The Physical Oceanography of the Bering Sea

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Introduction
The Bering Sea is a semi-enclosed, high-latitude sea that is bounded on the north and west by Russia, on the east by Alaska, and on the south by the Aleutian Islands (Fig. 1). It is divided almost equally between a deep basin (maximum depth 3,500 m) and the continental shelves (<200 m). The broad (>500 km) shelf in the east contrasts with the narrow (<100 km) shelf in the west. Seasonal extremes occur in solar radiation, meteorological forcing, and ice cover. Large interannual fluctuations exist in climate, due both to the Southern Oscillation and the Pacific-North American atmospheric pressure patterns (Niebauer 1988; Niebauer et al., chapter 2, this volume). An amplification of global warming is predicted in the Bering Sea (Bryan and Spelman 1985). Basin scale climate variability profoundly impacts both the physical and biological environment (Schumacher and Alexander, chapter 6, this volume).

Interactions among ocean, ice, and atmosphere dominate the physics of the Bering Sea. Large-scale weather patterns in both the tropical South Pacific (El Niño-Southern Oscillation events) and the North Pacific (Pacific-North America patterns) have strong connections to the Bering Sea, mainly via the atmosphere (Niebauer 1988; Niebauer and Day 1989; Niebauer et al., chapter 2, this volume). The mode of teleconnection appears to perturb the passage of storms (areas of low sea level atmospheric pressure with closed isobars) along the Aleutian Island chain. The migration of storms results in a statistical feature known as the Aleutian Low, one of the two main low pressure systems in the high latitude Northern Hemisphere. During summer with its long periods of daylight and high insolation, the Aleutian Low is typically weak and weather benign. During winter, a marked change occurs in atmospheric pressure fields. High sea level
pressure (Siberian High) dominates Asia, while the Aleutian Low intensifies and dominates weather over the North Pacific and Bering Sea. The juxtaposition of these features results in strong, frigid winds from the northeast. The frequency and intensity of storms in the southern Bering Sea decreases from winter to summer and frequency also decreases with increasing latitude (Overland 1981, Overland and Pease 1982). In the winter, an average of three to five storms per month move eastward along the Aleutian Chain forming the primary storm track, while less than two storms per month cross the northern Bering Sea. A secondary storm track curves northward along the Asian coast. A difference in the number of storms also occurs across the broad shelf, with more activity occurring over the outer shelf (Schumacher and Kinder 1983).

Due to the presence of the Aleutian Low, the wind torque over the Bering Sea is greater in winter than in summer by an order of magnitude. A climatology of the wind forcing (Bond et al. 1994) shows that eastward- and northward-propagating storm systems dominate the surface stress at short periods (<1 month), and their energy mixes the upper ocean. At longer periods (>1 month), the wind-driven (Sverdrup) transport accounts for roughly one-half of the observed transport in the Kamchatka Current.
Driven by changes in the location of the Aleutian Low, interannual variations in the Sverdrup transports occur that are ~25% of the mean.

Much of the physical oceanographic research in the Bering Sea has been coupled to fisheries research, since the combination of nutrient-rich slope waters and high summer solar radiation create one of the world’s most productive ecosystems (Walsh et al. 1989). Seasonal primary production often begins with a bloom on the shelf associated with ice-edge melt; annual production varies from >200 g C/m² over the southeastern shelf to >800 g C/m² north of St. Lawrence Island. Over the western shelf maximum annual production (>400 g C/m²) occurs over the continental slope (Arzhanova et al. 1995). These blooms often consume all available nutrients (Niebauer et al. 1995), so subsequent production depends on nutrients being supplied by mixing due to storms and/or advection. The blooms support strong higher trophic level production which in turn supports vast populations of marine mammals, birds, fish, and shellfish. During the mid-1980s, the pollock fishery was the world’s largest single species fishery, and during the mid-1990s the salmon run along the Alaska Peninsula was the world’s largest.

This paper both reviews and presents new data on the physical oceanography of the Bering Sea. It focuses on the basin, since a thorough review of the shelves has recently been completed (Schumacher and Stabeno 1998). Further, since this volume contains an extensive review of the water property characteristics (Luchin et al., chapter 3, this volume), the primary goal here is to review and update circulation over the basin. We begin with the known and estimated transports through each of the passes of the Aleutian Islands, followed by a discussion of the surface and deep currents of the Bering Sea basin and the flow on the Bering Sea shelves.

**General Circulation**

The circulation in the Bering Sea basin (Fig. 2) is often described as a cyclonic gyre, with the southward flowing Kamchatka Current forming the western boundary current and the northward flowing Bering Slope Current forming the eastern boundary current. Circulation in the Bering Sea is strongly influenced by the Alaskan Stream, which enters the Bering Sea through the many passes in the Aleutian Arc. The inflow into the Bering Sea is balanced by outflow through Kamchatka Strait, so that circulation in the Bering Sea basin may be more aptly described as a continuation of the North Pacific subarctic gyre. Circulation on the eastern Bering Sea shelf is generally northwestward. The net northward transport through the Bering Strait, while important to the Arctic Ocean, has virtually no effect on the circulation in the Bering Sea basin. It does, however, play the dominant role in determining the circulation on the northern shelf. The currents of the Bering Sea have been examined principally through inferred baroclinic flow and to a lesser extent by drifting buoys, current meter moorings, and models.
The passes play a primary role in determining both circulation and distribution of water properties. The Aleutian Arc forms a porous boundary between the Bering Sea and the North Pacific (Fig. 3). Even though only three passes (Amchitka Pass, Near Strait, and Kamchatka Strait) extend deeper than 700 m, there is significant flow through many of the 14 main passes. Transport into the Bering Sea can vary by more than a factor of two and this affects the transport in the Kamchatka Current. The strong variability of flow through the passes has time scales which vary from weeks to years. The source of most of the flow into the Bering Sea basin is the Alaskan Stream, the northern boundary of the cyclonic North Pacific subarctic gyre. It provides relatively fresh surface waters and warm subsurface water.
We first explore flow through the two passes on the eastern Bering Sea shelf, Unimak Pass (at the south), and Bering Strait (at the north). The influence of these passes is limited to the eastern Bering Sea shelf and plays virtually no role in modifying currents in the basin. We next examine flow through the major passes in the Aleutian Arc (starting on the eastern side and moving westward), which strongly influence the flow patterns and water properties of the basin.

**Shelf Passes**

Unimak Pass forms the only significant conduit between the shelves of the Gulf of Alaska and the eastern Bering Sea. This relatively shallow (<80 m) and narrow (~30 km) pass permits flow of a portion of the Alaska Coastal Current (ACC) into the Bering Sea. The ACC extends 1,000 km along the Gulf of Alaska coast from southern Alaska to Unimak Pass (Stabeno et al. 1995). A recent study of flow through Unimak Pass (Reed and Stabeno submitted) substantiates the earlier conjecture (Schumacher et al. 1982) that a strong seasonal signal in transport exists. The current appears to be uniform across the pass. Transport is predominantly baroclinic (>70%). Net baroclinic transport is northward with a maximum in the fall and winter (>0.50 × 10⁶ m³/s), a minimum in the late spring and summer (~0.10 × 10⁶ m³/s), and a mean transport of 0.23 × 10⁶ m³/s. Maximum daily-averaged barotropic flow is also northward and can exceed 100 cm/s in the winter (tidal currents can strengthen this, resulting in the flow exceeding 200 cm/s), although it is weak during the summer. The barotropic flow through this pass results from sea level changes induced by winds along the peninsula (Schumacher et al. 1982). While direct observations do not exist to establish the magnitude of the long-term variability, the portion of transport resulting from winds likely varies significantly.
Of all the passes that connect the surrounding oceans with the Bering Sea, only the physical oceanography of Bering Strait is well documented (Coachman et al. 1975, Aagaard et al. 1985b, Roach et al. 1995). An approximate 0.4 m mean sea level difference between the Bering Sea and Arctic Ocean drives a net northward transport (~0.8 × 10^6 m^3/s) through Bering Strait (Coachman 1993). Reversals of the northward flow occur during periods of southward winds. The strongest winds occur during the autumn/winter, resulting in a seasonal signal in the transport. The strongest monthly mean northward transport through Bering Strait occurs in July (~1.4 × 10^6 m^3/s), and the weakest in December (0.3 × 10^6 m^3/s). Thus the annual signal is out of phase with the inflow through Unimak Pass. There is also strong intra-annual variability, with mean annual transports ranging from 0.6 × 10^6 m^3/s to >1.0 × 10^6 m^3/s.

The northward flow through Bering Strait strongly influences the currents over much of the northern Bering Sea shelf. One consequence of the northward transport is that a supply of nutrient-rich water to the northern shelf upwells near St. Lawrence Island, thereby stimulating primary production (Nihoul et al. 1993). The northward flow also provides the only connection and exchange of water between the Pacific and Atlantic Oceans in the Northern Hemisphere. During ice formation, cold saline water (>34 psu and <=-1.5°C) is produced over the northern shelf and flows northward through Bering Strait. Globally, this water plays a role both in maintaining the Arctic Ocean halocline and in ventilation of the deep waters (Aagaard et al. 1985a).

**Deep Aleutian Passes**

The easternmost major conduit between the North Pacific Ocean and the Bering Sea is Amukta Pass (Figs. 1 and 3). Recent measurements have shown that the net baroclinic transport through this pass is highly variable (Reed and Stabeno 1997), ranging from 1.4 × 10^6 m^3/s northward to a weak (<0.10 × 10^6 m^3/s) southward flow, with a mean of 0.6 × 10^6 m^3/s. Maximum northward geostrophic speeds vary from 25 to 73 cm/s, while the southward speeds are weaker (7-32 cm/s).

This transport, along with the flow through Amchitka Pass, is the source of the Aleutian North Slope Current (Reed and Stabeno, chapter 8, this volume) and ultimately the Bering Slope Current. The width of the pass is sufficient (greater than the internal Rossby radius) that there is generally northward flow on the east side and southward flow on the west side. The relative strength of these two branches determines the direction and magnitude of the net transport. The source of the flow through this pass is usually the Alaskan Stream; however, at times (~10%) the eastward flow on the north side of the Aleutian Islands turns southward through the western side of the pass and retroreflects through the eastern side. A more common scenario, however, is that the Alaskan Stream enters on the eastern side of the pass with some retroreflecting, supplying a portion of the southward branch.
Two conductivity temperature depth (CTD) sections across Amukta Pass, done a year apart during August (Fig. 4), show water properties during a period of strong (Fig. 4a,b) and of weak (Fig. 4c,d) inflow. The warm water in the upper 100 m during both years resulted from summer warming. The warm water at depth during 1994 resulted from Alaskan Stream inflow, while during 1995 there was little inflow of Alaskan Stream water. While variability on time scales of days exists, the data presented in Reed (1995) and Reed and Stabeno (1997) suggest that inflow of Alaskan Stream water through Amukta Pass varies mainly on the time scale of months to years.

