

Circulation of Winter Water on the Chukchi Shelf in Early Summer

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Abstract

Using a variety of data sources we investigate the properties and pathways of Pacific-origin winter water as it spreads across the eastern Chukchi shelf in early summer. The focus is on the time period June-July 2011 during which an extensive shipboard hydrographic/velocity survey was undertaken as part of the Impacts of Climate on Ecosystem and Chemistry of the Arctic Pacific Environment (ICESCAPE) program. A revised circulation scheme is constructed revealing that the transport pathways on the Chukchi shelf are more complex than previously thought. Notably, the well known branch progressing northward from the Central Channel bifurcates as it reaches Hanna Shoal, flowing around both sides of the shoal and dividing into smaller filaments that continue towards Barrow Canyon. Mass is conserved in the circulation scheme, with approximately 1 Sverdrup flowing poleward across the Chukchi shelf within these pathways, then exiting Barrow Canyon. The salinity of the winter water varied on the shelf in 2011, with saltier water found in the upstream portion of what is defined as the central pathway. Using sea ice concentration data and atmospheric reanalysis fields, we argue that salinization of the winter water in the central pathway occurred via brine rejection as the parcels progressed north and passed through the Cape Lisburne polynya. This demonstrates that winter water pervading the interior shelf can be transformed by convective overturning north of Bering Strait, presumably stirring up nutrients from the sediments and thereby influencing primary productivity in the region of Hanna Shoal.

1. Introduction

Pacific water enters the Arctic Ocean through Bering Strait and flows northward across the wide and shallow Chukchi Sea. The characteristics of the water vary strongly with season, from warm and fresh water in summer to relatively salty and very cold water in winter and early-spring (Woodgate et al., 2005). The winter water is formed because of ice growth, which densifies the surface layer and leads to convective overturning that reaches the bottom. As such, the water is near the freezing point, weakly stratified, and high in nutrients that were stirred into the water column from the sediments. These properties together exert a strong influence on the physical state and ecosystem of the western Arctic Ocean. The dense winter water ventilates the

31 upper halocline of the Canada Basin (Aagaard et al., 1981), its vorticity structure influences the
32 manner in which the water is fluxed off the Chukchi and Beaufort shelves (Pickart et al., 2005;
33 Spall et al., 2008), and the high nitrate concentration fuels primary productivity throughout the
34 region (e.g. Hill et al., 2005). Hence, it is of high importance to determine the precise pathways,
35 mixing, and residence time of the winter water as it traverses the Chukchi shelf, none of which
36 are presently well known.

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38 Based on mooring data obtained through the years, and various numerical modeling studies,
39 the basic circulation pattern of the Chukchi Sea has become more clearly defined. To first order
40 the flow is dictated by the bottom topography of the Chukchi shelf (Figure 1). North of Bering
41 Strait the Pacific water is steered into three main branches: the western branch flows through
42 Hope Valley into Herald Canyon (Winsor and Chapman, 2004), the middle branch flows through
43 the topographic depression between Herald Shoal and Hanna Shoal known as Central Channel
44 (Weingartner et al., 2005), and the eastern branch follows the coastline of Alaska towards Barrow
45 Canyon. In summer and early-fall the eastern branch is referred to as the Alaskan Coastal Current
46 (Paquette and Bork, 1974), advecting the warm and fresh coastal water originating from the
47 eastern Bering Sea and Gulf of Alaska. Recent evidence suggests that, at this time of year, most
48 of the inflowing transport through Bering Strait is carried by this current together with the Central
49 Channel branch (Gong and Pickart, 2015; Itoh et al., 2015). Averaged over the entire year,
50 however, it is argued that the western branch transports roughly 50% of the Pacific water (some
51 of which gets diverted through Long Strait into the East Siberian Sea), while the other two
52 branches transport roughly 25% each (Woodgate et al., 2005). Modeling studies indicate a similar
53 annually averaged partitioning of the flow (Winsor and Chapman, 2004; Spall, 2007).

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55 Winds influence the circulation on the Chukchi shelf significantly on a variety of timescales.
56 Woodgate et al. (2005) demonstrated that the dominant observed variability in velocity over most
57 of the shelf is due to changes in the large-scale wind field (with the exception of Herald Canyon).
58 Assimilating Woodgate et al.'s (2005) mooring data and other available in-situ data into a
59 diagnostic model, Panteleev et al. (2010) showed how the shelf circulation is sensitive to the
60 seasonal wind patterns. During mid- to late-autumn, when the winds tend to be strongly out of the
61 northeast, more of the inflowing Pacific water is diverted towards Herald Canyon. This is

62 consistent with the numerical model results of Winsor and Chapman (2004) and also with the
63 inferences of Weingartner et al. (1998) based on data from the eastern shelf. In contrast, during
64 spring when the winds weaken, the flow is more evenly distributed among the three branches. On
65 shorter timescales the currents in the two canyons that cut into the shelf – Barrow Canyon in the
66 east and Herald Canyon in the west – are strongly influenced by synoptic weather systems.
67 Upwelling and reversed flow to the south readily occur in both canyons (Aagaard and Roach
68 1990; Pickart et al., 2010).

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70 The detailed circulation and ultimate fate of the Pacific water in the northern part of the
71 Chukchi shelf remain uncertain at this point. Numerical models imply that when the wind forcing
72 is weak much of the water ultimately ends up in Barrow Canyon (Winsor and Chapman, 2004;
73 Spall 2007). This is because the flow generally follows bottom depth contours (Figure 1). The
74 water in the eastern branch roughly parallels the coast of Alaska and flows directly into the
75 canyon. After the water in the middle branch passes through Central Channel the topography
76 steers it anti-cyclonically around Hanna Shoal towards the head of Barrow Canyon. For the
77 western pathway, north of Herald Shoal the bathymetric contours bend to the east (Figure 1)
78 which also brings some of the water in this branch around Hanna Shoal into Barrow Canyon. The
79 remainder of the western branch continues northward through Herald Canyon and reaches the
80 canyon mouth where it encounters the edge of the Chukchi shelf. Previous results suggest that
81 this outflow forms an eastward-flowing shelfbreak current (Pickart et al., 2005; Mathis et al.,
82 2007; Pickart et al., 2010) that eventually reaches the mouth of Barrow Canyon.

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84 The timing of the northward progression of Pacific water through the Chukchi Sea is also
85 uncertain. Both observations and models indicate that it takes roughly 3-4 months for water to
86 traverse from Bering Strait to the central portion of the shelf (Woodgate et al., 2005; Winsor and
87 Chapman, 2004; Spall, 2007). Not surprisingly, it takes significantly longer for the water in the
88 middle and western branches to reach Barrow Canyon than it does for the water in the eastern
89 branch. Spall's (2007) model indicates an advective time of 6-8 months for the flow through
90 Central Channel and Herald Canyon that bends to the east on the northern part of the Chukchi
91 shelf and drains through Barrow Canyon. The simulation of Winsor and Chapman (2004)
92 suggests that the portion of the western branch that forms the Chukchi shelfbreak jet takes more

93 than two years to reach the mouth of Barrow canyon. In contrast, Weingartner et al. (1998)
94 estimate that in the summer months the swift Alaskan Coastal Current transports water from
95 Bering Strait to the head of Barrow Canyon in just 2-3 months. It should be remembered,
96 however, that the travel times of all of the branches are significantly longer when under the
97 influence of northerly winds (Winsor and Chapman, 2004).

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99 In addition to the model studies quoted above, observational evidence also suggests a "long
100 route" of Pacific water on the northern part of the Chukchi shelf. Shipboard velocity data indicate
101 that some of the western branch through Herald Canyon is diverted to the east beyond Herald
102 Shoal (Weingartner et al., 2005; Pickart et al., 2010). As seen in Figure 1, this flow should join
103 the outflow from Central Channel and circulate anti-cyclonically around Hanna Shoal into
104 Barrow Canyon. Such a clockwise flow around the shoal is suggested by limited observations
105 (Weingartner et al., 2013a). As such, there should be eastward flow to the southeast of Hanna
106 Shoal, and the mooring data of Weingartner et al. (2005) confirm this. Furthermore, moored
107 temperature records at the head of Barrow Canyon have indicated the presence of Pacific winter
108 water well into the summer season (Woodgate et al., 2005; Weingartner et al., 2005), consistent
109 with such a long advective route. However, while supportive, the observational data to date are
110 sparse. Consequently, the circulation pattern depicted by the models – in particular the
111 topographically steered flow around Hanna Shoal into Barrow Canyon – remains to be confirmed
112 using in-situ measurements.

