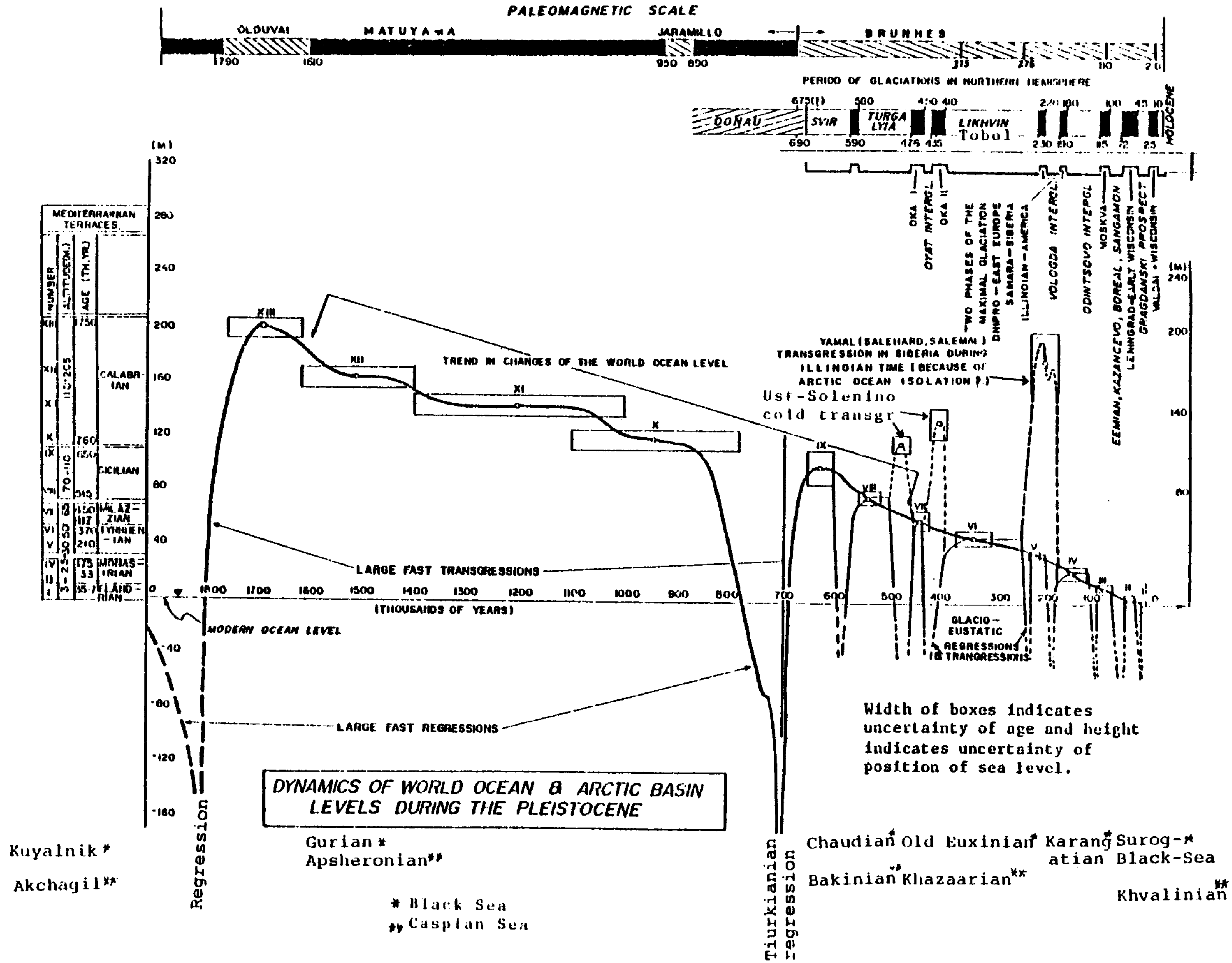


Figure 46.—Dynamics of World Ocean and Arctic basin levels during the Pleistocene. After Vinkovetsky and Vigdorchik (1971).

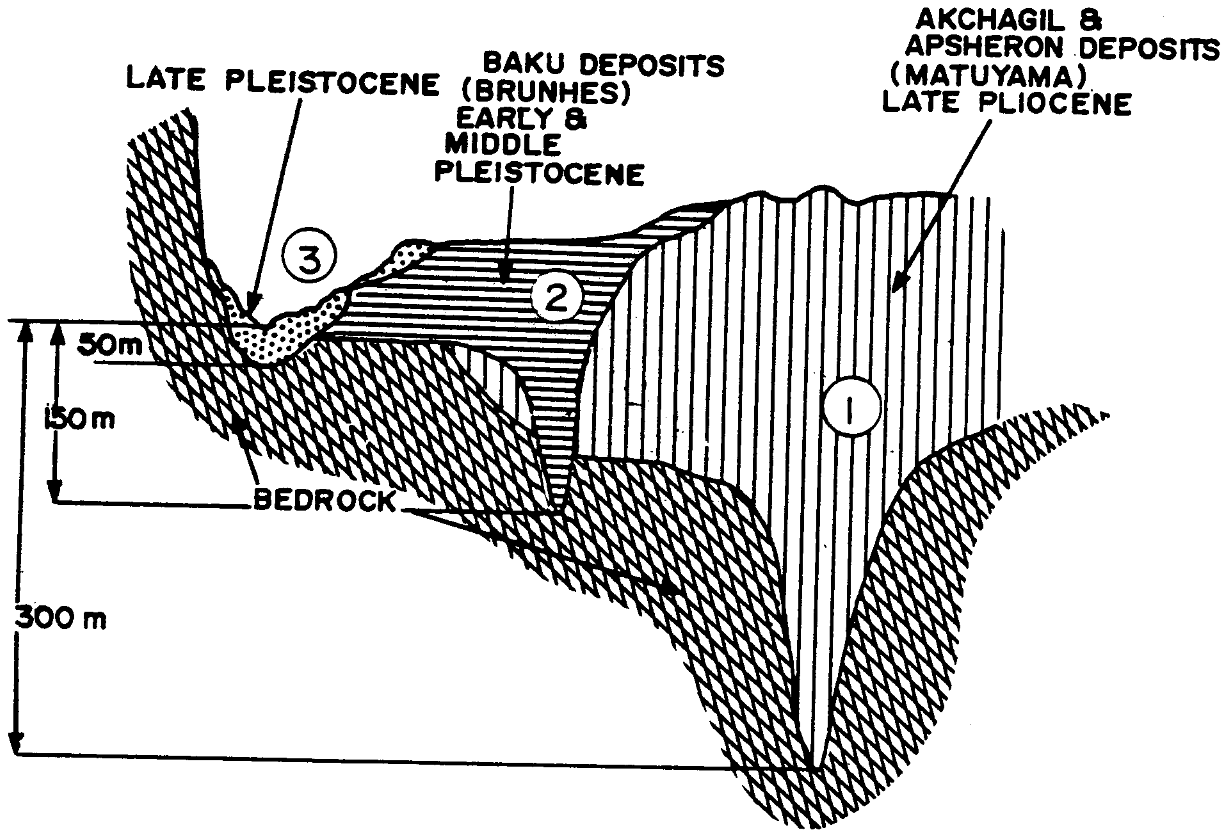


Figure 47.—Different generations of river valleys. After Obidientova (1971).

boundary). The evidence of this event is the numerous ancient valleys especially in coastal and shelf zones of the ocean; the bottom of these valleys reaches minus 150–300 m. They have typical canyon forms and are usually buried by the Pleistocene deposits (Figure 49). The pictures of the network of these canyons are shown in Figures 50–52. We see that all these data give evidence of a "catostraphically" fast ocean level drop with a magnitude of about 250–300 m at the time of the Matuyama–Brunhes boundary and also before Olduvdy time.

Pleistocene sea level oscillations are best preserved along coasts that are slowly rising, thereby carrying earlier sea levels and associated deposits beyond the reach of later wave action. The classical sequence of supposed glacio-eustatic terraces is that described by Deperét (1918) from the shores of the Mediterranean Basin and for the Russian geologists, it is typical to use his scheme. The terraces were considered to represent worldwide former sea levels and the general fall of sea level was explained by the progressive enlargement of the ocean basins. There were also many attempts to fit observed sea level changes into the classical European and North American four-stage glacial/interglacial sequence (Table 17). Recent work now shows evidence that few coasts and land areas are stable and in part because evidence from ocean cores indicates that the Quarternary has been characterized by at least eight temperature oscillations compared with the Wisconsin glacial maximum (Figure 53; Andrews 1975).

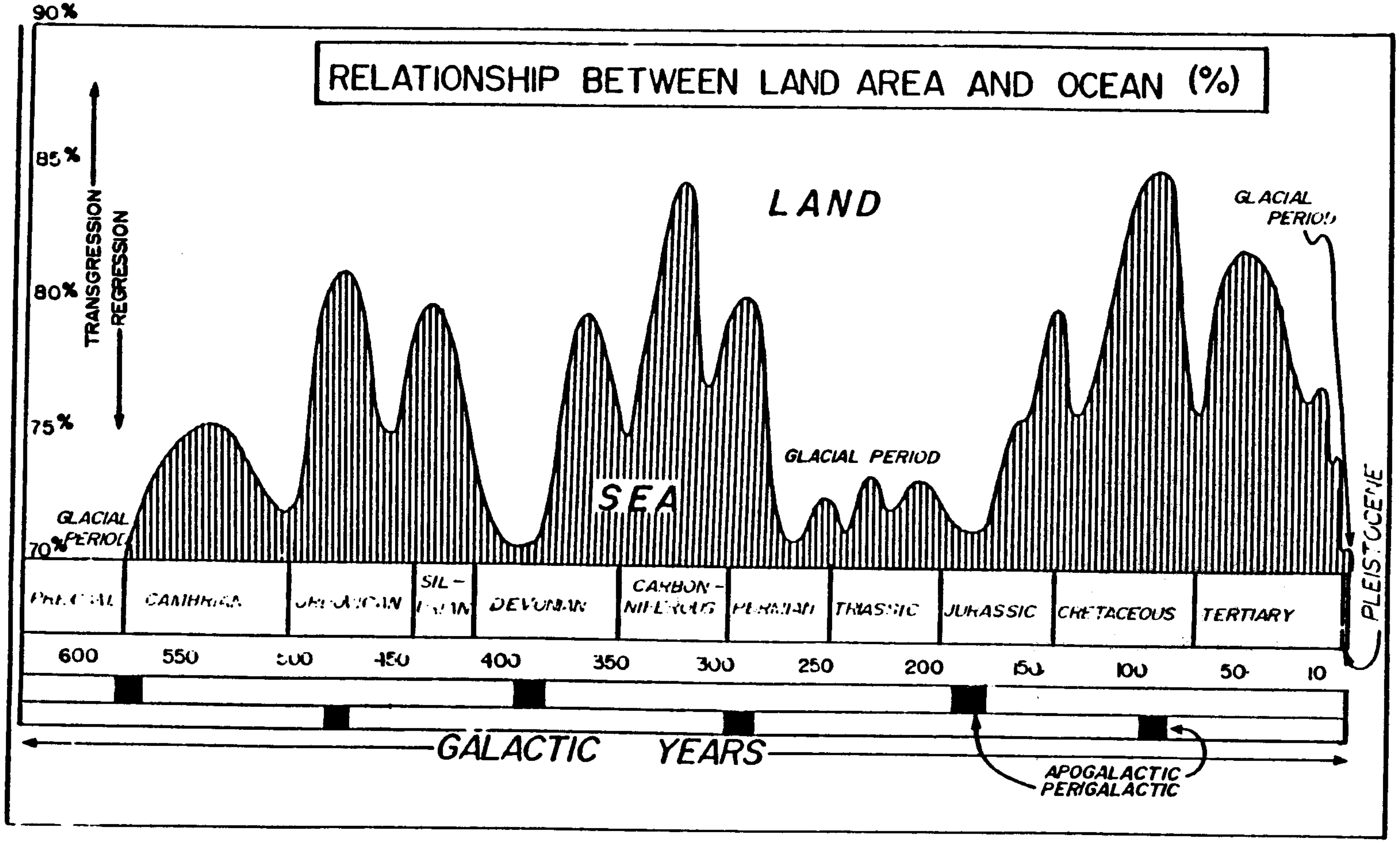


Figure 48.—Regressions and transgressions of the World Ocean in Earth history.

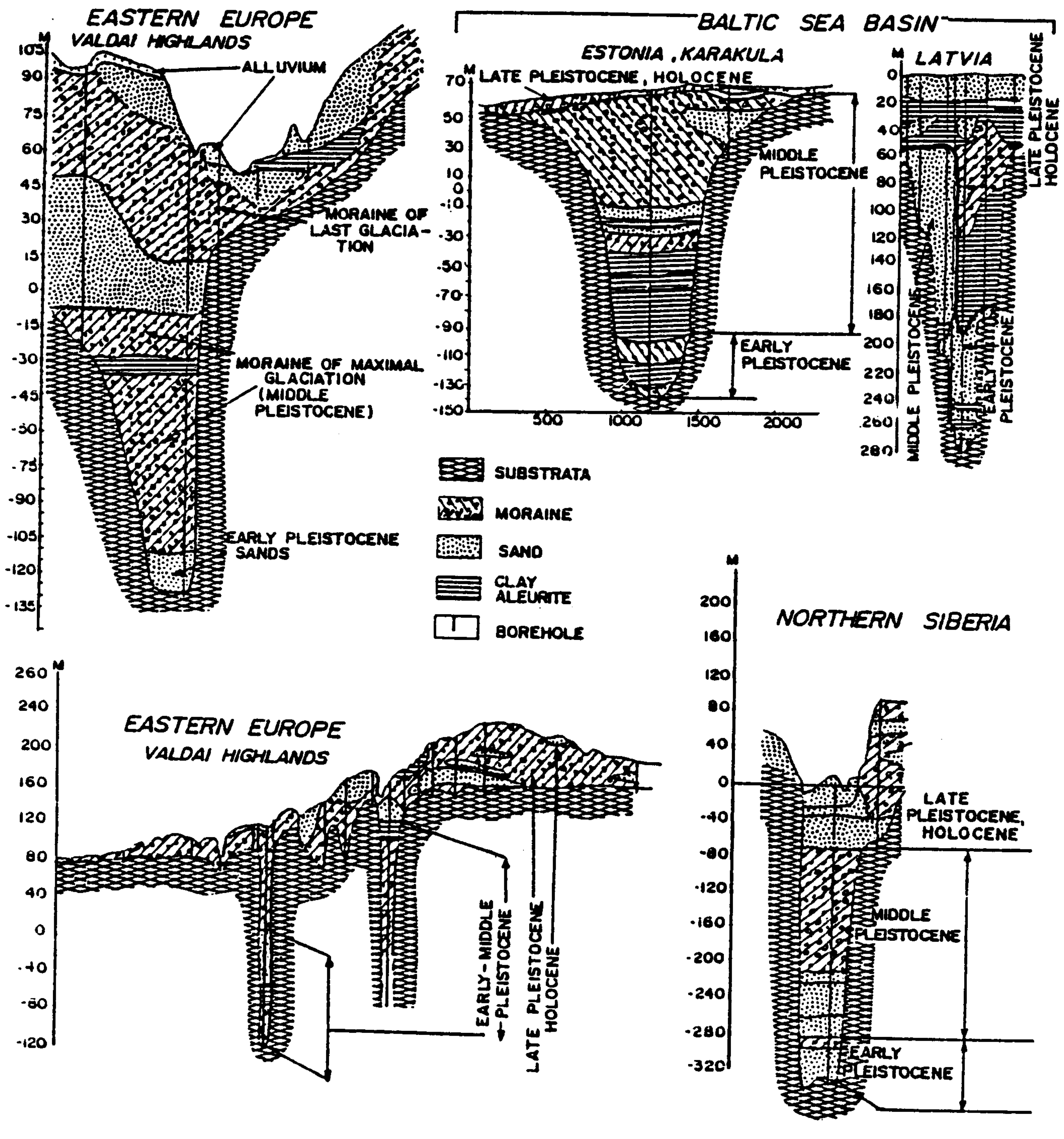


Figure 49.—Ancient canyons buried by Early, Middle, and Late Pleistocene deposits in northern Eurasia.

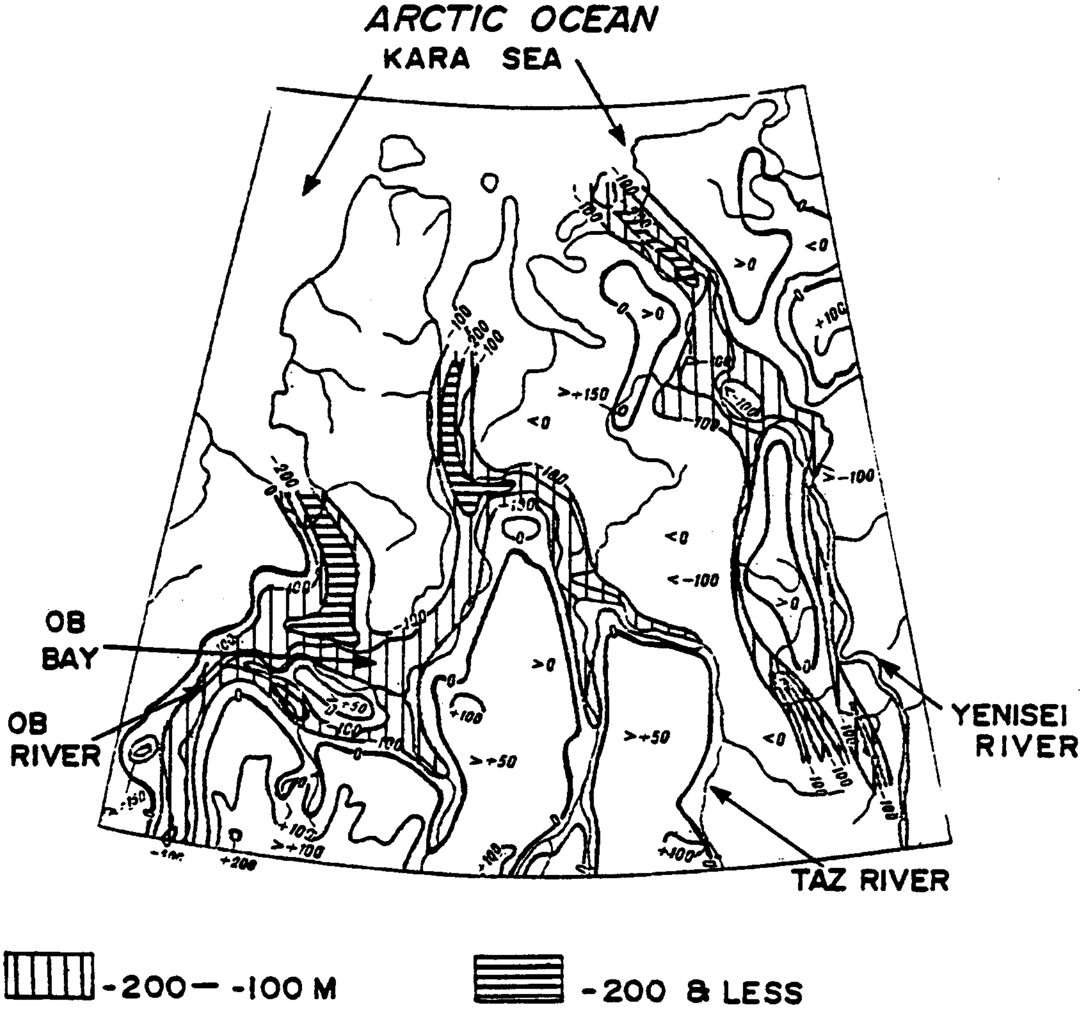
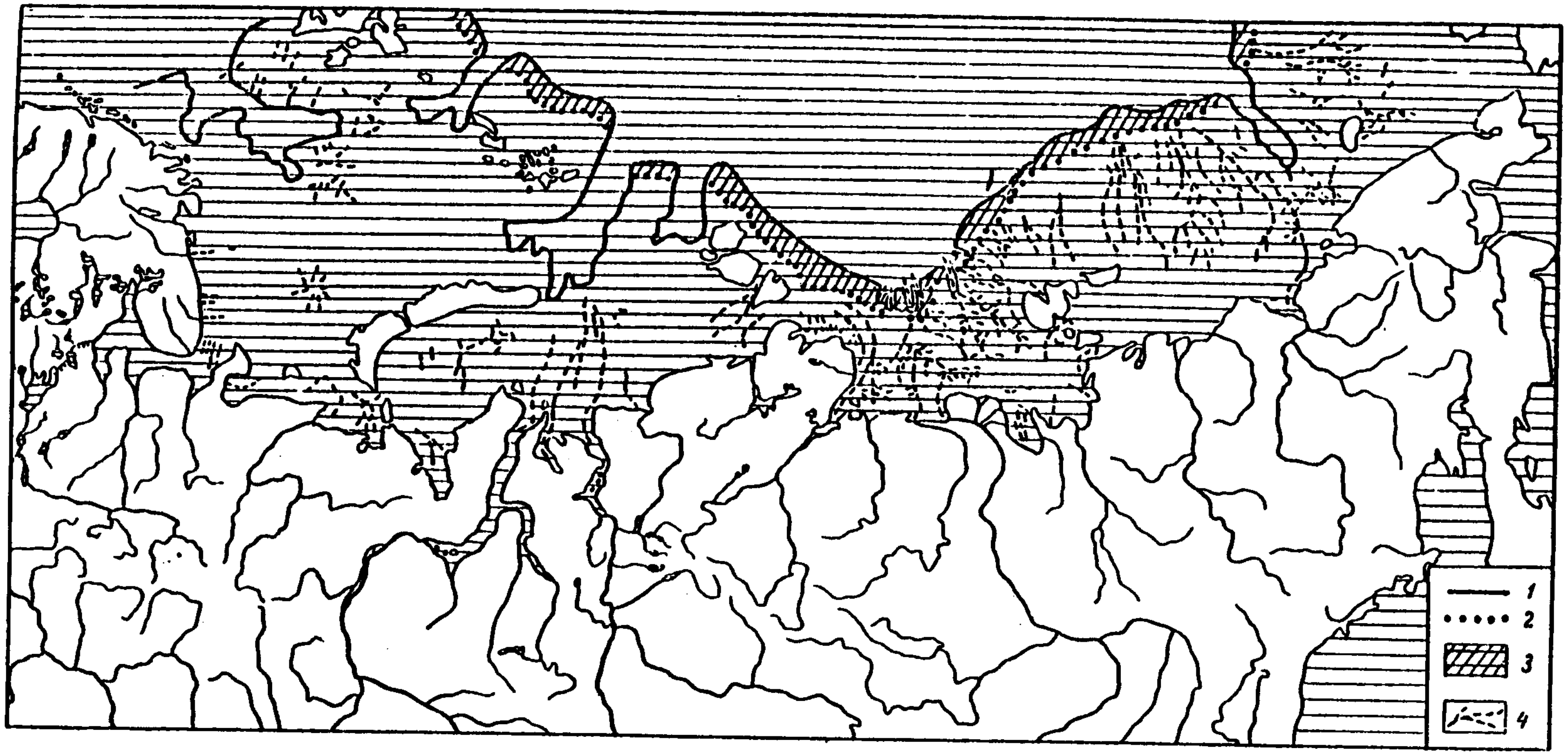


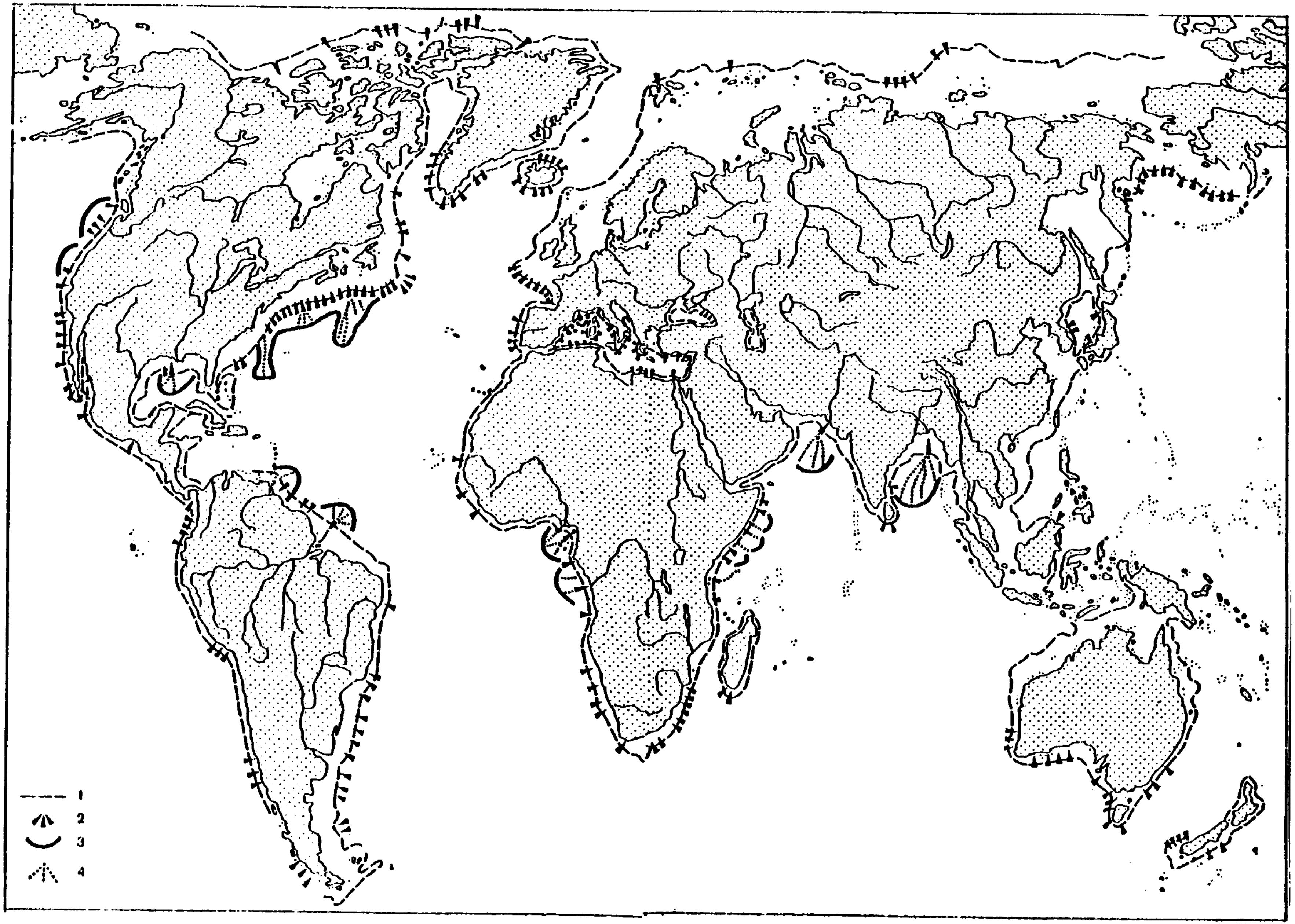
Figure 50.—Map of the ancient valleys (Cretaceous and paleogen rocks roof) in western Siberia (Suzdalsky 1972).

Some guide as to whether a former sea level terrace represents simply a high eustatic sea level or a tectonic uplift can be derived from the knowledge that much of the Antarctic ice sheet has remained quasi-stable during the Quaternary, with the possible exception of the West Antarctic ice sheet that is grounded below sea level; if that portion of the ice sheet melted, sea levels would rise about 5 m. Emiliani argues that world sea level may rise to a maximum more than once during each interglacial age in response to the melting of the Greenland ice sheet during general interglacial conditions when the northern summer solstice occurs at perihelion. If this argument is correct, the sea level would rise by an additional 7–10 m. Thus, any marine terrace over 15 m above sea level (the combined sea level storage of water in Greenland and West Antarctica) undoubtedly reflects uplift of the coast. Another possible



1. Boundary of the continental slope 2. Boundary of the shelf 3. Continental slope 4. Submarine canyons

Figure 51.—Network of ancient submarine canyons on the Arctic shelf (Lindberg 1970).



1. Shelf boundary 2. Submarine canyon 3. Alluvial fan of submarine canyon 4. Hollows of the turbid streams

Figure 52.—Ancient canyons on the World Ocean shelf (Leontiev 1970).

Table 17.—Mediterranean Terraces—Age and present elevation. After Andrews (1975), from Deperét (1918) and Guilcher (1969).

Terrace	Elevation (m)	Glacial Stages	Interglacial Stages
		Nebraskan	
Sicilian	80–100		Aftonian
		Kansan	
Milazzian	55–60		Yarmouth
Tyrrhenian	30–35		
		Illinoian	
Motastirian	15–20		Sangamon
	0–7		
		Wisconsin	

explanation for former high sea levels of 10–17 m above present is suggested by Hollin (1976), who calculated that a surge of the Antarctic ice sheet would raise sea level by that amount. Sea level records older than 200,000 years BP exist at numerous sites, including Alaska (Hopkins 1968, 1973). Of course the absolute dating is very difficult. The knowledge of marine changes in the interval between 250,000 and 45,000 year BP in the Soviet Union recently increased, owing to the development of the uranium-series methods of dating marine carbonates and the thermoluminescent method of dating the mineral deposits.

Summarizing the data, J. Andrews (1975) notes that the evidence for glacio-eustatic low sea levels during the Pleistocene (Brunhes time) is off New England, where five submerged shorelines have been described between 23 and 144 m lower than present. Oolithic beach ridges at 70–90 m lower than present were formed off eastern Florida during the Holocene transgressions. The maximum amount of glacio-eustatic sea level lowering during the Pleistocene in now known. Estimates range from 100 to 159 m below present, the latter on the basis of estimated ice volume during the Illinoian glaciation. Calculated ice volumes are usually maximum estimates of sea level lowering because of the isostatic compensation of the ocean basins; thus the estimate of 100–159 m should lead to an estimated lowering of sea level relative to the present of about 100 m. Best estimates for the Wisconsin glaciation place sea level between 100 and 135 m below the present (Andrews 1973).

Geologists are once again turning their attention to the correlation between variations in the incoming solar radiation and glacial/interglacial sequences. This trend has led to re-examination of the Milankovitch hypothesis, that glaciations are triggered by variations in solar radiation. Several authors have considered the relationship between high sea level stands and various refinements of the Milankovitch insolation curve. In particular, the three

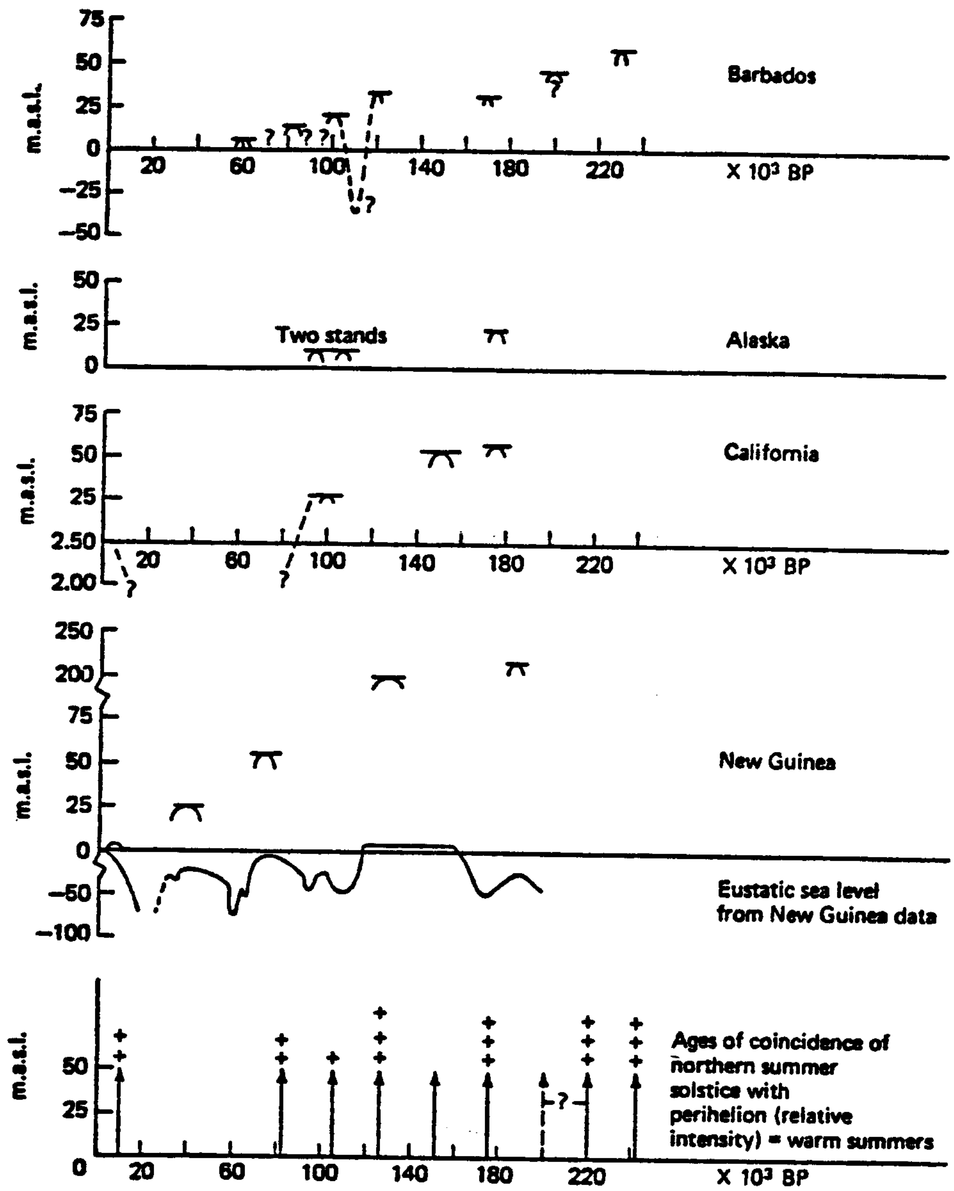


Figure 53.—Comparison of the marine terraces from various regions (Andrews 1975).

Barbados terraces at approximately 80,000, 100,000 and 120,000 BP correlate with three strong insolation maxima. The insolation peak at 150,000 BP is not strong and there is no evidence of a reef at this age in Barbados, Alaska, or New Guinea. These sea level fluctuations should, of course, match the sequence of glaciations, interglaciations, and interstades.

D. Hopkins (1967) presents a detailed sequence of the Quaternary marine transgressions that are discernible on the Alaskan coasts in Beringia (Figure 54). The Kotzebuan transgression is dated about 170,000 BP, whereas the succeeding Pelukian event is dated at about 100,000 BP and comprises two distinct sea level stands for Alaska. O'Sullivan (1961) developed a sequence of depositional history based on various land surfaces and lithologies and suggested at least four major marine transgressions. McCulloch (1967) investigated the depositional history of the northwestern part of the Arctic Coastal Plain and reported evidence for five transgressions. Hopkins' (1967, 1973) reviews of the Arctic Coastal Plains investigations and their correlation with other parts of coastal Alaska proposed that there appeared to be evidence for as many as seven transgressions during the Quaternary. The new data on the stratigraphy and diagenesis of perennially frozen sediments of the Barrow area was given in 1972 by P. Sellman and J. Brown (Figure 55, Table 18). T Péwé (1976) has given a summary of Alaskan Quaternary geology, including the problems of the history of Arctic Ocean transgressions and regressions.

Cold "Reservoir" Type of Marine Transgressions

Figures 56 and 57 show the extent of the cold* marine transgression (Yamal transgression) and the last interglacial warm marine transgression (Sangamon, Eemian, Boreal, or Kazantevo) in eastern Europe and Siberia. The greater extent of the cold marine transgression compared to the warm interglacial one is clear. Figure 58 shows the boundaries of the Yamal cold marine transgression and the areas covered with simultaneous glaciation on the Eurasiatic part of the Arctic basin after A. Popov and A. Kostyaev (Markov 1965). In Figures 59 and 60, we can see the cross sections with the cold marine transgression deposits between the two series of warm marine transgression (interglacial) deposits—Holstein and Eemian (Boreal, Sangamon). One cross section is from the confluence of the Irtish and Ob rivers about 800–1,000 km to the south of the modern Arctic Ocean, the second is from 100–200 km to the south of the ocean, near the Ob River.

The age and the facies relationship between the areas of transgression and glacial deposits are shown in Figures 61, 62, 63, and 64. The marine and glacial marine deposits are clay, loam, sandy loam, and sand. Their absolute age is 170–190,000 and 220–230,000 year ago, according to U^{234} and thermolumniscent analyses (Zubakov et al. 1974). Their thickness reaches about 220 m. The deposits usually have a normal magnetic polarity, but the study is continuing. The marine mollusks and cirripeds include *Balanus hameri, Nucula tenuis, Leda pernula,*

* The organic remains associated with these transgressions (shells, diatoms, foraminifera, pollen) were clearly indicative of sea and air temperature colder than during the Holocene and Boreal (Eemian/Sangamon) hypsithermals.

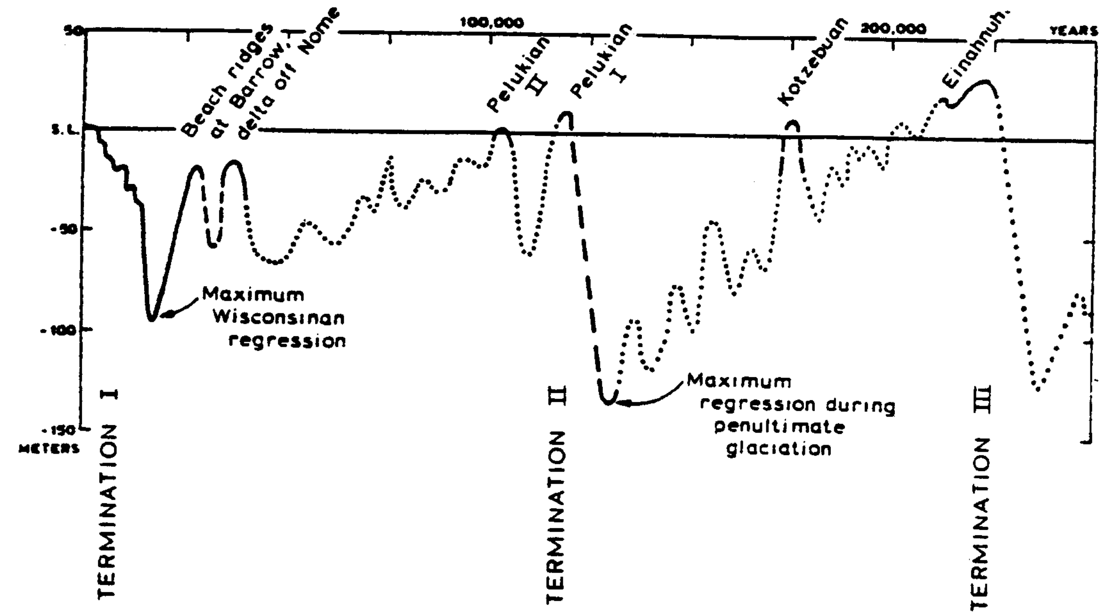


Figure 54.—Sea level history in Beringia during the last 250,000 years (Hopkins 1967).

Macoma calcarea, Mya truncata, Saxicava arctica, Astarte crenata, A. Montagui, Portlandia arctica, Joldiella lenticula, Propeamussium groenlandicum, and Cardium ciliatum. The foraminifera are *Quingueloculina arctica, Glandulina laevigata, Cibicides rotundatus, C.r. brononion obscurus, Globigerina bullo des, G. conglomerata, C. involuta, Protelphidium orbiculare, Elphidium granatam, E. obesum, E. subslavatum, Pursens na concava, Plano-cassidulina norcrossi, P. teretis, and Cassilamellina islandica.* Diatoms are fresh and brackish water taxa, both planktonic and benthic. The dominant forms are cold arcto-boreal with few boreal taxa (Aleshinskaya 1964).* Figure 65 shows diatom changes in Siberia during the Pleistocene; in the diagram we can see the correlation of the maximal cooling with transgression of the Arctic basin. This idea was expressed first by G. Lazukov in 1961, and it is shown in Figures 66 and 67 (latest variation). The pollen diagrams (Figures 68, 69, and 70) also show the cooling that took place during the maximal transgression in Siberia. We can see that

* "Severny Ledovity ocean I Yego Poberezie V Kaynozoye." Moskva, Nauka, 1970.

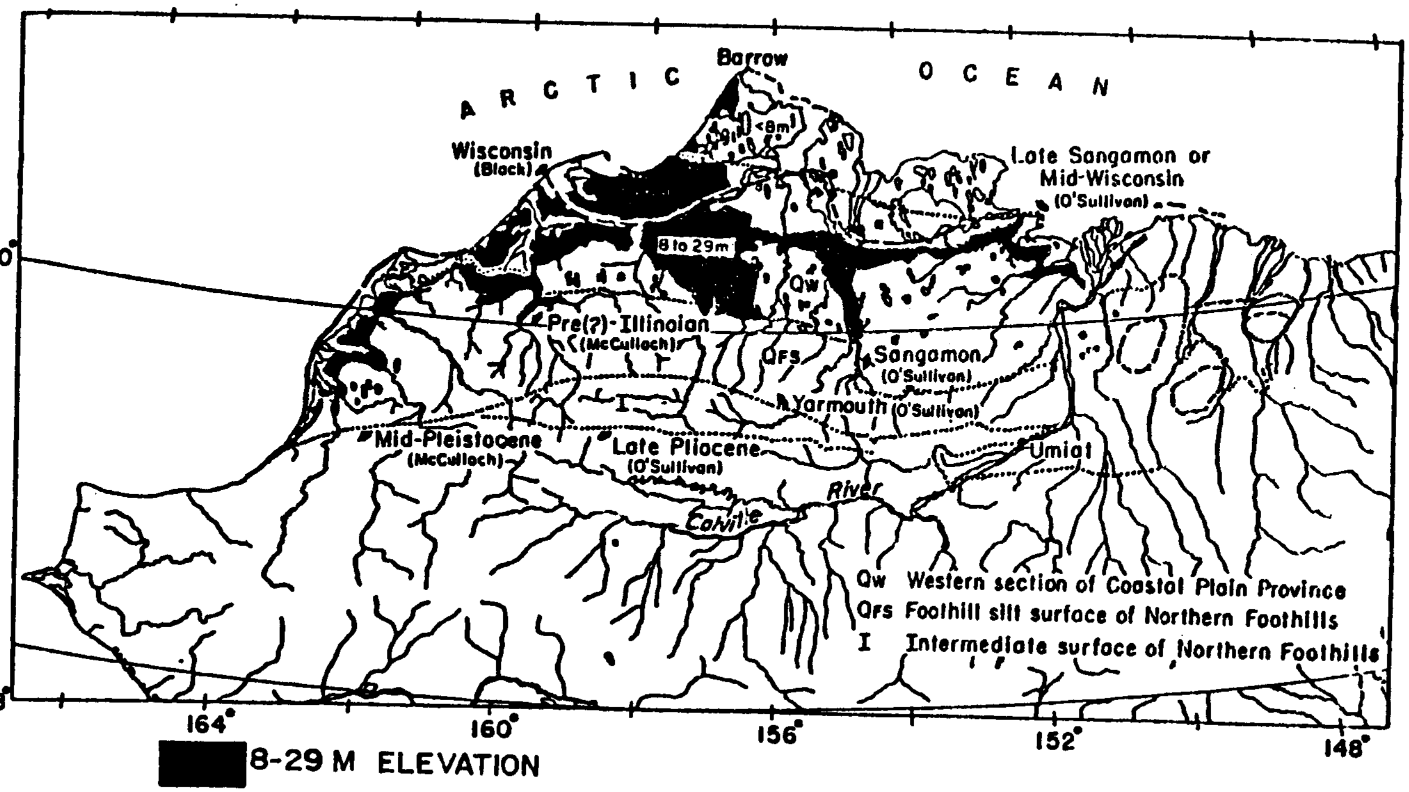
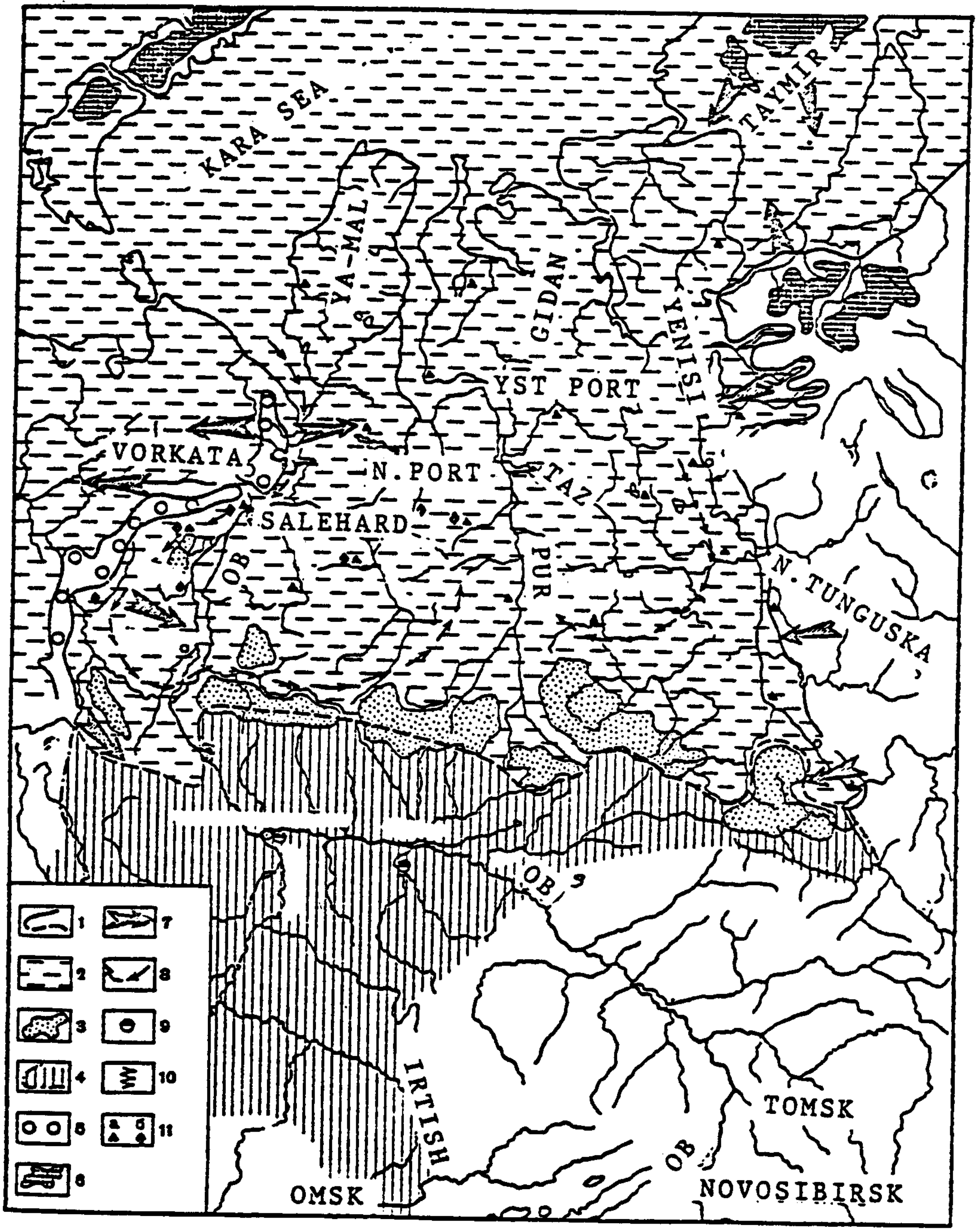


Figure 55.—Map of northern Alaska summarizing the major transgressions and their extent on the Arctic Coastal Plain. After Sellman and Brown (1973).