In the Aleutian Basin, a subsurface temperature maximum occurs between 150 and 400 m. The \sigma_t\text{-}density of the maxima is between 26.6 and 26.9 and the temperature ranges between 3.5 and 4.6°C (Fig. 5a,b). The variability in this temperature results from the absence or presence of flow of Alaskan Stream water through Amukta Pass. When flow is present, a signature of warm (>4.0°C) subsurface water is evident in the eastern basin and slope region (Reed 1995). Inferring lack of inflow from the absence of the warm subsurface water indicates that for periods of months inflow of Alaskan Stream water through Amukta Pass can be absent.

Few measurements have been made of transport through Amchitka Pass, the third deepest pass in the Aleutian Arc, and only three hydrocast sections have been reported in the literature through the narrow sill region at 51°28’N (Reed and Stabeno, chapter 8, this volume). Measured baroclinic transports range from 2.8 \times 10^6 m^3/s southward to 2.8 \times 10^6 m^3/s northward, with a mean of 0.3 \times 10^6 m^3/s northward. The flow pattern across the pass is similar to that observed through Amukta Pass, with inflow on the eastern side and outflow on the western side. The source of outflow is twofold. Some originates at Near Strait and Buldir Pass and flows eastward across Bowers Ridge, while the remainder consists of a retroflection of the inflow through the eastern part of Amchitka Pass.

A vertical section of temperature and salinity reveals the inflow of Alaskan Stream water on the eastern side of the pass where the depth of the 4°C contour dipped to ~400 m (Fig. 6a). As suggested by isopleths, the net baroclinic transport was weak (0.3 \times 10^6 m^3/s northward). Two satellite-tracked drifters were deployed shortly after the CTD section in the vicinity of Amchitka Pass. The trajectories (Fig. 7) reveal the presence of two eddies that inhibited much of the flow through the pass. The northern drifter remained in the cyclonic eddy for 22 days, making 2.5 circuits around the northern eddy. The other drifter made four circuits of the southern eddy before malfunctioning. Typical speeds of rotation were the same for each eddy, 30 cm/s. Flow of significant amounts of Alaskan Stream water was thus blocked from the pass. During 1991, there was also evidence from drifter trajectories (not shown) of an eddy in the central portion of the pass, but these drifters did not remain in the eddy for more than one rotation. Evidence of an eddy in the Alaskan Stream south of Amchitka Pass which affected the transport through Amchitka Pass, was revealed in altimeter data (Okkonen 1996).
Figure 4. (a) Vertical sections of temperature (°C) and (b) salinity (psu) in Amukta Pass in August 23, 1994, during a period of strong inflow (net $1.0 \times 10^6$ m$^3$/s) of Alaskan Stream water into the Bering Sea. (c) Vertical sections of temperature (°C) and (d) salinity (psu) during a period of weak (net flow $0.1 \times 10^6$ m$^3$/s) outflow from the Bering Sea.
Figure 5. (a) Distribution of temperature (°C) at the subsurface maximum during a period when the temperature was at a minimum. (b) Distribution of the temperature of the subsurface maximum when substantial areas were >4.0°C. (This figure is adapted from Figs. 3 and 5 in Reed 1995).
Results from the one current meter mooring, which was positioned north of the pass, have been reported in the literature (Reed 1990). The net flow (14 cm/s) is to the northeast (Fig. 8). Interpretation of measurements from a single mooring within a bi-directional flow is ambiguous; however, results from this current meter integrated with hydrographic surveys suggest a northward flow of $2-3 \times 10^6$ m$^3$/s through the eastern part of the pass. Through the western portion of the pass a southward flow occurs, resulting in a net transport northward of probably $1-2 \times 10^6$ m$^3$/s.

Virtually no direct measurements of current have been made in Buldir Pass. Trajectories of satellite-tracked drifters indicate that northward flow can occur through the pass. Stabeno and Reed (1992) estimated an inflow of less than $1.0 \times 10^6$ m$^3$/s.

Most of the transport into the Bering Sea occurs through Near Strait ($6-12 \times 10^6$ m$^3$/s). Historically, transport of Alaskan Stream water through Near Strait was estimated at $\sim10 \times 10^6$ m$^3$/s and believed to be relatively constant (Favorite 1974). More recent research indicates that the flow through this pass, as through the other passes, is extremely variable. Occasionally, instabilities (eddies) occur in the Alaskan Stream that inhibit

Figure 6. Vertical sections of (a) temperature (°C) and (b) salinity (psu) along 51°30'N in Amchitka Pass on September 15, 1992.
flow into the Bering Sea through Near Strait. Such an eddy in 1991 resulted in a net northward baroclinic transport of less than $3 \times 10^6$ m$^3$/s (Stabenon and Reed 1992). Near Strait is the only deep pass that has been monitored with current moorings (Reed and Stabenon 1993). A series of moorings deployed in the eastern portion of the pass indicate that the periods of reduced inflow vary on scales of months. Yearly average velocity at ~100 m was 6 cm/s into the Bering Sea in the eastern portion of the pass, and decreased to 1-2 cm/s in the central part of the pass. It was moderately surprising that the strongest, steadiest flow occurred near the bottom of the narrow passage on the eastern side of the pass, a result of tidal rectification.

Not only is Kamchatka Strait the location of the majority of outflow, but it is the only conduit into the Bering Sea that is deep enough (>2,000 m) to permit inflow of Deep Pacific Water (DPW). In the upper 1,500 m, flow is...
Figure 8. Daily currents measured at seven locations indicated in Fig. 2. A low pass filter (35-hour Lanczos squared) was applied to each series and the series were then rotated to their net direction (indicated in parentheses). The depth of each instrument was between 250 and 350 m. Site locations are shown in Fig. 1.
dominated by the southward-flowing Kamchatka Current on the western side of the strait and a weak inflow occurs on the eastern side (Stabeno and Reed 1994, Cokelet et al. 1996). The inflow of DPW occurs below 2,000 m predominately on the eastern side of the strait (Reed et al. 1993), and thus the highest salinities in the Bering Sea basin are observed here (Fig. 9b). Salinities decrease in the basin as the high salinity DPW mixes with surrounding water. The Bering Sea deep water contains the highest concentrations of silica in the world's ocean (Mantyla and Reid 1983). Between 2,000 and 3,000 m, high concentrations of silica (>225 m M/L; Fig. 9c) occur in the strait (Reed et al. 1993). Concentrations similar to these are observed in the Bering Sea basin farther to the northeast. This suggests a southward flow of deep Bering Sea water beneath the Kamchatka Current and above the inflow of DPW. Measurements of transport of the Kamchatka Current in the pass range from $5 \times 10^6$ m$^3$/s (Verkhunov and Tkachenko 1992) to $15 \times 10^6$ m$^3$/s (Ohtani 1970).

**Mean Upper Ocean Circulation**

General circulation over the deep basin is characterized by a cyclonic gyre, with three well-defined, distinct currents: the Kamchatka Current along
the western boundary; the Bering Slope Current along the eastern boundary; and the Aleutian North Slope Current connecting the inflow through Amukta Pass and Amchitka Pass with the Bering Slope Current (Fig. 2). Transport within the gyre can vary by more than 50%. Modeling studies have simulated such large changes in transport and identified the causal mechanisms to be fluctuations of Alaskan Stream inflow (Overland et al. 1994) and/or changes in the wind-driven transport within the basin (Bond et al. 1994).

The main features of the upper-ocean circulation are shown in Fig. 3. This figure has been modified from those in Stabeno and Reed (1992, 1994) and Reed (1995). Much of the information is from the water property and geostrophic flow results from Favorite (1974), Hughes et al. (1974), and Sayles et al. (1979). Information from satellite-tracked drifting buoys is taken from Stabeno and Reed (1994) and from the limited number of current meter moorings which have been deployed in the basin. A more detailed flow field (Fig. 10) derived from satellite-tracked drifters (following Stabeno and Reed 1994) from 1984 through 1996, reveal each of the main current systems along with flow through the passes and weak westward flow across the basin.

The Alaskan Stream

While the Alaskan Stream (AS) is not in the Bering Sea, it is the main source water for much of the flow that occurs in the Bering Sea and so merits a brief discussion. The AS, which is the northern boundary current of the Pacific subarctic gyre, extends from the head of the Gulf of Alaska to the western Aleutians. It is characterized as a narrow, high speed current, with small eddy kinetic energy to mean kinetic energy (KE'/KE) ratios (KE'/KE<1.0) from 165°W to 173°E (Stabeno and Reed 1990). It turns northwestward at Amchitka Pass (180°) following the bathymetry, slowing and broadening. Maximum daily averaged buoy speeds east of 180° were 70-95 cm/s, while to the west of 180° maximum speeds were 40-65 cm/s. Although the position of the Alaskan Stream relative to the Aleutian Islands is relatively constant, there is some variability, which, coupled with eddies, results in changes in transport through the passes.

Aleutian North Slope Current

Although there is some indication of an eastward flow along the north side of the Aleutian Islands between Kamchatka Strait and Near Strait, and also between Near Strait and Amchitka Pass (Stabeno and Reed 1994), the only well-documented persistent eastward flow occurs east of Amchitka Pass (Reed and Stabeno, chapter 8, this volume). This current is strongly influenced by flow through the passes. East of Amchitka Pass, it is not completely continuous, but rather originates from inflow on the eastern side of one deep pass and continues eastward until it (or some portion) exits through the western side of the next deep pass.
The Aleutian North Slope Current (ANSC) is best documented between 174°W and 167°W by hydrographic surveys (Reed and Stabeno 1997), satellite-tracked drifting buoys (Stabeno and Reed 1994), and more recently in current meter records (Reed and Stabeno 1997). It is here that the flow merits being called a current. The ANSC is a narrow (~20 km), stable current that appears to have an annual signal, as indicated by both the trajectories of satellite-tracked drifting buoys and time series from current meter moorings. Currents (measured at mooring site 5, Fig. 1) are northeastward following bathymetry. Few reversals are evident in these 13-month-long current records (Fig. 8). The strongest daily average currents (>40 cm/s at 100 m) occur during winter, and the mean velocity at 100 m exceeds 20 cm/s. The flow at 50 m above the bottom has no seasonal signal and is statistically significant at 4.5 cm/s northeastward. The KE'/KE ratio is very small (<0.5) in the upper ~300 m and near the bottom. This vertical structure, with decreasing speeds, is also evident in baroclinic estimates of flow from hydrographic surveys. Measurements of baroclinic transport range from $3 \times 10^6$ m$^3$/s (Reed and Stabeno, chapter 8, this volume) to $5.5 \times 10^6$ m$^3$/s (unpublished data from February 1997).

