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Thermal regimes in the Chukchi Sea from 1941 to 2008

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ABSTRACT

The summer (June–October) temperature observations in the surface (0 m) and subsurface (20 m-bottom) Chukchi Sea layers collected from 1941 to 2008 have been analyzed using the self-consistent data recovery procedure based on correlation analysis and iterative empirical orthogonal function (EOF) decomposition. The analysis of the surface and subsurface EOFs identified "cold", "normal", and "warm" thermal states with variability of 2–3 years, and also 4–7 years. We found that the Chukchi Sea water temperature has gradually increased since 1941. Warming in the surface layer since 1941 has been minimal in the Bering Strait (0.012 °C yr^{-1}, total 0.8 °C) and maximal in Long Strait (0.030–0.036 °C yr^{-1}, total 2.0–2.4 °C). In the subsurface layer, temperatures have increased about half as much; minimal (0.0030–0.0075 °C yr^{-1}, total 0.2–0.5 °C) in Long Strait and rather uniform (0.010–0.015 °C yr^{-1}, total 0.7–1.0 °C) for the remaining Chukchi Sea. Analysis of the satellite sea-surface height anomaly data shows that during the "warm" periods there is a stronger flow through the Bering Strait and intensification of the northwestward currents in the central Chukchi Sea. Extensive correlation analysis shows that the thermal state of the Chukchi Sea agrees well with the flow of Pacific water through the Bering Strait and by an increase of the global atmospheric temperature. In addition, typical circulation during "warm" and "cold" periods was reconstructed using four-dimensional variational (4Dvar) data assimilation into the ocean model, and estimates of volume and heat fluxes in the Chukchi Sea during "cold" and "warm" periods were derived which are consistent with EOF and correlation analyses.

1. Introduction

The shallow Chukchi Sea (CS) (Fig. 1) is located between the Arctic and Pacific oceans. Because of this location, CS thermal conditions are defined by the large-scale atmospheric and oceanic processes, which occur in both oceans. The Chukchi Sea is covered by ice for most of the year. According to long-term observations, the concentration of ice in the Chukchi Sea begins to decrease in June, and reaches its minimum in August–September. At that point, ice begins increasing again, and by November–December, new ice completely covers the Chukchi Sea. There is also significant interannual variability in ice cover in the Chukchi Sea in summer. During the warmest years, about 80–85% of the Chukchi Sea can be free of ice, though ice can cover up to 90% of the CS during the coldest periods (Plotnikov and Pustoshnova, 2012).

Large-scale ice, ocean, and atmosphere variability in the Arctic Ocean have been analyzed in many publications (e.g. Bjorgo et al., 1997; Comiso, 2002; Parkinson et al., 1999; Rothrock et al., 1999; Stroeve et al., 2007; Tucker et al., 2001; Wadhams and Davis, 2000). It was also shown that interannual variability as well as atmosphere and ice cover in the Arctic Ocean exhibit both a linear trend and oscillation with a period of 10, 20, and 50–60 years (e.g., Raspopov et al., 2004; Frolov et al., 2007). Several mechanisms were discussed. In particular, Empirical Orthogonal Function (EOF) analysis of the atmospheric pressure showed that the first (the Arctic Oscillation, AO) and second (the Dipole Oscillation, DO) components are associated for the observed diminishing of ice cover and ice thickness (e.g. Proshutinsky and Johnson, 1997; Wang et al., 2009). Meanwhile, the observed changes in ice cover are also related to global warming (Thompson and Wallace, 1998) and the culmination of an ice/ocean – albedo positive feedback (Ikeda et al., 2003; Wang et al., 2005). These large-scale factors obviously affect CS as well, though intense flow through the Bering Strait also applies strong additional forcing to the CS state (e.g., Coachman et al., 1975; Woodgate et al., 2005a, 2005b).

The flow of Pacific water through the Bering Strait brings to the Arctic Ocean about one-third of the total freshwater inflow to the Arctic Ocean (Serreze et al., 2006) and can melt out 1–2 × 10^6 km^2 of the one meter thick ice there (Beszczynska-Möller et al., 2011; Woodgate et al., 2010). For this reason, the Bering Strait flow has significant impact on the thermochaline and biological condition

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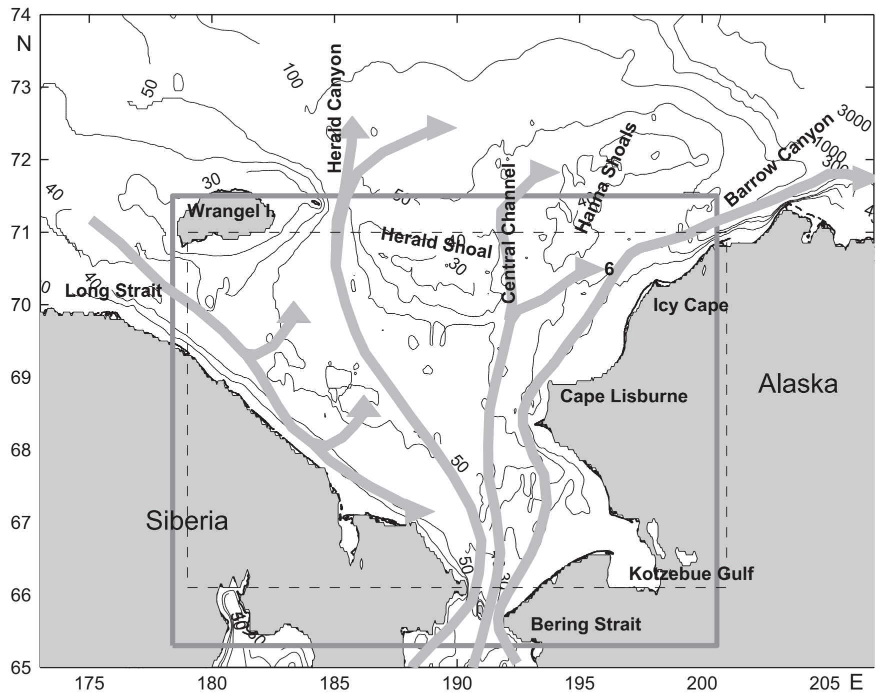