Table 18.—Summary of Quaternary transgressions for Alaskan coastal regions (Sellman and Brown 1973).

Late Pliocene Early Pleistocene							
				Years			
2,200,000	1,900,000 700,000	300,000	175,000 170,000	100,000	50,000 25,000	10,000 5,000	Investigators
		Einahnuhtan	Probably about 20 m				
Beringian. Two levels higher than present but lower than Anvilian.	Anvilian. Probably much higher than Kotzebuan and Einahnuhtan < about 100 m and > about 20m.		Kotzebuan. Probably about 20 m. Shorelines at about 33 m on western coastal plain.	Pelukian. Two highs at about 7–10 m.	Mid-Wisconsin. Probably a few meters below present.	Krusensternian. Within 2 m of present sea level for deposits <4,000 years old.	Hopkins
Marine transgression. Sediments on wavecut bedrock platform at Kivalina.	Mid-Pleistocene. Transgression-uplift south central part of coastal plain elevating marine sediments at least 100 m.		Pre-Illinoian. Transgressive beach deposit.	Sangamon Marine sediment on wavecut terrace along coast.	Mid-Wisconsin. Marine sediment and ice rafted boulders—deposits raised at least 8 m by later uplift.		McCulloch
Marine transgression. 95–160 m escarpment at thet 320 m elevation at southern margin of intermediate surface.			Yarmouth. Escarpment north of intermediate surface correlated with 95–m terrace near Umiut.	Sangamon. Inner margin of coastal plain causing alluviation of major drainages.	Mid-Wisconsin. Midway on coastal plain.		O'Sullivan
			Illinoian. Skull Cliff Unit.	Sangamon. Meade River Unit.	Wisconsin. Barrow Unit		Black
				Pelukian. Possible transgression indicated by more silty sediment under Mid-Wisconsin.	Mid-Wisconsin. Marine sediment dated near Barrow suggests extensive transgression on portions of Coastal Plain, to present level and possible depression of inland ridge.	Krusensternian. Small fluctuations in sea level in last 2,000 years forming and modifying present Barrow spit.	Sellman, Brown and others.



1. Boundaries of the Yamal maximal cold transgression (Sanchugovo-Salehard Samara-Dniper)
2. Depth greater than 50 m
3. Islands and shallow water area
4. Lacustrine-alluvial deposits
5. Areas of glaciation
6. Areas of glaciation
7. Direction of coarse material distribution
8. Current direction
9. Natural phenomena connected with glaciation
10. Natural phenomena connected with glaciation
11. Fauna and microfauna in the Yamal deposits

Figure 56.—Western Siberia during the Yamal transgression. After Kuzin (1961).

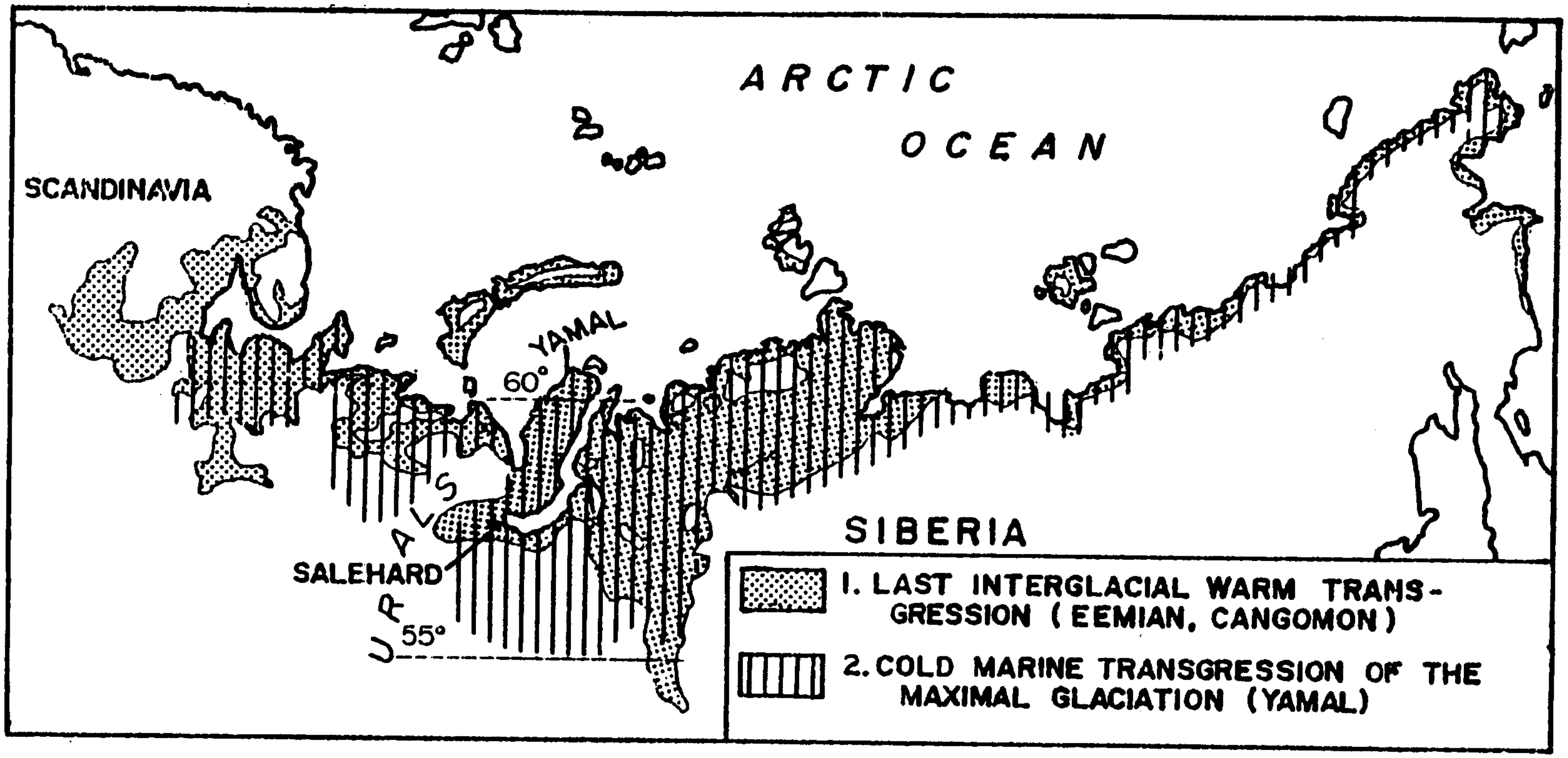


Figure 57.—Spreading of the two transgression deposits: last interglacial warm transgression (Eemian, Sangamon) and cold marine transgression of the maximal glaciation (Yamal, Illinoian) in Northern Eurasia.

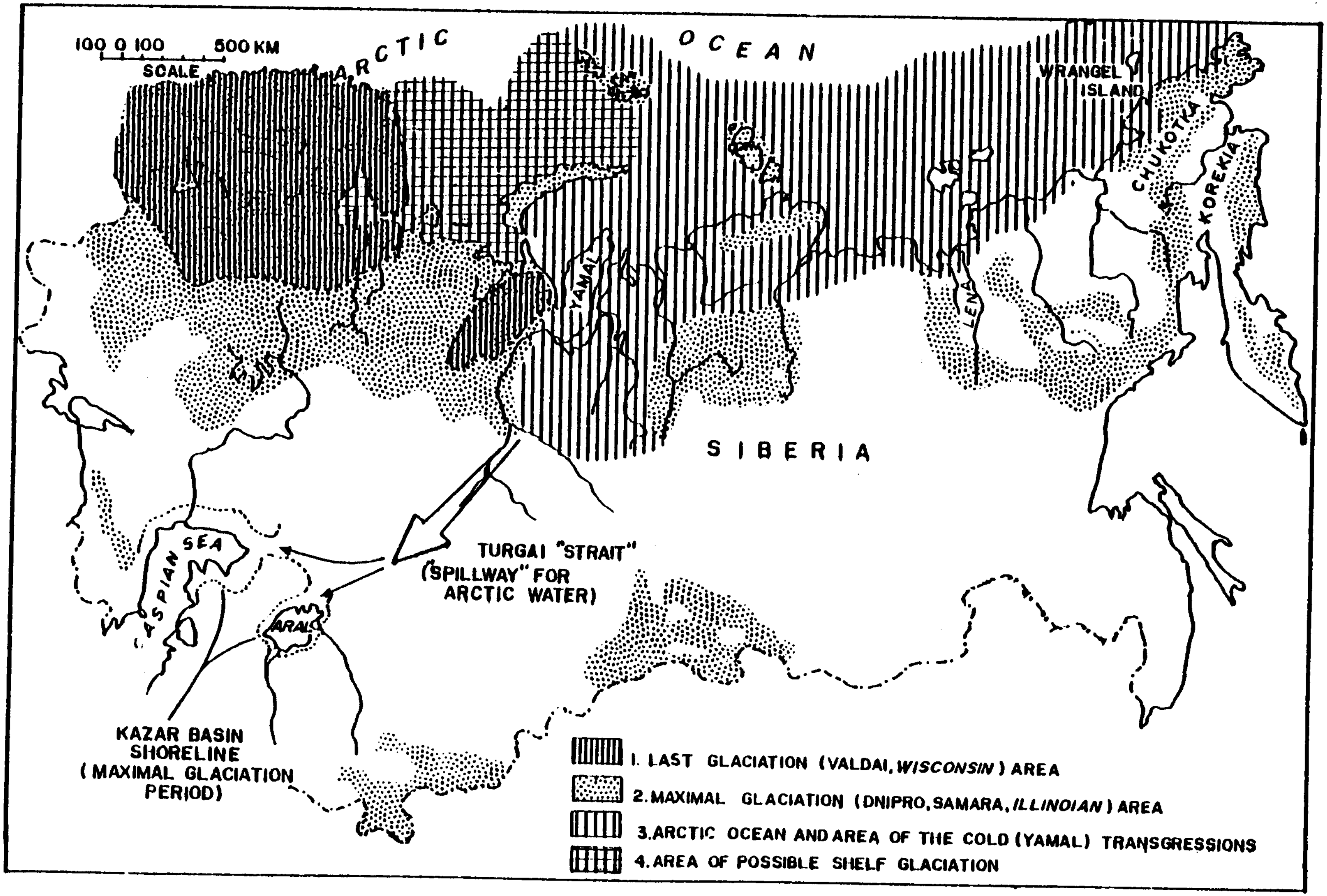


Figure 58.—Boundaries of the Yamal cold transgression and the areas simultaneously covered with glaciation.

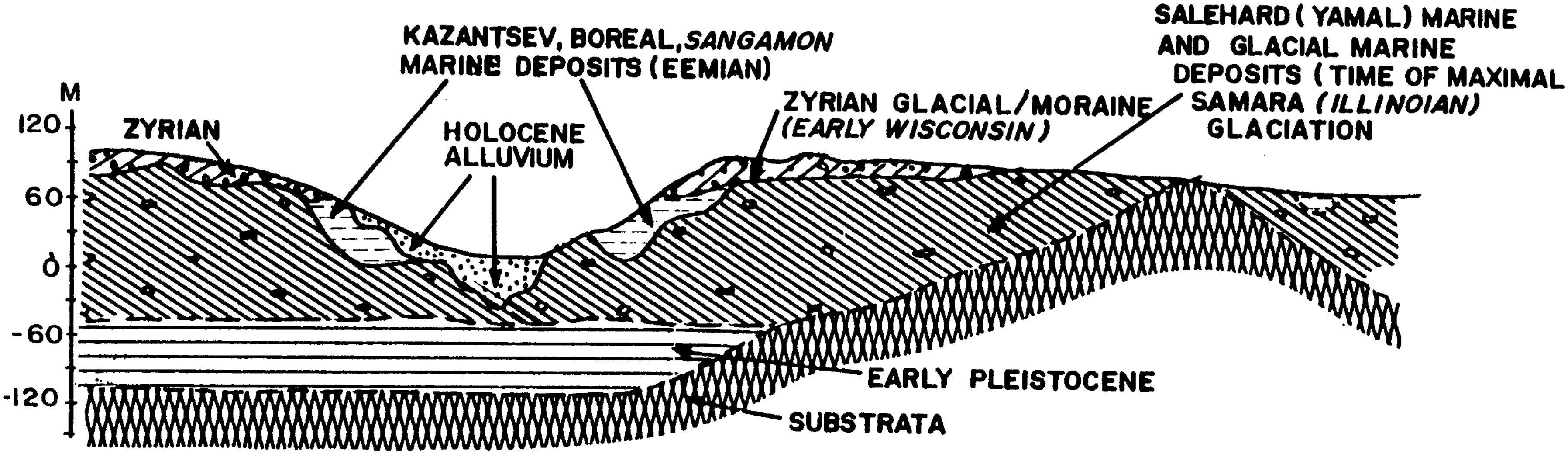


Figure 59.—Cross section of the Quaternary deposits at the Ob River basin about 100–200 km to the south from the Arctic (Lazukov 1965).

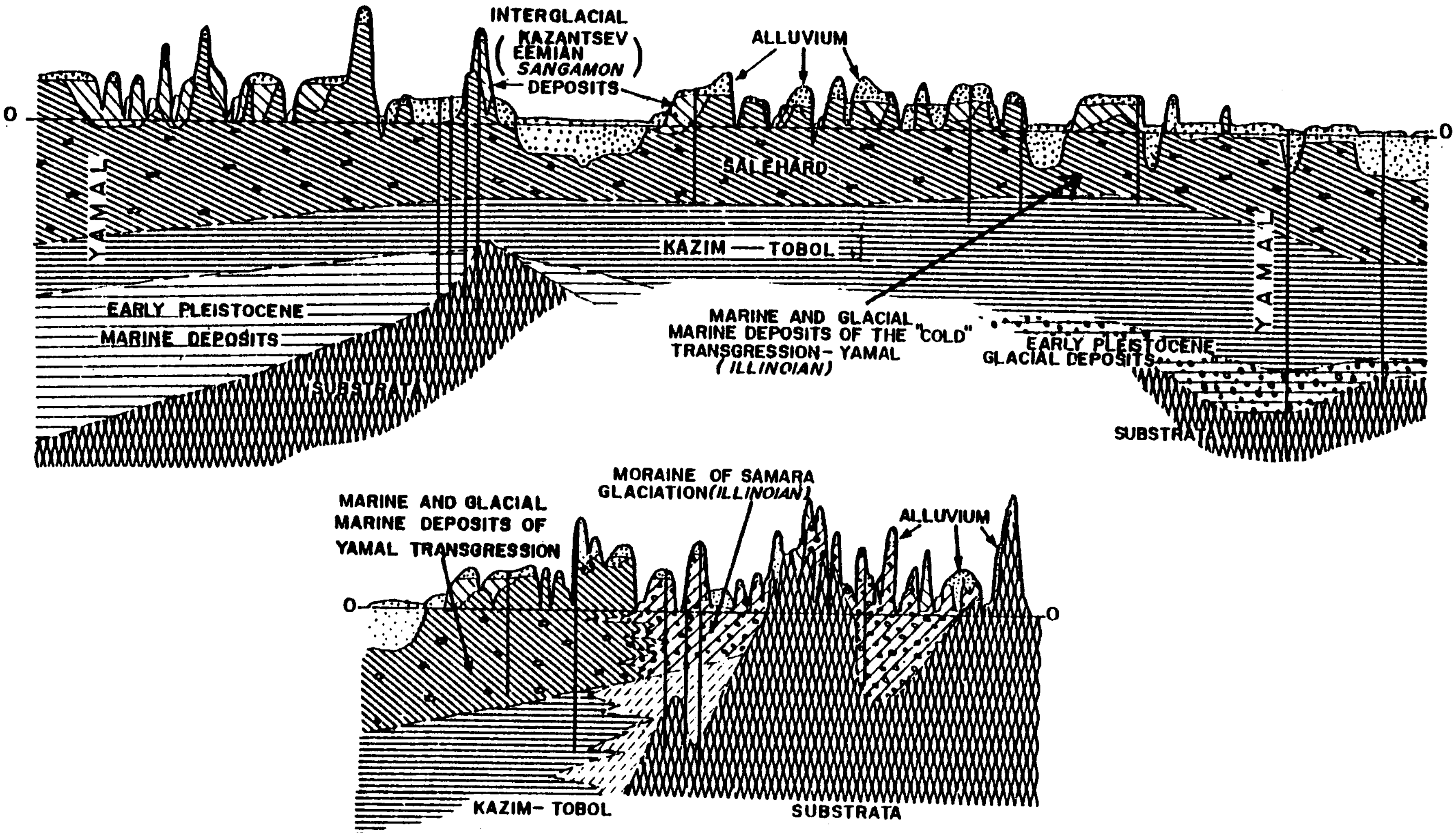


Figure 60.—Cross section of the Quaternary deposits at the mouth of the Irtish River about 800–1,000 km to the south from the Arctic Ocean (Lazukov 1970).

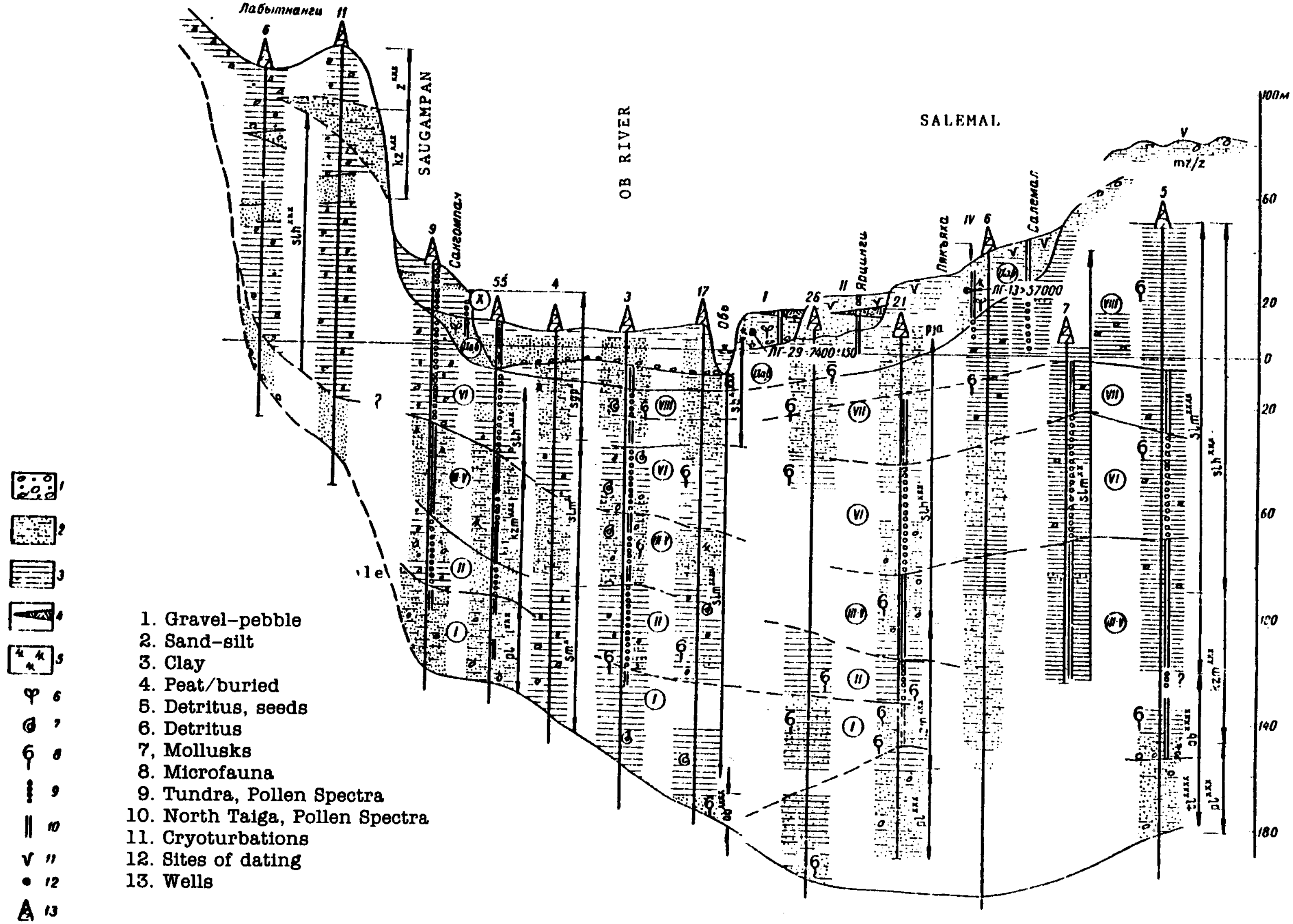


Figure 61.—Scheme of the Pleistocene deposits in the Ob River mouth (Zubakov 1972).

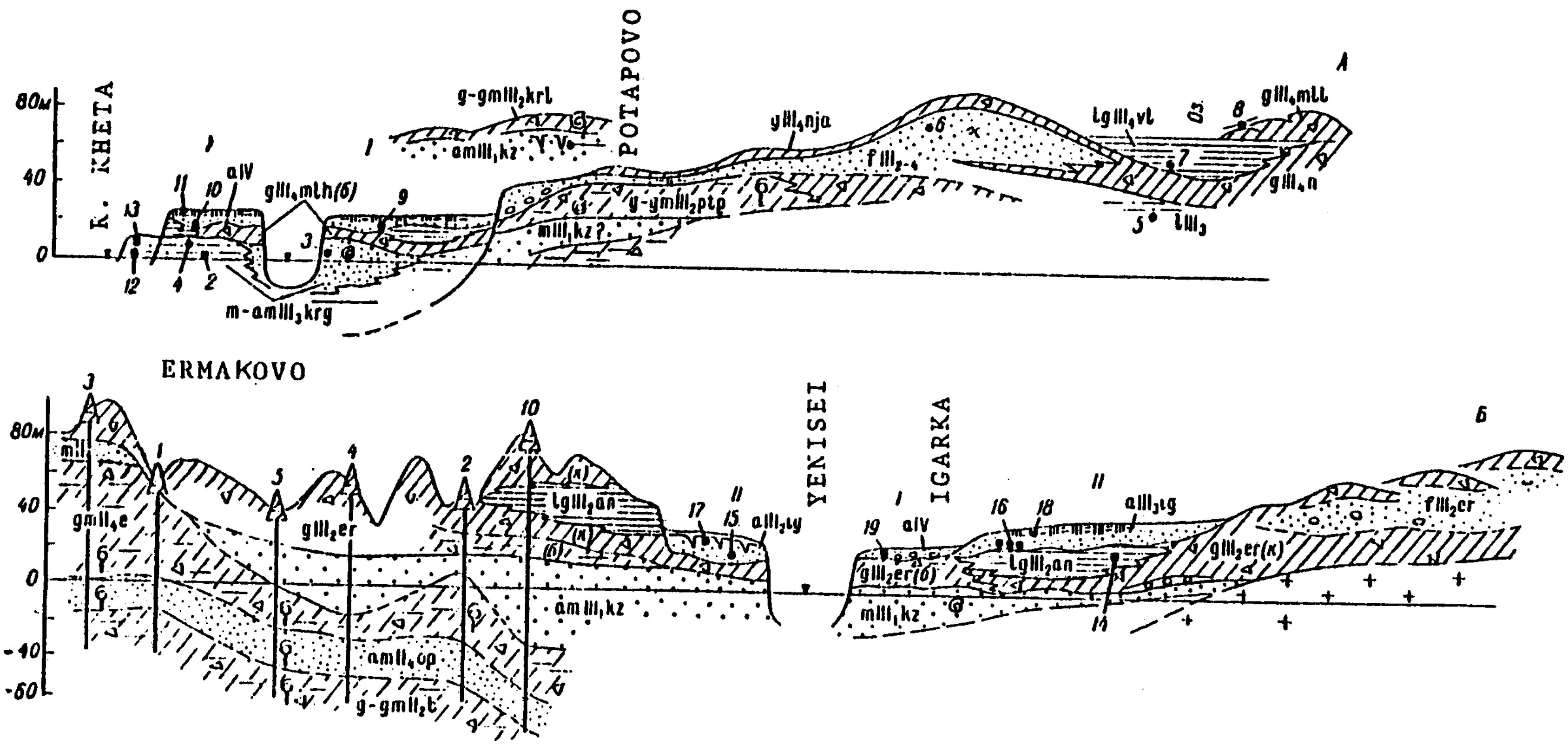


Figure 62.—Pleistocene deposits in the Yenisei River basin (Zubakov 1972).

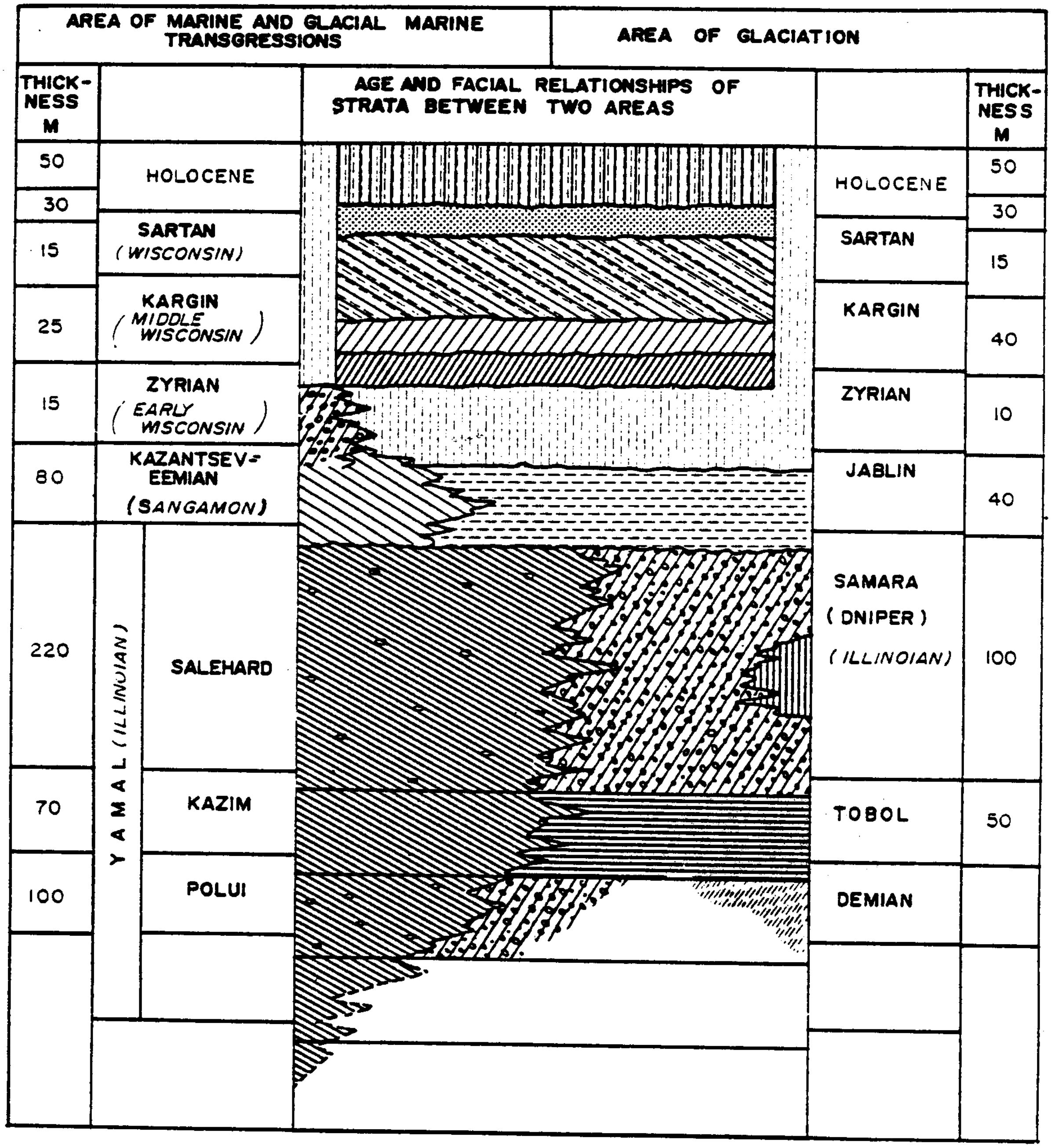


Figure 63.—Relationship between deposits of transgression and glaciation areas in Siberia (Lazukov 1970).

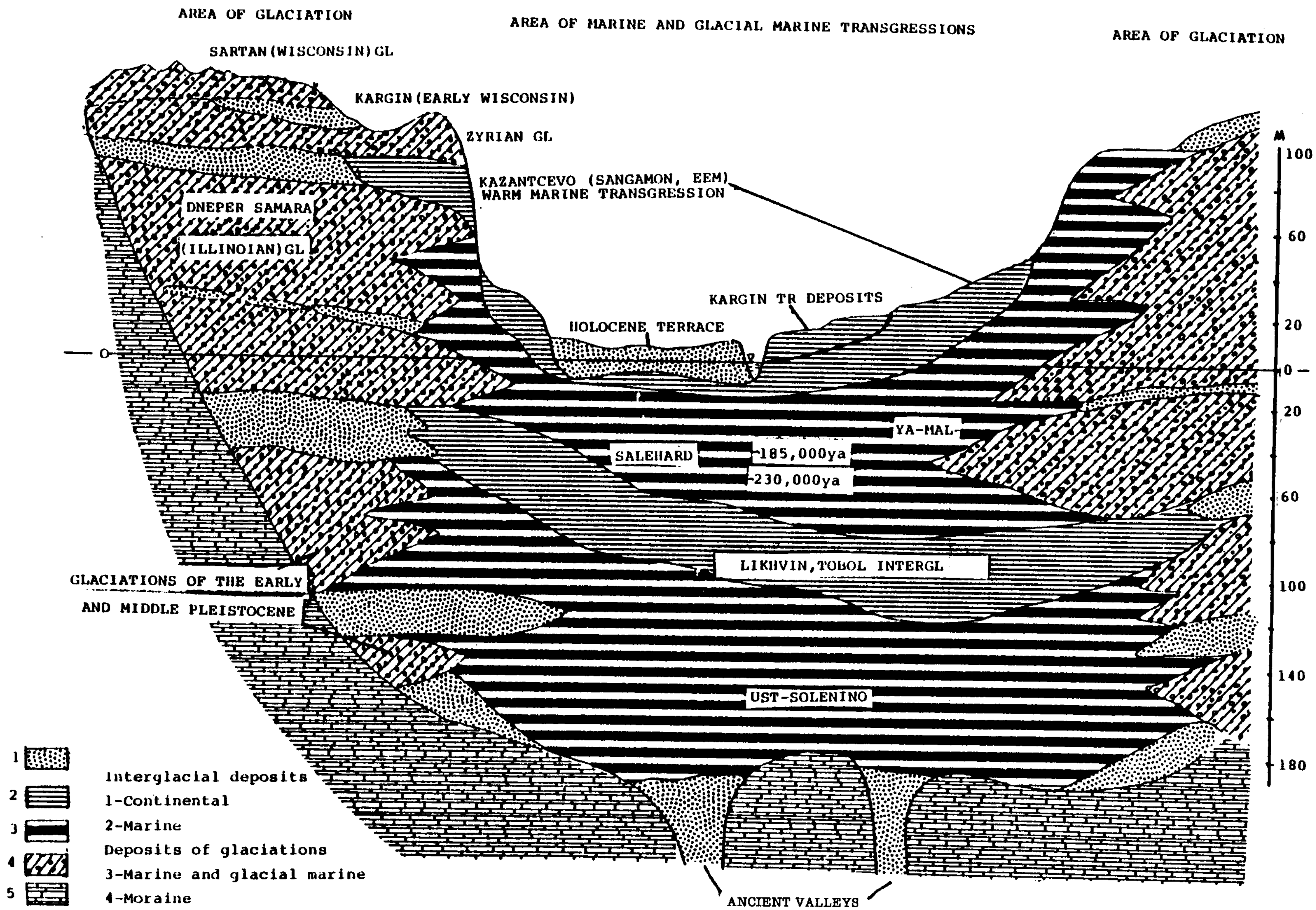


Figure 64.—General scheme of the Pleistocene deposits of the Eurasiatic coast of the Arctic.

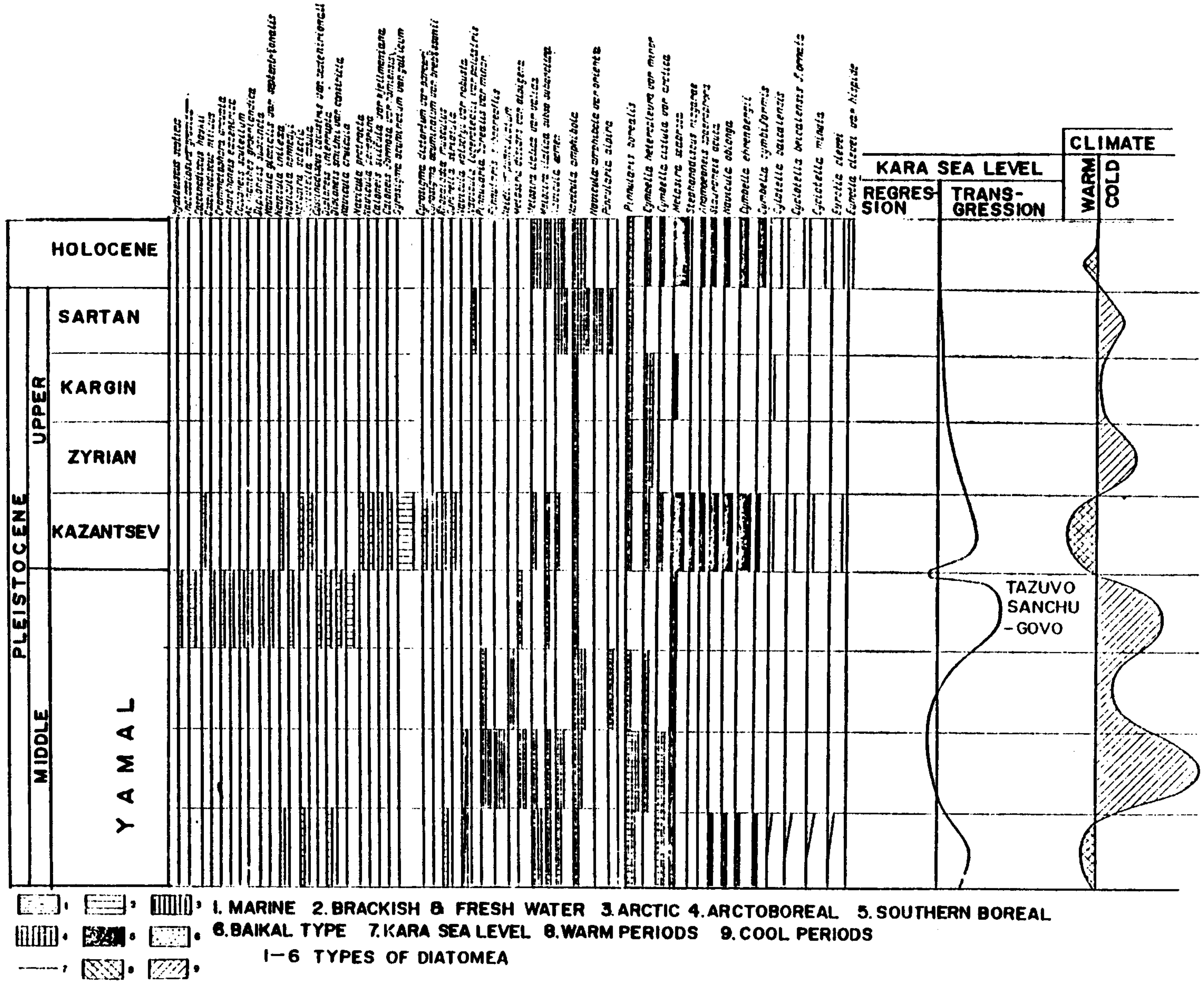


Figure 65.—Diatom diagram of the Quaternary deposits in western Siberia (Aleshinskaya 1964).

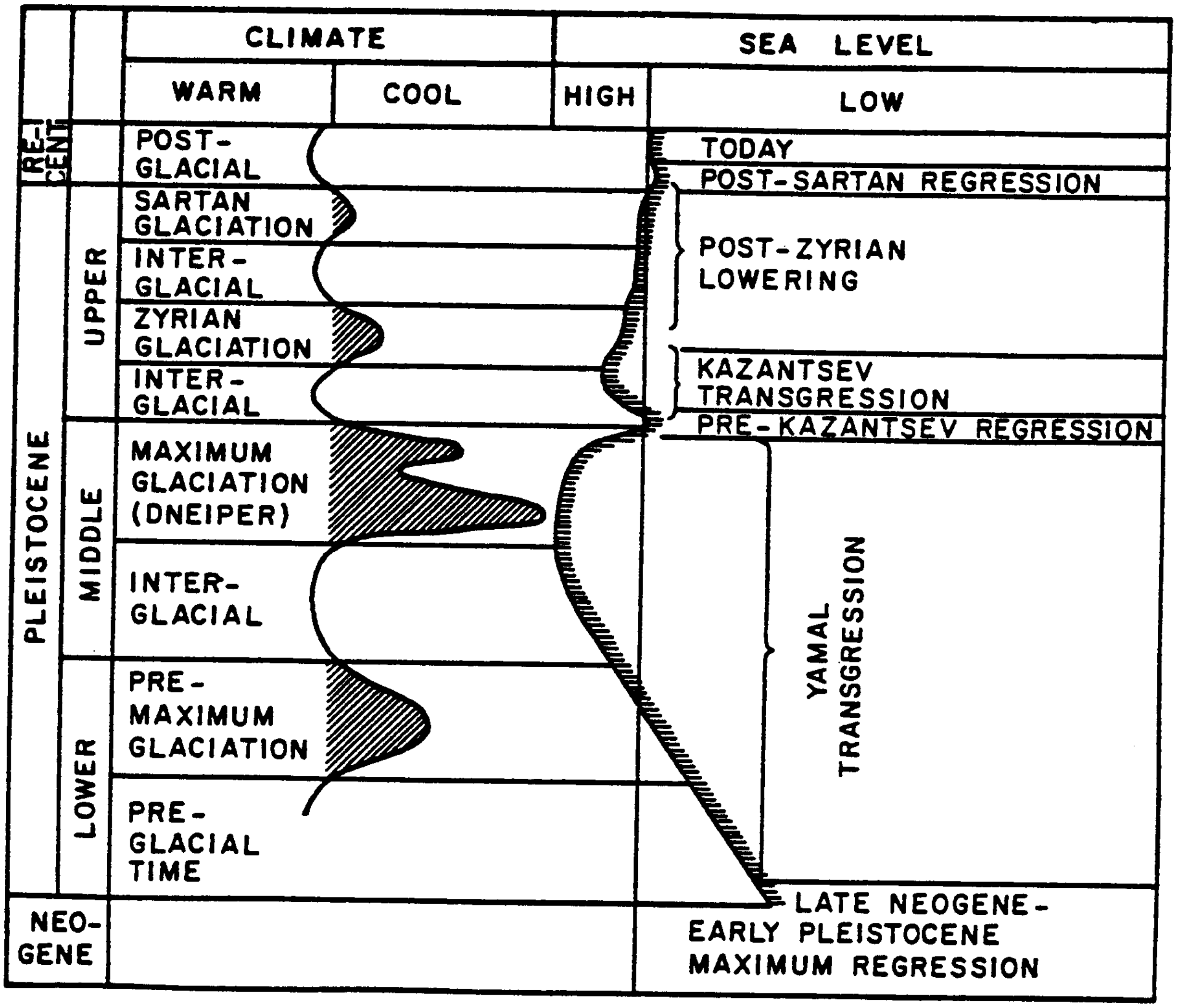


Figure 66.—Climatic changes and ocean level oscillations in western Siberia during the Pleistocene (Lazukov 1961).

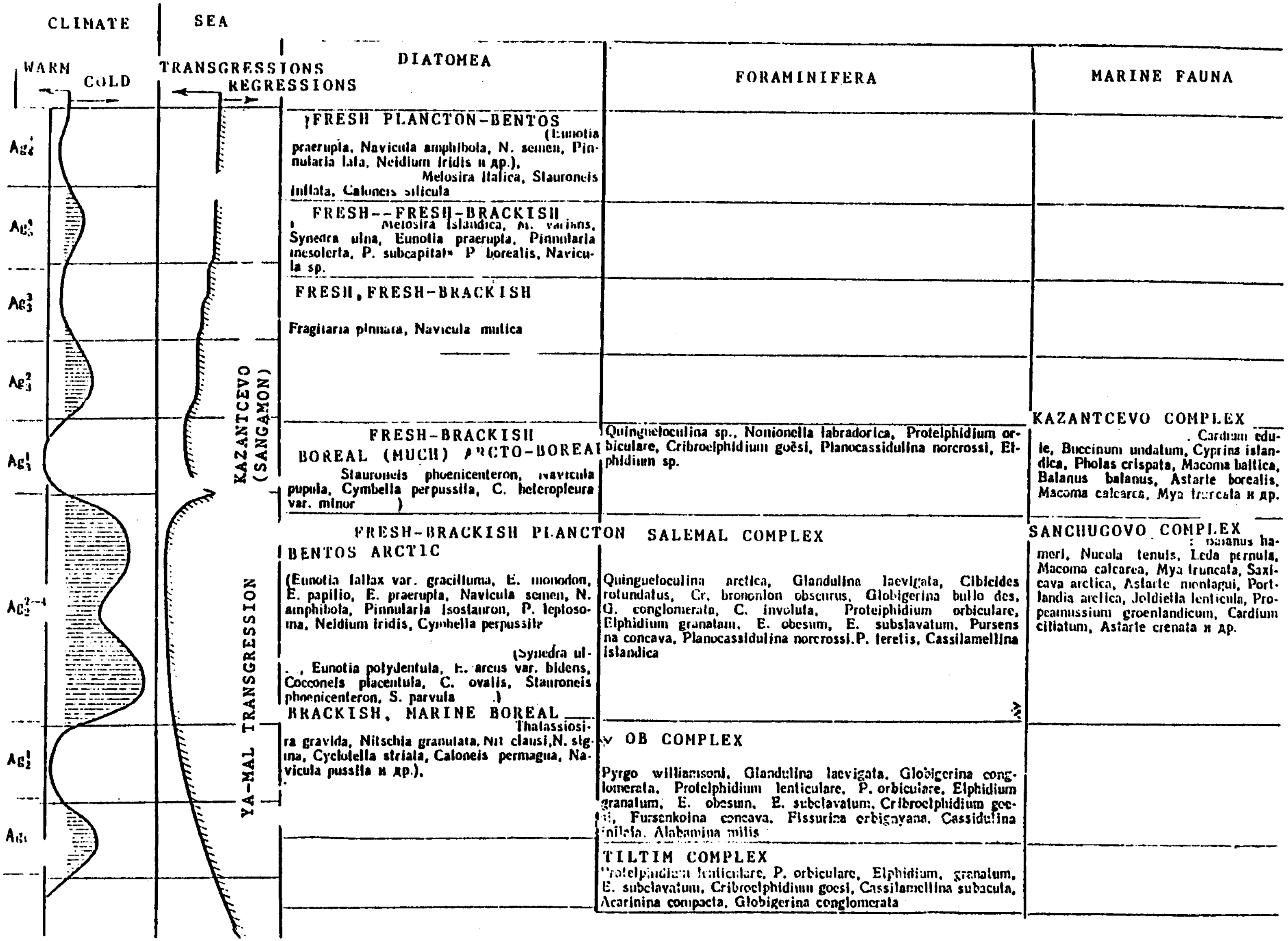
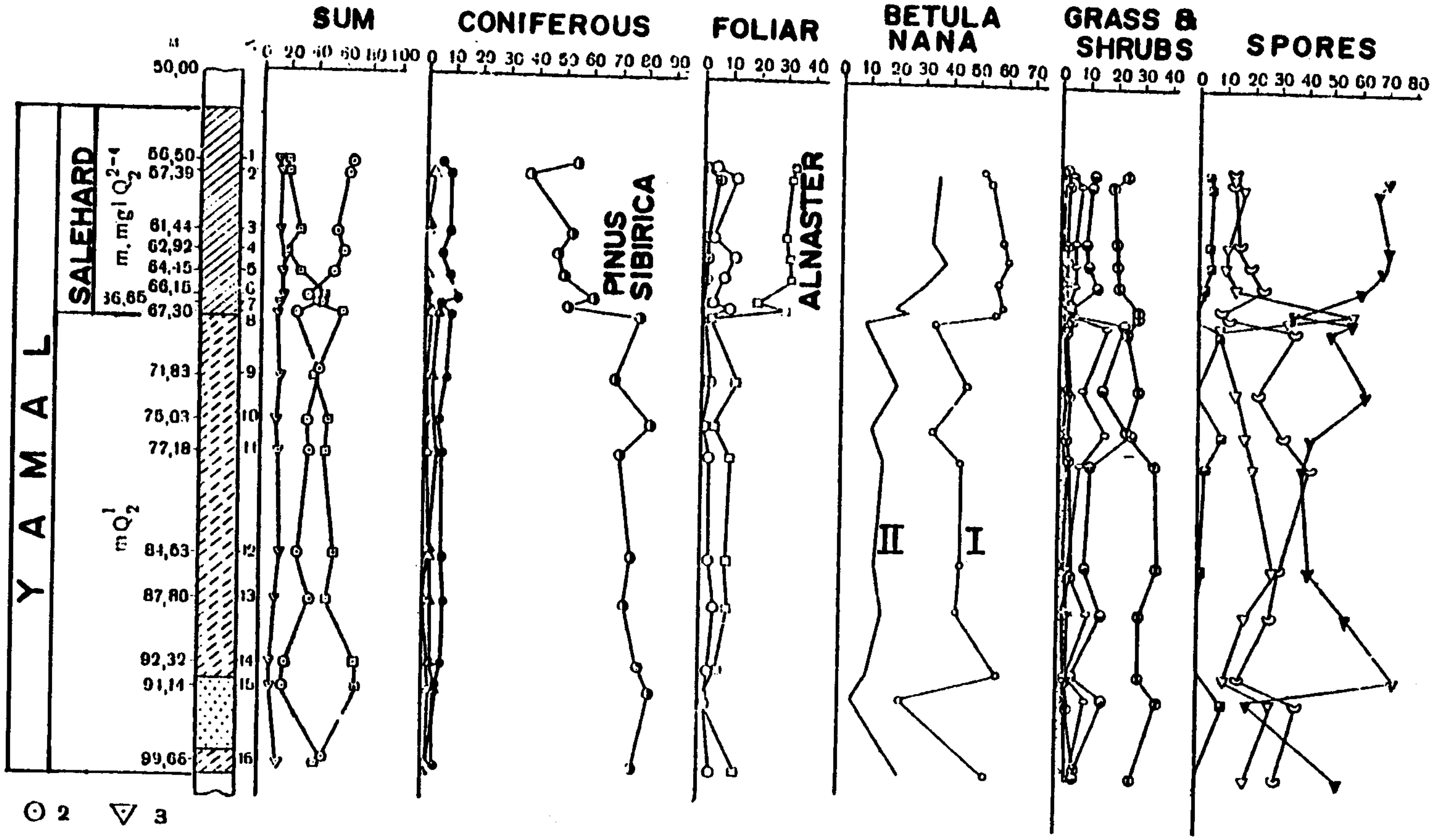
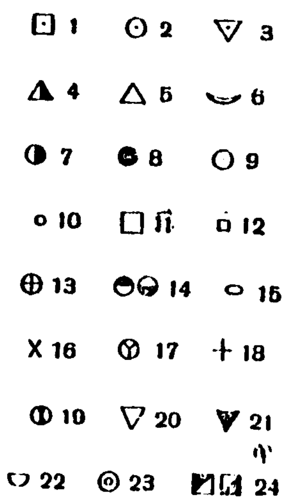


Figure 67.—Ocean level oscillations in Siberia. After Lazukov (1972).



