3.3 Physical Oceanography of the Fisheries Management Units

3.3.1 The Northeast Pacific Ocean

3.3.1.1 Description

Bounded on the north and east by the North America land mass, and essentially open to the west and south, the northeast "quadrant" of the Pacific Ocean includes the GOA and the Bering Sea (Figure 3.3-1). Although separated from the main ocean body by the Aleutian Islands, the Bering Sea is considered to be a northern extension of the northeast Pacific Ocean by virtue of hydraulic communication through the numerous passes and channels between the islands. On the west and south, the bounds of the northeast Pacific Ocean are generally considered to be the International Dateline and the northern 30th parallel, respectively.

Although dotted by numerous seamounts rising to within 1,000 meters (m) of the surface, seabed depths over most of the northeast Pacific Ocean tend to be greater than 4,000 m. Maximum depths of more than 7,000 m occur in the Aleutian Trench, which parallels and marks the southern base of the Aleutian Island chain (Figure 3.3-1). Along the land boundary, the continental shelf (depth less than or equal to [≤] 200 m) is relatively narrow (less than [<] 50 kilometers [km]) along the British Columbia and southeast Alaska coasts, and then broadens to 100 km or more along southcentral Alaska coast. Along portions of the Kenai and Alaska Peninsulas, the continental shelf attains a width of nearly 200 km.

3.3.1.2 Circulation

Surface currents in the Pacific Ocean are driven by the trade winds and westerlies, such that surface flows are predominantly westward in low latitudes (10° - 30°North [N]) and eastward in high latitudes (35° - 50°N). When these flows encounter the continents they are diverted both north and south to form coastal currents, which further serve to establish rotating water masses (gyres) that characterize the overall circulation patterns of the ocean (Figure 3.3-1).

The seaward "boundaries" of the northeast Pacific Ocean are arbitrarily, if not practically, determined by large-scale circulation features that result from these planetary driving forces. On the south, the North Pacific Drift transports surface waters eastward along (approximately) the 45th parallel. Upon reaching the North American continent, this flow splits into northbound and southbound branches known, respectively, as the Alaska Current and the California Current. As the Alaska Current tracks anti-clockwise along the continental margin, portions of it are known as the Alaska Coastal Current (ACC) and the Alaskan Stream (Figure 3.3-1). The anti-clockwise loop is closed by the Aleutian Current, which is a south-to-southeasterly extension of the Alaska Current (Figure 3.3-2). The resulting anti-clockwise circulation pattern is known as the Alaskan Gyre.

Winter intensification of the Aleutian Low leads to strong southeasterly winds along the southeast Alaska coast, which produce onshore Ekman transport and downwelling of coastal waters (Royer 1975). During summer, the North Pacific High tends to dominate the region such that lighter, more variable winds result in a relaxation of the coastal convergence. The overall anti-clockwise circulation is maintained by the introduction of fresh water along the coastline, as described in the following section.

3.3.1.3 Water Mass Characteristics

In the North Pacific high latitudes, surface waters have relatively low salinities because of the excess of precipitation and runoff over evaporation. Cooling these surface waters even to the freezing point does not make them sufficiently dense to cause them to descend any deeper than 200 m in the water column. Consequently, the deeper water in the North Pacific must originate elsewhere, and must flow in through the South Pacific because the connection with the Arctic Ocean, through the Bering Strait, is too narrow and shallow to be of consequence.

These deeper waters of the North Pacific originate in the southern (i.e. Antarctic) and North Atlantic Oceans, where the combination of surface temperatures and salinities produces very dense waters that subsequently sink to the sea floor. The Pacific Ocean has been described as a vast estuary, with low-salinity surface outflow from the North Pacific mixing with deeper, more saline water flowing in at depth through the south Pacific. Ultimately, the increasingly dense North Pacific water returns to the areas of sinking in the North Atlantic to complete the circuit, which is estimated to take centuries to complete.

Nutrients are distributed throughout the world's oceans by this system of deep circulation. For example, inorganic phosphates are consumed by plant growth at the surface and are regenerated at greater depths as the plants die, sink, and decay. Consequently, nutrients are in greater concentrations at depths of one to two km than at the surface. Inflow of the deeper water into the Pacific Ocean brings in water that is high in phosphate compared to the average concentration in the Atlantic Ocean. As a result, the accumulated phosphate in the Pacific Ocean has a concentration about twice that of the Atlantic.

The next two subsections describe in greater detail the physical oceanography of the two federal fisheries management units (FMUs) of the northeast Pacific Ocean. A final subsection addresses the sources and magnitude of variability in oceanic parameters.