Figure 10. Velocity at 40 m derived from satellite-tracked drifters. The method used was the same as described in Stabeno and Reed (1994).
Bering Slope Current

Eastward of 167°N, the ANSC turns northwestward, forming the Bering Slope Current (BSC), the eastern boundary current of the Bering Sea gyre (Schumacher and Reed 1992, Kinder et al. 1975). The characteristics of this flow differ markedly from the ANSC. This can easily be seen in comparing time series and statistics of two moorings, one in the ANSC and the other in the BSC (Fig. 1). Unlike the ANSC, the presence of an annual cycle is not clearly evident in the time series at any of the sites in the BSC (sites 1, 2, 3, and 4), although model simulations show an intensification in fall-winter (Overland et al. 1994). The reversals evident in the BSC time series are likely a result of eddies. Only weak flow occurs at the sites associated with the canyons (sites 2 and 4). At the northernmost site, the strong currents are associated with eddies and the background flow is weak. The eddy:mean kinetic energy ratios at site 3 best represent the BSC. Seaward of the 1000-m isobath the flow is more variable and the BSC is often less well defined. The mean velocity (site 3) at 100 m (11 cm/s) is approximately half that observed in the ANSC. The BSC is also shallower than the ANSC, with mean flow at 500 m ~1 cm/s. Total transport $(3.6 \times 10^6 \text{ m}^3/\text{s})$ is of course similar to that observed in the ANSC. These are consistent with the BSC being wider than the ANSC.

The BSC is strongest and most persistent south of 56.5°N. Most of the flow separates from the slope at the latitude of Pribilof Canyon or Zymchug Canyon, forming the source of the weak westward flow across the basin (Stabeno and Reed 1994). To the north of 58°N the flow is still generally northwestward, but is weaker, more intermittent, and confined to shallower water depth.

Studies of the BSC suggest that two significantly different structures or modes exist. Many CTD surveys reveal an ill-defined, highly variable flow interspersed with eddies, meanders, and instabilities (Kinder et al. 1975, Reed 1991). Other surveys, however, have revealed a more regular northwestward flowing current (Reed and Stabeno 1989). The trajectories from the more than 50 satellite-tracked drifters deployed in the southeast Bering Sea support this dichotomy in the structure of the BSC. They reveal the strong variability in the flow patterns that occur along the shelf break. In some trajectories, the BSC appears as a well-behaved current flowing northwestward along the shelf break (Stabeno and Reed 1994). At other times, chaos dominates the dynamics of the system (Reed and Stabeno 1990). At still other times, eddies, ranging in size from ~40 km to 150 km, are imbedded in the flow (Schumacher and Stabeno 1994). The structure of the BSC directly influences both the advection and dynamics along the slope, and also the occurrence of across-shelf fluxes. This dichotomy of structure, of either a series of eddies or a uniform flow, is supported by numerical model simulations of the subarctic North Pacific portion of the NRL Pacific Ocean model.
The Kamchatka Current

The third distinct current system, the Kamchatka Current, forms the western boundary current of the Bering Sea gyre. It originates near 175°E (Shirshov Ridge), as the weak westward flow (Stabeno et al. 1994). The source of this water is a combination westward flowing water (from the BSC) and northward flowing water entering the Bering Sea through Near Strait (Stabeno and Reed 1994, Khen 1989). The bathymetric feature of Shirshov Ridge causes a southward deflection and reduction in speed of the current (Stabeno and Reed 1994). As the Kamchatka Current continues southward along the Russian coast it both strengthens and deepens. Near Shirshov Ridge flow (>5 cm/s) is evident only in the upper 1,000 m, while in Kamchatka Strait it has deepened to >1,500 m. The maximum daily averaged speeds observed in the Bering Sea occur in the Kamchatka Current (40-77 cm/s). Daily averaged velocities as large as 100 cm/s occur in Kamchatka Strait. Meanders and eddies are common in the Kamchatka Current (Solomon and Ahlnas 1978, Stabeno et al. 1994, Cokelet et al. 1996). These features cause the long-term mean velocities to be much smaller than observed in the more stable Alaskan Stream.

Favorite (1974) estimated the transport in the Kamchatka Current as $\approx 10 \times 10^6$ m$^3$/s, but recent results indicate that it can be quite variable ranging from 7 to $15 \times 10^6$ m$^3$/s (Stabeno and Reed 1992, Reed and Stabeno 1993, Overland et al. 1994, Cokelet et al. 1996). Variations in wind-stress curl are substantial with a winter maximum and might affect the transport (Overland et al. 1994, Bond et al. 1994). At times flow in the Kamchatka Current recirculates in the Bering Sea, with only a portion flowing through Kamchatka Strait and the remainder flowing eastward along the north side of the Aleutian Islands (Reed et al. 1993).

Interior Flow

Resulting from the strong inflow through Near Strait, a weak northward flow is evident in buoy trajectories and model results over the basin (Figs. 2 and 10). The BSC is the source of the westward flow across the basin (Royer and Emery 1984), which originated at the two areas along the slope where the 1,000-m isobath is oriented east-west (Stabeno and Reed 1994). At the southern source (~56.5°N) the flow bifurcates with a significant portion forming a small sub-gyre in the southeast corner of the basin. At the northern source (~58°N) the flow continues eastward across the center of the basin.

There have been no long-term current moorings in the central basin, but a mooring (site 6) has been located in the southeast Bering Sea (2,200 m water depth) during spring and summer of 1992-1994 (Fig. 8). These records show generally weak flow interrupted by eddies (Cokelet and Stabeno 1997). The background flow contrasts sharply from the other two sites (one in the ANSC and the other in the BSC) discussed earlier. At 80 m the
Flow is weakly northward (~1 cm/s) with high eddy-mean kinetic energy ratios. Below 150 m the currents are negligible. The eddies that inter-sperse the weak background flow are far more energetic.

**Eddies**

As elsewhere in the world’s oceans, eddies are ubiquitous in the Bering Sea (Solomon and Ahlnas 1978, Kinder et al. 1980, Paluszkiewcz and Nebauer 1984, Schumacher and Stabeno 1994). They occur on horizontal scales ranging from ~10 km to 200 km. Proposed mechanisms for the creation of these eddies include instabilities, wind forcing, strong flows through the eastern passes, and topographic interactions (Schumacher and Stabeno 1994).

The influence of eddies in straits have already been discussed, as has the presence of eddies in the BSC. One of the best time series, showing eddies in the basin has been collected during the spring and summer over a 3-year period at site 6. Eddies are common in these current records. These observations characterize the eddies as often anti-cyclonic, 20-100 km in diameter, extending to a depth of 400-1000 m, and with rotational speeds >20 cm/s (Schumacher and Stabeno 1994, Cokelet and Stabeno 1997).

On the western side of the basin, eddies result from instabilities in the Kamchatka Current. The eddies are often anti-cyclonic, 20-100 km in diameter, and have rotational speeds of >40 cm/s. They have been observed in satellite-tracked buoy trajectories, hydrographic surveys, and satellite images (Solomon and Ahlnas 1978). In three embayments on the Kamchatka Peninsula anti-cyclonic eddies were evident in the tracks of many of the buoys. They resulted from the interaction of the Kamchatka Current with topographic features and are likely semi-permanent, since they appeared in trajectories from more than 1 year (Schumacher and Stabeno 1998).

West of Bowers Ridge a large (~200 km), energetic (velocities ~30-40 cm/s) eddy was observed in 1991 (Reed et al. 1993, Cokelet et al. 1996). Buoys were deflected around the feature, with none entrained into the center. Although this feature persisted for several months, it was only observed during 1991. Thus its persistence is unknown.

Eddies are common at the eastern shelf break and shoreward to depths of ~150 m. A recent interpretation of hydrographic observations (Reed, in press) suggests anticyclonic eddies exist in the region between 100 and 122 m <20% of the time. Shoaler than 100 m, eddies are uncommon.

**Deep Circulation**

Little comprehensive information is available on deep circulation in the Bering Sea. However, a study that examines all of the available information on deepwater properties, and infers flow, is nearing completion (T.E. Whitledge, University of Texas at Austin, pers. comm., November 1995
[now at University of Alaska Fairbanks, School of Fisheries and Ocean Sciences, Fairbanks, AK 99775-7220]). Sayles et al. (1979) presented limited data to 2,500 m and inferred flow at 1,000 and 1,400 m, referred to 2,500 m. The only significant flow at these levels was the Kamchatka Current.

Reed et al. (1993) presented vertical sections of temperature salinity sigma-t, and silica across the Kamchatka Strait. Inflow of deep water occurred below 2,500 m on the eastern side of the strait. Silica data suggested a return flow of upper deep water (near 2,500 m) on the western side of the strait. In July 1993, a World Ocean Circulation Experiment section was occupied in the Bering Sea from the continental slope southwestward through Amchitka Pass (Roden 1995). The deep Bering Sea is warmer, less salty, less dense, less oxygenated, and with greater concentrations of silica than the open waters south of the Aleutians. Silica increases somewhat near the continental slope (Roden 1995, Tsugonai et al. 1979). Clearly more data are needed, but the path of deep water must be northward and eastward from Kamchatka Strait with some return flow above 3,000 m on the western side of the strait. The map of deep flow is consistent with hydrographic data and water properties (Fig. 11).

We are not aware of any long-term measurements of flow below 1,000 m. A year-long record at ~1,000 m depth along the eastern shelf break indicates that all significant flow occurred above 500 m (Schumacher and Reed 1992). Reed (1995) reviewed previous efforts to establish reference levels for geostrophic flow computations in the western Bering Sea near the Kamchatka Current. Deep levels (to 3,000 m) appeared most suitable, but levels of 1,500 m or above were suggested for the eastern Bering Sea. Cokelet et al. (1996), compared measurements between an acoustic Doppler current profiler and geostrophic shear; these results supported the regional pattern suggested above.

**Shelf-Slope Exchange**

A connection exists between the basin and shelf that is fed by shelf-slope exchange. The northward transport through Bering Strait \((0.8 \times 10^6 \text{ m}^3/\text{s})\) and through Unimak Pass requires a net onshelf transport of \(~0.5 \times 10^6 \text{ m}^3/\text{s}\). One school of thought is that a “river” of water flows onto the northern shelf. This is manifest in some schematics (e.g., Shuert and Walsh 1993) that show the BSC flowing northward along the slope and bifurcating south of Cape Navarin. One branch (sometimes called the Anadyr Current) then flows across isobaths along the Gulf of Anadyr where a canyon exists which influences model simulations (e.g., Overland and Roach 1987) through Anadyr Strait and thence to Bering Strait. Direct observations (Stabeno and Reed 1994), however, show that the BSC tends to leave the slope near 59°N and then flow eastward. While temperature, salinity, and nutrient concentrations in the waters that eventually go through Bering Strait are all of slope origin, their source is likely not a “river” which flows from
Shelf-slope exchange can occur virtually anywhere along >1,200 km of shelfbreak north of Unimak Pass. Two regions exist where preferential transport onto the shelf has been observed. The first is Bering Canyon which lies along the Aleutian Islands near Unimak Pass. The enhanced concentration of nutrients observed near Unimak Pass likely originate from ANSC waters interacting with canyon topography and coming onto the shelf (Schumacher and Stabeno 1998). The second region occurs west of Pribilof Islands, where the narrowing of the shelf break accelerates the flow along the 100-m isobath, which then turns northward. The accelerated flow over the outer shelf between St. George Island and Pribilof Canyon results in water being entrained from the adjacent slope (Stabeno et al., chapter 9, this volume). Flow near the Pribilof Islands is evident in satellite-tracked drifter trajectories (drogue depth 40 m).