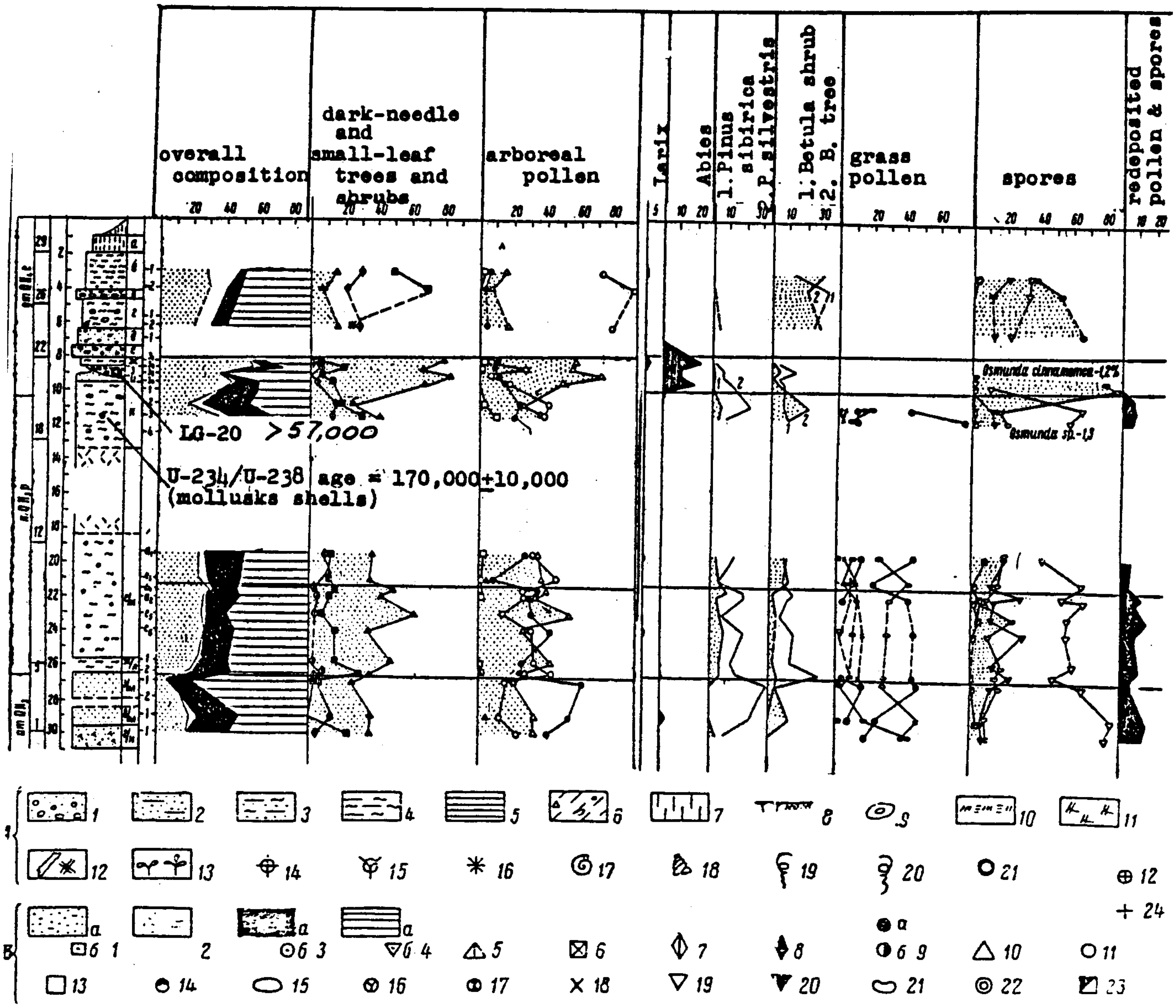


I. % Betula nana from total sum of grass and shrubs.

II. % Betula nana from total sum of pollen grains.

1. AP
2. NAP
3. Spores
4. Abies
5. Picea
6. Larix
7. Pinus sibirica
8. Pinus silvestris
9. Betula sect. Albae
10. Betula sect. Nanae
11. Alnus
12. Alnaster
13. Salix
14. Gramineae
15. Cyperaceae
16. Chenopediaceae
17. Ericales
18. Artemisia
19. Varia
20. Bryales
21. Sphagnum
22. Filicales
23. Equisetum
24. Lycopodium

Figure 68.—Pollen diagram of the marine deposits of the maximal cold transgression in Siberia near Salehard (Lazukov 1970).



A–Lithology
1. Sand with pebble & gravel
2. Sand
3. Silt
4. Loam
5. Varied clay
6. Facies similar to the till
7. Sandy and clayey loam
8. Soils & gyttja
9. Concretions
10. Peat
11. Plant detritus
12. Mammal bones
13. Carpoides
14. Marine diatoms
15. Freshwater diatoms
16. Diatomic water-plants
17. Marine mollusks
18. Freshwater mollusks
19. Foraminiferas
20. Ostracods
21. Radiocarbon samples

B–Spore-Pollen Data (1–4. Overall composition)
1. Trees
2. Shrubs
3. Grasses & shrubs
4. Spores
(5–13 Arboreal pollen)
5. Dark-coniferous in total
6. Total small-leaf trees
7. Shrubs
8. Broadleaf trees
9. Pinus Dyploxylon (a) and Pinus Haploxylon (b)
10. Spruce
11. Total birch
12. Willow
13. Alder
(14–18 Grass pollen)
14. Cereals
15. Sedges
16. Heathers
17. Various grasses
(19–23 Spores)
19. Green mosses
20. Sphagnum mosses
21. Ferns
22. Lycopodia
24. Single pollen and spores

Figure 69.—Pollen diagram of the deposits. After Zubakov (1972).

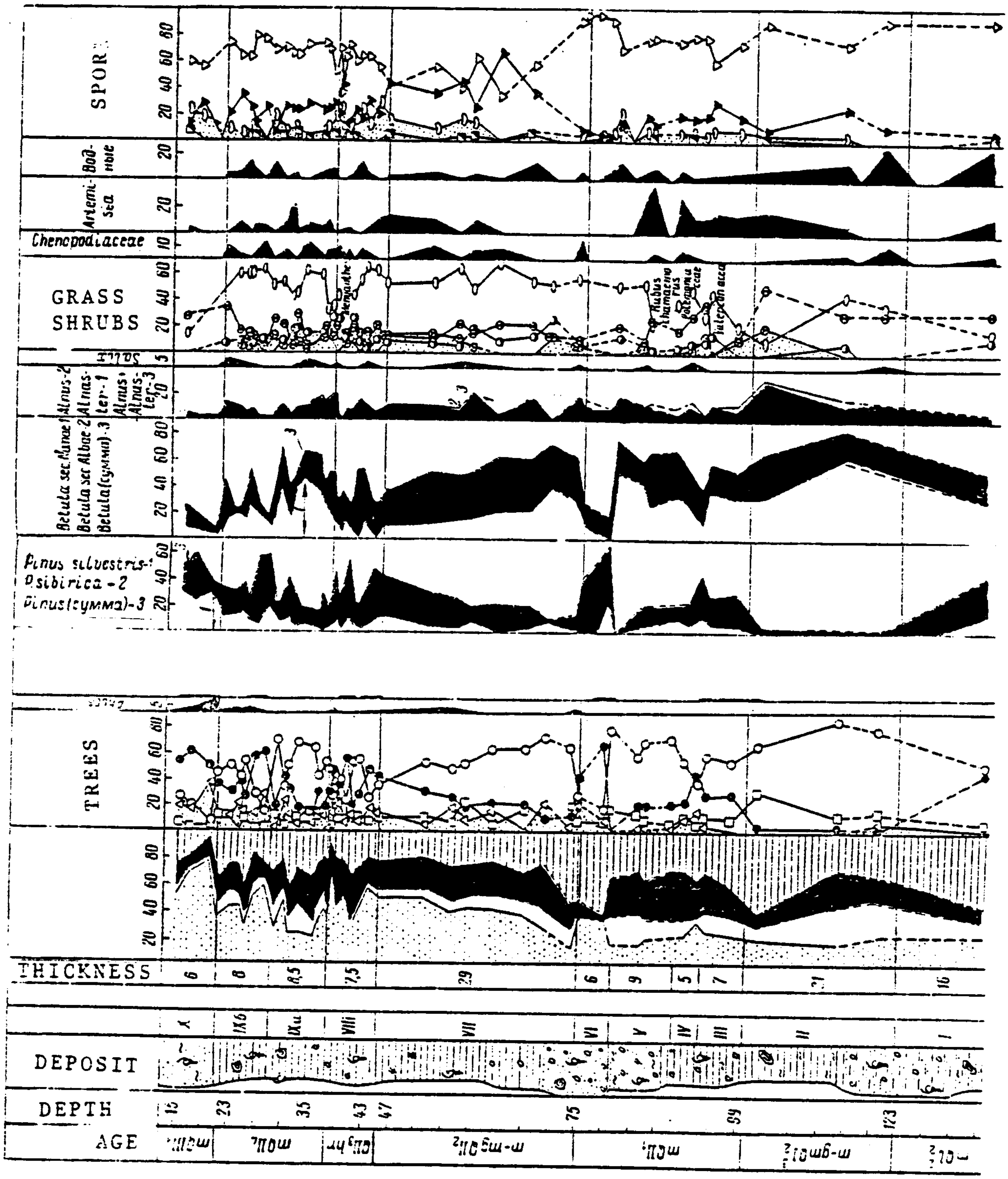


Figure 70.—Pollen diagram of the deposits. After Zubakov (1972).

the sums of trees, shrubs, and grass pollen are equal. *Betula nana, Alnaster,* and *Pinus siberica* show that arctic desert, periglacial tundra and forest-tundra, and partly northern taiga (lichen woodland) were the most characteristics types of vegetation along the coastal zone that was not covered by glaciation.

Figures 71–76 show the division of the Siberian marine Pleistocene according to Gudina and the different foraminifera complexes for the deposits of interglacial and glacial epochs. We can see the big difference between Kazantcev warm or Toboe (Turnhan) and Salemal–Salehard–Sanchugovo cold complexes of foraminiferas.

Figures 77–78 show the meridional correlation of the Pleistocene deposits in the Ob and Yenisei river basins after Zubakov (1972). We may see the extension of the cold marine transgression deposits far to the south all over western Siberia.

For the Chukotka Peninsula we have similar results in the Kiestov transgression. The maximal cold transgression there took place during the time of the maximal glaciation according to the data gathered by Petrov (1965) (Figures 79–82, Table 22) and Khoreva (1974) (Figures 83 and 84, Table 23). Position and altitudes of the marine terraces of the Arctic basin in Siberia and Chukotka and their age are very similar (Figures 85 and 86) and typical for the coastal areas in Eurasia. The Yamal and Krestov terraces are the result of cold marine transgression (Lazukov 1970; Petrov 1965) and they are higher than interglacial Boreal marine terraces of this very large region. We see that the data from the Eurasiatic part of the Arctic basin, including geological and geomorphological evidence and paleontological and palynological materials, provide the basis for the conclusion that the time of the maximal cold transgression of the Arctic basin took place during the maximal glaciation (Illinoian, Dnipro, Samara). The absolute age dating (Figure 87) also supports this conclusion (Zubakov et al. 1974; Svitoch et al.* 1976).

Problems of Arctic Shelf Glaciation

Explanations of cold marine transgression phenomena in the Arctic from evidence of Arctic Ocean isolation during glaciation differ. Some scientists (Degtyarenko et al. 1971) try to explain it as a result of tectonic movements (Figure 88) acting more intensively along the coastal areas close to the Ural Mountains, dividing Europe and Asia (northeast Europe and western Siberia). The weakness of this point of view is that new facts about the marine terraces of the Chukotka Peninsula contradict this idea. Other scientists (Archipov 1971; Troitsky 1975) think that most

* "Beringia in Cenozoic," Academy of Sciences of the USSR, Vladivostok, 1976.

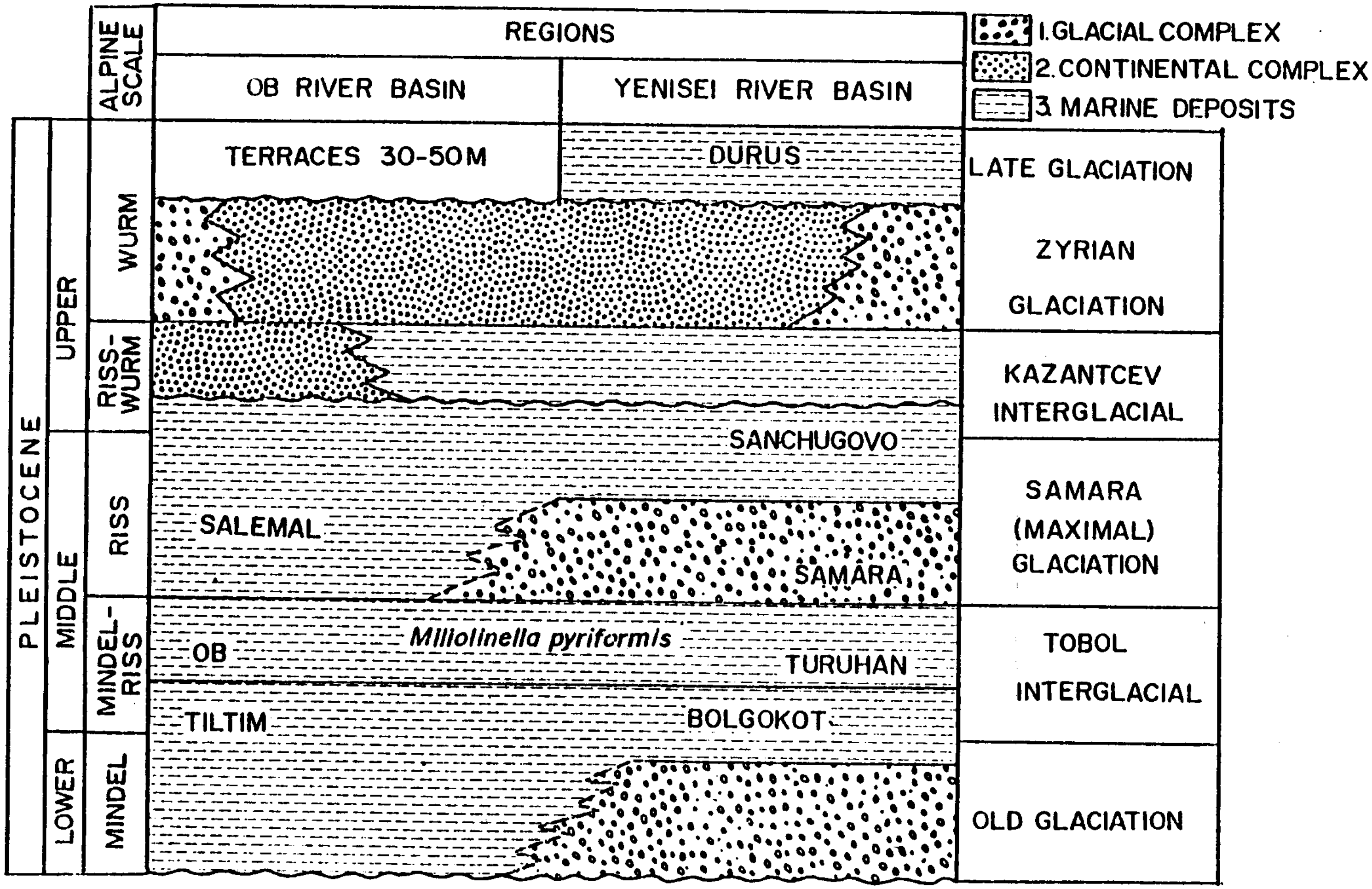


Figure 71.—Pleistocene of Siberia (Gudina 1969).

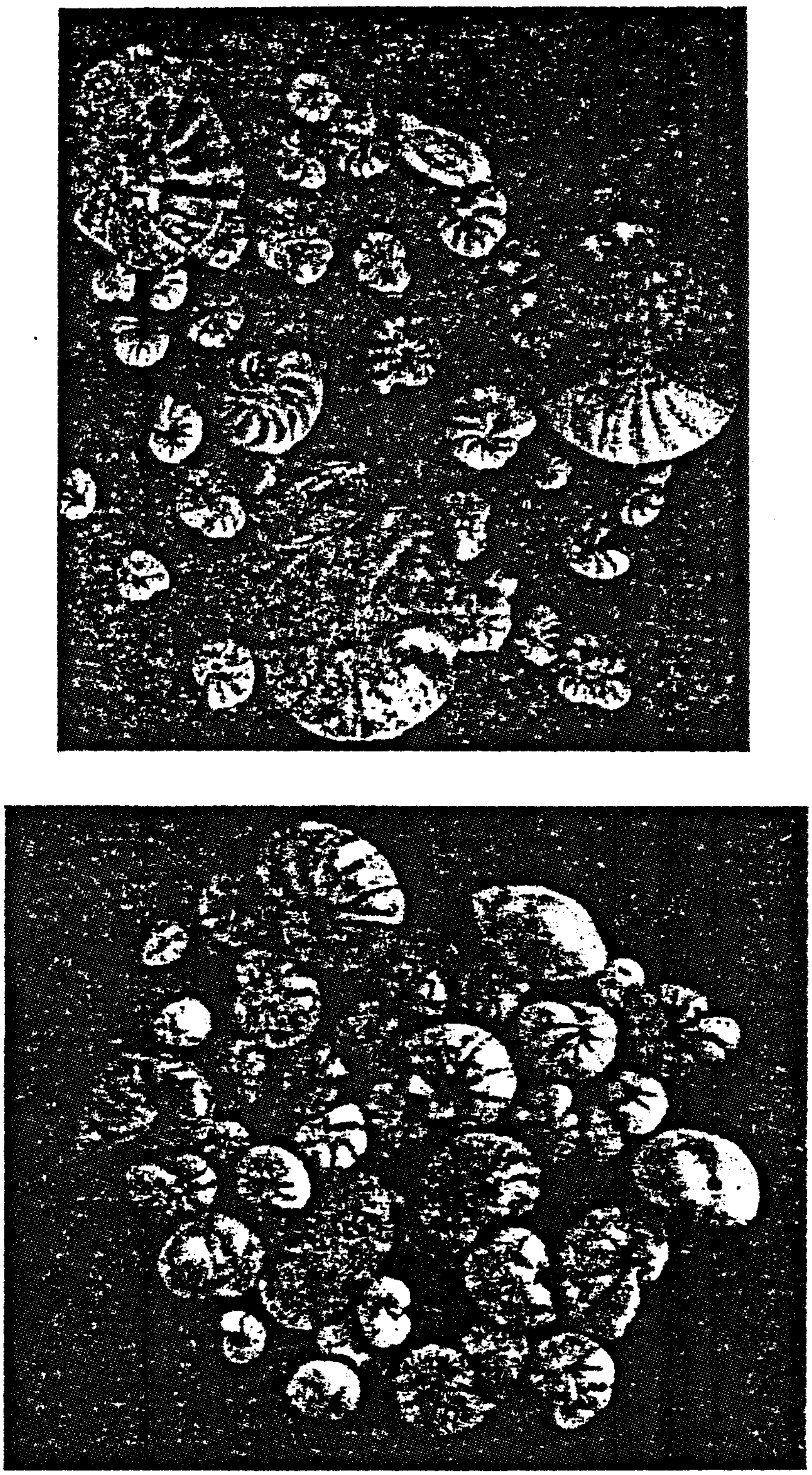


Figure 72.—Bolgohot foraminifera complex. After Gudina (1969).

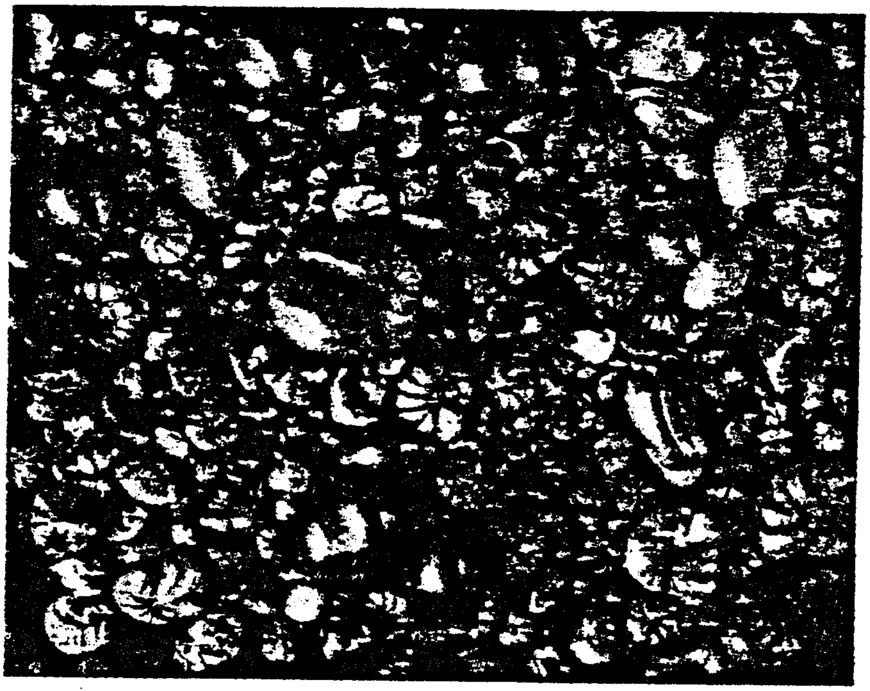


Figure 73.—Turuhan foraminifera complex. After Gudina (1969).

of the marine deposits were formed in the latest periods of glaciation. There is a good reason to agree with this, and in the model of the Arctic Ocean level changes (see below), we will try to show it. The third possible explanation of the cold marine transgression was used by Grosswald et al. (1973). These geologists proposed a great extension of the glaciation in the Barents Sea and to the east from the Urals (Figure 89), and we agree with the possibility of glaciation partly on the Arctic shelf and the development of such phenomena as shelf glaciation. The isolation of the Arctic Ocean from the warm water of the Atlantic created the possibility for shelf glaciation, but it is different to suppose that all the shelf was covered by an ice sheet for several reasons:

(1) Some of the islands in the Arctic Ocean, especially to the east from the Kara Sea, never have been glaciated at all (Wrangel Island; Yurtsev 1971). This scientist gives a tremendously rich list of the botanical species for this island, in contrast to the poor complexes of the continental area of Siberia (Wrangel Island as the refugium).

(2) The widespread development of glaciation on the shelf could change the marine character of the transgression and make it resemble a freshwater transgression; the data from the marine transgressions are against it.

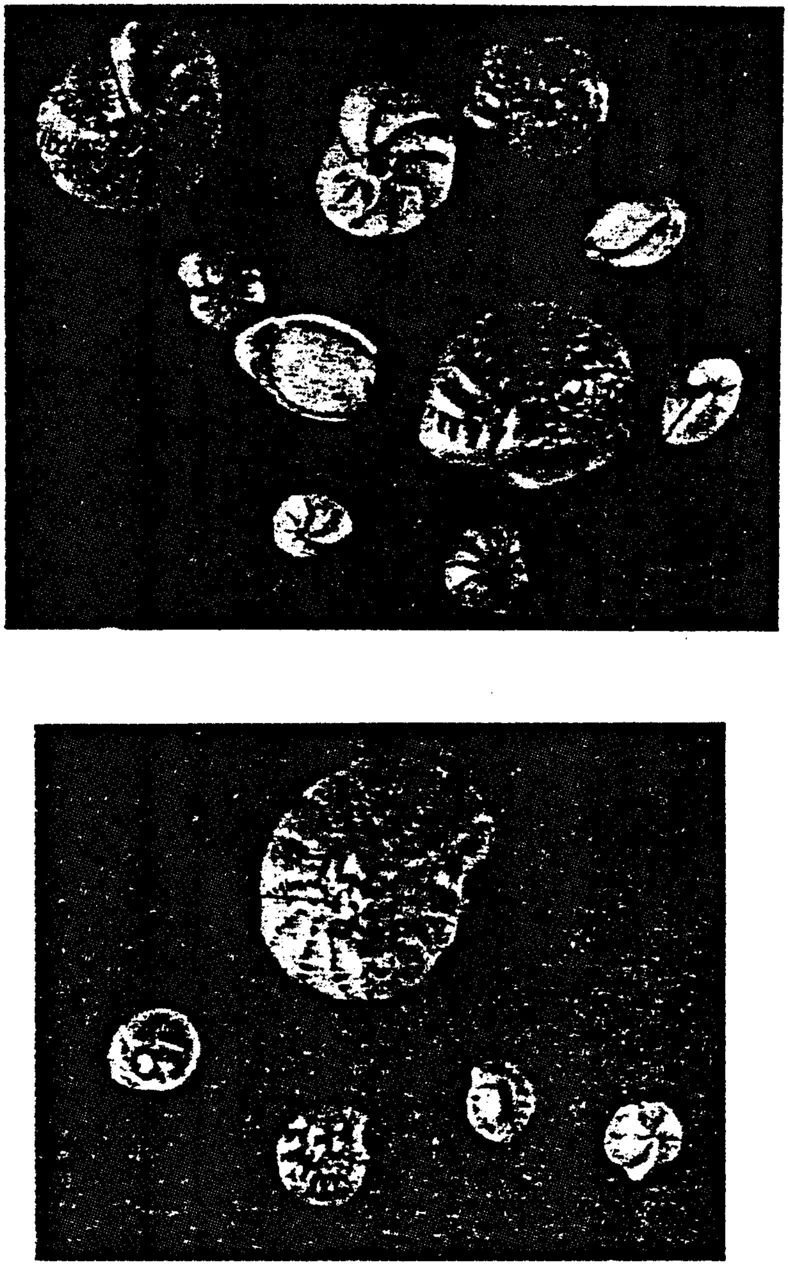


Figure 74.—Sanchugovo foraminifera complex. After Gudina (1969).

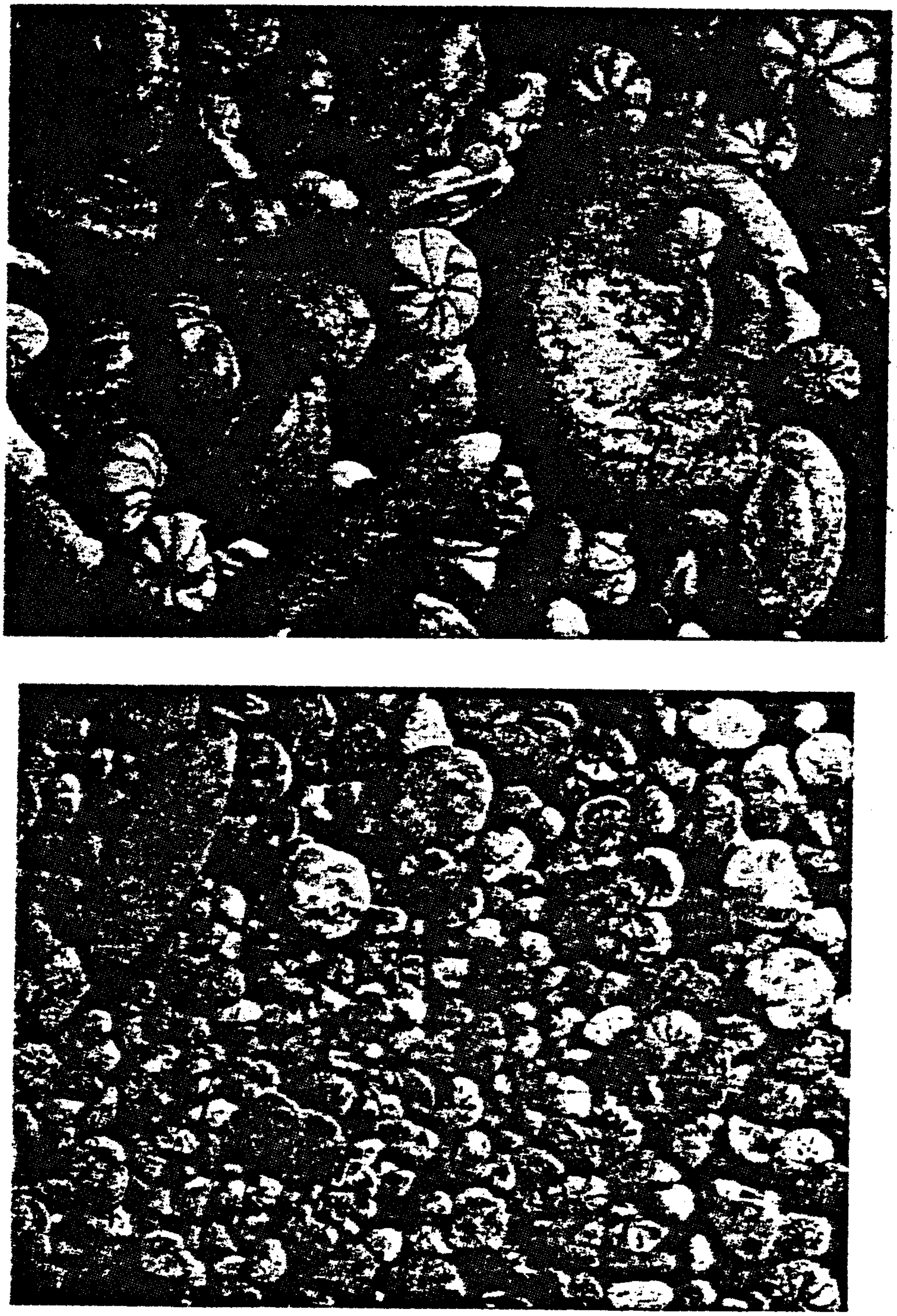


Figure 75.—Kazantcevo foraminifera complex. After Gudina (1969).

Figure 76.—Durus foraminifera complex. After Gudina (1969).

(3) There are no significant anomalies of isostatically uncompensated masses on the Arctic shelf of Eurasia that could be considered a result of glacio-isostasy (Figure 90; Artemiev* 1975).

(4) Latest study of the Barents Sea (the most possible area for such kind of glaciation) shows that the bottom of this sea was never covered by ice (Tore Vorren 1977; Matishov 1974 [Figure 91]; Baranowski 1977).

According to B. Koshechkin, S. Strelkov** (1976), the last Scandinavian glacial cover occupied Scandinavia, the Kola Peninsula, and the White Sea depression and spread far to the south and east. The continental ices descended toward the Norwegian Sea forming a narrow

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* M. W. Artemiev, "Isostasy of the USSR." Nauka, Moscow, 1975.

** Geomorphology and paleogeography materials of the Geography Congress, Moscow, 1976.

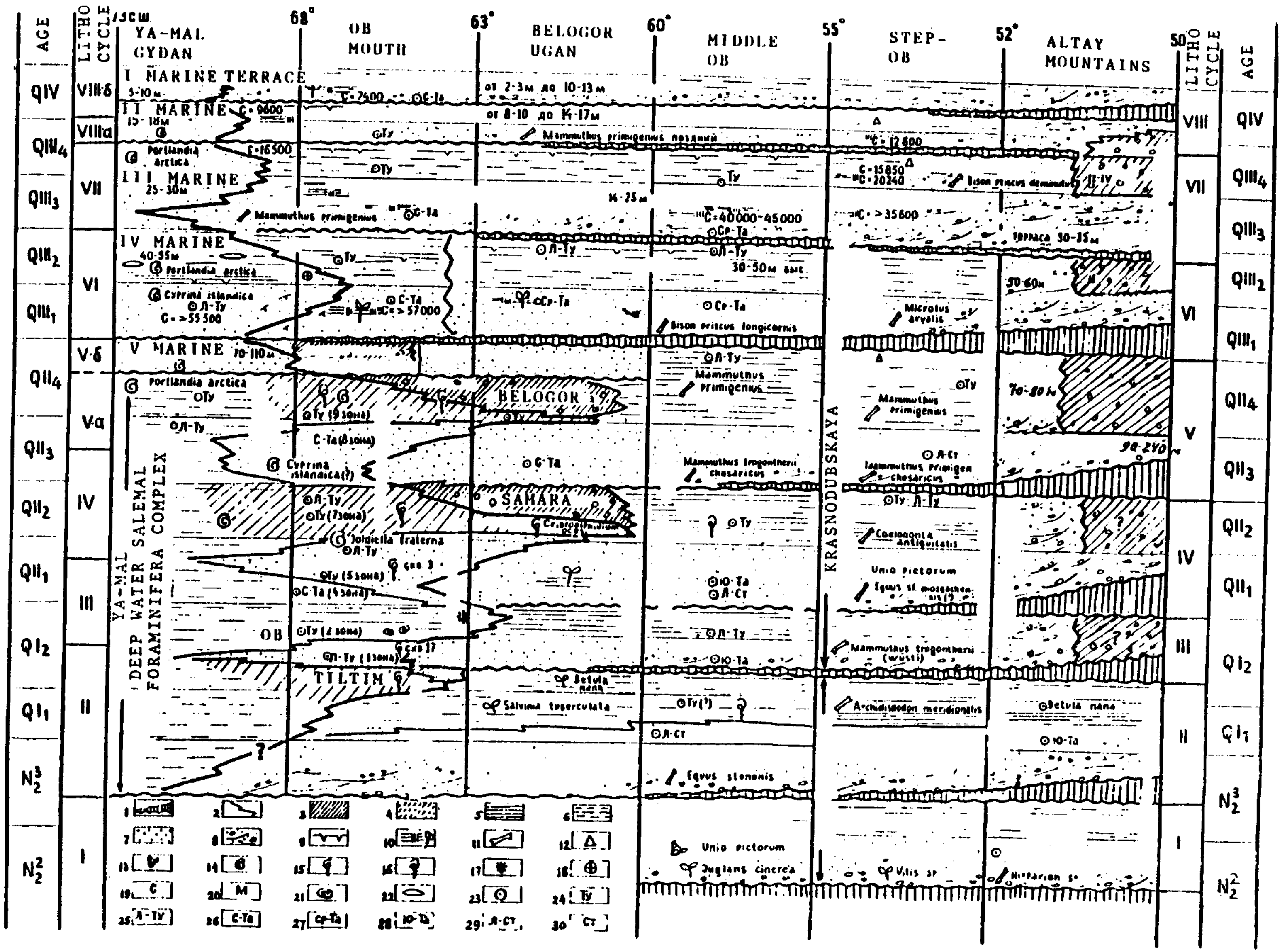


Figure 77.—Meridional correlation of the Pleistocene deposits in the Ob River basin. After Zubakov (1972).

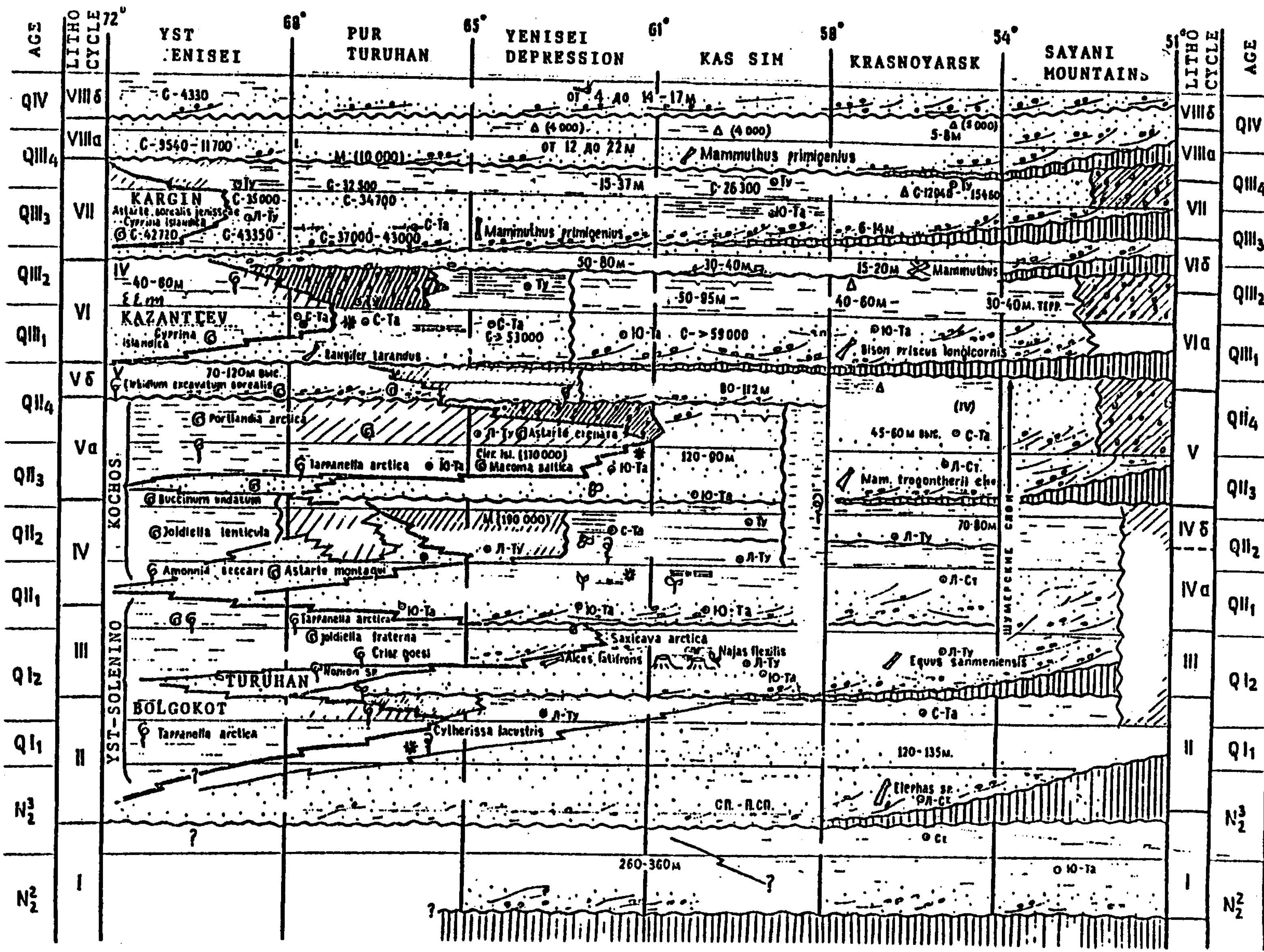
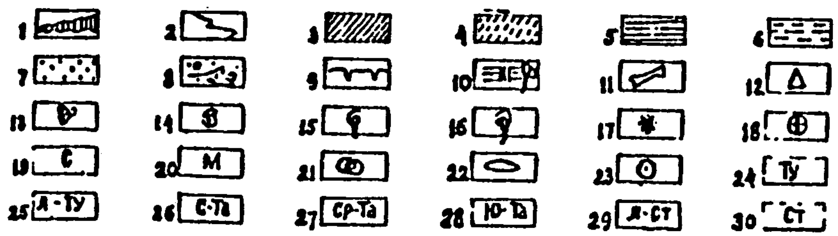


Figure 78.—Meridional correlation of the Pleistocene deposits in the Yenisei River basin. After Zubakov (1972).



1. Erosional break
2. Boundary of the marine and continental phases
3. Continental-glacial formation
4. Iceberg glacio-marine deposits
5. Varied clay
6. Clay aleurite
7. Sand
8. Pebble
9. Cryoturbations
10. Peat and detritus
11. Mammal remnants
12. Paleo and neolithic sites
13. Freshwater mollusks
14. Marine mollusks
15. Forminifera
16. Ostracoda
17. Freshwater diatoms
18. Marine and brackish diatoms
19. C^{14} data
20. Magnetic data
21. Concretions
22. Underground ice lenses
23. Pollen spectra
24. Tundra spectra
25. Forest-tundra spectra
26. North taigal spectra
27. Middle taiga
28. Southern taiga
29. Forest steppe
30. Steppe

Figure 78.—(Continued)

Legend for meridional correlation diagram of the Pleistocene deposits in the Ob and Yenisei river basins.

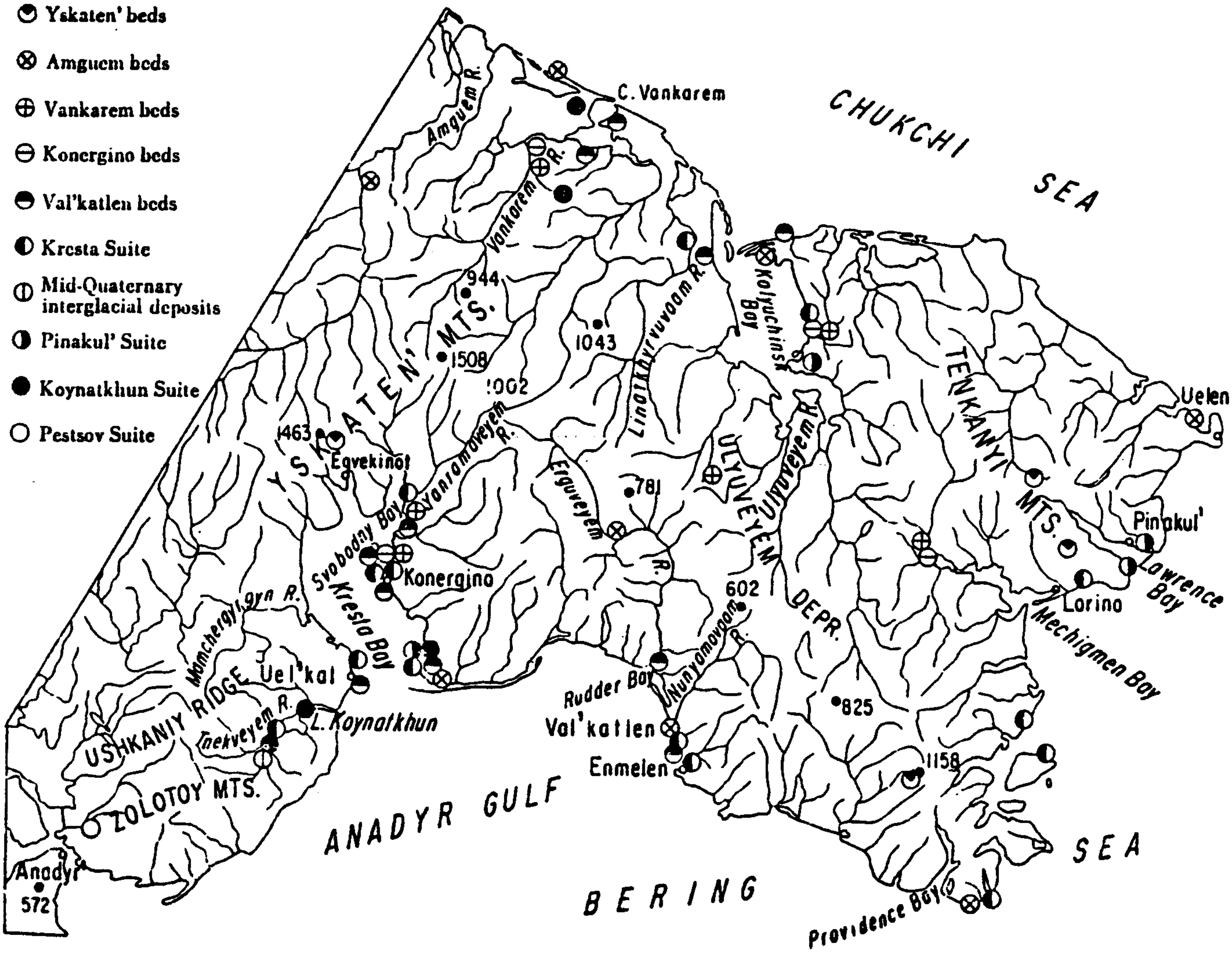


Figure 79.—Locations of the Pleistocene deposits of Chukotka. After Petrov (1965).