3.3.2 Gulf of Alaska Fishery Management Unit

3.3.2.1 Description

The GOA FMU includes all waters within the EEZ along the southeastern, southcentral, and southwestern coasts of Alaska from Dixon Entrance to Unimak Pass, a distance along the Alaskan coastline of more than 2,500 km (Figure 1.2-3). Greatest depths within the GOA FMU range from 3,000 m off southeast Alaska, to 4,000 m off southcentral, and to 7,000 m at the west end of the FMU, where the Aleutian Trench begins. However, the continental shelf areas (depths < 200 m) are of greatest importance in the context of fishery management issues.

As noted previously, the continental shelf within the GOA FMU is narrowest in southeast Alaska, ranging in width from less than 50 km between Dixon Entrance and Cape Spencer, and then broadening to 100 km or more along the southcentral coast to Seward. South of the Kenai Peninsula and west of Kodiak Island, the continental shelf is broadest, about 200 km, on Portlock Bank. Proceeding westward from Kodiak along the Alaska Peninsula, the shelf narrows gradually from 150 km to about 50 km at Unimak Pass. The progressive broadening and narrowing of the continental shelf from east to west plays an important role in the circulation of waters through the GOA FMU.

3.3.2.2 Circulation

Water movements within the GOA FMU are dominated by the ACC which changes character and direction three times and is joined by other narrower currents as it is forced by the coastline to change direction from northwestward to westward to southwestward as it flows through the unit (Figure 3.3-1). Starting off southeast Alaska like a wide river with imbedded eddies the main flow turns westward with the coastline and becomes two currents as it is joined by the faster ACC close to shore. As the coastline turns southwestward the flow seaward of the shelf break accelerates taking on the dynamics of a western boundary current, the Alaskan Stream, which reaches speeds of 60 to 100 centimeters per second (cm/sec) staying in a narrow jet over the continental slope to the end of the unit. This broad southwestward flow is now in four bands; the weak offshore portion, the swift Alaskan Stream, a weak tidally and bathymetry influenced flow on the outer shelf, and the moderate ACC inshore. Some of the offshore flow recirculates to the south then east forming the western branch of the GOA Gyre. This coastal circulation is driven in winter by the persistent anti-clockwise wind stress over the GOA and in summer by the immense fresh water input from coastal sources in British Columbia and southeast Alaska.

During the winter, when coastal runoff is minimal, anti-clockwise atmospheric circulation is most intense over the GOA, and wind stress maintains the coastal circulation with strong onshore convergence or downwelling. During summer when winds over the GOA slacken considerably, coastal runoff increases dramatically and creates a density gradient in nearshore waters that serves to maintain the anti-clockwise coastal circulation. Thus seasonal variations in wind stress and coastal runoff are balanced so that together they serve to maintain the generally steady westward movement of water through the GOA FMU.

Circulation near the continental shelf break (approximately [~] 200 m depth) generally follows the isobaths, with frequent eddies and meanders. Closer to shore the flow is more stable with fewer eddies and is more closely aligned with the coastline. Within the broader Alaska Current, the narrow and intense coastal current ACC extends from southeast Alaska to Kodiak Island. The ACC results from the density gradient produced by prodigious amounts of freshwater runoff that varies with the annual hydrologic cycle (Royer 1979 and 1983). The width of the ACC varies from only 5 to 10 km wide to as much as 40 to 50 km, depending on the rate of freshwater input. Current speeds within the ACC occasionally exceed 100 cm/sec, which has caused occasional reference to it as a coastal jet. The dilutional effects of the freshwater input are generally confined to the top layers (50 to 100 m) of the water column. The western segment of the ACC has been called the Kenai Current (Schumacher 1980).

West of Kodiak Island, where freshwater input is much reduced, the Alaska Current is driven more by prevailing winds. Accordingly, in winter a westward flow is maintained by wind stress, but in summer this driving force is somewhat lessened so current reversals and eddies occasionally occur (Schumacher and Reed 1986).

3.3.2.3 Water Column

The density structure of the water column is determined by its physical properties, most notably its temperature and salinity, as they vary with depth. At the temperatures typical of the northern GOA water (i.e., < 10 degrees Celsius [° C]), salinity is the dominant determinant of water density. Because of the plentiful coastal runoff and the excess of precipitation over evaporation, coastal waters of the GOA have salinities that are significantly lower than those of the North Pacific, which are already low relative to the world's oceans. Salinities at depths less than about 10 m in the GOA FMU are typically 25 to 30 practical salinity units (psu).

Salinity and density increase with depth, but the greatest rate of increase occurs within the pycnocline, which extends from about 30 m to 200 m depth. Above the pycnocline is the surface mixed-layer, in which the salinity is 32 to 33 psu. Below the pycnocline, salinity increases slowly to about 34.4 psu at a depth of 1,500 m. Temperatures in the mixed layer vary from 3° to 12°C seasonally. Below the pycnocline the temperature decreases slowly from 3° to 2.5°C near 1,500 m. These are relatively permanent features so significant changes occur only rarely, and then only as a result of large-scale changes in circulation. Ranges of physical properties of GOA waters are listed in Table 3.3-1.

