A Year in the Physical Oceanography of the Chukchi Sea: Moored measurements from Autumn 1990-1991.

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Abstract

Year-long time-series of temperature, salinity and velocity from 12 locations throughout the Chukchi Sea from September 1990 to October 1991 document physical transformations and significant seasonal changes in the throughflow from the Pacific to the Arctic Ocean for one year.

In most of the Chukchi, the flow field responds rapidly to the local wind, with high spatial coherence over the basin scale - effectively the ocean takes on the lengthscales of the wind forcing. Although weekly transport variability is very large (ca. -2 to +3 Sv), the mean flow is northwards, opposed by the mean wind (which is southward), but presumably forced by a sea-level slope between the Pacific and the Arctic, which these data suggest may have significant variability on long (order a year) timescales. The high flow variability yields a significant range of residence times for waters in the Chukchi (i.e. 1-6 months for half the transit) with the larger values applicable in winter.

Temperature and salinity (TS) records show a strong annual cycle of freezing, salinisation, freshening and warming, with sizable interannual variability. The largest seasonal variability is seen in the east, where warm, fresh waters escape from the buoyant, coastally trapped Alaskan Coastal Current into the interior Chukchi. In the west, the seasonally present Siberian Coastal Current provides a source of cold, fresh waters and a flow field less linked to the local wind. Cold, dense polynya waters are observed near Cape Lisburne and occasional upwelling events bring lower Arctic Ocean halocline waters to the head of Barrow Canyon.

For about half the year, at least at depth, the entire Chukchi is condensed into a small region of TS-space at the freezing temperature, suggesting ventilation occurs to near-bottom, driven by cooling and brine rejection in autumn/winter and by storm-mixing all year.

In 1990-1991, the ca. 0.8 Sv annual mean inflow through Bering Strait exits the Chukchi in four outflows - via Long Strait, Herald Valley, the Central Channel and Barrow Canyon - each outflow being comparable (order 0.1-0.3 Sv) and showing significant changes in volume and water properties (and hence equilibrium depth in the Arctic Ocean) throughout the year. The clearest seasonal cycle in properties and flow is in Herald Valley, where the outflow is only weakly related to the local wind. In this one year, the outflows ventilate above and below (but not in) the Arctic halocline mode of 33.1 psu and volumetric comparison with Bering Strait indicates significant cooling during transit through the Chukchi, but remarkably little change in salinity, at least in the denser waters. This suggests that, with the exception of (in this year small) polynya events, the salinity cycle in the Chukchi can be considered as being set by the input through Bering Strait and thus, since density is dominated by salinity at these temperatures, Bering Strait salinities are a reasonable predictor of ventilation of the Arctic Ocean.

Keywords: polar oceanography, shelf dynamics, volume transports, seasonal variations, ocean-ice-atmosphere system, coastal currents.

Regional Terms: Arctic Ocean, Chukchi Sea, Bering Strait, Long Strait, Herald Valley, Barrow Canyon.

1 Introduction

At the north end of the Pacific Ocean, the narrow (ca. 80 km across) Bering Strait opens into the ca. 500 km wide (east-west) and ca. 800 km long (north-south) Chukchi Sea. The Chukchi Sea, some 50 m deep, is typical of the arctic marginal seas in that it is seasonally ice-covered and experiences both arctic night and midnight sun. By its location, it is supplied by a seasonally varying input from the Pacific and is influenced by the autumn and winter atmospheric storms carried north through Bering Strait on the storm tracks. In winter, polynyas form along its coasts; in summer, buoyant boundary currents of riverine origin have been found on its east and west margins. As a shallow sea, even its depths are within reach of the influence of the atmosphere. Being fed by nutrient-rich waters from the south, it is one of the most productive areas of the world ocean. It has rich pelagic and benthic ecosystems and is a migration pathway and feeding ground for birds and marine mammals that seasonally visit the Arctic.

In a global sense, Bering Strait acts as the Pacific gateway to the Arctic Ocean, and all Pacific waters found in the Arctic must cross the Chukchi to reach the Arctic Ocean shelf-break at ca. 73°N. Once in the Arctic, the Pacific waters play three major roles: they bolster the halocline layers of the Arctic Ocean water column, they provide nutrients for Arctic ecosystems and they are an important part of the freshwater balance of the Arctic Ocean and finally the World Ocean. Each role depends critically on the volume and properties of the waters exiting the Chukchi into the Arctic.

Observations in Chukchi Sea are complicated by sea-ice, which is present for about 6 months of the year. In 1990-1991, for example, freeze-up started in the north near Wrangel Island in mid-October 1990. Although the central Chukchi remained comparatively free of ice through November, by December 1990 the entire Chukchi Sea was ice-covered. Melt-back in the Chukchi started in mid-June 1991 in Bering Strait and by mid-July, the ice edge was at ca. 70°N. However, the 1991 autumn ice-edge was only just north of Wrangel Island, some 300 km south of the 1990 extent.

Until the mid-1970s, it was widely believed that the main northward flow in the Chukchi Sea was a coastally trapped current which hugged the Alaskan coast and continued on into the Beaufort Sea (Sverdrup, 1929; LaFond and Pritchard, 1952; Paquette and Bourke, 1974). However, it is only later work that acknowledges this coastal current is just one part of the northward flow and Aagaard (1964), following earlier suggestions by Saur et al. (1954) and Fleming (1959), describes a broader northward flow (distinct from the coastally trapped current) that bifurcates west of Point Hope. In their comprehensive synthesis, Coachman et al. (1975) describe a basin-scale flow pattern, consonant with later results from a moored current meter array across the central Chukchi Sea made by Coachman and Aagaard (1981). As evident from the seasonal melt-back of ice (Paquette and Bourke, 1981; Ahlnäs and Garrison, 1984), and increasingly from direct measurements, this broad northward flow appears topographically steered into three main branches - one east of Hanna Shoal which feeds into Barrow Canyon and includes (but is not limited to) much of the coastally trapped current; one west of Herald Shoal feeding into Herald Valley; and one between the two shoals, referred to as the Central Channel flow (Weingartner et al., 1998, submitted) - with Taylor column circulations restricting flow over the shoals (Martin and Drucker, 1997). This separation is also evident in water mass properties. In Bering Strait, Coachman et al. (1975) differentiate Alaskan Coastal waters (eastern Bering Strait), Bering Shelf waters (central Bering Strait) and Anadyr waters (western Bering Strait), using salinity boundaries which vary from year to year. Within the Chukchi Sea, the Anadyr waters and the Bering Shelf waters merge into a water mass named Bering Sea Water. An important distinction here is the nutrient content of the waters - those with Anadyr origin (generally found in the western Chukchi) are far richer in nutrients than the Alaskan Coastal Water found in the eastern Chukchi (Walsh et al., 1989).

Extremely few direct measurements of flow in Herald Valley exist, although the outflow is traced by water mass properties within the Arctic, cf. the core of high nutrient water found against the Chukchi slope by the AOS-94 work (Swift et al., 1997) and others. Whilst Sverdrup (1929) inferred a northward flow through the valley from short-term current measurements, a view backed by other short-term measurements by Coachman et al. (1975), we present in this paper the first year-long measurements of flow through Herald Valley. The Central Channel flow, also sparsely measured, is discussed by Weingartner et al. (1998) and, more recently, by Weingartner et al. (submitted). In contrast, Barrow Canyon is extensively studied. The suggestion of Paquette and Bourke (1974) that Barrow Canyon drains dense winter waters from the Chukchi

Shelf is confirmed by observations (Garrison and Becker, 1976; Aagaard et al., 1985b; Weingartner et al., 1998), although in some years freezing in polynyas along the Alaska coast (Aagaard et al., 1981; Cavalieri and Martin, 1994) is insufficient to form very saline waters (Aagaard and Roach, 1990; Weingartner et al., submitted). Intermittent onshore flow, which also occurs, can upwell waters as deep as the Atlantic layer (Mountain et al., 1976; Garrison and Becker, 1976; Aagaard and Roach, 1990), and such upwelling may also occur in the Central Channel (Bourke and Paquette, 1976). Mixing during upwelling results in a net flow of heat and salt to the shelf (Garrison and Paquette, 1982; Aagaard and Roach, 1990; Münchow and Carmack, 1997). The upwelling, most prominent in autumn (Aagaard and Roach, 1990; Weingartner et al., 1998), is attributed variously to atmospheric pressure variations (Mountain et al., 1976), topographic interactions (Garrison and Becker, 1976; Signorini et al., 1997), shelf waves (Aagaard and Roach, 1990) and wind forcing (Münchow and Carmack, 1997). Indeed episodic wind-driven upwelling occurs along the northwest Alaska coast, displacing the easternmost branch of the Chukchi circulation seaward and weakening or reversing the along-coast flow (Johnson, 1989; Weingartner et al., 1998).

In addition to these three main pathways for outflow from the Chukchi (Barrow Canyon, the Central Channel and Herald Valley), exchanges between the Chukchi and the Arctic Ocean take place through Long Strait. Sverdrup (1929) suggested possible westward flow near the bottom, while Gorbunov (1957) and Coachman and Rankin (1968) found evidence for a variable wind-influenced flow. Codispoti and Richards (1968) and Walsh et al. (1989) find nutrient distributions on the East Siberian shelf to be of Pacific origin, suggesting northwestward flow through Long Strait. A counterflowing cold fresh current, the Siberian Coastal Current, is intermittently present in the south of Long Strait, the western Chukchi and even occasionally in Bering Strait (Weingartner et al., 1999).

The relative contributions of each of these four pathways has been uncertain. In the annual average, the Bering Strait inflow is ca. 0.8 Sv with higher flow in summer and lower flow in winter (Roach et al., 1995). A comparable northward transport across the central Chukchi Sea, with a similar annual cycle, has been described by Coachman and Aagaard (1981). The intermittent inflow of the Siberian Coastal Current into the Chukchi is estimated to be of order 0.1 Sv

(Weingartner et al., 1999), giving a total of ca. 1 Sv to distribute between the branches. Based on short-term measurements, Münchow and Carmack (1997) estimate the summer northward flow through Barrow Canyon at ca. 1 Sv, suggesting large variability to the flows and this, and the relative strengths of the branches, will be discussed both in this paper and in Weingartner et al. (submitted).