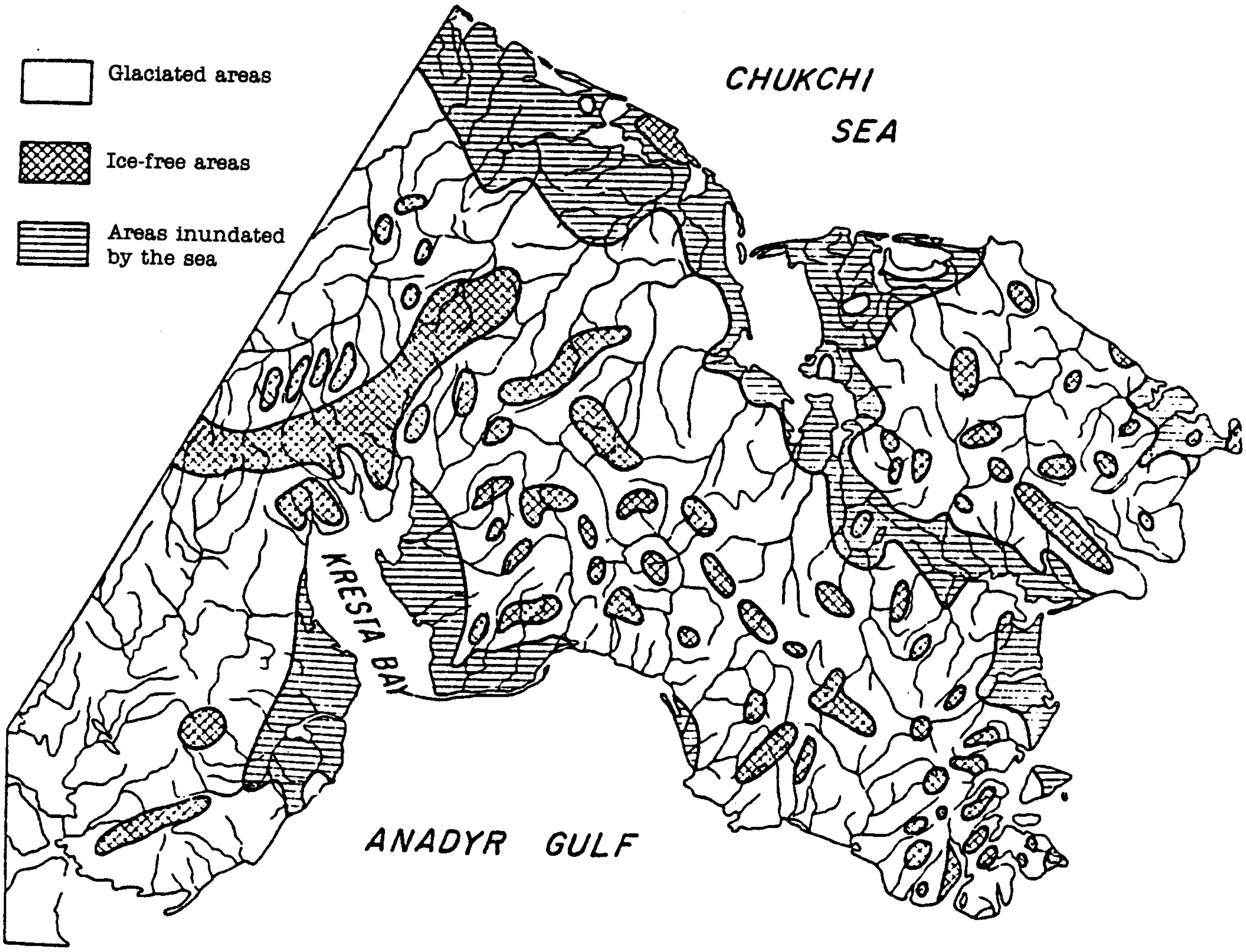


Figure 80.—Paleogeography of Chukotka during the maximal glaciation. After Petrov (1965).

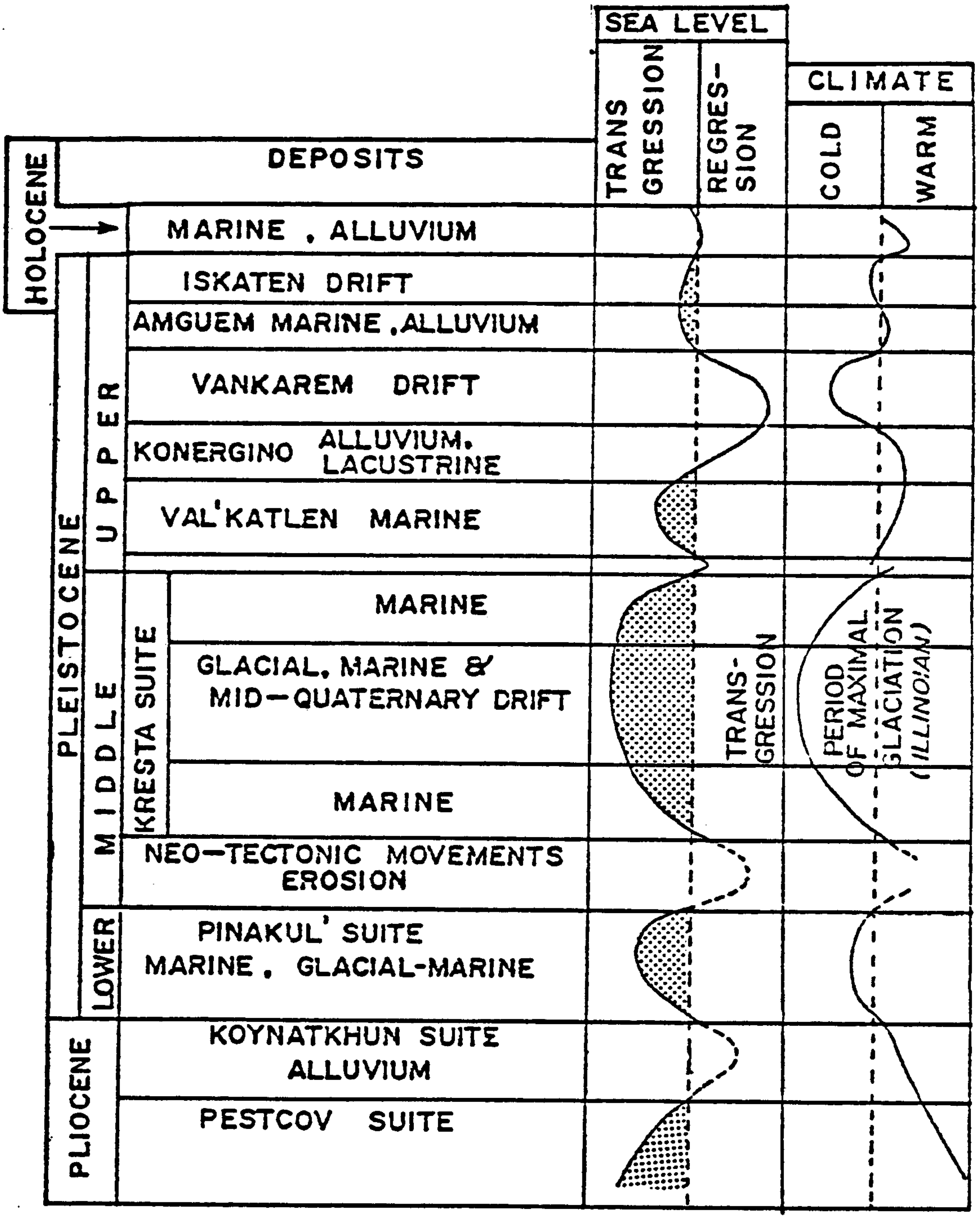
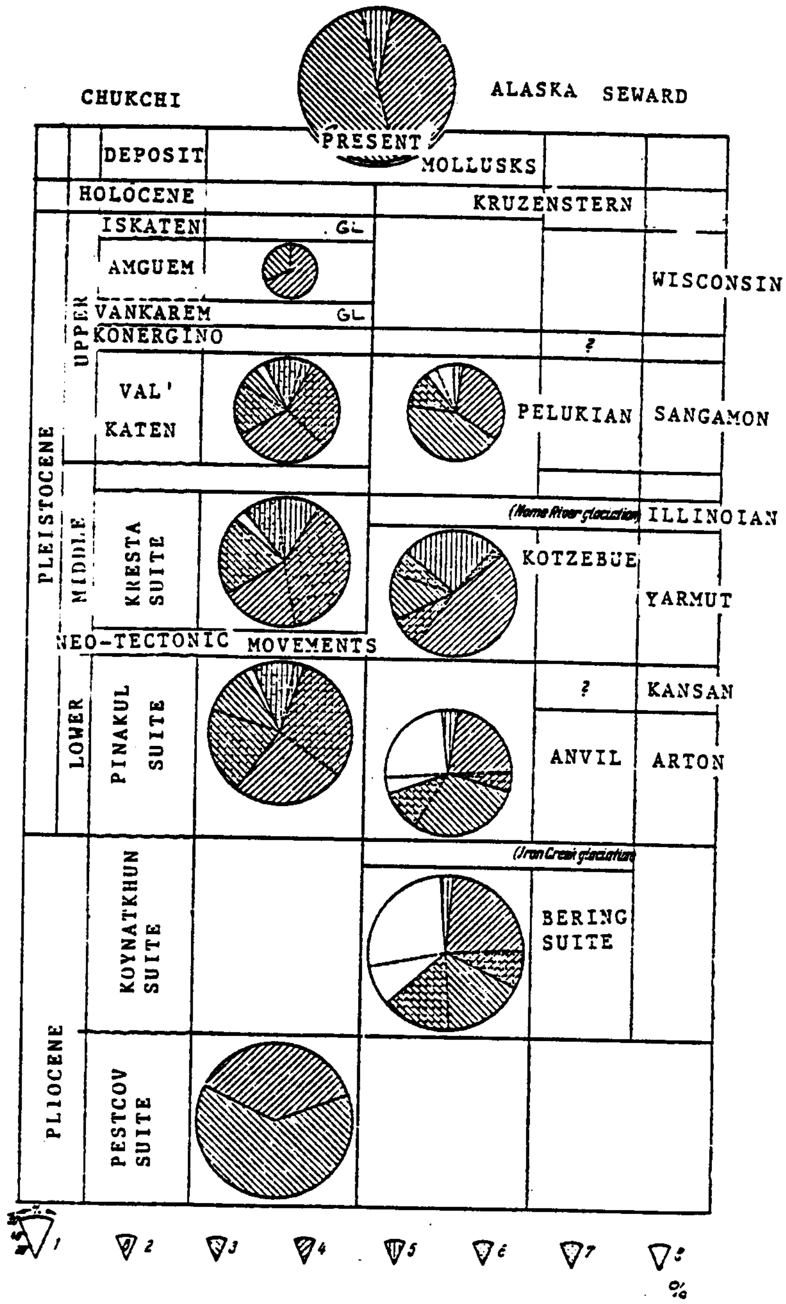


Figure 81.—Climatic changes and ocean level oscillation for the Chukotka. After Petrov (1965).



1. Radius–number of species, size of sector
2. Extinct
3. Boreal
4. Arcto-boreal
5. Arctic
6. Frequency
7. Seldom
8. Much

Figure 82.—Change in composition of fauna in the Bering and Chukchi seas. After Petrov (1965).

Table 19.—Fauna complex in the Chukchi and Bering seas (Petrov 1965).

Chukchi and Bering Seas Fauna Complex	Extinct	Boreal				Arcto-Boreal				Arctic				Total
		Seldom		Many		Seldom		Many		Seldom		Many		
		Number	%	Number	%	Number	%	Number	%	Number	%	Number	%	
Valkatlen	1	6	15	3	8	12	30	12	30	5	12	1	2.5	40
Kresta	1	11	20	–	–	18	36	10	20	5	10	6	12	51
Pinakul	1	11	20	6	11	16	25	14	25	6	11	1	2	35

Table 20.—Foraminifera, fauna, vegetation, and diatom complexes in Pleistocene deposits of Chukotka (Khoreva 1974).

	Foraminifera	Fauna	Vegetation	Diatomea
Valkatlen (Sangamon)	*Rhabdammina abyssorum* Sars, *Rheophax curtus* Cushman, *Recurvoides contortus sublittoralis* Saidova, *Ammotium cassis* (Parker), *Trochammina inflata* (Montagu), *Elphidiella recens* (Stschedrina), *E. urbana* Khoreva, *E. groenlandica* (Cushman), *Elphidium excavatum* (Terquem), *Cribrononion incertus* (Williamson), *Protelphidium orbiculare* (Brady), *Buccella frigida* Cushman, *Cibicides lobatulus* (Walker et Jacob), *Bulimina marginata* d'Orbigny, *Oolina borealis* Loeblich et Tappan, *Fissurina marginata* (Walker et Boys), *Nonionellina labradorica* (Dawson), *Cassidulina translucens* Cushman et Hughes.	*Neptunea beringiana, Argobuccinum oregonensis, Clynocardium californiensis* (Deshayes), *Cylichna occulta* (Mighels), *Astarte alaskensis* Dall, *A borealis borealis* (Schumacher), *A. montagui* (Dillwyn), *A. rollandi* (Bernardi), *Macoma calcarea* (Gmelin), *Mya elegans* (Eichwald), *M. pseudoarenaria* Schlesch, *M. truncata* Linne.	Shrub-Tundra	Arctic Arcto-Boreal Diatomea
Kresta Suite Cold Transgression	*Elphidiella arctica* (Parker et Jacob), *Protelphidium orbiculare* (Brady), *P. lenticulare* Gudina, *Elphidium subclavatum* Gudina, *E. subarcticum* Cushman, *Cribroelphidium granatum* Gudina, *C. goesi* (Stschedrina), *Stainfortia concava* (Höglund), *Pyrgo williamsoni* (Silwestri), *Fissurina marginata* (Walker et Boys), *Cassidulina islandica* Norvang, *Pseudopolymorphina curta* Cushman et Ozawa, *Bucella frigida* Cushman, *Cyclogura foliacea* (Philippi), *Cribrononion obscurus* Gudina, *Quinqueloculina borea* Gudina.	*Bathyarca glacialis* (Gray), *Yoldiella intermedia* (Sars), *Y. lenticula* (Möller), *Astarte borealis placenta* Morch, *A. alaskensis* Dall, *A. montagui* (Dyllwayn), *Neptunea satura heros* (Gray), *Tachyrhynchus erosus* (Gouthouy), *Hiatella arctica* (Lene).	Arctic Desert	Marine Arctic
			Forest Tundra	Arcto-Boreal
Pinaku	*Elphidium subclavatum* Gudina, *Protelphidium orbiculare* (Brady), *P. lenticulare* Gudina, *Cribroelphidium goesi* (Stschedrina), *C. granatum* (Gudina), *Elphidiella honnai* (Cu).	Portlandia arctica siliqua (Reeve), *Buccinum solenum* Dall, *B. terraehovae* Beck (Morch.), *Natica russa* Gould, *Clinocardium californiensis* (Deshayes).	Arctic and Shrub Tundra	Arcto-Boreal

Table 21.—Continental run-off of the Arctic Ocean (rivers and glacial water). After Antonov (1958).

Drainage Area	Volume of Annual Run-off (km^3)
Norwegia and northwest part of the USSR	153
Northern part of Europe	359
Northern part of Asia	2,442
Alaska and Canada with islands	1,053
Northern part of Greenland with islands	373
Average annual volume of water and ice received by the Arctic Ocean	4,380

Table 22.—Modern thermal balance of the Arctic Ocean (kcal/cm^2/year). After Borisov (1970).

Heat Income		Heat Expenditure	
1. Sum of radiation	72.6	1. Reflected radiation	54.3
2. Heat from Atlantic water	2.5	2. Effective radiation	16.8
3. Heat from Pacific water	0.4	3. Heat exchange with atmosphere	5.0
4. Heat from river water	0.2	4. Evaporation	5.0
5. Arctic heat conservation due to ice export	2.4		
6. Arctic heat conservation due to cool water export	3.0		
	81.1		81.1

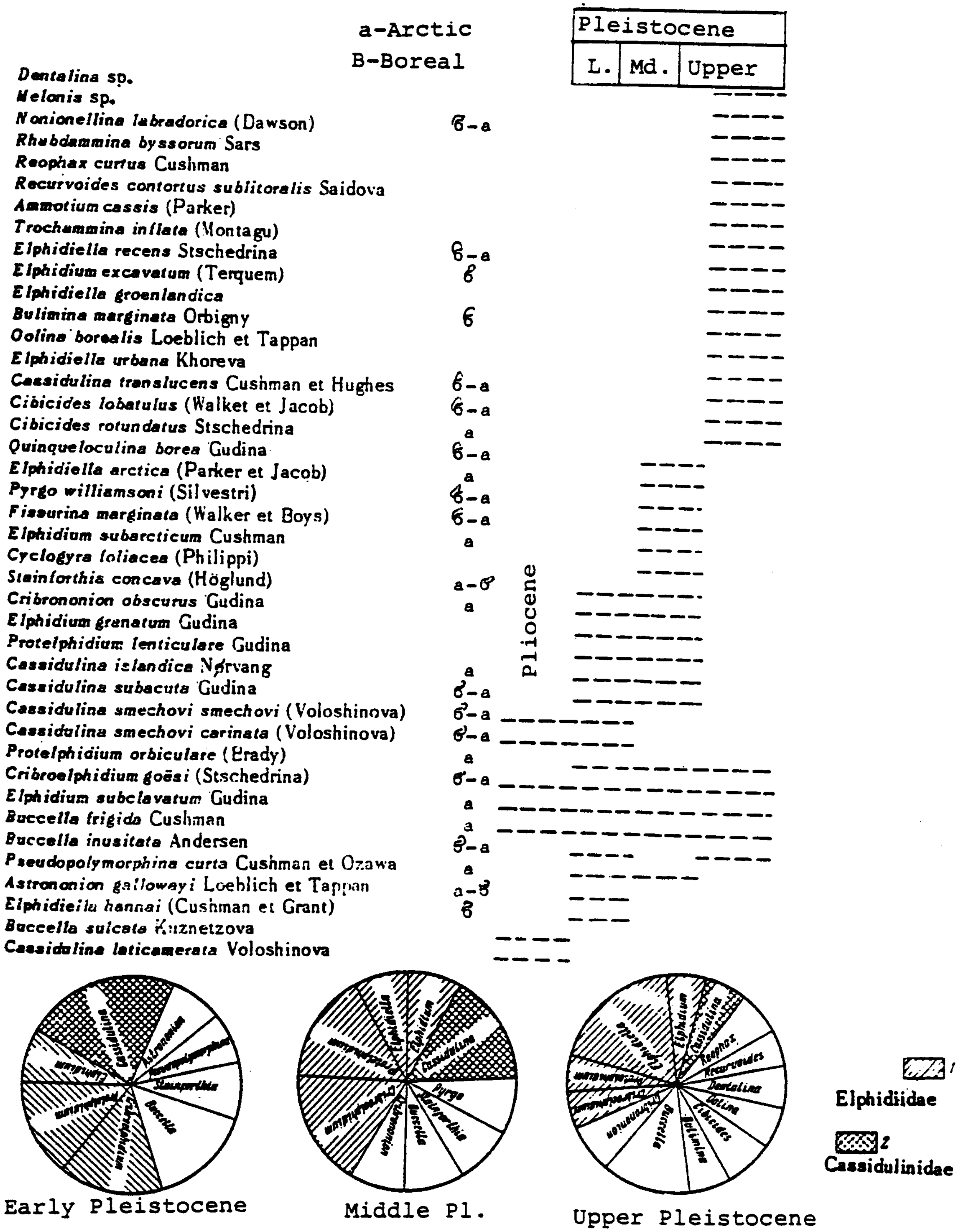


Figure 83.—Foraminifera complexes in the Bering Sea. After Khoreva (1974).

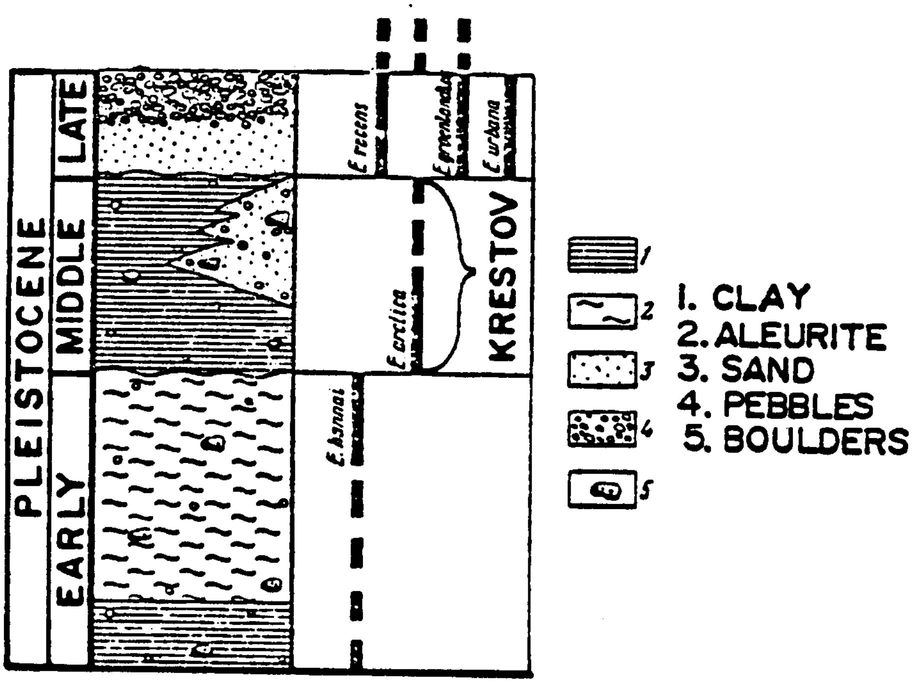


Figure 84.—Foraminifera in the different Pleistocene deposits.
After Khoreva (1974).

belt of shelf glacier. Glacial cover from the Kola Peninsula occupied the southwest part of the Barents shelf. Shelf glaciers might have dominated on the eastern part of the Barents Sea and particularly on the Pechora Sea shelf, but there are no reliable data to draw their limits in that region. The idea that there was a center of ice outflow on the Barents shelf (M. Grosswald 1967; V. Dibner 1967) is highly problematic, as no data of boulder composition in corresponding moraines have been obtained. Nevertheless, ice lobe outlines determined by marginal moraines confirm the possibility of ice spreading to the land from the White and Barents sea bottom along Vrangee and Rybachy peninsulas (western part of Barents Sea). Submerged ridges considered as marginal moraines were found only at depths of 100–150 m (O. Hotedabl 1953; G. Matishov 1974). The dominant part of the Barents Sea had never been occupied by an ice sheet.

We see that all these hypotheses have contradictions and do not satisfactorily explain the origin of the cold marine transgression which took place during the Illinoian (Dnipro, Samara) time.

The only alternative appears to be the isolation of the Arctic basin at that time, involving barriers along both the Bering Strait and North Atlantic areas.

Table 23.—Carbon-14 dates and sedimentation intensities in the Laptev Sea (M. L. Holmes amd I. S. Creager 1974).

Core	Depth in Core (cm)	Radiocarbon Age (Years B.P.)	Average Sedimentation Intensity (mg/cm/yr)*
137	3–17	3,410 ± 230	9
	103–117	18,400 ± 540	
143	24–32	14,200 ± 370	2**
144	3–20	6,050 ± 200	
	44–64	15,000 ± 460	
147	89–102	11,040 ± 310	15**

* Based on elapsed time between dates.

** Assumed surface interval (3 to 20 cm) date 6000 yr. B.P.

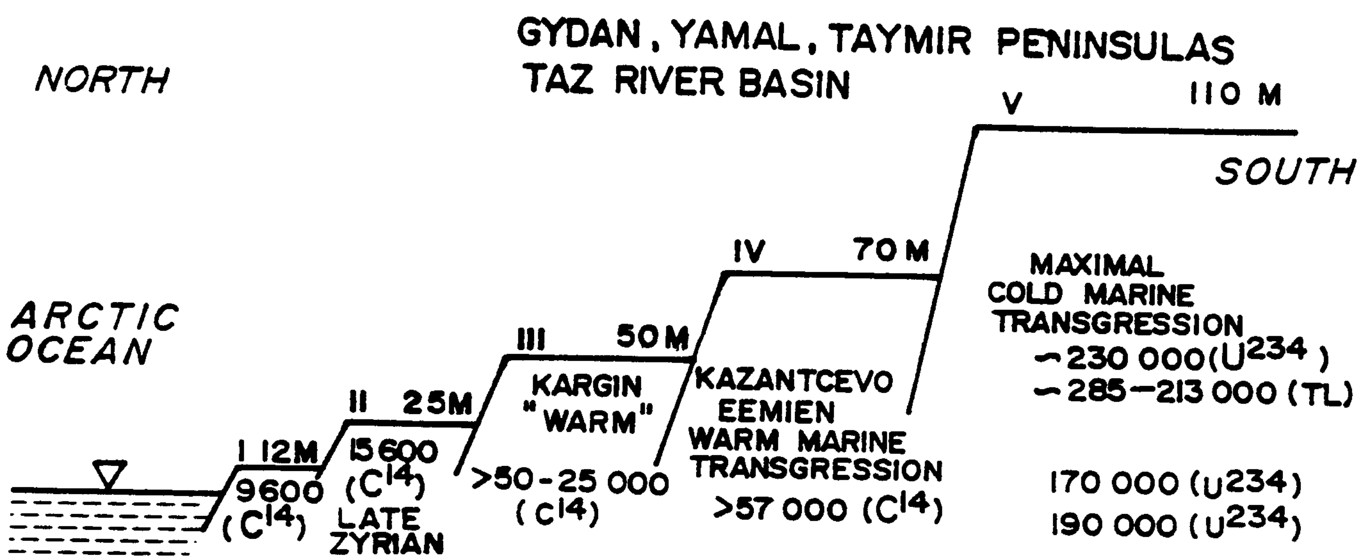


Figure 85.—Marine terraces of Siberia (Arhipov 1971).

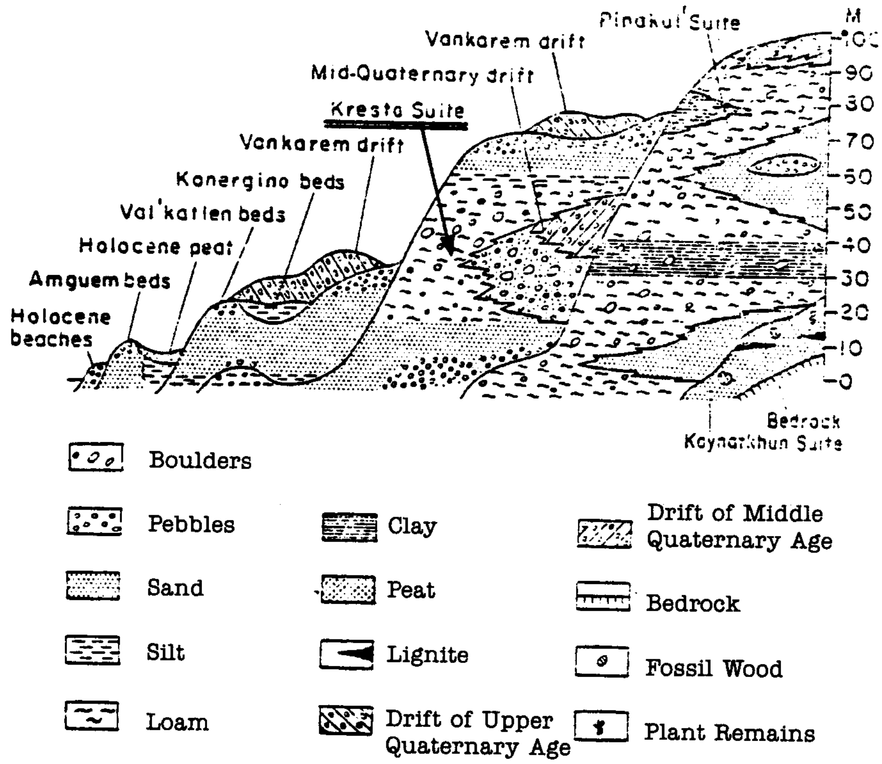
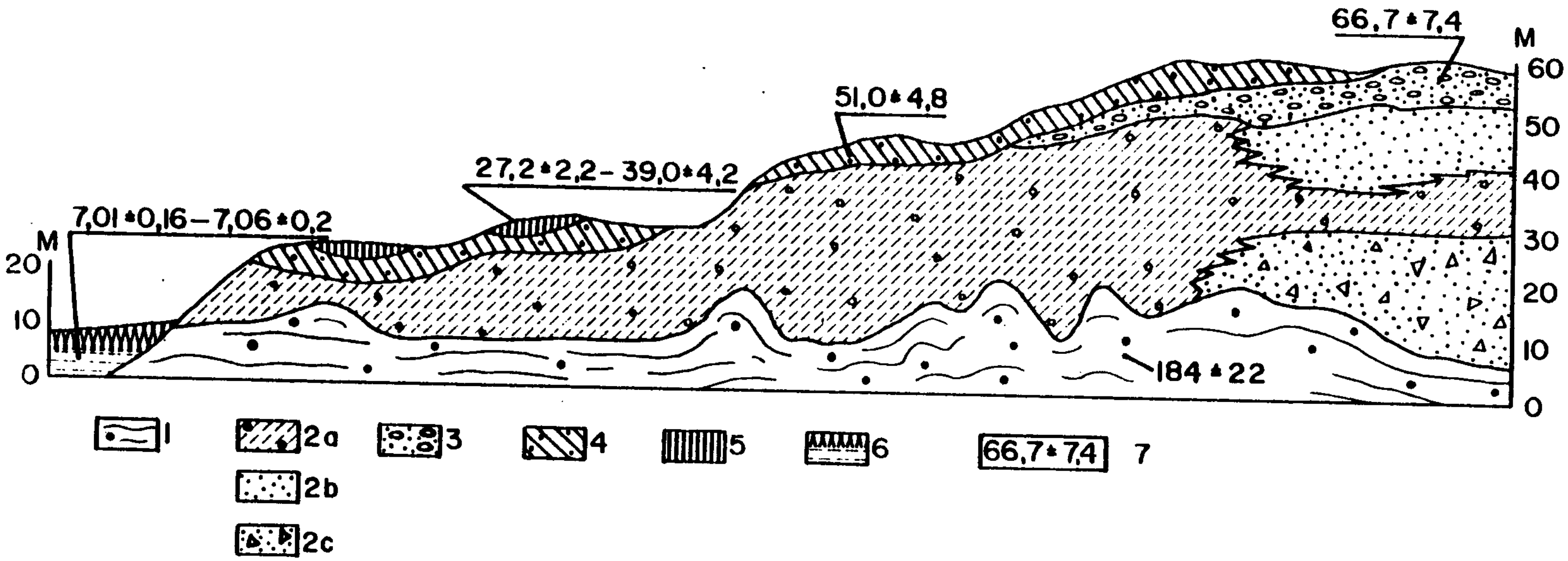


Figure 86.—Marine terraces of Chukotka Peninsula (Petrov 1965).



1	МГУ–КТЛ–207	39,0±4,2
2	МГУ–КТЛ–206	51,0±4,8
3	МГУ–КТЛ–209	66,7±7,4
4	МГУ–КТЛ–169	184±22
5	МГУ–320	7,06±0,2
6	МГУ–321	7,01±0,16
7	МГУ–201	27,2±2,2

1. Marine deposits of the Krestov-Skaya Suite (Middle Pleistocene)
2. a. Glacial marine deposits
   b. Beach sand
   c. Marine deposits (end of middle-beginning of Late Pleistocene)
3. Deposits of terrace 50–60 m (middle part of the Upper Pleistocene)
4. Glacial deposits of middle part of the Upper Pleistocene
5. Deposits of terrace 20–30 m (end of the Upper Pleistocene)
6. Deposits of terrace 5–7 m (middle part of the Holocene)
7. Absolute date, thousand years

Figure 87.—Age of the terraces in Chukotka. After Svitoch (1976).

Eurasiatic coastal area according to Degtyarenko et al (1971).

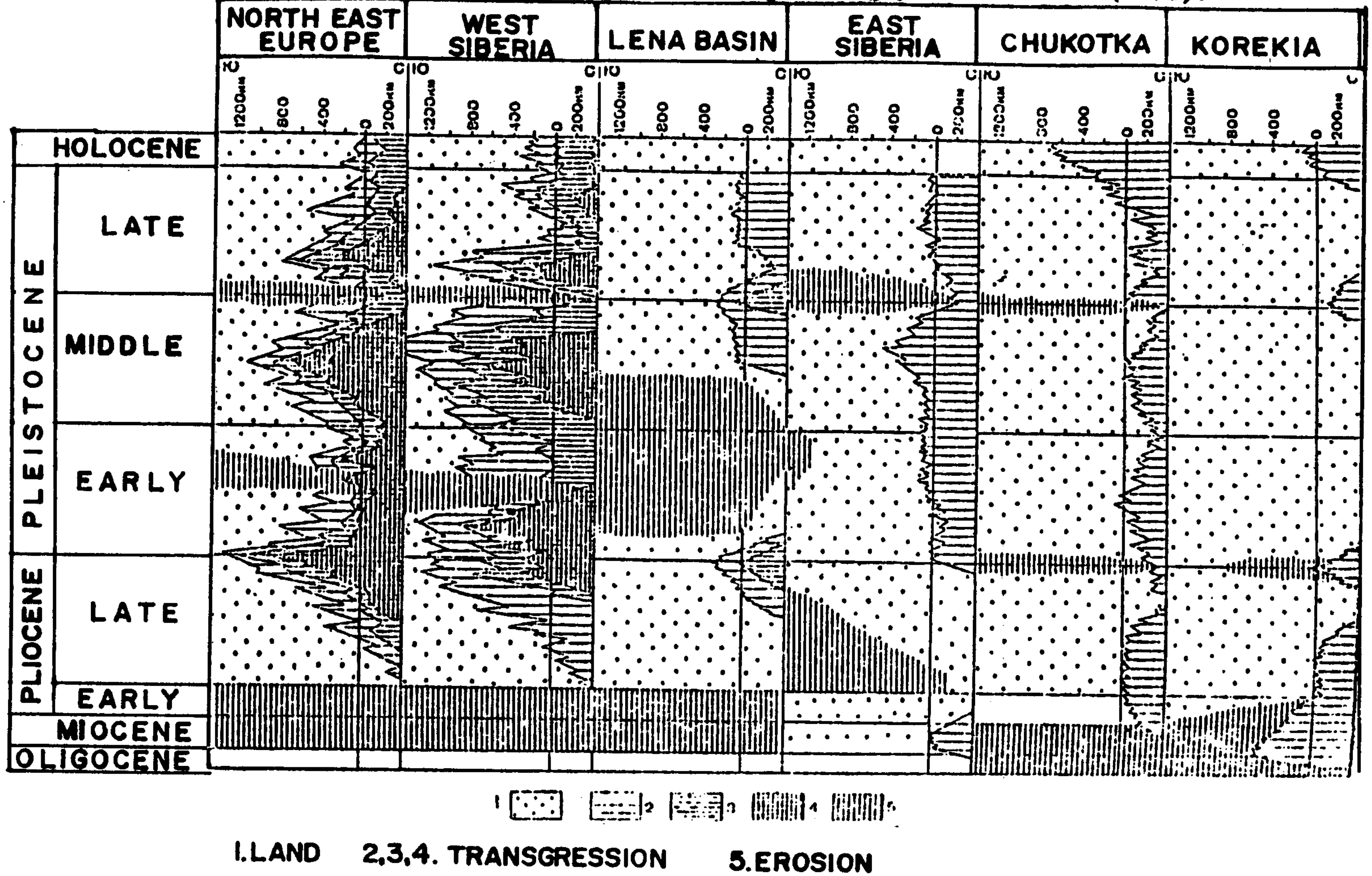


Figure 88.—The development of Pleistocene transgressions on the Eurasiatic coastal area (Degtyarenko et al. 1971).

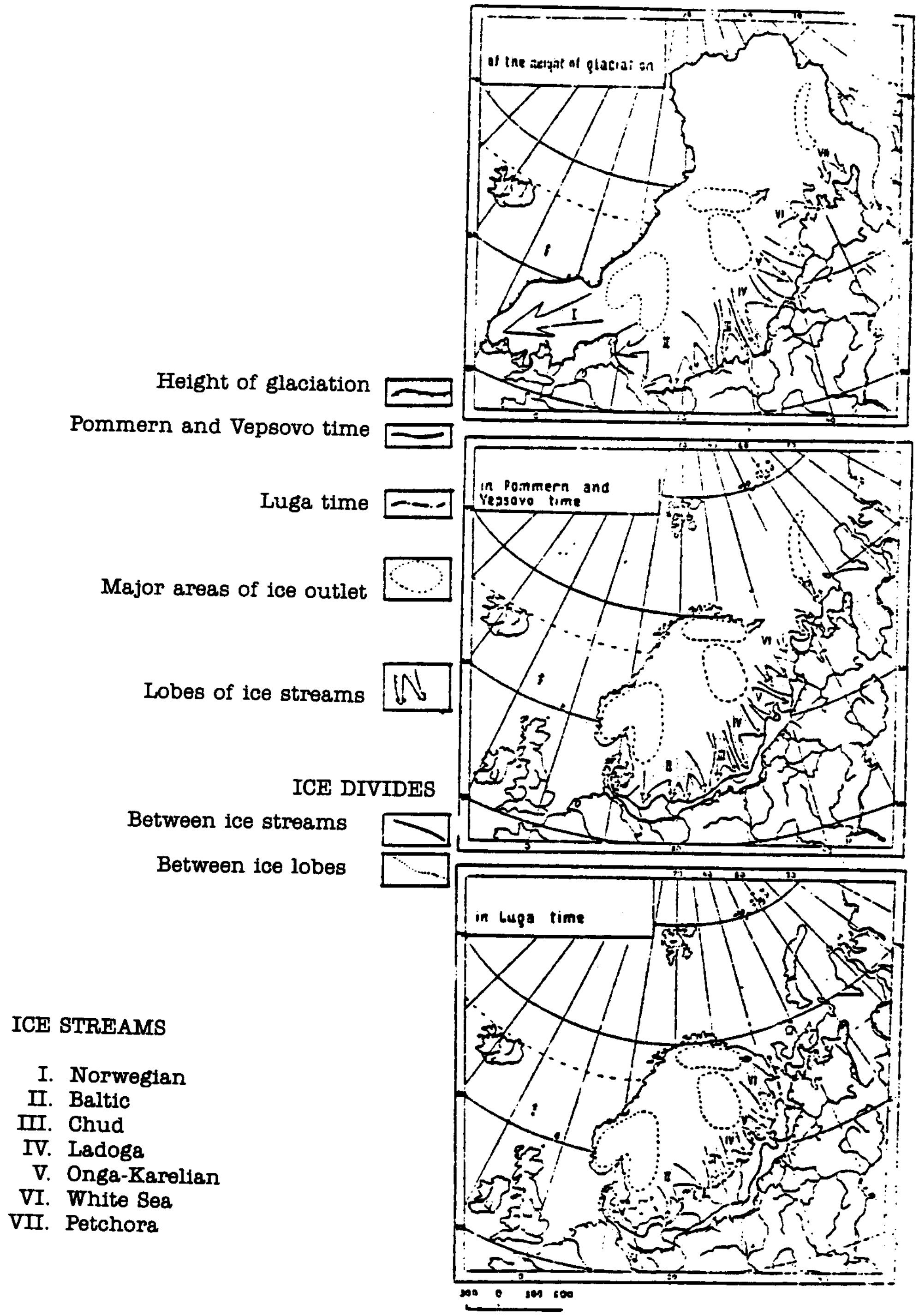


Figure 89.—Glaciation in the Barents Sea (Grosvald et al. 1973).

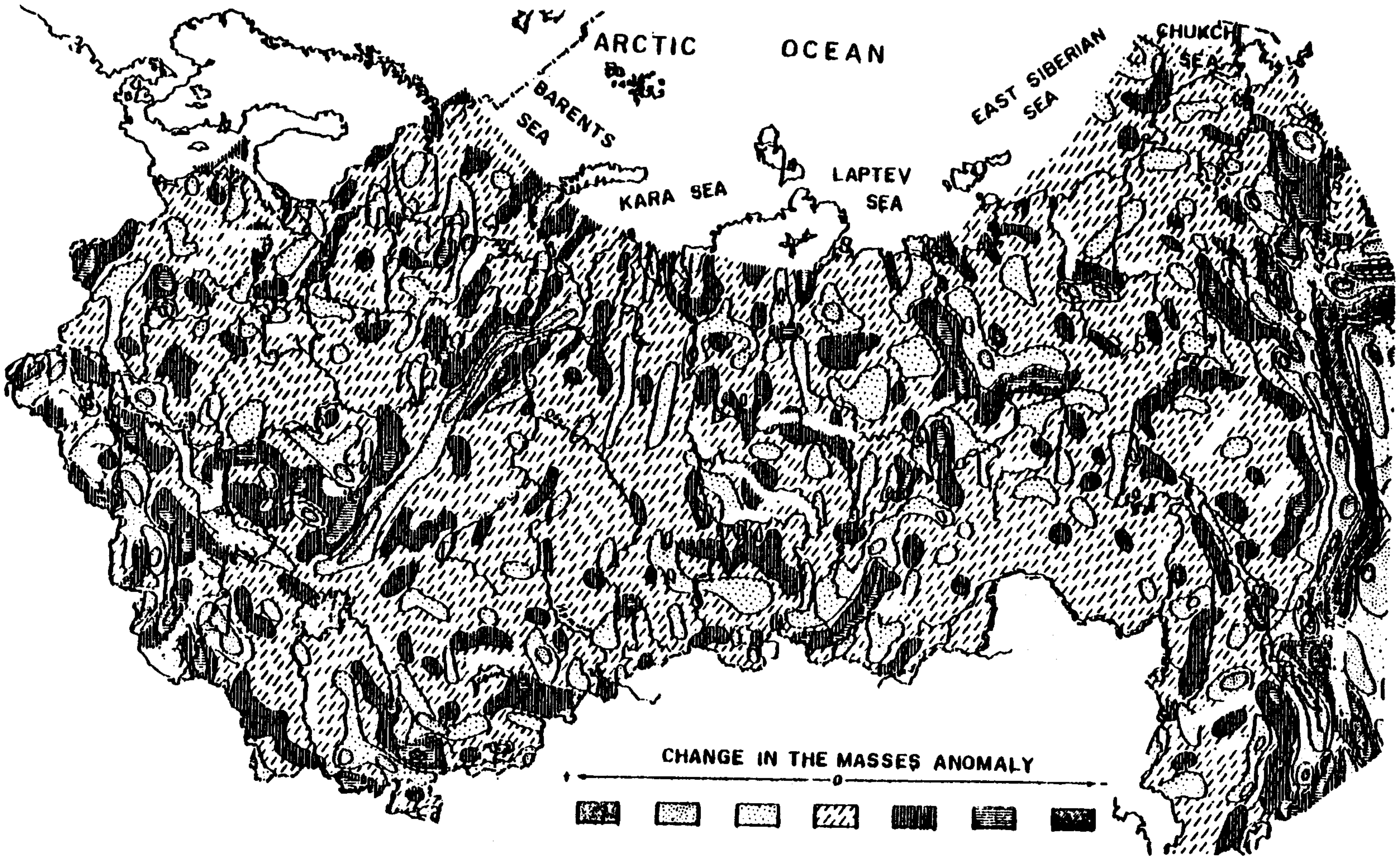
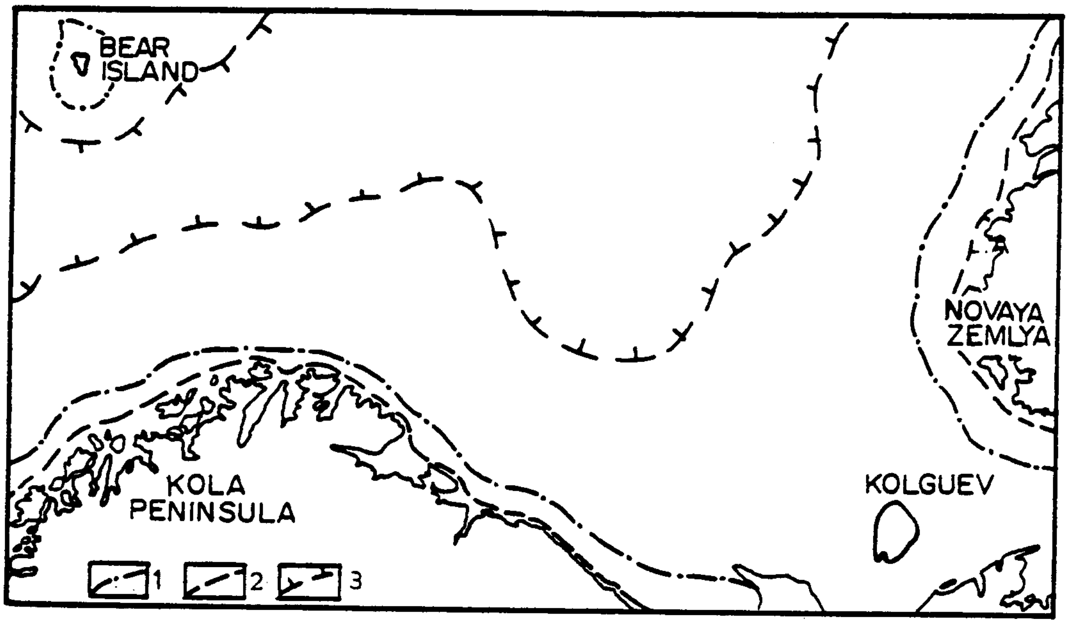


Figure 90.—Distribution of isostatically uncompensated masses in the Soviet Union. After Artemiev (1974).



1. Last glaciation
2. Last glaciation
3. Maximal glaciation (Illinoian)

Figure 91.—Boundaries of the glaciations in the Barents Sea. After Matishov (1975).

North Atlantic Barriers and Mechanisms of Dam Creation

Figure 92 shows the extent of the Quaternary glaciation in Alaska (T. Péwé 1976).* We see that the Bering Strait was closed by Illinoian glaciation. Glaciers from the Chukotka Peninsula (USSR) adjacent to the Bering Strait once extended southward and eastward more than 100 km onto the Bering shelf, reaching St. Lawrence Island (Hopkins et al. 1972, p. 125). Hopkins (1972) believed that the glaciers could hardly have extended this far if it were a time of high sea level when glacial margins must have been afloat, as suggested by some Soviet workers. Péwé notes that Grim and McManus (1970) found large-scale deformation structures on the sea floor probably caused by glacial ice in contact with the ground surface.

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* Modified from Coulter, Hopkins, Karlstrom, Péwé, Wahrhaftig, and Williams (1965). Additional information on Seward Peninsula and Bering Strait from D. M. Hopkins and C. L. Sainsbury (written commun., 1968) and Nelson and Hopkins (1972).