Small horizontal changes in water properties occur as the flow proceeds westward, but mainly in the mixed layer. Nearshore salinities in the eastern and northern GOA can be as low as 26.0 psu in the ACC in the fall, when precipitation is at its maximum. Along the edge of the shelf in the Alaskan Stream a low-salinity (<32.0 psu) tongue like feature protrudes westward. In Shelikof Strait and to the east, the range of temperatures (0° to 15°C) can be substantially greater than those farther west. Whereas surface salinity increases toward the west as sources of fresh water from the land diminish, salinity values at 1,500 m decrease very slightly. Temperatures at all depths tend to decrease toward the west.

Some chemical properties of GOA water make it unique in the world ocean. Compared to other ocean waters at similar latitudes, the deep water of the GOA has higher concentrations of silicate, phosphate, and nitrate and its well-developed oxygen minimum. The oxygen and phosphate distributions result from the decomposition of particulate organic matter sinking from the surface, as elsewhere, but the higher concentrations arise because of accumulation resulting from poor circulation of the deep water. Reeburgh and Kipphut (1986) examined GOA chemical profiles for dissolved oxygen, silicate, phosphate, and nitrate, and summarized available historical data in three distinct oceanographic domains: 1) the deep sea, 2) the continental shelf, and 3) fjords and estuaries. Of the three, the shelf domain has the least data.

Deep sea profiles show temperature decreasing continuously with depth, first in the main thermocline from 10°C at the surface to 6°C at 100 m, then gradually to 4°C at 350 m and even more slowly to 1.8°C at 2,500 m (Reeburgh and Kipphut 1986). Dissolved oxygen decreases from about 300 micromoles (μM) oxygen/kilogram (kg) at the surface to less than 50 μM oxygen/kg at 400 m, followed by a minimum near 900 m then a gradual rise to about 120 μM oxygen/kg at 4,000 m. Phosphate increases from 0.5 μM phosphoric acid-phosphorus/kg at the surface to a maximum of almost 3 μM phosphoric acid-phosphorus/kg from 500 to 1,500 m, then decreasing slightly to about 2.6 μM phosphoric acid-phosphorus/kg near 2,500 m. Nitrate increases from about 0.3 μM nitrate-nitrogen/kg at the surface to a maximum of about 40 μM nitrate-nitrogen/kg from 500 to 1,500 m, then decreases only slightly to about 35 μM nitrate-nitrogen/kg near 2,500 m. Silicate increases from about 5 μM dissolved silica-silica/kg to 150 μM dissolved silica-silica/kg at 500 m, then continues to increase slightly to 175 μM dissolved silica-silica/kg at 2,500 m. The dissolved oxygen minimum and the phosphate and nitrate maxima occupy similar depth zones. Some studies have investigated long-term variability in the deep sea using Ocean Station P data. Surface nitrate was never less than 10 μM, even during peak uptake. Hokkaido University (1981) confirmed measurable nitrate was always present and probably does not limit surface productivity. A well-established population of pelagic grazers appears to be responsible for the relatively high surface-nutrient concentrations (Miller *et al.* 1984).

The nutrients in the shelf waters interact horizontally and thus have similar properties to the shallow (< 250 m) range of the oceanic water described above. Seasonal changes depend upon the seasonal variations in the meteorological regime (Royer 1975). In the winter, southeasterly winds bring convergence and downwelling (Royer 1981) along with the winter cooling and replacement of warm, high-saline bottom waters. In the summer, the wind field reverses, bringing relatively warm, high-saline, low-oxygen, high-nutrient waters from the central GOA back onto the shelf at depths of 100 to 200 m. Nitrate profiles from near the mouth of Resurrection Bay show values 20 to 40 µM between 0 and 250 m depth during winter, and summer values of 1 to 30 µM over the same depth range.

Few nutrient studies have been done in fjords and estuaries, but exchange with the shelf water has been determined from a few localized intensive studies to be a function of sill depth. No anoxic conditions were observed in Alaskan fjords, indicating at least annual bottom water renewal (Muench and Heggie 1978). Shallow-silled (< 50 m) fjords renewed between February and April when surface waters were most dense. Intermediate sill depth (120 to 160 m) fjords followed shelf water density changes and led to fairly continuous flushing. Deep or unrestricted sill (greater than [>] 180 m) fjords are flushed between July and October, when warm, saline, higher-nutrient water returns to the shelf after the relaxation of convergence.

