Section I:
Selected Topics in Marine and Coastal Climatology

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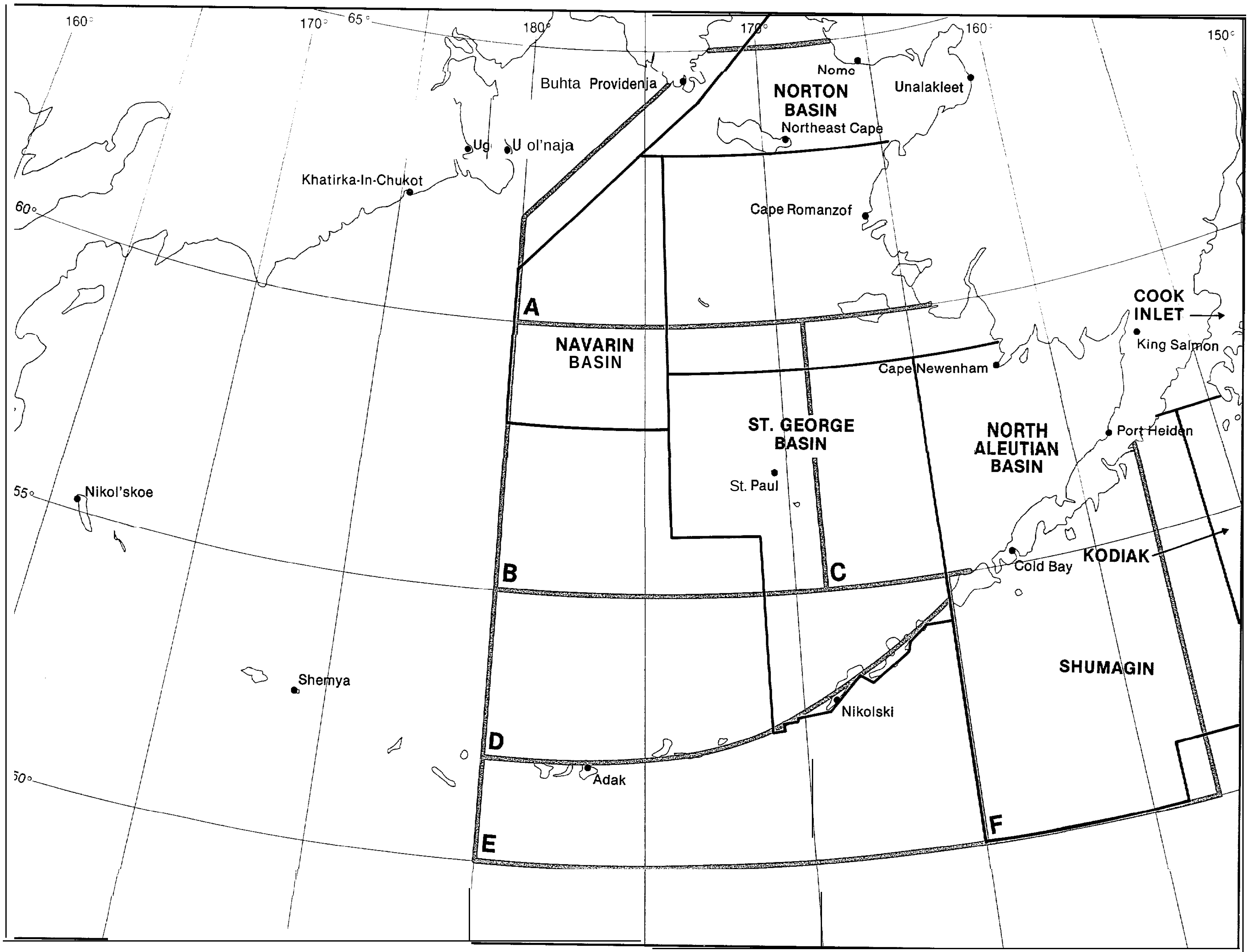


Figure 1. MMS Lease Sale Areas

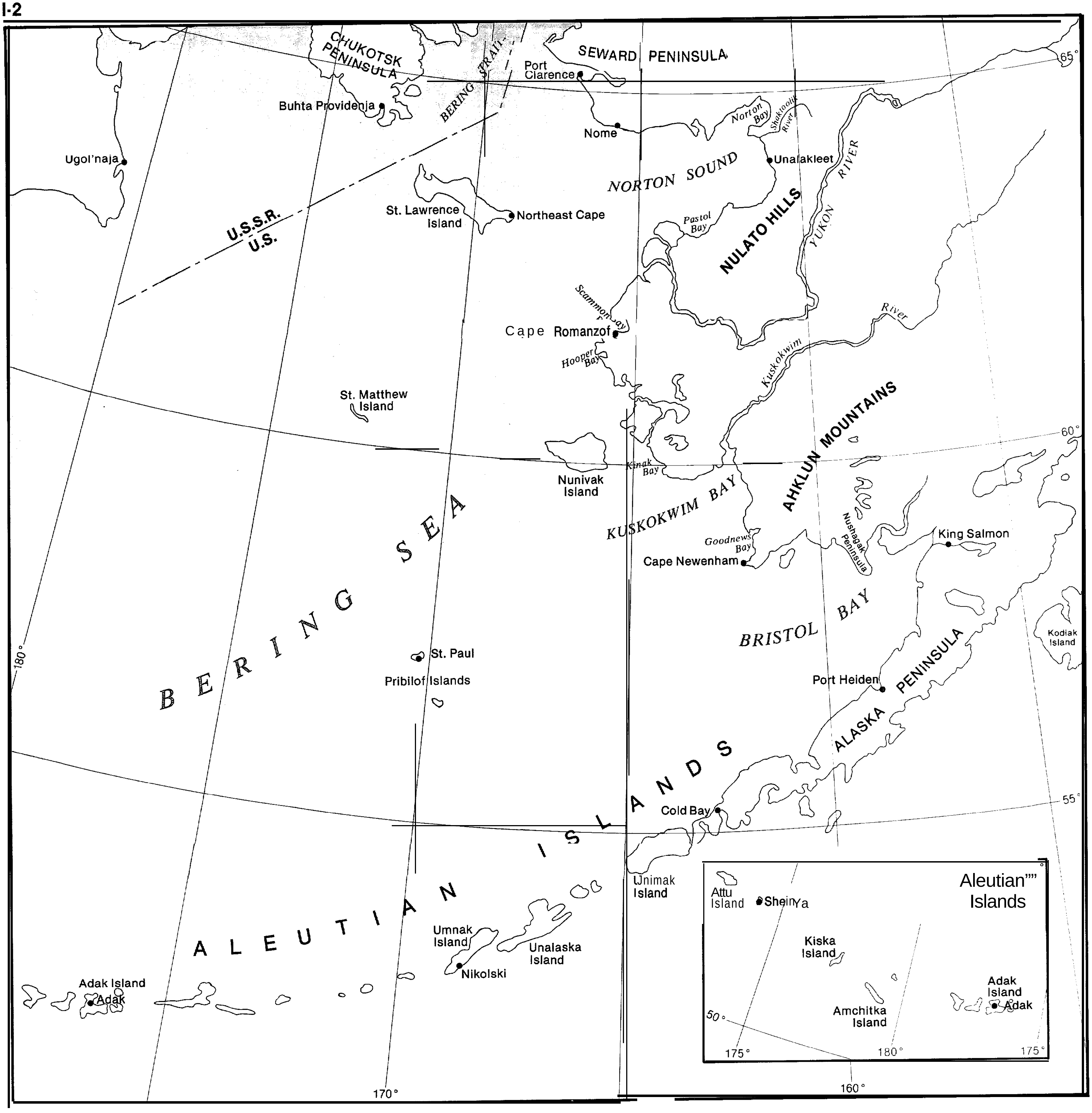


Figure 2. Place Names Map

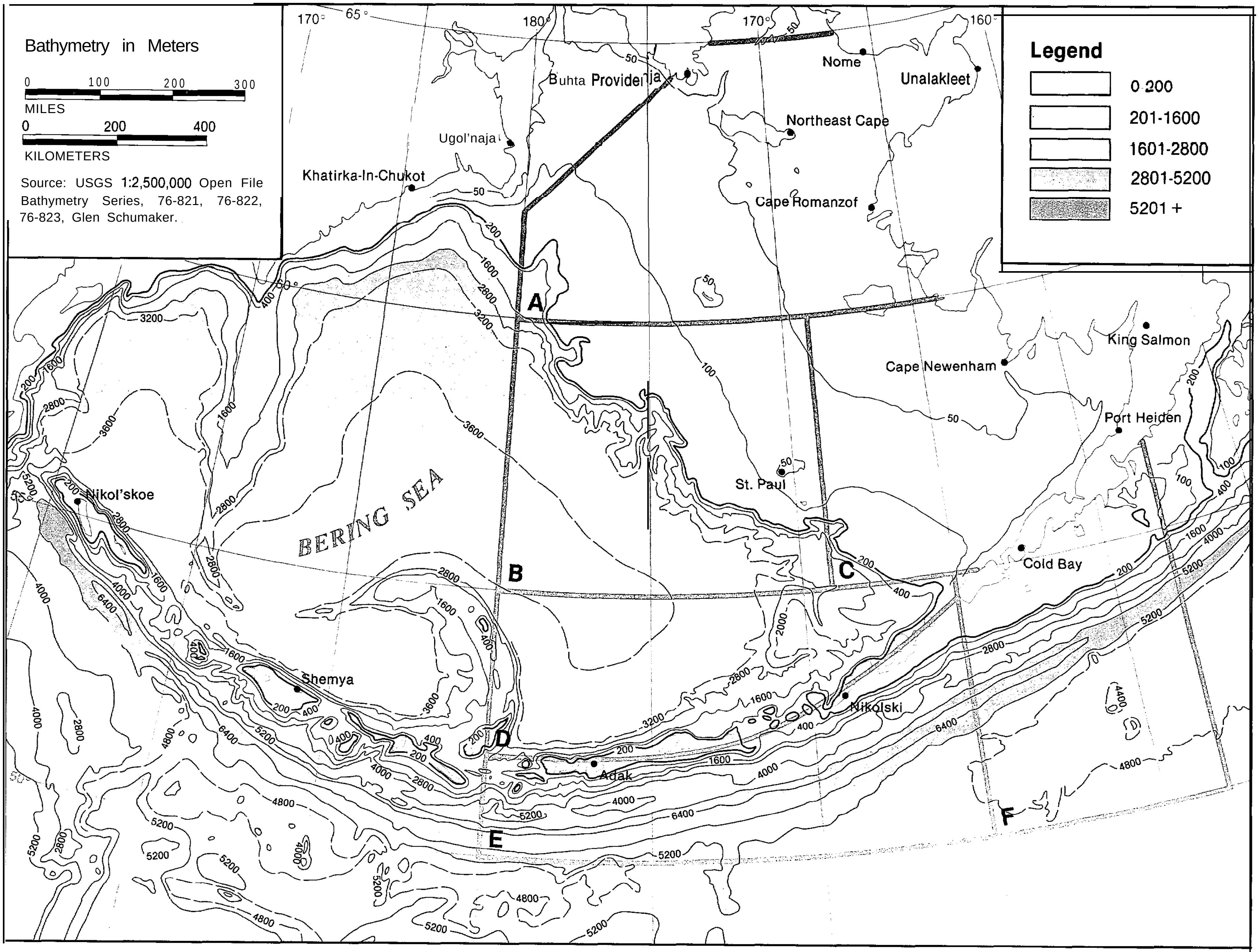


Figure 3. Bathymetry

Currents of the Bering Sea

North Aleutian Shelf

The primary flow of water into the Bering Sea originates at Unimak Pass. The source waterofthisflow is the Alaskan Coastal Current, from south of the Aleutians. Within the pass and north of Unimak Island, much of the coastal current is entrained into the wind-driven flow along the north Aleutian coast. Typically, this current flows to the northeast into Bristol Bay in the direction of the prevailing wind, following bathymetry contours along the coast. At times, the north Aleutian coastal current will undergo a reversal in direction due to changes in the large-scale and mesoscale wind direction. Because winds are highly variable, their contribution to net circulation is difficult to quantify, but the alongshore component of winds is highly correlated with both onshore and alongshore components of surface and subsurface currents.

Sea level changes on either side of Unimak Pass due to storm track and pressure cell movement are probably responsible for the fluctuation of magnitude and direction in the flow through the pass, which at times is southward. These reversals are more likely to occur when the flow from the seasonally variable Alaskan Coastal Current, from the Gulf of Alaska, is at its minimum. The shoaling bottom through Unimak Pass gives rise to vertical turbulence and mixes the water column.

On the north Aleutian shelf, the net northeasterly flow of approximately l-5 cm/s is present within the coastal zone (Baker 1983; Cline et al. 1982; Thorsteinson 1984). This current is believed to be continuous with a weak current past Nunivak Island (Kinder and Schumacher 1981). Near Port Moller, currents have smaller magnitudes and do not intensify near the coast. Close inshore, within 50 km, currents ranges from 1 to 6 cm/s (Kinder and Schumacher 1982).

A weak mean flow shows a cyclonic tendency around the perimeter of Bristol Bay, with maximum speeds (roughly 3.5 cm/s) found near and inside the 50-m isobath and in the coastal domain. Mean speeds observed in the central shelf domain were less than 1.0 cm/s, with no sense of an organized circulation (Kinder and Schumacher 1981). There is apparently a net westward convection of water from the central basin of Bristol Bay into the Bering Sea. However, flow in this central region is highly variable, atmospherically forced, and difficult to quantify. Coastal waters along the northern boundary of Bristol Bay, also called the coastal current, continue to follow the bathymetry. The coastal current flows northwesterly into the Bering Sea and then northerly along the Yukon/Kuskokwim Delta. Thus, the fundamental circulation in outer Bristol Bay consists of a typically unclosed, counterclockwise gyre open to the Bering Sea and driven by acombination of wind, tide, estuarian, and thermohaline effects.

Ninety to ninety-five percent of the velocity variance within the bay is tidal, with tidal currents an order of magnitude largerthan the mean flow. For example, on the north Aleutian shelf, where net currents are only l-5 cm/s and the typical wind-driven currents are approximately 10 cm/s at 5 m, the tidal currents are 40-80 cm/s or more (Thorsteinson 1984). Turbulence resulting from tidal currents causes mixing of the water column from the bottom to about 50-m above the bottom. Tidal currents in Bristol Bay are nearly reversing along the Alaska Peninsula and become more cyclonic and rotary offshore. National Ocean Survey current tables show a change in maximum ebb currents from 20-25 cm/s up to 30-40 cm/s in June near Amak Island. Near Port Moller, the tidal current speeds are as high as 100 cm/s (U.S. Department of Commerce 1980). At a depth of 2m the calculated tidal residual current is approximately 3-4cm/s, spatially highly variable, and directed to the northwest (Leendertse and Liu 1981).

Kinder and Schumacher (1981) identified three separate hydrographic flow regimes in the southeastern Bering Sea. The *Coastal* regime is present inside the 50-m isobath in the vicinity of Nunivak Island. It is characterized by generally warm, low saline, vertically well-mixed water which has typical currents on the order of 2-5 cm/s toward the northwest. The *Middle* regime is present in the central Bristol Bay region, where water depths are on the order of 50 to 100 m. It isdividedfrom thecoastal regime byafront with an enhanced salinity gradient and is characterized by a strongly stratified, two-layered structure extending approximately to the 100-m isobath. Mean flow is generally less than 1 cm/s, with no characteristic vector-mean direction. The *Outer* hydrographic region is divided from the middle region by a front along the 100-m isobath and is present out to the shelf break in the open waters beyond Bristol Bay. A fine vertical structure separates surface layers from the deeper, more well-mixed layers. The vector-mean current in this regime is directed to the northwest, with magnitudes on the order of l-10 cm/s and a statistically significant cross-shelf component of about l-5 cm/s.