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114 In this paper we use data from two shipboard hydrographic/velocity/tracer surveys, in
115 conjunction with mooring measurements, to elucidate the circulation of cold Pacific winter water
116 on the northeast Chukchi shelf. The data were collected in late-spring / early-summer 2010 and
117 2011. While the observations are not synoptic, the winds were relatively steady throughout the
118 measurement periods and so the results are not aliased by variable atmospheric forcing. Our data
119 indicate that the winter water circulation is more complex than previously envisioned, and not
120 entirely in line with the previous model results. We begin with a presentation of the various
121 datasets, followed by an overview of the atmospheric conditions during the fieldwork. Then we
122 present the characteristics and distribution of winter water on the shelf. A clear circulation pattern
123 emerges, which elucidates the long route for winter water to reach Barrow Canyon. We define a

124 main pathway on the central shelf and discuss the nature and cause of the variation in water mass
125 properties along this route.

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127 **2. Data and Methods**

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129 **2.1 Shipboard Hydrographic Data**

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131 Two cruises on the USCGC *Healy* were carried out in early-summer 2010 and 2011 in the
132 Chukchi Sea as part of the “Impacts of Climate on Ecosystem and Chemistry of the Arctic
133 Pacific Environment” (ICESCAPE) program (Arrigo et al., 2012). The first cruise took place
134 from 13 June – 22 July, 2010 (ICESCAPE I), and the second took place from 25 June – 29 July,
135 2011 (ICESCAPE II). Some of the sections were occupied on both cruises, but the lateral
136 coverage on the second cruise was more extensive. For this reason, and to maximize synopticity,
137 we focus on the 2011 dataset. As seen in Figure 2, the 2011 survey nicely covered the northeast
138 Chukchi shelf, including the southern half of Barrow Canyon. For completeness we include one
139 transect from the 2010 cruise which crosses the Central Channel, since this section was not
140 occupied in 2011. We emphasize that both surveys took place during the same time of year
141 (June/July), and the distribution of winter water on the shelf was very similar in each. Hence this
142 supports the inclusion of the 2010 Central Channel section in the study.

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144 On both cruises a Sea-Bird 911+ conductivity/temperature/depth (CTD) instrument was
145 mounted on a 12-position rosette with 30-liter Niskin bottles. The temperature sensors
146 underwent laboratory calibrations before and after each cruise, and the conductivity sensors were
147 calibrated using the in-situ salinity data. The resulting accuracies are 0.008°C for temperature
148 and 0.004 for salinity for waters on the shelf, and 0.002°C and 0.002, respectively, for waters on
149 the continental slope. During one of the transects on the 2011 cruise, expendable CTD probes
150 were used (XCTDs). The accuracy of the XCTD data are taken to be 0.02°C for temperature,
151 0.04 for salinity, and 1 m for depth (see Kadko *et al.*, 2008). Vertical sections of potential
152 temperature, salinity, and potential density were constructed using a Laplacian-Spline
153 interpolator with a lateral grid spacing of 5-15 km and a vertical grid spacing of 2-5 m.

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156 **2.2 Shipboard Velocity Data**

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158 Velocity data were obtained on both ICESCAPE cruises using the hull-mounted Ocean
159 Surveyor 150 KHz Acoustic Doppler Current Profiler (ADCP) on *Healy*. The data were acquired
160 using the University of Hawaii's UHDAS software and were post-processed with the CODAS3
161 software package (see <http://currents.soest.hawaii.edu>). The two main challenges for obtaining
162 useful data were the presence of ice under the hull, which often times blocks the transducer, and
163 the occurrence of bubbles under certain conditions. The CODAS3 visual editing tools were
164 applied to minimize both of these error sources.

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166 Low frequency (i.e. days to weeks) ADCP errors can result from the misalignment of the
167 ADCP transducer or from small errors in the ship's heading record. *Healy's* ADCP installation
168 was assessed using bottom track and water track calibrations. These compare the ship track or
169 ship's accelerations derived from the ADCP with those derived from the GPS data. The extensive
170 periods of shallow water encountered during the ICESCAPE cruises allowed for long timeseries
171 of bottom tracking. As such, the time-invariant hardware alignment was well determined, and the
172 time-dependent features of the timeseries helped quantify heading device fluctuations. The
173 resulting small bottom track standard deviation of 0.1 degrees suggests that fluctuations of the
174 ship's heading unit were small. We therefore estimate that the low-frequency ADCP errors were
175 less than 2 cm/s. Following the CODAS3 processing, the velocities were de-tided using the
176 Oregon State University barotropic tidal prediction model <http://volkov.oce.orst.edu/tides>
177 (Padman and Erofeeva, 2004).

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179 The ADCP data do not extend over the full water column. There is a surface blanking region
180 of roughly 15-20 m, and a bottom blanking region of 15% of the local water depth (e.g. 7.5 m
181 for a shelf depth of 50 m). Since we are interested in the circulation of Pacific winter water
182 throughout the water column, we computed absolute geostrophic velocities by referencing the
183 thermal wind shear using the ADCP data. This was done as follows. Using the gridded vertical
184 sections of temperature and salinity, we constructed vertical sections of dynamic height (relative
185 to the sea surface). Analogous sections of velocity perpendicular to each transect were made

186 using the de-tided ADCP data. Then, at each cross-stream grid point across the section, the
187 absolute geostrophic velocity was calculated by matching the vertically-integrated relative
188 geostrophic velocity to the integrated ADCP velocity over their common depth range. This was
189 done for the XCTD transect as well. Finally, vertical sections of absolute geostrophic velocity
190 were constructed using the same Laplacian-Spline interpolator employed for the CTD variables.

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192 **2.3 Shipboard Nitrate data**

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194 Nutrients were measured during both cruises at standard depths (typically five levels for a 50
195 m cast: 5 m, 10 m, 25 m, just above the sea floor, and at the depth of the fluorescence maximum).
196 A Seal Analytical continuous-flow AutoAnalyzer 3 was used, with modifications according to
197 Armstrong et al. (1967). Here we use only the nitrate data (NO_3^-), which were typically finalized
198 within 24 hours of the water sample collection at a station. For detailed presentations of the
199 ICESCAPE nutrient data see Mills et al. (2015) and Lowry et al. (2015).

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201 **2.4 Shipboard Wind Data**

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203 During the ICESCAPE cruises *Healy* had four meteorological sensors mounted on various
204 parts of the ship, providing one-minute averaged true wind speed and direction by accounting for
205 the motion of the vessel using the GPS data. In general, the four sensors gave similar results for
206 both speed and direction, but the instrument located on the starboard main mast was considered to
207 be most reliable and subject to the least amount of blocking due to the vessel's superstructure (S.
208 Roberts, pers. comm., 2012). Accordingly, we used the data from the main mast starboard sensor.
209 To assess the accuracy of this product, we compared it to timeseries from the meteorological
210 station in Pt. Barrow, AK (marked by the star in Figure 3). Although the ship was often a fair
211 distance from the weather station, the two timeseries were generally similar. During the
212 occupation of the transect across the central part of Barrow Canyon (section BCC in Figure 2),
213 which was the closest section to the meteorological station, the agreement was near perfect. This
214 provides confidence in the accuracy of the wind data obtained from *Healy*.

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2.5 Reanalysis Fields and Ice Concentration Data

To assess the large-scale atmospheric setting during the ICESCAPE program, and to estimate salt production due to ice formation within polynyas, we use fields from the North American Regional Reanalysis (NARR) (Mesinger et al., 2006) and sea ice concentration data from the Advanced Microwave Scanning Radiometer (AMSR-E) (Spreen et al., 2008). The spatial and temporal resolution of the NARR is 32 km and 6 hours, respectively. Previous work has shown that although the surface meteorological fields from the NARR are in good agreement with observations at high latitudes, the sensible and latent heat fluxes are biased high (Renfrew et al., 2009). As a result, we use a well-established bulk parameterization to estimate the turbulent fluxes (Smith, 1988; DeCosmo et al., 1996) with the NARR surface meteorological fields as inputs (Moore et al., 2014).