3.3.3 Bering Sea and Aleutian Islands Fishery Management Unit

3.3.3.1 Description

The Bering Sea is a semi-enclosed, high-latitude, subarctic sea and is considered to be a northern extension of the North Pacific Ocean. Shaped somewhat like a sector of a circle with its apex at the Bering Strait, the Bering Sea has a total area of 2.3 million square kilometers (km^2). Forty-four percent is continental shelf (depth < 200 m), 13 percent is continental slope, and 43 percent is deepwater basin where depths reach as much as 3800 m along the western margin of the sea. The broad continental shelf on the east side of the Bering Sea is one of the most biologically productive areas of the world. The BSAI FMU comprises most of the continental shelf and consists of the entire EBS from the Alaskan coastline westward to the international boundary. Also, those waters within the EEZ south of the Aleutian Islands from Unimak Pass to the international boundary are included in the BSAI FMU (Figure 1.2-2).

3.3.3.2 Circulation

Numerous straits and passes through the 2,000-km arc-shaped Aleutian-Komandorski archipelago connect the Bering Sea to the North Pacific Ocean. The amount of water exchanged between the North Pacific Ocean and the Bering Sea through passes between the various Aleutian Islands is uncertain. Waters from the Alaska Current enter the Bering Sea at Unimak Pass and, to a lesser extent, through other passes between Aleutian Islands. Major exchanges of water occur at the west end of the Aleutian-Komandorski archipelago, with large inflow to the Bering Sea through Near Strait and outflow through Kamchatka Strait. Some additional leakage into the Bering Sea occurs through passes between the islands just east of Near Strait.

As the warm Alaska Stream water enters the Bering Sea and is cooled and transported through the anti-clockwise Bering Sea Gyre, large upwellings occur bringing cold deep waters to the surface (Ohtani 1970). Eddies, ranging in diameter from 10 to 200 km, can be found throughout the Bering Sea and contribute to the vertical mixing of waters. These eddies are thought to be caused by instabilities, wind forcing, strong flow through passes in the U.S., and topography (Schumacher and Stabeno 1998).

To the north, the Bering Sea is connected with the Chukchi Sea and Arctic Ocean through the Bering Strait which separates the Seward Peninsula (Alaska) from the Chukotka Peninsula (Russia). At the Bering Strait, there is a relatively small net annual outflow of water (Coachman and Aagaard 1988), although this flow can be reversed by relatively rare combinations of meteorological conditions (Coachman and Aagaard 1981).

Patterns of circulation in the Bering Sea have been inferred mostly from distributions of water properties, but some knowledge has also been obtained from drifter studies (Stabeno and Reed 1994). The overall circulation pattern is generally anti-clockwise within the basin, with the most prominent feature on the east side being a weak and variable northwestward flow over the broad continental shelf adjacent to Alaska. Along the edge of this shelf the Bering Slope Current transports water northwest at speeds of 10 to 20 cm/sec (Kinder *et al.* 1975, 1986), although Royer and Emery (1984) found this flow to be somewhat slower in winter. The Bering Slope Current intensifies as it approaches the Asian continent, bifurcating into a northerly flow through the Gulf of Anadyr and a southwesterly flow that is the origin of the Kamchatka Current. The Kamchatka Current is an intense western boundary current that continues southwestward along the Russian coast (Figure 3.3-2).

Flow over the North Aleutian Shelf (adjacent to Alaska) is characterized by Schumacher and Reed (1992) as weak and variable, with low current speeds (< 5 cm/sec). Mean speeds observed in the central shelf area are less than 1.0 cm/sec and reveal no organized circulation (Kinder and Schumacher 1981). Within Bristol Bay the *mean* flow is weak and shows an anti-clockwise tendency along the perimeter of the bay. Maximum speeds (~ 3.5 cm/sec) occur near the 50-m isobath and near the coast. However, the vast majority of the velocity variance within the bay is tidal, with tidal currents an order of magnitude larger than the mean flow. For example, on the north Aleutian shelf, where net currents are 1-5 cm/sec and the typical wind-driven currents are approximately 10 cm/sec at 5-m depth, the tidal currents are 40-80 cm/sec or more (Thorsteinson 1984). Turbulence resulting from these tidal currents causes mixing of the water column from the seabed to about 50 m above it.

3.3.3.3 Hydrography

Hydrographic structure over the U.S. is well-defined and consists of three domains that are separated by physical fronts (Kinder and Coachman 1978, Schumacher *et al.* 1979, Kinder and Schumacher 1981). The inner front is aligned approximately with the 50-m isobath, the middle with the 100-m isobath, and the outer at the shelf break (~ 200 m). The associated oceanographic domains are referred to as the coastal domain, middle domain, and outer domain. Two other distinct domains exist off the shelf: a narrow, energetic shelf break domain and the deep-ocean domain. The Pribilof Islands and the Unimak Islands provide distinct separate habitats within the Bering Sea. These domains will be referred to repeatedly in the following sections that describe the characteristics of the Bering Sea.