Yukon Delta

The dominant current near the Yukon Delta is the northward flowing Alaskan Coastal Water. The current is thought to bifurcate at the northwest corner of the delta, with one fork flowing inland, toward Norton Sound, and the remaining flow continuing northward (U.S. Navy 1958). Local and seasonal effects can produce variability in the prevailing flow directions. In winter, when winds are from the north, flow offshore of the delta can actually reverse for days or weeks at a time (Aagaard and Coachman 1981). This situation accounted for the flow of the Alaskan Coastal Water about 30% of the time between September 1976 and March 1977 (Zimmerman 1982). The surface currents offshore of the delta tend to flow in the same general direction as the synoptic and mesoscale winds, from the north or northeast in winter and from the southwest during open waterseason.The typical summerwind frequently produces downwelling and shoreward transport of water, which results in a raised water level and increased wave energy near the coast.

Norton Sound

The currents in Norton Sound are dominated by regional wind and surface pressure patterns. The highest observed flow was measured at about 50 cm/s; flow decreased with increasing depth (Muench 1981). These atmosphere-driven flow events may differ from the mean flow and produce uncertain, intermittent variability in the circulation pattern. Oceanographic data from the mouth of Norton Sound indicate a net northward water transport, with strong seasonal differences in movement rates. Currents between the mouth of the sound and St. Lawrence Island to the west are characterized by somewhat pulsive north-south flow events having speeds of 50-100 cm/s (Muench, Pearson, and Tripp 1978). These speeds contrast with reported mean flow rates of 15 cm/s observed in relative synchrony with major meteorological events. The mean circulation pattern within the sound iscyclonic incharacter(Drakeet al. 1980). A typical feature is westerly flow of water mass, varying in extent and intensity over time, along the northern coast (Cline, Muench, and Tripp 1981). The tidal component in the sound is on the order of 50 cm/s and reverses either diurnally or semidiurnally. The reversals are roughly northeast/southwest within Norton Sound.

The upper- and lower-layer circulation is decoupled in the eastern sound, but less so in the western sound, where there is a monotonic decrease in speed along with a slight rotation of flow as depth increases. Northwesterly surface flow rotates to westerly near the bottom. In summer, easterly flow enters the sound along its southern shore, curves cyclonically to the north, and is then deflected to the west at the north coast, roughly following the bathymetry. This flow varies in intensity and extent from year to year. In the summer of 1979, a westerly mean flow paralleled the coastline and was superimposed upon a highly variable flow which included reversals (Muench 1981).

Bering Strait

The Bering Sea is characterized by an open shelf south of St. Lawrence Island. Mean currents are variable in direction and range from 1 to 4 cm/s, with the tidal current accounting for 55 (± 31)% of the fluctuation (Coachman, Salo, and Schumacher 1983). Near St. Lawrence Island, the Bering Sea narrows into two straits, the Shpanberg and Anadyr. North of the island the two straits merge to form the Bering Strait. Circulation here is dominated by a northward mean flow ranging from 4 to 15 cm/s, with very small tidal influences 24 (± 13)% variability (Coachman, Salo, Schumacher 1983). Flow in both the Anadyr and Shpanberg is to the north, approximately parallel to the local bathymetry. The flow appears to come from around both ends

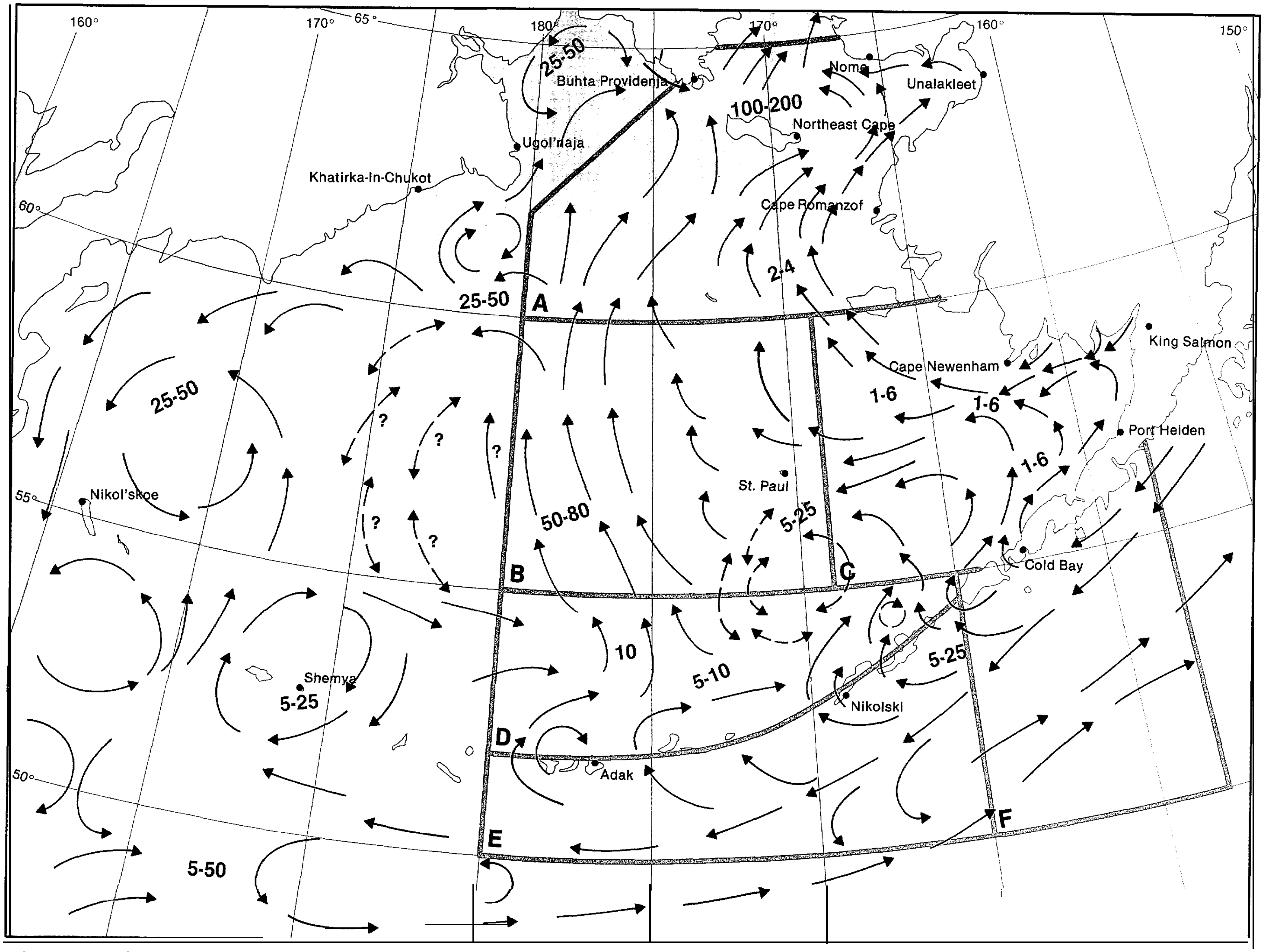


Figure 4. Bering Sea Currents-Summer

Legend

Bering Sea surface currents. Numbers indicate mean speed in cm/s. Arrows depict flow as follows:

Prevailing current direction

Variable current direction

Bering Sea surface currents synthesized from Arsen'er 1967; Goodman et al. 1942; Kinder and Schumacher 1981; LaBelle 1983; Marine Advisory Program, University of Alaska: Notorov 1963; Pelto 1981; Takenouti and Ghtahi 1974; and U.S. Navy 1977.

of St. Lawrence Island. Frequent reversals are coincidental with meteorological events. These reversals can affect the flow over vast regions covering thousands of square kilometers. The presence of ice appears to dampen the impact of wind stress forcing. The major driving force for the northward flow through Bering Strait is the sea surface sloping down to the north (Aagaard and Coachman 1966). A slope of 2×10^{-6} is associated with average summer northerly transport of approximately 1.6 x10” m^3/s. The normal condition is, thus, one in which sea level in the southern Chukchi Sea (in summer) is about 0.5 m lower than in the northern Bering Sea. A major cause of variations in the sea level difference must lie in fluctuations of the regional wind distribution. It is also possible that the atmospheric pressure field may itself directly modify the oceanic pressure field (Aagaard, Coachman, and Tripp 1975).

An examination of recent meteorologic data (Aagaard and Coachman 1981) showed the following results. In every case of southerly flow through the Bering Strait, the large-scale atmospheric pressure patterns were the same. One day

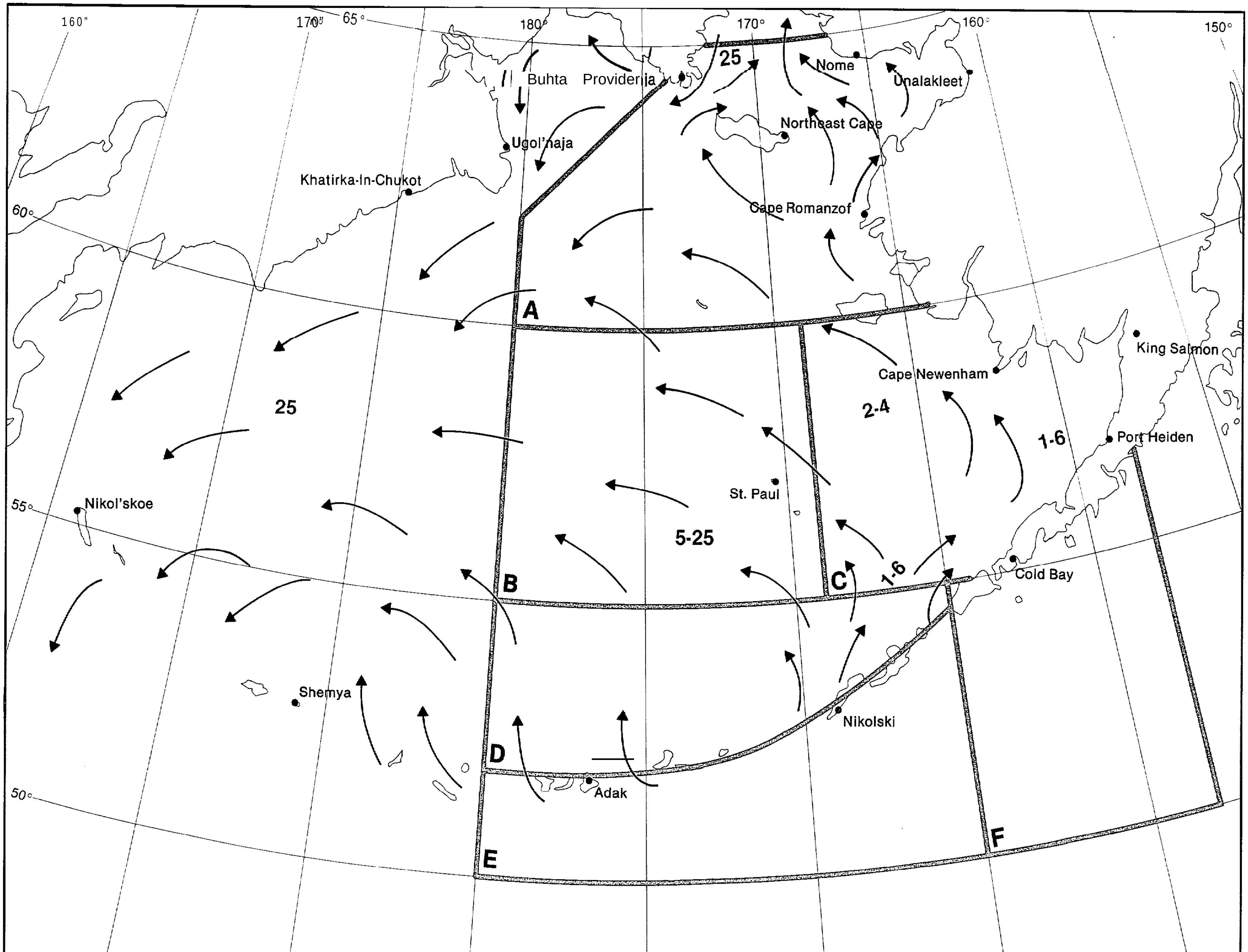


Figure 5. Bering Sea Currents-Winter

before a peak in southerly flow, a strong low-pressure system was centered some distance to the southeast of Bering Strait in the area of Bristol Bay, Kodiak, Anchorage, and the northern Gulf of Alaska. At the same time the Siberian high was centered some distance west or west-northwest of the strait. The isobars signifying the strongest pressure gradient between pressure centers were located precisely over the Bering Strait region. Most significantly, they had a nearly north-south orientation which extended from over the Chukchi Sea south into the central Bering Sea-completely across the northern Bering Sea shelf. If the north-south orientation of the isobars did not extend totally across the northern shelf or if the isobars were oriented northeast-southwest (the nearest typical configuration), strong southerly flow events did not occur.

The mechanism which drives major south flow events now seems clear. Strong north winds must develop over the entire northern Bering Sea, not just over the immediate region of Bering Strait. Large-scale, strong atmospheric pressure cells are required: a low far to the southeast and a high well to the west. The strong northerly winds generated thereby move water southward off the entire northern Bering Sea shelf. Removal of sufficient water off the northern shelf generates a sea-level slope down to the south-sea-level slope has been shown to be the major force driving transport through the strait (Coachman et al. 1975). This, together with the strong north winds caused by the east-west atmospheric pressure gradient, drives enhanced southerly transport. These conditions apparently require about one day to develop, so that maximum south transport occurs the following day. Because the system behaves to a marked degree as a coherent unit, water levels at both St. Lawrence Island and Cape Lisburne fall together and are nearly in phase with the transport.

Northward transport stands in contrast to the southerly transport events. Periods of northerly flow tend to be more persistent and not so great in magnitude, nor do they show the marked episodic character of the southerly flows. The greater persistence of northerly flow must reflect the basic driving force, a higher sea level in the Bering Sea than in the Arctic Ocean (Coachman et al. 1975), which still remains unexplained. There were, however, a number of relatively rapid northward accelerations of transport during the seven months of record which appear to have two basic causes:

(1) After strong south transport events, rapid accelerations commonly occur which can be thought of as compensatory. When atmospheric conditions causing the southerly transport event dissipate, water is not being removed from the northern Bering shelf, but there is still voluminous southerly transport in the system. Water "piles up" in the region around St. Lawrence Island and Norton Sound, a condition reflected by a strong, positive difference in water level. Following this by about one day, a strong northward acceleration occurs.

(2) Occasionally, major northward accelerations appear to be, at least in part, directly driven by atmospheric conditions. Specifically, these are a strong low pressure centered in the western Bering Sea southwest of Bering Strait, or a deep trough from the central Aleutians toward the northwest, so that the isobars in the strong pressure gradient are directed northward from the central Bering Sea along the axis of the system. This configuration creates strong, southerly winds which can move water from the central Bering Sea onto the northern Bering Sea shelf, raising the water level in the vicinity of St. Lawrence Island and enhancing the sea-level slope down to the north.

Central Bering

West and northwest of the North Aleutian Basin and Yukon Delta lies St. George Basin, the Central Bering Sea, and still further west, the Navarin Basin. Circulation in these regions is not as well understood as in the coastal basins. Fewer studies have been conducted in the offshore Bering. Data are site-specific and sporadic over decades. No consistent flow patterns have emerged as representative of the regional circulation. In fact, there is little consensus among investigators that the principal flow is north-south, east-west, or cyclonic or anticyclonic in nature (See Natarov 1963; Arsen'ev 1967; Tak enovti and Ohtani 1974; Goodman 1942; Ratmanov 1937). The northward flowing, eastern boundary current is roughly balanced by a southward flow along the Soviet coast. Within the central region, flow is probably dominated by the location and strength of large-scale atmospheric pressure cells. Response times, directions, and persistence are probably of a similar scale as those controlling flow through the Bering Strait (Aagaard and Coachman 1981). Thus, a dominant regional flow pattern is not readily observed nor easily quantified.