The fidelity of the radiative fluxes in all reanalyses is a function of the ability of the particular reanalysis to capture the cloud fraction (Walsh et al., 2009). In this regard, the NARR is one of the few reanalyses to correctly capture the transition in the change in sign of the net radiative forcing that occurs in the Arctic during the spring period (Walsh et al., 2009). On the other hand, in a comparison to observations at the ARM site at Point Barrow, the NARR downwelling longwave flux is biased high by ~10% during the winter and biased low by ~10% during the summer with the transition occurring in May (Walsh et al., 2009). As a result, the net longwave cooling in the NARR is underestimated during the winter and overestimated in the summer.

The spatial resolution of the daily AMSR-E ice concentration data is 6.25 km, and the values are deemed accurate to $\pm 10\%$ (Cavaleiri et al., 1991). For the salt production calculations, daily mean values of the NARR fields were generated and then interpolated to the AMSR-E spatial grid.

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2.6 Mooring Data

2.6.1 Chukchi Shelf

To increase the data coverage on the Chukchi shelf we used timeseries from a collection of moorings that were maintained as part of other measurement programs (see Figure 2 for the locations of the moorings).

(i) Barrow Canyon Array

An array of 6 moorings was maintained across the head of Barrow Canyon during 2011 by the Bureau of Ocean Energy Management, Shell, and Conoco Phillips. For a detailed description of the array the reader is referred to Weingartner et al. (2013b). Each mooring contained an upward-facing ADCP and near-bottom MicroCat CTD sensor. After low-passing the velocity records using a 36-hour filter to remove the tidal signal, the velocities were averaged vertically over the depth of the water column. All variables were then subsampled to produce hourly timeseries. Since our study focuses on winter water, it was necessary to consider a broader time period for the Barrow Canyon mooring records than that coincident with the ICESCAPE II cruise. This is because the canyon is situated along the eastern-most pathway of Pacific Water (see Figure 1), which is characterized by swift northeastward flow. Hence the flushing of winter water along this branch is faster than along the other pathways. In order to capture the normal seasonal presence of the winter water at the head of Barrow Canyon, we used the mooring data from May – July 2011. The ADCP velocity data are available for this period for all 6 moorings. However, the MicroCat CTD sensor at the offshore-most mooring experienced technical problems in early May, limiting the temporal coverage to 11 days at that site.

(ii) Interior Moorings

Data from three additional moorings were used to identify the flow of winter water in the region between the southern two ICESCAPE hydrographic transects (Figure 2). Detailed information about the moorings, as well as the data availability for the three sites, is available at <https://workspace.aos.org/group/6316/project/56008/files>. The western site (Crackerjack) and

279 middle site (Burger) were maintained by Shell, and the eastern site (Klondike) was maintained
280 by Conoco Phillips. At each of the sites velocity was measured using an upward-facing ADCP,
281 and pressure, temperature, and salinity were measured using ancillary CTD sensors.
282 Unfortunately, only data from 2009-10 are available for the interior moorings. However, since
283 we are considering only time periods that are characterized by relatively steady winds (i.e. no
284 storm events, see below), it is consistent to use data from the same season during these previous
285 years (similar reasoning applies for including the southern-most CTD transect from ICESCAPE
286 I). The ADCPs on the interior moorings measured the velocity in three bins (surface, middle, and
287 bottom), and the data were similarly low-passed and subsampled to produce daily values. The
288 salinity values at the Klondike site were unrealistically large, hence this timeseries was omitted
289 from our analysis.

290

291 *(iii) Removing Storm Events*

292 As detailed below in Section 3, the ICESCAPE II cruise occurred during a time when the
293 winds were relatively steady. As such, it was necessary to remove any time periods in the
294 mooring records that were influenced by strong storms so as not to bias the results. This was
295 accomplished using the hourly-averaged Pt. Barrow weather station wind record in conjunction
296 with the mooring velocity data. In particular, for the Barrow Canyon moorings we used the
297 component of wind and water column velocity along 56 degrees true (the along-canyon direction),
298 since the two variables are most significantly correlated in this direction (Nobre and Pickart,
299 2014). The most prevalent wind events are upwelling storms when the winds blow strongly out of
300 the northeast for a period of several days or longer, associated with up-canyon (southwest) flow
301 with a lag of roughly 12 hours. Accordingly, we scrutinized the Barrow Canyon mooring data
302 and Pt. Barrow wind record for the signature of such wind-forced upwelling, and found two
303 events in May 2011 and one in May 2010. These periods were removed from the mooring data.
304 An analogous procedure was applied to the three interior mooring sites, and one upwelling event
305 in both May 2009 and May 2010 were removed from those records.

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307 **2.62 Bering Strait**

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309 To help interpret the variation in winter water properties in the shipboard data we used
310 timeseries data from mooring A3 in Bering Strait during spring/summer 2011 (see Figure 1 for
311 the location of the mooring). The instrumentation included a MicroCat and an upward-facing
312 ADCP at 48 m depth. As with the other timeseries data, the records were low-pass filtered then
313 subsampled hourly. Detailed information about the mooring can be found at
314 http://psc.apl.washington.edu/HLD/Bstrait/CruiseReportKhromov2010wEL_verNov11.pdf.

316 **3. Results**

318 **3.1 Atmospheric Setting**

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320 We first describe the atmospheric conditions during which the ICESCAPE program took
321 place. Since this study focuses on the 2011 cruise, we present results from that time period, i.e.
322 25 June to 29 July, 2011. Throughout the five week cruise the winds were predominantly out of
323 the east/northeast (Figure 3). In fact, the only brief period of westerly winds occurred during the
324 occupation of the BCC transect (Barrow Canyon, Figure 2). The mean wind speed over the
325 domain was 6.6 m/s along 246°T. How does this compare to the typical early summer
326 conditions? Using the NARR data we constructed maps of the mean sea level pressure (SLP)
327 and 10 m winds for the period 2000-2011 (Figure 4, top row). At this time of year there is a
328 signature of the Beaufort High (Moore, 2012) associated with anti-cyclonic winds in the
329 northern Chukchi Sea and Canada Basin. However, there is very little evidence of the Aleutian
330 Low in the Bering Sea and Gulf of Alaska during the summer, which is consistent with the
331 seasonality of this feature (Favorite et al., 1976).

332
333 Compared to the 12-year climatology, in summer 2011 the Beaufort High was strong and
334 displaced to the northwest, and there was a well-developed Aleutian Low situated in the northern
335 Bering Sea (Figure 4, middle row). Together, these two centers of action led to enhanced
336 easterly winds in the northern Chukchi Sea. Spall et al. (2014) argue that these winds drove
337 upwelling at the shelfbreak which led to the massive phytoplankton bloom observed at the
338 northern end of transect CSW on the 2011 cruise. The anomaly fields indicate, however, that the

339 very southern portion of the Chukchi shelf and the Bering Strait region experienced weaker
340 winds than normal (Figure 4, bottom row).

341
342 To obtain an overall characterization of the winds on the Chukchi shelf we considered a box
343 encompassing the shelf (see Figure 4) and computed the mean wind vector for each of the 12
344 years corresponding to the early summer period (Figure 5). Consistent with the results of earlier
345 studies (Brugler et al., 2014; Spall et al., 2014) one sees that, in general, the winds in summer
346 increased over the latter half of the period (with the exception of 2010 when the winds were
347 quite weak). Brugler et al. (2014) attributed this to a combination of a stronger Beaufort High
348 and deeper Aleutian Low (primarily the latter), and this is consistent with the SLP anomaly map
349 of Figure 4 for 2011. In addition to the increased wind speed in recent years, the meridional
350 component has become northerly, as opposed to earlier in the decade when it was mostly
351 southerly.