Fig. 1. Bottom topography in the CS. Thin dashed and solid thick lines designate the domains for EOF analysis and numerical modeling. Thick gray lines designate the schematic of local circulation.

in the entire Western Arctic Ocean (e.g., Killworth and Smith, 1984; Shimada et al., 2001; Steele et al., 2004; Walsh et al., 1989; Aagaard and Carmack, 1989) and plays an important role in the global freshwater cycle (e.g., Aagaard and Carmack, 1989; Wijffels et al., 1992; Woodgate and Aagaard, 2005), and possibly even the world climate (De Boer and Nof, 2004).

According to Aagaard et al. (2006), the observed warming and freshening of the upper halocline in the Pacific sector of the Arctic Ocean is the result of the penetration of Atlantic and Pacific water into the Arctic Ocean. Further, the salinity of Pacific water entering into the deep Arctic Ocean is defined by processes in the deep regions of the Bering Sea, Eastern Bering Sea shelf and Chukchi Sea.

Due to geographical proximity (close geographical locations), thermohaline structures in the Northern Bering and Chukchi Seas are formed by very similar processes. These include the persistence of northward flow in both regions, similarity in the seasonal cycle, and processes of ice formation in the winter polynyas located in the Gulf of Anadyr and along the Alaskan Coast between Cape Lisburne and Cape Barrow (Aagaard et al., 1981; Cavaleri and Martin, 1994; Gladyshev, Khen, 1999; Hu et al., 2011; Itoh et al., 2012; Schumacher et al., 1983).

It has also been shown that oceanic processes in the CS and Northern Bering Sea are regulated externally by the sea-level gradient between the Pacific and Atlantic oceans and by wind (e.g. Coachman and Aagaard, 1966; Shtokman, 1957). It is generally hypothesized that this gradient is of steric origin (Coachman et al., 1975; Stigebrandt, 1984) due to evaporation and precipitation rates that differ between the Pacific Ocean (precipitation prevails over evaporation and sea level is elevated) and the Atlantic Ocean (evaporation dominates precipitation and sea level is reduced). This sea-level gradient is believed to drive the mean northward flow of approximately $0.8–1.0 \times 10^6\ m^3\ s^{-1}$ from the Bering Sea towards the Arctic Ocean. The schematic of the Chukchi Sea circulation is shown in Fig. 1. The inflow of $\sim 0.9 \times 10^6\ m^3\ s^{-1}$ through the Bering Strait generates three branches through the Herald and Barrow Canyons, and the Central Channel. On average, there is also a weak ($\sim 0.1 \times 10^6\ m^3\ s^{-1}$) inflow through the Long Strait, though this flow can vary significantly—from $1 \times 10^6\ m^3\ s^{-1}$ inflow to $1 \times 10^6\ m^3\ s^{-1}$ outflow (Münchow et al., 1999; Panteleev et al., 2010).

Basic understanding of CS dynamics has improved since earlier studies in the 1950s, 1960s, and 1970s (Coachman et al., 1975; Gudkovich, 1961, 1962; Shtokman, 1957) due to both observational (Aagaard et al., 1985; Weingartner et al., 2005; Woodgate et al., 2005a) and modeling studies (Nihoul et al., 1993; Overland and Roach, 1987; Panteleev et al., 2010; Proshutinsky, 1986; Proshutinsky et al., 1995; Spall, 2007; Spaulding et al., 1987; Winsor and Chapman, 2004). In particular, it was shown that local wind significantly modifies the Bering Strait inflow at time scales from synoptic to inter-annual, sometimes blocking it completely or even reversing it (see Aagaard et al., 1985; Panteleev et al., 2010; Roach et al., 1995; Woodgate et al., 2005a). It was also shown that bottom topography is responsible for the three branches of Pacific water that transit the CS; the Alaskan Coastal Current and the Central and Herald outflows (Coachman and Aagaard, 1966; Coachman et al., 1975; Weingartner et al., 1998; Woodgate et al., 2005a). Synoptic and seasonal variability are regulated by the wind and sea ice conditions.

It is well known that warm Pacific water defines CS thermal conditions (Carmack, 1986; Coachman and Aagaard, 1974; Coachman et al., 1975; Codispoti, 1979; Fedorova and Yankina, 1963; Overland and Roach, 1987; Proshutinsky, 1986; Wilson and Wallace, 1990; Woodgate et al., 2005a, 2005b). Fedorova and Yankina (1963) estimated that Pacific water brings heat which, averaged annually, is enough to melt approximately 50% of the entire winter sea-ice area in the CS. Another example of this influence was given by Woodgate et al. (2006), who showed that the increase of Bering Strait heat input alone between 2001 and 2004 could have melted 640,000 km^2 of one-meter-thick ice.

During the 2001–2007, there has been a gradual increase of heat flow through the Bering Strait (Woodgate et al., 2006, 2010) and a corresponding decrease of the observed ice thickness (Steele et al., 2004; Woodgate et al., 2010). According to