Sea Ice

Introduction

The annual cycle of formation and dissipation of sea ice in Alaska waters has widespread effects on a number of phenomena. When the ice forms, the coastal climate changes in character from maritime to continental with much colder temperatures and lower humidities than would be the case if open water were present. The ice also interferes and even stops water transportation with the possible exception of icebreakers and otherspeciallydesigned ships. It makes the cleanup of oil spills difficult, if not impossible, by hampering the operation of cleanup equipment and by trapping oil under the ice. Sea ice also has important effects on the life cycles of living creatures in and near the sea.

In the Bering Sea, the sea ice generally begins as fast ice formation along the shores of the Seward and Chukotsk peninsulas in October. As the season progresses and waters in the more open portions of the Bering Sea cool off, the pack ice generally begins its seasonal southward formation in November. An estimated 97% of the ice in the Bering Sea is formed within the Bering Sea(Leanov 1960); very little is transported south through the Bering Strait. During periods of increasing ice and prevailing northerly winds, the ice apparently is generated along the south-facing coasts of the Bering Sea and moves southward with the wind at as much as 1 knot before melting at its southern limit (Pease 1971). During periods of southerly winds, ice coverage generally decreases in the Bering. Prevailing winds can persist in one direction for weeks at a time in winter in the Bering Sea, causing a wide variation in ice cover from month to month and from year to year(see Figure6 and map set 17, Section II of this volume).

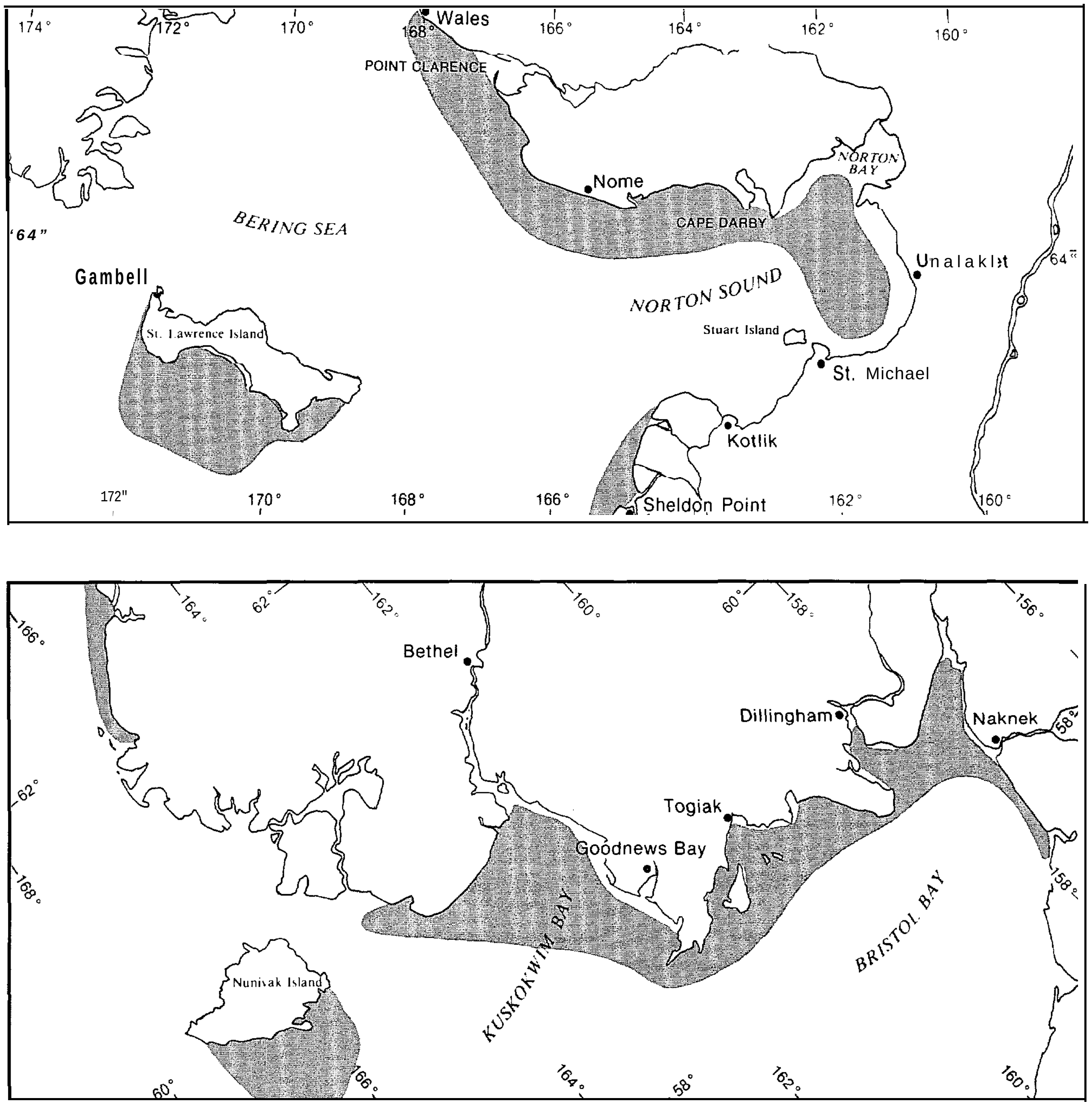


Figure 6. Recurring Polynyas

Synthesized from: McNutt 1981; Stringer, Barrett, and Schreurs 1980; Wohl 1982.

Recurring Leads and Polynyas

Wind and current stresses on the ice can cause tension or divergence and open relatively narrow, long stretches of open water in an otherwise dense ice cover. In the absence of strong currents, the wind induces leads which run perpendicular to the wind direction. Flaw **leads** generally occur just seaward of the stable fast ice zone when strong offshore winds develop.

In the Bering Sea a wind-induced polynya (Figure 6) immediately south of St. Lawrence Island is a frequent but undependable feature (McNutt 1981; Wohl pers. comm.). Northerly winds cause the polynya to form in the lee of the island as sea ice is advected to the south. The polynya can extend more than 160 km and is frequently covered with thin ice. However, the feature is temporal, and a wind shift to southerly flow can close this area rapidly. At such times, a corresponding polynya to the north of St. Lawrence Island is sometimes observed, but it is generally much smaller and occurs less frequently.

A polynya can form on any side of Nunivak Island, depending upon the prevailing wind direction. Usually the feature is located to the north or south, under southerly or northerly winds, respectively. Like the polynya off St. Lawrence Island, the appearance of this polynya is variable, but it is usually observed at least once each year, often more. Its extent is variable, **and** thin ice commonly covers the polynya quickly during cold, northerly wind storms.

The polynyas shown for Norton Sound, Bristol Bay, and Kuskokwim Bay were taken from Stringer, Barrett, and Schreurs (1980). These features were mapped from LANDSAT scenes collected between 1973 and 1976. Generally, the major polynyas in these areas open in response to northerly winds, which cause all but landfast ice to move toward the south. As the polynyas are opened by the wind, new ice forms and is, in turn, advected southward. This mechanism for new ice production can be very efficient under the proper circumstances (Pease 1980). None of the polynyas can be considered even semipermanent since a reversal in the wind direction can completely close them. Furthermore, many of the areas shown are partially covered with very thin ice when the northerly winds bring below-freezing temperatures.

Fast Ice and Shear Zones

According to World Meteorological Organization sea ice nomenclature, fast ice includes all ice that has become attached to the shore, even multiyear pack ice. A common feature at the seaward boundary of the fast ice is an area of shear ridges. Shear ridges in the Bering Sea tend to be more localized and of lesser extent and magnitude than farther north. Figure 7 shows the various kinds of ice near shore. Bering Sea ice does not have any multiyear ridges. The accompanying fast ice boundary maps, from the Alaska Marine Ice Atlas, were synthesized from Stringer, Barrett, and Schreurs (1980).

Any significance accorded to trends apparent on these maps must be tempered by consideration of the variability exhibited in the ice-edge data. At some locations, the edge of the fast ice varies considerably in position during each period. Although the average edges along the coast show a temporal trend, it has only minor significance. In other locations, the variability of the fast-ice edge of each period is small compared to the changes in the average position from period to period (Stringer 1981). The intraseason and interseason variability of the fast-ice edge are very dependent on the meteorology and associated wind patterns as well as the offshore bathymetry. Although the prevailing winds in winter are generally northeasterly, there are often periods of a week or more with southerly winds. During northeasterly winds, shore leads and polynyas open up, only to be closed again when winds shift to south or southwest. Also, the high tide ranges in Bristol Bay and along the coast south of Norton Sound tend to break up extensive areas of fast ice, except where it is grounded on mud flats or offshore shoals.

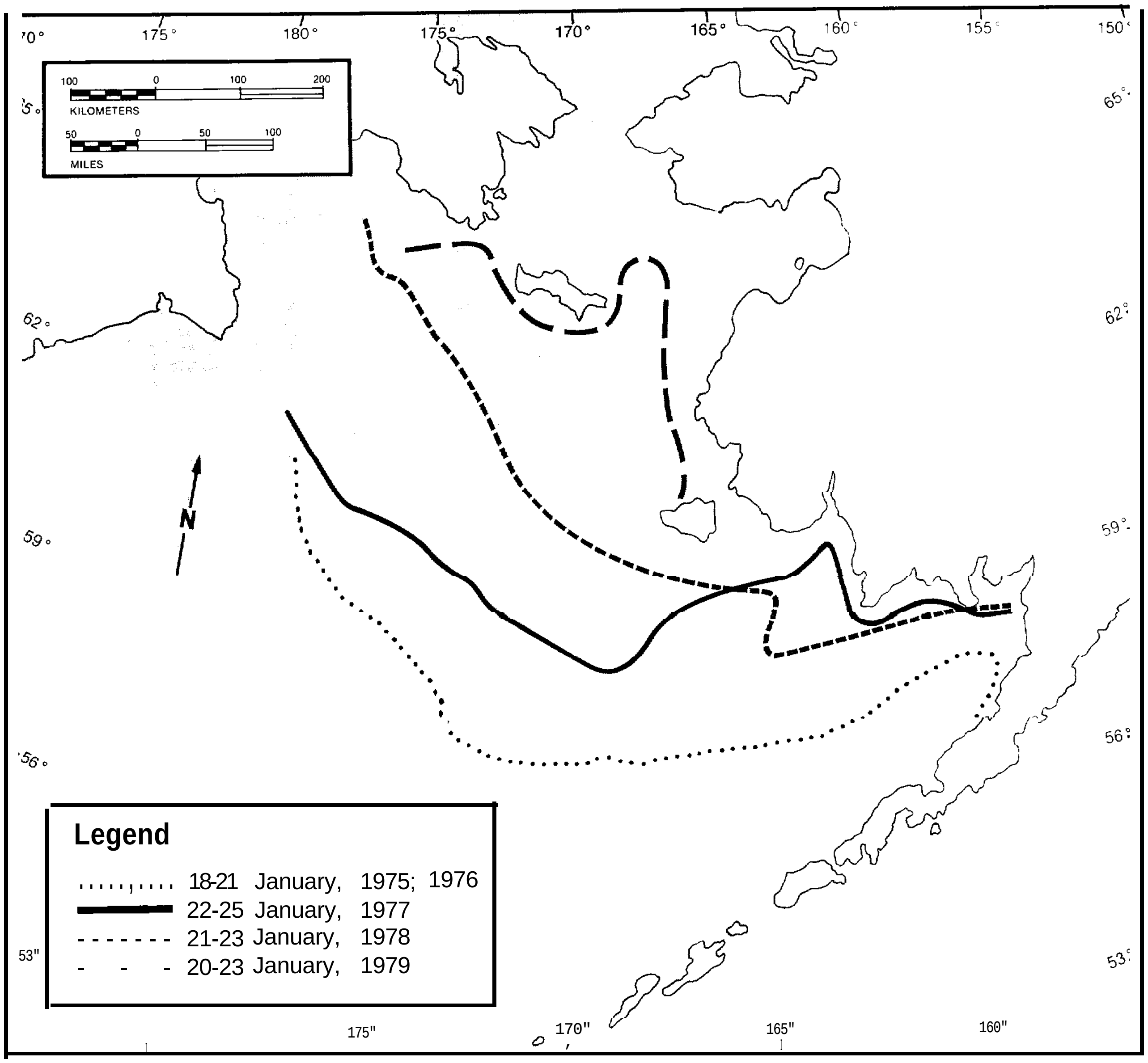


Figure 8. January Southern Ice Limit for 1975-1979. Source: Niebaur 1981.