Whilst there is some knowledge of the inputs and outputs from the Chukchi, including the imprint of Pacific waters on the upper Arctic Ocean (e.g. Coachman and Barnes, 1961; Kinney et al., 1970), the extent to which the Pacific waters are modified on the Chukchi shelf and the details of their movement and mixing near the shelf break and over the slope are unknown, and these processes are a focus of the recent and on-going Shelf Basin Interaction program. Coachman and Barnes (1961) cited extensive mixing of the Pacific waters on the Chukchi shelf, and Coachman et al. (1975) described significant mixing in the central Chukchi (but not along the Alaska coast). The winter salinisation of the waters along the northwest Alaska coast is well documented, however, and these waters may enter the Arctic Ocean through Barrow Canyon (Aagaard et al., 1985b; Weingartner et al., 1998). Modification on the shelf of the waters that exit through the Central Valley and Herald Valley has not been documented. Furthermore, the seasonality of the outflows is largely unknown, although Cooper et al. (1997) argue that the properties of the Arctic Ocean halocline require that it be maintained by winter outflow from the shelf, while Walsh et al. (1997) predict very large seasonal variability in the biochemical properties of the outflow. Finally we note that while the mean motion of the exiting shelf waters over the slope to the north has been thought to be eastward (Aagaard, 1984; Johnson, 1989; Pickart, 2004), several authors (Sverdrup, 1929; Münchow et al., 2000; Shimada et al., 2001) find evidence for sustained westward flow over the outer shelf or upper slope.

This paper presents mooring data from the most comprehensive measurement program yet undertaken in the Chukchi Sea, a set of one-year mooring deployments covering both U.S. and Russian waters, with CTD cruises in the months of mooring deployment and recovery. Since the Chukchi Sea is split by the U.S.-Russian Exclusive Economic Zone, even in ice-free seasons, surveys of the entire sea are rare. The present study focuses on the time-series measurements, with emphasis on the physical processes in evidence in the year 1990-1991. Although it is snapshot of a system that has significant interannual variability (e.g. Coachman et al., 1975; Weingartner et al., 1998, submitted), this work illustrates the different physical processes that affect this region and, by quantifying their varying influences for this one year, allows an understanding of possible future configurations of the system. **Section** 2 describes the data and data limitations. **Section** 3 considers the strong spatial coherence of the flow system and its relationship to the wind field. **Section** 4 examines water mass transformations in the records and relates them to local and non-local processes. **Section** 5 quantifies the transports through the system, including the outflows to the Arctic Ocean and discusses the role of the Chukchi in the modification of the Pacific throughflow.

2 Mooring data - accuracies, errors and tides

A set of twelve moorings, each carrying an Aanderaa Recording Current Meter (RCM) and a Seabird SeaCat (SBE) approximately 10 m above the sea floor (**Table** 1) was deployed in three east-west lines in the Chukchi Sea from September 1990 to September 1991 (**Figure** 1). The southernmost moorings (MA1, 2 and 3) measured the Bering Strait region. Five moorings (MC1, 2, 3, 4, 6) stretched across the central Chukchi Sea, whilst the northernmost four moorings (ME2 in Long Strait, MF1 and MF2 in Herald Valley, and MK1 in Barrow Canyon) measured the outflows from the Chukchi Sea. The SBEs recorded temperature and conductivity data on the hour, with calibrated accuracy (from pre- and post-deployment calibrations) of 0.02°C, 0.0012 Sm^{-1} (equivalent to 0.02 psu). The RCMs also recorded hourly (an 1-hour average speed and a single measurement of direction) with accuracy of 1 cm/s in speed and 5° in direction.

Except for the RCM on MF1, which failed after ca. 4 months, and the SBE on MC1, which flooded, the instruments recorded a complete year of data. In this highly productive regime, rotor fouling is common and RCMs at MA1 and MA2 in Bering Strait record a progressive slowing of speeds for the last ca. 160 days of the time-series. Since the correlation of the hourly velocity records between MA3 and MA1 and MA2 is high (greater than 0.9 for the first 200 days), a linear regression with MA3 has been used to create two surrogate velocity time-series for the last 160 days at MA1 and MA2. These are used in **Section**s 3 and 5 where noted. Temperature and salinity (TS) records from the RCMs and SBEs (where both were available at a single mooring) agreed well. Thus, on MC1, where the SBE failed, the RCM TS records are used with confidence, although with an increase in error, i.e. to ± 0.05°C and 0.1 psu.

Since all instrumentation was in the bottom 20% of the water column (RCM at 9 m and SBE at 8 m above the bottom), when information on overlying water properties is required, it must be based on CTD data or extrapolated. The simplest extrapolation is to treat the water column as homogeneous. For velocity this is a reasonably good approximation. An ADCP moored near MA2 in 1990-1991 (Roach et al., 1995) shows comparatively small velocity shears over the lower water column (ca. 5 cm/s over the bottom 30 m) with larger shears in the upper layers (ca. 20 cm/s in the top 15 m), consistent with the synoptic velocity profiles of Coachman et al. (1975) from Bering Strait. (Note that the errors associated with the monthly means[1] of Roach et al. (1995)'s Figure 4 are ca. 20 cm/s.) Annual averages from the ADCP indicate that the velocity at 9 m above bottom is ca. 10% less than the depth-averaged velocity, although both agree to within errors, as do the monthly means, although for the latter the nominal shear has larger uncertainty. Ship's ADCP data from the strait region in summer/early autumn shows a similar vertical profile in velocity other than near the coast, where a surface intensified boundary current is seen (Woodgate, 2003; Woodgate and Aagaard, submitted). Further north in the Chukchi, similarly small vertical shears are observed in ship ADCP data (Weingartner et al., 1999, submitted). For water mass properties, the assumption of homogeneity may be expected to hold in autumn/winter, when surface cooling or brine rejection from ice formation will vertically mix the water column, or when storm events stir the water column to depth. CTD surveys (e.g. Coachman et al., 1975) indicate that, at least in summer, vertical variability is of the same order of magnitude as horizontal variability. Thus, where we have no further data to constrain the vertical stratification, we will extrapolate water properties to the surface by assuming homogeneity, remembering that the assumption will, certainly in summer, underrepresent the lighter waters.

A tidal analysis (**Table** 2) shows the dominant variability of the velocity records is not the tidal signal. Although the tidal currents are greatest on the western side of the Chukchi Sea and near Wrangel Island, maximum amplitudes are less than 5 cm/s, generally less than 2 cm/s.

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[1] Note the labels on their Figure 4a are one month too early, e.g. the extreme low flow month is February, not January.

Where stated, the velocity data have been filtered to remove the tidal and inertial signals by using a Lanczos (square-taper) low-pass filter with a cut-off period of 40 hours and resampling at 6 hour intervals.

3 What drives the flow in the Chukchi Sea?

3.1 General flow patterns

In the annual mean, throughout the Chukchi waters flow from the Pacific towards the Arctic Ocean with mean velocities of order 5 cm/s in the central Chukchi and 10-20 cm/s in Bering Strait, Herald Valley and Barrow Canyon, (**Table** 1 and the progressive vector diagrams of **Figure** 2). At each mooring the flow is dominantly aligned with isobaths, the mean being to the north (northwest, northeast, westnorthwest) although frequent flow reversals are common. This bi-directionality is most marked in Bering Strait, Herald Valley and Barrow Canyon. Elsewhere in the basin (e.g. the MC-line), the flow is less constrained (see the fraction of the variance explained by the first principal component of velocity, **Table** 1).

Whilst the annual mean velocities are small, the variability is high. The strongest flows are found in Bering Strait, where 6-hour mean filtered currents can reach 80-120 cm/s and individual 1-hour averages peak at 100-150 cm/s. By continuity (since the region opens to the north from the ca. 80 km wide Bering Strait to the ca. 400 km long section at the MC-line), one would expect (for the same transport) the velocities at the MC-line to be less and indeed they are, with speeds (6-hour filtered mean and 1-hour mean) rarely exceeding 40 cm/s and 60 cm/s respectively. The Herald Valley and Barrow Canyon moorings show higher flows, again as might be expected by continuity. In general, the flows are strongest in summer and autumn and weakest in winter and spring, this being particularly marked in Herald Valley, and to a lesser extent the eastern and central Chukchi, and almost if not entirely absent in Barrow Canyon. The transports of these outflows to the Arctic are discussed in detail in **Section** 5.

The flow reversals (see e.g. **Figure**s 7, 8, 9, 10, 15, 16) can last for a week or more and are generally correlated with the local wind field (see **Section** 3.2) and with flows at neighbouring moorings. For example, the velocities at the MA moorings are closely correlated with each other (0.95 for first 200 days, see **Section** 2), and with the velocities in the central Chukchi. Similarly, MC4 correlates well with its neighbours (MC3 at 0.73, MC6 at 0.69) and with the MA-line (0.63). (These significance levels are all greater than 0.99). An EOF analysis of the first principal component of velocity indicates two underlying spatial patterns (**Figure** 3). The first EOF (38% of the variance) shows the high correlation between Bering Strait and the central Chukchi Sea and a circulation pattern consistent with northward flow through the strait being coherent with flow in the central and northeastern Chukchi. The second EOF (29% of the variance) suggests northward flow in the eastern Chukchi is correlated with southward flow in the western Chukchi and vice versa, in particular, linking the flow variance at MC1 and MC2 with the Long Strait mooring ME2. Together, the two EOFs explain 67% of the variance, but note that the flow in Herald Canyon (MF2) is poorly correlated with either EOF (see **Section** 5.4) and that this analysis also masks the seasonality of the Siberian Coastal Current (**Section** 4.3.3).