Circulation over the shelf is related to domain structure (Coachman 1986). In the outer domain (100 to 200 m), tidal currents account for about 80 percent of the flow, with a mean of about 5 cm/sec along shore to the northwest, and an onshore-offshore flow of 1 to 5 cm/sec that is quite variable. Tidal mixing is very important in the outer domain. In the middle domain (50 to 100 m) the important flow is due to tides and inertial currents. There is little net motion, and vertical mixing due to tides is important here. Tidal currents account for about 95 percent of the flow energy in the coastal domain (< 50 m), but as already noted, the mean flow has a speed of 1 to 5 cm/sec in a generally northwest direction. In contrast to circulation of the GOA, the circulation of the Bering Sea shelf has relatively small net flows and relatively large tidal forcing.

There are two main water masses on the shelf: Alaska Coastal water and central Bering water. Coastal water is found shoreward of the 50-m isobath in the south U.S., while central Bering water is found in the middle domain, from the inner front (~ 50 m) to the middle front (~ 100 m). Alaska Coastal water is a combination of coastal freshwater discharge and more saline water from the deep basin, and is generally well-mixed by winds and tides. The central Bering water in the middle domain has a lower layer that is isolated from seasonal heating and thus has temperatures that reflect prior winter conditions. Water of the outer domain

(100 to 200 m) is not really an identifiable water mass, but instead is a mixture of central shelf and deep Bering Sea water. Because of greater tidal and advective energy, it is less strongly stratified than the middle domain, but exhibits considerable small-scale vertical variation in properties that originate in the middle domain. These vertical variations, known as vertical "fine structure," are important to the flux of water properties horizontally and vertically through the water column.

Hattori and Goering (1986) summarized the available data on the distribution of salinity, temperature, phosphate-phosphorous, nitrate-nitrogen and ammonia-nitrogen, and silicic acid (Table 3.3-2) and characterized the four domains according to nutrients. Because the fronts inhibit lateral fluxes of water and dissolved materials between the four domains, nutrient zones are consistent with the physical domains. The vertical physical system also regulates the biological processes that lead to separate cycles of nutrient regeneration. The source of nutrients for the outer domain is the deep oceanic water, and for the middle domain, it is the shelf-bottom water. Starting in winter, surface waters across the shelf are high in nutrients. Spring surface heating stabilizes the water column, then the spring bloom commences and consumes the nutrients. Steep seasonal thermoclines over the deep Bering Sea at depth of 30 to 50 m, the outer domain at 20 to 50 m, and the middle domain at 10 to 50 m restrict vertical mixing of water between the upper and lower layers. Below these seasonal thermoclines nutrient concentrations in the outer domain are invariably higher than those in the deep Bering Sea water with the same salinity. Winter values for nitrate-nitrogen/phosphate-phosphorous ratios are similar to the summer ratios which suggests that, even in winter, the mixing of water between the middle and outer domains is substantially restricted (Hattori and Goering 1986).

Spring and summer storms can increase the total seasonal productivity by mixing to depths sufficient to resupply nutrients to the euphotic zone, but by the end of summer, nutrient depletion in the euphotic zone is common all across the shelf. Year-to-year consistency of trends between summer nutrient distributions in 1975 and 1978 was shown by Hattori (1979).

3.3.3.4 Effects of Sea Ice

Oceanic conditions, both physical and biological, can be profoundly influenced by the presence of sea ice. During extreme winter conditions, sea ice covers the entire eastern shelf of the Bering Sea; however, interannual variability of coverage can be as great as 40 percent (Niebauer 1988). The growth of ice over deep water is limited by relatively warm water in the central basin, so the maximum extent of the ice is restricted to the shelf.

The ice generally begins its seasonal southward formation in November. It is estimated that about 97 percent of the ice in the Bering Sea is formed within the Bering Sea itself (Leonov 1960). Very little ice is transported south through the Bering Strait (Tabata 1974). The ice apparently forms like a giant conveyor belt, being generated along the south-facing coasts in the Bering Sea and moving southward at as much as 0.5 meters per second before finally melting at its southern limit (Pease 1981). On average, seasonal ice formation progresses at an average rate of 12 to 13 percent per month over the area of the eastern shelf, reaching 60-65 percent coverage by late March (Niebauer 1981). The ice advance generally consists of a short, rapid advance (approximately 24 percent per month) in November-December, before slowing to approximately 6 to 7 percent in December-March. With the exception of the rapid advance in November and part of December, the ice appears to dissipate faster than it forms, at about 18-20 percent per month in late March to early July. Lisitsyn (1966) reported that, during the period of ice retreat, 63 percent of the ice melts within the Bering Sea basin. The remainder leaves the Bering Sea by way of various straits and passes.