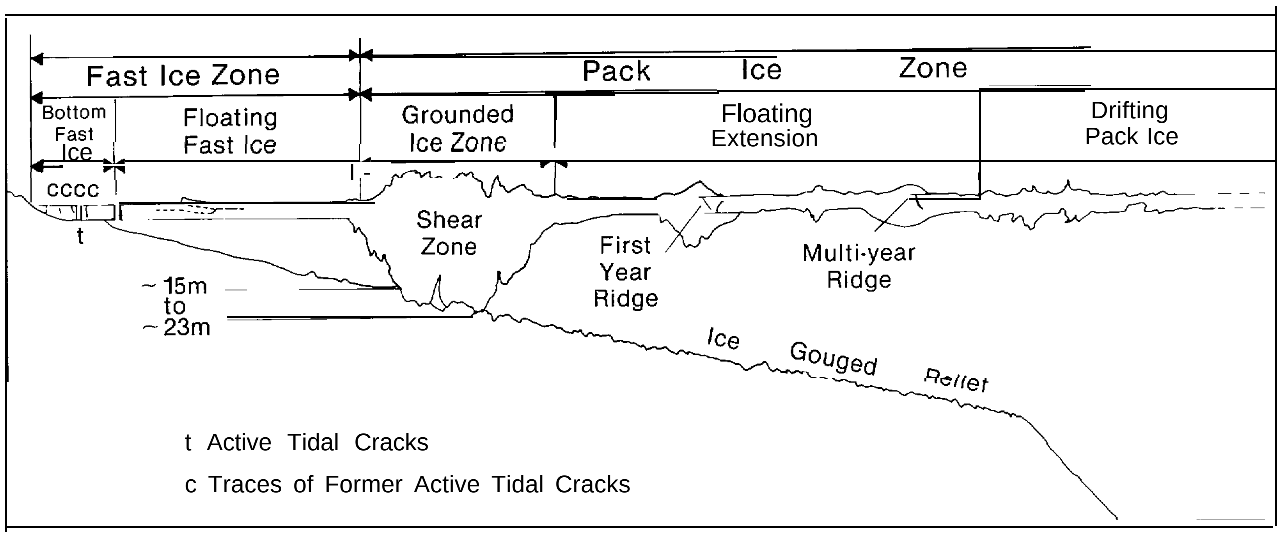


Figure 7. Sea Ice Zones and Types

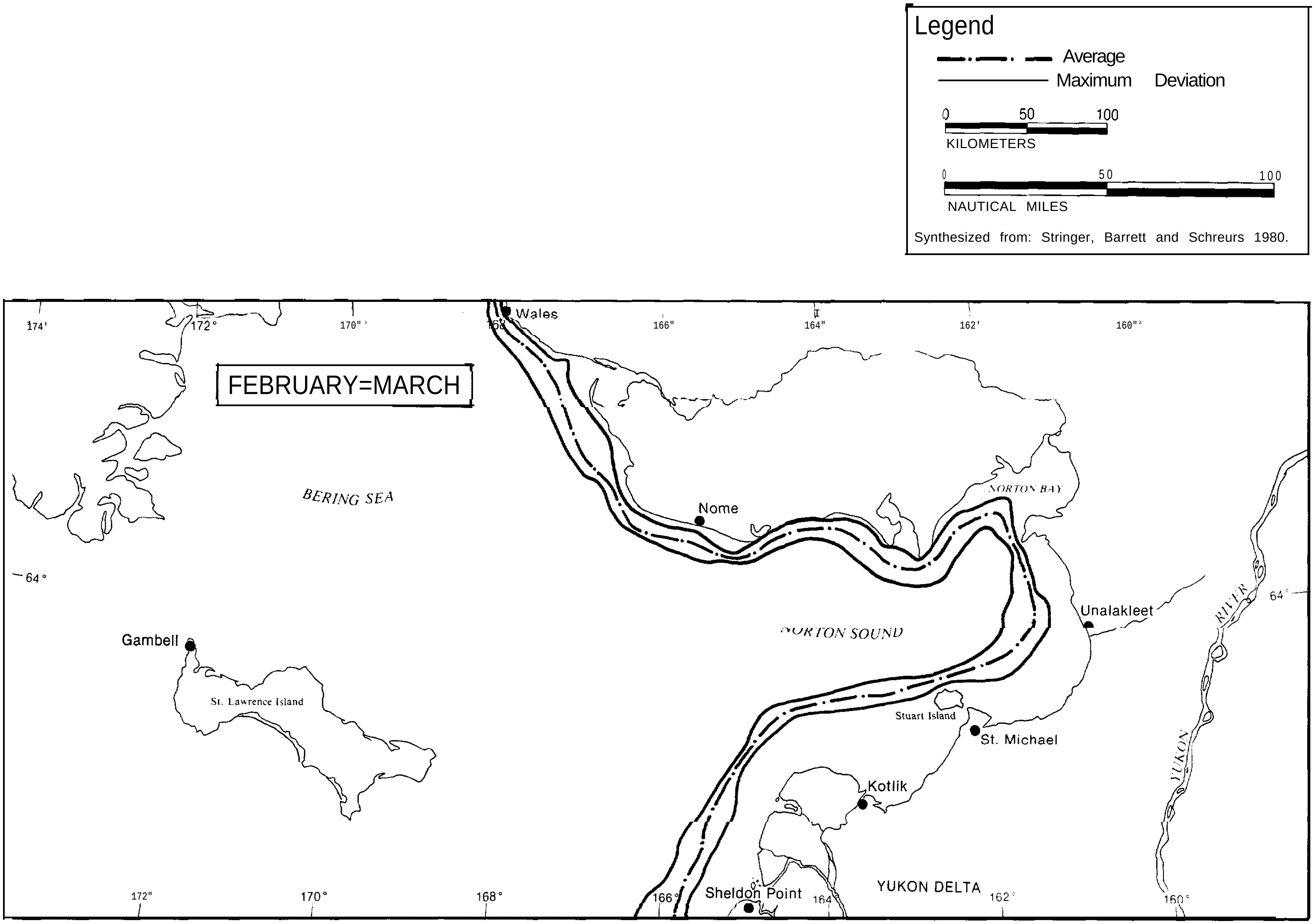


Figure 9. Seasonal Fast Ice Boundary — Norton Sound (February/March)

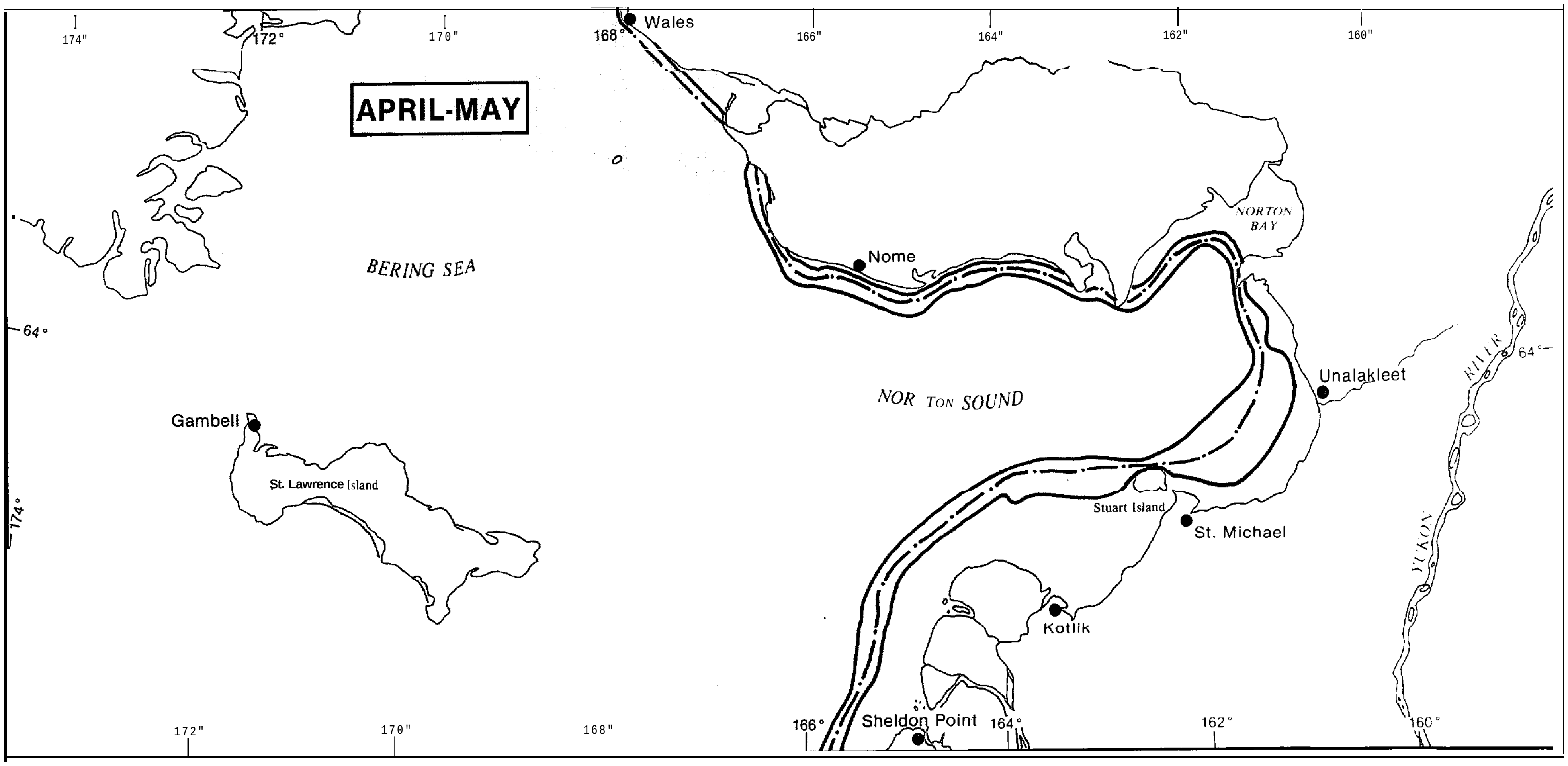


Figure 10. Seasonal Fast Ice Boundary-Norton Sound (April/May)

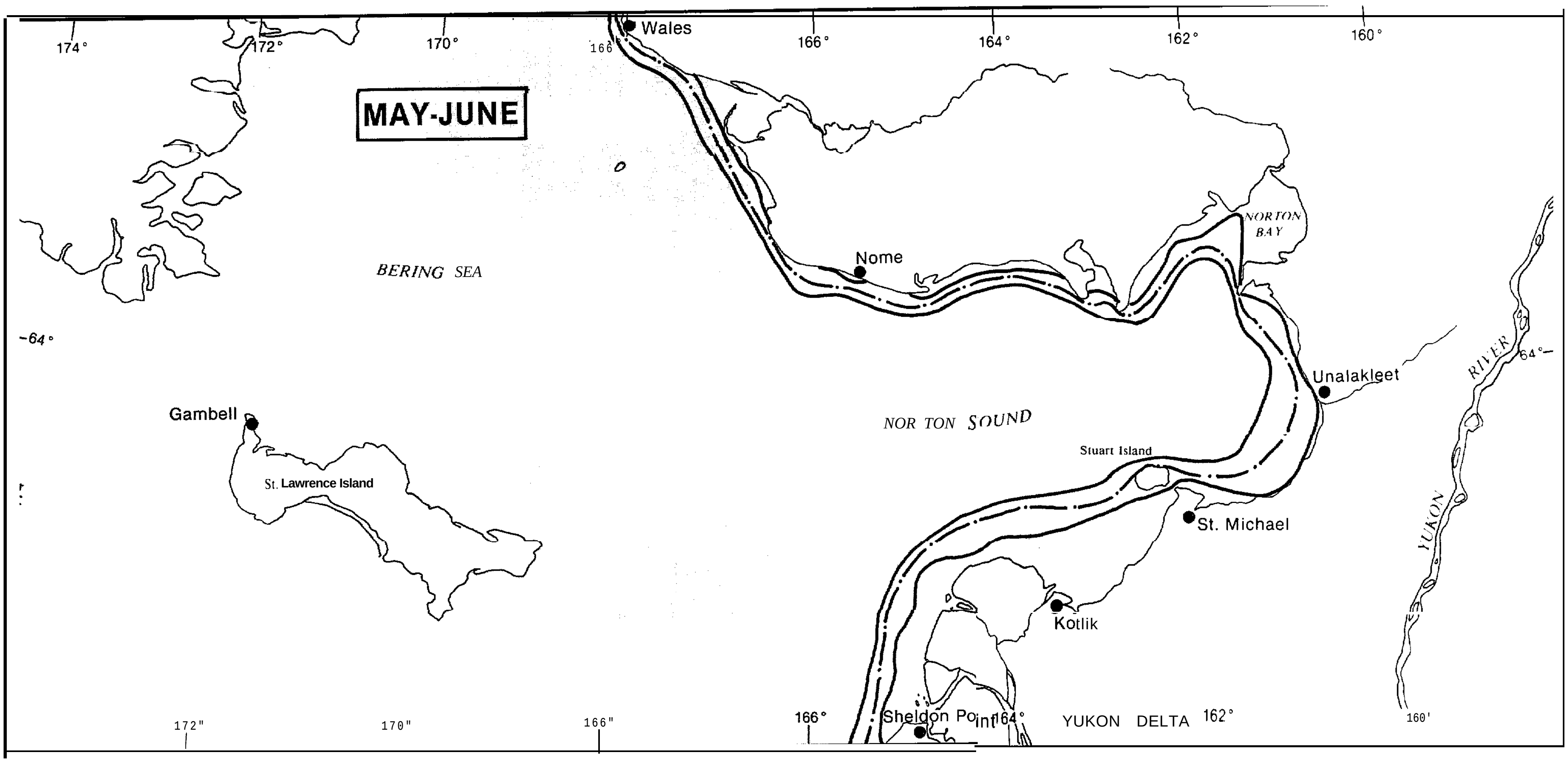


Figure 11. Seasonal Fast Ice Boundary-Norton Sound (May/June)

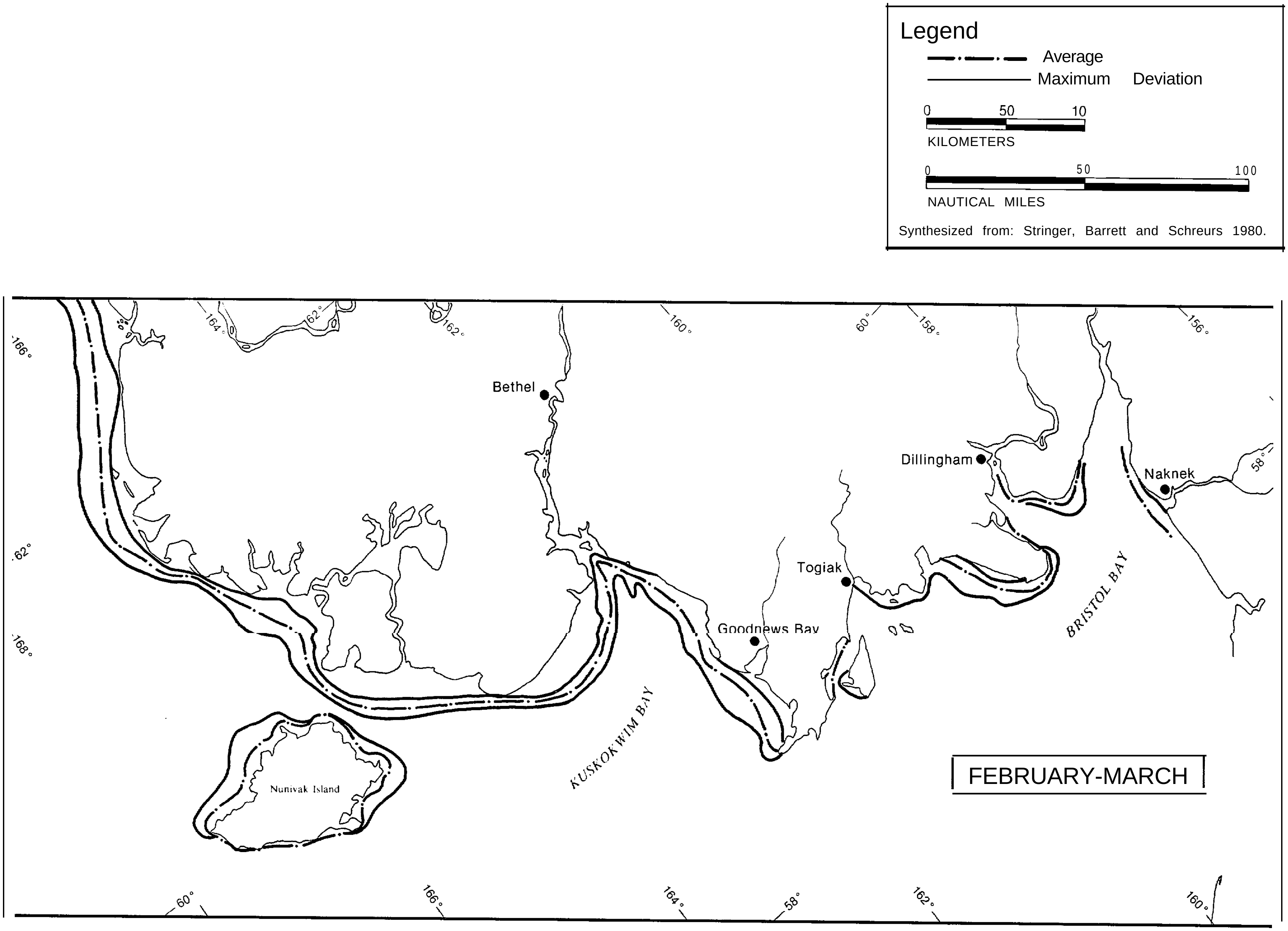


Figure 12. Seasonal Fast Ice Boundary -Southeast Bering (February/March)