Since the moorings are generally separated by between 40 and 60 km, these results show that the correlation length scales in the Chukchi Sea are much greater than the local baroclinic Rossby radius (order 5-10 km). **Section** 3.2 indicates that wind is the dominant forcing for the ocean system and we conclude that, in the Chukchi, the ocean thus adopts the much longer correlation length scales of the atmosphere.

3.2 Driving mechanisms - wind and pressure head

The flow through Bering Strait is generally attributed to two factors, viz.:
(a) a sea surface pressure gradient (assumed constant) order 10^{-6} between the Pacific and the Arctic oceans (Coachman and Aagaard, 1966; Stigebrandt, 1984), and
(b) varying atmospheric and wind effects, frequently represented as a linear function of the atmospheric pressure difference across the strait (Coachman and Aagaard, 1981) or the local wind (Aagaard et al., 1985a; Spaulding et al., 1987; Coachman and Aagaard, 1988; Roach et al., 1995).

With the exception of Aagaard et al. (1985a), sea-level set-up/set-down against the local topography is usually neglected due to paucity of data. An empirical linear fit to the local wind does, however, explain ca. 60% of the variance in the flow (Aagaard et al., 1985a; Roach et al., 1995), including flow reversals, and this and atmospheric pressure indices are often used to estimate the Bering Strait flow over several decades (Aagaard et al., 1985a; Coachman and Aagaard, 1988; Roach et al., 1995).

An important, frequently understated, feature of the system is that the annual mean winds in Bering Strait and throughout the Chukchi are significant (surface[2] mean winds being ca. 4 m/s in Bering Strait and 2-4 m/s in the central Chukchi) and towards the southwest (Bering Strait and the MC-line) or west (the northern line), i.e. opposing the Pacific-Arctic pressure head. Thus, model simulations which assume no wind (e.g. parts of Overland and Roach, 1987) firstly overestimate the flow through the Strait (by ca. 20%, see below) and secondly neglect a significant term in the vorticity balance of the flow. Another understated assumption is the presumed (but unproven) constancy of the Pacific-Arctic pressure difference.

Using the FNOC (Fleet Numerical Oceanography Center) 6-hour geostrophic winds for 1990-1991 (cf. Aagaard et al., 1985a), a standard conversion to surface winds (rotation counterclockwise by 27°and reduction by 0.64, Aagaard et al., 1991) yields a time-series with high coherence (r=0.85 and 0.50 for northward (v) and eastward (u) wind components) and very similar mean magnitude to an observational wind record from Prince of Wales (Pease, pers.comm.). A little better agreement (similar for v (r=0.86), but better for u (r=0.7)) is obtained using a slightly different (by ca. 10°) turning angle, and a more severe (by ca. 10%) reduction, reflecting limitations of the model and/or the conversion. Thus, we use surface winds (i.e. with reduction and rotation) but select the direction of the model wind which gives the best correlation with each current meter record[3]. **Table** 3 gives the results of correlations between the local model wind and the principal component of the water velocity at each mooring. A high correlation (significant at over the 99% level) is evident in Bering Strait (ca. 0.7 at MA1, MA2 and MA3), in the eastern central Chukchi (ca. 0.6 at MC4), in Barrow Canyon (0.6 at MK1) and Long Strait (0.6 at ME2). In each case the lag is the timestep of the wind data (6 hours) and thus not distinguishable from zero.

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[2] i.e. reduced by 0.64 from the geostrophic wind (Aagaard et al., 1991)

[3] All these correlations are significant at the 99% level.

Figure 4 shows the scatter plot of the principal component of the measured water velocity and model wind for Bering Strait (mooring MA3), and a comparison between the same observed velocity and a proxy velocity time-series reconstructed from the model wind. The reconstruction is very good, although a dominant discrepancy is the frequent significant underestimation of the strength of the southward flow. This property is seen at all the moorings and is reflected in the crescent structure of the scatter plots (e.g. **Figure** 4). The asymmetry in this response implies the importance of another mechanism or a non-linearity, but further analysis requires higher quality wind data.

Returning to the simplest approach, we consider the principal component of the water flow (v_{pc}) to be approximated by a linear function of a wind velocity (w_r) component (picked to give the best correlation), i.e.

$$v_{pc} = P_h + m \times w_r$$

with empirical constants P_h (related to pressure head) and m (related to the coupling of the current and the wind) determined from a least squares fit between the wind and the current data (**Table** 3). The offsets at zero wind forcing (P_h), which represent the flow that would be driven by the pressure head forcing alone, are larger in Bering Strait ($\sim$ 40 cm/s northward flow at MA1 and MA2 and $\sim$ 30 cm/s at MA3) and smaller in the central Chukchi ($\sim$ 5 cm/s northward flow at MC2 increasing to $\sim$ 10 cm/s at MC6), as expected from the geometry. Note that these zero-wind flows are larger than the actual annual mean flows (compare **Table** 1, $\sim$ 30 cm/s at MA1 and MA2; $\sim$ 20 cm/s at MA3; $\sim$ 5 cm/s at the MC line) indicating that, in the annual mean, the wind significantly slows the northward flow through Bering Strait. The winds, which are generally stronger in winter, oppose the pressure head driven flow, resulting in the weaker water flow observed in winter. Indeed, a southward wind greater than 10 m/s (ca. 20 knots) can reverse the flow through Bering Strait. The slope of the fit (m) also shows a factor of four reduction between Bering Strait and the MC line, though this reflects at least in part that the wind model misses the real-world intensification of winds in the Strait due to funneling effects.

By considering best-fits to 30-day segments of data[4], we can consider the variability in m

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[4] Despite the shorter time-span, the correlations are still significant at the 95% level up to day 520.

and in P_h over the year (**Figure** 5). Note first the lack of seasonal cycle in these coefficients, suggesting that the seasonal presence of ice does not significantly change the driving, a result consistent with one year of *in situ* ice-drift measurements from a moored upward-looking ADCP in Bering Strait. These indicate the ice generally to be moving with the water at this location. Secondly, see that values for m (effectively the coupling response of water to wind) fluctuate within standard Ekman values (0.02-0.05, e.g. Pond and Pickard, 1983) with errors being the same order. Thirdly, and most interestingly, note that in contrast, there does appear to be a statistically significant trend in P_h, implying a slow increase in the pressure head driving of the flow through Bering Strait. Whilst this conclusion is at the limit of this data set (and is based on a very simple assumption), it is the first indication that the pressure head forcing may not be constant but could have significant variability on timescales of order a year.

Whilst a statistical analysis cannot determine the physical mechanism - note for example that both the wind velocity and the wind stress correlate equally well with the flow - this straightforward analysis does yield some interesting conclusions.

- As discussed by many authors, the flow through Bering Strait can be considered to be driven by a pressure head forcing and a local wind forcing.
- This relationship also holds for much of the Chukchi. Thus, the ocean effectively takes on the natural length scales of the wind, explaining the large correlation between neighbouring moorings and the large-scale patterns of the flow EOFs. (We will use this coherence in **Section** 5 to justify the calculations of water mass transport through various sections.)
- Whilst the wind forcing contains much short timescale variability, there are indications of longer (multiple month) variability in the pressure head forcing.
- Whilst local wind is a good predictor for water velocity in many areas, this is not true in the western Chukchi (MC1 - **Section**s 4.3.3) and in Herald Valley (MF1, MF2 - **Section** 5.4) for reasons we do not fully understand.

3.3 Residence times for the Chukchi Sea

Since the flow variability (both spatially and seasonally and weekly) is so high, residence times for the Chukchi vary significantly depending on starting time and place. The annual mean velocity for the central Chukchi (5 cm/s, **Table** 1), yields a ca. 4 month total transit time from Bering Strait to the MC-line and on to Herald Valley or Barrow Canyon, each leg $\sim$ 300 km. A water parcel starting, however, at MC6 at year day 300 (end of October) encounters over a month of northward flow and (if the velocities at MC6 can be extrapolated northward as is implied by **Section** 3) would cover 300 km in ca. 1 month. A month later, the flow is more variable and the same transit would take ca. 6 months. Stated another way, after 60 days the chance of a water parcel having gone 300 km from the MC-line is 30-50% (column 3, **Table** 4).

Figure 6 compares this variability at the different mooring sites. A common feature is the much longer transit times, e.g. MC6 (above) or MC3 (compare 90 days with 40 days), for parcels starting in the winter months (e.g. November through March), consistent with the mean flow being weaker and more variable in winter (e.g. Roach et al., 1995). The especially long transit times at MC1 reflect the seasonal presence of the Siberian Coastal Current (see **Section** 4.3.3). This large seasonal variability will have impacts on the shelf ecosystem (which has significant seasonality in primary and secondary production, Walsh et al., 1997), on the cycling of carbon and nutrients on the shelf, and also on the export of waters and nutrients into the Arctic.

4 What determines temperature and salinity in the Chukchi Sea?

4.1 How temperature and salinity change over a year

To zeroth order, the Chukchi cools in autumn, salinises in winter, warms and freshens in spring/summer. Details of and deviations from these transitions lead to a greater understanding of the physical processes affecting water mass properties in the Chukchi.

Figures 7, 8, 9 and 10 show time-series of temperature, salinity and ice cover (obtained from SSM/I 25 km resolution daily ice images from NSIDC) from Bering Strait, the central Chukchi, Herald Valley and Barrow Canyon. The seasonal cooling in autumn is associated with a gradual initial freshening of ca. 1 psu. CTD profiles show this to be consistent with storm events and surface cooling gradually mixing the overlying lower salinity waters down to the depth of the moored instruments, suggesting that the water column is well mixed in autumn. As ice starts to form above the moorings, a salinisation is seen in the records and this increase in salinity continues well into the winter, yielding an overall increase in salinity of ca. 2 psu. In 50 m of water, this change would require local ice formation of ca. 4-5 m, significantly thicker than the mean ice thickness of ca. 1.5 m measured by a moored, upward-looking ADCP in Bering Strait in 1990-1991[5]. This implies an advective source of salt, an issue to be addressed in **Section** 4.2. In late winter, while the ice coverage is still ca. 100%, in most of the Chukchi a freshening of 1-2 psu is observed in the salinity records and in general this trend continues until all the ice is melted. Again this is a greater salinity change than expected from just the melting of *in situ* ice. The moorings nearest to the coasts and some of those on the northern mooring line do not show the same trends and we will return to these below (**Section** 4.3), particularly, the western central Chukchi (MC1), where the freshening occurs as the ice starts to break up (**Section** 4.3.3), and the eastern Chukchi near the coast (MC6), which records sudden increases in salinity (**Section** 4.3.1). Note that both in western Herald Valley (MF1) and Long Strait (ME2) no significant freshening occurs after the winter.