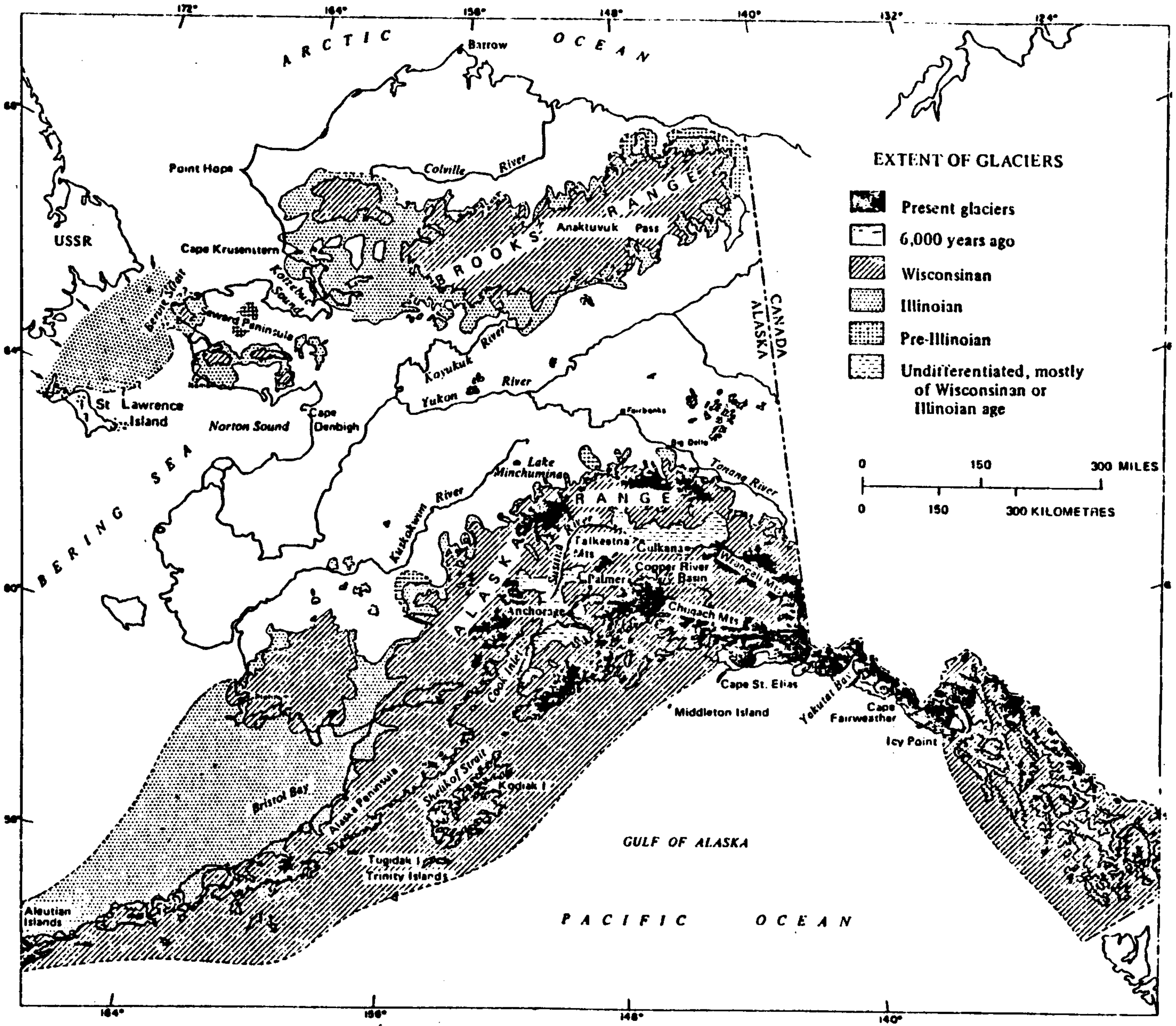


Figure 92.—Extent of Quaternary glaciation in Alaska (Péwé 1976).

Hopkins (1972) further concluded that this extent of Siberian ice may have occurred during Illinoian time rather than during a time of high sea level, the Kotzebuan transgression. New work by Hopkins has extended the limits of the Nome River Glaciation in the southern part of the Seward Peninsula and in the Bering Strait (cited in Péwé, 1976).

On the map of the Atlantic barriers (Figure 93) we see that the water depth here now is not more than 500 m and the shapes of the geological structures look like blocks. During the glaciations the ice sheets were developed all over Iceland, the Faeroe Islands, and Scotland (Figures 94 and 95). These ice sheets had to give a loading impulse for glacio-isostatic movements, and the earth core blocks under the ice sheets may have been submerged. On the paleotectonic scheme of the Cenozoic, made by M. Muratov** (1975), the North Atlantic barrier areas are shown as an alternation of young and old platforms and of continental oceanic blocks. On the tectonic map of the Arctic basin made by V. Dibner et al. (1965) the regions represented by Greenland, the Faeroes, and Scotland are separated by activated deep faults (Figure 96). We can consider them as transform faults, also according to plate tectonics. Another map by the same authors shows the thickness of the earth's core in this region (Figure 97).

Recently the geomorphology of the floor of the Norwegian and Greenland seas was described in detail by V. M. Litvin (1973) according to the new results of the investigations carried out by Soviet and foreign expeditions during the last fifteen years. The relief of the floor of the Norwegian and Greenland seas has been fairly well studied. Very extensive data have been obtained through expeditions by the Polar Research Institute of Fishery and Oceanography aboard the ships *Sevastopol, Akademik Knipovich,* and others (Vinogradova et al. 1960; Litvin 1964, 1968). During the IGY, investigations of the structure of the floor of the northern part of the seas were carried out through expeditions by the Arctic and Antarctic Institute aboard the *Ob* and the *Lena* (Voekov et al. 1968; Laktionov 1960); in the central and southern parts, Norwegian expeditions operated aboard the *Yogann, Yort,* and the *G. O. Sars* (Eggvin 1961). In recent years, American expeditions on icebreakers carried out geological and geophysical investigations of this region (Johnson et al. 1966–69). It should be mentioned that detailed operations carried out in the earlier period by Norwegian scientists on the Norwegian Shelf yielded clues for an understanding of the structure and origin of glaciological shelves (Holtedahl 1940, 1956). Finally, the latest investigations in the region of Iceland and the Jan Myan Islands were carried out in 1971 by an expedition of the Academy of Sciences of the USSR aboard the *Akademik Kurchatov.*

** M. Muratov (1975) Proishogdenie Materikov i Okeanicheskih vpadin, Moskva, Nauka.

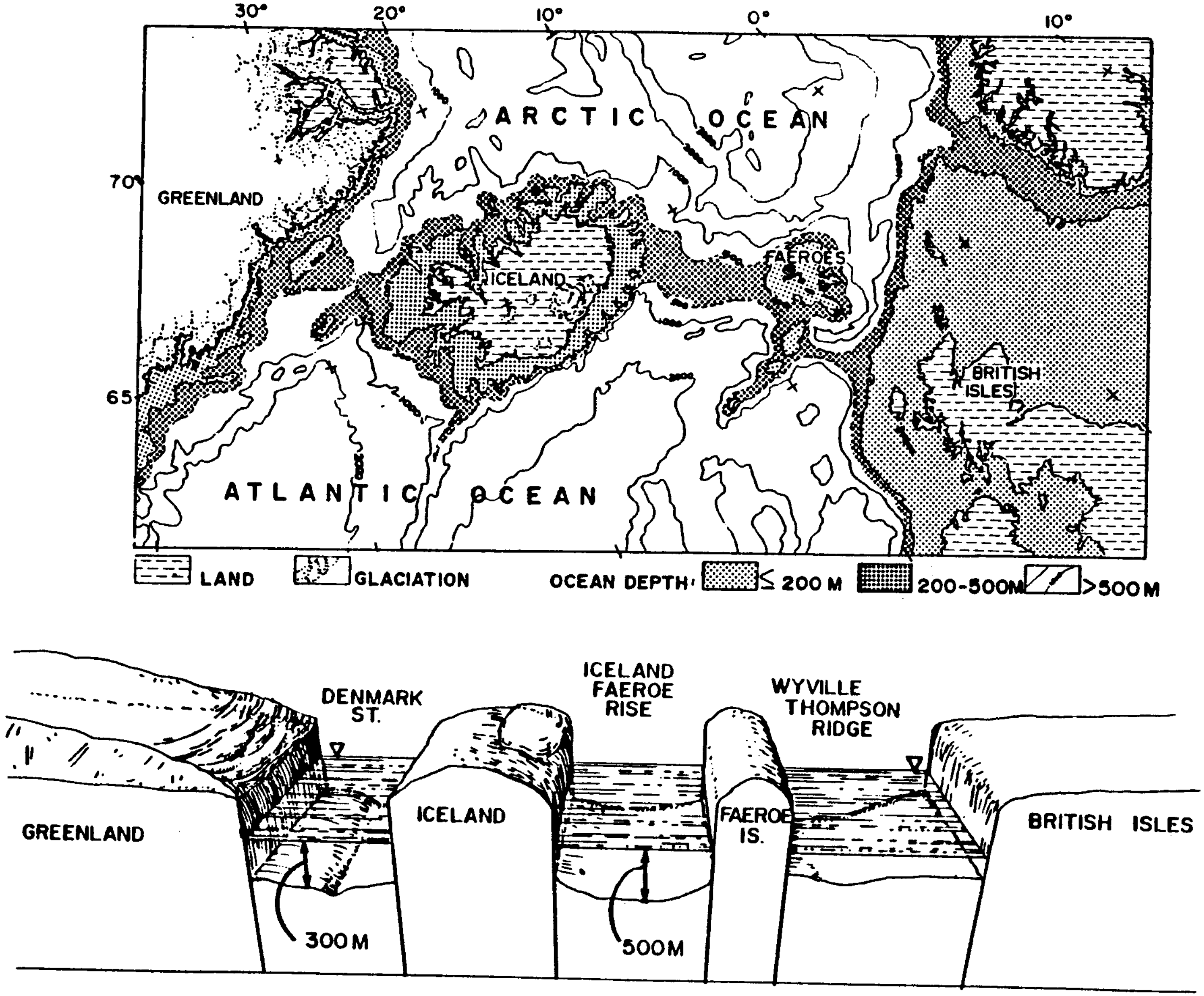


Figure 93.—North Atlantic barriers.

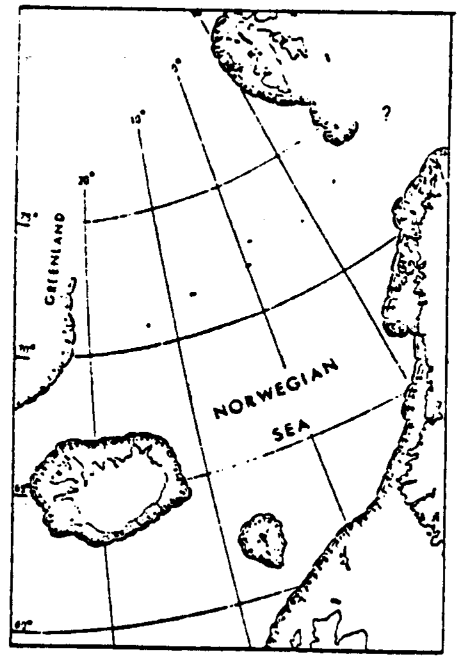
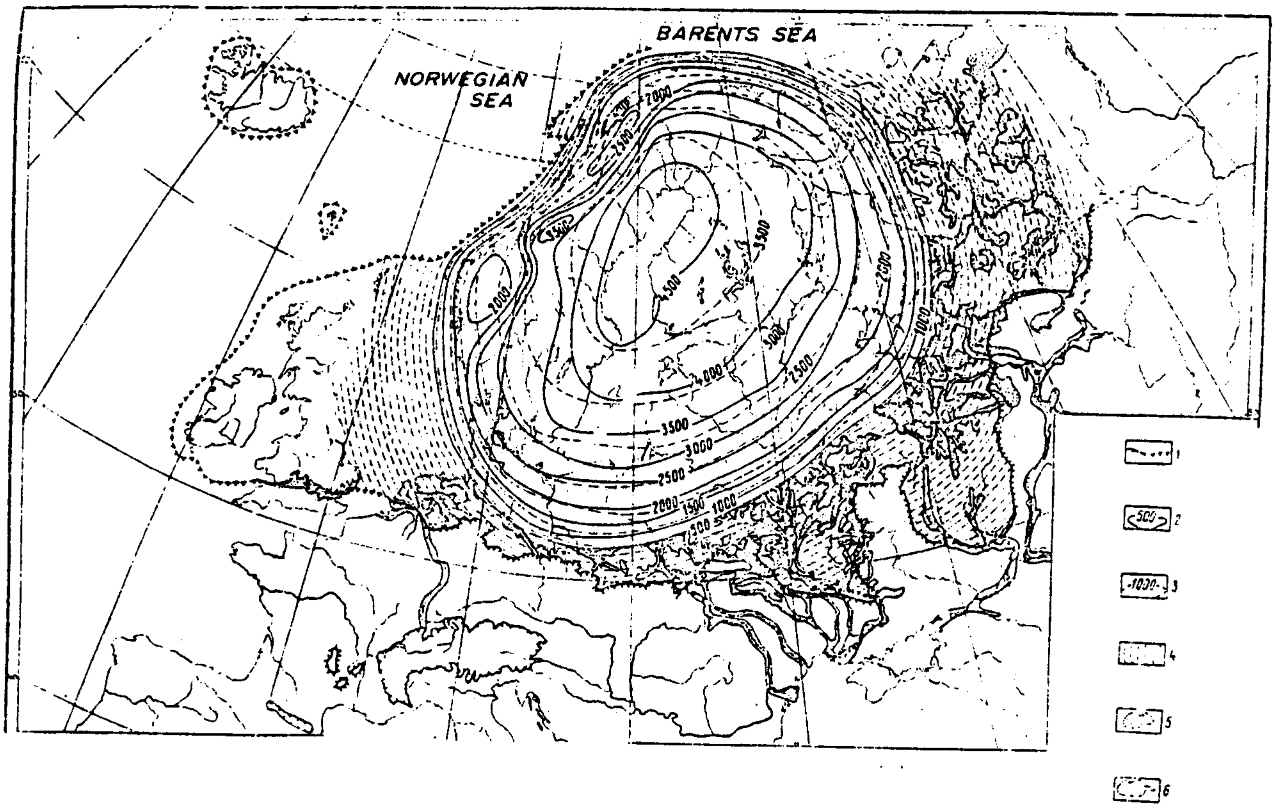


Figure 94.—Probable extent of the northwestern European ice sheet (Hoppe et al. 1972).

The present geomorphologic outline and the geomorphologic map (Figure 98) constructed by V. Litvin (1973) are based on the data of Soviet and foreign investigators.

The Norwegian and Greenland seas are bounded on the east and west by the continental borders, including the shelf and the continental slope. In the south and southwest, the seas are separated from the Atlantic Ocean by a chain of rises which include the islands of Iceland and the Faeroe Islands with their shelves and the underwater Wyville Thomson and Faeroe–Iceland Ridges, and the Iceland–Greenland Rise.

According to geophysical data, the lithosphere of the Norwegian and Greenland seas within the limits of the shelf has a continental type structure (Demenitskaya et al. 1964; Ewing et al. 1959; Johnson et al. 1969). The thickness of the crust here reaches 30–33 km. In the zone of the continental slope, the crust generally becomes thicker and, at its foot, reaches 10–15 km. Here the "granite" layer tapers off. In the deep-water troughs the thickness of the crust is less than 10 km, and its section similar to the oceanic type. In the zone of the mid-oceanic ridge, the lithosphere becomes appreciably thicker reaching, in the regions of Iceland and the Jan Myan Islands, a thickness of 25–28 km, due to the increased thickness of the "basalt" layer. The section of the crust here is almost identical to the section of the crust of the mid-Atlantic Ridge, which is evidence for the genetic link of these morphostructures.



1. The limit of spreading of Middle Pleistocene glaciation (established and tentative)
2. Contour lines of equal thickness of ice in the Scandinavian ice sheet (m)
3. Contour lines of the surface of the Scandinavian ice sheet, meters a.s.l.
4. Periphery of the ice sheet (altitude of the surface and ice thickness under 500 m)
5. Outwash plains of the beginning of ice degradation epoch
6. Largest alluvial plains mainly formed by salt water

(The paleogeography of Europe during the late Pleistocene, reconstructions and models, Institute of Geography of the USSR Academy of Sciences, Moscow, 1973).

Figure 95.—Reconstruction of the maximum (Middle Pleistocene) European ice sheet (Aseev et al. 1973).

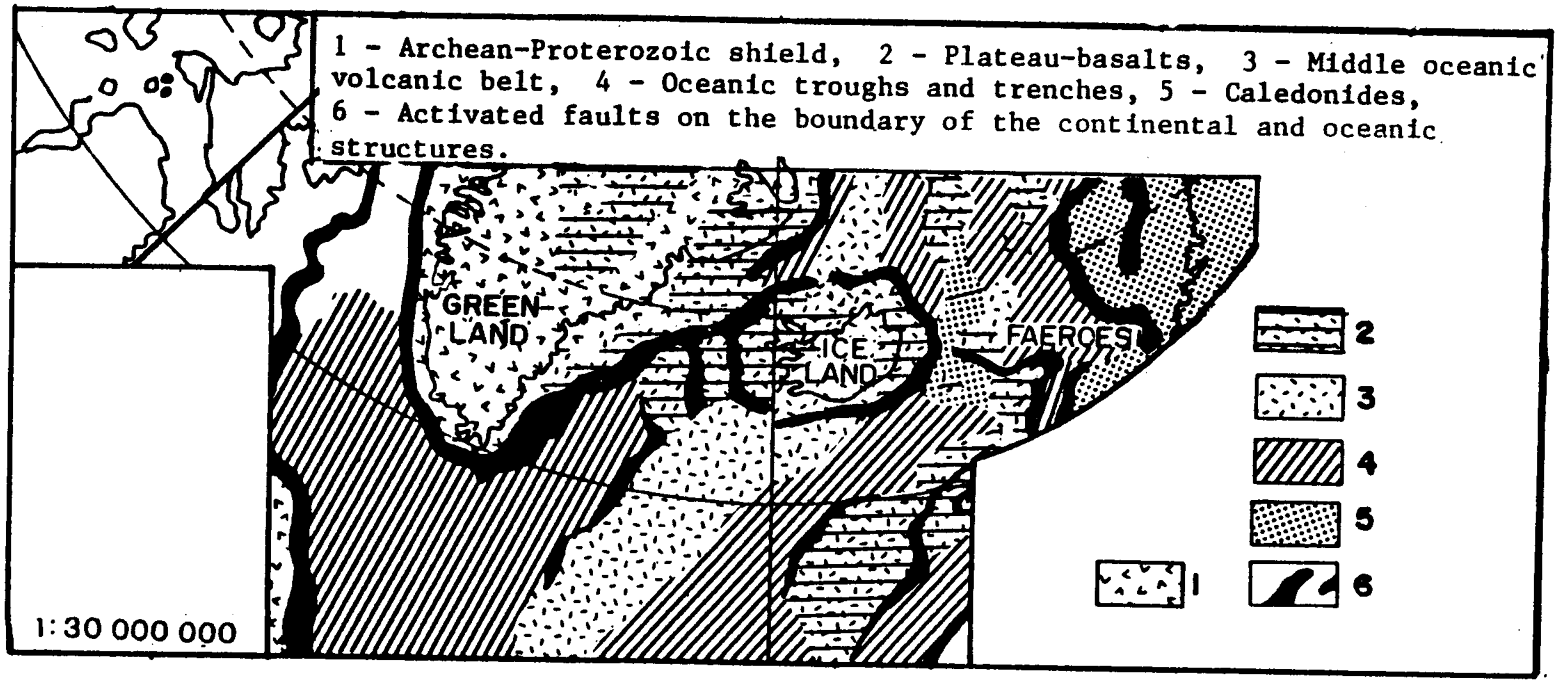


Figure 96.—Tectonic map of the Arctic Ocean (Dibner et al. 1965).

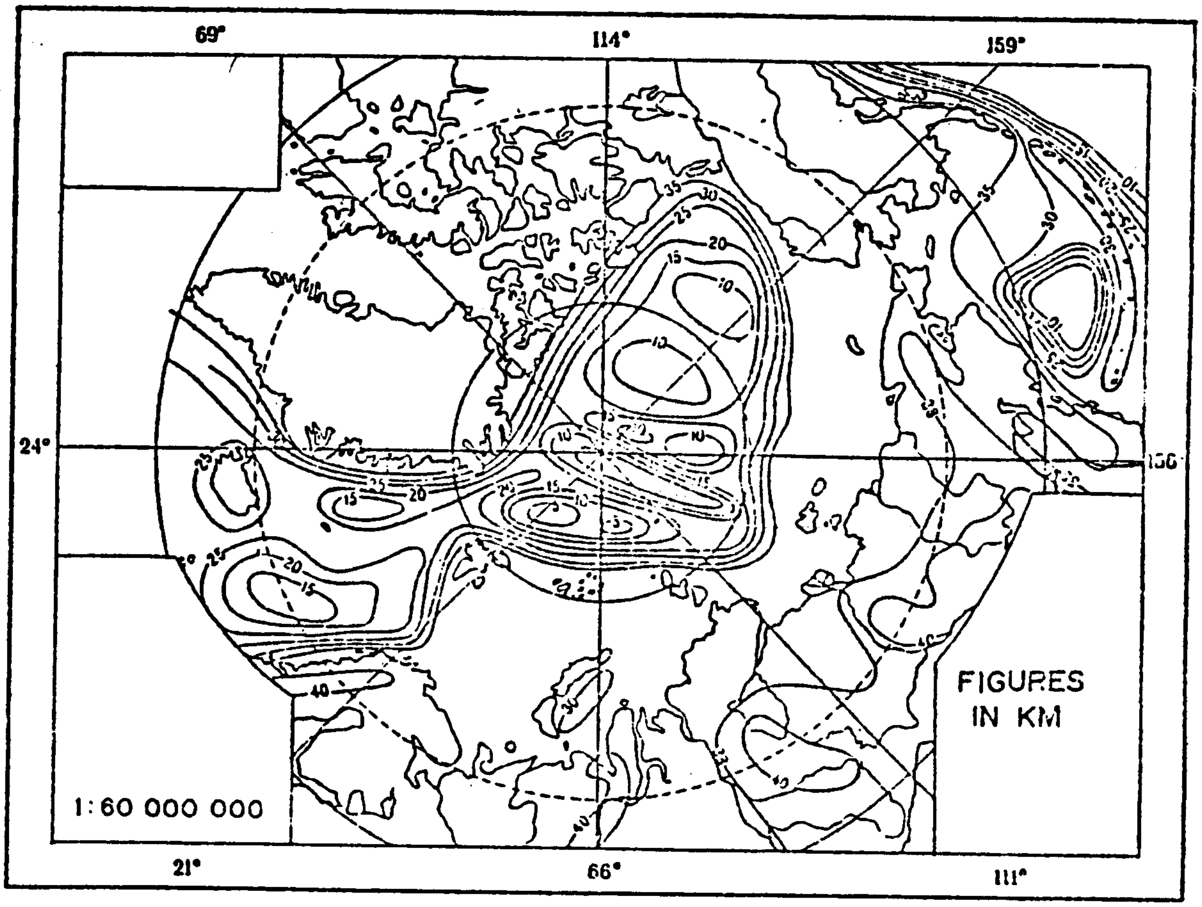
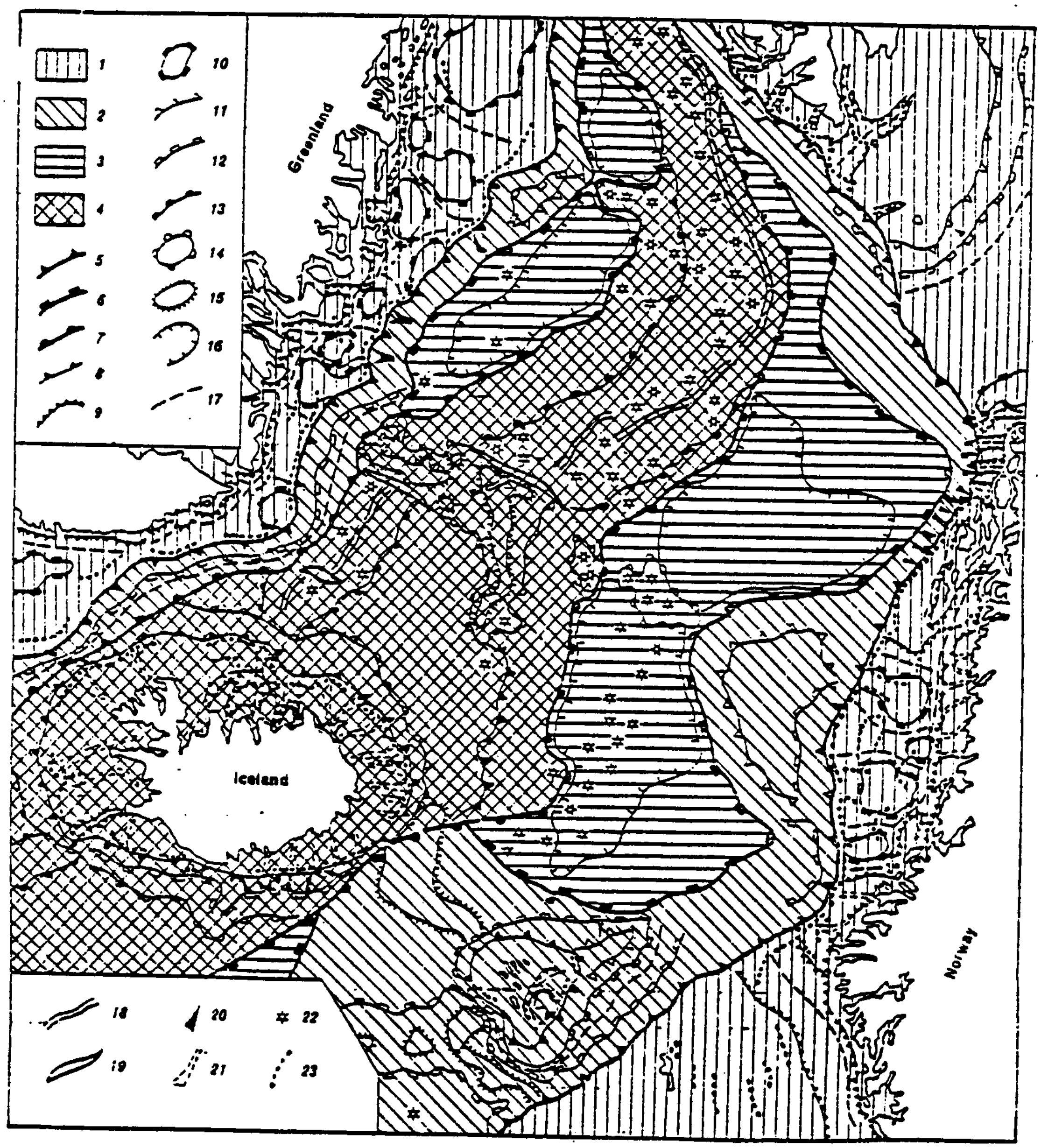


Figure 97.—Thickness of the earth's core in the Arctic basin (Dibner et al. 1965).

The continental shelves extending along the coasts of eastern Greenland, western Spitsbergen, Norway, and the western borders of the Barents Sea, despite some differences, have many common features. Almost all of them are separated by a series of tectonically defined grooves into a number of raised banks. The longitudinal grooves are most clearly expressed along the coasts of Greenland and Norway and in the remaining regions are negligible. They are the morphologic expression of the edge fractures of the continental blocks embedded in the Tertiary period, and rejuvenated by glacio-isostatic movements in the Pleistocene epoch. The transverse grooves are developed nearly everywhere but concentrated, for the most part, opposite the large fiords on land, which is indicative of their genetic link.



1. Continental shelf
2. Continental slope
3. Basin bed
4. Zone of mid-oceanic ridge
5. Edge of continental shelf
6. Foot of continental slope
7. Contour of mid-oceanic ridge
8. Edge of island shelf
9. Small tectonic scarps
10. Contour of rises (banks)
11. External edge of large steps
12. Foot of large scarps
13. Contour of fluted ridges
14. Contour of volcanic massifs
15. External edge of surfaces of rises, block-shaped ridges, and flat-topped mountains
16. Bottom contour of deep-water troughs
17. Axes of tectonically defined grooves
18. Rift valleys
19. Transverse grooves (fractures)
20. Underwater canyons
21. Glacial valleys
22. Volcanic underwater mountains
23. Terminal moraine ranges

Figure 98.—Geomorphologic scheme of the Norwegian–Greenland Basin (Litvin 1973).

The coastal belt of the shelf up to a depth of nearly 50 m, in general, forms an underwater abrasive platform with uneven glacial exertions, the platform being known as a strandflat (Holtedahl 1956). It is cut across by narrow valleys which are underwater extensions of fiords. The surfaces of the banks of the shelf at depths less than 200 m are, for the most part, leveled by abrasive-accumulative processes during the postglacial rise of the ocean level. The floors of the grooves at depths up to 350–450 m are also leveled by the intense accumulations of deposits. The remaining territory of the shelves of the Norwegian and Greenland seas has a shallow, undulating relief caused by extensive development of relict glacio-accumulative formations. The terminal moraine ranges in the transverse grooves, and along the external edge of the shelf, are particularly marked, and indicate the maximum extent of glacial cover within the limits of the shelf. Terracing of the surface of the shelf of the Norwegian and Greenland seas is also observed. The terraces at depths of 70–90, 150–180, and 240–270 m are most distinctly traced. Their development took place when the ocean level was lower. The last terrace is dated as pre-glaciation time in the Northern Hemisphere (Matuyama–Brunhes boundary) when the shelf, for the most part, was land.

The continental slope along both sides of the Norwegian and Greenland Seas has, on the whole, a fairly simple structure. Its dismemberment by underwater canyons, most numerous near the coast of Norway, is typical. The continental slope is less steep along the coast of Greenland and near the southwestern coast of Norway where it has a noticeable block-like dismemberment.

The formation of the continental slope is due to tectonic processes. It probably represents an edge flexure which developed at the boundary of the continental and oceanic crusts, complicated in a number of places by systems of fractures and faults. The canyons of the slope lie along lines of tectonic disturbances and are shaped by underwater exogenic processes (for example, sludge flows). The edge plateaus represent blocks separated from the shelf and immersed, covered later by a train of deposits carried away from the land and the shelf. Accumulative trains pile up at the foot of the continental slope forming its characteristic concave profile and, on the inclined plains of the foot of the continent, extending into the seabed. Portions of the very flat continental slope opposite mouths of shelf channels represent, evidently, hugh debris cones of sedimentary material, which were carried over a long period from the shelf along these channels, particularly in the period of their subaerial development during the large, fast regressions of the ocean (see Figure 47).

The island shelves of Iceland and the Faeroe Islands are similar in many respects to the continental shelves of the seas. Traces of the action of Pleistocene glaciers in the form of glacio-erosional (valleys and channels) and glacio-accumulative (small hills and ridges) structures of underwater relief are most pronounced on the island shelves. The surface of the underwater Wyville Thomsom and Faeroe–Iceland ridges, as also of the adjoining flat-topped banks of the Rockall Rise, is distinguished by a relatively smooth, slightly undulating form of relief. It is quite probable that they are composed of covers of Tertiary plateau-basalts known in Iceland, on the Faeroe Islands, in Greenland, and in Scotland (Gakkel et al. 1968; Muratov 1961). On the other hand, judging by the geographical position, the morphology, and the finds, together with the basalts of the bedrocks of the sedimentary-metamorphic complex on the Faeroe–Iceland Ridge, this region should probably be considered a submerged part of the old Eria platform. Near the continental shelf the rise is crossed by the U-shaped Faeroe–Shetland trough. A similar trough crosses the Iceland–Greenland elevation. There is no doubt that these troughs are caused by tectonic processes and represent components of the system of disjunctive dislocations framing the continental borders.

The depth of the dismemberment on the Iceland Ridge ranges from 300 to 500 m. At a number of places, the rift zone is crossed with deep troughs which are the morphologic expression of transverse fractures along which the displacement of neighboring rift structures takes place. These troughs include the Lena trough on the boundary of the Arctic basin, a system of two troughs in the region of Jan Myan Island, and a less clearly expressed trough at 69°N. The present tectonic activity of the rift zone is indicated by the epicenters of numerous near-surface earthquakes concentrated exactly along the rift valleys and transverse troughs (Heezen et al. 1961; Sykes 1965). Calculations of the stresses in the earthquake focus indicate the presence of stresses directed transversely to the trend of the rift zone (Mucharina 1967).

A large submarine volcanic massif is seen in the region of Jan Myan Island, which has a wavy or somewhat undulating surface. South of this stretches the Jan Myan Ridge, which has a block-shaped structure, a smooth surface, and steep side slopes. A similar plateau-like structure is found west of Jan Myan Island, together with massive, flat-topped mountains. Probably all these forms of relief were caused by the development of covers of Tertiary plateau-basalts in much the way as the basalt plateaus were formed on both sides of the rift zone in Iceland (Muratov 1961). Judging from its structure, the Iceland plateau is also composed of covers of basalts, through an older series (Johnson et al. 1967). It is highly probable that in the Tertiary period a hugh basalt plateau existed between Greenland, Jan Myan, and Scotland*, which was later fractured into a series of blocks, part of which was submerged

* Tor H. Nilsen, "Lower Tertiary laterine on the Iceland–Faeroe Ridge and the Thulean land bridge. *Nature,* Vol. 274, August, 1978.

in the ocean. As the mid-oceanic ridge developed toward the end of the Tertiary and during the Quaternary, the Iceland–Jan Myan region was drawn into its sphere, and the rift zone made its way across the plateau in the form of the Central Graben of Iceland and the Iceland Ridge.

In Figures 98, 99, and 100 the recent seismic activity or active tectonic life of this area is shown. This history involves block tectonic processes with displacement as an element of the plate movements. These secondary displacement processes under oceanic conditions produce the mosaic of plateau-basalt blocks as shown in Figure 101 (V. Dibner et al. 1965). The general correlation scheme of neotectonic and glacio-isostatic movements after N. Nikolaev (1967) made on the materials from the Scandinavian ice sheet area is shown in Figure 102.

The simple model (Figure 103) shows the process stages and possible result of the glacio-isostatic movements in the briefly described special conditions of the geological structure in the active barrier zone of the North Atlantic.

At present the maximal water depth in the North Atlantic barrier area is more than 500 m. Note that the depth between Greenland (now glaciated) and Iceland is less—about 300 m. Using a simple formula (see below) we can calculate the magnitude of submergence of the blocks under ice and the compensated emergence of the blocks between them. One method of preliminary simplified computation is shown in Figure 104. The isostatic uplift of the unglaciated block situated between ice-loaded blocks can reach 250 m. Glacio-eustatic lowering of the World Ocean at that time might reach 150 m. In this case the water depth in the North Atlantic barrier area was only about 100 m, and in this condition the ice spread across the uplifted area. The result could be the following: the raised block plus the shelf glaciation created the dam and isolated the Arctic Ocean from the Atlantic.

This method of preliminary simplified calculation is based on thickness data for the continental and oceanic blocks of the earth's core in the North Atlantic region for each block: Scotland, Faeroes, Iceland and so on, involving density of the core (δ), density of ice (δ_i), and the sizes of blocks (thickness–B, areas–S). The preliminary equation looks like this:

$$B_1 S_1 \rho_1 + B_i S_1 \rho_i = B_2 S_2 \rho_2 + hS_2 \rho_2 \qquad (10)$$

$$h = \frac{B_1 S_1 \rho_1 + B_i S_1 \rho_i - B_2 S_2 \rho_2}{S_2 \rho_2}$$

Calculation shows that the uplift for the block between Scotland and the Faeroes was about 0.5 km (h^1), and for the block between the Faeroes and Iceland it was about 0.3 km; this is sufficient for dam creation. Of course, all the calculations only show some way and point out one of the ways for a possible model development.

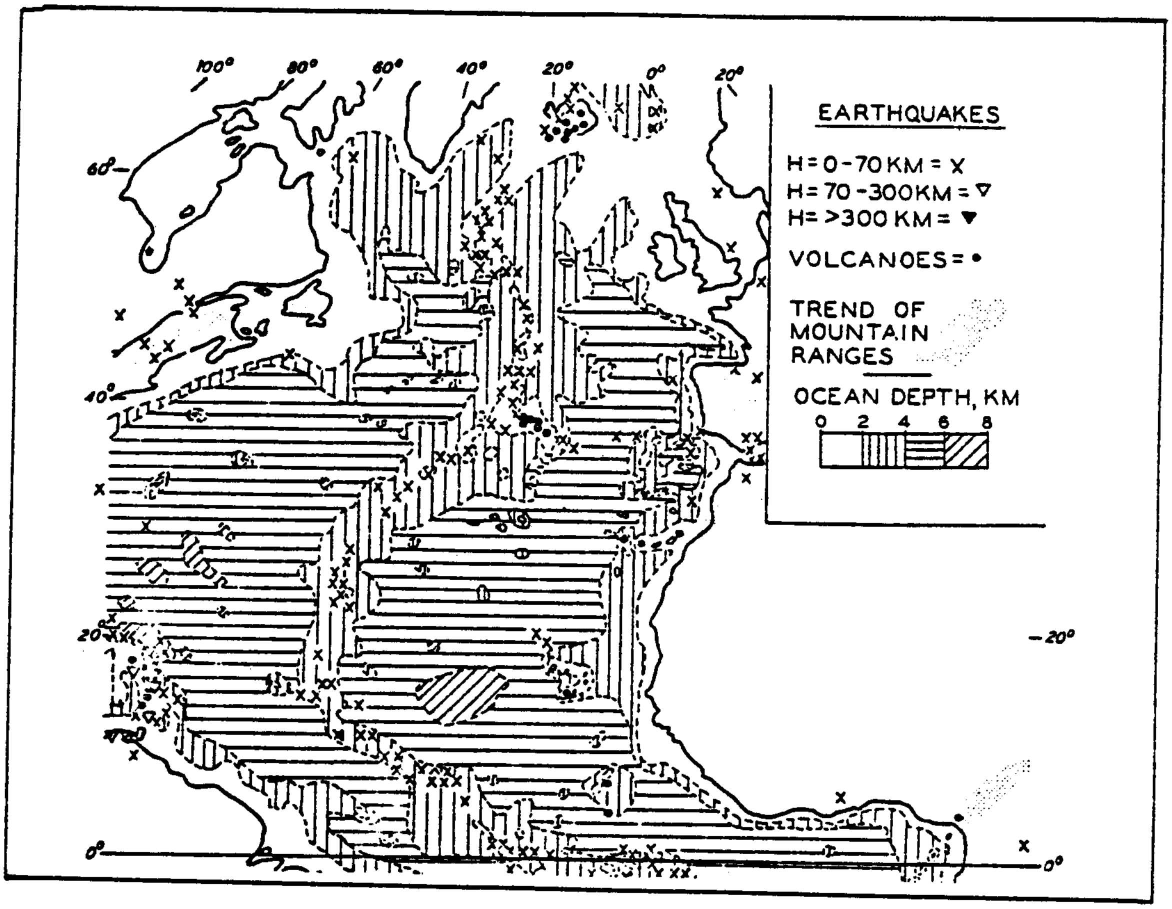


Figure 99.—Representative seismicity map (Cutheberg and Riechter 1954).

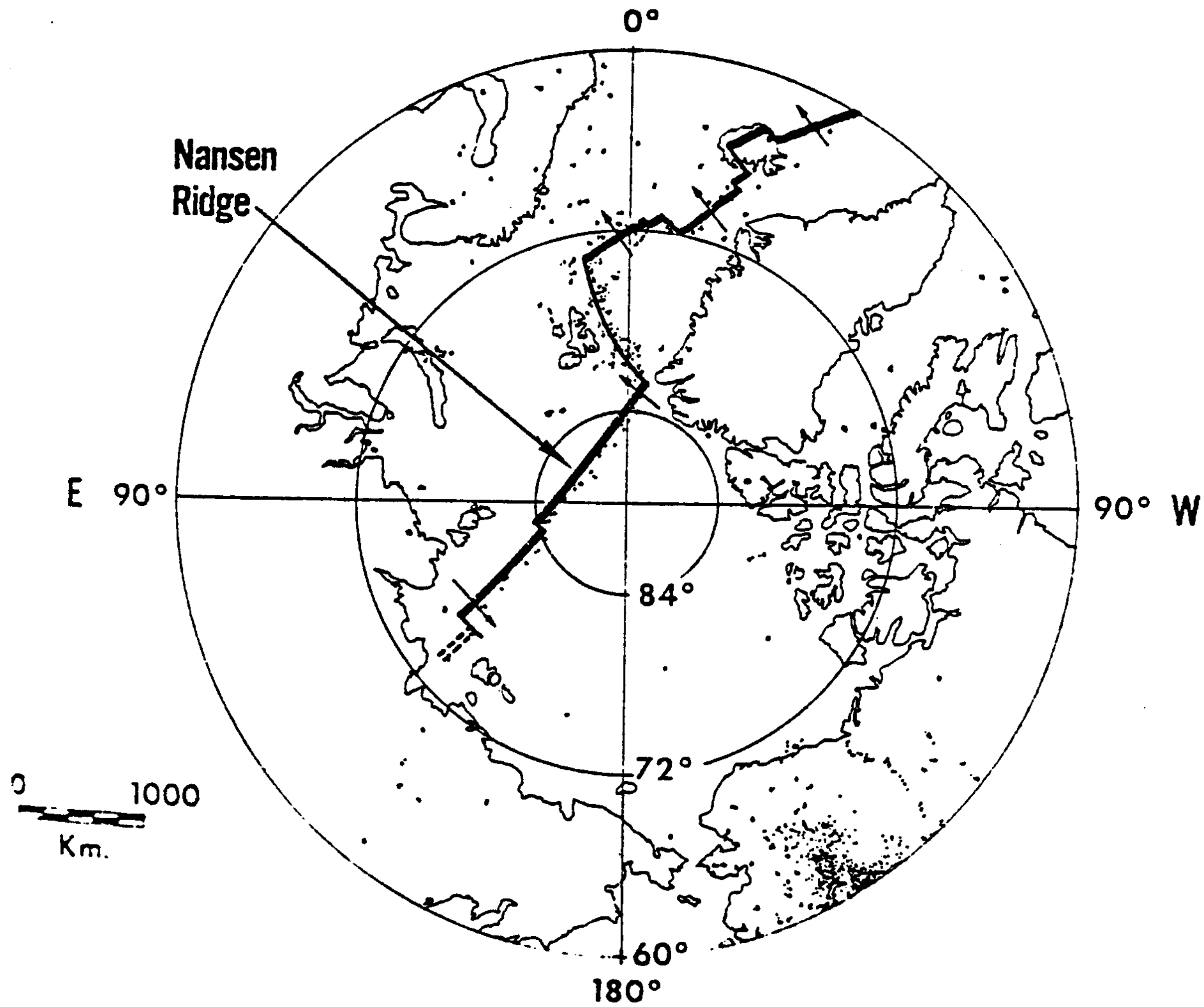


Figure 100.—Postulated rotation of Nahsen ridge. Epicenters marking plate boundaries. ESSA, CGS data, 1961–1969.

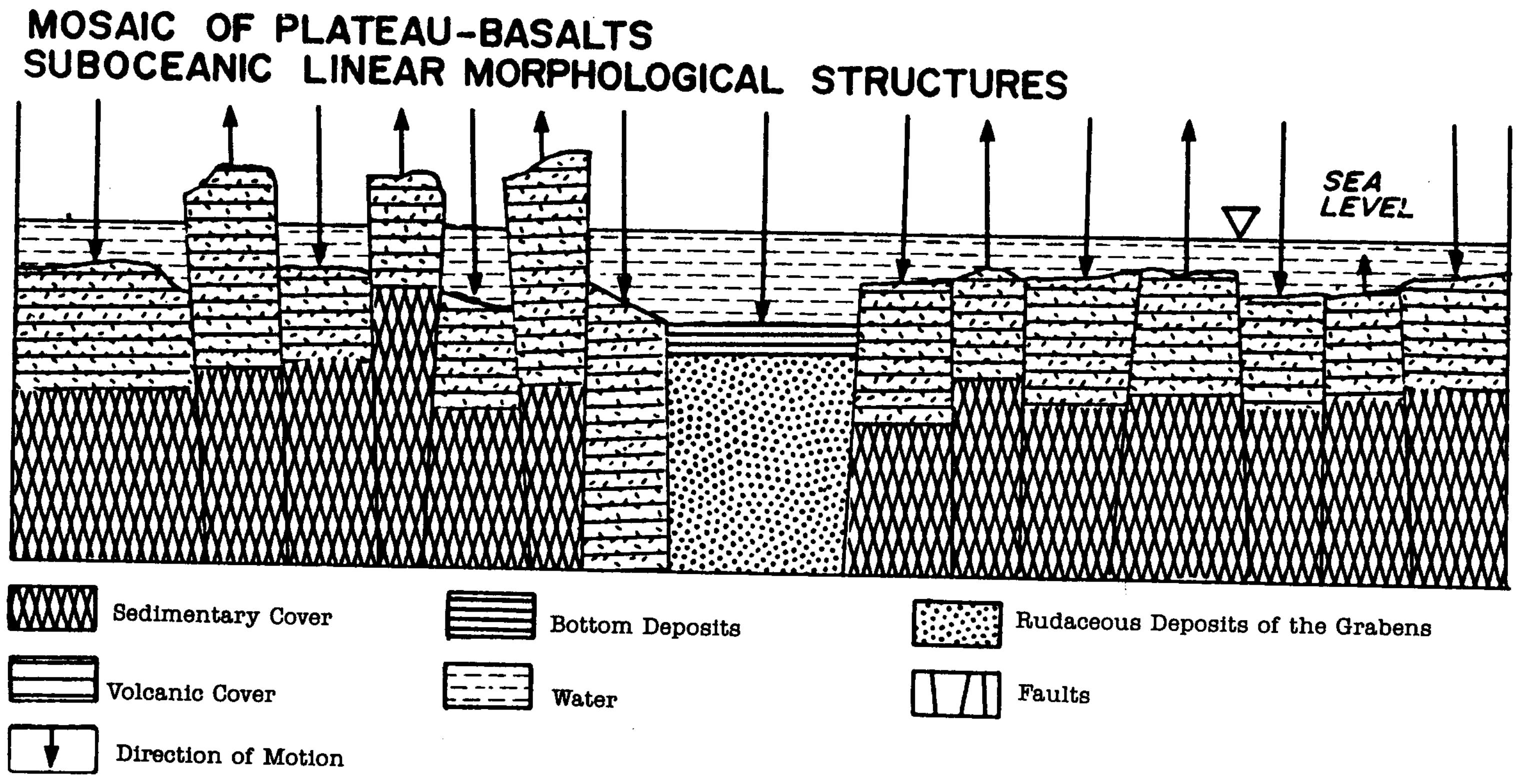
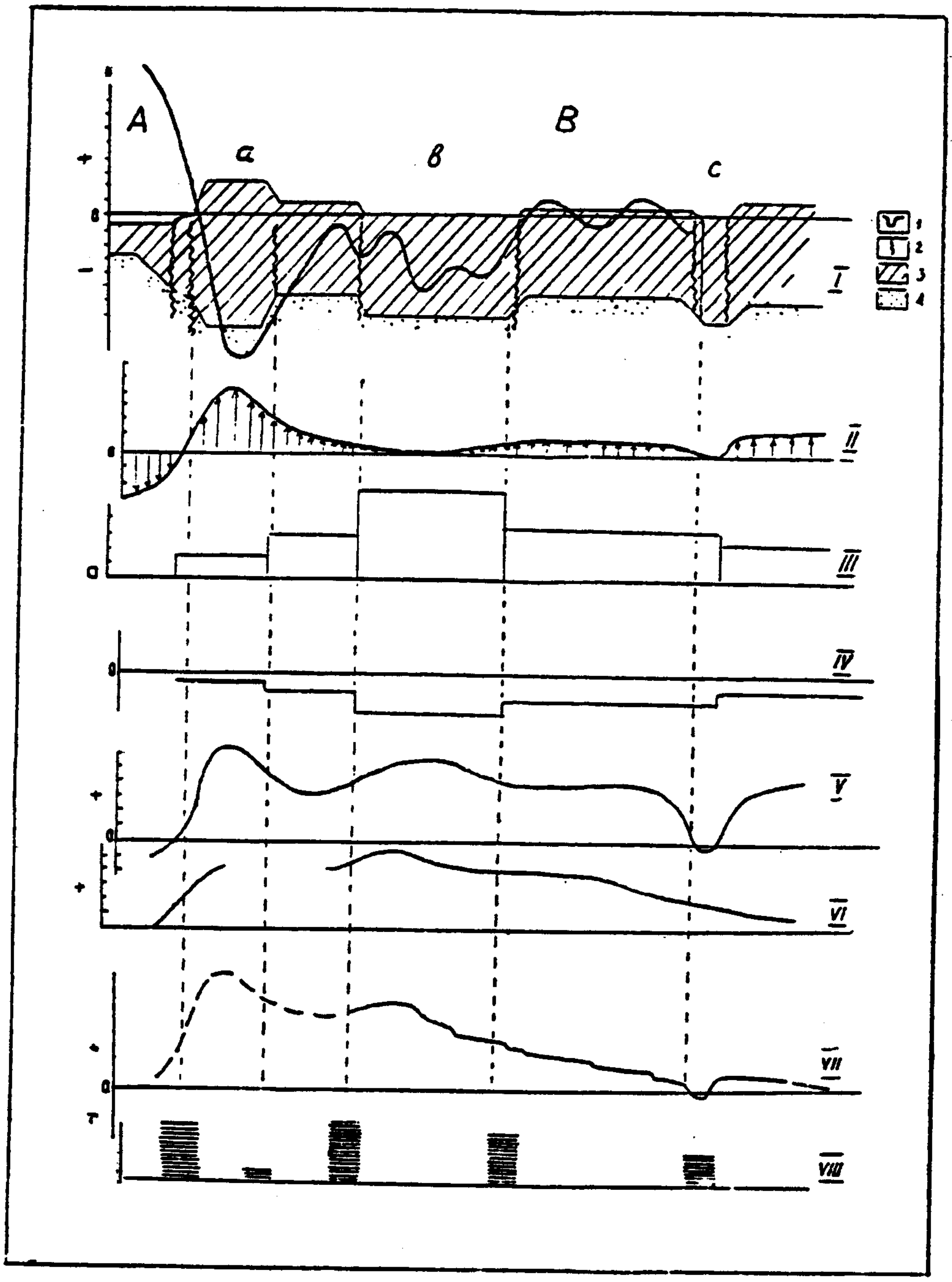


Figure 101.—Block tectonics of plateau basalts (Dibner 1965).