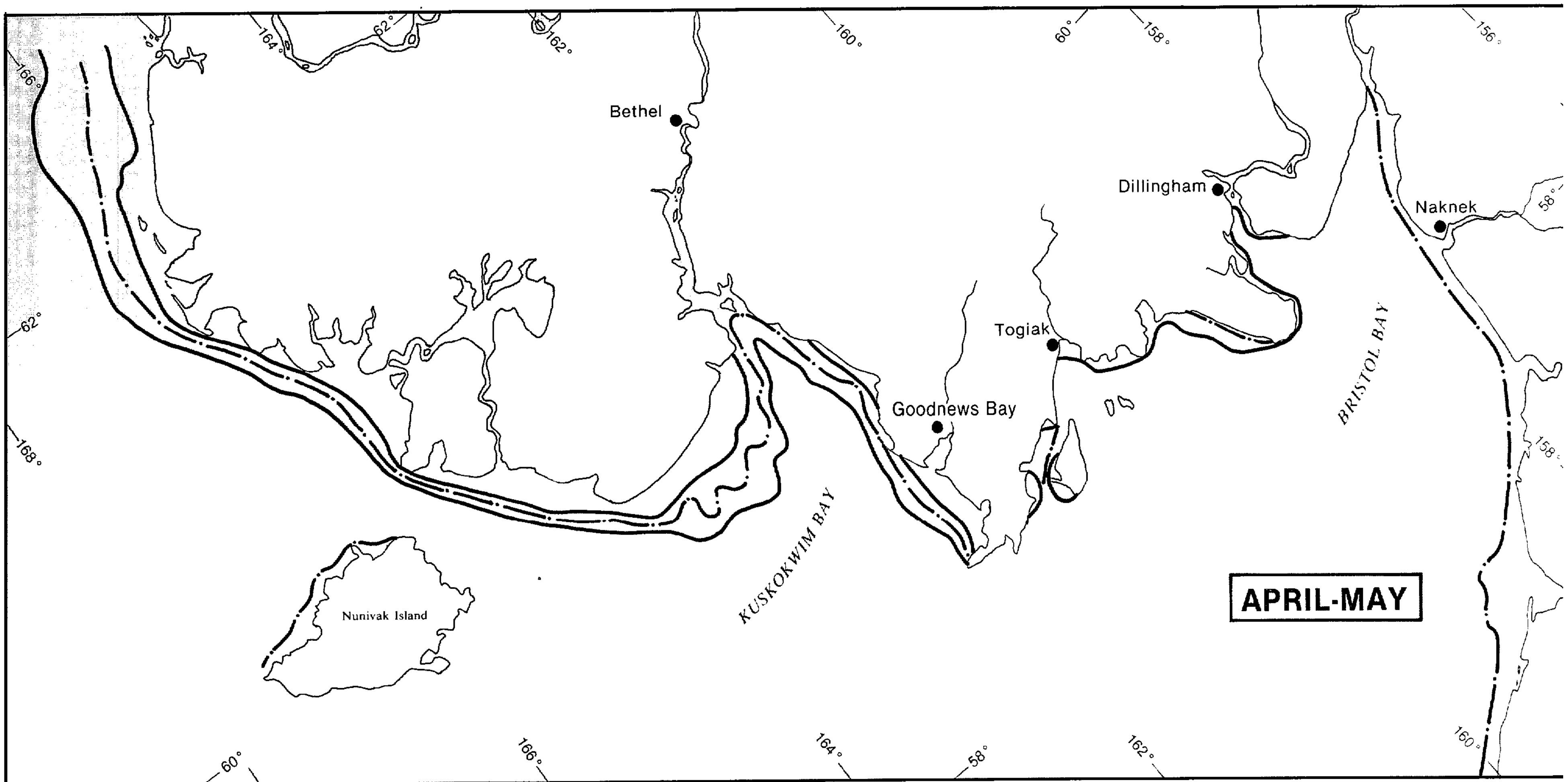


Figure 13. Seasonal Fast Ice Boundary— Southeast Bering (April/May)

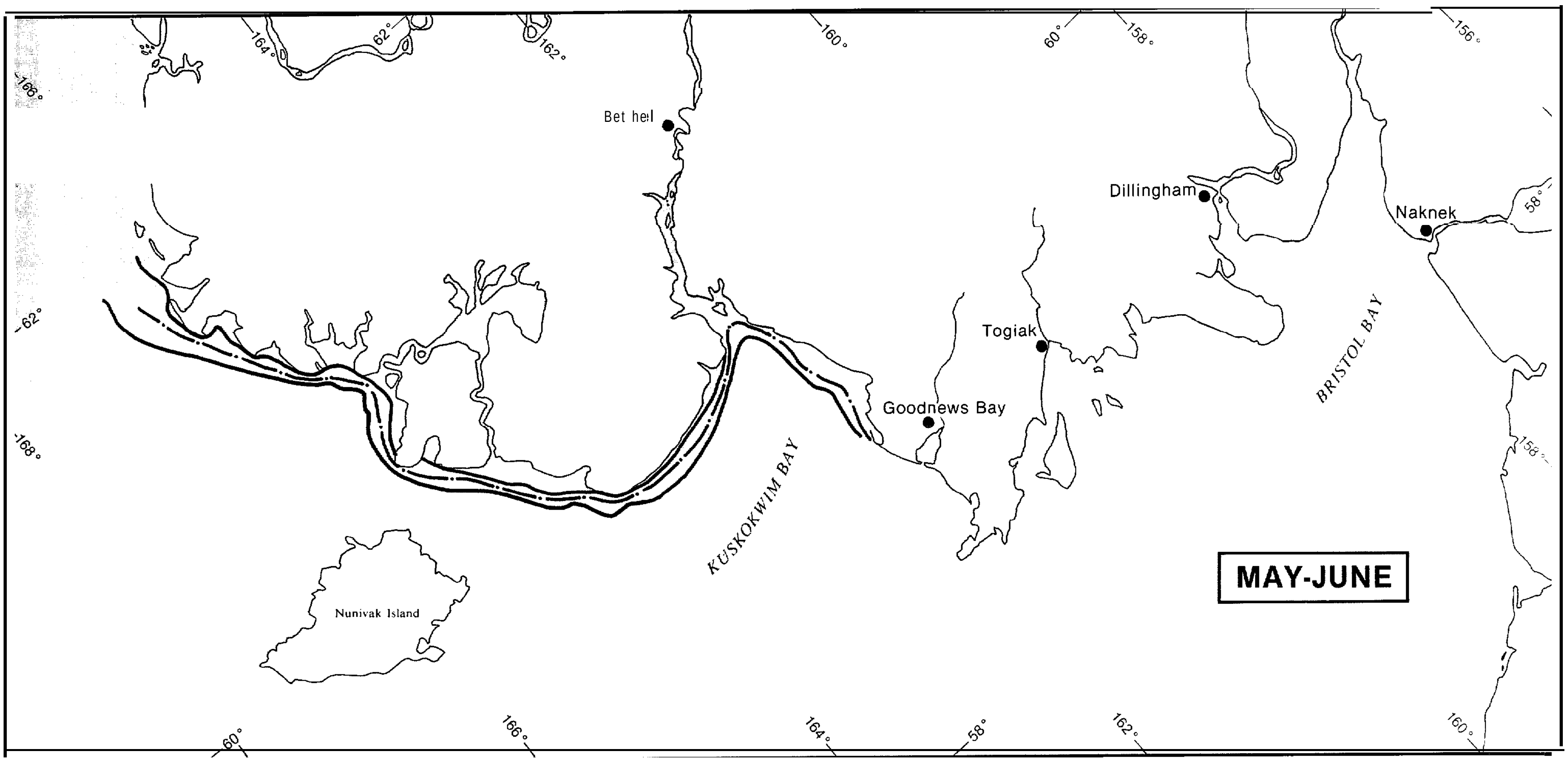


Figure 14. Seasonal Fast Ice Boundary-Southeast Bering (May/June)

Tides

The practical study of tides, aimed at predicting surface elevations and times, involves the empirical treatment of observations made at the desired location over an extended period of time. The motion of the heavenly bodies, particularly the sun and the moon, relative to the earth is known with great precision, so the tide generating potential at any place and time can be computed. Mathematically the potential can be resolved into a finite number of strictly periodic components which, upon addition, produce the total potential of hundreds of tide-generating components that are listed by various authors. Many of the components are of insignificant amplitude and can be excluded from consideration. In practice only seven components are widely used: four semidiurnal (M_2, S,, N,, K,) and three diurnal components (K,, O_1, P_1) (McLellan 1965). The names and relative weights of these components are shown in Figure 15.

Theoretical models of tides must beverified on the strength of observations at tidal stations where the tide wave has been distorted through passage over a continental shelf of complex topography. The nature of tides in a particular area is highly dependent on the bathymetry, shape, and direction of the coastline and latitude. In the Bering Sea in the last 10 years there have been a large number of presuregauge and current meter observations taken so that tides can be modeled with a high degree of acuracy. The following discussion paraphrases material contained in Pearson, Mofjeld and Tripp, 1981 in which theoretical tide model results are compared to observations. The tides most concerned with are the principal tidal constituents N_2 and M_2 in thesemi-diurnal band and 0 and K in the diurnal band. Ordinarily the S_2 would be included in thediscussion, however, S_2 is anonymously small throughout the Bering Sea, possibly because it has small amplitudes in the adjacent North Pacific Ocean. The complicated distributions of semi-diurnal and diurnal tides in the Bering Sea produce a rich variety of tidal types, ranging from fully semi-diurnal in some regions to fully diurnal in others.

The tide wave enters the Bering Sea as a progressive wave from the North Pacific Ocean, mainly through the central and western passages of the Aleutian-Komandorski Islands. The Arctic Ocean is a minor secondary source of tides which propagete southward into the north Bering Sea where they complicate the tidal distributions.

Tides in the Bering Sea are considered to be the result of cooscilation with large oceans. Once inside the Bering Sea, each tidal constituent propagates as a free wave subject to Coriolis effect and bottom friction.

SYMBOL	NAME OF PARTIAL TIDE	COEFFICIENT RATIO
M_2	Principal Lunar	100.0
S_2	Principal Solar	46.6
N_2	Larger Lunar Alliptic	19.2
K_2	Luni-Solar Semi-Diurnal	12.7
K_1	Luni-Solar Diurnal	58.4
O_1	Principal Lunar Diurnal	41.5
P_1	Principal Solar Diurnal	19.4

Contracted from table 15.1,
Elements of Oceanography (McLellan 1965).

Figure 15. Major Tide Components

The tide wave propagates rapidly across the deep western basin. Part of it then propagates onto the southeast Bering shelf where large amplitudes are found along the Alaska Peninsula and in Kvichak and Kuskokwim Bays (Figure 17). Another part propagates northeastward past St. Lawrence Island and into Norton Sound. Over most of the Eastern Bering Shelf region the tide is mainly semi-diurnal, but in Norton Sound diurnal tides predominate. Over the remainder of the Bering tides tend to be mixed. In the Aleutians diurnal rather than semi-diurnal components are stronger.

Legend

Diurnal Range

Max diurnal	Min diurnal
Max tide	Min tide

Diurnal range is the average difference in height between mean higher high water and mean lower low water in feet on a single day.

Max diurnal and Min diuranl are the maximum and minimum differences in feet respectively between the higher high water and lower low water that are predicted to occur during the year.

Max tide is **the** highest tide predicted to occur at the location in feet above the datum level generally taken to be the mean of the lower of the two low waters of each day.

Min tide is the lowest tide predicted to occur at the location in feet above the datum level generally taken to be the mean of the lower of the two waters each day. A negative number indicates a level below the datum level.

Prepared by AEIDC from West Coast *of North and South America Tide Tables, High and tow Water Predictions, 1986, West Coast of North and South America,* NOS/NOAA, 1985.

Station	Diurnal range	Max diurnal	Min diurnal	Max tide	Min tide
1 Ogliuea Island	3.5	7.1	0.2	5.5	-2.2
2 Hot Springs Bay	3.1	6.3	0.2	4.9	-2.0
3 Sweeper Cove	3.7	7.5	0.2	5.8	-2.3
5 Martin Harbor	3.2	6.5	0.2	5.0	-2.0
4 Shoal Point	3.2	4.2	0.2	5.0	-2.0
6 Inanudak Bay	3.7	6.4	0.3	4.9	-2.0
7 Chernofski Harbor	3.8	6.3	0.1	5.1	-1.8
6 Kashega Bay	4.0	6.2	0.2	5.3	-1.7
9 Dutch Harbor	3.7	6.4	0.1	5.0	-1.9
10 Cape Sarichef	5.0	7.4	1.1	6.3	-1.6
11 Port Moller	10.8	17.8	5.9	13.4	-4.8
12 Sand Point	17.3	10.4	2.5	9.3	-2.5
13 Port Heiden	12.3	19.9	6.6	14.9	-5.2
14 Egegik R. Entrance	18.2	26.1	11.0	22.1	-4.6
15 Naknek R. Entrance	22.6	31.6	11.0	26.9	-4.7
16 Naknek Air Base	3.2	3.4	2.0	3.1	-0.4
17 Kvichak	16.5	21.3	10.9	19.9	-1.8
16 Nushagak Bav	19.5	28.4	12.0	24.0	-5.0
19 Goodnews Bav	8.5	11.9	2.6	11.3	-0.6
20 Kuskowak Creek	12.2	17.3	4.4	16.8	-0.5
21 Bethel	4.0	1.2	0.3	1.0	-0.0
22 Village Cove	3.2	5.5	0.0	4.3	-1.6
23 St. Matthew Island	2.1	3.7	0.1	2.9	-1.1
24 Northeast Cape	1.7	6.6	0.1	2.9	-4.0
25 Meghoweyik River	1.7	2.8	0.6	2.1	-0.6
26 Cape Romanzoff	6.8	11.2	2.3	8.6	-2.3
27 Kwikluak Pass	2.3	2.5	1.0	2.1	-0.3
26 Apoon Mouth	4.0	6.8	0.9	5.8	-1.0
29 St. Michael	3.9	6.7	0.9	5.7	-1.0
30 Carolyn Island	1.8	3.1	0.4	2.6	-0.5
31 Nome	1.6	1.6	0.4	2.1	-0.6

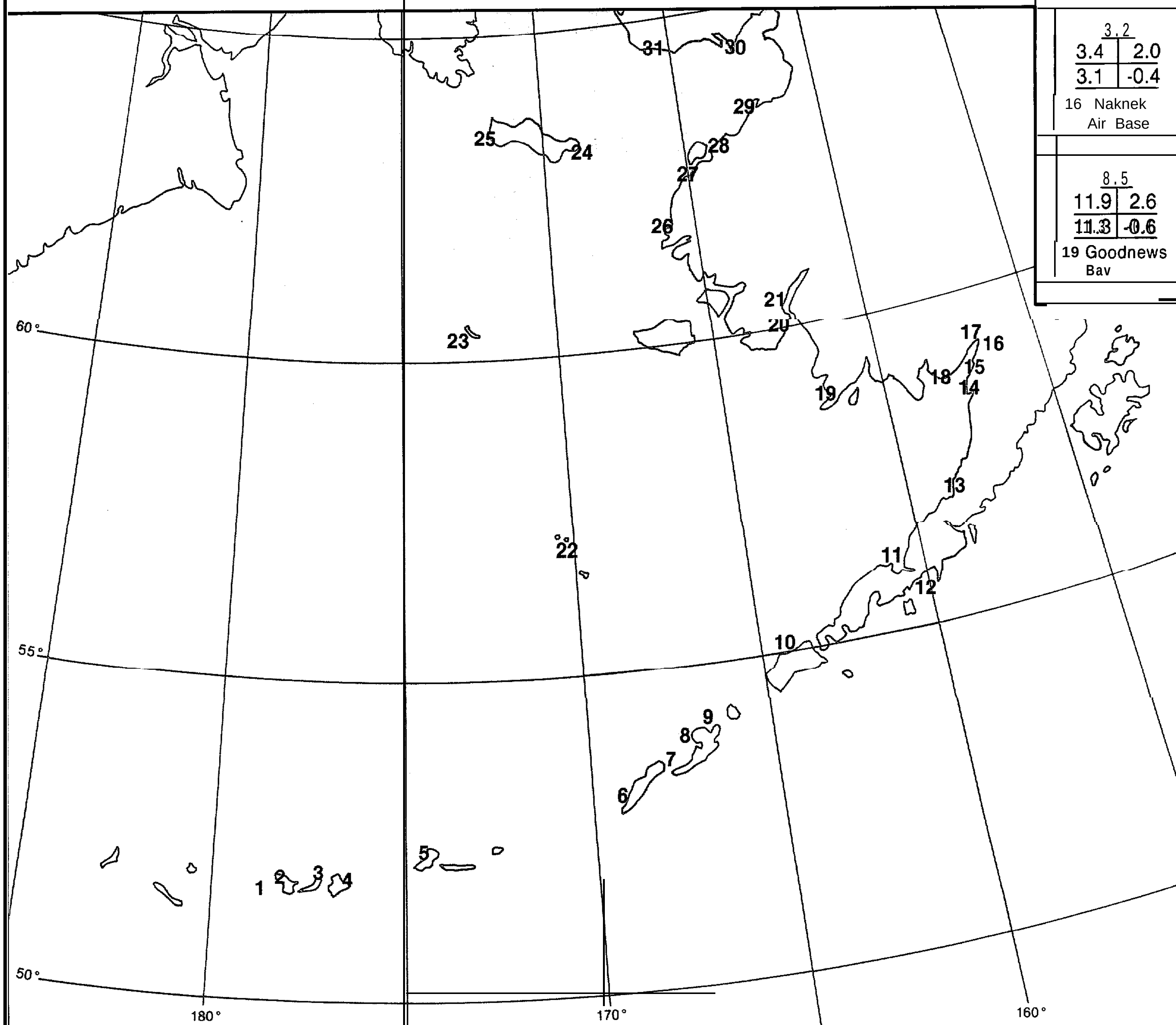


Figure 16. Tide Data

Legend

Type of Tide

- Semi-diurnal
- Diurnal
- Mixed
- —8— Corange (Feet)

Adapted from Bering, Chukchi, and Beaufort Seas Coastal and Ocean Zones Strategic Assessment Data Atlas, Prepublication Edition, and U.S. Navy, 1972.