The temperature records show an even clearer seasonal evolution. Temperatures start to fall before ice is present, and reach the surface freezing temperature before ice is at 100% coverage. For the first part of the winter (while the *in situ* salinity is increasing) at all moorings except Long Strait (ME2), the waters are always slightly supercooled, i.e. 0.01-0.02°C[6] below the freezing temperature (except for rare isolated events). As the salinity starts to drop in the second part of the winter, the water temperature tracks the corresponding freezing temperature very well. Generally the temperature rises above the freezing point as the ice concentration falls from 100% and the waters continue to warm after all the ice is gone. The exceptions are at the northernmost moorings, Long Strait (ME2) and Barrow Canyon (MK1), where the temperature never rises significantly above the freezing point between winter and the end of the record in September 1991, even though the ice concentrations start to fall ca. 4 months earlier. The western side of Herald Valley (MF1) is similar, exhibiting only a short rise in temperature almost at the end of

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[5] Within this record, there are only 15 days of ice thickness greater than 4.5 m, and this will include ridged ice as well as flat ice.

[6] This difference is at the limit of the accuracy of the calibration. All moorings are consistently colder than the surface freezing temperature, suggesting either a systematic error in the calibration (not yet found to be the case) or a real effect. Note that the adiabatic change moving surface water at these temperatures and salinities to 50 m depth is a warming of < 0.001°C.

the record, whilst the eastern side of Herald Valley (MF2), although part of the northernmost line of moorings, does exhibit a warming trend as per the central Chukchi Sea.

These seasonal changes in TS result in an anticlockwise loop structure on a TS-plot for each mooring (**Figure** 11), which demonstrates the ubiquity of this cycle through the central Chukchi. As the record is one-year long, non-closing loops (e.g. Long Strait, ME2) are an indication of interannual change. This visualisation emphasizes comparative variability across the Chukchi - the summer temperatures increase towards the east, where the warmest and freshest waters are found near the Chukchi coast, MC6. This increased temperature span is found also in the east in Bering Strait and in Barrow Canyon and is indicative of the influence of the fresh, warm waters of the Alaskan Coastal Current (**Section** 4.3.2). The coldest, saltiest waters are also found at the eastern moorings (**Section** 4.3.1). The cycle in the extreme western Chukchi (MC1) is enhanced by exceptionally fresh cold waters typical of the seasonal Siberian Coastal Current (**Section** 4.3.3).

Figure 12 replots the same data, this time by month, emphasizing the Bering Strait data. The wide range of TS-variability that exists in the Chukchi in the summer is reduced to a very narrow region of TS-space for almost 6 months of the year, during which time the temperatures are effectively at freezing and the salinity and density are increasing. If these changes are surface forced, this implies that for almost 6 months of the year, the Chukchi Sea is effectively mixed from top to bottom. We reiterate here, that these measurements are from $\sim$ 40 m depth and so in spring and summer, the later parts of the record, we expect surface waters to be less dense, although mixing, including storm events, will act to homogenize the water column. In autumn and winter, in addition to storm events, vertical mixing will also be initiated by cooling and salinisation at the surface and the rapid response of these records to individual events in the ice-cover record indicate that the 40 m TS does indeed significantly feel surface forcings.

Generally, from January through April, during the freeze, the Bering Strait waters are the saltiest and densest in the Chukchi. An exception to this occurs in February, where the eastern Chukchi (MC6) has the saltiest, densest water, and this ‘salinity nose’ is related to coastal polynyas in **Section** 4.3.1. The other exception is the upward curving TS-trace, with temperatures of ca. -0.5°C at salinities of ca. 34 psu. These occur at Barrow Canyon (MK1) and indicate lower halocline water and the shallow portions of the Atlantic layer upwelling in the Canyon (**Section** 5.4 and e.g. Mountain et al., 1976; Garrison and Becker, 1976; Aagaard and Roach, 1990). At other times in the year, the TS at the Bering Strait moorings is in the middle of the range of TS-properties within the Chukchi. Both in Bering Strait and in the Central Chukchi, the warmer, fresher (and therefore less dense) waters are generally found on the eastern side, with most variability being seen at MC6 (**Figure** 11). A comparison of October 1990 and September 1991 indicates some of the interannual variability to be expected in the region. This can be signficiantly greater than implied by these data, as discussed by Weingartner et al. (1998) and Weingartner et al. (submitted).

Thus we learn that:

- large seasonal cycles in TS exist in the Chukchi, with the most variability being seen in the east, and these seasonal changes occur on the same timescales as the transit time for water through the Chukchi;
- for ca. 6 months of the year, the water properties of the Chukchi (at least at depth) are condensed into a small region of TS-space and this suggests that the deeper waters are connected to the atmosphere by convection driven by storm mixing, surface cooling or brine rejection;
- whilst atmospheric driving and vertical mixing play important roles in the Chukchi, solely local effects cannot drive the seasonal cycles observed (the salinisation is too great to be purely local and the warming too spatially varying);
- the Chukchi is also strongly influenced by advection of waters from all directions, both significant large-scale currents (i.e. the Bering Strait inflow, **Section** 4.2) and coastal currents (i.e. the Alaskan Coastal Current, **Section** 4.3.2, and the Siberian Coastal Current, **Section** 4.3.3), and episodic events (i.e. coastal polynyas, **Section** 4.3.1, and inflow from the Arctic, **Section** 5.4). In the next sections, we will address these issues.

4.2 Dominance of Bering Strait on the Chukchi Sea salinities

Use of monthly smoothed time-series of T, S and ice concentration (**Figure** 13) allows us to separate longer term trends from the shorter timescale variability. The onset of ice formation (left column of plots) occurs first, unsurprisingly, at the northernmost line and later (by ca.

20 days and 40 days respectively) at the MC-line and in Bering Strait. The melt-back shows the same trend (south melting first) although with slightly longer delays (ca. 20 days and 60 days) and these changes are reflected closely in the water temperature records (middle column of plots). Apart from the seasonal asymmetry (i.e. 40 days versus 60 days) this is consistent with a purely local seasonal forcing. (Bering Strait has daylight, 1 - 24 hours, all year whilst the MC-line, e.g. MC3, and the northernmost line, e.g. MF2, experience 1 and 2 months of Arctic night respectively.) Were local effects, in particular ice formation, dominant also for salinisation (i.e. consider an ocean with no meridional velocity), we would expect the same nesting of curves of salinity, i.e. the largest change in salinity would be seen at the northernmost sites, where the ice season is longer, and salinity would start to increase first in the north and later in the south (see **Figure** 14).

The right column of plots in **Figure** 13 show this not to be the case. The maximum Bering Strait salinities are as high as those found in the rest of the Chukchi. Also, rather than the salinity curves nesting as per **Figure** 14, the curves overlap. The salinity increase starts first in the south in Bering Strait at ca. day 300 (late October 1990), with the peak salinity occurring at ca. day 430 to 460 (March 1991). At the MC-line (with the exception of MC6, discussed below), salinities start to increase later than in Bering Strait (ca. day 320, end of November), with the salinity peak occuring some 40 days after that in Bering Strait at ca. day 480 to 500 (early May 1991), by which time the Bering Strait salinity is already decreasing. Rather than matching the nesting of **Figure** 14, these curves strongly imply advection of a seasonal signal from Bering Strait to the MC-line (see bottom right plot in **Figure** 13). At the northernmost line the situation is more confused. One mooring (the eastern Herald Valley mooring, MF2) exhibits the same seasonal cycle in salinity as the rest of the Chukchi and, at this mooring, the salinity peak occurs at ca. day 520 (early June), i.e. lagging the Bering Strait line by ca. 60 to 120 days. The Barrow Canyon mooring (MK1), whilst showing more variability, also has its maximum salinity at about the same time[7]. These timescales are consistent with the advective timescales calculated in **Section** 3.3 and are strongly suggestive that, for salinity, it is the input through Bering Strait that determines the general properties of the Chukchi. This will be

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[7] Note that the salinity peaks at MK1 related to Atlantic water can be separated out on grounds of temperature. This removes the peaks around day 400 (early February) and after day 550 (early July). In contrast, both peaks at day 450 and 520 (late March and early June) are associated with near-freezing temperatures.

quantified with transport estimates in **Section** 5. Advection could also explain the asymmetry in freeze-in and melt-back, although other factors could equally well give this result, for example, spatially varying seasonal forcing or a consideration of ice thickness (e.g. in the simple model of **Figure** 14, the northern ice would be thicker and thus take longer to melt). In reality, the ice edge is, however, held to be determined by the advection of warmer waters northward (e.g. Paquette and Bourke, 1981; Martin and Drucker, 1997), supporting the dominance of advection in this system.

We will quantify these conclusions (i.e. that although significant cooling may occur in the Chukchi, the seasonal timescale salinity changes seen in the Chukchi are, at least in this year, dominated by inflow through Bering Strait) in **Section** 5.

4.3 Freshening and salinisation from coastal currents and polynyas

Whilst our data indicate little locally-driven net change in the Chukchi salinities in this year, coastal polynyas and coastal currents are known to be sources and sinks of heat and salt and, though not dominant, these signals can been seen in the mooring data.