Episodic events of onshelf flow have been observed resulting from both eddy-topography interactions and as a result of instabilities in the BSC (van Meurs and Stabeno, in press). Along the slope of the central shelf (~56.7°N), current records reveal that eddies are common and estimates of salt and temperature fluxes indicate that significant onshore fluxes exist (Schumacher and Reed 1992). In 1992 an excellent set of data was
collected when an eddy translated onto the shelf. (Schumacher and Staben 1994). Three satellite-tracked drifters had been deployed in the center of an eddy. The drifters remained in the eddy for weeks, until the eddy moved eastward onto the shelf. The drifters were deposited onto the shelf as the eddy spun down. Both of these episodic events of onshelf flow were related to the stability of the BSC.

The net effect of onshelf fluxes from the slope, either regionally or by intermittent processes, is the presence of slope waters over the outer shelf of the eastern Bering Sea. Using the generally accepted value of mean speed along both the southeastern and central outer shelf (0.05 m/s in the upper 100 m of the ~100 km wide domain), we compute a transport of $0.5 \times 10^6$ m$^3$/s. It is these waters that are westward intensified and become the strong steady current that flows through Anadyr Strait, providing nutrients for the high productivity observed there.

**Shelf Flow**

**Low Frequency Currents**

The schematic of flow on the eastern shelf (Fig. 2) is a synthesis of moorings, satellite-tracked drifting buoys, and hydrographic sections (Schumacher and Staben 1998). Flow through Unimak Pass results in a weak, although persistent, flow along the 50-m isobath. Measurements are not sufficient to determine if there is a seasonal signal in this flow, reflecting the variability of the flow through Unimak Pass. The flow along 50-m isobath continues through Shpanberg Pass and accounts for about a third of the flow through Bering Strait (0.35 m$^3$/s). Some flow through Unimak Pass follows along the 100-m isobath (Reed and Staben, submitted). This flow continues northwestward along the ~100 m isobath, although its position appears to vary on annual or longer scales (Reed, in press). It is augmented by the onshelf flow associated with Bering Canyon and episodic events. Additional flux of onshelf flow occurs south of the Pribilof Islands. North of the Pribilof Islands a portion of northward flow along the 100 m isobath separates and flows eastward across the shelf, augmenting the flow through Unimak Pass. The flow that continues along the 100-m isobath appears as a broad, slow northwestward flow.

This northward flow intensifies along the east coast of Siberia forming the only strong low-frequency currents on the shelf (the Anadyr Current). The flow turns eastward across the shelf along the southern coast of Siberia and eventually exits the Bering Sea through Bering Strait. The flow through Bering Strait dominates the current dynamics of the northern shelf.

The freezing and melting of sea ice can result in a distribution of mass that generates baroclinic flow. As with all high latitude seas, sea ice is one of the primary characteristics of the Bering Sea. For approximately half the year the Bering Sea is free of ice, but during November ice begins to form or be advected into the Bering Sea through Bering Strait. Although
the ice is generally limited to the Bering Sea shelves, the extent of the seasonal advance is the largest observed in any arctic or subarctic region. The formation of the ice is best described by “conveyor belt” analogy (Pease 1980, Overland and Pease 1982). Ice forms along the leeward side of the coasts and islands in polynyas (open regions of water). The ice is then advected southward (or southwestward) where it is melted by the warmer water. At the polynyas the freezing sea water produces regions of high salinity (>34 psu) and at the melting edge the ice freshens the water (~30 psu). The amount of production and advection of ice depends upon which storm track dominates in a given winter, with greatest ice production occurring in years when the Aleutian Low is well developed and storms migrate along the primary storm track. By late March or April, the ice generally has begun to retreat, leaving behind a freshened water column and shelf with relatively cold water.

Cold saline water from the Gulf of Anadyr and Anadyr Strait polynyas provides a substantial fraction of the total salt advected into the Arctic Ocean (Cavalieri and Martin 1994). The ice melt produces a lens of fresh water that can assist in the formation of a strong two-layer system over the middle Bering Sea shelf. Associated with ice melt is a bloom of phytoplankton (during April) that accounts for 10-65% of the total annual primary production over the eastern shelf (Niebauer et al. 1990, Stabeno et al. 1998) with estimates of 45-70% for the western shelf (Mordasova 1995). During years when the ice is not present over the southeastern Bering Sea shelf, the spring bloom is delayed (May).

Far less is known about the flow along the western shelf. Satellite-tracked drifter trajectories reveal a narrow, westward flowing current (15-25 cm/s) along the south coast of Siberia. These observations support the inferred flow from water property distributions.

**Tidal Currents**

Tides enter the Bering Sea from the Pacific Ocean, and to a much lesser degree from the Arctic Ocean through Bering Strait (Pearson et al. 1981, Mofjeld 1986). Most of the kinetic energy south of 60°N is provided by tidal currents. North of this latitude the strength of the tides decreases and the mean flow strengthens (Schumacher and Stabeno 1998, Coachman 1986).

Tidal currents play a vital role in the physical oceanography over the shelves. They provide sufficient energy to mix the bottom ~40 m of water over the southeastern shelf, thus setting up a two-layer density structure in water depths of 50-100 m. Shoaler than this the water column is weakly stratified or well mixed (Schumacher and Stabeno 1998). Tidal currents may contribute to the cross-shelf flux of salt, nutrients and heat necessary to maintain the high levels of primary production over the eastern shelf. Tidal currents also provide a mechanism for the generation of subtidal flow resulting from rectification of tidal currents with topography (Kowalik, chapter 4, this volume).
Future Directions

While our knowledge of the physical oceanography of the Bering Sea has expanded greatly over the past few decades, many phenomena exist which are not understood primarily because observations are limited or do not exist. That a vast percentage of the Bering Sea lies within the domain of two different nations has not facilitated research programs that could provide the needed observations. Further, while the eastern continental shelf continues to have ongoing research programs and interest in the role of physical processes over the western shelf is growing, the deep basin remains largely unexamined.

The Bering Sea is a vital part of the general circulation of the North Pacific Ocean; fluxes of heat, salt, nutrients, and other dissolved constituents and planktonic material are exchanged through the passes. Primary questions, however, remain about the variability, magnitude and mechanisms influencing transport through the Aleutian Passes. In contrast, the flux northward through Bering Strait, which is important to conditions on the Arctic shelves and ocean, is relatively well described and understood.

While it is recognized that flow through the passes is a primary source of circulation within the basin, many questions remain regarding the current systems of the Bering Sea. What is the nature of the annual signal of the ANSC, and if there is a significant annual signal, does it occur in both speed and transport? The BSC apparently can be characterized by two modes, yet the phenomena that generate these are not known. While some studies have elucidated the nature of the Kamchatka Current, little is known of the temporal variability in transport and eddy kinetic energy. What is the magnitude and variability of inflow of DPW into the Bering Sea Basin? The flow patterns of the deep basin have been inferred from a limited number of hydrocasts, so we know neither the temporal nor spatial variability. The behavior of the source waters for deep circulation, the inflow of the DPW through Kamchatka Strait, is not known.

The processes that result in the exchange of slope and shelf waters have not yet been determined. Hence, we do not know the mechanisms that provide nutrients to the euphotic zone and are responsible for the region of prolonged biological production known as the “Green Belt” (Springer et al. 1996). While the processes are unknown, the results of their interactions are evident. The continental shelf of the Bering Sea exhibits extremely high productivity, and this richness applies throughout the food chain. Not only are there vast quantities of commercially valuable species, but the eastern shelf is the summer feeding ground for numerous marine bird and marine mammal populations of the North Pacific Ocean. The eastern Bering Sea provides an ideal location to examine exchange mechanisms between slope water of an eastern boundary current and a continental shelf. Because the coast and its inherent topographic and coastal convergence processes are far removed from the slope, the processes involved in shelf/slope exchange should provide a clear signal. The conti-
mental shelf of the western Bering Sea is bounded by a typical western boundary current, so that contrasts of processes between the eastern and western shelves should be fruitful.

Future studies that focus on how the extant physical phenomena affect marine populations offer the best opportunity to enhance our understanding of ecosystem dynamics. This, in turn, could lead to management strategies aimed at sustainable production to ensure a rich ecosystem for our future generations. The observational database for the Bering Sea is not adequate, in both spatial and temporal coverage, to answer most of the questions noted above. In addition to further observations, modeling efforts need to be improved. A primitive equation basin-shelf model coupled to both outflow through Bering Strait and exchange in the North Pacific Ocean is a likely starting place. Once the model provides accurate simulations of the physical features, then biophysical processes and rates can be incorporated. Some of the questions that must be addressed to understand the ecosystem are best investigated by modeling efforts.

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An Update on the Climatology and Sea Ice of the Bering Sea

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Abstract
The Bering Sea is a region of extreme seasonal as well as substantial interannual variability in its air-ice-ocean environment. This chapter updates work on the weather/climate and ice environment of the Bering Sea and its surrounding environs. Topics with extra emphasis include polynyas, analyses of ~50-year long time series of the air-ice-ocean parameters, analysis of a climate “regime shift,” and the connection of the Bering Sea environment with interannual variability in the Aleutian Low (i.e., indices of interannual atmospheric variation such as the North Pacific [NP] and western Pacific [WP] oscillations in the North Pacific as well as the Southern Oscillation Index [SOI] in the tropical Pacific).

Introduction
This chapter is an update and expansion upon, but does not replace, the reviews of synoptic climatology and the effects on sea ice of the Bering Sea by Overland (1981) and Niebauer (1981a,b). This chapter is also meant to compliment the Bering Sea atlases of Brower et al. (1977) and Brower et al. (1988), the satellite microwave sea ice atlases of Parkinson et al. (1987) and Gloersen et al. (1992), as well as a recent review of the physical environment of the Bering sea by Niebauer and Schell (1993). One of the topics
that we emphasize is the update of time series, of ~50 years in length, of the air-ice-ocean environment of the Bering Sea as well as indices of the Aleutian Low (both North Pacific and western Pacific) and El Niño–Southern Oscillation (ENSO). In this time series analysis, we also update the quantitative relationships among the time series outlining the role of the Aleutian Low and its impact on the Bering Sea environment before and after a climatic “regime shift” (e.g., Ebbesmeyer et al. 1991) that occurred in the late 1970s. Our focus is primarily the southeastern Bering Sea shelf, but we do include some Russian material on the western Bering Sea.