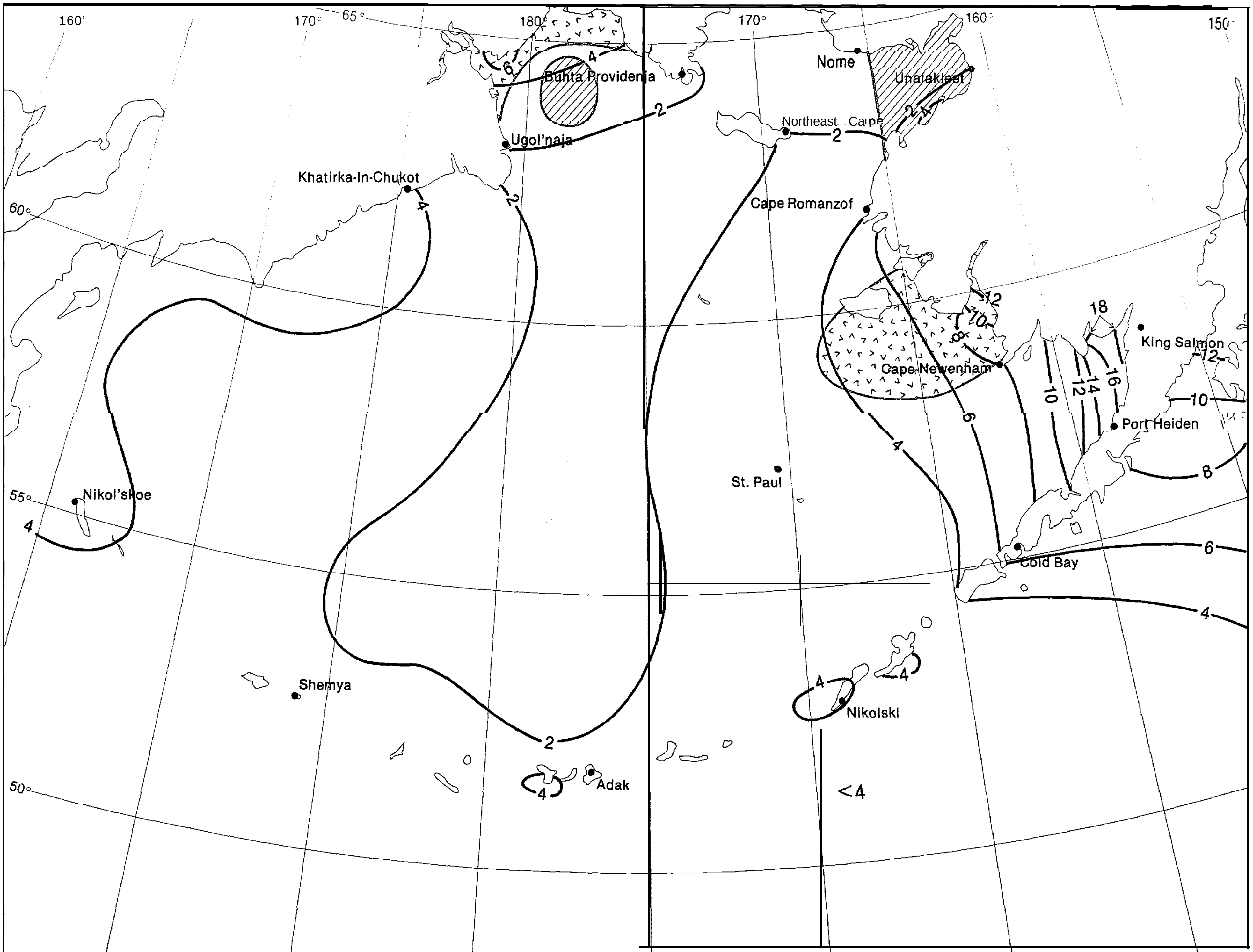


Figure 17. Tide Co-Range

Storm Surges

Storm surges are waves oscillating in the period range of a few minutes to a few days, in a coastal or inland water body, resulting from forcing from atmospheric weather systems (Murty 1984). By this definition, wind-generated waves (often referred to as wind waves) and swell, which have periods of several seconds, are excluded. The spectrum of storm surge waves is centered around 10^{-4} cycles per second (CPS), which gives a period of about three hours. However, depending mainly on the topography of the water body and secondarily on other parameters, such as the direction of movement of the storm, strength of the storm, stratification of the water body, presence or absence of ice cover, and nature of tidal motion in the water body, the periods of the water level oscillations may vary considerably. Even in the same water body, storm surge records at different locations can exhibit different periods.

Although storm surges belong to the class known as long waves, as do astronomical tides and tsunamis, there are at least two important differences. First, whereas tides and tsunamis occur on an oceanic scale, storm surges are simply a coastal phenomenon. Second, significant tides and tsunamis cannot occur in an enclosed, small, coastal or inland water body, but storm surges can occur even in lakes, or in canals and rivers. The range or height of a storm

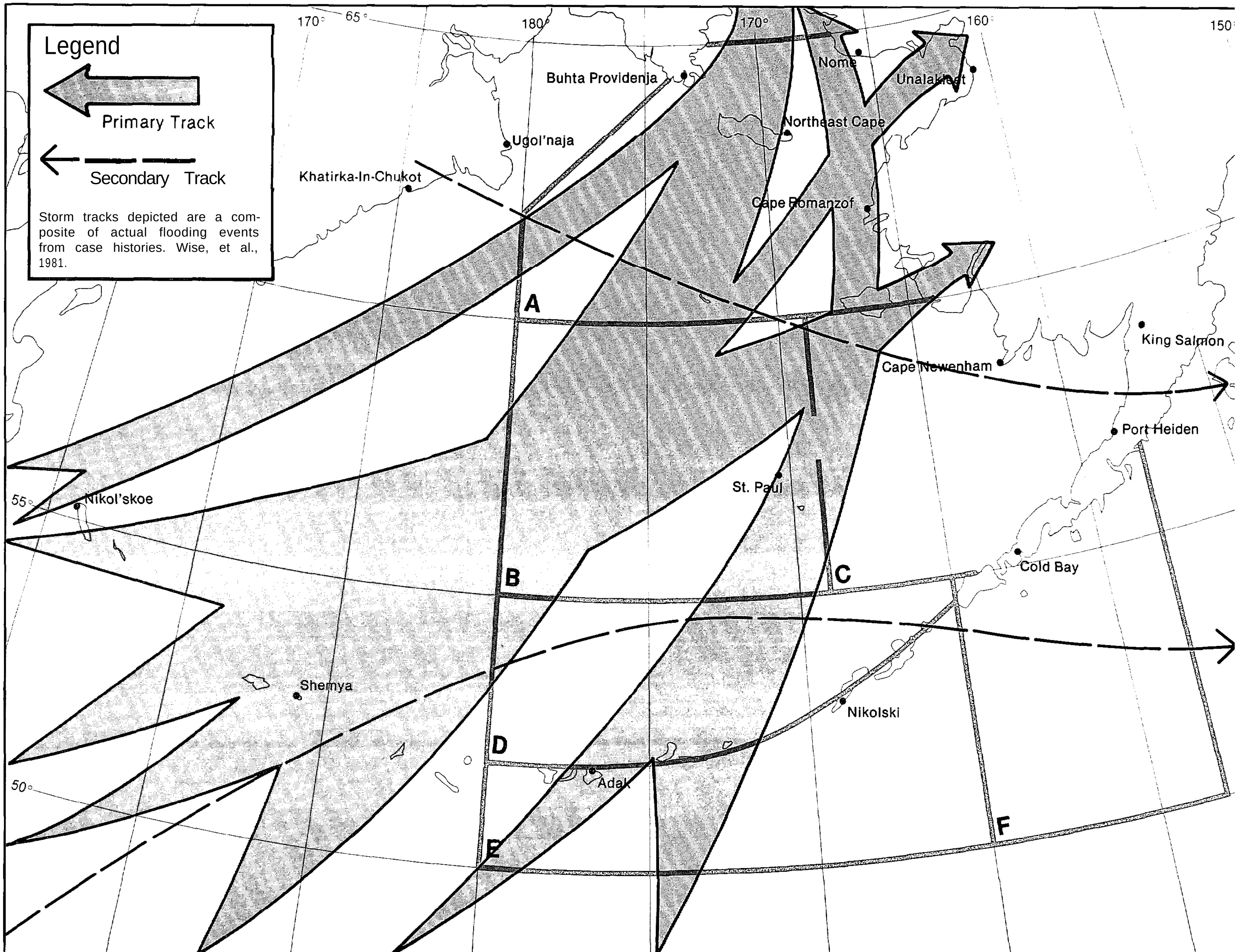


Figure 18. Storm Tracks with Storm Surge Floods

surge depends not only on characteristicsof the storm but also on the topography onshore and bathymetry offshore. Shallow water bodies generally experience surges with greater ranges. Also, the height of a storm surge is less if the sea floor is steep than if there is a shallow slope to the sea floor (Murty 1984). Storm characteristics that effect the height of a surge include atmospheric pressure; wind speed, direction, and length of fetch; the latitude; and the direction and speed of storm movement. Air and water temperature differences also affect the height of surges.

Following is a discussion of the storm surge potentials from astudydone in 1981 (Wise, Comiskey, and Becker), supplemented by storm statistics since then (NOAA Storm Data, 1981-1986) and a modeling study for surges in Norton Sound (Wise, Comiskey, and Becker 1981; Kowalik and Johnson 1985).

Along the southwest and south coasts of the Seward Peninsula the terrain is generally of moderate relief, with the exception of Port Clarence and the east end of Norton Sound. The waters offshore are shallow with a gently sloping sea floor. The open waters of the Bering Sea provide a long fetch for the development of storm waves. Sea ice restricts the development of storm waves from about the first of December to the first of June on the average; however, there can be a high degree of annual variability. Of 13 known flooding events in Nome, all except two occurred in the fall. One destructive storm for which little factual information is available occurred in April 1906, and another occurred in July 1969. Surges above 4 m(12 ft) haveoccurred along this section of coast. The most recent was in November 1974; a 3 m (10-ft) surge brought water into the town over the sea wall. This particular storm caused widespread flooding all along the Bering Sea coast.

With the exception of the Shaktoolik River mouth, which is of low relief and marshy, the coast at the east end of Norton Sound is generally rugged due to the proximity of the Nulato Hills. The south coast of Norton Sound is generally of low relief. The sound itself is shallow, with a gently sloping sea floor that is very favorable for the development of storm surges. The range of wind directions for the development of storm surges is limited to westsouthwest to west. However, the east end of Norton Sound often experiences minor flooding, despite unfavorable winds, due to rising sea levels all over the sound.

A storm surge modeling study(Kowalik and Johnson 1985), determined that for the November 1974 storm the highest surge in Norton Sound was in Norton Bay, with a modeled surge of more than 3 m (10 ft). The same study also determined that negative surges of more than a meter can occur in the eastern end of Norton Sound in the presence of strong, persistent northeast winds in winterwith an ice cover present. The winds tend to move the pack ice away from the shorefast ice and reduce ice cover from 0.7 or 0.9 coverage to less than 0.55 coverage over much of Norton Sound.

Eleven of the twelve storm surge cases at Unalakleet occurred in the fall; the other was in July. Sea ice and shorefast ice limit the fetch for the development of positive storm surges from about the first of December to the first of June.

The shores of Pastol Bay and the north coast of the Yukon River delta do not have long fetches favorable for the generation of waves and storm surges. However, this area experiences surges due to increases in the height of the water in Norton Sound. The remaining coast of the Yukon Delta is exposed to the open waters of the Bering Sea, where conditions are very favorable for the development of storm surges due to low retief onshore, shallow water offshore, and thousands of miles of open sea. Most surges causing property damage occur in the fall or in August. However, early summer surges can be very hard on nesting birds in the salt flats. In June 1963 80% to 90% of the black brandt production was lost due to flooding of the nesting area after eggs were laid.

The coastal area is generally of low relief from the Kuskokwim Delta to Goodnews Bay, with numerous lakes, sloughs, and marshes. From Goodnews Bay to the Nushagak Peninsula the coastline is more rugged due to the proximity of the Ahklun Mountains. The remainder of the coastline of Bristol Bay is similar to the stretch from the Kuskokwim River to Goodnews Bay. Offshore the shape of the sea floor is conducive to the formation and enhancement of storm surges. From Goodnews Bay northward an adequate fetch can be generated with storm winds from south through west to northwest. East of Goodnews Bay, westsouthwest to west are the only directions from which an adequate fetch can develop.

Autumn and late summer are the seasons for destructive storm surge flooding in this area. There are ten known cases of storm surge flooding of populated areas; seven were in autumn and three were in August. Two storms, in November 1979 and in August 1980, account for most of the factual reports of storm surge flooding. The November 1979 storm caused storm surge flooding from Cape Newenham to Scammon Bay. Surges were estimated at 2.5 m (8 ft) in exposed locations in the Kuskokwim Delta. The storm was on a track from westsouthwest to eastnortheast, and a long fetch of more than 640 km (400 mi) developed with the storm. The August 1980 storm, one of the few summer flooding events, caused flooding on the shore of Bristol Bay. The storm was on a track from south-southwest toward north-northeast from near Atka Island, in the Aleutians, to Kuskokwim Bay. Exposed locations showed surge flooding up to 4 m (12 ft).

The coastal area from Hooper Bay to Kinak Bay is a favored nesting area of migratory birds in the spring and summer. Minor storm surges that cover nests in this area at the wrong time can be detrimental to the annual production of several species of birds. Five cases of minor summer flooding of the salt flats were documented in an annual report for the Clarence Rhode National Wildlife Refuge (USFWS 1964). One event (June 22, 1963) caused a loss of black brandt offspring estimated at 80% to 90% of the year's production.

Most of the north shore of the Alaska Peninsula east of Cold Bay is favorable for the occurrence of storm surge flooding, with low, marshy terrain onshore and a moderately sloping sea floor offshore. West of Cold Bay the Aleutian Islands and the south shore of the Alaska Peninsula conditions are not favorable due to rugged terrain onshore and steep ocean floor offshore. The only storm surges discovered in this area were at Meshik, or Port Heiden, and St. Paul. Damage to structures in this area is more likely to be from strong winds and beach erosion caused by wave action than from flooding.