4.3.1 Salinisation from the Lisburne coastal polynya - MC6

At the eastern end of the central Chukchi line, MC6 exhibits a substantial peak in salinity ca. 2 months before the maximum salinity at the other moorings on the MC-line. The 1-hour data shows a rise of almost 2 psu over 2 days (**Figure** 15, bottom plot) at day 413 (mid-February), a reduction back to almost the original salinity and then a second burst of high salinity water. The water is supercooled relative to the surface freezing point at this salinity (**Figure** 15, third plot), suggesting ice formation and brine rejection as the source. The winds and currents at this time (see **Figure** 15, top plots) are strongly southwestward, consistent with offshore advection of ice from north of Cape Lisburne, a region known for midwinter polynya formation (Cavalieri and Martin, 1994; Winsor and Björk, 2000). Indeed, satellite SSM/I ice data (Martin, pers.comm.) predict salinity pulses from the polynya region between Cape Lisburne and Point Barrow at days 411-417 (mid-February) and three pulses between days 417 and 432 (end of February to early March), in remarkably good agreement with the mooring data. Summing current meter velocities over this period suggests a lateral extent of this water of ca. 40 km (NS) by ca. 30 km (EW). (Note the distance from MC6 to the coast is ca. 35 km and from MC6 to MC4, where salinisation is not observed, is ca. 110 km). Taking this as a volume of shelf water formation (estimating the layer thickness at ca. 10 m as per observations of Aagaard et al. (1985b)) and converting into an annual flux, yields a contribution of 0.0004 Sv from this one patch of water alone. This is only 0.06% of the Bering Strait throughflow, but about 4% of the estimate made by Winsor and Björk (2000) of salty ($>$ 34 psu) polynya-formed waters in the Chukchi and Bering seas. This water is equal in density to the core of Atlantic water in the Arctic Ocean and thus, though entrainment will likely prevent it reaching these depths, has the potential to ventilate at least the lower halocline. Other anomalous salinity peaks in the data might also be related to coastal effects. In this year, however, the volume of dense polynya waters observed exiting the Chukchi is not large. For reference, winter 1990-1991 was not an extremum (either high or low) of the 39 winters examined for polynya formation by Winsor and Björk (2000).

4.3.2 Freshening and warming from the Alaskan Coastal Current - MC6

In summer, both the warmest and the freshest waters in the Chukchi are found on the eastern side of the Chukchi. The extreme values can be related to mixing of waters found in the central Chukchi with the warm, fresh waters of the Alaskan Coastal Current (Paquette and Bourke, 1974), a buoyant boundary current of riverine origin, ca. 10-20 km wide, surface trapped with intensified currents, on the eastern side of the Chukchi in the summer months. Its northward flow is confirmed by ADCP data (Woodgate, 2003; Woodgate and Aagaard, submitted) and local shipping information[8], and its transport can be estimated to be of order 0.05 Sv. It can be followed both by ship data (Woodgate, 2003) and by satellite data (Ahlnäs and Garrison, 1984) north along the coast and into the Arctic at Point Barrow. [Note here a confusion in nomenclature. Coachman et al. (1975) distinguish Bering Sea Water in the western Chukchi (formed from a mix of Anadyr waters in western Bering Strait and Bering Shelf Water in central Bering Strait) and Alaskan Coastal Waters (in the eastern Chukchi), on the basis of

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[8] A shallow ($\sim$ 10 m deep) bar (Prince of Wales Shoal) extends northwards for ca. 50 km just north of the Cape of Prince of Wales, and shipping travelling north often hugs this bar to ride the current north.

annually varying salinities, defining Alaskan Coastal Water (ACW) as fresher than 32.1 to 32.4 psu, depending on the year. The waters of the buoyancy-driven Alaskan Coastal Current are significantly fresher than this, e.g. 30 psu (Woodgate, 2003; Woodgate and Aagaard, submitted). Thus, whilst waters of the Alaskan Coastal Current meet the traditional definition of ACW, they are only a small fraction of the ACW and most of the ACW is not part of the buoyant Alaskan Coastal Current.] The Alaskan Coastal Current (ACC) is presumed to be be highly seasonal, both in properties and volume transport. Moorings MA2 and MC6 are sufficiently far from the coast (ca. 20 and 35 km respectively) to be generally outside this boundary current and hence track ACW, not the ACC. Similarly, MK1 is also measuring predominantly ACW, not the ACC. Thus the mooring data of this paper say very little about the ACC flow. Yet SeaWifs images of the Chukchi indicate that plumes can carry these ACC waters into the central Chukchi Sea.

4.3.3 Freshening and cooling from the Siberian Coastal Current - MC1

The western end of the central Chukchi line, MC1, contains the freshest water observed by these moorings at any time in the Chukchi excluding MC6. The correlation of the flow at MC1 with the local wind is extremely low (< 0.3, see **Table** 3). Its contribution to the first spatial EOF of the Chukchi flow field (**Section** 3.1, **Figure** 3) is negligible and, although it shows some small covariance with ME2, its input to the second EOF is again almost vanishingly small. The annual mean flow is only barely northward (ca. 1 cm/s) and the progressive vector diagram (**Figure** 2) shows major events of southeastward flow. Two such events occur before freeze up. During these, southeastward flow advects cold fresh water over the mooring (**Figure** 16, see e.g. day 290-297, mid-October). These excursions are not reflected in the wind field. In fact, southward winds, when they occur, appear to invoke a northwestward flow at the mooring (e.g. days 278 and 297), reminiscent of the mechanism of the 'poleward undercurrent' (e.g. Yoshida, 1980)[9].

Fresh, cold waters in this location are usually related to the Siberian Coastal Current, see e.g. Weingartner et al. (1999), who describe hydrographic and ship's ADCP measurements of this seasonally present current from 1992-1995. They find a fresh, cold, wind- and buoyancy-driven

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[9] In simplest form, taking the example of the Siberian coast, a northwestward wind invokes off-shore Ekman transport, driving upwelling at the coast. The resultant density structure drives a baroclinic current which intensifies southeastwards with depth.

current, with an unstable, eddy-rich, alongshore southeastward flow. Whilst the core of the current can be fresher than 27 psu, a smaller freshening effect can be seen up to 100 km offshore. Since the current rarely reaches Bering Strait, it must deflect/mix into the Chukchi north of the strait and Weingartner et al. (1999) describe mechanisms of jets (squirts) to accomplish this. The location of MC1 is ca. 45 km off the Russian coast. Thus it is highly likely the southeastward, cold, fresh flows experienced at MC1 have their origin in the seaward edge of the Siberian Coastal Current, either attached or detached from the main current. They illustrate the variability of the current, but can do little to quantify its properties.

5 Mass balances of the Chukchi Sea

5.1 Summary of the main flows of the Chukchi Sea

The mean flow through the Bering Strait and the Chukchi Sea is northwards, driven by a pressure head from the Pacific and opposed by the local winds. Flow occurs primarily aligned with local isobaths, and variability of flow in the Bering Strait and within the Chukchi is strongly related to the wind (and possibly to variability in the Pacific-Arctic pressure head). The narrow Bering Strait spatially constrains the flow, thus the highest velocities in the Chukchi are found in the strait. Since the forcing is large scale (atmospheric or ocean-basin scale), the ocean responds generally on large scales and the flow through much of the Chukchi appears to be coherent, certainly as far north as the MC-line, where the bottom topography allows the Chukchi to be considered as a broad (ca. 400 km), comparatively flat channel. This coherence extends further northeast to Barrow Canyon.

North of the MC-line and at the northernmost line, the bottom topography is no longer simple. Two major topographic rises (Herald Shoal and Hanna Shoal, see **Figure** 1) are associated with trapped Taylor column circulations (Martin and Drucker, 1997), the effect of which is to separate the northward flow through the Chukchi into three streams, which are seen in the seasonal melt back of ice (Paquette and Bourke, 1981; Martin and Drucker, 1997). The flow east of both shoals feeds into the head of Barrow Canyon at mooring MK1. The sharp topography in Herald Valley and Barrow Canyon also has a steering effect on the currents. The flow between the shoals, through a valley named the Central Channel by Weingartner et al. (submitted), is unmeasured by the moorings described in the present paper. The flow west of both shoals is split by topography around Wrangel Island. Some of this flow exits the shelf via the eastern side of Herald Valley (MF2). Aligned with the axis of the valley is a sharp front in water mass properties (Weingartner et al., submitted) between Chukchi waters in the east and Arctic waters in the west, suggesting that outflow of Pacific waters from the Chukchi is limited to the eastern side of the valley, whilst the western side is dominated by Arctic waters. South of Wrangel Island (Long Strait), the north side of the channel indicates a consistent westnorthwestward flow (i.e. out of the Chukchi), although previous authors suggest the net flow through Long Strait to be zero, with the outflow being balanced by the occasionally-present Siberian Coastal Current, estimated to be of order 0.1 Sv by Weingartner et al. (1999). (Note that although the mean volume flow may be zero, the outflow and inflow differ significantly in properties, so there may well be net fluxes of heat, salt and nutrients through Long Strait.)

In addition to the large scale flows presented in the preceding paragraphs, there are also the coastal currents - the Alaskan Coastal Current (**Section** 4.3.2) in the east and the Siberian Coastal Current (**Section** 4.3.3) in the west, both of which are present at least occasionally at the central Chukchi line and in Bering Strait.

The large scale features of the Chukchi flow are reflected in modelling efforts. The early models of the region (e.g. Overland and Roach, 1987; Spaulding et al., 1987) both favour the eastern branches in the Chukchi Sea. More recent studies, e.g. the barotropic model of Winsor and Chapman (2004), and the 23-year mean of a multilevel General Circulation Model (Maslowski, pers.comm.) give more dominance to the branch through Herald Valley. Both the latter models suggest that Barrow Canyon drains a significant part of the Chukchi Sea, including some of the waters from the Central Channel and from Herald Valley. However, Winsor and Chapman (2004) analyse these pathways under different wind forcings and for westward wind[10] (the mean wind in the Chukchi Sea in 1990-1991) the streamlines of their model indicate the Herald Valley outflow continues northwards into the Arctic Ocean, rather than turning east.