352
353 According to the idealized model study of Winsor and Chapman (2004), the circulation in
354 the Chukchi Sea is most sensitive to easterly winds (relative to the no wind case). For an easterly
355 wind speed of 7.7 m/s their model streamlines veered to the western side of the shelf, and the
356 coastal pathway on the northeast shelf (i.e. the model equivalent of the Alaskan Coastal Current)
357 reversed to the south. In light of the fact that the mean zonal wind speed measured by the ship
358 (Figure 3) was 6 m/s out of the east, this would suggest that the circulation during the 2011
359 cruise should have been significantly influenced by the wind. However, as seen below, the
360 coastal pathway was quite strong, and the combined transport of the two main flow branches on
361 the eastern shelf was comparable to that measured earlier in the decade under weaker winds
362 (Gong and Pickart, 2015). This apparent discrepancy is resolved by noting that in Winsor and
363 Chapman's (2004) model the winds were applied over the entire domain. As seen in Figure 5,
364 according to NARR, the average windspeed over the entire Chukchi Sea was smaller (zonal
365 component of 2 m/s) due to the weak winds in the southern part of the shelf (Figure 4, second
366 row). In any event, we characterize the winds during the ICESCAPE II cruise as being generally
367 steady and of moderate strength, but not strong enough to reduce the transport of Pacific water in
368 the eastern Chukchi Sea nor cause any flow reversals on the shelf.

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372 **3.2 Identification and Characteristics of Winter Water**

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374 As discussed in the introduction, winter water is formed when surface water freezes and the
375 resulting brine rejection leads to convective overturning of the water column. Consequently, the
376 newly ventilated winter water is close to the freezing point, and such water is commonly
377 observed in near-bottom mooring records on the Chukchi shelf during winter (e.g. Weingartner
378 et al., 1998; Woodgate et al., 2005). By the time that the two ICESCAPE surveys took place in
379 June/July the water had warmed somewhat. As such, in this study we define newly ventilated
380 winter water (hereafter referred to simply as winter water) as being colder than -1.6°C with a
381 salinity between 31.5 and 33.5.¹ The volumetric T/S plot of Figure 6a demonstrates that this
382 water was the most common water mass present on the Chukchi shelf during the 2011 survey
383 (true as well for 2010, not shown). Roughly 35% of the water on the shelf was colder than -
384 1.6°C (see also Lowry et al. (2015)). As previously mentioned, this water is of primary
385 importance to the ecosystem of the western Arctic Ocean: it is dense enough to ventilate the
386 upper halocline of the Canada Basin, and its elevated nitrate concentration (Figure 6b) spurs
387 primary production throughout the region.

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389 We are interested in determining the pathways of the winter water, so for each vertical
390 section in the survey we identified those stations containing winter water that were also located
391 within a well-defined jet as deduced from the absolute geostrophic velocity. For example,
392 consider the Central Channel section (labeled CC in Figure 2), which is shown in Figure 7a
393 (oriented such that the viewer is looking north). One sees that there are three distinct regions of
394 winter water (middle panel of the figure): one within the channel, one on top of the ridge that
395 forms the eastern side of the channel, and a third narrower lens at the base of the slope leading to
396 the Alaskan coast. The first and third lenses are embedded, respectively, in two poleward jets
397 (right-hand panel of Figure 7a, where the -1.6°C isotherm is marked by the thick black line). The
398 western jet is the well-known Central Channel branch and is associated with downward-sloping

¹ There was no Atlantic water or hyper-saline Pacific winter water (Weingartner et al., 1998) observed on the Chukchi shelf during either ICESCAPE cruise, hence none of the water sampled was saltier than 33.5.

399 isopycnals to the east resulting in a bottom-intensified flow. The eastern jet is the Alaskan
400 Coastal Current (ACC). One sees that at this early point in the summer the ACC is advecting the
401 last vestiges of winter water at depth along with warm Alaskan coastal water in the upper layer
402 (in this snapshot the current is weakly baroclinic). Curiously, the third lens of winter water in the
403 CC section is centered in a region of weak southward flow.

404

405 At the next section to the north (CN in Figure 2), the Central Channel is no longer a
406 topographic feature. Now there are only two lenses of winter water. The first is located on the
407 outer shelf, and much of it is embedded in a poleward jet (Figure 7b). This is clearly the Central
408 Channel branch after it has emerged from the channel, but it also appears to include a
409 contribution from Herald Canyon that has been diverted eastward along the northern shelf.
410 This is because of the relatively large width of the feature (shown below), and the fact that the
411 inner part of the jet contains substantially warmer water in the upper layer. This suggests a
412 shorter residence time from Bering Strait for that portion of the jet (transporting warm water
413 from the Bering Sea). The second lens of winter water in the CN section straddles the top of the
414 ridge in the middle of the transect. This is the same ridge that was present in the CC section; as
415 seen in Figure 1, the ridge extends to the southwest from Hanna shoal (and contains numerous
416 gaps). As was the case in the CC section, the lens of winter water on the ridge in the CN transect
417 is progressing to the south (the transport is greater here, see below). We believe that this is not a
418 coincidence, and that winter water flows back towards Bering Strait along the ridge. This is
419 further supported by the fact that the same feature was present in the 2010 occupation of CN (not
420 shown). We discuss the origin and fate of this pathway in Section 3.3.

421

422 It is instructive to show two more vertical sections from the 2011 ICESCAPE survey: the
423 CSC transect and the HSSE transect (Figures 7c and 7d, respectively). The former crosses the
424 shelfbreak into the Canada Basin, and one sees that a thin layer of winter water extends across
425 the shelf onto the upper-slope. All of this water is progressing eastward, transported by two
426 distinct currents – the shelfbreak jet and a jet on the shelf. The shelfbreak jet is characterized by
427 isopycnals sloping downward offshore which leads to a bottom-intensified flow. This is
428 consistent with the shelfbreak jet observed in the Beaufort Sea at this time of year (i.e. a bottom-
429 intensified flow of winter water, see Spall et al., 2008). However, unlike the Beaufort shelfbreak

430 jet which arises largely from outflow from Barrow Canyon (Pickart et al., 2005), the shelfbreak
431 jet in Figure 7c is being fed from the Herald Canyon outflow (Pickart et al., 2010). The mid-
432 shelf jet at section CSC is likely the continuation of the poleward jet at section CN. The final
433 section shown (HSSE) extends from Barrow Canyon to Hanna Shoal, and in Figure 7d the
434 viewer is looking southward. We include this transect because modeling studies suggest that the
435 Pacific water emanating from Bering Strait through the Central Channel should flow anti-
436 cyclonically around Hanna Shoal. As seen in Figure 7d, there is indeed a well-defined jet on the
437 shelf advecting winter water to the southwest (although the amount of winter water at this
438 location is small).

439

440 **3.3 Circulation Diagram and Volume Transports**

441

442 Using the information from the complete set of vertical sections in Figure 2, together with
443 the mooring data, we now construct a map of winter water circulation on the northeast Chukchi
444 shelf. Figure 8 shows the lateral distribution of potential temperature averaged over the bottom
445 15 m of the water column (excluding stations with bottom depth deeper than 85 m). Overlaid on
446 this are vectors denoting the transport per unit width associated with the winter water at each
447 CTD station (i.e. water colder than -1.6°C). The reader should keep in mind that these are not
448 true vectors, but instead represent the flow normal to each transect from the absolute geostrophic
449 velocity. (We note that the winter water transport was so small at some of the stations that the
450 vectors are not visible in Figure 8.) At the mooring sites we have plotted the mean near-bottom
451 velocity vectors (true vectors) for the winter water (excluding storms, as explained above). From
452 this figure some clear patterns emerge.

453

454 The coastal pathway is characterized by the warmest temperatures on the shelf. This
455 pathway extends into Barrow Canyon (along the eastern flank of the canyon) and is broadest at
456 the IC section where the bathymetry becomes nearly flat going offshore. Note also that the
457 bottom water on the top of Hanna Shoal is relatively warm. The shoal is thought to be a region
458 of stagnant flow (Martin and Drucker, 1997), although this could not be verified here because
459 the water depth on top of the shoal was too shallow for *Healy's* hull-mounted ADCP to return
460 useful data. Aside from these two regions, the deep portion of the shelf was largely filled with

461 winter water. The anti-cyclonic pathway of this water from Central Channel around Hanna Shoal
462 is clearly evident, providing the first unambiguous observational evidence of this circulation
463 feature predicted by the models. However, close inspection of Figure 8 reveals some unexpected
464 aspects of the circulation not seen in previous numerical simulations.