**Updated General Background**

The Bering Sea is the semi-enclosed sea separating Alaska and Siberia and is the only direct ocean link between the Arctic and Pacific oceans (Fig. 1). The dimensions of the Bering Sea are ~1,500 km north-south by ~3,000 km along the Aleutian Chain. The northeast half of the basin is the widest

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**Figure 1.** The Bering Sea is about half abyssal with depths greater than 3.5 km while the northeast half is a wide, shallow continental shelf with depths less than 200 m.
continental shelf outside the Arctic Ocean at ~500 km wide (Fig. 1). However, most of this shelf is shallow at <100 m deep with the shelf break ~170 m. The southwest half of the Bering Sea is deep ocean basin with maximum depths of ~4,000 m. The net ocean flow through the Bering Sea is from south to north, that is from Pacific to Arctic. This flow is somewhat restricted by the Aleutian Chain between the Pacific and the Bering Sea, but is more restricted by the Bering Strait (~85 km by ~45 m) between the Bering Sea and Arctic Ocean. The long-term mean transport through the Bering Strait is 1-2 Sv (Coachman and Aagaard 1988), or ~15% of the world ocean inflow to the Arctic Ocean. However, recent estimates are ~1.2 Sv (Roach et al. 1995) or even lower at ~0.7 Sv for the 5-year period 1990-95 (K. Aagaard, Applied Physics Lab, University of Washington, Seattle, WA; and T. Weingartner, SFOS, University of Alaska, Fairbanks, AK, pers. comm.).

The Bering Sea is subjected to large seasonal variations in daylight and insolation that result in large variations in weather and ice patterns.

Figure 2. Eastern Bering Sea shelf region with southern ice limit for January 1975-1979 (from Niebauer 1981).
Figure 3. Maximum, mean, and minimum extent of the ice edge for: (a) March 15 (typically the maximum winter ice extent), and (b) September (typically the minimum summer ice extent) for the region of the Bering Sea. Ice data are means for 1973-1986. (From U.S. Naval Oceanography Command 1986, after Niebauer and Schell 1993.)
Overland (1981) has suggested that the weather and climate of the Bering Sea are related to the presence and fluctuations in sea ice. Fig. 3 shows the region-wide distribution of sea ice in winter and summer. Approximately the northern 75% of the shelf region is ice covered in most winters and the entire Bering Sea is typically ice free in summer. This seasonal sea ice advance and retreat is the largest of any of the Arctic or subarctic regions, averaging ~1,700 km (Walsh and Johnson 1979), while interannual variability has been as great as ~400 km, or ~25% of the seasonal range (Niebauer 1983). Maximum ice cover (Fig. 3a) occurs in March or early April and lags minimum insolation (late December) by ~3 months mainly due to the heat capacity of the ocean. At this point, nearly all the Arctic Ocean is ice covered as well as ~1/2 to 1/3 of the Bering Sea. In any one year, heaviest ice is usually found in the western Bering Sea. In the mean, the maximum ice edge is ~900 km south to southeast of the Bering Strait but this can vary from ~700 km in light (in extent) ice years to ~1,100 km in heavy (in extent) ice years. In an average year, the maximum amount of the Bering Sea that is ice covered is ~37%, but in an icy year, this can go to ~56% while in a light year, ~20% (Plotnikov 1990). Duration of ice cover also varies from a minimum of ~3 months to a maximum of isolated instances of some ice actually lasting through the summer in isolated areas (Plotnikov 1990). Minimum sea ice occurs in September (Fig. 3b), again lagging maximum insolation by ~3 months, with the ice edge in the Chukchi Sea far to the north of the Bering Sea.

Winter ice formation in the Bering Sea has been described by a “conveyor belt” analogy (e.g., Muench and Ahlnas 1976, Pease 1980, Burns et al. 1981). Sea ice forms along the south facing coasts where polynyas are formed as the predominantly northerly winds carry sea ice southward away from the coasts (e.g., Fig. 4). Major polynyas occur south or downwind of the Chukchi Peninsula, St. Lawrence and St. Matthew islands, and the Seward Peninsula (Fig. 4). Sea ice is blown toward the south until it reaches warmer water and melts. The ice edge thus advances southward as the melt water cools the upper ocean layer, but seldom farther than the warmer deep water off the shelf break. Ice drift rates vary from 17 to 22 km/day to as fast as 28-32 km/day (Shapiro and Burns 1975, Muench and Ahlnas 1976, Weeks and Weller 1984, Glueck and Niebauer 1996). In the Bering Strait speeds as fast as 50 km/day have been observed, although there are reversals driven by wind events (Pease 1980). Undeformed ice thickness is ~0.5-1.0 m in the open Bering Sea but rafting does occur.

While most of the ice is formed in situ, there is transport of ice into and out of the Bering Sea through the Bering Strait in the north and through Kamchatka Strait in the far western Aleutians (Yakunin 1987). The average annual ice transport into the Bering Sea from the Arctic Ocean through the Bering Strait is ~3 km³/year for the period 1964-1987, but with a range of ~54 km³/year out of the Bering Sea into the Arctic Ocean (1964) to ~44 km³/year in from the Arctic Ocean (1985). For comparison, in a typical year at maximum ice in late March, the average ice content of the Bering
Figure 4. Recurring polynyas in the eastern Bering and Chukchi seas. (From Niebauer and Schell 1993 as redrawn from Stringer et al. 1982.)
Sea is ~850 km$^3$ (note that this is not the total amount of ice formed because of the “conveyor belt” ice formation analogy described above). In addition to interannual variability through the Bering Strait, there is also an intra-annual pattern superimposed on the ~3 km$^3$/year net inflow through the Bering Strait, with mean ice transport into the Bering in ~January-March but mean transport of ice out ~April-May. Week to week variations are very much larger than this ~3 km$^3$/year, ranging from 45 km$^3$ (in) to 54 km$^3$ (out). To the south, in Kamchatka Strait, the net flow of ice is out of the Bering Sea into the North Pacific in all years with a mean of ~36 km$^3$/year and a range of ~15-90 km$^3$/year (Yakunin 1987).

As mentioned above, polynyas are formed as winds carry sea ice away from the coasts (Fig. 4). Polynyas (defined here as areas of persistent and/or recurring mesoscale size areas of ice-free water despite air temperatures well below the freezing point for sea water) are an important component in air-ice-ocean interaction in the Bering Sea. Polynyas are sites of high rates of heat exchange between atmosphere and ocean (Maykut 1978) resulting in zones of ice formation and brine production. This is true of ice covered seas in general. The Bering Sea polynyas are “latent” heat polynyas (Pease 1980) in which winds or currents cause open water resulting in formation of new ice (as opposed to “sensible” heat polynyas in which ice melts resulting in open water).

As an example of Bering Sea polynyas, the St. Lawrence Island Polynya (SLIP) as shown in Fig. 4 is one of the largest island polynyas in the Bering Sea. In a recent analysis of ERS-1 Synthetic Aperture Radar (SAR) imagery and SAR-derived ice vectors, Glueck and Niebauer (submitted) outlined several SLIP events (openings and closings of a week or two duration). While the SLIP does develop along the north coast of the island under southerly winds, SLIP events occurring off the south coast under the predominant northerly winds are more prevalent and dramatic. At maximum size, the southside SLIP extends tens to hundreds of kilometers offshore by >100 km in the long-shore direction. The SLIP outline has been observed at least 235 km downwind (Glueck and Niebauer, submitted). They also observed maximum ice speeds in the SLIP >30 km/day away from the island. Because they are episodically wind-driven, latent heat polynyas are often characterized by an outline or gradient of ice types becoming older with distance away from the land and open water.

Sea ice production rates are reported in the range of 10-12 cm/day in the SLIP from Cavalieri and Martin (1994) using SMMR, from Glueck and Niebauer (submitted) SAR observations, and from results of Pease’s (1987) analytical model of polynya growth (Glueck and Niebauer, submitted). The calculations agree within ~20%. Associated calculations of brine production rates of ~0.016 Sv (range of 0.006-0.042 Sv) for the February 1992 southern SLIP event (Glueck and Niebauer, submitted) compare favorably with the results of Cavalieri and Martin (1994) of ~0.017 Sv (0.008-0.028 Sv). For comparison, this brine production is nearly identical in magnitude with the mass transport of the Mississippi River of ~0.018 Sv.
Figure 5. Mean (1946-1983) atmospheric surface pressure chart for the Northern Hemisphere for (a) July and (b) January. Isobars are labeled as mb + 1,000 except for pressure below 1,000 mb, which are labeled as actual pressure. (U.S. Naval Fleet Numerical Weather Central in Monterey, CA.)
Dense brine formed as a result of salt rejection during ice formation in polynyas in the Bering Sea helps maintain the Arctic Ocean halocline (Aagaard et al. 1981, Cavalieri and Martin 1994). This brine becomes part of the northward flow across the shelf through the Bering Strait into the Arctic Ocean (Schumacher et al. 1983, Overland and Roach 1987). Aagaard et al. (1981) and Björk (1989, 1990) suggest that 1-2 Sv of brine are required to maintain the Arctic Ocean halocline with an associated renewal time of ~10-40 years (Aagaard et al. 1981, Wallace et al. 1987). Cavalieri and Martin (1994) estimate that the brine contribution from all the polynyas in the Arctic, including the Bering Sea, is ~0.9 Sv (0.7-1.2 Sv). For the western Arctic (i.e., Bering, Chukchi, Beaufort, and East Siberian seas), they estimate that average annual dense water contribution from polynyas is ~0.51 Sv of which the Bering Sea contribution is ~0.059 Sv. Altogether, these calculations suggest that the Bering Sea polynyas contribute ~6% of the brine to the Arctic halocline. SLIP brine production is ~30% of the Bering Sea contribution, ~3% of the western Arctic contribution, and ~1.8% of the total contribution to the maintenance of the Arctic Ocean halocline.

The weather of the Arctic and Bering Sea is dominated by major pressure systems that change seasonally as shown in Fig. 5. In summer (Fig. 5a), the Bering Sea is between the northern portions of a high pressure system over the North Pacific and low pressure over Asia. In the mean (and also generally in a day-to-day sense) the pressure gradients and winds over the Bering Sea in summer are relatively weak.