A. Atlantic Ocean
B. Baltic Sea
a. Epiplatform, Scandinavian Mountains
b. Bothnia-Kandalakshan depression
c. Graben
I. Thickness of Earth's core
 1. Gravity anomaly curve (in Bouguer reduction)
 2. Main regional fault zones
 3. Core
 4. Mantle

II. Diagram of summary neotectonic movements
III. Diagram of ice sheet loads
IV. Diagram of glacio-isostatic movements
V. Curve of Late and Post-Glacial movements with interaction of neotectonic and glacio-isostatic compounds
VI. Curve of Late and Post-Glacial movements after A. Högbom
VII. Curve of recent movements
VIII. Diagram of seismicity level

Figure 102.—The general correlation scheme of neotectonic and glacio-isostatic movements for the Scandinavian ice sheet area. After Nikolaev (1967).

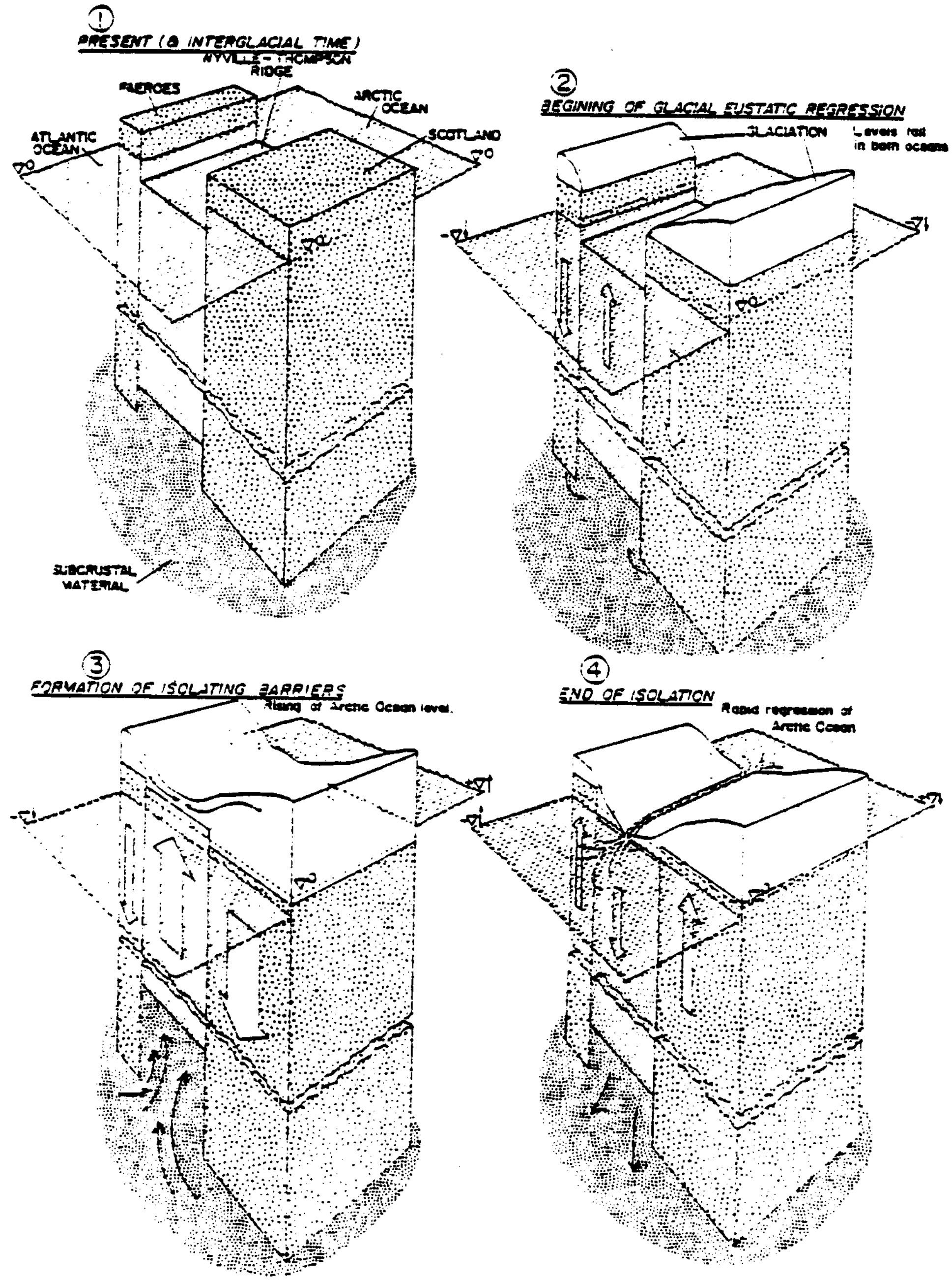


Figure 103.—Isostatic, eustatic, and glacial barrier between Greenland, Iceland, Faeroe Islands, and Scotland during maximal glaciation.

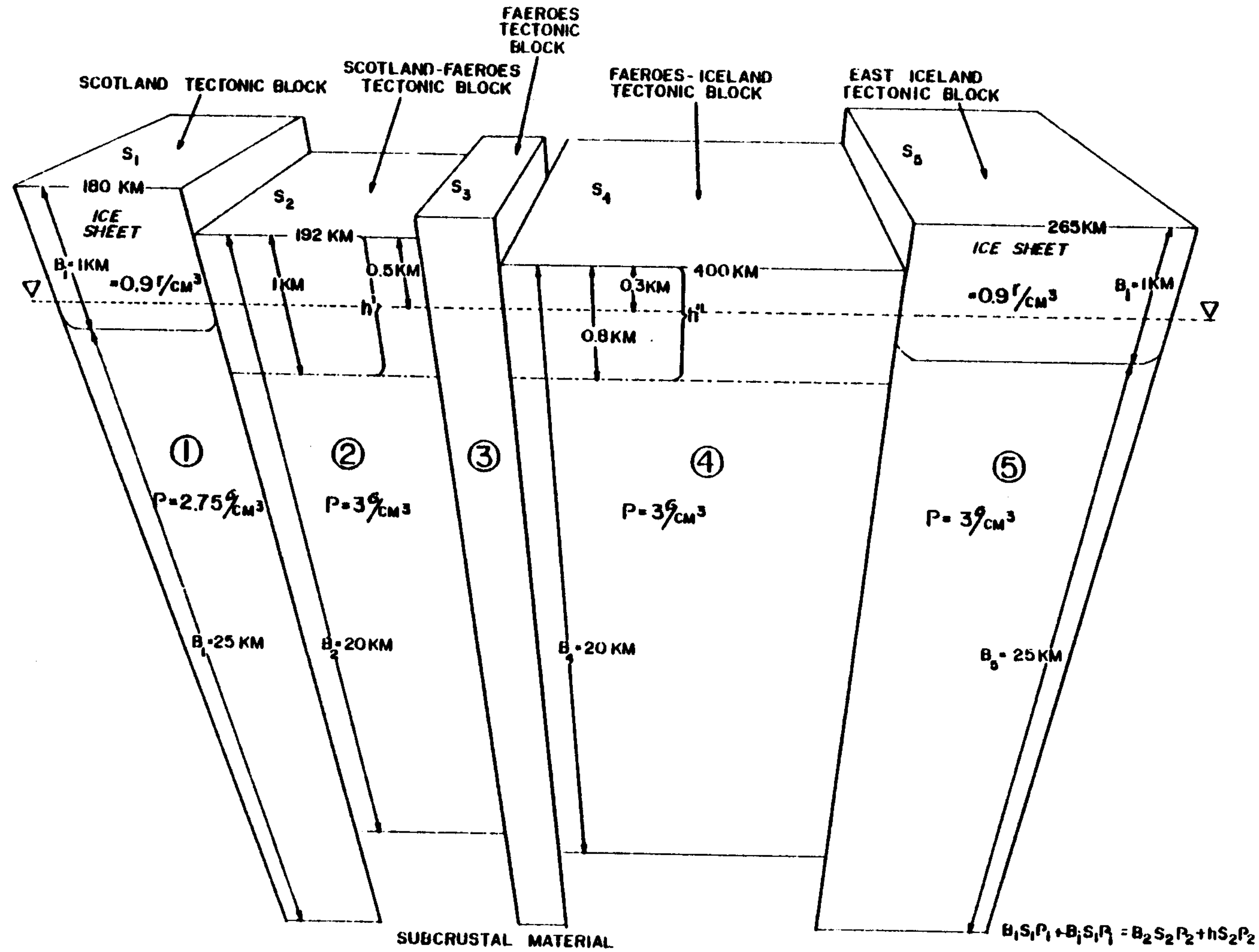


Figure 104.—Preliminary calculation of block glacio-isostatic uplift in the North Atlantic barrier region.

Conceptual Model of Changes in Arctic Ocean Level during the Glacial–Interglacial Cycle

We see that it is possible to consider the formation of the isolating barriers in the northern Atlantic (Greenland, Iceland, Faeroe Islands, Scotland) and Bering Strait areas as a cumulative result of (a) glacial-eustatic regression; (b) isostatic uplift of peripheral blocks around the glacial shields in a mosaic of plateau-basalt suboceanic structures (Daly 1934; Fairbridge and Newman 1968; Newman and Fairbridge 1971); and (c) glaciation of emerged and semi-emerged areas. In Figure 105 we can see the evidence for asthenospheric motion in this area (Vogt 1971). The possibility for large scale asthenospheric movements during and as a result of the ice load from both sides of the North Atlantic barriers may be seen in Figure 106.

The transgression of the isolated Arctic Ocean could be about 200–250 m higher than the World Ocean. After the beginning of deglaciation further transgression occurred due to ice-melt, and complete isolation of the Arctic basin was ended in the course of deglaciation. The dams finally ceased to be effective and regression of the Arctic Ocean was rapid. Normal glacio-eustatic transgressions were resumed with warm waters spreading from the Atlantic to the Arctic.

The correlations between the dynamics of glaciation and changes in levels of the World Ocean and Arctic Ocean are shown in Figure 107. In the framework of this model, two regression–transgression sedimentation cycles in the Arctic Ocean correspond to one cycle in the World Ocean, for the same time period (Vigdorchik 1973, 1978; Vigdorchik and Vinkovetsky 1976).

River Inflow and Rise of Arctic Ocean Level during Glaciation

The rise in Arctic Ocean level during maximal glaciation (Illinoian time) could be explained by the Arctic basin receiving river water inflow during several thousand years.

Let us see how the data about continental surface run-off into the Arctic Ocean could be helpful for our preliminary calculations of this ocean level rise during glaciation. This modern run-off is 4,380 km^3 per year (Antonov 1958). As Table 21 shows, the amount during glaciation was less than a third of the modern figure because of ice sheet extension in a large part of the Arctic basin; so at that time it was about 1,000 km^3 per year.

We saw before that isolation had only to take place at the end of glaciation, occupying about 3,000–4,000 years. The area of the Arctic Ocean is 13.1 million km^2. This means that rising of level $H = (1{,}000\ km^3 \times 3{,}500\ yr)/13.1 \times 10^6\ km^2 = 270$ m. But this figure could not have been reached because of the "Turgai Strait," the big valley to the south of western Siberia; the "strait" that connected the Siberian basin with the Aral and Caspian sea basins. This strait transferred the surplus ocean water entering the Siberian plains to the Caspian pluvial sea. The terrace from the maximal glacial age in this strait is about 180–190 m above

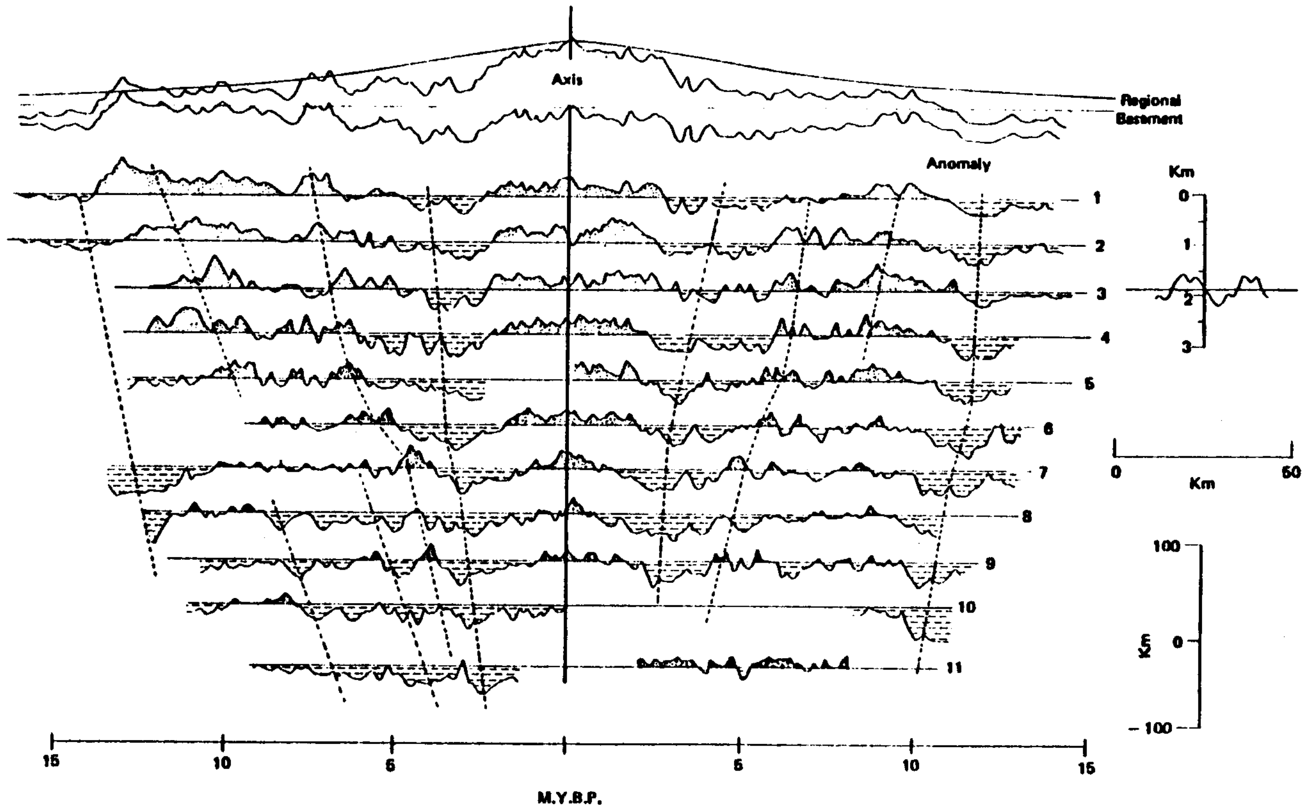


Figure 105.—Evidence for asthenosphere motion. After Vogt (1971). When the normal regional subsidence curve (Sclater et al. 1971) is removed from basement profiles across the Reykjanes Ridge (Talwani et al. 1971), time-transgressive trends under the Reykjanes Ridge become apparent. Implied mantle flow rates are about 20 cm/yr (Vogt 1971). North is at the top, and isochrons parallel the axis. Time-transgressive ''events'' are dashed. The absolute level of the basement is at about –1.8 km, which is above the norm of Sclater et al. (1971).

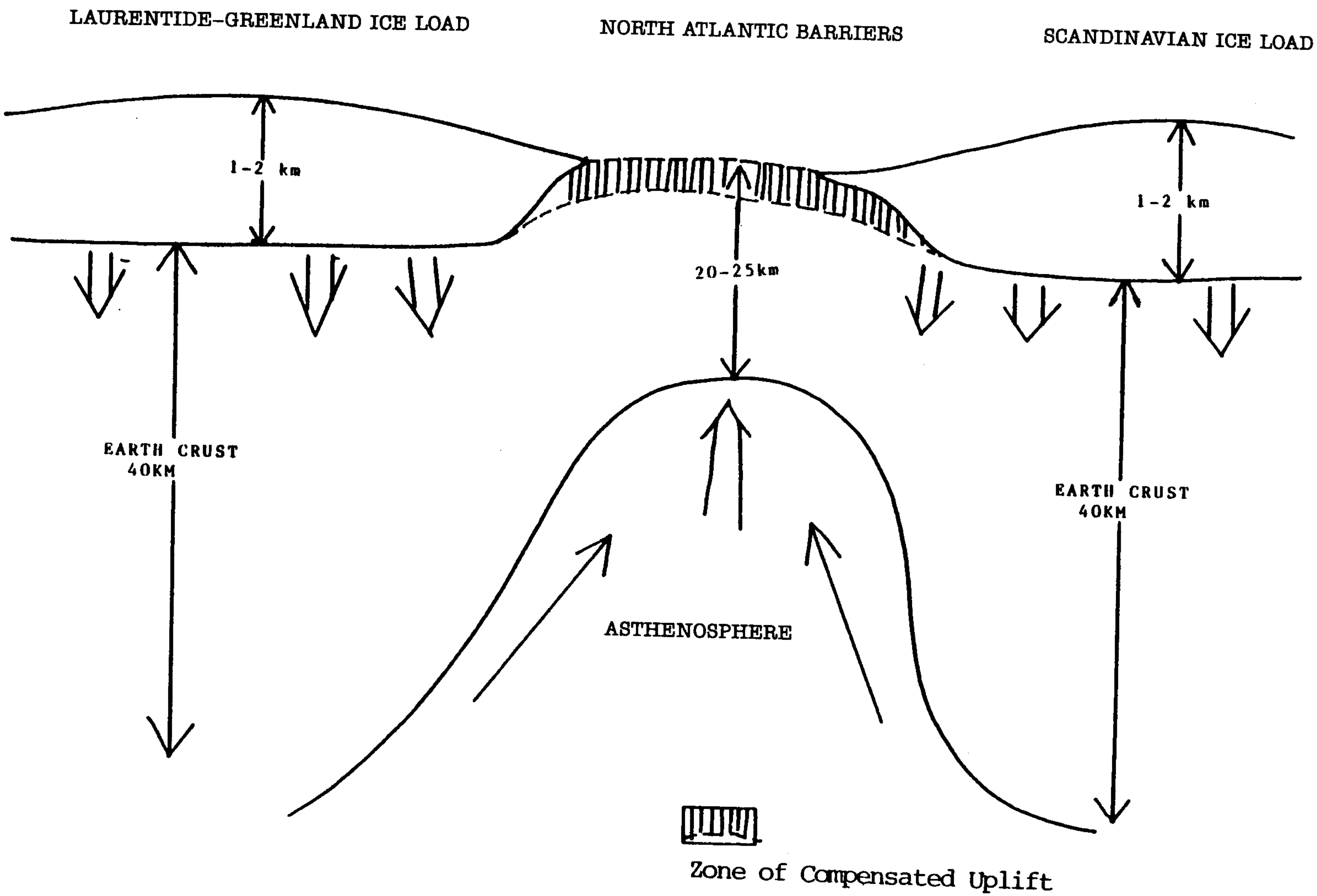
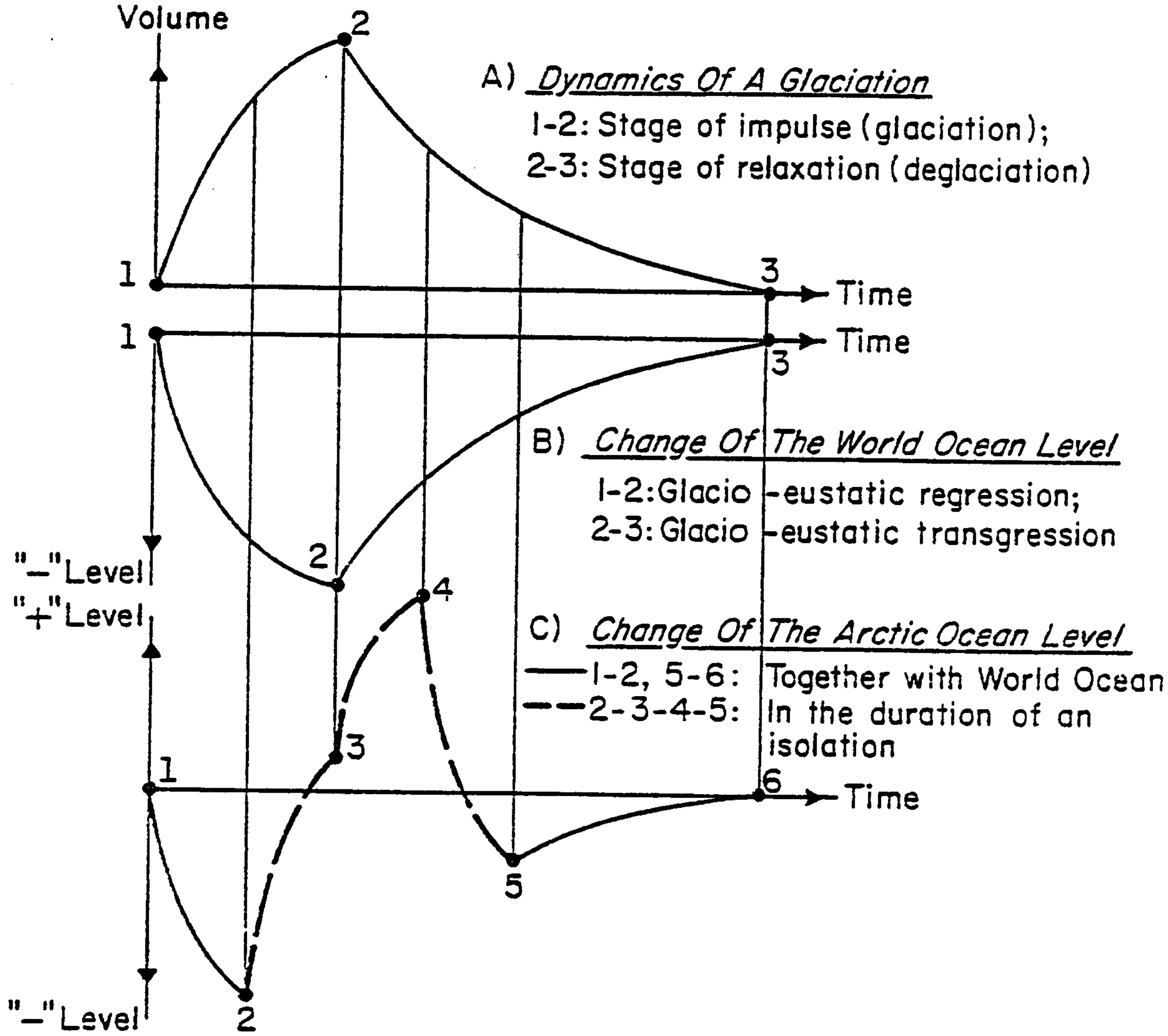


Figure 106.—Large scale asthenosphere movements in the North Atlantic barriers during the glaciations.



1: Beginning of glaciation, glacio-eustatic regression and glacio-isostatic movement.
1–2: Glacio-eustatic regression together with World Ocean.
2: Formation of isolating barriers (Greenland–Iceland–Faeroe Islands–North Sea shelf; Bering Strait area) as a cumulative result of the (a) regression; (b) isostatic uplift of a peripheral bulge; (c) glaciation of emerged and semi-emerged areas.
2–3: Transgression of isolated Arctic Ocean because of river water inflow.
3: Maximum glaciation, and the beginning of deglaciation.
3–4: The more intensive transgression, with the addition of ice-melting waters.
4: The end of a complete isolation in course of deglaciation; the barrier and Arctic Ocean levels are equal, and then lower together.
4–5: Rapid regression of Arctic Ocean.
5: Barriers finally ceased to be effective.
5–6: Normal glacio-eustatic transgression up to full relaxation of a glaciation impulse.

Figure 107.—Correlations between the dynamics of glaciation and changes in levels of the World Ocean and the isolated Arctic Ocean. After Vinkovetsky and Vigdorchik (1971).

sea level (Figures 58 and 108). We could suppose that this figure is the controlling one for the Arctic Ocean level rise during the maximal glaciation. The transgression of the Caspian Sea during that time, made possible by the Arctic water surplus inflow, was shown in Figure 108. Thus, in conclusion we see that the figure for the Arctic Ocean rise would be about 150–180 m in comparison with the modern one.

Some geologists in the Soviet Union estimate that the levels of the cold marine transgressions reached 240 m and more during the maximal glaciation (Suzdalsky 1971; W. Zagorskaya 1972; and others). These scientists think that cold marine transgressions took place during each glaciation, including the last one (Valdai, Wisconsin). Let us emphasize, however, that in our opinion, the data concerning "cold marine transgressions" are persuasive only for the time of the glaciations before and during the maximal (Illinoian) glaciation.

From the investigation of the composition and microfauna of the Arctic Ocean bottom* sediments up to a depth of 4 m, Belov and Lapina (1960, 1973) were able to define a number of horizons formed during warming trends in the Arctic, and a number of other horizons formed during sudden cooling trends associated with the glacial periods. Horizons formed during the warming trends are characterized by the presence of microfauna of the North Atlantic type and by increased concentration of calcium carbonate, iron, manganese, and organic matter. These sediments are also finer grained. The horizons formed during the cooling trends are characterized by the absence of microfauna of the North Atlantic type, and a lower percentage composition of all other components (Fe_2O_3, MnO, $CaCO_3$). Thus, the analysis of the character of the longer bottom cores has shown that variations with depth are related to alternation of warming and cooling trends in the Northern Hemisphere. Not only climatic but also hydrological changes and sea level fluctuations are reflected in the sedimentary record.

Data have shown that during the cooling trend the total influx of warm Atlantic waters into the Arctic basin decreased. The possibility that the connection between the Atlantic Ocean and the Arctic basin was broken at various times is not excluded. During the warming trend the connection between the Atlantic Ocean and the Arctic basin was re-established and Atlantic waters flowed into the Arctic basin over a wide front, just as they do at the present time.

From the structure of the sediments it was possible to calculate their rate of accumulation and absolute age, and to reconstruct in a general way the geological history of the Arctic Ocean during that time. Briefly, the history of the Arctic Ocean may be recapitulated in the following manner. The longest bottom cores have uncovered the tops of mid-Quaternary deposits, which

* Teplovaia melioratsia severnyeh Shirot, 1973, Moskva, Nauka.

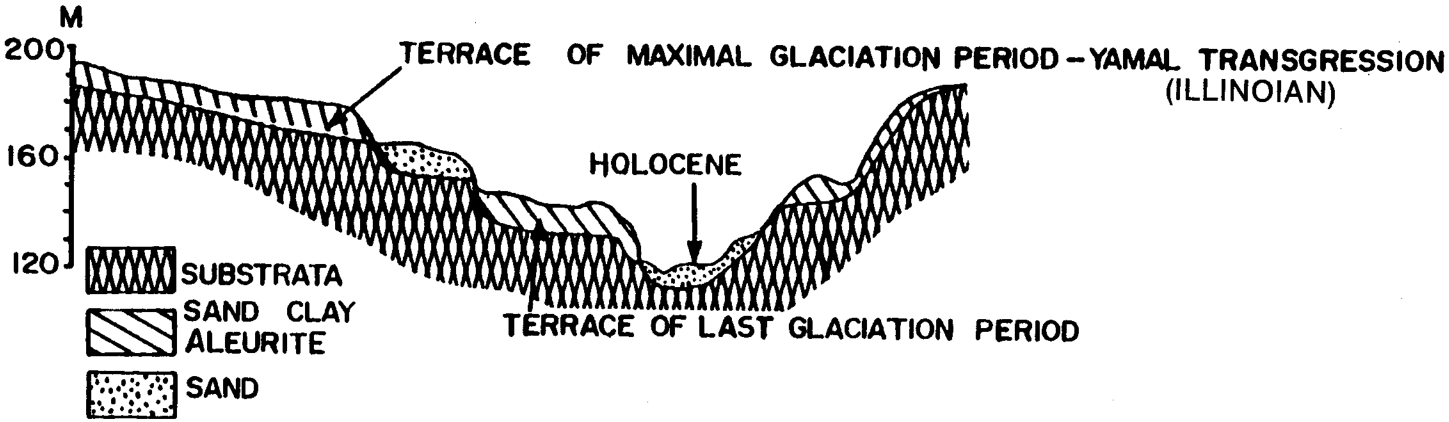


Figure 108.—Geological cross section through Turgai Valley (Boboedova 1975).

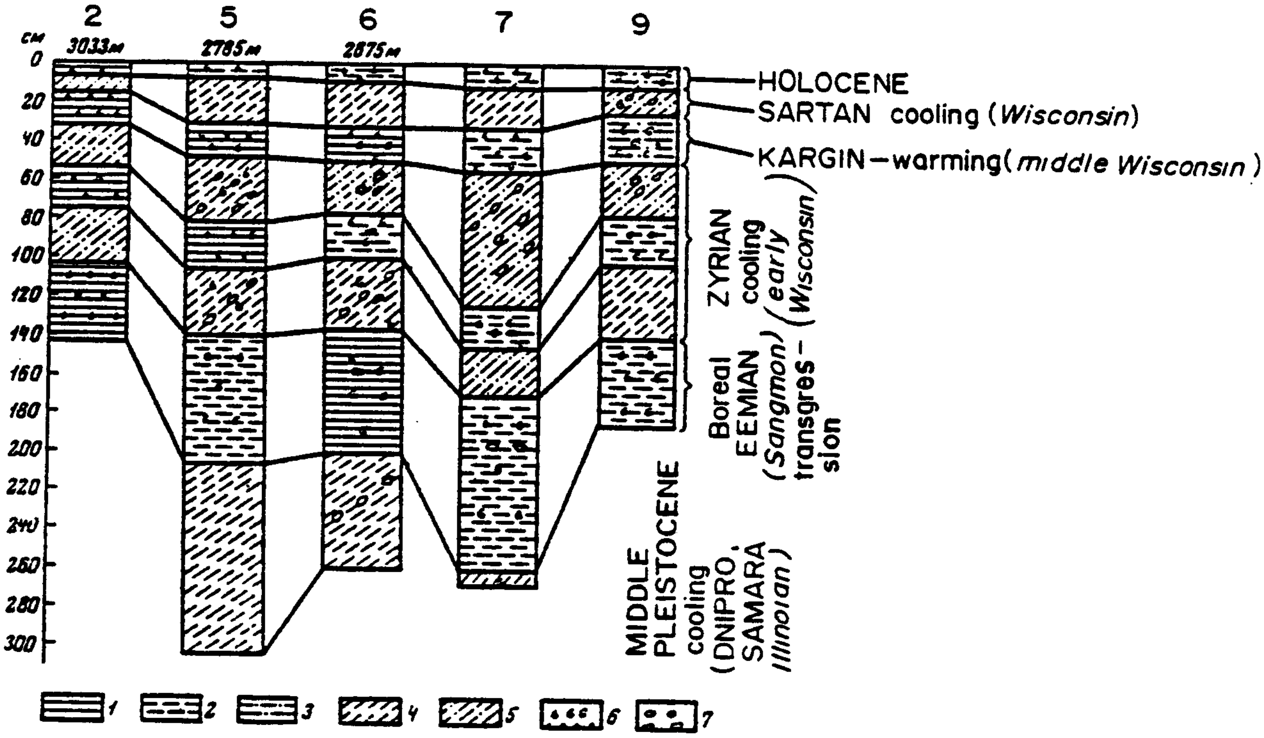
were formed at the close of the maximum glaciation in Siberia, and are represented by clayey silt with layers of sand and inclusions of gravel and pebbles (Figure 109). The nature of the sediments of this period, which are characterized by a negligible amount of iron, manganese, and calcium carbonate, indicate that the climate was severe. The absence of microfauna of the North Atlantic type is an indication of the abrupt termination of the influx of warm Atlantic waters into the Arctic basin. The total duration of the period was not determined by Belov and Lapina.

The Norwegian Sea and North Atlantic deep-sea sediments are relatively unstudied. A number of workers have dealt with the description of a fossil assemblage retrieved from one to several cores (Saito et al. 1967; Stadum and Ling 1969; Bjorklund and Kellogg 1972). Other workers have studied limited numbers of cores covering only a small portion of the region (Böggild 1907; Holtedahl 1959; Ericson et al. 1964a; Olausson 1972; Schreiber 1967). The most comprehensive of these studies is the work of Ericson et al. (1964a). In 26 cores from the Norwegian and Greenland seas, they noted the predominance of "glacial-marine sediment" and suggested, on the basis of frequencies of foraminiferal species in six cores, that the last ice age ended 11,000 years ago in this region. Kellogg (1973) analyzed 34 surface-sediment samples and six cores spanning the past 150,000 years and made climatic interpretations based on frequency of foraminiferal species. The results seemed to correspond to the possibility of the periodic isolation of the Arctic (Figures 110 and 111).

The latest data on bottom deposits gathered during the eleventh expedition of the ship "Professor Zubov" (N. Kulikov and E. Shkatov [1973] in "Geology of Sea," 3, Leningrad, USSR) show a similar picture (Figure 112). Note the relatively small amount of deposit on the slope of the Faeroe Islands in the Holocene (upper layer of the bottom deposits, core 22). This fact could possibly reflect the compensated uplift of the Faeroe block after the ice unloading. Likewise, the relative thickness of the bottom deposits in the trench between the Faeroes and the Shetland Islands could be explained by compensated submergence of this block (cores 2 and 18).

The data of Ruddiman and McIntyre* (1976) are of special interest for our conclusion about the block mechanism of dam creation in the North Atlantic barrier zone (Figure 113). There are certain "anomalies" in their data. For instance, core V27–110 from Hatton–Rockall basin on Rockall plateau provides the converse proof of the genesis of northeast Atlantic barren zones. Sediments in this core, which was taken from a depth of 1,264 m, were deposited far

* Ruddiman and McIntyre. Northeast Atlantic paleoclimatic changes. GSA Memoir 145, 1976.



1. Dark brown and brown layer mud
2. Brown mud
3. Brown and sandy mud
4. Gray and yellow-gray mud
5. Gray and yellow-gray sandy mud
6. Foraminifer of the North Atlantic types
7. Gravel
8. Number of core and depth

Relief of the floor of the Arctic Basin.

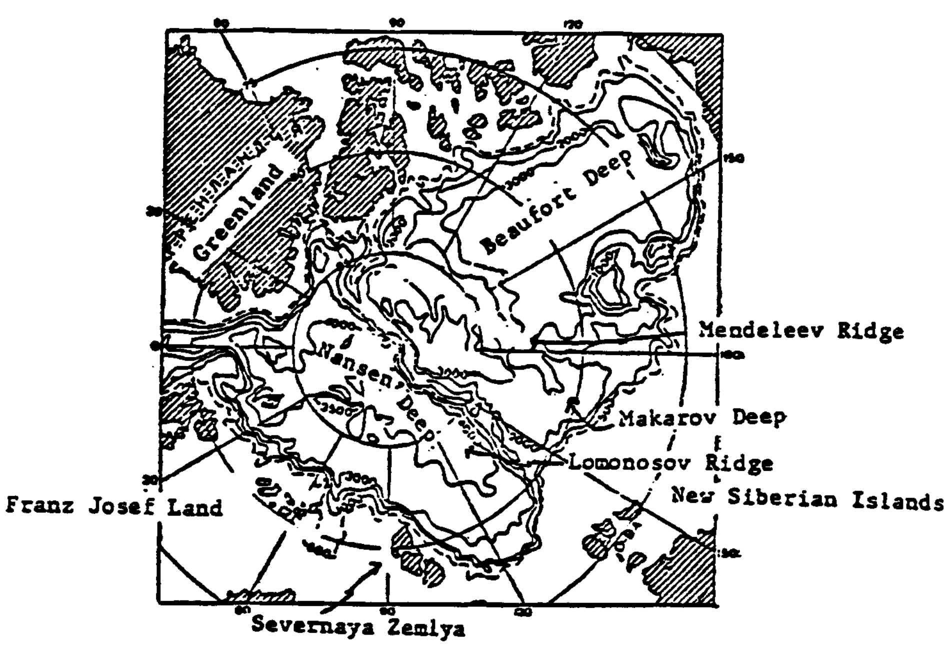


Figure 109.—Bottom deposits of the Arctic Ocean (Belov and Lapina 1973).

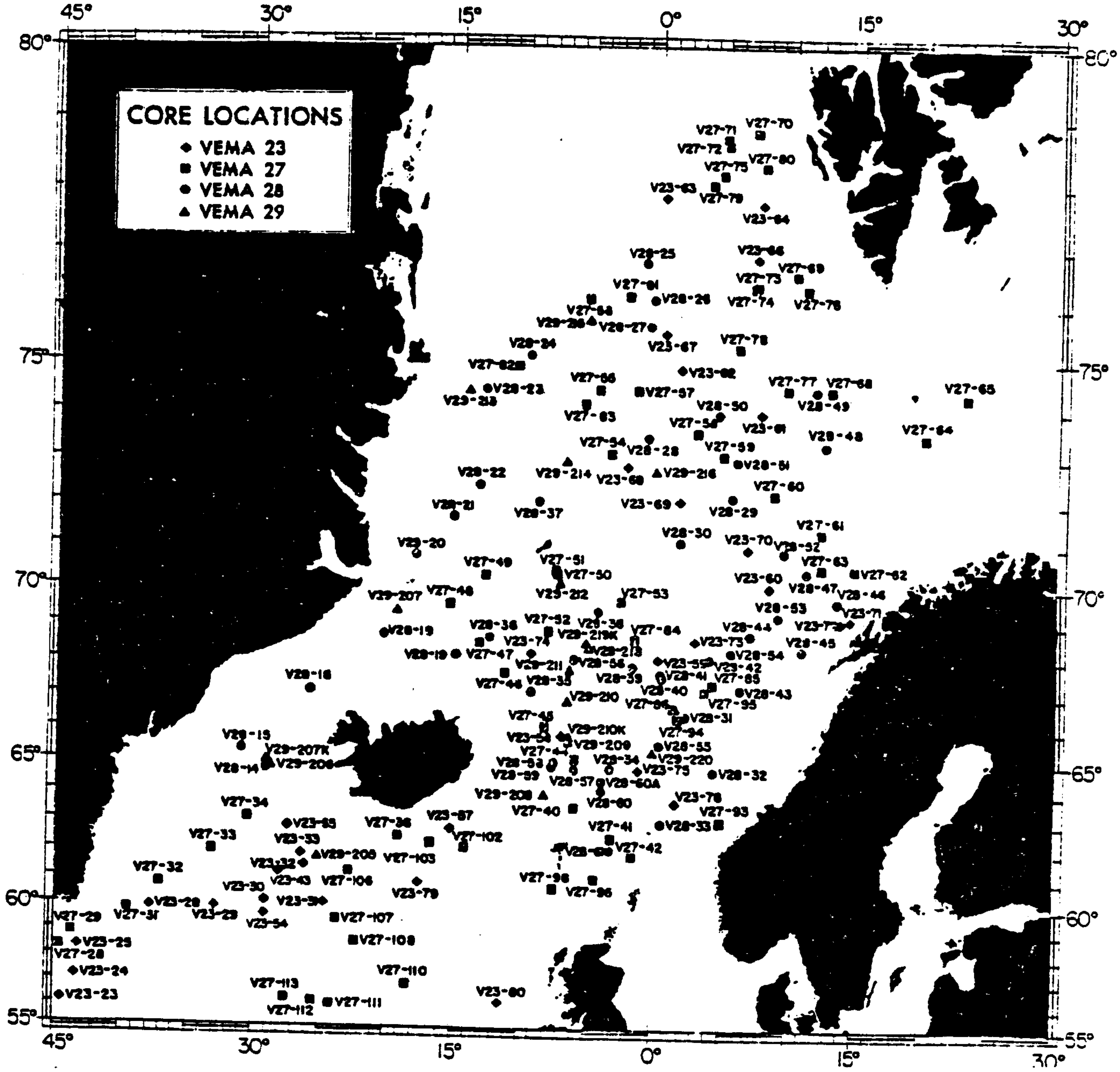


Figure 110.—Locations of cores taken in the Norwegian and Greenland seas by R. V. Vema (Kellogg 1973).

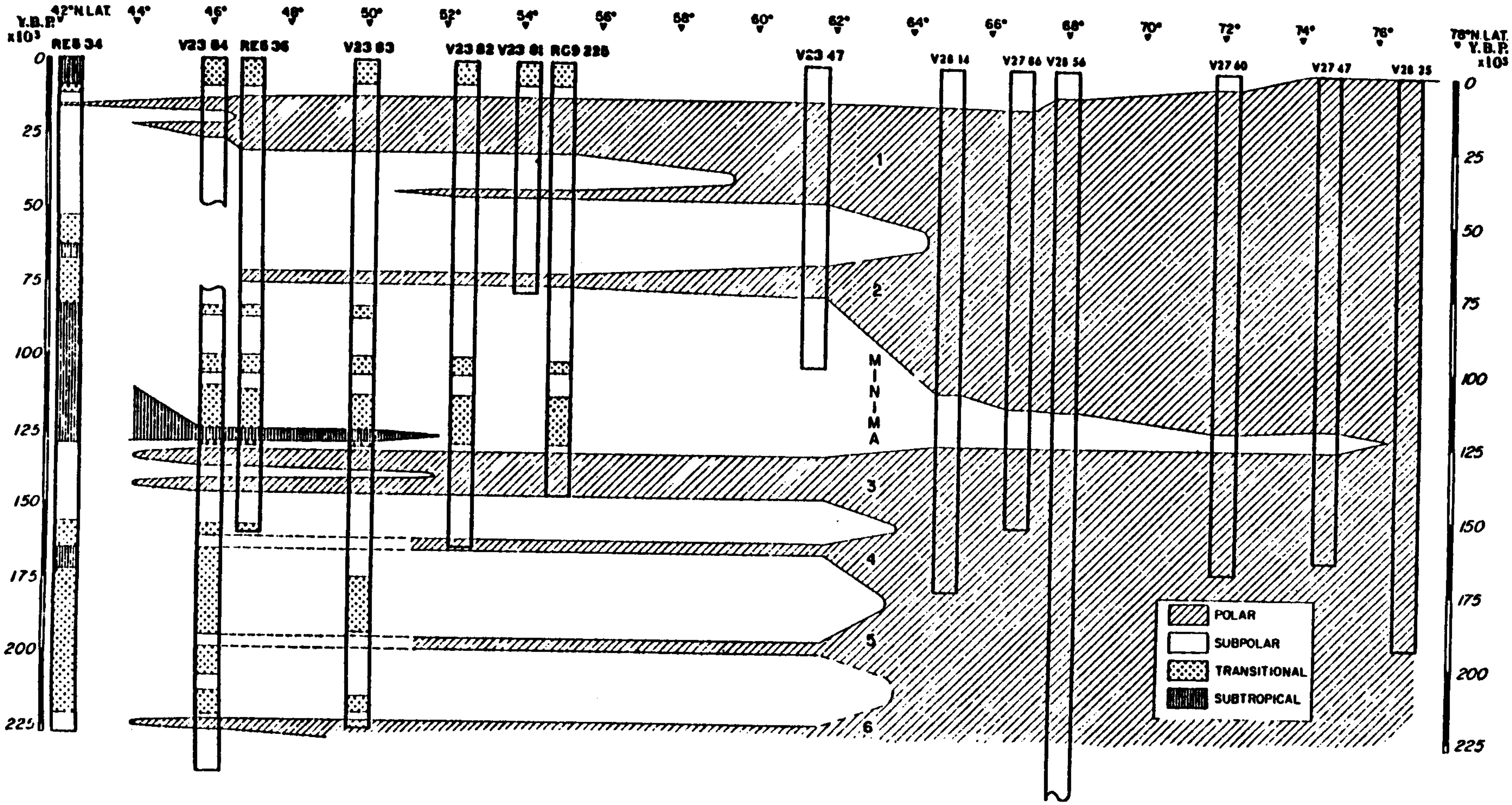


Figure 111.—Variations in the faunal and floral composition in the Norwegian Sea and northern North Atlantic. Climatic zones south of 62°N are based on coccoliths (McIntyre et al. 1972). Note that subpolar faunas have been present in the Norwegian Sea only twice in the last 150,000 years: at present and about 120,000 years BP, corresponding to the Recent and Eemian. After Kellogg (1973).