465
466 Based on this figure we constructed a flow schematic showing the progression of winter
467 water across the shelf (Figure 9a). This circulation diagram is true to the data, with the following
468 caveats. First, although there was no water colder than -1.6°C observed at section IC during the
469 ICESCAPE II cruise, it is likely that, earlier in the season, winter water was advected along the
470 coastal pathway through this section. This is consistent with the fact that the water within the
471 ACC at section IC was the coldest in the entire transect ($<0^{\circ}\text{C}$), and the fact that winter water
472 was observed at the head of Barrow Canyon earlier in the season in the mooring data. Secondly,
473 at section BCX there is negligibly small transport of winter water in the two pathways identified
474 on this section (see Figure 9b). However, the two jets that were observed on the transect were
475 transporting anomalously cold water ($<-1.4^{\circ}\text{C}$), which we take to be the leading edge of the
476 winter water along this route. Finally, during the occupation of the CSW section the shelfbreak
477 jet was in the process of spinning back up after a wind-driven flow reversal (see Spall et al.,
478 2014), so the eastward transport of winter water in the jet was somewhat small at this time.

479
480 The flow map of Figure 9a is the first observationally-based depiction of winter water
481 circulation on the northeast Chukchi shelf. The two well-known pathways of Pacific water on
482 the eastern portion of the shelf– the ACC and the Central Channel branch – advect water to the
483 north. The latter is then joined by water that was diverted from Herald Canyon, and together this
484 merged flow approaches Hanna Shoal. However, rather than simply flowing around the northern
485 side of the shoal as predicted by the models, this branch splits and the Pacific water progresses
486 around both sides of the shoal. These two branches subdivide further as the water progresses
487 towards the head of Barrow Canyon. There is also evidence of an offshoot from the Central
488 Channel pathway just north of channel. Thus, there are two means by which winter water feeds
489 Barrow Canyon: (1) a direct coastal route which flows along the eastern flank of the canyon (fed
490 by some water diverted from the Central Channel); and (2) a circuitous pathway that divides into
491 a number of smaller filaments together providing water to both sides of the canyon. The coastal

492 route is swift and the winter water is largely flushed through this pathway by early summer,
493 while the interior route is substantially longer with slower flow speeds feeding Barrow Canyon
494 well into the summer (see also Nobre et al., 2014). Finally, winter water flows to the east in the
495 shelfbreak jet which will ultimately reach the mouth of Barrow Canyon and likely interact with
496 the outflow from the shelf.

497
498 The synoptic volume transports computed along these different flow paths (marked in
499 Figure 9b and listed in Table 1) give a remarkably consistent view. We present the transport of
500 winter water for each section (black numbers), along with the total transport of the flow feature
501 in which the winter water is embedded (red numbers). Starting with the coastal jet, the total
502 transport is the same at sections CC and IC (0.43 Sv and 0.44 Sv, respectively; $1 \text{ Sv} = 10^6 \text{ m}^3\text{s}^{-1}$),
503 but, as noted above, there was little to no winter water remaining in this pathway by the time that
504 the ICESCAPE surveys took place. The total transport of shelfbreak jet varied from 0.10 Sv to
505 0.28 Sv. However, as noted above, the CSW transect was occupied near the end of a period of
506 enhanced easterly winds, so the jet (particularly its upper layer) was still largely reversed at that
507 location (see Spall et al., 2014). Assuming that the undisturbed transport is significantly larger
508 than this, it appears that there was a general diminishment of transport of the jet towards the east.
509 This was true as well for the winter water transport (which on average accounted for
510 approximately 60% of the total transport of the shelfbreak jet).

511
512 The total transport of the mid-shelf flow evolves as follows. Progressing along the Central
513 Channel pathway, the volume flux remains roughly constant as it goes from CC to CSW
514 approaching Hanna Shoal (order 0.5 Sv).² East of here, where the flow bifurcates, there is
515 roughly an equal split in transport around the two sides of the shoal. In particular, 42% of the
516 transport progresses anti-cyclonically around the northern side of the shoal, and 58% of the
517 transport progresses cyclonically around the southern side. Encouragingly, the sum of these two
518 branches (eastward transport at HSN + southward transport at CSW = 0.53 Sv) is nearly
519 identical to that approaching the shoal (eastward transport at CSW = 0.51 Sv). As these two
520 branches flow around Hanna Shoal they split again into smaller filaments. On the southern side

² There is an increase in transport where the Herald Canyon branch merges with the Central Channel branch, but keep in mind that the CC section was occupied in 2010.

521 the split is nearly even in transport (0.16 Sv at CN versus 0.19 Sv at HSS) and roughly conserves
522 mass (total of 0.35 Sv compared to 0.31 Sv before the split). On the northern side, the transport
523 remains constant from HSN (0.22 Sv) to CSE (0.23 Sv), but then decreases in value at HSSE
524 (0.11 Sv). This decrease likely arises because some of the flow is diverted to the outer-shelf or
525 into the shelfbreak jet due to the presence of a small canyon upstream of HSSE. We note that
526 there is a small amount of water (including winter water) flowing southward on the eastern end
527 of this section (not shown), which has also been observed in larger amounts in other
528 hydrographic surveys (R. Pickart, unpublished data). Beyond section HSSE the anti-cyclonic
529 branch splits into two filaments, with most of the transport going into the northern filament.

530

531 Mass is conserved over the locus of the flow branches in Figure 9 (within the uncertainty
532 of the transport measurements), and we can account for the volume flux observed through the
533 head of Barrow Canyon. At section BCH, 0.99 Sv of Pacific water is flowing northward through
534 the canyon, which is approximately equal to the sum of the ACC transport in the coastal route
535 and the transport in the Central Channel branch (together totaling 0.96, using the average value
536 for CC, CN, and CSW). Furthermore, we can make inferences about the partitioning/origin of
537 the flow through the canyon. As seen in Figure 9b, 0.20 Sv is flowing northward on the western
538 flank of the canyon at BCH, and 0.79 Sv is flowing northward on the eastern flank. This implies
539 that the water flowing anti-cyclonically around the northern side of Hanna shoal (order 0.2 Sv)
540 feeds the western flank of Barrow Canyon, while the combination of the water flowing
541 cyclonically around the southern side of Hanna Shoal plus the ACC (which together sum to
542 order 0.8 Sv) feeds the eastern flank. We note that this is consistent with the mean vectors from
543 the Barrow Canyon mooring array which show southeastward flow emanating from south of
544 Hanna Shoal to the eastern side of the canyon (Figure 8).

545

546 The transport of winter water on the mid-shelf is partitioned similarly to the full transport of
547 the branches (Figure 9b), although mass doesn't need to be (nor is it) conserved over the whole
548 domain because of the seasonal presence of this water mass. As was the case for the full
549 transport, roughly half of the winter water from the Central Channel branch flows on either side
550 of Hanna Shoal (roughly 0.25 Sv approaches the shoal, and 0.13 Sv continues east while 0.13 Sv
551 is diverted south). The difference now is that, in the region where the filaments form, the

552 transport of winter water is quite small. Upstream of here 46% of the total transport (on average)
553 is due to winter water, which decreases to only 17% in the filaments. This suggests, as noted
554 above, that in the 2011 survey this region was at the leading edge of the winter water, which is
555 consistent with the small amount of water colder than -1.6°C observed flowing north in Barrow
556 Canyon (most of the winter water in the coastal branch had likely passed through the canyon
557 prior to the survey). Later in the season the winter water from the interior pathways around
558 Hanna Shoal should begin draining through the canyon. This is in line with the observation that
559 winter water is prevalent at section BCC during August but not July (Nobre and Pickart, 2014).