The change to winter is drastic (Fig. 5b); now the Asian continent is dominated by high pressure while the Aleutian and Icelandic lows dominate the North Pacific (including the Bering Sea) and North Atlantic respectively. (As a working definition, the Aleutian Low is a region of primarily wintertime low atmospheric pressure covering much of the North Pacific and Bering Sea and typically centered along the Aleutian Island chain as shown in Fig. 5b.) The winter pressure gradients are much stronger resulting in intensified winds, especially in storms. The Aleutian Low actually represents a composite of the migrating storms (e.g., Anderson and Gyakum 1989) that dominate the day-to-day weather, and so the Aleutian Low is statistical in nature. In the winter months, three to five storms per month (defined as areas of low sea level pressure [SLP] with closed isobars) move along the Aleutian Islands and into either the Gulf of Alaska or the Bering Sea–Bristol Bay. By contrast, less than two storms per month cross the northern Bering Sea (Overland 1981). This storm track is reflected in the mean wind stress curl (Bond et al. 1994), which exhibits a maximum extending from the southern tip of Kamchatka along or just south of the Aleutians into the Gulf of Alaska. The gradients in the root-mean squared surface stress are smaller, implying a more uniform distribution of wind mixing in the North Pacific and Bering Sea. The wind forcing of the Bering Sea is greater by an order of magnitude in winter than summer due to the seasonal modulation of the Aleutian Low.
The large annual variations in the Bering Sea are accompanied by substantial interannual variations in the air, ice, and ocean environment (e.g., Walsh and Johnson 1979; Niebauer 1980, 1983, 1988; Walsh and Slater 1981; Roger 1981; Overland and Pease 1982) especially during the winter. These interannual variations are driven largely by the atmosphere, specifically through its influence of the horizontal transports of heat, water vapor, and momentum, and the amount of energy radiated to space. The overall sense of the atmospheric circulation anomalies can be described using an index based on the intensity and position of the wintertime Aleutian Low. During the summer the atmospheric circulation tends to be weak and interannual variations are generally not prominent, with the summer of 1997 providing a notable exception. The interannual variability in the summer is expected to be controlled principally through radiative effects, in particular, via the fractional coverage of low cloud decks and their impact on insolation. Interannual fluctuations do not appear to be driven by the Bering Sea ocean circulation because ocean flow, at least on the shelf, is so sluggish (Reed 1978) and because flow from the Pacific through the Aleutian passes is restricted.

While Bering Sea variability appears to depend so much on the Aleutian Low, the Aleutian Low, in turn, responds to variations on a Pacific-wide and global scale (e.g., Barnston and Livezez 1987, Wallace et al. 1990, Trenberth 1990, Bond et al. 1994, Latif and Barnett 1994, Trenberth and Hurrell 1994). During El Niño events, in the Northern Hemisphere winter, the Aleutian Low has been shown to deepen and move southeastward of normal (Bjerknes 1966, 1969, 1972). These El Niño patterns in the Aleutian Low have been related to warming in the Bering Sea (e.g., Niebauer 1988, Niebauer and Day 1989). In some winters, the atmospheric teleconnection mode called the Pacific–North American pattern (PNA, Wallace and Gutzler 1981) is prominent, with substantial effects on the Aleutian Low. Positive PNA events are associated with an intensified Aleutian Low along with ridging over western Canada and Alaska so that warmer air is driven northward. In connecting PNA and El Niño, Niebauer (1988) calculated a correlation coefficient of \(-0.27\) (significant at a 0.01 level) with PNA lagging SOI by \(\sim 2\) months, suggesting that at times positive PNA events are related to negative SOI, or El Niños. Conversely, during La Niña events, the Aleutian Low appears to become less intense and to move westward of normal and this pattern has been related to cooling in the Bering Sea (e.g., Niebauer 1988, Niebauer and Day 1989). In addition, atmospheric circulation over the North Pacific is also sensitive to local SST anomalies, as argued by Namias (1978) among others, and as suggested by recent numerical model results (e.g., Latif and Barnett 1994, see also Discussion below).

These physical weather-climate patterns have also been related to fluctuations in Bering Sea fisheries (e.g., Quinn and Niebauer 1995). For example, a major event, or regime shift, in environmental parameters in the North Pacific and Bering Sea in the late 1970s (e.g., Niebauer 1988, Tren-
berth 1990, Trenberth and Hurrell 1994, Ebbesmeyer et al. 1991) has had major implications for the entire biophysical environment, as found in zooplankton abundance (Brodeur and Ware 1992, Roemmich and McGowan 1995) and groundfish stocks (Hollowed and Wooster 1992).

**Time Series**

**Data and Methods**

The annual cycles for each of the environmental parameters described below and in the following sections are shown in Fig. 6. The actual time series of anomalies are shown in Fig. 7.

The ice data are for the eastern Bering Sea and Chukchi Sea and are a composite of monthly data for 1953-1977 from Walsh and Johnson (1979) and the weekly data for 1978-1994 from the Navy-NOAA Joint Ice Center in Suitland, Maryland. The weekly data are averaged to monthly. The monthly mean percent ice cover was calculated as the ratio of ice cover to the total area (see Niebauer 1981). This gives a semiquantitative estimate of ice cover as it considers all the ice enclosed by the southern limit of ice but without regard to thickness or concentration of ice. Note that the method of data collection, and thus the uniformity of the data collection, have varied over time. Satellite observations were not available until ~1973. The Electrically Scanning Microwave Radiometer (ESMR) only operated from 1973 to 1976 (Parkinson et al. 1987) but Scanning Multichannel Microwave Radiometer (SMMR) began in fall 1979 (e.g., Gloersen et al. 1992). While there was a shift in the way satellite data were collected in the period of the regime shift, the effects on the time series of anomalies appear minimal.

Monthly mean sea surface temperature (SST) for the period 1947-1994 were obtained from the Climate Research Group, Scripps Institution of Oceanography, UCSD. The SSTs used here are 5° × 5° means centered south of the Pribilof Islands (55°N, 170°W) on the edge of the Bering Sea shelf but ~500 km from shore.

Air temperatures and surface winds for 1947-1994 and 1965-1994 respectively were obtained from the National Weather Service Local Climatological Data for St. Paul Island in the Pribilof Islands. The magnitude of the wind is resolved along due north-south but the wind anomalies still require caution in interpretation because, as plotted in Fig. 7, the actual direction of the anomaly is not indicated. In the Pribilofs, the monthly mean winds are from the north in all months except July and August (Fig. 6). Therefore, a negative anomaly can mean either less-strong winds from the north (i.e., less than the long-term mean for that month) or actual southerly flow.

The interannual variations in the weather patterns affecting the North Pacific and Bering Sea are described here using the North Pacific (NP) and West Pacific (WP) indices. These two indices characterize the teleconnection modes identified by Barnston and Livezey (1987) that have large

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Figure 6. The annual cycles for each of the Bering Sea air, ice, and ocean physical environmental parameters and the length of each data set.
Figure 7. Deviations from monthly mean air, ice, and ocean physical environmental parameters from the Bering Sea, as well as the Southern Oscillation Index (SOI) from the south Pacific. NP = North Pacific index, WP = western Pacific index, SST = sea surface temperature, and WIND (n/s) = north-south component of the wind. For the wind, positive anomalies mean either more strongly from the north or less strongly from the south (see Fig. 6). All of these data are smoothed with a 9-month running mean.
amplitudes in the Bering Sea. The NP index is defined (after Trenberth and Hurrell 1994) as the areal averaged monthly mean surface air pressure for the mid- to North Pacific (27.5-60°N by 140°W-160°E). The actual sea level pressure used to calculate the NP index for 1947-1994 are from the Climate Research Group of Scripps Institution of Oceanography. Previously, in Niebauer (1988) and Niebauer and Day (1989), the Pacific-North American (PNA) index of Wallace and Gutzler (1981) was used but this index uses points from Hawaii, the North Pacific, western Canada, and the southeast United States. We wanted to concentrate in the North Pacific. The PNA and NP indices appear to measure the same phenomenon as their correlation coefficient = –0.82 (the 99% confidence level = –0.14) with no time lag, but falls off precipitously to –0.18 within about one month.

WP is a north-south dipole over the western North Pacific. The WP index is defined following Wallace and Gutzler (1981) as the difference in the normalized 500 mb height anomalies between two points (60°N, 155°E and 30°N, 155°E). Since the pattern associated with the WP index has an equivalent barotropic structure, it also characterizes the primary fluctuations in the sea-level pressure field in the western Pacific. A negative WP anomaly indicates a stronger Aleutian Low west of 180° with the stormtrack displaced northward, while a positive anomaly indicates a weaker Aleutian Low west of 180° with the stormtrack displaced southward.

The Southern Oscillation Index (SOI) is calculated as the difference between normalized sea level pressure (SLP) from Tahiti in the Society Islands in the South Pacific, and Darwin, Australia. These data were obtained from the Climate Diagnostics Bulletin of NOAA. El Niño events are negative anomalies (e.g., 1951-52, 57-58, 65-66, 72-73, 77-78, 82-83, 86-87, and 91-93) while La Niña events are positive anomalies (e.g., 1949-51, 55-56, 70-71, 73-74, 75-76, and 88-89).

In all of the time series in Fig. 7, we removed the mean annual cycles (Fig. 6) in an attempt to isolate the monthly deviations from the mean (except for the SOI which is already an anomaly series). We did this simply by averaging all the Januarys, all the Februarys, etc., and then subtracting the mean January, February, etc., from each individual January, February, etc. The resultant time series are shown in Fig. 7 smoothed with a 9-month running mean. Each of the unsmoothed time series were cross-correlated with all the other time series, leading and lagging one dataset with the other for up to 3 years (Fig. 8 and Table 1). Significance levels were calculated (Zar 1984) with an effective number of degrees of freedom ($N_{eff}$) calculated according to Trenberth (1984). $N_{eff}$ depends upon autocorrelation such that the stronger the autocorrelation, the fewer the degrees of freedom.

**Annual Means**

The North Pacific index (NP) reaches its maximum in July, falling to a minimum in January (Fig. 6). This is indicative of the North Pacific High dominance in the Northern Hemisphere (NH) summer (Fig. 5a) followed by the
Figure 8. Cross-correlation plots of selected Bering Sea and north and south Pacific time series of Fig. 7: (a) ICE = sea ice vs. Bering Sea air temperatures, (b) SST = Bering Sea surface temperature vs. ICE, (c) ICE vs. WP = west Pacific index, (d) ICE vs. SOI = Southern Oscillation Index, (e) NP = north Pacific index vs. SOI, and (f) WP vs. SOI. The x-axis is the time lag between datasets. N is the number of data points while $N_{eff}$ is the number of effective or independent data points available for use in calculating the confidence levels shown on each plot.
Table 1. Table of correlation coefficients matrix of monthly mean times series of deviations from mean percent ice cover (ICE), sea surface temperatures (SST), winds, air temperatures (AT), North Pacific (NP) and West Pacific (WP) indices from the Bering Sea, Niño 1+2 SST (N1+2) and the southern oscillation index (SOI) from the south Pacific as illustrated in Fig. 6.