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[10] Winsor and Chapman (2004) use a steady wind of ca. 8 m/s. For the timespan of our data set, the annual mean FNOC wind at MF2 is ~ 2 m/s. Over the year, some 20% of the 6-hourly FNOC model winds exceed 8 m/s westward.

5.2 Summing the transports - methods and errors

Using this general circulation scheme, we construct transport estimates for three lines in the Chukchi, namely Bering Strait (the MA moorings), the central Chukchi (the MC moorings) and the northern edge of the Chukchi (moorings ME2, MFs and MK), by defining suitable sections and multiplying the across-section velocity at each mooring on the section with an appropriate cross-section area (see below for discussion, **Figure 1** for sections and **Table** 5 for estimates and errors). By picking boundaries midway between moorings, this is equivalent to assuming barotropic flow and linear interpolation between moorings. This approach, whilst crude, is the most complex justifiable given the available data. For each section, we examine the validity of interpolating the currents both in the vertical and in the horizontal, to obtain an estimate of the errors. The large coherence between neighbouring moorings suggests that interpolation between moorings is reasonable at the MA- and MC-lines. At the northernmost line, things are obviously more complex and instead of a continuous east-west section, we consider only the separate outflows, as outlined below.

An ADCP record from Bering Strait 1990-1991 (Roach et al., 1995) indicates that although significant shears may exist in the upper water column (e.g. order 10 cm/s in the upper 15 m), the 40 m velocity record is a fair representation (good to ca. 10%) of the depth-mean velocity, especially in the long-term mean since the monthly results show the surface flow sometimes weaker and sometimes stronger than the flow at depth. (To maintain consistency throughout the lines, despite having this shear information, we chose not to use it for these transport calculations.) Ship's ADCP data from eastern Bering Strait (Woodgate, 2003) also suggest that other than in the Alaskan Coastal Current discussed above (**Section** 4.3.2), an assumption of uniform flow is reasonably good (again, ca. 10%). (Neglecting both the northward flowing Alaskan Coastal Current and the occasionally present southward flowing Siberian Coastal Current is considered to add order 10% error to the flow estimates.) The combination of MA1 and MA2 (corrected for rotor fouling, **Section** 2) agrees well with the transport estimate obtained using MA3 and the full cross-section area of Bering Strait.

The MC-line constructed through the MC moorings is close to perpendicular to the principal flow directions at each site. Near the MC-line, ship's ADCP data from summer (Woodgate, 2003) suggest that away from the coast, the vertical shear is comparable to that in the strait and a barotropic assumption is good to ca. 10%. The waters from the boundary currents (the Siberian Coastal Current in the west near MC1 and the Alaskan Coastal Current in the east near MC6) are seen at least occasionally at the moorings and thus, to some extent, are included in the transport estimates. (Again we consider that omitting these currents adds ca. 10% to the errors in the transport.) Horizontal interpolation is justified by the high spatial correlations between moorings (**Section** 3.1).

The northernmost section is more complex. Long Strait contains (at least occasionally) the southeastward flowing Siberian Coastal Current, of which there is little trace at ME2. Lacking better information, we estimate the transport using half the cross-section area of Long Strait. In Herald Valley, based on CTD data we assume the division between northward and southward flow lies down the middle of the valley. For MF1, thus, the cross-section area is bounded by the middle of the valley in the east and Herald Island in the west. The transport from MF1 (only 3 months long) is almost zero, and neglecting it in the annual means is certainly within our error bars. For MF2, we define the boundaries as the middle of Herald Valley in the west and the top of rise just to the east of Herald Valley, reasoning that a change in sign of topographic slope marks a significantly different flow regime to that within the valley. In Barrow Canyon, we use as boundaries the deepest part of the canyon in the west (since CTD data suggest the outflow is concentrated over the southeast flank of the canyon) and the Alaskan coastline in the east. Note that the Barrow Canyon mooring MK1 is much further south than the measurements of Münchow and Carmack (1997) and also further south than some of the recirculations implied by the models of Winsor and Chapman (2004) and Maslowski (pers.comm.). Finally note that we are missing observations from the Central Channel (**Figure** 1) and we will return to this in **Section** 5.4.

5.3 Seasonal transport variability within the Chukchi Sea

The annual mean of the transports calculated by the methods outlined above are given in **Table** 5, with estimated errors for the section sums. The Bering Strait mean (0.8 Sv ± 0.2 Sv) agrees with the 4-year mean presented by Roach et al. (1995). The transport through the MC-line is also reassuringly 0.8 Sv ± 0.2 Sv, indicating that to within errors, mass is conserved by our calculations.

These means hide a large variability, however. The 6-hour mean transport in Bering Strait (not shown) whilst mostly less than 2 Sv northwards or southwards, can at times peak at almost 4 Sv in either direction. Whilst the 2 Sv flows can be maintained for over 10 days, the maximum flows occur for only a day or so at a time. There is no obvious phase difference between the Bering Strait flow and the flow across the MC-line. Differences are usually small (less than 0.5 Sv), but may be as large as 1 or 2 Sv maintained for a maximum of 2 days. A 1 Sv flow imbalance implies a 1 m sea-level change over the area between the two lines, though it is more probable this discrepancy reflects the inaccuracy of our transport estimates.

A strong seasonal cycle (**Figure** 17) is evident in the monthly running mean transports both in Bering Strait and in the central Chukchi, with the mean transport in winter not differing significantly from zero, whilst the autumn maximum lies at ca. 1.5 Sv (± 0.2 Sv). Thus, again, the data demonstrate the significant seasonal changes within the Chukchi. Seasonality at the northern line is less marked, but by continuity, must also be present, i.e. not only does the output to the Arctic vary seasonally in TS, but it also varies substantially in volume.

5.4 Four comparable outflows from the Chukchi Sea

Table 5 gives estimates of the individual outflows exiting from the Chukchi over the northern line and indicates that the outflows through Barrow Canyon, Herald Valley and Long Strait are all comparable ($\sim$ 0.1 to 0.3 Sv.) The sum ($\sim$ 0.6 Sv) is less than the total transport through Bering Strait or the MC-line. Although the error bars could account for the discrepancy, at least some of the missing transport is the Central Channel branch seen in the seasonal melt-back of ice (**Section** 5.1, Paquette and Bourke, 1981; Martin and Drucker, 1997; Weingartner et al., submitted). Our measurements suggest this fourth outflow is order 0.2 Sv, which agrees with the estimate of Weingartner et al. (submitted) who use all currently available data from the channel. Thus, we have four main exit routes for water from the Chukchi - Herald Valley, the Central Channel, Barrow Canyon and Long Strait - all with similar transports, but varying flow characteristics and water properties.

In contrast to the rest of the Chukchi, the flow on both sides of Herald Valley shows extremely little correlation with the local wind. The northward flow on the eastern side of the Valley is remarkably unidirectional (see **Figure** 9), a consistently northward outflow which is much weaker in the winter months (**Figure** 17). Whilst in TS-space, the waters resemble the central Chukchi (**Figure** 11), the flow field implies different dynamics to the wind-driven Chukchi Sea, presumably linked to the topography. A residual anticyclonic circulation resulting from the Taylor column dynamics believed to trap water over the neighbouring Herald Shoal (Martin and Drucker, 1997) could act to rectify the current in this way. Note also that Herald Valley is the one deep exit from the large area of the central Chukchi Sea. Examination of density differences between the MC-line and MF2 lead to no consistent pattern, and in any case, the large spatial separation and discrepancies in measurement depth make such a comparison possibly hopeless.

On the other hand, the flow at MK1 at the head of Barrow Canyon (**Figure** 10 and **Figure** 17) correlates well with the local wind, has many reversals and little seasonal variability. This outflow is the most studied of the Chukchi outflows and our observations echo previous conclusions of a wind-driven flow carrying Chukchi waters into the Arctic, with flow reversals (again wind-driven) being strong enough to bring lower halocline or upper Atlantic layer water significantly up-canyon. In October 1990 (ca. day 290), and January and February 1991 (days 370-410), ca. 5-day up-canyon flows with (6-hourly mean detided) speeds of up to 100 cm/s bring lower halocline waters usually found at ca. 200 m depth to mooring MK1, which is measuring at 71 m depth in 80 m of water, some 100 km from the shelf-break (**Figure** 12). These flows, though small, could in principle produce dense mixing products if combined with shelf-waters. In 1990-1991, however, the densest waters found are the freezing product in the eastern central Chukchi (at MC6), which we relate above (**Section** 4.3.1) to polynya formation.

The Long Strait flow is more variable than the Herald Valley flow, although still dominantly out from the Chukchi westnorthwestwards. Whilst it shows zero or small net inflow to the Chukchi at the time when the flow in the rest of the Chukchi is almost zero (i.e. winter), there is another minimum in the transport in the summer indicating a more complex flow response. The maximum outflow (westnorthwestwards) is in November 1990 and coincides with a maximum southward transport in Herald Valley, consistent with an anticyclonic circulation around Wrangel Island, but this circulation scheme is merely an hypothesis. Whilst inflow (eastsoutheastwards) to the Chukchi does occur, in TS properties the waters at ME2 never reflect the fresh, cold waters of the Siberian Coastal Current. Throughout the winter, significant steps in the salinity record may (as elsewhere) indicate brine rejection from the wind-initiated polynyas often seen in ice charts to form around Wrangel Island, e.g. http://www.natice.noaa.gov/westarct1.htm and Winsor and Björk (2000). Most notable in the record is that although waters around 0°C are found at ME2 in September 1990, the following September the water temperature does not rise above -1.7°C. It does depart from the freezing point in May 1991, but by the recovery of the moorings at the end of September 1991, the warmer waters of the Chukchi have not penetrated that far northwestward, consistent with the idea of Long Strait and western Herald Valley belonging more to the Arctic than to the Chukchi Sea.

