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Sea ice and water column structure on the eastern Bering Sea shelf

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ABSTRACT

Seasonal sea ice is a defining characteristic of the eastern Bering Sea shelf, and plays a critical role in determining the vertical structure of temperature and salinity over this shelf. Ice movement relative to local winds, ice composition, and the impact of both arrival and retreat of ice on the water column at four mooring sites over the middle shelf are examined. Ice forms primarily in coastal regions and is advected over the southern and outer shelves. Ice drift from satellite data for two representative years, 2003 and 2007, was ~2% of local NCEP wind speed and oriented 44° to the right of the winds ($r^2 = 0.25$). Measurements from 30 ice cores collected in 2006–2009 gave an average salinity of 5.62 ± 0.88, and an average nitrate concentration of 0.99 ± 0.83 μM. Time series data collected at the biophysical moorings in the Bering Sea (1995–2012) were used to explore the evolution of the water column under ice. At the northern mooring, M8, the water column had mixed and cooled to ~ -1 °C prior to the arrival of ice. Little melt occurred after ice arrival. At the other three moorings, the ocean temperature was 2–4 °C when ice arrived, resulting in extensive melt. Melting ice freshened and cooled the upper water column, resulting in stratification, which persisted for 10–25 days. Wind-induced water-column mixing occurred more slowly under the ice than in ice-free waters. An estimated 1.4 m of ice melted with the first arrival of ice at the three southern moorings where the latent heat of fusion accounted for approximately half the observed cooling. During ice retreat, there appeared to be little ice melt around the southern two moorings, but an estimated 0.8 m at M5 and M8. The extent of ice melt sets up the water column for the following summer.

1. Introduction

Situated between the North Pacific and the Arctic Ocean, the eastern Bering Sea shelf is ice-free in summer. Ice begins to form along the coast in the northern Bering Sea as early as November (Pease, 1980). During subsequent months, the sea ice is advected southward with maximum ice extent typically occurring in March or April (Stabeno et al., 2012b). In extensive ice years, sea ice can advance more than 1000 km, which, prior to the extensive melting now occurring in the Arctic Ocean, was the largest ice advance in any arctic or subarctic region (Niebauer, 1998). Sea ice typically begins to retreat in late winter or early spring, and the Bering Sea is again ice-free by late June.

Pease (1980) describes the expansion of ice into the southern Bering Sea middle shelf as a conveyor belt in which ice forms in polynyas in the northern Bering Sea, and is advected southward by winds. These polynyas are most common on the southward- and westward-facing coastlines due to prevailing northerly winds in winter, but can also form off northward- and westward-facing coastlines when winds are southerly (Niebauer and Schell, 1993; Niebauer et al., 1999). The St. Lawrence polynya occurs along the southern coast of St. Lawrence Island, and is a dominant ice-producing feature (Danielson et al., 2006; Drucker et al., 2003; Schumacher et al., 1983; Stringer and Groves, 1991), with ice advected primarily toward the south and southwest (McNutt, 1981). The leading edge of the ice melts as it encounters warmer ocean temperatures in the south or along the shelf break (McNutt, 1981; Zhang et al., 2010).

The eastern Bering Sea continental shelf is broad (>500 km), less than 180 m deep, and extends more than 1200 km south from Bering Strait to the Aleutian Islands. The shelf is divided into inner (water depth <50 m), middle (~50–100 m), and outer (~100–180 m) domains (Coachman, 1986), which are separated by transition zones or fronts (Kachel et al., 2002; Schumacher et al., 1979; Stabeno et al., 2001). Ice typically appears first in the coastal

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domain, and latest in the outer domain. On average, sea ice covers the middle shelf north of 58°N for five months each year, with ice persisting, on average, for less than one month on the shelf south of 57°N (Stabeno et al., 2012a).

North–south differences in timing of ice retreat and advance, together with a north (weaker)–south (stronger) gradient in tidal energy, result in distinct differences in the water-column structure on the northern and southern shelves during the summer (Stabeno et al., 2012a). The northern shelf, with a more extensive and persistent ice cover, remains cold until solar input warms the ocean, initiates ice melt, and sets up summer stratification. This results in a warmer, fresher surface layer overlaying a colder, saltier bottom layer separated by a ~10 m interface. The cold bottom layer (<2 °C) over the middle shelf, known as the cold pool, remains largely well mixed due to the tides and persists through the summer. In contrast, in the south, ice retreat typically occurs in March or April with southerly winds; these winds are usually still strong enough to mix the water column (Stabeno et al., 2010). The water column becomes thermally stratified when winds weaken and solar insolation increases (Stabeno et al., 2007). In years with extensive ice in March or April, the bottom temperatures in ice-covered areas in the south are <2 °C, and part of the cold pool (Stabeno et al., 2012b; Wyllie-Echeverria and Wooster, 1998). During years when the sea ice does not extend onto the southern shelf, the bottom temperatures remain above 2 °C, and the cold pool exists only over the northern shelf (Stabeno et al., 2012a, 2012b).