560

561 **3.4 Evolution of Winter Water Properties along the Central Pathway**

562

563 **3.41 In-situ Observations**

564

565 While all of the winter water on the shelf during the ICESCAPE II cruise was confined to a
566 relatively narrow temperature range (-1.78°C to -1.6°C), the salinity of this water mass displayed
567 substantial variation along the central pathway. This is seen in Figure 10, where we have
568 excluded the 2010 data (to remain synoptic) and have isolated the branch that flows from the
569 Central Channel anti-cyclonically around the north side of Hanna Shoal. We have also attempted
570 to exclude the portion of the flow that emanates from Herald Canyon by not considering the
571 outer-shelf portions of the sections in question. One sees that there is an abrupt change in salinity
572 of the winter water starting at the CSC line; i.e., the water becomes markedly fresher. It is of
573 interest to know whether this change was due to variation in the properties of the winter water
574 entering Bering Strait as the season progressed, or the result of local processes on the Chukchi
575 shelf. To answer this we need to know when the water that was measured at each section passed
576 through Bering Strait. The simplest approach is to assume a constant advective speed along the
577 pathway. From the absolute geostrophic velocity, the average speed for the six 2011 sections in
578 Figure 10 is 11.7 ± 2.8 cm/s (varying between 8.1 cm/s at HSSE and 15.6 cm/s at HSN). For
579 reasons explained below, we chose an advective speed of 10.5 cm/s, and Figure 11 shows the
580 mean potential temperature and salinity of the winter water at each section as a function of
581 elapsed time from Bering Strait. One sees that the salinity decreases and the temperature
582 increases for an elapsed time greater than 100 days (i.e. downstream of section HSN along the

583 pathway). If this is due to changes in the winter water entering the Chukchi Sea, it means that,
584 later in the season, the winter water flowing through Bering Strait was saltier and colder.

585
586 To investigate this we used the timeseries data from mooring A3 in the northern part of
587 Bering Strait in 2011. During the time period from late-December 2010 to early-May 2011, the
588 water entering the strait was at/near the freezing point, with very little variation in temperature
589 (not shown). The salinity, on the other hand, fluctuated quite a bit over this time. Using the
590 chosen advective speed of 10.5 cm/s, this implies that the winter water sampled between sections
591 CN and HSSE in Figure 9 passed through Bering Strait between 20 March and 11 April
592 (indicated by the grey shading in Figure 12). This range of dates of course depends on the
593 advective speed chosen, and, because of the uncertainty in our mean estimate, we considered a
594 range of speeds. However, as seen below in Section 3.42, a choice of 10.5 cm/s leads to the most
595 sensible results, and this value is well within the uncertainty of the calculated mean along the
596 central pathway.

597
598 One sees in Figure 12a that there was very little change in temperature of the winter water
599 in Bering Strait over this three week period (ranging from -1.77°C to -1.79°C). The temperatures
600 measured at the six CTD transects, on the other hand, ranged from -1.73°C to -1.62°C (Figure
601 11). Hence the winter water warmed during its transit along the pathway on the central shelf,
602 which is not surprising (likely due to solar warming as the ice receded and/or lateral mixing with
603 water outside of the pathway). The salinity of the winter water in the strait did vary significantly
604 over the three week period (Figure 12b); however, there was no trend. In particular, the water
605 did not become systematically saltier over this time which would have had to be the case in
606 order to explain the along-pathway variation seen in Figure 11. This implies that there was some
607 means by which the later-arriving winter water was salinized on the Chukchi shelf subsequent to
608 flowing through Bering Strait.

609
610 To see this more clearly, we plotted the temperature and salinity of the winter water
611 measured at each section, adjusted back in time to when the water passed through Bering Strait,
612 compared to the mooring timeseries (Figure 13). This shows the degree of warming that took
613 place on the shelf (Figure 13a), indicating that the later-arriving water (after 27 March) warmed

614 less. The salinity comparison reveals that the earlier-arriving water (prior to 27 March) is in line
615 with what was observed at the northern end of the central pathway in Figure 9 (i.e. sections
616 CSC, CSE, and HSSE). That is, the spatial variation in these sections generally follows the
617 temporal variation of the water passing through Bering Strait during this time. However, the
618 salinity at the southern three transects (CN, CSW, and HSN) is significantly larger than that of
619 the water which flowed through the strait in the latter part of the time period. Hence we need to
620 identify a mechanism that increased the salinity of this later-arriving winter water by 0.2 – 0.3
621 north of Bering Strait. We now argue that freezing and brine rejection, associated with polynya
622 formation, likely modified the winter water that was sampled in the southern part of the pathway
623 during the ICESCAPE II cruise.

624

625 **3.42 Polynya-driven water mass transformation**

626

627 A number of polynyas are observed to form in the Chukchi Sea during the winter and
628 spring. These include the Seward Peninsula Polynya, the Kotzebue Sound Polynya, the Cape
629 Lisburne Polynya, and the Northeast Chukchi Polynya (Stringer and Groves, 1991; Cavalieri and
630 Martin, 1994; Martin et al., 2005). Previous work has estimated the ice production and resulting
631 salt flux into the ocean that occurs in these polynyas (Winsor and Bjørk, 2000). We follow this
632 methodology to assess changes in salinity of the winter water as it passed through the southern
633 Chukchi Sea in the early part of summer of 2011. We adopt a Lagrangian approach in which the
634 salt enhancement is estimated for parcels that are initiated at the approximate location of the A3
635 mooring (66.3°N) and transit northwards to the CN line along central pathway. We use 167°W as
636 the meridian along which the parcels travel³ and consider three different speeds: 10 cm/s, 10.5
637 cm/s, and 11 cm/s (outside of this range, the results are very much at odds with the data). The
638 AMSR-E sea concentration data are used to identify regions along the parcel trajectories where
639 brine rejection occurs, and, as outlined in Section 2.5, the NARR fields are then used to estimate
640 the heat loss that leads to sea ice formation and the resulting salt flux into the water column.

³ This is the approximate longitude of the pathway. Results are not overly sensitive to this choice, and, due to the other uncertainties in the calculation, it was deemed unnecessary to make the pathway vary with latitude in this part of the domain (the precise pathway is in fact unknown).

641 In Figure 14 we show the sea ice concentration along 167°W between 66°N (near Bering
642 Strait) to 72°N (the approximate latitude of the CN section along the central pathway) from
643 March to July, 2011. During this period the Chukchi Sea transitions from 100% ice cover to open
644 water. Also shown are the locations of the parcels started at 66.3°N on March 20, April 8, and
645 April 20 with velocities of 10, 10.5 and 11 cm/s for each start date. As one can see, the parcels
646 released on March 20 transit the region under 100% ice cover except for two short periods around
647 April 1 and April 15 when they pass through regions of reduced ice cover near 68°N. This
648 implies minimal salinization of the water, indicating that parcels leaving Bering Strait near that
649 time would not be significantly transformed. This is consistent with Figure 13b. However, the
650 parcels released on April 8 pass through the second of these regions of reduced ice cover during
651 the latter half of April, as well as another region of reduced ice cover in May between 69°N and
652 71°N, resulting in an extended period of time within a polynya. Finally, the parcels released on
653 April 20 spend the first month of their transit under sea ice before passing into permanent open
654 water later in the season. This suggests that the middle of the three release times (or thereabouts)
655 is most conducive for modification of the winter water in polynyas while re-freezing is still
656 occurring; this is again consistent with figure 13b.

657

658 A series of snapshots of the sea ice concentration in the southern Chukchi Sea, from mid-
659 April to early-June, reveals that it is within the Cape Lisburne polynya where the salinization of
660 the winter water likely occurs (Figure 15). In particular, the early parcels (blue symbols) pass by
661 Cape Lisburne before the polynya opens up substantially, and stay north of the region of reduced
662 ice cover as they advect poleward. By contrast, the middle parcels (magenta symbols) first enter
663 the polynya south of Pt. Hope and remain within the area of reduced ice concentration for a
664 prolonged period, staying near the western edge of the polynya. The later-leaving parcels (black
665 symbols) don't reach the polynya until it is subsumed into the region of melt-back expanding
666 from south to north. Hence, the timing of salinization based on the polynya activity is in line with
667 the data (Figure 13b). We now ask, is the degree of salinity enhancement reasonable as well?