5.5 Arctic Ventilation

Regardless of the details of the exit pathways, it is clear that the Chukchi must provide ca. 0.8 Sv of waters of varying density, salinity and temperature to the Arctic. **Figure**s 18 and 19 (top panels) show the volume contribution within different salinity and density ranges as estimated from the moored data. We stress again these observations are from the bottom layer and so represent the densest water fractions in the Chukchi. Very little water saltier than 34 psu is observed entering the Arctic in this year. Two peaks in the salinity distribution reflect the ca. 32.8 psu contribution from Herald Valley (a summer value of salinity reflecting the higher outflow in summer) and the more saline (ca. 33.3 psu) contribution from Barrow Canyon (generally a contribution from summer/autumn 1991). The minimum in the salinity distribution between these peaks lies close to 33.1 psu, the traditional value of Arctic upper halocline water (Aagaard and Carmack, 1989), suggesting firstly that inputs from the Chukchi fall both above and below this traditional halocline and secondly that (in this year) the Arctic 33.1 psu water is not renewed to any great extent by the Chukchi - indeed, the lack of 33.1 psu water is evident throughout the Chukchi (see lower panels on **Figure** 18).

The same calculations (not shown) for temperature, show that 80% of the water observed entering the Arctic in this year is colder than 0°C and, in fact, 50% is within 0.1°C of the freezing point. (Note here that since our measurements are 10 m above bottom, these results are skewed to the colder, denser fraction of the water column, and that when stratification is present, i.e. especially in spring and summer, excluding storm events, we will be underestimating the warmer, fresher components.) The result is a variation in input density (**Figure** 19) of 25.5 - 27.5 kg/m^3 (sigma-0), which coincides with equilibrium depths in the Arctic Ocean of ca. 50 - 150 m (e.g. AOS data, Swift et al., 1997).

Thus, the picture emerges of the outflows from the Chukchi as a set of streams, each varying in volume, temperature and salinity throughout the year, with resultant varying input depths to the Arctic Ocean. Despite the flow being generally enhanced in summer, the shortness of the warm period in the northern Chukchi means the majority of the water provided to the Arctic is extremely cold. The density (and thus the equilibrium depth in the Arctic) is dependent primarily on salinity, however, and thus selection of salinity (**Section** 5.6) is a critical factor in Arctic ventilation.

5.6 How much does the Chukchi Sea modify the Pacific inflow?

Thus we return to a fundamental question for the role of the Chukchi in ocean climate, i.e. how much are the waters from the Pacific modified during their multi-month transit through the Chukchi? Quite surprisingly, **Figure** 18 illustrates that in 1990-1991 the range of salinities at each mooring line in the Chukchi is very similar, remarkably so given the errors involved - compare the mean salinity (weighted by volume of water) at the MA-line and the MC-line (both $\sim$ 32.8 psu, with errors of order 0.1-0.2 psu). Whilst the same mean calculated for the northernmost line is $\sim$ 33 psu, this sum is missing ca. 0.3 Sv of unknown salinity and thus the mean salinity is subject to a larger error and does not differ significantly in these data from the MA- or MC-line[11]. This tallies with the conclusions of **Section** 4.2, i.e. that at least the deep waters propagate through the Chukchi with comparatively little net change in salinity due to

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[11] Note we do not include the transit-time lag in this calculation, preferring to consider the inventory for a whole year at each location, thus implicitly assuming (the defensible fact) that interannual variability is less than the seasonal variability. The alternative would be to estimate transit times (say 4 months) and sum the last 8 months at the northern line and the first 8 in the Bering Strait. In practice this introduces much larger errors, indeed to the extent that the total volume is no longer in balance, reflecting the errors in the transit time estimation.

local effects. This could be due either to little local ice formation, or to a balance of salinisation from ice formation with the freshwater inputs from the boundary currents. In temperature, however, the mean temperatures (again weighted by transport) for Bering Strait, the MC-line and the northern mooring line are -0.3°C, -0.6°C and -1.0°C respectively, errors order $\sim$ 0.1-0.2°C. Thus, the cooling between Bering Strait and the northernmost line is significant and consistent with loss of heat to the atmosphere.

Thus, at least for 1990-1991, we conclude with respect to temperature and salinity that:

- the seasonal variation of the inflow from the Pacific to the Chukchi is far larger than the modifications of the Pacific water on the Chukchi shelf;
- at least at depth, waters flowing through the Chukchi are significantly cooled during their transit of the shelf, but do not generally undergo dramatic salinisation, and
- since density is predominantly a function of salinity at these temperatures, the salinity of the inflow through Bering Strait is a good indicator of the salinity of waters found in the central Chukchi and furthermore a reasonable proxy for the outflow to and especially the ventilation of the Arctic Ocean.

6 Summary and Concluding Remarks

Time-series data from a set of 12 moorings measuring temperature, salinity and water velocity from autumn 1990-1991 allow us to describe a year in the physical oceanography of the Chukchi Sea, an arctic marginal sea significantly driven by both the atmosphere and the oceanic advection from the south. Although the data represent only lower-layer measurements and are only from one year of a system with much interannual variability, they allow significant insight into the physical processes important within the Chukchi Sea.

We find a flow field predominantly slaved to the local wind, which acts to impose atmospheric length scales on the ocean. Thus the circulations are generally on the scale of the Chukchi, with flows in Bering Strait correlating well with flows in the central Chukchi Sea and even Barrow Canyon. The Herald Valley outflow is largely uncorrelated with the local wind, possibly a result of trapped anticyclonic circulation around the adjacent Herald Shoal. In the western Chukchi, the seasonally present Siberian Coastal Current also has different, more baroclinic dynamics than the rest of the Chukchi. A similar buoyant coastal current, the warm, fresh Alaskan Coastal Current, is found on the eastern side of the Chukchi, although it is barely, if at all, described by these mooring data. (Note the traditional definition of Alaskan Coastal Water includes waters that are not in the Alaskan Coastal Current.)

The annual mean flow everywhere is to the north. It has long been suggested that the Pacific throughflow is driven by a sea-level difference between the Pacific and the Arctic, with flow variability mediated by the wind field. The data presented in this paper indicate that the mean wind opposes this pressure-head driving, significantly slowing the mean flow through the strait. They also imply that, whilst the coupling of wind to water shows no significant seasonal change, the pressure-head driving does vary on timescales of order a year. As expected by continuity, velocities are greatest in Bering Strait, and as befits a dominantly wind-driven system, the variability of the flow is much greater than the mean. The large variability gives rise to a wide range of residence times for waters in the Chukchi, ranging from a shortest half-transit time of 1 month to a longest half-transit time of 6 months. In general, residence times are greater in winter, when the southward winds are stronger and thus the northward flow is weaker.

Whilst the flow field is coherent over much of the Chukchi, the water properties are not. The western Chukchi, which is fed mostly by the Anadyr waters from south of the Bering Strait, is much richer in nutrients and is usually colder and saltier than the eastern Chukchi, which shows the influence of the Alaskan Coastal Current and the eastern Bering Sea. In temperature and salinity, large seasonal cycles are almost ubiquitous, with the largest ranges, especially the warmest and freshest waters, being found in the eastern Chukchi. The seasonal cycle of cooling, salinisation, freshening and warming shows the waters of the Chukchi to condense into a small region of TS-space at the freezing temperature for almost 6 months of the year. This suggests a ventilation of the entire water column of the Chukchi Sea from the surface by cooling and brine rejection for almost half of the year. Even in spring and summer, when the water properties are generally stratifying, storm events may mix to near bottom.

Whilst there is evidently a substantial loss of heat from waters transiting the Chukchi, it is somewhat surprising that the salinity output from the Chukchi is not significantly different from the salinity input measured through Bering Strait. The seasonal change of salinity through the Chukchi can, in this year at least, be explained as a dominantly advective effect, indicating that, since density is determined mostly by salinity at these temperatures, the Bering Strait salinities are a reasonable predictor of equilibrium depth in the Arctic. Coastal and shelf salinisation events - dense waters formed from polynyas near Cape Lisburn, upwelling of saline lower halocline waters through Barrow Canyon - compete with the low salinity coastal current inputs, however each are comparatively small terms compared to the order 0.8 Sv input through Bering Strait.

The fate of the Bering Strait inflow within the Chukchi is thus largely one of transit and cooling with possible summer warming. There is evidence of four branches of outflow from the Chukchi - the flow through Long Strait, possibly balanced in volume by the Siberian Coastal Current input, though there are undoubtedly net fluxes of salt, heat and nutrients; the Herald Valley outflow; the Barrow Canyon outflow and the inferred outflow through the Central Channel. All are of comparable magnitude, but differing properties, i.e. temperature, salinity and seasonal variation. Some correlate well with the local wind (e.g. Barrow Canyon and Long Strait) whilst others do not (e.g. Herald Valley). Certainly the Pacific flux into the Arctic Ocean varies seasonally in temperature, volume and equilibrium depth. Quantifying this suggests that at least in 1990-1991, the Arctic halocline mode of 33.1 psu was not renewed to any great extent by the Chukchi outflow.

The Chukchi Sea is one of the most productive of the world's ocean. By proxy, it is the link between the Pacific and the Arctic and provides nutrients, stratification and freshwater important to the Arctic ecosystems and circulation. This study provides a description and explanation of one year in the life of the Chukchi Sea, exhibiting the variety of competing physical processes that are found there. The outcome of these competitions results in an extremely rich interannual variability which is largely undocumented by these data, yet obviously is critical to understanding the role of the Chukchi in the world ocean.

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References

Aagaard, K., 1964. Features of the physical oceanography of the Chukchi Sea in the autumn. Master's thesis, University of Washington.

Aagaard, K., 1984. The Beaufort Undercurrent. In: Barnes, P., Reimnitz, E. (Eds.), The Alaskan Beaufort Sea: Ecosystems and Environment. Academic Press, New York, pp. 47–71.

Aagaard, K., Carmack, E. C., 1989. The role of sea ice and other fresh water in the Arctic circulation. Journal of Geophysical Research 94 (C10), 14,485–14,498.

Aagaard, K., Coachman, L. K., Carmack, E. C., 1981. On the halocline of the Arctic Ocean. Deep-Sea Research 28, 529–545.