The sea ice affects exchanges with the atmosphere and inhibits the transfer of freshwater (salt) and heat. It changes the coupling of the oceanic and atmospheric momentum exchanges by altering the surface roughness. The creation and melting of the sea ice alters the horizontal and vertical density gradients in the water column. Increases or decreases in the vertical density gradient affect the mixing and transport of nutrients and organisms in the euphotic zone. The ice edge also serves as both a source and sink of freshwater that can affect productivity. In fall during freeze-up, freshwater is extracted from the seawater, while during the spring, melting supplies freshwater to the ice edge.

One might reasonably assume that primary productivity in a winter ocean covered to a large extent by ice would be low and uncomplicated. However, McRoy and Goering (1974) reported on studies that revealed a complex productivity system in the water column and ice that makes a measurable contribution to the total annual production of the Bering Sea. The annual increase in production in the Bering Sea begins in late February with the development of the algal community in the sea ice. The production of this community increases with the passing of winter and probably reaches a maximum just before the ice melts completely. The ice algae comprise the first spring bloom that occurs in the Bering Sea, preceding the bloom that occurs in the open water farther south.

In April, as the ice melts, a second spring bloom develops in the wake of the receding ice. This begins along the southern ice front, coinciding approximately with the edge of the continental shelf. This bloom is promoted by the stability associated with the low-density water around the melting ice. As a result of the seasonal ice cover, the annual primary production of the Bering Sea is actually increased. Furthermore, the annual spring increase in algal standing stock begins in the middle and northern Bering Sea rather than the expected southern waters. Niebauer *et al.* (1990) subsequently estimated that the ice edge bloom of phytoplankton accounts for between ten percent and 65 percent of the total annual primary production.

Sea ice also influences bottom temperatures, and hence influences many species on the shelf. In winter, there is little stratification, and the sea is cold from top to bottom. In colder years there is more sea ice than in warmer years. The ice helps to cause and maintain density stratification when it melts. After the ice has melted, solar heating causes further stratification, and thus bottom temperature changes very slowly. Consequently, in cold years—years with extensive sea ice—the colder-than-normal bottom temperature is even more persistent than usual (Coachman 1986). Thus, the distribution and abundance of temperature-sensitive bottom-dwelling species and some nearshore species are related to the extent of sea ice. Variability (1972-1998) of sea ice arriving and departing the southern middle shelf is discussed by Stabeno *et al.* (2001).

3.3.4 Sources and Magnitude of Oceanic Variability

3.3.4.1 Atmosphere-Ocean Time Scales and Forcing Mechanisms

Atmospheric and oceanic parameters in the North Pacific and Bering Sea have variability that exists on several time scales and is due to many different forcing mechanisms (Table 3.3-3). Short-term (daily to annual) fluctuations in atmospheric and oceanic conditions are familiar and generally well-understood, to the extent that cause-and-effect relationships are well established. Fluctuations having longer (interannual) time scales are becoming better documented, due to extensive environmental monitoring activities, but definition of causal relationships for most remains an elusive challenge. The focus of this section is on atmosphere-ocean interactions that occur on time scales of several months to several years, or even decades. No attempt is made to catalogue all possible sources of variability. Rather, only the few that are well-known are identified and their possible influences are described.

3.3.4.2 Mesoscale Eddies

Eddies are rotating masses of water that are formed when an ocean current is deflected or pinched off by a topographic feature on the seabed or at the continental margin. Eddies can also form as a result of velocity shear on the fluid boundary between a relatively swift current and a much slower moving water mass. Rotating around generally vertical axes, mesoscale eddies have diameters of tens to hundreds of kilometers and, depending on their size, have rotational periods measured in days, or even weeks. Because they dissipate their energy only very slowly, these eddies can have lives measured in months to years, and their trajectories can be traced by the persistence of water properties in their cores. Movement of an eddy past a fixed current meter is evidenced by a cycle of flow acceleration, deceleration, and then acceleration back to the mean flow speed (or vice versa).

Mesoscale eddies are ubiquitous features of oceanic circulation and occur frequently on continental shelves and slopes. Kinder and Coachman (1977) described observations of an isolated eddy of high-salinity water nestled in the outer reaches of the Pribilof Canyon and partially in water depths greater than 1,000 m. The temperature-salinity characteristics of the eddy were those of the Bering Slope Current. The authors attribute its formation as evolving from a pinching off of a meander of this current in a manner similar to that which occurs when the Gulf Stream (Atlantic Ocean) forms warm eddies that travel northward along the U.S. east coast. Similar eddy events have been observed in the northern GOA and reported by Royer *et al.* (1979). They describe a persistent clockwise 100-km feature lying off the continental shelf and attribute its formation to instabilities of the Alaska Current.