Superstructure Icing

Structural icing on ships, offshore structures, and port facilities isawintertime hazard in open waters and coastal sections of Alaska. The icing causes slippery decks, renders moving parts inoperable, and, in extreme cases, causes uneven loading and raises the center of gravity on small ships. Accumulation of ice on rigging and on deck equipment such as crab pots also increases wind effects because a larger surface area is presented to the wind. Ice forming on structural surfaces above or close to a body of water arises principally from seaspray(Nauman and Tyage 1985; Liljestrom 1985), with lesser amounts from atmospheric precipitation (freezing rain and wet snow) and fog (arctic sea smoke, white frost, black frost). Sea spray, the most dangerous source of icing, is produced by the breaking of waves against obstacles such as ships' hulls, other floating objects, shore structures, and, possibly, other sources (Minsk 1977).

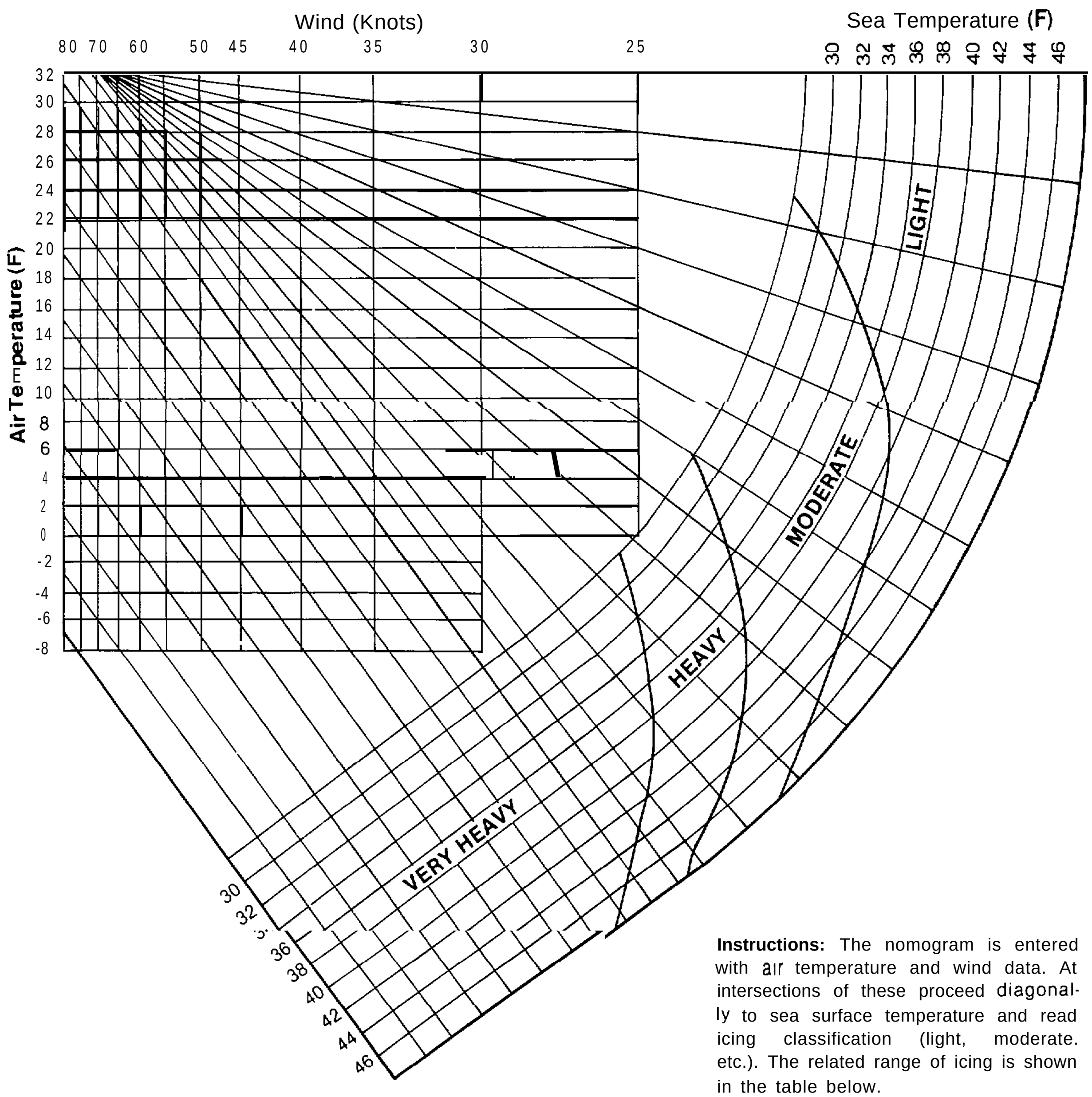
Statistical analysis (Borisenkov and Panov 1972) of more than 3,000 cases of ship icing indicates that in 86% of the cases icing was caused by ocean spray alone. Spray combined with fog, rain, or drizzle (liquid sources) accounted for only 6.4% of the cases, and spray combined with (solid source) snow only 1.1%. The cases of icing attributable only to fog, rain, or drizzle account for 2.7% (Minsk 1977). In the remainder of icing cases data were not sufficient to determine the cause.

Since the overwhelming majority of superstructure icing on ships and offshore structures is from sea spray, the remainder of this section will concentrate on this type of icing. Since a ship can present different aspects to the wind and spray, it is to be expected that the amount of spray reaching the ship will vary: Russian observations (Kultashev, et al. 1972) showed that the greatest frequency of spray and, therefore, icing occurs when a ship is heading into the wind at an angle between 15" and 45". Asymmetrical icing occurs under this condition, with the greater accumulation on the windward side. Less icing occurs with the ship headed directly into the wind, and then accumulation tends to be uniform. With ships heading downwind, spray icing is generally much less than at other angles. In developing the nomogram for forecasting spray icing potential, downwind cases (those for which the ship's heading was 120° or greater off the wind) were not used.

Meteorological/oceanographic conditions necessary for significant spray icing are water temperatures less than 8°C, winds of 25 knots (13 meters per second) or more, and air temperatures less than - 2 °C (28°F, the freezing temperature of seawater of average salinity). Generally, the stronger the wind, and the colder the air and water, the higher the rate of icing on comparable vessels or structures. In some cases, however, where the wind fetch is not sufficient to fully develop waves, icing rates are lower.

The accompanying potential superstructure icing rate nomogram (Figure 19) is a modification of that shown in Wise and Comiskey (1980), using the open ocean cases appearing in Pease and Comiskey (1985), developed

Figure 19. Superstructure Icing Rate Nomogram



Instructions: The nomogram is entered with air temperature and wind data. At intersections of these proceed diagonally to sea surface temperature and read icing classification (light, moderate. etc.). The related range of icing is shown in the table below.

Icing Classification	Rate per Hour
Light	0 to .3 inches
Moderate	.3 to .8 inches
Heavy	.8 to 2 inches
Very Heavy	2 inches

Adapted from Wise and Comiskey, 1980, using data from Pease and Comiskey, 1985.

from icing case histories in the Gulf of Alaska and southern Bering Sea. Icing intensities in inches per hour are also from Pease and Comiskey (1985). If a vessel experiencing icing takes evasive action (i.e., changes heading, reduces speed, seeks shelter, etc.), icing rates experienced would probably be less.

Reported cases of ship icing (Figure 20) in the northern Gulf of Alaska and the Bering Sea are shown from two sources; Borisenkov and Panov (1972) and WBH29 (Dyson 197583). The lack of reported icing in the northern Bering may be a result of reduced ship traffic as well as conditions not favorable for icing. Kozo (1983) estimates a potential for superstructure icing for the northern Bering Sea in September, extending into Norton Sound in October, and even into the southern Bering during the most extreme conditions. Icing potential decreases in the north Bering as the sea ice cover advances; however, the potential for moderate or heavy icing downwind of the sea ice edge persists throughout the winter.

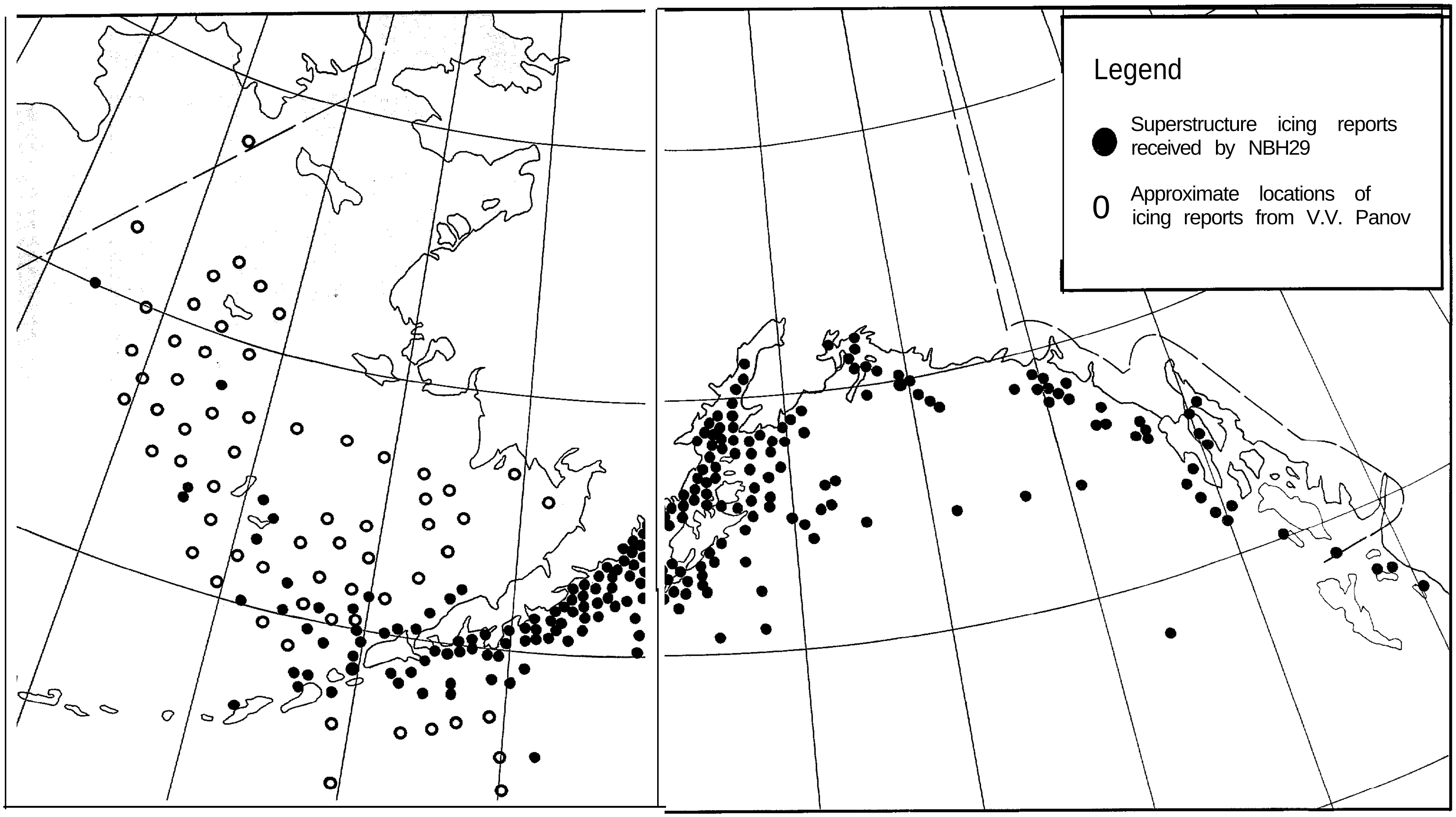


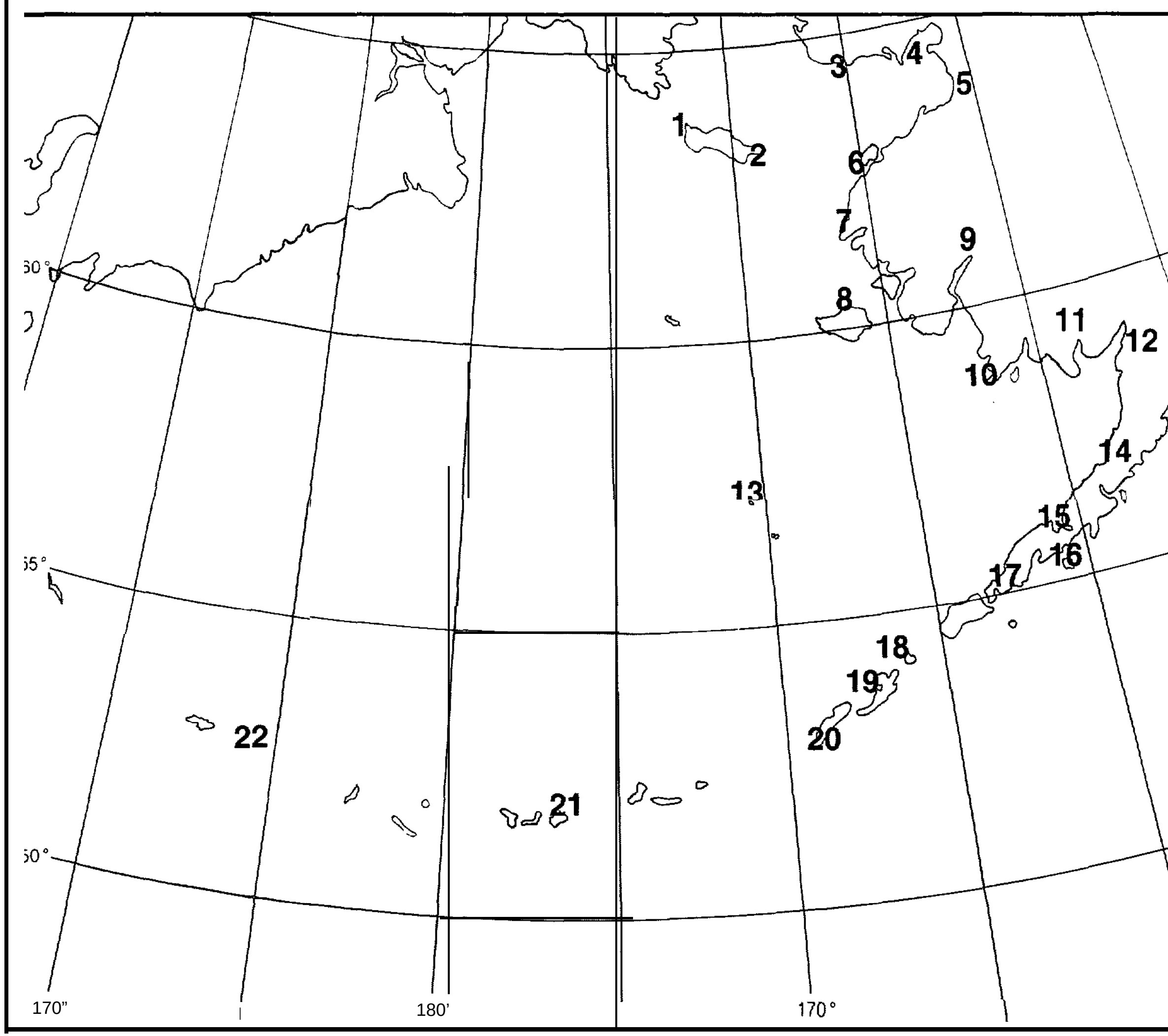
Figure 20. Related Occurrences of Superstructure Icing on Ships