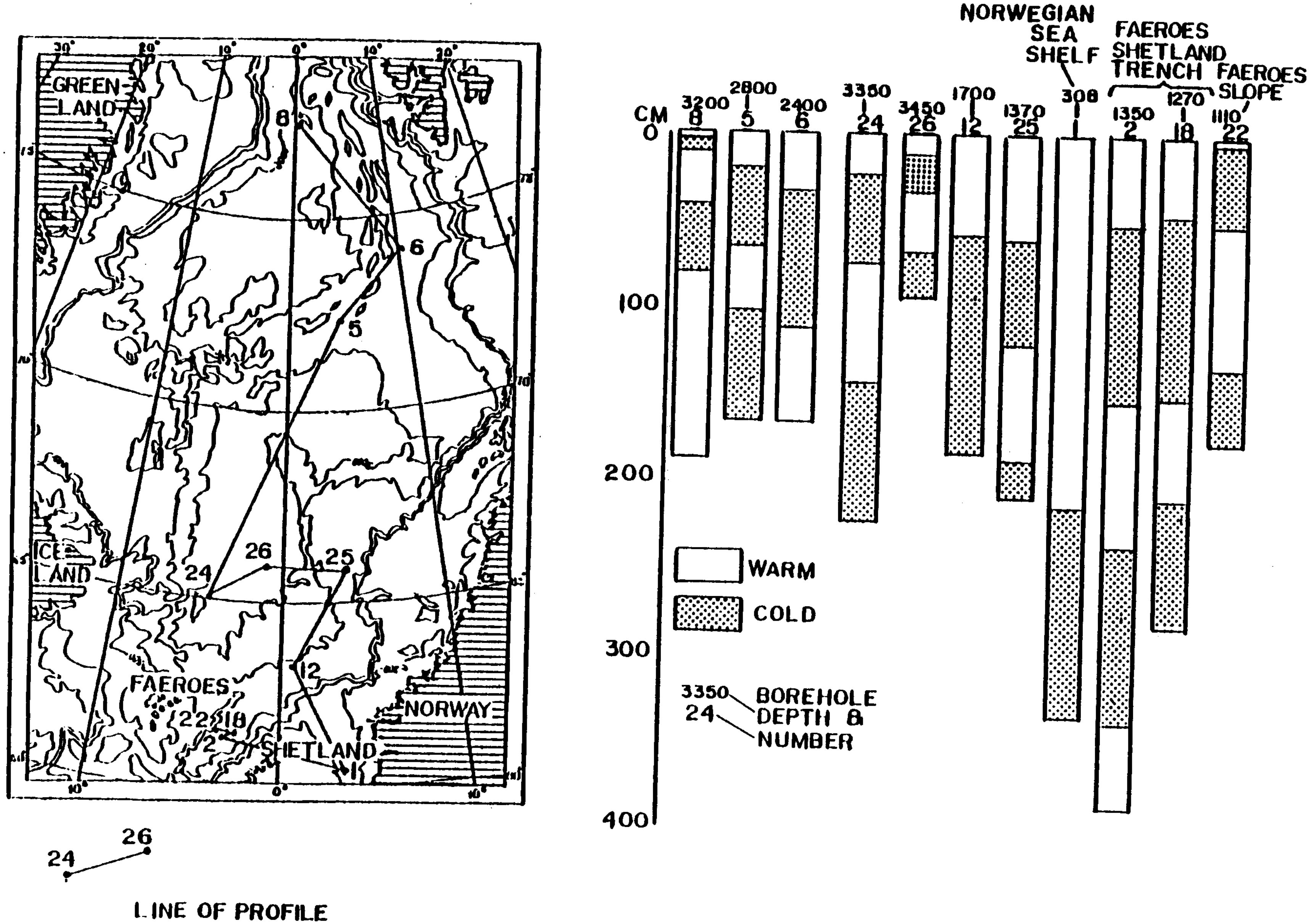


Figure 112.—Bottom deposit cores of Norwegian–Greenland basin (Kulikov and Shkatov 1973).

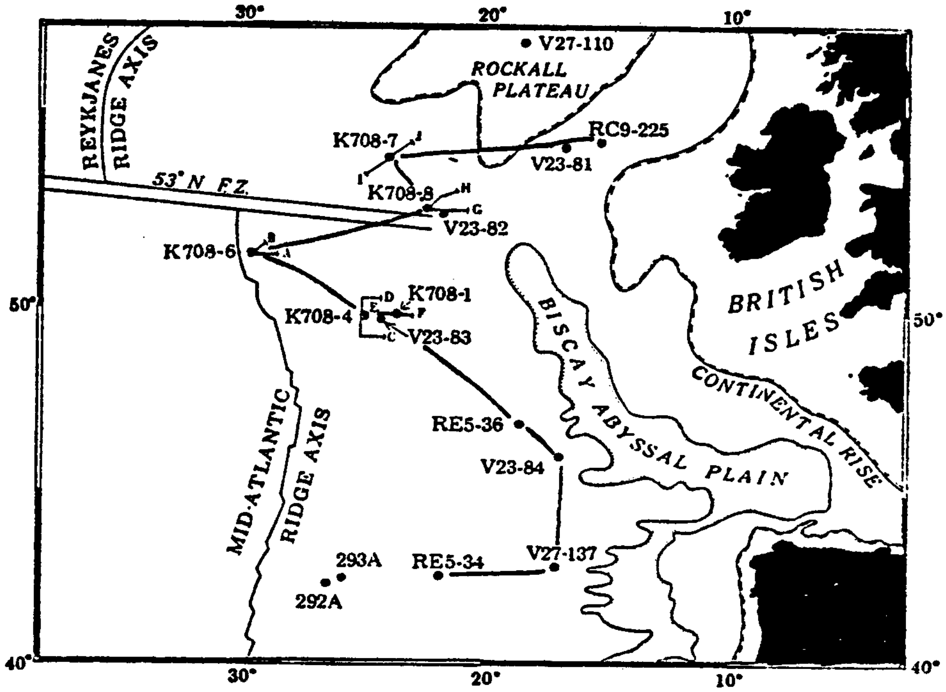


Figure 113.—Locations of northeast Atlantic cores. After Ruddiman and McIntyre (1976).

above the present compensation depth. A total carbonate curve for the section of core through termination II is shown in Figure 114. Even at this shallow depth and reduced sedimentation rate, a coccolith-barren zone occurs between 44 and 49 cm, with thin, almost barren zones at 53, 55, and 59 cm. In addition, the coccolith counts in core V27–110 show no significant increase of solution-resistant species near the barren zones or within the almost barren zones. Both *Coccolithus pelagicus* and *Cyclococcolithus leptoporus* occur in lower percentages in the last glaciation than in the Holocene Epoch or the last interglaciation. In the North Atlantic Ocean today, where solution is known to occur on the bottom beneath subpolar surface water in the Labrador Basin, the core tops contain between 50 and 100% of the resistant *C. pelagicus*. The highest percentage in the glacial part of core V27–110 is only 1.7%. Ruddiman and McIntyre conclude that "barren zones exist in northeast Atlantic depths too shallow for major carbonate solution, and conversely, that faunal and floral indices in and near barren zones even at 4,000 m indicate very little solution. Core intervals that are coccolith barren must represent periods during the Pleistocene glacial maximums when the overlying surface waters were devoid of coccoliths. We thus reiterate that productivity variations are the primary cause of coccolith-barren zones in the Atlantic Ocean north of lat. 40°N." We think that such anomalies as the barren zone on Rockall plateau could be explained as a result of compensated uplift of the unglaciated Rockall block during the glaciation and submergence of the neighbor

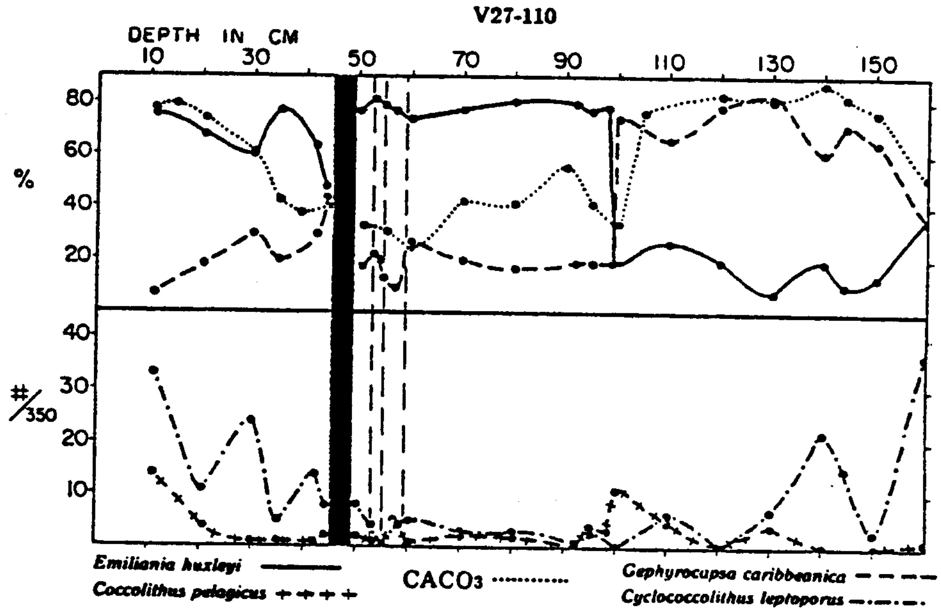


Figure 114.—Comparison of coccolith species. After Ruddiman and McIntyre (1976).

block at the north. It seems to us that the oceanographic and paleoceanographic maps of these authors for the northeast Atlantic Ocean, depicting inferred surface currents and ecologic water masses (Figure 115) give a picture that perfectly corresponds to our idea of the possibility of Arctic Ocean isolation during the glaciations. Isolation of the Arctic Ocean would have had some influence on important aspects of the natural environment including development of permafrost in the Arctic basin and the Northern Hemisphere (Figures 116 and 117). Dam creation would have contributed not to the cooling of the Atlantic Ocean during the glaciations, but on the contrary, the conservation of the ocean water heat. As a result, the warm, wet air and water masses were directed to the coastal areas of Africa and Asia. Let us consider this example from the modern thermal balance of the Arctic Ocean (Table 22). We may suppose that the heat of the Atlantic water (2.5 kcal/cm^2/year) would not be lost by the Atlantic in the case of isolation. We can also add that currents such as Kebot, Labrador, and eastern Greenland would not bring cool water (2.4 kcal/cm^2/year and ice (3 kcal/cm^2/year) to the Atlantic. This means that the Arctic loss of heat was about 2.5 + 2.4 + 3.0 = 7.9 (kcal/cm^2/year). On the contrary, the figure shows the approximate surplus of heat for the Atlantic if the Arctic was isolated. But what was the figure of real cooling influence because of ice spreading in the North Atlantic and Bering barriers area? The consequences of this heat balance

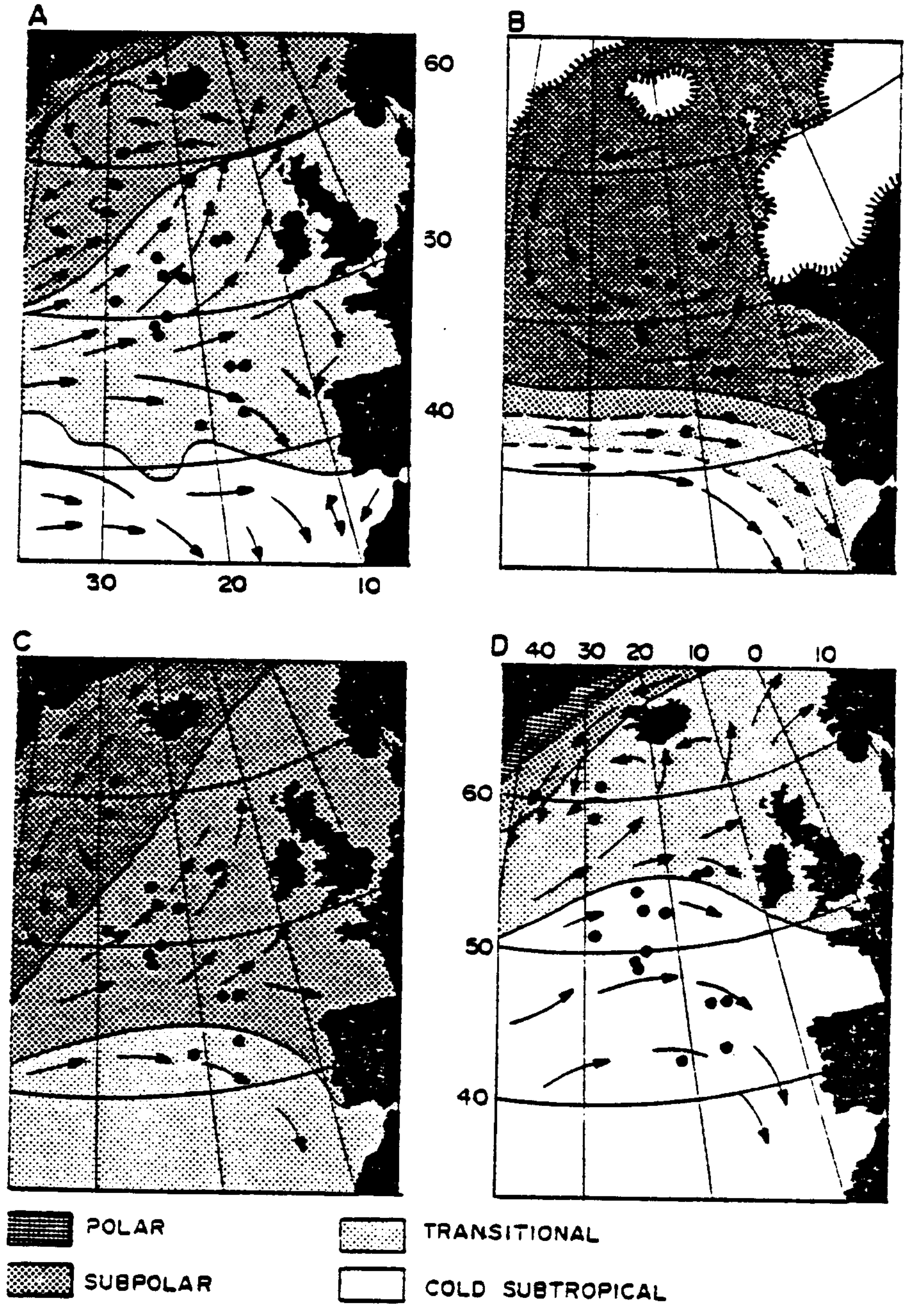
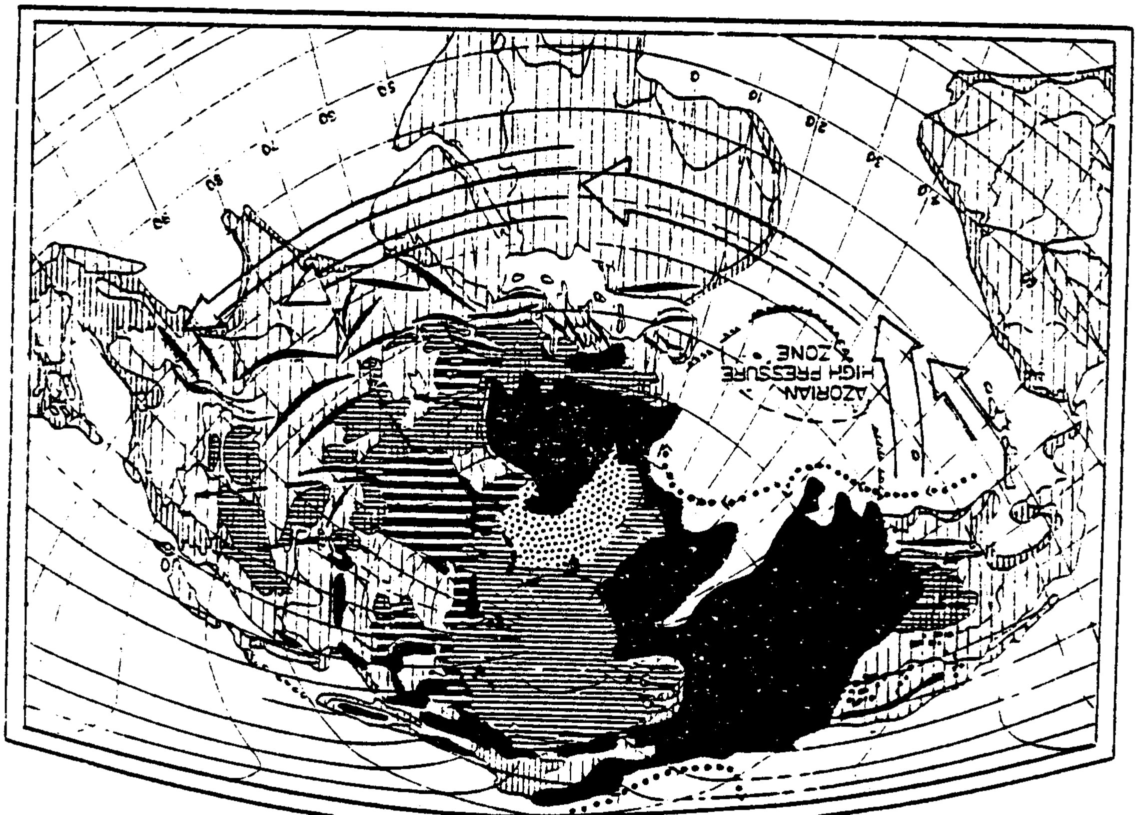
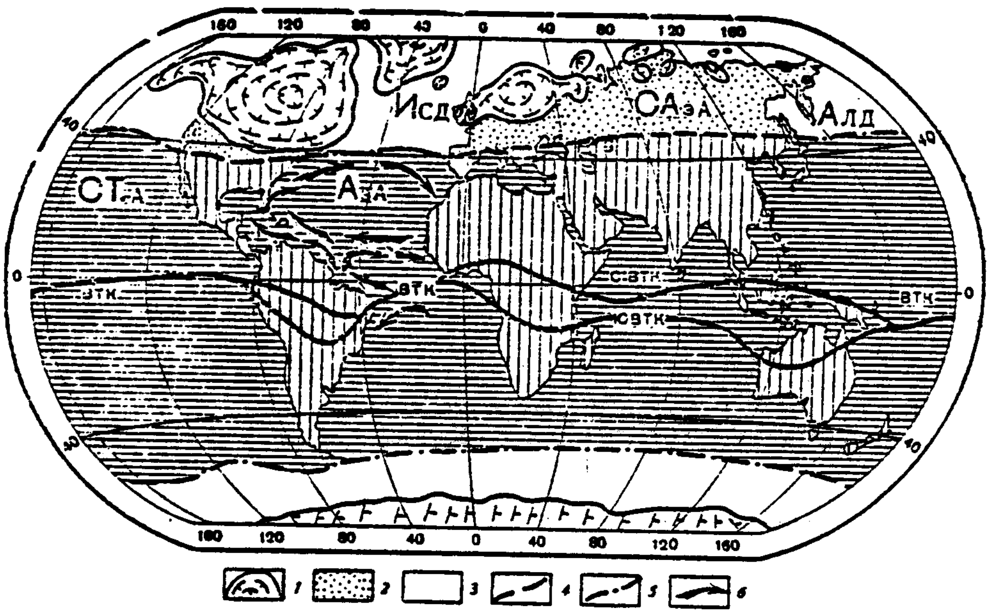


Figure 115.—Oceanographic and paleoceanographic maps of the northeast Atlantic Ocean, depicting inferred surface currents and ecological water masses at the following time levels: A–today; B–maximum glaciation, cycle B, approximately 18,000 BP; C–deglacial part of cycle V, ash zone I, 9,300 BP; D–maximum interglaciation in cycle B, Barbados III, approximately 120,000 BP. After Ruddiman and McIntyre (1976).



1. Areas of glaciation extent
2. World Ocean shelf exposed because of glacio-eustatic regression
3. Loess formation areas
4. Mountain glaciations
5. Pluvial regions
6. World Ocean
7. Isolated Arctic Ocean
8. Territories inundated by Arctic Ocean transgression
9. Boundary of marine ice extent
10. Main direction of winds connected with loess formation
11. Main trajectory of cyclones
12. North Atlantic current direction
13. Area of possible shelf glaciation

Figure 116.—Paleogeographic situation in the Northern Hemisphere during maximal (Illinoian, Dnipro) glaciation. After Vinkovetsky and Vigdorchik (1971).



1.Ice sheet
2.Permafrost
3.Sea ice
4.Boundary of the permafrost
5.Boundary of the sea ice
6.Gulf stream

ИсД-Islandic depression
АлД -Allutian depression
АзА -Worth-Asiatic anticyclon
СТА -North Pacfic anticyclon
АзА -Asorian anticyclon

ВТК-Inner tropical zone of convergence
СВТК-Northern
NВТК-Southern

Figure 117.—Cryogenic area of the earth during the Pleistocene. After Velichko (1977).

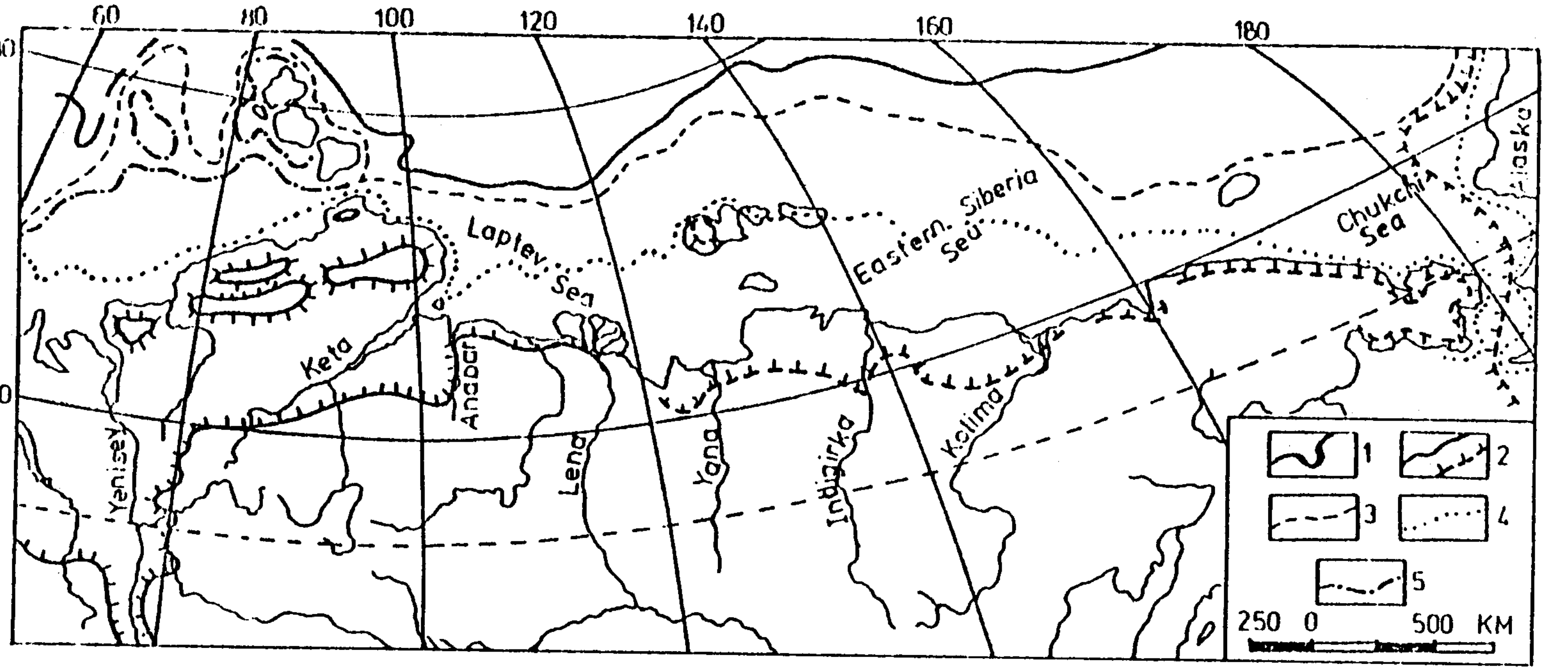
peculiarity are the points of interest for paleoclimatical and paleoceanographic reconstructions, including the questions of temperature and salinity of the Arctic water and permafrost development on the Arctic shelf. The further reconstruction of the atmospheric circulation and calculations of the thermal radiation and ocean water balances would give a picture of natural conditions very different from the existing consensus. Of course, further investigation could help a great deal with the comprehension of a therodynamic model of the natural processes that took place during glaciations on the Arctic shelf in connection with permafrost development.

We tried to show the data and assumptions that characterize the development of the Arctic Ocean during the earlier and middle Pleistocene glaciations (Dnipro, Samar, Illinoian). We saw the possibility of an ocean isolation at that time. If this was the case, it would have interrupted the exchange of Pacific and Atlantic water with Arctic water, and changed the direction and vertical redistribution of the currents, temperatures, and salinity of the Arctic waters. It would also have influenced submarine permafrost development. However, there are data and ideas emphasizing the aspects of an exposure of the Arctic Ocean shelf during the post-Illinoian glaciation (Wisconsin) and its effects on Arctic water conditions and permafrost development on the shelf; here this exposure was caused by glacio-eustatic lowering of levels together with World Ocean levels (an unisolated condition).

ARCTIC SHELF EXPOSURE, GLACIO-ISOSTASY, BLOCK TECTONICS, AND PERMAFROST DEVELOPMENT, AND SPREADING OF THE ARCTIC BASIN

The assumptions of glaciologists who consider the lower position of the Arctic Ocean level as only a result of a glacio-eustatic process are summarized, for instance, by S. Strelkov and here illustrated by Figures 118–120. The map and two curves show the location of the Arctic Ocean shoreline in the Pleistocene according to this scientist's conclusion. The Eurasiatic shelf is shown as being exposed during glaciation including the maximal one (Illinoian). The amplitude of the shoreline oscillations reached 400–450 m in the Kara Sea (minus 250 for Samar glaciation time [Illinoian] and plus 200 for the boreal [Sangamon] transgression. The –250 is related to the Ziryan glaciation (early Wisconsin), and changes of the shoreline position along the Arctic coast (Figure 120) are explained as a result of tectonics.

Figures 121–123 and Table 23 show the data about sea-level-stillstands and fluctuations during the last glaciation and the Holocene in the Laptev Sea according to Soviet and American data gathered and analyzed by M. L. Holmes and I. S. Creager (1974). Their proposed sea level curve of the Laptev Sea for the period 17,500–13,500 BP looks similar to some other World Ocean curves, but a radiocarbon date for core 137 (depth 103–117 cm) does not correspond to this curve and, together with cores 143 and 144, shows a lowering of the level after the last glaciation maximum (about 18,000 BP) The authors explain this fact by the contamination of core 137 by inactive carbon.



1. Position of the Arctic Ocean shoreline during the maximum glaciation (Illinoian)
2. Established (a) and postulated (b) position of the shoreline during the boreal transgression (Sangamon)
3. Position of the shoreline before and during the Zyranskoye glaciation (early Wisconsin)
4. Position of shoreline during the Karginskoye period (mid-Wisconsin interstade)
5. Position of the Kara Sea shoreline during the Sartanskoye period (Wisconsin)

Figure 118.—Map of the location of the shoreline of Arctic seas at individual times during the Quaternary period (S. Strelkov 1965).

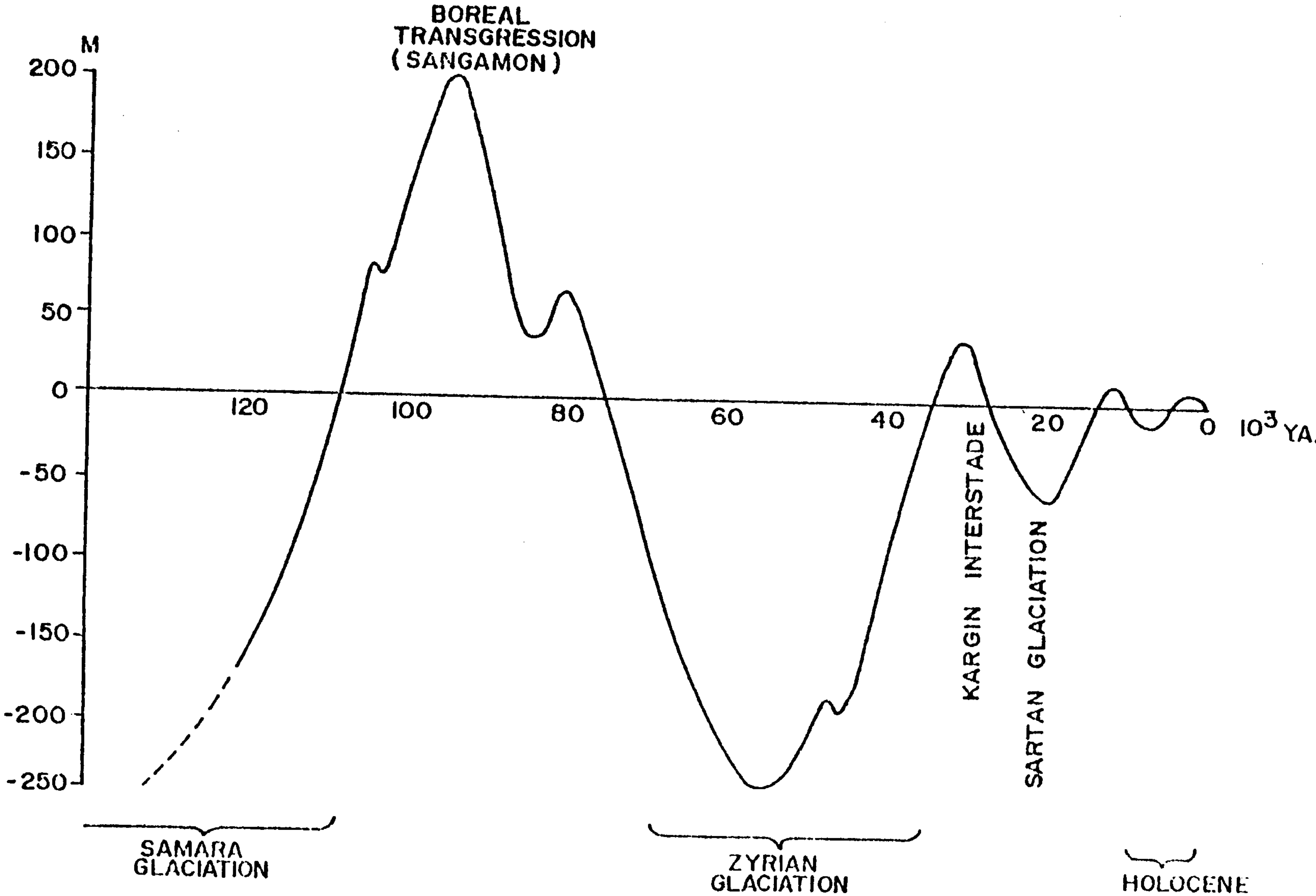


Figure 119.—Sangamon (Boreal) shorelines in the Kara Sea (S. Strelkov 1965).

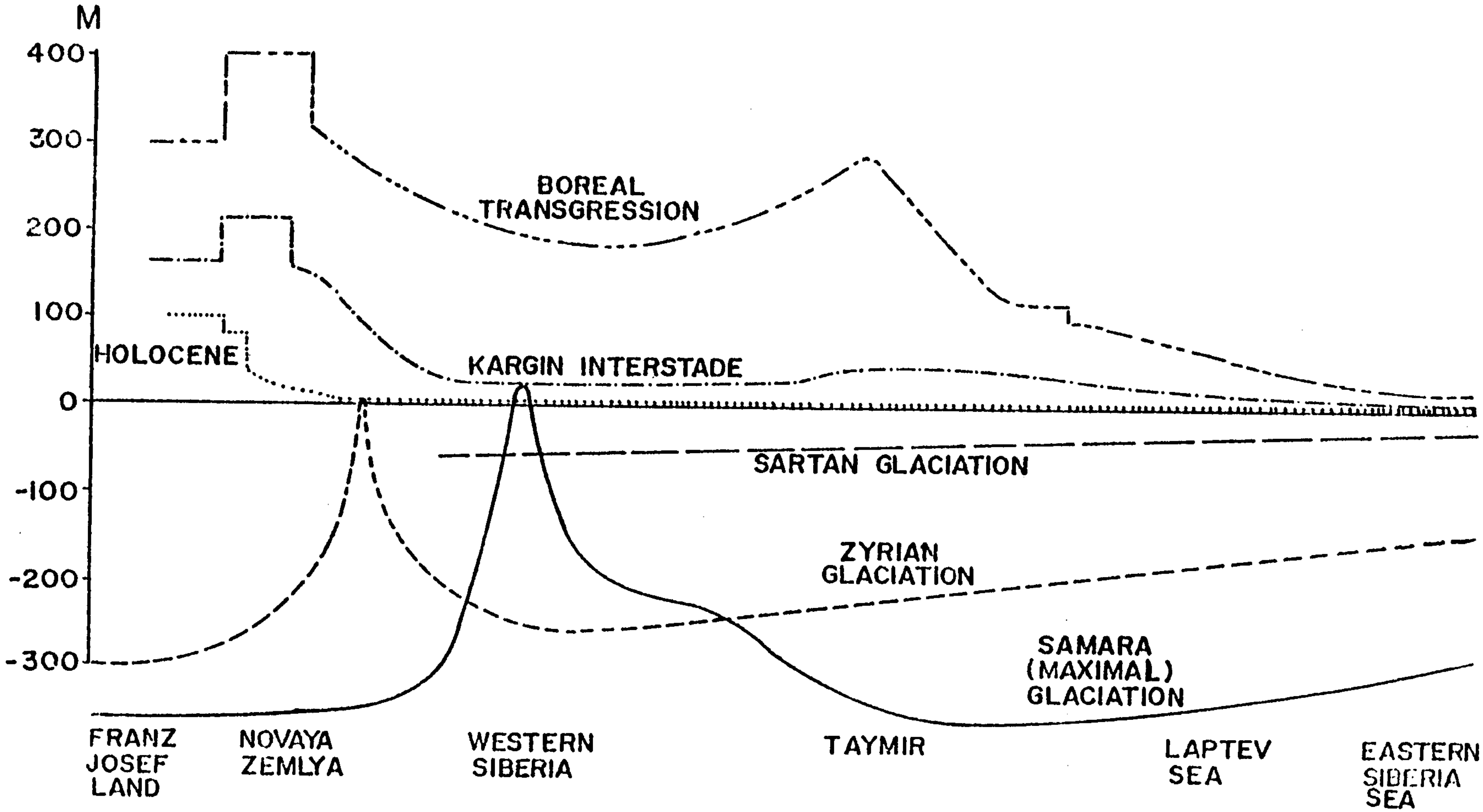


Figure 120.—Changes in shoreline positions during the Pleistocene (S. Strelkov 1965).

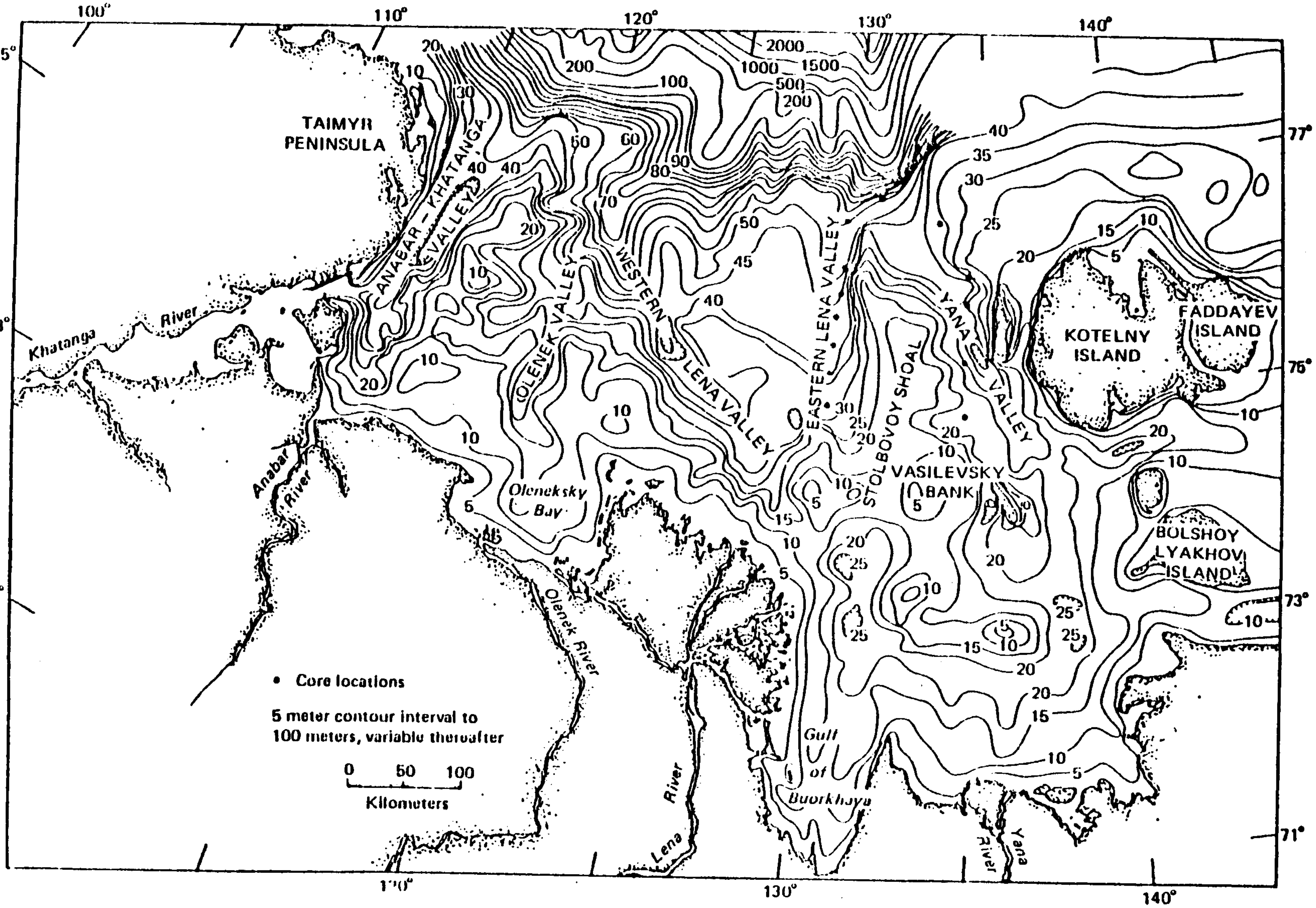


Figure 121.—Bathymetric chart of the Laptev Sea (Holmes and Creager 1974).

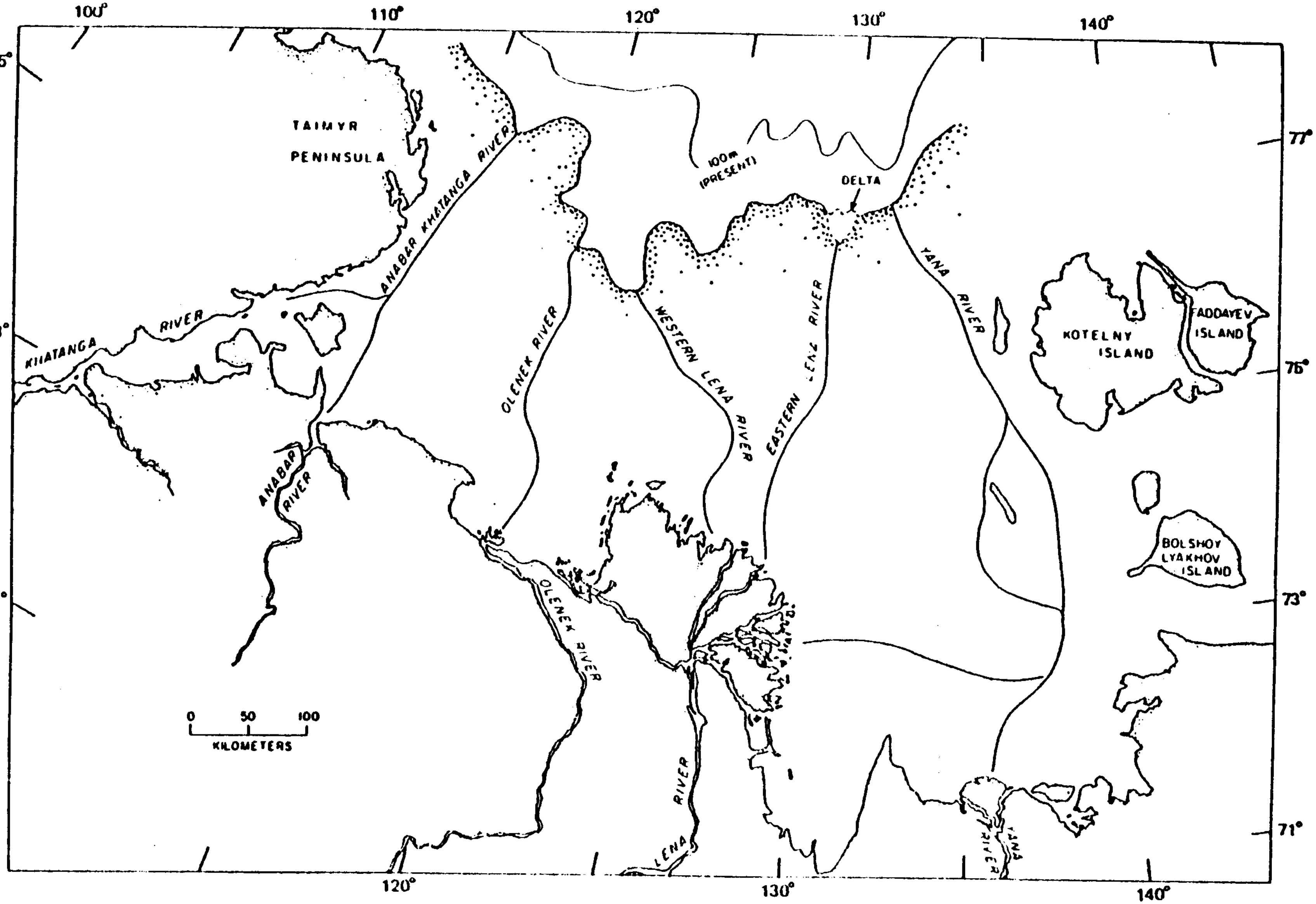


Figure 122.—Paleogeography of the Laptev Sea at about 15,000 years BP (Holmes and Creager 1974).

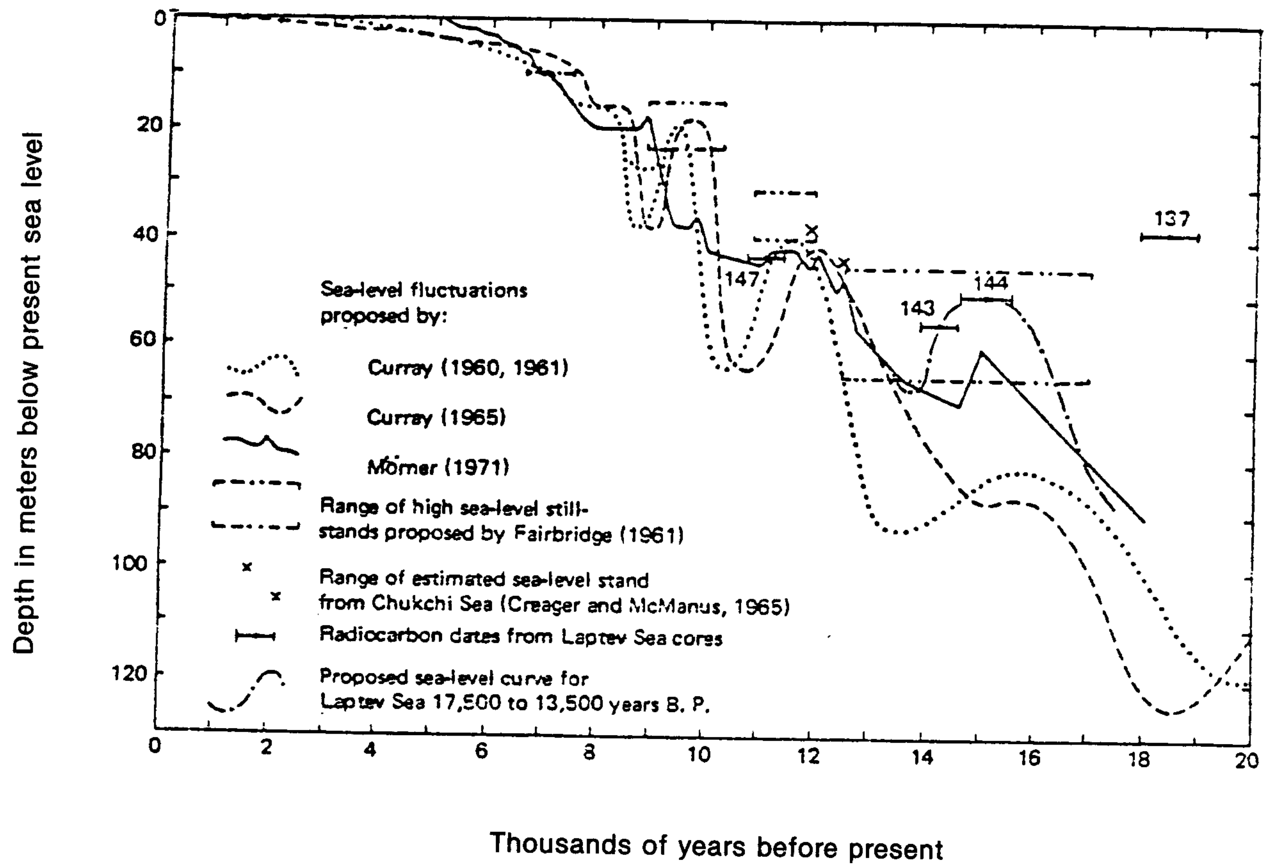


Figure 123.—Sea-level stillstands and fluctuations during Late Pleistocene and Holocene times (Holmes and Creager 1974).