The role of mesoscale eddies in the ocean and, more specifically, in the GOA and Bering Sea, is not determined. However, eddies could play an important role in controlling exchange of water between the North Pacific and Bering Sea (Okkonen 1993). Eddies have an important role in mixing water masses, so they might be providing microclimates that enhance or deter productivity. The interaction of eddies with other longer-term oceanic processes can serve to confound further comprehension of the overall circulation and its ecosystem-level effects. Accordingly, mesoscale eddies are essentially noise that is superimposed over the combined signal of longer-term quasi-periodic processes that are evident in the overall picture of oceanic variability.

3.3.4.3 Interannual Variability

The phenomenon known as El Niño – Southern Oscillation (ENSO), as described by Philander (1990), has long been recognized as a significant factor in the interannual variability of atmospheric-oceanic response. ENSO events radiate from the equatorial regions at irregular intervals, but ranging most commonly from three to seven years between events. ENSO events account for approximately one-third of the ice and sea surface temperature variability in the Bering Sea (Niebauer and Day 1989). ENSO forcing in the oceans at high latitudes is primarily through poleward propagation of Kelvin waves (Jacobs *et al.* 1994). This conclusion is supported by data of Enfield and Allen (1980) who found poleward-propagating, coastal-trapped disturbances along the west coast of North America that were correlated with equatorial disturbances. Royer (1994) reported that ocean temperature fluctuations at depth at GAK 1 (an oceanographic observation station near Seward) are well-correlated with ENSO events.

In addition to fluctuations associated with ENSO forcing, the water temperature variations at GAK 1 have been found to be associated with the lunar nodal tide component, which has a period of 18.6 years (Royer 1994). This tide component is the twelfth largest of all tidal components and is related to the 18.6-year periodicity of the lunar declination. Equilibrium tide theory predicts that this tidal component will vary with latitude, with amplitudes increasing with latitude (Parker *et al.* 1995). Because the interdecadal sea surface variability seems to occur simultaneously in the GOA and Bering Sea, it is expected that this component forces Bering Sea parameters in a similar fashion as in the GOA. Temperature anomaly patterns are similar with no phase shift, which suggests that the forcing is simultaneous.

3.3.4.4 Interdecadal Variability

A chronology of interdecadal climatic changes affecting the North Pacific Ocean was compiled from available measured atmospheric pressure data by Minobe (1997) for the period 1899-1997. A climatic regime shift was defined as a transition from one climatic state to another within a period substantially shorter than the lengths of the individual epochs of each of the (two) climatic states. Data used included the North Pacific index (NPI). The NPI is the area- and time-averaged sea level pressure anomalies in the region of 160°East (E) to 140°West (W) by 30° to 60°N for winter to spring (December to May), which illustrated rapid strength changes in the Aleutian low in the winter and spring seasons. Bidecadal pressure averages during 1899-1924 showed that the Aleutian low was about one millibar (mb) weaker than average, then strengthened to one mb below normal during 1925-1947. Similar behavior occurred in the later part of the Twentieth century as the Aleutian low shifted back to one mb above normal from 1948 to 1976, then strengthened back to one mb below normal during 1977-1997.

Using late-nineteenth century data for spring air temperature in western North America, Minobe (1997) identified 1890 to be the first regime shift. This extended the length of the first period to 34 years in comparison to the 22-, 26-, and 20+-year regimes to follow. The 50- to 70-year interdecadal variability, a two-regime cycle, has been prevalent from the nineteenth century to the present in North America. Minobe (1997) speculated that the likely cause of this variability is an internal oscillation in the coupled atmosphere-ocean system. This suggests that the next climatic regime shift is likely to occur between 2000 and 2007.

Long-term changes in fish populations around the North Pacific Ocean have apparently been influenced by climatic change of the same 50- to 70-year variability. Alaska salmon decreased in the 1940s and increased in the 1970s. Larger Japanese sardine catch amounts occurred in the regimes with the deepened Aleutian low. Baumgartner *et al.* (1992) found evidence of an approximately 60-year variability in sardine and northern anchovy populations in the eastern North Pacific from sediments in the Santa Barbara basin dating back to A.D. 270.

3.3.4.5 Regime Shifts

An update of evidence for regime shifts in the North Pacific Ocean in the 1920s, the 1940s, a major one in the winter of 1976/1977, and a minor one in 1988/1989 was presented recently at the North Pacific Marine Science Organization (PICES) symposium (Hare *et al.*, Hare and Mantua, McFarlane *et al.*, Zhang *et al.*, Park and Oh, Kang *et al.*, Suga *et al.*, Yasuda *et al.*, Savelieva *et al.*, Rogachev, Overland *et al.*, Miller and Schneider, and Minobe 2000). Coincidently, the beginnings of another large change in 1998/1999 were mentioned at the symposium; these are discussed in more recent papers by Minobe (2002), Conners *et al.* (2002), Mantua and Hare (2002), and Schwing *et al.* (2002).