Figure 21. Climatic Means and Extremes

Station	Mean Annual Maximum (°C)	Mean Annual Minimum (°C)	Highest (°C)	Lowest (°C)	Total Precipitation Average Annual (in)	Greatest Month (in)	Greatest Day (in)	Snowfall Average Annual (in)	Snowfall Greatest Month (in)	Snowfall Greatest Day (in)	Snow depth (in)	Prevailing Wind Direction / Average Annual Speed (knots)	Fastest Wind Direction / Speed (knots)
Gambell	-1.7	-6.1	18.3	-32.2	18.61	5.86	12.01	98.1	48.6	14.1	77	NNE / 15.9	NA / NA
2 Northeast Cape	-1.1	-6.2	20.0	-37.8	16.87	13.68	6.60	67.2	32.8	6.9	62	N / 11.2	SE / 94
3 Nome	0.3	-7.5	30.0	-43.3	14.77	7.82	2.99	55.8	25.9	9.0	74	N / 9.1	WSW / 48
4 Moses Point	-0.1	-7.8	30.6	-47.8	18.95	9.56	2.06	75.3	39.2	20.6	108	NNE / 10.8	SW / 50
5 Unalakleet	0.7	-7.0	30.6	-45.6	14.16	8.07	1.53	39.9	37.3	14.0	42	E / 10.5	E / 60
6 Emmonak	2.3	-5.8	26.7	-45.6	19.75	5.65	12.00	63.7	27.5	NA	63	E / 8.6	NA / NA
Cape Romanzof	0.7	-4.1	26.1	-32.2	25.24	10.50)	2.77	69.4	68.7	12.5	116	NE / 11.7	NE / 55
8 Nunivak	1.7	-4.5	24.4	-44.4	15.16	5.45	2.80	56.9	23.0	NA	48	NW / 13.6	NA / NA
9 Bethel	2.0	-6.1	30.0	-43.3	16.13	5.81	12.02	46.8	21.6	8.2	45	NNE / 11.0	SE / 60
10 Cape Newenham	2.8	-1.6	23.9	-33.3	36.42	11.42	3.13	76	20.7	13.1	51	N / 10.0	SE / 99
11 Dillingham	5.6	-3.3	33.3	-37.8	25.08	8.02	2.19	73.3	44.8	NA	57	N / 9.6	NA / NA
12 King Salmon	4.8	-3.9	31.1	-43.3	19.33	7.30	2.00	43.7	20.0	11.0	19	N / 9.2	E / 58
13 St. Paul Island	3.6	-1.0	17.8	-32.2	22.71	9.32	2.00	56.4	55.8	13.8	32	N / 15.8	N / 65
14 Port Heiden	5.4	-1.1	30.6	-32.2	15.68	5.55	13.00	49.4	37.7	18.0	24	ESE / 12.9	E / 65
5 Port Moller	5.5	-2.9	23.3	-22.8	43.51	NA	15.00	92.1	NA	10.4	36	S / 8.9	S / 44
16 Sand Point	7.8	1.8	20.6	-13.9	51.57	13.39	3.90	40.5	23.3	9.0	11	NW / 12.0	NW / 74
17 Cold Bay	5.8	0.6	25.6	-25.0	35.01	9.97	4.90	60.6	54.3	21.4	29	SSE / 14.7	SSE / 63
18 Dutch Harbor	7.4	2.2	26.7	-16.7	60.53	16.04	7.58	72.2	52.6	NA	25	SE / 9.6	E / 82
19 Driftwood Bay	6.3	-0.7	24.4	-16.1	20.55	7.63	13.08	70.2	40.1	30.0	25	NW / 8.3	WSW / 55
0 Nikolski	5.0	0.8	15.6	-11.7	23.41	NA	2.50	33.5	NA	10.4	12	NW / 13.8	NNW / 60
1 Adak	6.9	2.3	23.9	-16.1	61.19	17.34	5.49	99.2	54.8	17.0	34	WSW / 13.1	SW / 109
22 Shemya	5.1	2.1	17.2	-13.9	30.44	8.71	15.20	69.6	42.1	NA	28	N / 13.9	SW / 108

Legend

Temperature	6.9 | 2.3 / 23.9 | -16.1
Total Precipitation	61.19 / 17.34 | 5.49
Snowfall	99.2 / 54.8 | 17.0
Snowdepth	34
Surface Wind	WSW / 13.1; SW / 109



17.6 | 3.6	**Temperature(°C)** Mean Annual Maximum	Mean Annual Minimum
22.2 | -14.4	Highest	Lowest
114.30	**Total Precipitation (inches) (Rain and Snow)** Average Annual	
34.87 7.59	Greatest Month	Greatest Day
58.5	**Snowfall (inches)** Average Annual	
46.9 | 14.2	Greatest Month	Greatest Day
40	**Snow depth (inches)** Annual Maximum on Ground	
SSE / 9.3	**Surface Wind (knots)** Prevailing Direction Average Annual Speed	
SSE / 52	Fastest Direction and Speed	

Where the mean annual temperature (average of the mean annual maximum and minimum) is less than zero some type of permafrost will probably be present.

These data can be used for long-range planning and design criteria. More detailed information can be obtained from the National Climatic Data Center and the Arctic Environmental Information and Data Center in Anchorage.

NA = Information is not available

Prepared from NOAA/NESDIS and Canadian AES data.

Hypothermia

Hypothermia is the cooling of the body's core temperature to 95°F or below. It can cause shivering, numbness, and disorientation. In the extreme it can cause death. The body loses heat gradually in cold, dry conditions, but quickly becomes hypothermic in wet conditions. Rain, immersion in cold water, and perspiration can all cause rapid heat loss. However, the evaluation and treatment of hypothermia, whether wet or dry, on land or water, is essentially the same, namely to warm the victim by whatever appropriate means are available.

The following discussion was taken in part from Peters (1982).

The body loses heat in five ways:

- A large amount of heat is lost from the body in respiration. Exhaled warm air is replaced by cooler inhaled air, producing a net heat loss.The amount of the net heat loss can be reduced by covering the mouth/nose with wool or fur, thereby "prewarming" the inhaled air as it passes through the material which has been warmed by exhaled air and by heat radiating from the body.

- Evaporation of perspiration from the skin and moisture from the lungs contributes greatly to the amount of heat lost by the body. Although evaporation cannot be prevented, the amount of evaporation (and therefore cooling) can be controlled. Wearing clothing that can be opened or removed easily for ventilation will let water vapor escape and not condense to liquid water in the clothing. Keeping clothing dry preserves its insulating value and reduces heat loss.

- Sitting on snow, touching cold equipment, and being rained upon are all examples of how heat can be lost as a result of conduction. If an individual becomes wet a tremendous amount of body heat is lost rapidly. Deaths have occurred as a result of immersion in water below 40°F—body temperature could not be maintained. Although not as immediately serious, perspiration, rain, or wet snow should never be allowed to saturate articles of clothing, as this seriously reduces their insulating properties.

- Radiation causes the greatest amount of heat loss from the body from uncovered surfaces, particularly the head, neck, and hands. Coverage of these areas, therefore, is extremely important in keeping warm.

- The body continually warms (by conduction) a thin layer of air next to the skin. If the warm layer is removed by wind or air currents (advection), the body is cooled. The primary function of clothing is to retain this layer of warm air next to the skin by enclosing air in cell walls or between numerous fibers, while allowing water vapor to pass outward. Heat is lost rapidly with the lightest breeze unless the proper type of clothing is worn to prevent the warm air from being advected away.

Deaths have been attributed to a loss of body heat at temperatures of 40°F, with a 30 mph breeze. Undertheseconditions, the cooling effect on the skin is equal to that of much lower temperatures due to increased evaporation and convection. With lower temperatures and/or strong winds, cooling occurs even more rapidly. Wind protection and insulation (dead air space) can help ensure that body heat is retained at a safe level.

Treatment

Recognition and proper treatment of hypothermia must be prompt. Delays even after rescue can cost a person his life. Low body temperature is the best indication of hypothermia. Blood pressure and pulse are also good indicators. The pulse is generally slow and irregular, while blood pressure is low.

The hypothermia victim is pale in appearance, the pupils are constricted and react poorly to light, and respiration is slow and labored. He will usually be shivering violently, with frequent muscular rigidity. There may also be an appearance of intoxication.

Emergency treatment must begin as soon as possible to stop the drop in body temperature. Wet clothing should be removed. If the body temperature is 97°F or above, no treatmentother than dry clothing and moving the victim to a warm area is generally necessary. If these are not available, the wet clothing should not be removed.

Combatting "afterdrop" in the core body temperature is extremely important. When heat is applied to the arms and legs, it causes those blood vessels to relax. This allows cold blood to flow back into the body core, further cooling the vital organs. Warming of the trunk of the body should be the prime concern.

During experiments in conjunction with the U.S. Coast Guard, researchers determined that the best warming technique was from the inside out, by having the victim breathe moist, warmed oxygen (Wilson 1976).

The next best treatment is a hot bath, with the water temperature between 90 and 100°F. If a tub is not available, an inflated life raft could be used. If possible, the limbs should remain out of the water. When no tub-type facility is available, a hot (115°F) shower while wrapped in towels or blankets is preferable.

When hot water for a tub or shower is unavailable, wrap the victim in blankets in a warm room with a heating pad or well-wrapped hot water bottle on the chest, or apply body warmth by direct contact with a rescuer.

Warm liquids may be given, but care must be taken to insure the victim is conscious and does not breathe the liquid into his lungs. Alcohol should never be given because it causes "afterdrop." Observe the victim's respiration closely and monitor for vomiting.

It has been learned in studies done in Alaska that victims of wet hypothermia can survive for a prolonged time in cases of deep cooling. Apparently, in the rapid cooling which occurs with wet hypothermia, physiological changes undergone by the body are more likely to be reversible than in the slower cooling of dry hypothermia. There have been victims of immersion hypothermia who were apparently dead but revived with proper treatment.

Wind Chill (Equivalent Temperatures)

The temperature of the air is not always a reliable indicator of how cold a person will feel outdoors. Other weather elements, such as wind speed, relative humidity, and sunshine (solar radiation), also exert an influence. In addition, the type of clothing worn, togetherwith the state of health and the metabolism of an individual, influence how cold a person will feel. Cooling may be described as loss of heat from exposed flesh. Freezing occurs when there is such total heat loss that ice forms in the exposed tissues. Thecooling poweroftheatmosphere(bywind)is primarily heat transfer by advection-in human cases, by exposure of uncovered flesh to the environment. Even small amounts of air movement have considerable chilling effect because this movement disruptsorremoves the thin layer of warmed air that builds up near and about the body. This air movement leads to loss of total heat, since heat is transferred from the core of the body to rewarm the new colder air, replacing that blown away. Therefore, wind chill not only leads to frostbite locally, but may contribute to general hypothermia.

Equivalent Wind Chill Temperature

Cooling Power Of Wind Expressed As "Equivalent Chill Temperature"

Wind Speed		Temperature (°C)																		
Knots	km/hr																			
Calm		12	8	4	0	- 4	8	- 12	- 16	- 20	- 24	- 28	- 32	- 36	- 40	- 44	48	52	56	- 60
		Equivalent Chill Temperature																		
3	6	12	8	4	0	- 4	- 8	- 12	- 16	- 20	- 24	- 28	- 32	- 36	- 40	- 44	- 48	- 52	- 56	- 60
5	10	9	5	0	- 4	- 8	- 13	- 17	- 22	- 26	- 31	- 35	- 40	- 44	- 49	- 53	- 58	- 62	- 67	- 71
11	20	5	0	- 5	- 10	- 15	- 21	- 26	- 31	- 36	- 42	- 47	- 52	- 57	- 63	- 68	- 73	- 78	- 84	- 89
16	30	3	- 3	- 8	- 14	- 20	- 25	- 31	- 37	- 43	- 48	- 54	- 60	- 65	- 71	- 77	- 82	- 88	- 94	- 99
[illegible]	[illegible]	1	5	11	17	- 23	- 29	- 35	- 41	- 47	- 53	- 59	- 65	- 71	- 77	- 83	- 89	- 95	- 101	- 107
[illegible]	50	0	6	- 12	- 18	- 25	- 31	- 37	- 43	- 49	- 56	- 62	- 68	- 74	- 80	- 87	- 93	- 99	- 105	- 112
[illegible]	[illegible]	0	7	13	19	26	- 32	- 39	- 45	- 51	- 58	- 64	- 70	- 77	- 83	- 89	- 96	- 102	- 109	- 115
[illegible]	[illegible]	1	- 7	- 14	- 20	- 27	- 33	- 40	- 46	- 52	- 59	- 65	- 72	- 78	- 85	- 91	- 98	- 104	- 111	- 117
[illegible]	80	1	8	14	- 21	- 27	- 34	- 40	- 47	- 53	- 60	- 66	- 73	- 79	- 86	- 92	- 99	- 105	- 112	- 118
[illegible]		1	8	14	- 21	27	34	40	-47	- 53	-60	-66	- 73	-79	- 86	-92	-99	- 105	- 112	- 118
[illegible]	[illegible]	1	8	14	-21	27	34	40	-47	- 53	-60	-66	-73	- 79	- 86	- 92	- 99	- 105	- 112	- 118
		Little Danger						Increasing Danger (Flesh May Freeze Within 1 Minute)							Great Danger- (Flesh May Freeze Within 30 Seconds)					
Danger of Freezing Exposed Flesh For Properly Clothed Individuals																				

Figure 22. Equivalent Wind Chill Temperature Adapted from NWS/NOAA Technical Procedures Bulletin No. 165. Effective Temperature (Wind Chill Index) 1976.

During the antarctic winter of 1941 Siple and Passel developed a formula to determine wind chill from experiments made at Little America (Siple and Passel 1945). The formula relates heat loss (H) from an object or person to wind speed and to the difference in temperature between the air and the object or person (DT). It is measured in heat units (calories) per unit area over time. The skin temperature of most people is approximately 33°C (91.4°F). Heat losses for the human body can then be computed for any combination of wind and temperature. Equivalent temperature is based on calm conditions and a person walking vigorously at 3 knots (4 mph). Each combination of wind and air temperature produces a heat loss H. The equivalent temperature is that temperature that would compute the same heat loss at a wind of 3 knots. The accompanying chart, figure 18, shows equivalent wind chill temperatures in °C for various combinations of winds in knots or kmlhr and temperatures.

Concepts in the following discussion of wind chill are from an appendix to an article by William J. Mills, Jr., M.D., as published in Alaska Medicine (1973). Dr. Mills is still active in the treatment of cold injuries in Alaska.

Almost everyone knows that the increased speed of wind may cause increased danger of skin freezing. Many assume that the increase in wind speed causes the ambient air temperature to fall lower. This is not so. What does occur is air movement, so that warmed air is moved away from the individual exposed to the wind, causing first local, then general body cooling. Any resultant decrease of skin temperature is due to heat loss, insidious or sudden. Local vasoconstriction, vascular shunting, and cellular changes take place; eventually ice forms in the tissues, with true tissue freezing or frostbite.

This phenomenon can be readily proved. Place a laboratory recording thermometer with a thermistor attached (or any outdoor thermometer) out your car window on a calm day when the temperature is, say, – 20°C (– 4°F)—just a nice winter day in Anchorage, Alaska. Let it sit for a few minutes until the temperature reading has stabilized. This temperature, as read, will remain at the ambient air temperature level. Now slowly accelerate your vehicle to 80 kmlhr (50 mph); the temperature remains unchanged at -20°C (-4°F). Now attach the thermistor to your bare hand. Place your ungloved hand out the same car window in the same ambient temperature of – 20 °C (– 4 "F). After a few minutes at 0 kmlhr, the skin temperature may be read at approximately 93 °F (normal skin temperature in the nonsmoker). Your skin temperature will drop as heat is lost to the exterior, sometimes falling as low as 8.5" to 80 °F very rapidly. As the car is accelerated and the warmed air layer is moved away, the thermistor records skin heat loss. If you continue driving, your skin temperature may drop to a level near 23°F (– 5°C), the temperature at which freezing of skin may actually occur.

Wind chill may occur not only from natural wind, but also with air movement generated by automobile, snowmobile, aircraft, or helicopter rotoblade. These vehicles may predispose passengers to frostbite or general hypothermia.