<table>
<thead>
<tr>
<th></th>
<th>SST</th>
<th>WIND</th>
<th>AT</th>
<th>NP</th>
<th>WP</th>
<th>N1+2</th>
<th>SOI</th>
</tr>
</thead>
<tbody>
<tr>
<td>SST</td>
<td>-0.17</td>
<td>-0.12</td>
<td>0.26</td>
<td>0.14</td>
<td>0.24</td>
<td>-0.17</td>
<td>0.17</td>
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<tr>
<td></td>
<td>(0-1, SST lags, 0.95)</td>
<td>(3, SST lags, 0.95)</td>
<td>(0, no lag, 0.99)</td>
<td>(3, SST lags, 0.95)</td>
<td>(0, no lag, 0.99)</td>
<td>(-1, SST lags, 0.95)</td>
<td>(2, SST lags, 0.95)</td>
</tr>
<tr>
<td>WIND</td>
<td>0.32</td>
<td>-0.48</td>
<td>0.26</td>
<td>0.14</td>
<td>0.24</td>
<td>-0.17</td>
<td>0.17</td>
</tr>
<tr>
<td></td>
<td>(0-1, ice lags, 0.99)</td>
<td>(0, no lag, 0.99)</td>
<td>(0, no lag, 0.99)</td>
<td>(3, SST lags, 0.95)</td>
<td>(0, no lag, 0.99)</td>
<td>(-1, SST lags, 0.95)</td>
<td>(2, SST lags, 0.95)</td>
</tr>
<tr>
<td>AT</td>
<td>-0.51</td>
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<td>-0.11</td>
<td>-0.11</td>
<td>-0.41</td>
<td>0.13</td>
<td></td>
</tr>
<tr>
<td></td>
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<td>(0, no lag, 0.99)</td>
<td>(4, AT lags, 0.95)</td>
<td>(0, no lag, 0.99)</td>
<td>(0, no lag, 0.99)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>NP</td>
<td>0.14</td>
<td>-0.1</td>
<td>0.26</td>
<td>0.24</td>
<td>0.19</td>
<td>-0.48</td>
<td></td>
</tr>
<tr>
<td></td>
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<td>(0, no lag, 0.99)</td>
<td>(0, no lag, 0.99)</td>
<td>(0, no lag, 0.99)</td>
<td>(1 to 3, N1+2 lags, 0.99)</td>
<td></td>
</tr>
<tr>
<td>WP</td>
<td>0.24</td>
<td>-0.09-0.14</td>
<td>0.55</td>
<td>0.15</td>
<td>0.15</td>
<td>-0.48</td>
<td></td>
</tr>
<tr>
<td></td>
<td>(0, no lag, 0.99)</td>
<td>(3, SST lags, 0.95)</td>
<td>(0, no lag, 0.99)</td>
<td>(1, AT lags, 0.95)</td>
<td>(0.6, WP lags, 0.99)</td>
<td>(1 to 3, N1+2 lags, 0.99)</td>
<td></td>
</tr>
<tr>
<td>N1+2</td>
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<td>0.17</td>
<td>0.11</td>
<td>0.15</td>
<td>-0.16</td>
<td>0.19</td>
<td></td>
</tr>
<tr>
<td></td>
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<td>(-17, SST lags, 0.95)</td>
<td>(7, AT lags, 0.90)</td>
<td>(1, AT lags, 0.95)</td>
<td>(-1 to 2, NP lags, 0.99)</td>
<td>(0, WP lags, 0.99)</td>
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</tr>
<tr>
<td>SOI</td>
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<td>-0.12</td>
<td>INCONCLUSIVE</td>
<td>-0.21</td>
<td>0.24</td>
<td>-0.48</td>
<td></td>
</tr>
<tr>
<td></td>
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<td>(0-5, SST lags, 0.90)</td>
<td>(18, AT lags, 0.99)</td>
<td>(0, no lag, 0.99)</td>
<td>(0, WP lags, 0.99)</td>
<td>(1 to 3, N1+2 lags, 0.99)</td>
<td></td>
</tr>
</tbody>
</table>

The data in the first set of parentheses indicate the lag, in months, when maximum correlation occurred, which dataset lags, and the significance level. The numbers in the second set of parentheses indicate the number of months the correlations are greater than the 95% significance level. "INCONCLUSIVE" indicates that there are so many correlation peaks that the analyses are inconclusive.
dominance of the Aleutian Low dominance in the NH winter (Fig. 5b). The climb from the winter minimum to the summer high is not monotonic but rather plateaus for April-June before hitting the July high. The WP index is defined in terms of monthly anomalies and so its annual mean is identically zero.

Air temperatures in the Pribilof Islands lag the NP by one month with maximum temperatures of ~9°C occurring in August and minimum temperatures of ~−5°C occurring in February. There is no plateau in temperature as in the pressure index.

The north-south component of the mean monthly surface wind from the Pribilof Islands is more complex than the other annual cycles (Fig. 6). The winds are southerly only in July and August and undergo a relatively rapid shift to northerly during early autumn. The minimum (maximum) in the northerly component observed in November (February) reflects a relatively high (low) incidence of cyclonic storms passing to the north and west.

Minimum sea ice occurs in September (Figs. 3, 6) with the ice edge in the Chukchi Sea far to the north of the Bering Sea. The Bering Sea is sea ice–free in summer. Maximum ice cover occurs in February-March (Figs. 3, 6). The seasonal latitudinal excursion of the ice is ~1,400 km.

Finally, sea surface temperature (SST) from the southeast Bering Sea (Fig. 6) peaks at just under 10°C in August before falling to the annual minimum of ~3.5°C in March.

**Time Series Analyses**

As remarked above, the time series shown in Fig. 7 have all been cross-correlated with one another within the Bering Sea, with indicators of ENSO activity in the South Pacific (SOI and Niño 1+2), and with time series of variability in the Aleutian Low (NP, WP). These cross-correlation statistics are shown in Table 1. The purpose of these analyses are to try to understand the extraordinary interannual variability, including the regime shift observed in the air, ice, and ocean environment of the Bering Sea as pointed out above. For comparison and update, a similar analysis was reported on these same datasets (except for the NP and WP indices) by Niebauer (1988) and Niebauer and Day (1989) when the datasets spanned 20-33 years (~1953-1985) as compared to the present >45 years (~1947-1995).

Most of the basic conclusions (summarized in this section) have not changed. However, a regime shift that occurred in the mid-late 1970s (e.g., air temperature and ice in Fig. 7), has affected the correlations shown in Fig. 8 and Table 1. For example, the magnitude of the correlation coefficients dropped an average of ~28% between the years 1974-1982 (Niebauer 1988) and this study. The reasons are related in the next section.

The interannual variability in the Pribilof air temperatures, surface winds and SST, and sea ice are nearly in phase with some lags (e.g., Fig. 8 and Table 1). The hierarchy of lags are that air temperatures, ice, and SST
lag winds by 0-4 months, SST and ice lag air temperatures by 0-3 months, and SST lags ice by 0-1 month. This suggests that the interannual variability in the southern Bering Sea is primarily driven by the atmosphere rather than the ocean. This is consistent with the sluggish flow of 0.01-0.03 m/s or 0.9-2.6 km/day over this 500 km wide shelf and with the heat budget studies on the Bering Sea shelf by Reed (1978). On the other hand, stronger currents tend to occur in the western portion of the Bering Sea, and here oceanic heat transports may have a significant effect on ice extent.

The Bering Sea time-series are also significantly correlated with the El Niño-Southern Oscillation (ENSO) signals (Fig. 8 and Table 1) from the south Pacific as well as the NP and WP (Fig. 8 and Table 1) for the North Pacific. Because a significant amount of the Bering Sea air-ice-ocean climate depends upon the Aleutian Low weather system and because it has been known for at least 30 years that the Aleutian Low depends, at least in part, upon the Southern Oscillation (Bjerknes 1966, 1969, 1972), it is worthwhile to discuss briefly the El Niño-Southern Oscillation (ENSO).

The Southern Oscillation is a basin-scale seesaw in anomalous sea level pressure between the extreme western portion and the central and eastern portion of the tropical Pacific. Extremes in this dipole in anomalous pressure are associated with El Niño and La Niña events, with the former characterized by high pressure to the west and low pressure to the east, and vice versa. During El Niños, these pressure anomalies cause the normal easterlies along the equator to weaken, or with intense cases, to reverse to westerly. On the other hand, La Niñas feature an enhancement of the normal west to east pressure gradient, and hence stronger than normal easterlies. These anomalies in the surface winds affect the upper tropical Pacific in the following manner. Easterly surface winds tend to cause higher sea level, a deeper thermocline, and relatively warm SSTs in the western Pacific. These easterlies contribute to upwelling along the South American coast and the equator, causing a shallower thermocline and lower SSTs in the eastern Pacific. This SST pattern in the tropical Pacific, in turn, can positively reinforce the winds through its impact on atmospheric boundary layer temperatures and deep cumulus convection and hence sea level pressure (e.g., Wyrtki 1982). In essence, El Niños (La Niñas) represent a relaxation (intensification) of the east-west asymmetry in the coupled atmosphere-ocean system of the tropical Pacific. A typical evolution for the tropical Pacific (e.g., the canonical ENSO event as composited by Rasmusson and Carpenter 1982) includes a positive SOI anomaly, or La Niñas, followed by a negative SOI anomaly, or El Niño (e.g., 1970-1972 or 1975-1977 in Fig. 7). But the fluctuations of the tropical Pacific climate system are only quasi-periodic on time scales of 2-7 years; substantial decadal variations occur in the background state and in the frequency of individual El Niño and La Niña events (Fig 7).

The mechanism(s) by which ENSO anomalies affect the rest of the globe have been studied using observations (e.g., Horel and Wallace 1981, Lau and Wallace 1979, Mantua et al. 1997, Minobe 1997, Trenberth 1990,
Trenberth and Hurrell 1994) and numerical models (Graham et al. 1994, Graham 1995, Kumar et al. 1994, Meehl et al. 1993). These results suggest that it is SST anomalies along the equator that force anomalies in deep cumulus convection, which in turn are associated with anomalies in the upper-tropospheric flow gyres that straddle the equator. These perturbations at upper levels interact with the mid-latitude westerlies, bringing about the wide-ranging impacts of ENSO. An important point is that ENSO’s effect on the higher-latitude circulation depends on the higher-latitude base state, which is subject to other competing factors. This means the forcing by ENSO of the global circulation is not highly deterministic.

Comparing the Bering Sea time series with the ENSO time series, the Bering Sea time series are significantly correlated with, but typically lag behind, the SOI (Fig. 8 and Table 1), with perhaps the exception of the winds. The lag ranges 0-12 months for correlation of CI > 0.95 (Table 1). The correlations with N1+2 tend to be poorer but are significant for Bering Sea SST and AT. The correlations suggest that warming in the Bering Sea (i.e., below normal ice and winds and above normal SST and AT) follows negative anomalies in the SOI, or El Niños. La Niñas, or positive anomalies, in the SOI tend to precede cooling in the Bering Sea as illustrated by above normal ice and northerly winds and below normal SST and AT.