In the late 1970s a step change in climate, referred to as a regime shift, occurred in the North Pacific Ocean. While there is evidence to suggest that there have been previous regime shifts, as noted in the previous section, it was the 1970s regime shift that stimulated extensive research on the topic, and especially how oceanic ecosystems were responding to these phenomena. Although more than a decade was required to recognize the pattern, the regime shift of 1976/1977 is now widely acknowledged, as well as its associated far-reaching consequences for the large marine ecosystems of the North Pacific Ocean. The 1989 regime shift has been studied extensively by Hare and Mantua (2000) who assembled and examined 100 environmental time series of indices (31 climatic and 69 biological) to obtain evidence of regime shift signals. A few examples of these illustrate that such signals are evident in the BSAI and GOA data.

Sea surface temperature anomalies, relative to long-term averages, around the Pribilof Islands indicate that the BSAI environmental regime appears to have shifted. The dominance of positive anomalies (warmer than average) from 1977 to 1988 switched abruptly to negative anomalies (colder than average) in 1989, which prevailed at least through 1997. Further evidence of a regime shift is seen in the time series of the southern extent of sea ice in the Bering Sea.

Niebauer (1998) reports that prior to the late 1970s ENSO regime shift below-normal sea ice cover in the Bering Sea was typically associated with ENSO conditions. These conditions caused the Aleutian Low atmospheric pressure center to move east of its average or normal position, with the result that warm Pacific air was directed over the Bering Sea. Conversely, above-normal sea ice cover was associated with La Niña conditions, during which the Aleutian Low moves west of its normal position, allowing higher pressure and colder weather in the Bering Sea. However, since the1970s regime shift, ENSO conditions are causing the Aleutian Low to move even farther east, causing winds to blow from the east and north off Alaska, and resulting in above-normal ice cover in the Bering Sea.

Before the regime shift, ENSO and La Niña conditions occurred with about the same frequency. Since the regime shift, ENSO conditions are about three times more prevalent. Both Mantua *et al.* (1997) and Minobe (1997) present evidence that this regime shift is the latest in a series of climate shifts that date back at least to the late 1800s and might be attributable to a 50- to 70-year oscillation in a North Pacific atmospheric-ocean coupled system.

Abundant evidence suggests that the coupled atmospheric-oceanic system of the North Pacific is subject to multiple forcing factors, each having characteristic behaviors and different frequencies of occurrence. The evidence also indicates that, rather than there being a single average or normal condition, the overall system appears to stabilize periodically around two or more normal states, changing from one to another abruptly in what has been termed a regime shift. These are the characteristics of systems whose dynamics are addressed by chaos theory, which is a body of mathematical theory that focuses on systems that have multiple states of equilibrium. Chaos theory attempts to define the mechanisms that cause the systems to change from one equilibrium state to another and to predict all such equilibrium conditions.

Using available sea level pressure and sea surface temperature data, along with coastal air temperature data from Sitka, Overland *et al.* (2000) formulated a conceptual chaotic model for the North Pacific. They were able to determine that the energy content of North Pacific time series of these parameters is broad-banded (i.e. over a broad frequency range) and temporally irregular (i.e. non-steady with respect to time). They reported that their conceptual model reflects the observed irregular behavior and suggests that the transitions from one equilibrium state to another are rapid rather than gradual.

Use of the word chaos in this context is not to imply the more common definition of great confusion or disorder. Rather, its use invokes the mathematical implication that there is order behind the irregularity of the system. A chaotic model may lead to a better understanding of the low-frequency relationship between the physical and biological systems in the North Pacific. One characteristic of a chaotic system is that , near the time of major interdecadal transition, there could be several years of extreme, and perhaps opposite, anomalies in the physical system. These extremes provide opportunities for change in the biological system. Recent experience with North Pacific fisheries may provide examples of such transition periods.

Although the Bering Sea is not discussed, a new review paper summarizes many details and the big picture of multidecadal (about 50 years) change in the Pacific Ocean (Chavez *et al.* 2003) characterized by about 25-year boom and 25-year bust cycles in the opposing anchovy-sardine populations. In the mid-1970s the change was from a cool anchovy regime to the warm sardine regime. Satellites have recently confirmed an increase in basin-wide sea-level slope after the 1997/1998 ENSO coincident with a dramatic increase in chlorophyl off California, indicating a shift back to a cool anchovy regime that occurred in the middle to late 1990s. The effects of ENSO in the tropics which radiate north on a shorter cycle of three to seven years and some unmeasured anthropogenic effects may tend to mask some of the synchronicity of changes in the physical and biological systems.