The history of the shelf development is always considered when Russian geologists are trying to evaluate the extension and thickness of submarine permafrost. A typical example is the work of A. I. Chehkhovsky (1972) on offshore permafrost in the Kara Sea. In this work the problems of permafrost development (or the negative temperature zone on the areas with porous water of high mineralization) were considered on the Kara Sea shelf down to 200 m in depth. The present temperature conditions for permafrost formation on the whole of the Kara Sea shelf are different from the conditions of the sea sector with a water depth of 1–2 m. The shallow areas freeze to the bottom during the winter. The winter temperature of the deposits reaches –4.2° to –5.5°C at a depth of 3–4 m from the bottom (water depth of 1–2 m). This is low enough for the shelf area to be influenced by the rivers' thermal inflow. If the thermal conditions of the shelf at 1–2 m depth depend on climatic conditions, the thermal regime and extension of submarine permafrost deeper will be connected with such factors as bottom relief and temperature and salinity of the seawater; and these factors contribute along with the history of shelf development in the Pleistocene and Holocene.

The Kara Sea shelf relief is very irregular with vast shallow areas and deep depressions. In the middle of the sea, at about 80°E, the central Kara upland extends with an average depth of 50–100 m. The east Novozemelskaya depression, with depths down to 500 m, is situated along the east coast of Novozemelskaya. The trenches Santa Anna and Voronin are to the west and east of the central Kara upland and are located in the submeridian direction up to 79°. The hydrological regime of the Kara Sea is defined by a small amount of solar radiation, of cold water inflow from the Arctic basin, relatively low penetration of water from the Atlantic basin, and considerable salinification of surface water due to runoff of the Ob and Enisey rivers. The bottom roughness also has a strong effect on water mass circulation.

The water mass of the Arctic basin and Kara Sea is inhomogeneous in its physical characteristics, and consists of several kinds of water differing from one another by some characteristic peculiarities. In the Kara Sea shelf area one can define three principal water mass categories: Arctic, Atlantic, and river water. Arctic water formation is connected with migration of Atlantic water, water of the Pacific Ocean, river water, and water from Arctic basin melting ice; and in addition, the hydrometeorological processes occurring in the sea affect them. Atlantic water inflow in the Kara Sea, with the warm Gulf Stream currents, occurs at depths from 200 to 800 m in the Arctic basin. The river water occurs as far as the Yamal and Gydan peninsulas. The river water influence exists in the Kara Sea up to 75° N. Their deepest penetration is limited by warm water intermixing.

The considered area was under epeirogenic movement of different directions in the Quaternary period, but the major one, according to A. Chekhovsky, was extremely slow subsidence. From the viewpoint of A. Chekhovsky, the northern part of the lowland was inundated by Arctic basin water several times. The predominance of negative movement shows that almost all the present-day Kara Sea shelf during the Quaternary period was under water. This is why, according to A. Chekhovsky, the permafrost, which can be compared with thick permafrost found today in the northern part of western Siberia, could not be formed. During this subsidence, there were some positive movements. The maximum amplitude (range) of positive tectonic movement took place at the beginning of the Holocene (Kuzin 1961). Sea regression led to draining of the shelf down to a depth of 15–20 m. Therefore, most of the shallow shelf sectors, which were emerging from the sea at the beginning of the Holocene, had favorable climatic conditions for forming permafrost of considerable thickness. This thick permafrost formation under the Kara Sea shelf could also take place during glaciations. The author thinks that the total Kara Sea ice thickness was increased during the glaciation epoch. V. Zubakov (1972) considers the ice connected with the ice-shelf subformation, which corresponds in appearance to the coastal zone glaciation with typical development of both ground ice and sea ice.

In severe climatic conditions, the ice-shelf forms from surface precipitation and from the freezing seawater below the surface. As a result of mutual growth, the ice can go down to 200 m in depth. One can assume that the Kara Sea ice thickness during glaciation attained this figure depending on the shelf depth. At the present time such ice thickness and more can be observed in the Antarctic (Shackleton and Ross ice shelves). The extreme point of view, expressed by M. Grosswald (1970), is that the (shelf) ice thickness of the Barents Sea and, in part, the Kara Sea reached 1,800–2,000 m during glaciation. A. Chekhovsky supposes that the necessary conditions for forming ice 800 m thick existed on the Kara Sea shelf during the Ziryan glaciation (earlier Wisconsin). The climatic conditions were probably very severe on the whole present-day Kara Sea water area, and analogous to Greenland and Antarctica. One can assume that the average annual temperature could go down to –30° to –35°C. Simple calculations show that under the geothermal gradient of about 3° the permanent thickness can reach 400–700 m. Maximum permafrost thicknesses were formed under shallow shelf areas as they have less ice thickness and therefore the lowest temperature at the bottom. After the ice regression and inundation of the shelf area for seawater with a negative temperature (approximately 35,000–50,000 years ago) and ground freezing stopped and melting began because of the heat flow. Thawing would continue until complete permafrost disappearance. In the area with positive bottom temperature, it is only a question of time. Under the negative bottom temperatures the thawing would continue as far as the thickness comes into conformity with these temperatures. Considering only the heat loss for a phase change from ice to water, the decrease in permafrost thickness can be calculated this way:

$$G t \ \lambda_t = \frac{QPH}{T} \tag{11}$$

where Gt - geothermical gradient of the thawing zone (0.030 gr/m)

λ_t - coefficient of the thermoconductivity $(1{,}1 \ \frac{\text{k cal}}{\text{M Hour} \ ^{o}\text{C}})$

QP - phase change heat $(24000 \ \frac{\text{k cal}}{\text{m}^3})$

T - thawing time (50,000 years)

H - the value of the perennial melting from below (M)

The calculations of A. Chekhovsky show that during 50,000 years, about 500 m of permafrost could be thawed.

Analyzing the conditions of permafrost existence in the Kara Sea and using for the division of the shelf such characteristics as (1) the different kinds of water distribution, (2) morphology of the bottom shelf, and (3) history of the paleogeographical development, A. Chekhovsky made the following division of the Kara Sea shelf (Figure 124):

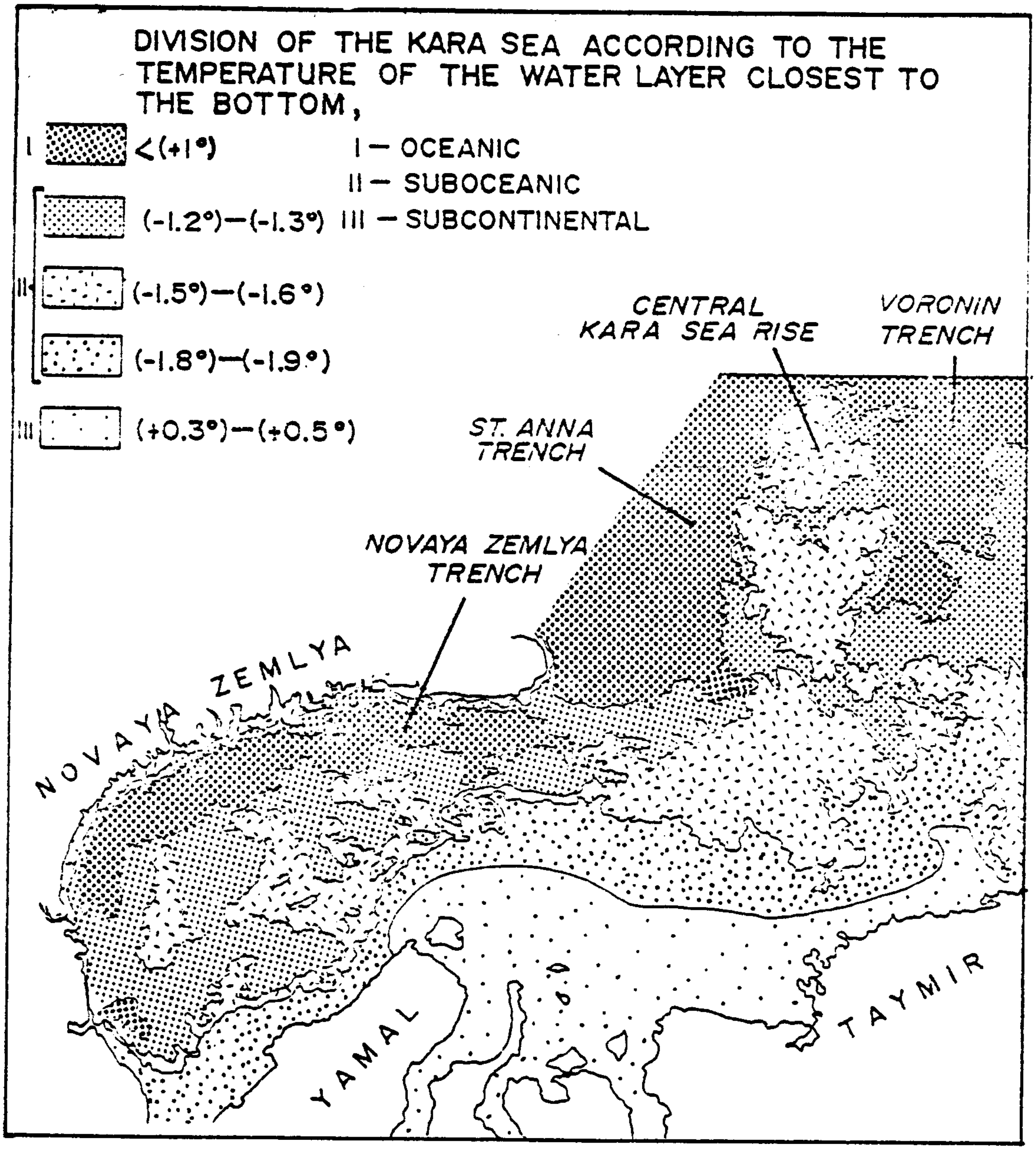


Figure 124.—Kara Sea division according to the temperature of the water layer closest to the bottom (A. Chekhovsky 1972).

I. *Oceanic Region:* Includes the St. Anna and Voronin trenches and Novaya Zemelskaya depression with depths of more than 200 m. The bottom of these depressions is formed by non-wave accumulation processes and they have flat, practically horizontal surfaces. The water filling them can be then characterized with distinctive stratification according to the temperature and salinity.

Arctic Waters can be divided into three layers:

1. The water layer at the depth of 50 m has a uniform (in a vertical direction) negative temperature and a salinity of 29 to 32 ppt. For this layer the convective mixture of water and the Arctic basin sea ice influence on the temperature regime and salinity are typical.
2. Below the first layer, within 50–100 m, there are waters with increasing salinity (up to 34 ppt). These waters have formed on the Arctic basin continental slope.
3. At depths from 100 m to 200 m, there are waters with a salinity of about 35 ppt. The water temperature of this layer forms under the influence of Atlantic water and the second layer waters.

Atlantic Water exists at depths of more than 200 m. It can be characterized by a positive temperature and salinity of about 35 ppt. This water flows to the Kara Sea along the St. Anna and Voronin trenches. Each water layer temperature for the oceanic region is presented in Table 24. Chekhovsky thinks that during the whole Quaternary period the bottom water layer temperatures of the ocean depression were positive. Therefore, permafrost could not have formed at the bottom.

II. *Suboceanic Region:* Occupies the shelf from depths of 20 to 200 m. In the eastern and western parts and also above the "Kara highland" the predominant depths are 50 to 100 m. Forming of the suboceanic bottom relief is mostly linked with an abrasion process. Tectonic subsidence caused the displacement of these surfaces from the wave-affecting zone to greater depths.

Table 24.—Relationship of temperature and salinity with water depth in the oceanic region (Chekhovsky 1972).

Sea Depth (m)	Salinity (ppt)	Temperature (°C)
0–50	29–30	–1.2 to –1.3
50–100	33–34	–1.5 to –1.6
100–200	35–36	–0.4 to –0.6
200	35	<1.0

The whole oceanic region can be characterized by negative temperatures of the bottom water layer. Temperature and water salinity to a depth of 1,000 m completely correspond to the waters of the first and second layers of the oceanic region. Water occurring below 100 m has considerably lower temperatures, which approach the freezing temperature of water with similar salinity (Tables 24 and 25). Permafrost can occur on the bottom of the suboceanic region. The thicknesses could be in accordance with present-day bottom temperature conditions or even greater. This situation probably can take place on the shelf area from depths of 20 to 50 m. During the conditions of the shelf glaciation epoch such areas could have the greatest thickness of permafrost (600–650 m). During the post-Zyrian period, permafrost of about 500 m in thickness could have thawed. Therefore, the present-day permafrost thickness in this part of the shelf can be found as a difference between thicknesses which were developed during the Zyrian period and were thawed after that time. This thickness is about 100–150 m. In the rest of the suboceanic shelf region the permafrost thicknesses depend upon present-day bottom temperature conditions. Permafrost thicknesses for different sea depths are shown in Table 26.

Table 25.—Relationship of temperature and salinity with water depth in the suboceanic region (Chekhovsky 1972).

Sea Depth (m)	Salinity (ppt)	Temperature (°C)
0–50	29–30	−1.2 to −1.3
50–100	33–34	−1.5 to −1.6
100–200	35–36	−1.8 to −1.9

Table 26.—Thickness of submarine permafrost in the suboceanic region of the shelf (Chekhovsky 1972).

Sea Depth (m)	Temperature (°C)	Submarine Permafrost Thickness (m)
0–50	−1.2 to −1.3	100–150
50–100	−1.5 to −1.6	60–80
100–200	−1.8 to −1.9	80–100

III. *Subcontinental Region:* This is situated to the north from the Yamal and Gydan peninsulas, and occupies a large part of the Yamal–Gydansky bank to a depth of 20 m. Here one can define the present-day abrasion and abrasion-accumulation surfaces. Abrasion surfaces form on the zone of wave influence and occur at depths of 15–20 m. In the coastal part of the shelf, the sedimentation of clastic material is derived from river drift or shore abrasion. The abrasion-accumulation surfaces form here and are covered by relatively thick layers of consolidated sands. The temperature and salinity of the water are affected by the Ob–Enisey current. The water of this current warms (+6 to +8°C) with a salinity of 10–15 ppt. The Ob–Enisey current keeps an extremely large reserve of the advective heat caused by continental run-off. In the winter the water of the Ob–Enisey current drops to –0.4° to –0.6°C, which corresponds to freezing for water with a salinity of 10–15 ppt. The calculated average annual temperatures are 0.3–0.5°C. As mentioned above, the influence of the Ob–Enisey current is limited by the depth of 10 m (to the north from Yamal and Gydan peninsulas). This isobath was taken by A. Chekhovsky as a border between the subcontinental and suboceanic regions.

Permafrost in the subcontinental region is both Pleistocene and Holocene in age. Its formation is connected with Zyrian glaciation and with the beginning of the Holocene, when part of the shelf had the characteristic of littoral plains. The present-day geothermal conditions of the bottom make it possible to assume that there is no permafrost here now. The permafrost relict layer occurs deeply from the bottom in this region and its formation is connected with the Zyrian glaciation. The base of the frozen layer in this region is 150–200 m. During the winter the bottom temperature in the subcontinental region is –0.4° to –0.6°C, affecting the formation of thin seasonal permafrost. Proceeding on the basis of the point of view described in his article and taking into account the Kara Sea paleogeographical development and the performed calculations, A. Chekhovsky came to the following conclusions about permafrost and its thickness in the defined regions:

1. In the oceanic region permafrost is absent.
2. In the suboceanic region permafrost exists under the sea bottom.
3. In the subcontinental region permafrost exists at considerable depths from the bottom.
4. In a large part of the shelf, permafrost thickness corresponds to present-day bottom temperature conditions. Shallow shelf areas, to 50 m in depth (subcontinental part of suboceanic region), have an abnormally great thickness of permafrost, which does not correspond to present-day temperature conditions. Here the present degradation of frozen rocks is taking place.
5. The greatest permafrost thickness (up to 150 m) should be expected in the suboceanic region to a depth of 50 m.

The usual attempts to estimate the thickness of submarine permafrost on the base of sea level history also meet serious difficulties because of the complicated tectonic movements of the coastal and shelf areas of the Arctic in the Pleistocene and Holocene. We have tried to make some graphics representing the change of the old shoreline's position along the Eurasiatic coast. Figure 125 shows the present condition of the boreal (Sangamon, Kuzenkova) transgression "shoreline" on the northern Eurasian Arctic coast. We see that the most shoreline raised are situated on the Kola Peninsula (180 m), Zemlya Franza Josifa (300 m), Novaya Zemlya (420 m), Polar Ural (280–300 m), Taymir (200 m), and northern slope of the middle Siberian Range (200 m). These data correspond to the concept of the limited glaciations of these areas and their glacio-isostatic rebound. On the contrary, the position of the Sangamon shoreline is low enough at the very large area of plains continuing into the shelf of the Arctic.

We have tried also to evaluate the general and relative character of the Arctic coast movement from Sangamon to the present; Figure 126 represents this evaluation. We see that the average level of the Sangamon shoreline position all along the Eurasiatic coast is about 130 m. This level, obviously, shows the main trend of the Arctic coastal area movement—uplift. The figures mentioned above 130 m give the numerical value of this process. In Figure 126 we may also see the relative uplift and relative submergence of the different geomorphological and tectonic structures. We see that the magnitude of the relative "block" movements reaches hundreds of meters and some morphostructures are always late in their motion in comparison with adjacent structures. That is why the geochronological estimations connected with submarine permafrost development in the Arctic may be completely wrong if they are based only on the glacio-eustatic ideas without concrete knowledge of the local tectonics for each given area.

The formation history of the cryogenic series in the territory of the USSR, including the series transformed by seawater and lying subaqually, was described recently by Fotiev (1976)* (Figure 127). This scientist distinguished four stages in the natural evolution of northern Eurasia (after A. A. Velitchko 1973)**. According to his opinion, in the central, northern, and northeastern regions of the Asiatic part of the USSR, the average negative annual temperature of the air, which caused the formation of cryogenic series, occurred in the end of the first stage (about 0.7×10^6 yr BP). The second stage (700,000 to 30,000 yr BP) was marked by a freezing of rocks in the European part as well, but here and in the south of the Asian part

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* S. M. Fotiev, "Formation history of the cryogenic series in the territory of the USSR," International Geography, Geomorphology and Paleogeography, Moskva, 1976.

** A. A. Velitchko, "Prirodny process Pleistocena," Nauka Moskva, 1973. (Natural Process of the Pleistocene.)

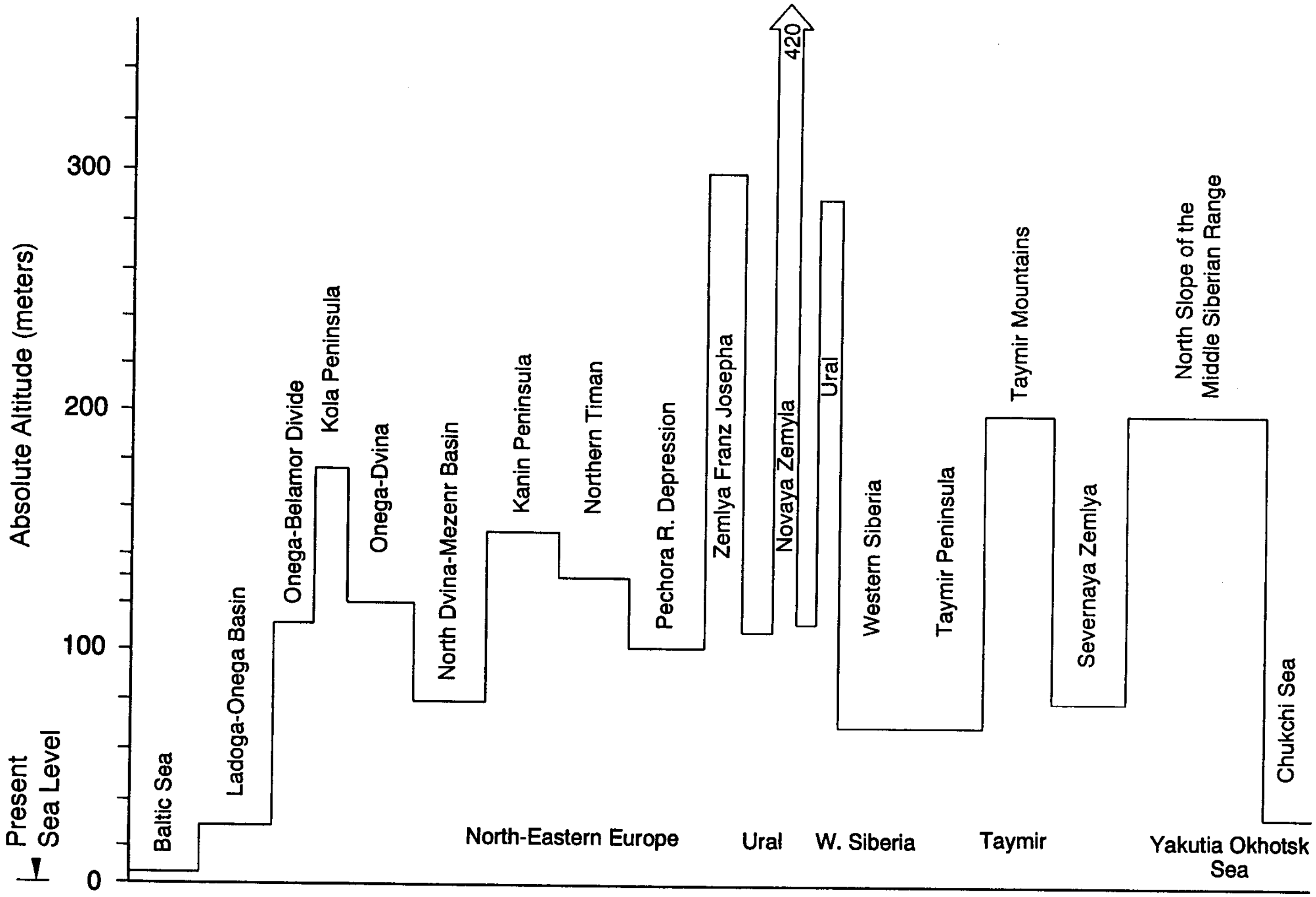


Figure 125.—Present position of the Boreal (Sangamon, Kazantcevo) shoreline on the Eurasian coast.

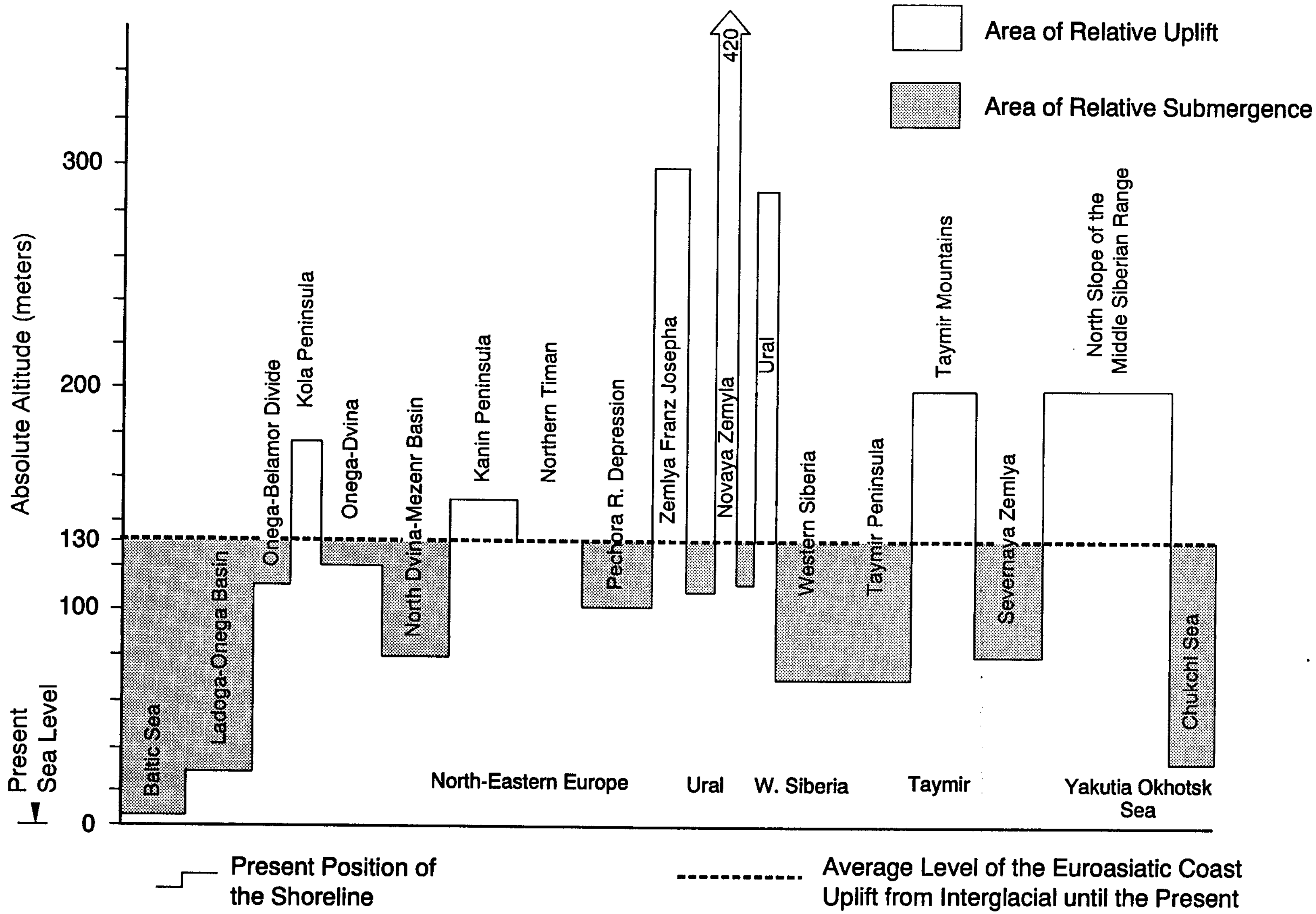
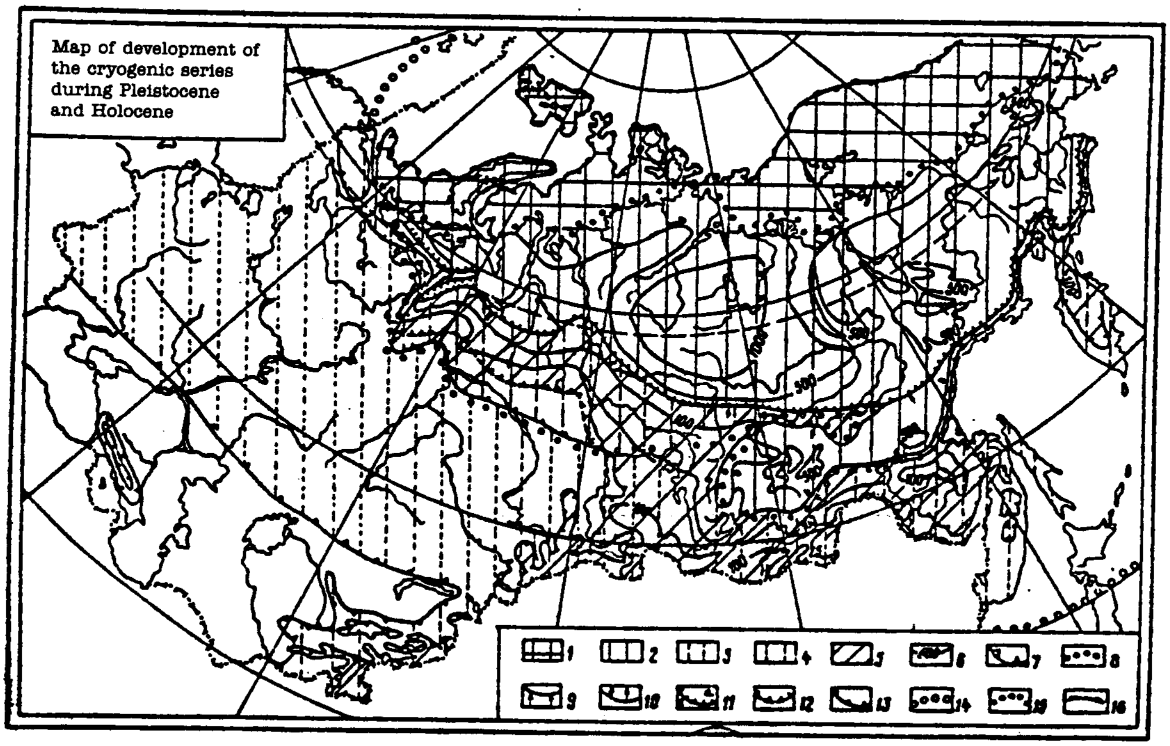


Figure 126.—Relative uplift and submergence of the different Eurasiatic Arctic areas, geological and geomorphological structures from Sangamon to the present.



1–4: The area of distribution of the cryogenic series of Pleistocene age:

1. Transformed by sea waters and lying subaqually
2. Not degraded from the surface during Holocene
3. Degraded from the surface during Holocene (the roof is at the depth of 80–300 m)
4. Completely degraded during Holocene

5. Distribution area of the cryogenic series of Upper Holocene age
6. Thickness isolines of the cryogenic series

Boundaries:

7. Cryogenic regions during Pleistocene climatic minimum (After A. A. Velichko)
8. Areas of completely degraded cryogenic series during Holocene; areas of the cryogenic series which did not degrade from the surface
9. Latitudinal and zonal heat exchange peculiarities
10. Altitudinal and zonal heat exchange peculiarities
11. Between the Northern (a) and Southern (b) geocryological zones

12. Distribution area of the Upper Holocene cryogenic series in the west Siberian and Pechorian basins (After V. V. Baulin and N. G. Oberman)
13. Area of the recent cryogenic series (After I. Ya. Baranov, V. V. Baulin and S. M. Fotiev)
14. Area of marine ices during Upper Pleistocene (After A. A. Velichko)
15. Land area during Upper Pleistocene (After N. N. Nikolaev)
16. The land boundary during Upper Pleistocene

Figure 127.—Map of development of the cryogenic series during the Pleistocene and Holocene (Fotiev 1976).

of the country, the epochs of perennial freezing were repeatedly followed by epochs of melting. During the third stage (30,000 to 9,000 yr BP) a homogenous sharp continental climate developed there over the Northern Hemisphere. According to Fotiev (1976), it was in the epoch of the chief climatic minimum of the Pleistocene that the severest geocryological environment formed on the huge territory of Eurasia. Rocks over the entire territory cool down to the lowest negative temperatures. The cryogenic area reaches its maximum expanding not only because of perennial freezing of the rocks in the low latitudes, but in the high latitudes as well, where a huge area of the land had emerged from the sea level due to a sea regression. S. M. Fotiev thinks that during the fourth stage, because of a less severe and less continental climate with increased precipitation, the cryogenic series began to degrade and this degradation reached its maximum about 4,500 yr. BP. Along the northern margin of the cryogenic area, due to the sea transgression, the seawaters began to change the cryogenic series formed in the subaerial conditions. Both on land and on sea the natural complexes changed more essentially in the European sector of Euroasia affected by the Atlantic. To the north the melted rocks became thinner down to complete disappearance. In the northern and northeastern regions the temperature at that time increased only within the negative values.

A new stage of perennial freezing began during the late Holocene. The southern boundary of the cryogenic area within the limits of the platform again moved to the south: in the European part of the USSR to 150–200 km; in its Asian part, to over 1,000 km.

Taking into account the history of the perennial freezing of rocks and considering also the warm epochs of the Holocene, S. M. Fotiev distinguishes two geocryological zones, namely, the northern and the southern ones, and identifies them within the recent area of cryogenic region. Their boundary coincides with the line that distinguishes the limits of the cryogenic series of the Pleistocene from those of the late Holocene. This boundary passes between the territories with essentially different parameters of the cryogenic series. In the northern zone the cryogenic series has been existing continuously during scores and hundreds of thousands of years. Here it is distributed worldwide and is very thick (from 150 to 1,500 m and even more). It has low temperatures (from −2 to −16°C). In the southern zone (except the relicts) the cryogenic series has been continuously existing during some thousand years. Its distribution is of very interrupted and island nature. Its thickness is rather small (from 10 to 100 m, very seldom more), its rock temperature is rather high (from 0 to −2°C). On the S. M. Fotiev map of development of the cryogenic series during the Pleistocene and Holocene (Figure 127) the series transformed by seawater and lying subaqually, extend all over the Eurasiatic part of the Arctic shelf about 400–600 km to the north from the present Arctic Ocean shoreline. The area of the submerged cryogenic series under the ocean according to Fotiev's evaluation is much more extended than after Baranov (1960), Figure 12.

We see that S. M. Fotiev defines the third stage (30,000–9,000 yr BP) as a time "when the huge area of the land has emerged from the sea level due to a sea regression". The materials given above show also that paleogeographical conditions of the shelf during the last glaciation time (25,000–10,000 yr BP) were favorable for submarine permafrost development. However, during the Early and Middle Pleistocene, especially during the maximal (Illinoian) glaciation epoch the conditions of the shelf developed another way possibly because of the cold marine transgression in the Arctic Ocean, induced by its isolation from the World Ocean.

As it may be seen in our Figure 46, showing the dynamics of the Global Ocean and the Arctic basin levels during the Pleistocene, the most appropriate periods for the development of submarine permafrost were the following:

1. Time of the great regression of the World Ocean between the end of Jaramillo times and the beginning of Brunhes times with magnitude of sea level oscillations of several hundred meters about 890–690 thousand years ago.
2. Periods of the glaciations and glacio-eustatic falls of the Arctic Ocean 115–100, 72–45, and 25–10 thousand years ago.

The period of time from the beginning of the Brunhes to the end of the maximal (Illinoian) glaciation, and following interglacial, was inappropriate for the exposure of the Arctic shelf because of the high stands of the Arctic Ocean during glaciations that were connected with the isolation from the World Ocean and the cold marine transgressions of the "Reservoir" type due to the river inflow.

TYPES OF SUBMARINE PERMAFROST ZONES

The problem of submarine permafrost zone typification was recently considered by V. Kudrysvtsev (1975), I. N. Romanovsky (1975), I. Danilov and Y. Zhigarev (1977), and in part by F. Are (1976). The most comprehensive typification was made by V. Kudryavcev and N. Romanovsky. The cryolithozone of the Arctic shelf could be divided into three types:

I. The part of the shelf with under-sea deposits saturated with saline water, the water temperature being below 0°C. The origin of the water is marine.

II. The layers of ice-bonded deposits in the shelf structure. They are the result of the continental development of the shelf during its exposure.

III. The perennially frozen rocks of the shelf which were formed under the coastal conditions, often beneath a layer of low temperature saline water.

Type I of the cryolithozone extends over large areas of the Arctic shelf, continental slope, and deep trenches, where the seawater temperatures are negative. The colder and more saline seawater infiltrates into the permeable rocks. During periods of ice development, this process became more active due to higher salt concentration and density of the water under new ice, essentially in shallow areas. In the area close to the coast, especially near river deltas, this convection may become weaker due to the fresh river water inflow. That is why the periods of continental conditions of climate are less favorable for the development of the cryolithozone (1). The second means of creation for this zone could be connected with the supply of underground saline and cold (negative temperature) water from the continent to the shelf. An understanding of the hydrological structure is very important here. If these structures are open to the sea there is a good possibility for this water to recharge at the shelf, and this could be the source of the positive and negative temperature anomalies of the seawater close to the bottom. We see that the character of the hydrological structures can greatly influence the development of this cryolithozone. Usually the coastal area of the shelf is only part of the continental hydrological structure. All artesian basins in the Soviet Arctic are usually open to the North Pole. Ya. Neizvestnov, Iu. Semenov (1973), Ya. Neizvestnov, N. Tolstikhin, O. Tolstikhin, S. Tomirdiaro (1972), and others have published some works on hydrological peculiarities of the Arctic shelf and hydrological division of this area. They have emphasized the importance of the ''criopegi'' study. This name has been given by the authors mentioned above to saline water with negative temperature. The divisions of ''criopegi'' are: moderate cold, –0 to –2°C; cold, –2 to –8°C; very cold, –8 to –23°C; the most cold, –23 to –36°C; and super cold, <–36°C. First three gradation were found at the shelf deposits of the northern Eurasia. Let us emphasize the role of the ''criopegi'' in the permafrost development and existence in the first cryolithic zone.

Cryolithozone II is situated in the areas having been once coastal and exposed one or more times before and later submerged under seawater. We have seen the periods that are the most suitable for this due to Arctic Ocean regression, especially for those areas with a high rate of uplift. We think the combination of both processes could give the effect of exposure and vice versa. Of course, in the case of submergence, thermoabrasion took place and was especially intense. The presence of ice lenses and ice layers in the deposits (often in the dusty aleurites of alluvial-lacustrine origin) helped greatly in terms of thermoabrasion. For example, the rate of coastal retreat is about 4–5 m/year on the average and in some places reaches 100 m/year. Some islands in the Soviet Arctic, for instance, have disappeared from sight in one generation. In comparison with coastal areas where the temperature of the surface rocks is –5 to –6°C to –10 to –12°C, on the continental slope, temperature rises by 1–1.7°C. This rising of temperature influences the degradation of the upper layers of the permafrost.

In this second zone two different trends take place: first, a thickening of the cryolithic zone due to the influence of the convection of the saline cold water; second, a decrease in thickness due to an increase in temperature and the processes of the ice melting under the influence of saline water. Usually the thickness of the permafrost here changes from several to 40–60 m. The age of the permafrost depends on the beginning of the continental regime and could be varied from the Pliocene until now. If there were several periods of conditions (exposure), the cross section of the permafrost includes several layers and thick lenses (~20 m) of underground ice. In this case the general thickness of the perennially frozen rocks of the shelf can reach 200 m. The complicated history of these parts of the second Cryolithic zone is usually reflected in the equally complicated structure and texture of the deposits due to the dramatic thermal regime changes in time. The contact of saline water and fresh ice is possible only in conditions of high permeability of the rocks. That is why the ice, formed during the continental regime, may be preserved in impermeable clayey and silty deposits for a long time, but not in sandy deposits. Because of their high permeability, gravel and sand fractions help to transfer the ice-bonded deposits into the unbonded but cold deposits saturated with saline water of negative temperature.

Cryolithozone III is situated at the shallow parts of the shelf where the processes of sedimentation take place or the areas progressively become shallow due to recent positive movement. The intensive process of permafrost creation is very active in the territories near river deltas, where the deposits are saturated mostly by the fresh, cold water. The evaluation of the freezing temperature of this water according to salinity may help in the prediction of areas with permafrost. Development of grounded ice and sea ice positively influences permafrost development, especially if the bottom deposits are clay and impermeable to the relatively saline water under ice. It is important to pay attention to the impermeable deposits because they are suitable for permafrost existence in this zone limiting infiltration of seawater with higher salinity after development of the sea ice. Looking at the permafrost data for the Eurasiatic shelf, we may see the third cryolithic zone extension in the deltas of the Ob, Eyinsey, Hatanga, Lena, Yana, Indigirka, and many other rivers. This zone is closely connected with the first and second zones on the Arctic Shelf but more favorable conditions for the third zone do exist at the eastern part of the Eurasiatic shelf of the Arctic.

CONCLUSIONS

After looking through all available materials on subsea permafrost on the Eurasiatic shelf of the Arctic Ocean, we may more critically evaluate not only the existing data but also published summaries of that problem. In our opinion the latest summary, "Cryolithozone of the Arctic Ocean" by F. Are (1976), CRREL draft translation 686, gives no comprehensive and up-to-date picture of the subsea permafrost study level on the Eurasiatic shelf. It seems to us that F. Are's summary is primarily, and only, a report on the formal history of investigations, especially on sea level history, without paying attention to the principal differences and contradictions of the main ideas playing the decisive role in explaining submarine permafrost origin, development, and relict state.

We can make the following conclusions.

1) The materials discussed in this report show that Arctic Ocean oscillations were of three major types: (1) "catastrophic,", (2) glacio-eustatic (usually known), and (3) "reservoir." Only the periods of catastrophic fall in the time of the Mathuyama–Brunhes paleomagnetic changes and the glacio-eustatic falls at the Upper Pleistocene were appropriate for permafrost develpoment on the exposed shelf. Most of Brunhes time, including the period of maximal glaciation (Illinoian), was unsuitable for permafrost development because of the alternations of the cold marine glacial and warm interglacial transgressions far southward of the modern ocean shoreline. We think the evidence from northern Europe and Siberia, Chukotka Peninsula, Bering Strait, and other areas, including the Arctic and Atlantic bottom deposits, gives a picture of the isolation of the Arctic basin from the global ocean and explains the cold marine transgression phenomena. The combination of the glacio-isostatic and block tectonic processes could be used as a mechanism for dam creations in the North Atlantic. Simple ice cap development in the Bering Strait and Bering Sea is a good explanation of the isolation from that side of the Arctic.

2) We could not find any positive data on the development of the ice sheets and caps on the Eurasiatic shelf. The direct geological materials and the data on mass anomalies and postglacial movements, the marginal glacial forms and paleobotanical forms (data on refugiums) all along the shelf may be interpreted as the negative evidence in this case.

3) We consider the concept of shelf glaciation as probable but only for a limited and shallow part of the Barents Sea and less probable for certain small areas of the Kara Sea. We think that periods of Arctic Ocean isolation and rising of its level were appropriate for ice shelves, especially because of river inflow. At the same time, the rising of the sea level to 100–150 m was an obstacle for the interaction of the ice shelf or floating ice with the bottom due to a substantial increase in the water depth. On the other hand, the increase in the thickness of the ice-shelf influenced the increase of seawater salinity under the ice during the development of this ice. This process as the inverse one was involved with a system of connections limiting the rise of floating ice thickness. This fact agrees well with the normal marine conditions existing during the cold marine transgressions and reflected in the faunistic and microfaunistic complexes. It is understandable that these periods were not suitable for submarine permafrost development.

4) One of the important lessons of the Eurasiatic shelf study is the recognition of the highly contrasting block movements of the different geological structures (differential tectonics) on the coast and shelf. The magnitude of such a relative movement for adjacent blocks

often reached hundreds of meters for the period from the last big interglacial until now. The rates of movement reach several centimeters, or millimeters a year, and could be seen on maps of recent tectonics made from geodetic measurement. The contrasting movements are especially active at the areas along the active faults in the active zones. This means that any calculation of the age and thickness of permafrost made without evaluation of local tectonics could not be of great value. This is especially important to keep in mind during the search for permafrost on the shelf in the zones of deep faults, dividing the different structures, in the zones of high seismicity, for instance, such as the Prudhoe Bay area with the monstrous oil field lying exactly on the boundary of the two plates. The graphics shown in this report help us to understand also that the main trend of Arctic shelf and coast movement was mainly positive during the Pleistocene. It appears as though this process is part of the contrasting movement of the two biggest structures: submergence of the central deep parts of the Arctic basin and rise of the continents with shelves surrounding it. The block-tectonics of the coastal and shelf areas are only the second, third and so on levels related to this long developing process of Arctic basin spreading.

5) Knowing more about the specific and main peculiarities of Arctic submarine permafrost geological history we might concentrate more on the summary of the direct data, gathered during the special investigations. We saw on the maps of the Eurasiatic shelf that the extension of permafrost reaches 400 km and more, practically covering all the shelf and partly the continental slope. This extension is shown mostly from the geophysical data and partly from the drilling data. The published data on drilling consider only a distance of about 25 km, sometimes 35 km. The thickness of the submarine permafrost in some places reaches 120–150 m. Sometimes there are several layers of permafrost, divided by talik. The upper and possibly the lower boundaries of permafrost are highly irregular, as seen in the results of seismic refraction investigations. Results of the available materials about Soviet geophysical investigations and related bibliography were mostly gathered and used by Hunter, Judge et al. (1976) in their work for the eastern Beaufort Sea.

6) The submarine permafrost of the shelf consists of three different "cryolithic" zones, described above. This division is conditioned not only by geological development but also by environmental peculiarities, especially geohydrological and oceanographic factors. The role of the geohydrological structures and chemistry (composition and salt concentration) and temperature of the underground water is extremely important. The same is true with seawater. Other factors include ice–bottom interaction and salt transfer from ice to water during ice development. The grain size of the deposits, their permeability, moisuture, and pore water salinity balance play very significant roles in the changes

of the thermal regime of the rocks and development of the two main types of permafrost: frozen rocks containing ice (ice-bonded permafrost) and cold (frost) rocks whose temperature is below zero but which do not contain ice (unbonded permafrost). The results of investigations on the Arctic shelf of Eurasia allow us to emphasize that the boundary between these two types of permafrost is extremely variable in the vertical and horizontal directions, because of the frequent changes of the parameters mentioned above. We can add also that this boundary is highly flexible in space and time. The scale of the time change could be variable in many ways: seasonally, annually, connected with short climatic cyclicity (4, 8, 12 years, etc.) or related to longer cycles (~1,800 years, 4,000 years, and others—those connected with the inter-Holocene changes) or the glacial-interglacial cycles and rhythms.

7) The empirical and experimental data show some interesting specifics for the shelf rock thermal regime; for instance, the connection between the pore water salt concentration and freezing temperature of the deposits. Another important result is the fact that the freezing point for deposits with some salt concentration is highly dependent on their grain size composition and mineralogy. The character of the relationship between changes in salinity, moistness, and temperature of the rocks and the amount of unfrozen water is also a point of interest. In addition, the experimental data show that salinity of the deposits, the factor that decisively influences their phase conditions, does not have much effect on their thermophysical characteristics. These characteristics depend mostly upon the moistness of the deposits. The empirical formulas for the definition of (1) the freezing temperature of the rocks due to their salinity and moistness; (2) the amount of frozen water in its connection with the salinity and moistness of deposits.; (3) the coefficient of thermal conductivity of the rocks relating to their moistness; (4) the volumetric thermal capacity of the deposits according to the different moisturization; and (5) the seasonal changes of seawater salinity in its connection with sea ice development—all these formulas could be very helpful in continuing investigations and might have a serious practical significance.

The available results of the submarine permafrost study on the Eurasiatic part of the shelf are concrete input into the efforts to assess the environmental hazards of the American part of the Arctic shelf. This is important, particularly now because of exploration for oil and gas reserves.

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