The presence or absence of sea ice impacts spatial and temporal variability of biological production across the shelf and slope (Sigler et al., 2014; Stabeno et al., 2012a, 2012b). Constituents of sea ice, including iron and phytoplankton, are advected southward, and influence the amount and fate of spring production (Aguilar-Islas et al., 2008; Schandelmeier and Alexander, 1981). In addition, sea ice impacts the timing of the spring phytoplankton bloom, with an under-ice bloom occurring if ice is present after mid-March, and an open-water bloom occurring in May or June if no ice is present after mid-March (Brown and Arrigo, 2013; Sigler et al., 2014; Stabeno et al., 2012a). Vertical stratification of the water column is important to support under-ice phytoplankton blooms (Alexander and Niebauer, 1981; Mundy et al., 2009). The southern extent of the cold pool and the timing of spring primary production influence the food chain and higher trophic level predator–prey relationships for the following 6–8 months (Sigler et al., 2014), and over longer time scales (Mueter and Litzow, 2008).

The purpose of this paper is to examine the impact of sea ice on the vertical structure of the water column. Measurements from satellites, ice cores and four long-term biophysical moorings distributed along the 70-m isobath from 56.9°N to 62°N are used to determine the rate and direction of ice movement, the arrival of ice at moorings, and the extent of ice melt and resulting stratification at each of the mooring sites (Fig. 1). First, calculations of the rate and direction of ice movement in the region using satellite images are shown. Next, the characteristics of 30 ice cores collected from 2006 to 2009 are presented. Using time series of temperature, salinity and currents from the biophysical moorings (M2, M4, M5 and M8; Fig. 1), the impact of sea ice on the water column structure is examined. Using selected time series from the ~42 yearly records, the thickness of the ice melted at the mooring sites is estimated.

2. Methods

2.1. Satellite data

Satellite measurements of ocean color and sea surface temperature (SST) were obtained using Moderate Resolution Imaging Spectroradiometer (MODIS) data files from the ocean color website at NASA (http://oceancolor.gsfc.nasa.gov). SeaDAS, a NASA image-analysis package, was used to map true-color and SST images. Individual ice floes were identified in a series of true-color images and tracked from one image to another until they were no longer recognizable (Holt et al., 1992; Leberl et al., 1983). Ocean color and SST observations cannot be made through clouds, severely limiting the number of MODIS images conducive to floe-tracking over a stormy Bering Sea. In addition, increased cloud cover is associated with southerly winds, so the set of tracked floes were biased against times when southerly winds were prevalent. This limited the analysis to years with a sufficient number of consecutive days clear enough to follow a floe, and to large floes in regions where horizontal shear was weak enough that floes remained intact.

For 2002–2011, ice concentration data from the Advanced Microwave Scanning Radiometer-Earth Observing System (AMSR_E) sensor aboard the MODIS Aqua satellite were downloaded from the National Snow and Ice Data Center website (http://nsidc.org/data/amsre/; Cavalieri et al., 2003). Ice concentrations from the NASA Special Sensor Microwave Imager (http://nsidc.org/data/docs/daac/ssmi_instrument.gd.html; Maslanik and Stroeve, 1999) were used for fall 1994–2002 to extend the time span prior to the start date of AMSR-E data. Daily-mean ice concentration values were used in this paper. The data were used to calculate mean daily ice concentrations within boxes centered on the four moorings (M2, M4, M5, and M8 in Fig. 1). The grid size of the ice data is approximately 12.5 km × 12.5 km. Following Stabeno et al. (2012a) we used 100 km × 100 km boxes.

2.2. Wind stress and winds

The National Center for Environmental Prediction – Department of Energy Reanalysis 2 (NCEP-2) uses a state-of-the-art analysis/forecast system to perform data assimilation with data ranging from January 1979 to August 2012 (http://www.cpc.ncep.noaa.gov/products/wesley/reanalysis2/; Kanamitsu et al., 2002), and is an update to the National Center for Atmospheric Research (NCEP/NCAR) reanalysis. Six-hourly wind and wind-stress data were extracted from the NCEP-2 at the grid points nearest to the four mooring location**s**. These data were obtained from the website maintained by NOAA Earth System Research Laboratory, Physical Sciences Division in Boulder, Colorado, USA (http://www.esrl.noaa.gov/psd/). NCEP-2 winds are well correlated with observed winds in the Bering Sea (Ladd and Bond, 2002).

QuikSCAT wind data were downloaded from the Jet Propulsion Laboratory, Physical Oceanography Distributed Active Archive Center (http://podaac.jpl.nasa.gov). Scatterometer winds are derived from ripple patterns on the surface of the ocean, and are therefore limited to ice-free areas. Although the instrument sees through clouds, measurements are affected by heavy rain. The records identify pixels where rain was suspected, and these records were excluded from our analysis. Grid spacing for QuikSCAT winds is 12.5 km (0.25° longitude), and the data were binned into 2° bins for regional plots.

2.3. Ice core measurements

During four springs (2006–2009), a total of 30 ice cores were collected from floes distributed over the eastern Bering Sea shelf by NOAA and University of Washington scientists (Fig. 1B). In 2006, sampling was conducted from the R/V *Thomas* G. *Thompson* (Cruise TN193). The *Thompson* is not an icebreaker and thus, sampling was restricted to the ice edge. Cruises in 2007–2009 were conducted on the icebreaker U.S.C.G.C. *Healy* as part of the