The lag-correlations have been used to indicate which parameters are most strongly related, but additional information is generally necessary to identify the causal mechanisms for these relationships. We have suggested previously that a significant connection between the ENSO events in the Southern Hemisphere and interannual variability in the Bering Sea is atmospheric and is the wintertime intensity and position of the Aleutian Low. In the mean from December to March, the center of the Aleutian Low is located over the Aleutian Islands (Fig. 9a). During El Niño winters (Fig. 9b), the Aleutian Low is located slightly southeastward of its usual position, and is ~2 mb lower in central pressure. During La Niña winters (Fig. 9c), the Aleutian Low is displaced westward of its usual position, and is ~3 mb higher in central pressure. These differences in mean sea-level pressure, as also shown using anomaly maps (Figs. 10a,b), represent the cumulative effects of variations in the tracks and character of high-frequency weather disturbances. There is the tendency for more or stronger low-pressure cyclonic storms, and fewer or weaker high-pressure anticyclones, in the Bering Sea during El Niño winters, and vice versa. These transient disturbances are responsible for large meridional heat fluxes (e.g., Lau and Wallace 1979). A net poleward flux of warmer air tends to accompany cyclones, and a net equatorward flux of colder air tends to accompany anticyclones. Therefore, El Niño winters with their preponderance of cyclonic systems tend to be warm in the Bering Sea, especially in its southern portion, and La Niña winters favoring anticyclones tend to be cold. The heat fluxes due to the transient disturbances can be enhanced or suppressed by the temperature advection associated with anomalies in the mean meridional wind; as shown by Table 1 and Fig. 8, a systematic
Figure 9. Sea level pressure (SLP) patterns for the north Pacific and Bering Sea winters (December-March) for (a) all the winters for 1947-94, (b) all the El Niño winters (i.e., 1951-52, 57-58, 65-66, 72-73, 77-78, 82-83, 86-87 and 91-93), and (c) all the La Niña winters (i.e., 1949-51, 55-56, 70-71, 73-74, 75-76 and 88-89). Pressure less than 1,004 mb is shaded for comparison. (After Niebauer, submitted.)
Figure 10. (a) El Niño winters minus mean winters for 1947-1994 (i.e., Fig. 9b minus Fig. 9a). Note the large area of drop in pressure (mb) over the Bering Sea and eastern North Pacific as the Aleutian Low intensifies and moves east and south of normal in teleconnection with El Niño events in the Southern Hemisphere. (b) La Niña winters minus mean winters for 1947-1994 (i.e., Fig. 9c minus Fig. 9a). Note the large area of rise in pressure (mb) over the southeastern Bering Sea and north Pacific as the Aleutian Low weakens and moves westward of normal in teleconnection with La Niña events in the Southern Hemisphere. (After Niebauer, submitted.)
relationship between ENSO and the mean meridional wind is lacking, but during any particular winter the latter’s (i.e., meridional wind) effects can be significant to the mean temperature.

There is additional evidence supporting this seesaw relationship between north-south interannual variability in the ice extent in the eastern Bering Sea, and the interannual variability in the position of the Aleutian Low (e.g., Niebauer 1988, Parkinson 1990, Fang and Wallace 1994). Sea ice extents in the Bering Sea and the Sea of Okhotsk are negatively correlated (Fig. 11), implying that the air temperatures in these regions are also negatively correlated. This result can be attributed to the typical horizontal scale (1,000-2,000 km) of the individual weather disturbances that account for most of the wintertime forcing of the Bering Sea. When storms are causing relatively warm conditions in the southeastern Bering, relatively cool air tends to be drawn down over the Sea of Okhotsk, and vice versa. The tendency for one state or the other will be reflected in the longitudinal position of the Aleutian Low. We do note that this is not the whole explanation as the correlation analysis explains only 10% of the variability. Further, Plotnikov (1990) points to a recent (~10 years) increase

Figure 11. Same as Fig. 8 except the data are deviations from mean ice cover for the Sea of Okhotsk and the Bering Sea for January 1973-March 1984. (From Niebauer 1988.)
in frequency of cases where this relationship does not hold very well, suggesting possible climatic changes in the hydrometeorology.

**Interannual Variability for 1975-94**

While a "typical" ENSO event has been described (e.g., Rasmusson and Carpenter 1982) as a coupled La Niña, or high intensity event, followed within a year or so by an El Niño, or low intensity event (e.g., 1956-1958 or 1970-1972 in Fig. 7), we note that there has not been a "typical" ENSO event since the mid-1970s (i.e., since the regime shift). This change in the ENSO patterns appears to have caused changes in the position and intensity of anomalies in the Aleutian Low associated with El Niños (Niebauer submitted; Fig. 12). To illustrate, there have been ∼4 El Niños but only two La Niñas (see the SOI panel in Fig. 7) since 1976. The ENSO event of 1974-1978 actually had two La Niñas preceding an El Niño. This is also the ENSO event that occurred during extraordinarily strong shifts in the Bering Sea environment (e.g., especially air temperature in Fig. 7, and ice in Figs. 2, 7, and 13). Following that, the intense El Niño of 1982-1983, as well as the El Niño of 1986-1987, did not have a preceding La Niña. While there was a strong La Niña in 1988-1989 followed by an El Niño, the El Niño lasted
about 4 years. (It is interesting to note that this 4-5 year-long El Niño coincides with the reduced flow of ~0.7 Sv through the Bering Strait over this same period, 1990-1995, as observed by Aagaard and Weingartner [pers. comm.]). Finally, the ratio of El Niño:La Niña has gone from 0.8 for 1947-1977 to 2.8 for 1978-1996 (Niebauer, submitted).

This shift has had a strong effect on the relationship between the Bering Sea ice cover and the SOI. The cross correlation between the SOI and Bering Sea ice for the entire period 1953-1995 is 0.17 (Fig. 8 and Table 1). However, for the period 1953-1978 (i.e., before the regime shift), the correlation coefficient was ~0.26, while after the regime shift, for the period 1978-1995, the correlation coefficient actually changes sign to ~ −0.18 (Niebauer, submitted). The change in sign of the correlation, combined with the lack of La Niña events, means that since the regime shift, increases in ice are now associated with El Niño events. However, this is consistent with El Niño-associated Aleutian Lows becoming even more intense and moving even farther east following the regime shift. These more easterly Aleutian Low pressure patterns are forcing southerly winds off the
Pacific into interior Yukon and Alaska and easterly winds carrying continental air from Alaska out over the Bering Sea. Before the regime shift, the Aleutian Low was farther west in association with a high frequency of storms and southerly winds, and hence warmth, over the Bering Sea.

**Discussion**

As mentioned earlier, in some winters, fluctuations associated with other teleconnection modes (e.g., the western Pacific oscillation [WPO or WP; Wallace and Gutzler 1981, Barnston and Livezey 1987] and/or the Pacific-North American pattern [PNA, Wallace and Gutzler 1981]) can have substantial effects on the Aleutian Low seemingly independent of ENSO events. For example, there is significant correlation between WP and SOI undoubtedly through association with the Aleutian Low. For the entire period 1947-1993, as well as for the period before the shift (1947-1977), the correlation between WP and SOI is ~0.2 with WP lagging SOI by 0 and 4 months (e.g., Fig. 8 and Table 1). After the shift, the correlation at 0 lag dropped a little to ~0.17 while the 4-month lag peak was greatly diminished. For WP and ice (1953-1996), the correlation is 0.23 (ice lagging 0-2 months) which is higher than between SOI and ice (0.18; Fig. 8 and Table 1). For the period after the regime shift (1978-1996), the correlation between WP and ice (0.30) is greater than between SOI and ice (~0.18), but of opposite sign. However, for the period before the regime shift (1953-1977), the correlation between SOI and ice (0.26) is slightly greater than the correlation between WP and ice (0.21 with no lag between WP and ice). (All of these cross correlations are significant at >0.1 level although none explain more than 10% of the variability.)

An important point here is that the significant correlation between SOI and ice lies over a broad period of lag (at least 12 months) as compared to a narrow range (0-2 months) for WP and ice. The narrow, sharp peak of WP and ice correlation (as well as WP vs. AT and wind in Table 1) suggests that WP and ice fluctuations are probably parts of the same or a local process (i.e., shift in location of the Aleutian Low) but different from the SO teleconnection; that ENSO events are a significant driving force for both WP and ice through teleconnection of the Aleutian Low with the Southern Oscillation. Niebauer (submitted) showed that average winter (Nov.-Mar.) ice was ~2.2% above normal for the 1950s to the regime shift while the average winter SOI was ~0.12 mb above normal. After the regime shift, both ice and SOI dropped to ~3.1% and ~0.7 respectively. This relationship between SOI and ice provides the possibility of predicting ice conditions further in the future because of this longer duration lag of statistically significant correlation, and because of (or perhaps in spite of) the change in sign of correlation between ice and SOI through the climate shift.

Against the larger picture of decadal time scales (i.e., Pacific-wide response and global warming), Trenberth and Hoar (1996) consider two questions: (1)"Is this pattern of change a manifestation of the global warming
and related climate change associated with increases in greenhouse gases in the atmosphere?”, or (2) “Is this pattern a natural decadal-timescale variation?” For the second question, Trenberth and Hoar (1996) analyze normalized SOI (negative Darwin sea level pressure) anomalies from the 1880s-1990s. From a statistical viewpoint, they suggest that the shift to more El Niños vs. La Niñas since 1976-1977, as well as the ~5 year long El Niño from 1990 to 1995, is unprecedented with a probability of occurrence about 1 in 2,000 years.

However, more recently, Mantua et al. (1997), Minobe (1997), and Zhang et al. (1997) show evidence of four reversals, or climatic regime shifts, around 1890, 1925, 1947, and 1977, in a recurring pattern of very large scale, interdecadal atmosphere-ocean climate variability centered over the midlatitude North Pacific. Mantua et al. (1997) call this the Pacific (inter) Decadal Oscillation or PDO. Minobe (1997), using reconstructed climate records (tree rings) for the last three centuries, presents evidence of a 50-70 year oscillation from the Eighteenth Century to the present. Minobe (1997) suggests that 50-70 year oscillation seen in ~300 year record is likely an oscillation of the ocean-atmosphere coupled system, although he suggests it may be modulated by solar radiation in the Twentieth Century.

The causes of decadal variability in the North Pacific is currently attracting a great deal of interest. Essentially, two different explanations have been offered. Certainly decadal or longer variations in ENSO are important, through their influence on the Northern Hemisphere weather via atmospheric teleconnections (e.g., Kumar et al. 1994, Graham et al. 1994, Graham 1995). The long-term variations in the systematic sense of ENSO, or the relative frequency of individual events, may be due both to ENSO’s intrinsic variability (e.g., as modeled by Zebiak and Cane 1987) and to external forcing, such as increased concentrations of greenhouse gases (e.g., as modeled by Meehl et al. 1993). Alternatively, or in conjunction, long-term fluctuations in the North Pacific may be associated with local, positive feedbacks between the atmosphere and ocean. Anomalous SSTs may cause significant surface heat and moisture flux anomalies that in turn reinforce the atmospheric anomalies impacting the ocean (e.g., Latif and Barnett 1994, Trenberth and Hurrel 1994). The natural time scales for the North Pacific gyre circulations may then be responsible for the decadal time scales that have been observed. Regardless of their cause, these decadal variations of the North Pacific need to be recognized for their role in providing a varying background or “basic state” for the Bering Sea atmosphere-ice-ocean system.

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