668

669 To answer this, we estimated the change in salinity along a parcel trajectory as follows. Sea
670 ice production was deemed to occur when the sea ice concentration was less than 100% and the

671 sea surface temperature was at the freezing point. When these conditions were met, sea ice
672 production was assumed to occur at a rate:

673
$$P = \frac{Q_{net}}{r_{ice} L_h},$$

674 where P is the ice production rate (m^3/s); Q_{net} is the daily mean net cooling (W/m^2), i.e. the sum
675 of the turbulent (latent and sensible) heat flux and the net longwave radiative flux; r_{ice} is the ice
676 density (kg/m^3); and L_h is the latent heat of fusion (J/kg , see Cavalieri and Martin, 1994).

677 Following Cavalieri and Martin (1994), the resulting salt flux is given by:

678
$$F_s = r_{ice} P (S_w - S_i),$$

679
680 where F_s is the salt flux (kg/s); S_w is the salinity of the ocean water (32.6, based on the
681 observations); and S_i is the salinity of the frazil ice (formed under turbulent conditions) that is
682 given by:

683
$$S_i = 0.31 S_w.$$

684 To provide an estimate of the uncertainty in the calculations, cases were considered in which
685 the ice concentration was varied by $\pm 10\%$ with the proviso that ice concentration cannot be
686 higher than 100% or lower than 0%. The resulting salt flux was assumed to be mixed through the
687 volume of open water at that particular grid point, assuming a constant depth of 40 m (typical
688 shelf depth along the pathway), to generate a daily change in salinity. The net change in salinity
689 along the parcel trajectories was then calculated as the sum of these daily changes. In this
690 calculation we do not consider the effect of heating of the sea surface by the net shortwave
691 radiative or turbulent heat flux. The impact of this surface warming is implicitly taken into effect
692 by restricting the generation of the salt flux to conditions where the sea surface temperature is at
693 the freezing point.

694
695 The estimated net change in the salinity of the winter water progressing northward along
696 the central pathway from the Bering Strait mooring site to the CN section is shown in Figure 16
697 for a range of start dates from March 20 to April 23, 2011. There is an overall trend for the
698 salinization to increase for parcels leaving between late-March and early-April, before decreasing
699 the latter half of April. This is expected based on figures 14 and 15, i.e. due to the opening of the

700 Cape Lisburne polynya, followed by the melt-back on the southern shelf. Note that the start date
701 for maximum salinization is earlier for the slower moving parcels due to the longer time it takes
702 these parcels to reach the location of the polynya. Also, the magnitude of the change in salinity is
703 larger for the slower moving parcels; this is the result of the additional time that they spend in the
704 regions of reduced ice cover.

705

706 Based on these results and the analysis of Section 3.41, a parcel velocity of 10.5 cm/s
707 provides the best fit to the data. For this choice, the largest change in salinity is ~ 0.25 for parcels
708 leaving the Bering Strait mooring site around April 8. Due to the uncertainty in the actual
709 advective speed (which surely varied over the course of the central pathway), and in light of the
710 various assumptions made in the salt flux calculation, the agreement with figure 13b is quite
711 good. It suggests that we have correctly identified the cause of the increased salinity of the winter
712 water on the three upstream ICESCAPE II transects. This is also consistent with the slightly
713 colder temperature observed on these transects, indicative of more recent convective overturning.
714 We note that the integrated effect of changing the ice cover by $\pm 10\%$ along this trajectory
715 results in an approximate 10% uncertainty in the salinity change. A similar uncertainty is
716 expected to occur as a result of biases in the air-sea heat fluxes. The results are also sensitive to
717 the vicinity of the trajectories to the location of the Cape Lisburne polynyas, which of course is
718 not precisely known.

719

720 It is worth noting that longwave cooling played a major role in the salinization of the winter
721 water in the southern Chukchi Sea in 2011. In Figure 17 we have plotted the relative
722 contributions of the longwave radiative and turbulent heat fluxes to the change in salinity along
723 167°W for a parcel velocity of 10.5 cm/s as a function of start date. (Other choices for parcel
724 velocity resulted in similar results.) For early start dates, the turbulent heat flux makes the largest
725 contribution to the total change in salinity. However, after early-April the contribution from the
726 longwave radiative flux dominates. The cooling due to the turbulent heat fluxes is highly episodic
727 and is typically associated with cold air outbreaks where there is a large temperature difference
728 between the atmosphere and the ocean (Renfrew et al., 2002a). The longwave cooling, on the
729 other hand, is typically lower in magnitude but more continuous in time (Renfrew et al., 2002b;
730 Morales Maqueda et al., 2004). The transition in Figure 17 can therefore be understood as being

731 the result of the reduction in the air-sea temperature contrast that occurs as the region warms in
732 April and May, as well as the more steady nature of the radiative cooling. Recall from figure 14
733 that the salinization for parcels leaving in early-April (i.e. the best fit to the data) occurred during
734 two distinct periods. The first period of brine rejection was dominated by cooling due to the
735 turbulent heat fluxes, while that during the second period was driven mainly by longwave
736 cooling.

737

738 **4. Summary and Discussion**

739

740 The ICESCAPE hydrographic and velocity data, together with contemporaneous mooring
741 data from other programs, have allowed us to construct the most complete circulation diagram to
742 date of the northeast Chukchi shelf during the early-summer time period when newly-ventilated
743 winter water is progressing poleward. The observations suggest that as the Central Channel
744 branch approaches Hanna Shoal it bifurcates and flows around both sides of the shoal before
745 dividing into smaller filaments that enter Barrow Canyon. Encouragingly, mass is conserved in
746 this scheme: approximately 1 Sv flows into the northeast part of the shelf via the coastal pathway
747 (the Alaskan Coastal Current, ACC) and the Central Channel pathway, and an equal amount
748 exits through the head of Barrow Canyon. It appears that the branch circulating anti-cyclonically
749 around the northern side of Hanna Shoal feeds the western flank of the canyon, while the branch
750 passing on the south side of the shoal, together with the ACC, feeds the eastern flank.

751

752 While all of the winter water in the 2011 ICESCAPE survey was relatively close to the
753 freezing point, its salinity varied. In particular, the winter water sampled in the upstream portion
754 of the central pathway was saltier than that measured farther downstream. Mooring timeseries in
755 Bering Strait indicated that this pattern was not due to variation of the inflowing winter water
756 from the Bering Sea. We attributed the increase in salinity to transformation of the water within
757 the Chukchi Sea, as it passed along the western side of the Cape Lisburne polynya in early May.
758 Applying a sea-ice formation / brine rejection algorithm in a Lagrangian framework, the
759 predicted increase in salinity – and its timing – was on par to that measured in the ICESCAPE II
760 survey.

761

762 Our results have raised additional questions, and, at the same time, may offer some clues
763 regarding different aspects of the ecosystem of the Chukchi Sea. A primary question is, what
764 causes the bifurcation of the flow impinging on Hanna Shoal, which is not seen in the models.
765 One possibility is the presence of a canyon to the east of the HSN line (this can be seen as an
766 incursion of the isobaths in Figure 2, including the 50 m isobath). To first order the flow wants
767 to follow the topography, which means it should be deflected to the south here which may cause
768 it to recirculate to the west of the shoal. Such small scale topographic features may not be
769 present in the models. As noted earlier, another such small canyon is located to the north of the
770 HSSE line, which could cause some of the flow to divert from the shelf in that area. Canyons
771 such as these offer the chance for interaction between the shelfbreak jet and the circulation on
772 the outer shelf. Another question (for which we don't have an answer) is, why does the flow on
773 either side of Hanna Shoal break up into smaller filaments? Here again topography may be a
774 contributing factor. For example, several distinct jets were observed crossing the corrugated
775 ridge between Hanna and Herald Shoals. It could be that the water is seeking gaps in the ridge
776 through which to flow to the east. Stability of the flow may be another factor. It should also be
777 remembered that the wind was mostly from the east during the ICESCAPE II survey and this
778 would tend to decelerate the eastward flow on the shelf, which perhaps played a role.