Aagaard, K., Pease, C. H., Roach, A. T., Salo, S. A., 1991. Beaufort Sea mesoscale circulation study: Final report. Tech. Rep. ERL PMEL-90, NOAA Technical Memorandum.

Aagaard, K., Roach, A. T., 1990. Arctic ocean-shelf exchange: Measurements in Barrow Canyon. Journal of Geophysical Research 95, 18,163– 18,175.

Aagaard, K., Roach, A. T., Schmacher, J. D., 1985a. On the wind-driven variability of the flow through Bering Strait. Journal of Geophysical Research 90, 7213–7221.

Aagaard, K., Swift, J. H., Carmack, E. C., 1985b. Thermohaline circulation in the arctic mediterranean seas. Journal of Geophysical Research 90, 4,833–4,846.

Ahlnäs, K., Garrison, G. R., 1984. Satellite and oceanographic observations of the warm coastal current in the Chukchi Sea. Arctic 37, 244–254.

Bourke, R. H., Paquette, R. G., 1976. Atlantic water on the Chukchi shelf. Geophysical Research Letters 3, 629–632.

Cavalieri, D. J., Martin, S., 1994. The contribution of Alaskan, Siberian and Canadian coastal polynyas to the cold halocline of the Arctic Ocean. Journal of Geophysical Research 99, 18,343–18,362.

Coachman, L. K., Aagaard, K., 1966. On the water exchange through Bering Strait. Limnology and Oceanography 11(1), 44–59.

Coachman, L. K., Aagaard, K., 1981. Re-evaluation of water transports in the vicinity of Bering Strait. In: Hood, D. W., Calder, J. A. (Eds.), The Eastern Bering Sea Shelf: Oceanography and Resources, vol. 1. National Oceanic and Atmospheric Administration, Washington, D. C, pp. 95–110.

Coachman, L. K., Aagaard, K., 1988. Transports through Bering Strait: Annual and interannual variability. Journal of Geophysical Research 93 (C12), 15,535–15,539.

Coachman, L. K., Aagaard, K., Tripp, R. B., 1975. Bering Strait: The Regional Physical Oceanography. University of Washington Press, Seattle.

Coachman, L. K., Barnes, C. A., 1961. The contribution of Bering Sea water to the Arctic Ocean. Arctic 14, 146–161.

Coachman, L. K., Rankin, D. A., 1968. Currents in Long Strait, Arctic Ocean. Arctic 21, 27–38.

Codispoti, L. A., Richards, F. A., 1968. Micronutrient distributions in the East Siberian and Laptev seas during summer 1963. Arctic 21 (2), 67–83.

Cooper, L. W., Grebmeier, J., Whitledge, T., Weingartner, T., 1997. The nutrient, salinity and stable oxygen isotope composition of Bering and Chukchi Seas waters in and near the Bering Strait. Journal of Geophysical Research 102, 12,563–12,578.

Fleming, R. H., 1959. Oceanographic survey of the eastern Chukchi Sea, 1 August to 2 September 1959, Preliminary report of Brown Bear Cruise No. 236. Tech. Rep. 59-30, Department of Oceanography, University of Washington, Seattle.

Garrison, G. R., Becker, P., 1976. The Barrow Submarine Canyon: A drain for the Chukchi Sea. Journal of Geophysical Research 81, 4,445–4,453.

Garrison, G. R., Paquette, R. G., 1982. Warm water interactions in the Barrow Canyon in winter. Journal of Geophysical Research 87, 5,853–5,859.

Gorbunov, Y. A., 1957. On the water exchange between the East Siberian and Chukchi seas through Long Strait. Problemy Arktiki 1, 35–40.

Johnson, W. R., 1989. Current response to wind in the Chukchi Sea: A regional coastal upwelling event. Journal of Geophysical Research 94, 2,057–2,064.

Kinney, P., Arhelger, M. E., Burrell, D. C., 1970. Chemical characteristics of water masses in the Amerasian Basin of the Arctic Ocean. Journal of Geophysical Research 75, 4,097–4,104.

LaFond, E. C., Pritchard, D. W., 1952. Physical oceanographic investigations in the eastern Bering and Chukchi seas during the summer of 1947. Journal of Marine Research 11, 69–86.

Martin, S., Drucker, R., 1997. The effect of possible Taylor columns on the summer ice retreat in the Chukchi Sea. Journal of Geophysical Research 102, 10,473–10,482.

Mountain, D. G., Coachman, L. K., Aagaard, K., 1976. On the flow through Barrow Canyon. Journal of Physical Oceanography 6, 461–470.

Münchow, A., Carmack, E. C., 1997. Synoptic flow and density observations near an arctic shelf break. Journal of Physical Oceanography 27, 1,402–1,419.

Münchow, A., Carmack, E. C., Huntley, D. A., 2000. Synoptic density and velocity observations of slope waters in the Chukchi and East-Siberian seas. Journal of Geophysical Research 105, 14,103–14,119.

Overland, J. E., Roach, A. T., 1987. Northward flow in the Bering and Chukchi seas. Journal of Geophysical Research 92, 7,097–7,105.

Paquette, R. G., Bourke, R. H., 1974. Observations on the coastal current of arctic Alaska. Journal of Marine Research 32, 195–207.

Paquette, R. G., Bourke, R. H., 1981. Ocean circulation and fronts as related to ice melt-back in the Chukchi Sea. Journal of Geophysical Research 86, 4,215–4,230.

Pickart, R. S., 2004. Shelfbreak circulation in the alaskan Beaufort Sea: Mean structure and variability. Journal of Geophysical Research 109 (C4), 10.1029/2003JC001912.

Pond, S., Pickard, G. L., 1983. Introductory Dynamical Oceanography. Pergamon Press, Oxford, U.K.

Roach, A. T., Aagaard, K., Pease, C. H., Salo, S. A., Weingartner, T., Pavlov, V., Kulakov, M., 1995. Direct measurements of transport and water properties through Bering Strait. Journal of Geophysical Research 100, 18,443–18,457.

Saur, J. F. T., Tully, J. P., LaFond, E. C., 1954. Oceanographic cruise to the Bering and Chukchi seas, summer 1949, Part IV: Physical oceanographic studies: Vol. 1. Descriptive report. Tech. rep., U. S. Navy Electronics Laboratory Report 416, San Diego.

Shimada, K., Carmack, E. C., Hatakeyama, K., Takizawa, T., 2001. Varieties of shallow temperature maximum waters in the western Canadian Basin of the Arctic Ocean. Geophysical Research Letters 28, 3,441–3,444.

Signorini, S. R., Münchow, A., Haidvogel, D., 1997. Flow dynamics of a wide arctic canyon. Journal of Geophysical Research 102, 18,661–18,680.

Spaulding, M., Isaji, T., Mendelsohn, D., Turner, A. C., 1987. Numerical simulation of wind-driven flow through the Bering Strait. Journal of Physical Oceanography 17, 1,799–1,816.

Stigebrandt, A., 1984. The North Pacific: A global-scale estuary. Journal of Physical Oceanography 14, 464–470.

Sverdrup, H. U., 1929. The waters on the north Siberian shelf. Scientific Research of the Norwegian North Polar Expedition 4, 1–131.

Swift, J. H., Jones, E. P., Aagaard, K., Carmack, E. C., Hingston, M., MacDonald, R. W., McLaughlin, F. A., Perkin, R. G., 1997. Waters of the Markarov and Canada basins. Deep-Sea Research II 44, 1,503–1,529.

Walsh, J. J., Dieterle, D. A., Muller-Karger, F. E., Aagaard, K., Roach, A. T., Whitledge, T. E., Stockwell, D., 1997. CO2 cycling in the coastal ocean. II. Seasonal organic loading of the Arctic Ocean from source waters in the Bering Sea. Continental Shelf Research 17, 1–36.

Walsh, J. J., McRoy, C. P., Coachman, L. K., Goering, J. J., Nihoul, J. J., Whitledge, T. E., Blackburn, T. H., Springer, A. M., Tripp, R. D., Hansell, D. A., Djenidi, S., Deleersnijder, E., Henriksen, K., Lund, B. A., Andersen, P., Muller-Karger, F. E., Dean, K. K., 1989. Carbon and nitrogen cycling within the Bering/Chukchi seas: Source regions for organic matter effecting AOU demands of the Arctic Ocean. Progress in Oceanography 22, 277–359.

Weingartner, T. J., Aagaard, K., Woodgate, R. A., Danielson, S., Sasaki, Y., Cavalieri, D., submitted. Circulation of the north central Chukchi Sea shelf. Deep-Sea Research II.

Weingartner, T. J., Cavalieri, D. J., Aagaard, K., Sasaki, Y., 1998. Circulation, dense water formation and outflow on the northeast Chukchi Sea shelf. Journal of Geophysical Research 103, 7,647–7,662.

Weingartner, T. J., Danielson, S., Sasaki, Y., Pavlov, V., Kulakov, M., 1999. The Siberian Coastal Current: A wind and buoyancy-forced arctic coastal current. Journal of Geophysical Research 104, 29,697–29,713.

Winsor, P., Björk, G., 2000. Polynya activity in the Arctic Ocean from 1958 to 1997. Journal of Geophysical Research 105 (C4), 8,789–8,803.

Winsor, P., Chapman, D. C., 2004. Pathways of Pacific water across the Chukchi Sea: A numerical model study. Journal of Geophysical Research 109 (C3), 10.1029/2003JC001962.

Woodgate, R. A., 2003. Alpha Helix HX274 Cruise Report, Bering Strait Mooring Cruise June-July 2003, available at psc.apl.washington.edu/AlphaHelix2003.html.

Woodgate, R. A., Aagaard, K., submitted. Revising the Bering Strait freshwater flux into the Arctic Ocean. Geophysical Research Letters available at psc.apl.washington.edu/HLD/Bstrait/BSFWpaper.html.

Yoshida, K., 1980. The Coastal Undercurrent - a role of longshore scales in coastal upwelling dynamics. Progress in Oceanography 9, 83–131.

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