779
780 The data presented here highlight the fact that there is a fast coastal pathway (which was
781 nearly void of winter water by the time the ICESCAPE surveys took place) and a set of slower
782 pathways on the interior shelf. This is potentially of importance for the distribution of nutrients
783 and the subsequent growth of phytoplankton on the shelf. As shown above, the winter water is
784 high in nitrate, and a long residence time means that such nutrients remain on the shelf as the ice
785 retreats and more sunlight penetrates the water column. Of note is the fact that there is a
786 convergence of several slow pathways in the region south of Hanna Shoal. This is the area where
787 enhanced numbers of walrus have been observed (Hannay et al., 2013). Prolonged primary
788 production and carbon export to the bottom is conducive for a rich benthic habitat which should
789 attract walrus. Similarly, our results imply that much of the water encircling Hanna Shoal drains
790 into the western side of Barrow Canyon, and this location is characterized by extraordinarily
791 high levels of benthic biomass (Grebmeier, 2012). A sustained supply of winter water to this
792 region could help support this "hot spot".

793

794 Finally, there are strong ramifications to the notion that brine rejection can further
795 transform the winter water flowing poleward in the central pathway. Salinization of winter water
796 on the Chukchi shelf has been documented previously for the coastal pathway (e.g. Weingartner
797 et al., 1998; Itoh et al., 2002). Our results suggest that this can happen for the water in the slower
798 pathways as well. The convective overturning that results from the brine rejection will likely
799 reach the bottom (since the Chukchi shelf is so shallow) which will stir regenerated nutrients
800 from the sediments into the water column (there is evidence of this; R. Pickart, unpublished
801 data). In turn, levels of primary production on the Chukchi shelf could be enhanced because the
802 water has such a long residence time (compared to the swift coastal current that advects the
803 nutrients quickly into the basin). It remains to be determined if such brine rejection in the central
804 pathway occurs on a regular basis.

805

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807

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	Mid-Shelf Jet		Coastal Jet		Reversed Flow		Shelfbreak Jet	
	Total	winter water	Total	winter water	Total	winter water	Total	winter water
Central Channel (CC)	.44	.22	.43	.08	-.02	-.01	–	–
Chukchi North (CN)	.60	.27	–	–	-.16	-.06	–	–
Chukchi Shelfbreak West (CSW)	.51	.24	–	–	-.31	-.13	.10	.06
Hanna Shoal North (HSN)	.22	.13	–	–	-.03	-.02	.28	.23
Chukchi Shelfbreak Central (CSC)	.26	.12	–	–	–	–	.15	.10
Chukchi Shelfbreak East (CSE)	.23	.08	–	–	–	–	.17	.04
Hanna Shoal Southeast (HSSE)	.11	.02	–	–	–	–	–	–
Barrow Canyon Head Extension (BCX)	.18	.01	–	–	–	–	–	–
Hanna Shoal South (HSS)	–	–	–	–	-1.9	-.2	–	–
Icy Cape (IC)	–	–	.44	–	–	–	–	–
Barrow Canyon Head (BCH)	.20	.03	.79	–	–	–	–	–

977

978 Table 1. Volume transport (in Sverdrups) for the different flow paths associated with the vertical sections
979 in the ICESCAPE II survey. Positive transports are poleward. For each section the total transport is
980 listed along with the transport of winter water. See text for details.

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986 **Figure Captions**

987

988 **Figure 1:** Schematic of the circulation of Pacific water in the Chukchi Sea, prior to this study.
989 Place names are listed along with the location of the Bering Strait mooring A3 (black star).

990

991 **Figure 2:** Locations of the shipboard measurements and moorings used in the study. See the
992 legends for details. The US/Russian convention line is plotted.

993

994 **Figure 3:** Wind vectors measured from *Healy* at each of the CTD station locations. The location
995 of the Pt. Barrow weather station is indicated by the star.

996

997 **Figure 4:** NARR fields averaged over the time period of the ICESCAPE II cruise, 25 June –29
998 July. The left-hand panels are sea level pressure (color and contours, mb) and 10 m wind vectors.
999 The right-hand panels are 10 m wind speed (color) and 10 m wind vectors. Top row: mean
1000 conditions for 2000-11. Middle row: Conditions during 2011. Bottom row: Anomaly (2011 –
1001 climatology). The black box in the lower right-hand panel marks the region over which the mean
1002 winds were computed (see the text and Figure 5).

1003

1004 **Figure 5:** Mean 10 m wind vectors for the Chukchi Sea from NARR (within the box shown in
1005 Figure 4) over the time period of the ICESCAPE II cruise, 25 June –29 July, shown for each year
1006 from 2000-11.

1007

1008 **Figure 6:** (a) Volumetric T/S diagram using the CTD data from the ICESCAPE II cruise. For
1009 each bin corresponding to 0.05°C in temperature and 0.05 in salinity, the number of observations
1010 (in log counts) is colored. The thick black lines mark the boundary between the water masses.
1011 WW = newly ventilated winter water; RWW = remnant winter water; BSW = Bering Summer
1012 Water; ESMW = early season melt water. The thin dashed line is the freezing line. (b) Scatter
1013 plot of the bottle data in T/S space where the nitrate concentration is colored ($\mu\text{mol/kg}$).

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1018 **Figure 7:** Vertical property sections for selected transects (indicated by red in the left-hand
1019 panels). The middle panels show potential temperature (color) overlain by salinity. The right-
1020 hand panels show absolute geostrophic velocity (color, where positive velocities are into the
1021 page) overlain by the -1.6°C isotherm (thick black line).

1022

1023 **Figure 8:** Average potential temperature (color) in the bottom 15 m of the water column at each
1024 station whose bottom depth is shallower than 85m. The vectors at the station sites are the
1025 transport per unit width of the winter water. The vectors at the mooring sites are the near-bottom
1026 velocities (see the legend).

1027

1028 **Figure 9:** (a) Winter water flow pathways deduced from the data. (b) Volume transports of the
1029 pathways. The red numbers are the total transport and the black numbers are the winter water
1030 transport. The station locations are marked by the dots.

1031

1032 **Figure 10:** Average salinity (color) of the winter water for the stations comprising the central
1033 pathway around the north side of Hanna Shoal. The vectors denote transport per unit width of the
1034 winter water (from Figure 8).

1035

1036 **Figure 11:** Average salinity and potential temperature of the winter water along the central
1037 pathway of Figure 10, plotted versus elapsed time from Bering Strait (see text for details).

1038

1039 **Figure 12:** Timeseries of temperature (top panel) and salinity (bottom panel) for mooring A3 in
1040 Bering Strait (see Figure 1 for the location of the mooring). The grey shading denotes the time
1041 period when the winter water, sampled at the sections in Figure 11 along the central pathway,
1042 passed through the strait.

1043

1044 **Figure 13:** Zoomed-in view of the timeseries of temperature (top panel) and salinity (bottom
1045 panel) from Figure 12. The green circles denote the properties of the winter water measured
1046 along the central pathway at the sections in Figure 11, adjusted for when the water passed
1047 through Bering Strait (see text for details).

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Figure 14: Time-latitude variation of sea ice concentration (%) from the AMSR-E along 167°W during March -July 2011. Also shown are the daily location of parcels starting at 66.3°N on March 20 (blue symbols), April 8 (magenta symbols) and April 20 (black symbols). Locations are shown for parcel velocities of 10 cm/s ('+'), 10.5 cm/s ('o') and 11 cm/s ('x').

Figure 15: AMSR-E sea ice concentration (%) on: (a) April 15, (b) May 1, (c) May 15, and (d) June 1 2011. Also shown are the locations of parcels starting at 66.3°N on March 20 (blue symbols), April 8 (magenta symbols), and April 20 (black symbols) and travelling along 167°W. Locations are shown for parcel velocities of 10 cm/s ('+'), 10.5 cm/s ('o') and 11 cm/s ('x').

Figure 16: Change in salinity along 167°W for parcels starting at 66.3°N and ending at 72°N as a function of start date in March/April 2011. Changes in salinity are shown for parcel velocities of 10 cm/s, 10.5 cm/s and 11 cm/s. Error bars indicate the change in salinity resulting from a +/- 10% change in ice cover along the trajectory of the parcels.

Figure 17: Contribution of the longwave radiative flux (blue curve) and turbulent heat flux (red curve) to the total change in salinity (black curve) along 167°W for parcels with a velocity of 10.5 cm/s starting at 66.3°N and ending at 72°N as a function of start date in March/April 2011.