

The Beaufort Gyre Fresh Water Reservoir: state and variability from observations

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Abstract

We investigate basin-scale mechanisms regulating anomalies in fresh water content (FWC) in the Beaufort Gyre (BG) of the Arctic Ocean using historical observations and data collected in 2003-2007. Specifically, the mean annual cycle, interannual and decadal FWC variability are explored. The major cause of the large FWC in the BG is the process of Ekman Pumping (EP) due to the Arctic High anticyclonic circulation centered in the BG. The mean seasonal cycle of liquid FWC is a result of interplay between the mechanical (EP) and thermal (ice transformations) factors and has two peaks. One peak occurs around June-July when the sea ice thickness reaches its minimum (maximum ice melt). The second maximum is observed in November – January when wind curl is strongest (maximum EP) and the salt input from the growing ice has not yet reached its maximum. Interannual changes in FWC during 2003-2007 are characterized by a strong positive trend in the region varying by location with a maximum of approximately 170 cm per year in the center of EP influenced region. Decadal FWC variability in the period 1950 – 2000 is dominated by a significant change in the 1990s forced by an atmospheric

circulation regime change. The center of maximum FWC shifted to the southeast and appeared to contract in area relative to the pre-1990's climatology. In spite of the areal reduction, the spatially-integrated FWC increased by over 1,000 km³ relative to climatology.

1. Introduction

The state of the Arctic Ocean and its influence on the global climate system strongly depend on the fresh water budget. Changes in the fresh water balance influence the extent of the sea-ice cover, the surface albedo, the energy balance, the temperature and salinity structure of the water masses, and biological processes. The first reasonably-comprehensive estimates of the Arctic Ocean salt and heat balances were published by *Timofeev* [1960] and later by *Mosby* [1962]. Both authors considered all possible sources and sinks of fresh water and attempted to balance them using the relatively sparse observational information available at that time. Unfortunately, there remains substantial uncertainty in current estimates of Arctic Ocean FWC and its seasonal and interannual variability. The significance of this statement, which has repeatedly concluded all earlier studies of the Arctic Ocean fresh water balance [*Timofeev*, 1960; *Mosby*, 1962; *Coachman and Aagaard*, 1974; *Aagaard and Greisman*, 1975; *Aagaard and Carmack*, 1989; hereafter *A&C*] and recent [*Carmack*, 2000; *Serreze et al.*, 2006; *White et al.*, 2007; *Carmack et al.*, 2008], is increasing with societal demand to explain the rapidly changing Arctic climate.

Investigations of the fresh water balance are closely related to the “Toward an ice-free Arctic Ocean” theme that has been considered and discussed by generations of Arctic scientists [e.g. *Zubov*, 1943; *Budyko*, 1969; *Aagaard and Coachman*, 1975; *Stroeve et al.*, 2007]. Such studies have pointed out that the volume of fresh water stored in the Arctic Ocean and its vertical distribution (ice thickness and concentration, cold oceanic halocline thickness and strength) are regulated by atmospheric dynamics and

thermodynamics, fresh water sources and sinks, and ocean circulation and mixing processes - the key processes influencing Arctic climate.

Substantial international interest in Arctic Ocean FWC developed in the 1970s, driven by the Former Soviet Union plans to divert Siberian Rivers for additional irrigation of Kazakhstan and Central Asia. A series of publications [e.g. *Aagaard and Coachman*, 1975; *Semtner*, 1987] discussed potential changes in Arctic conditions that might arise as a result of the expected reductions in river runoff, and the consequential reduction in the strength of the oceanic cold halocline and vertical stability of the upper ocean. Further studies highlighted the impact that fresh water exported from the Arctic Ocean has on deep-water formation at subpolar latitude and in turn, the intensity of the Atlantic Ocean Meridional Overturning Circulation (MOC) [*Aagaard et al.*, 1985; *Lazier*, 1980; *Clarke and Gascard*, 1983; *Talley and McCartney*, 1982; *Dickson et al.*, 1988; *Hakkinen*, 1993, 1999; *Curry and Mauritzen*, 2005; *Rennermalm et al.*, 2006].

Significant changes in the fresh water flux from the Arctic Ocean were documented by *Dickson et al.* [1988]; most notable was the “Great Salinity Anomaly”, GSA. The GSA appeared in the Greenland Sea in 1968 and circulated around the subpolar ocean for at least 15 years. Various explanations for the GSA have been offered [*Dickson et al.*, 1988; *Walsh and Chapman*, 1990; *Mysak et al.*, 1990; *Hakkinen*, 1993, 1999]. *Proshutinsky and Johnson* [1997] (hereinafter *P&J*) gave evidence that such salinity anomalies may originate inside the Arctic Ocean due to a redistribution of fresh water. *Belkin et al.* [1998] revisited the GSA of the 1970s and documented the existence of two additional subpolar salinity anomalies in the 1980s and 1990s.

Other interesting results were obtained from modeling studies by *Delworth et al.* [1993, 1997] who pointed out that salinity anomalies in the Arctic Ocean surface layer precede anomalies of the MOC intensity by 10-15 years. *Hakkinen* [1993] examined the cause of salinity anomalies using a 3-D coupled ice-ocean model and concluded that strong northerly winds over the Greenland Sea and Fram Strait area lead to an increase in the thick ice transport from the Arctic Ocean and to increased fresh water discharge. While analyses of observational data by *Wohlleben and Weaver* [1995], *Walsh and Chapman* [1990], and *Mysak and Venegas* [1998] supported Hakkinen's model results, the possible contribution of liquid fresh water to the flux increase was not considered.

Another development in the investigation of the Arctic Ocean fresh water budget and its variability began in the 1990s following a decision by the Arctic Ocean Sciences Board (AOSB) in July 1996 to organize a “Symposium on the Freshwater Balance of the Arctic”. These activities resulted in the NATO advanced research workshop “The Freshwater Budget of the Arctic Ocean”, Tallinn, Estonia, 27 April - 1 May, 1998. The workshop proceedings [*Lewis*, 2000] outlined the state and variability of the Arctic Ocean fresh water budget including a traditional investigation of the fresh water budget of the Arctic Ocean (i.e. detailed analysis of its major components including river runoff, inflow from the Atlantic and Pacific oceans, outflows through Fram Strait and the Canadian Archipelago, the atmospheric moisture flux, and the annual cycle of ice formation and melt). A comprehensive inventory of the fresh water budget of the Arctic Ocean was also presented by *Carmack* [2000]. The report from a later meeting of the AOSB (<http://www.aosb.org/mtgs/XVII.html>) included recommendations to (1) study shelf processes contributing to the formation of sea ice and of dense bottom water; (2)

study the outflow of fresh water and ice through the Canadian Archipelago; (3) study the impact of a reduction or cut-off of Pacific inflow; and (4) continue to monitor the total FW exchange with the North Atlantic (NA), including the export of ice. Most of these recommendations have been implemented by the international ASOF program (Arctic - Subarctic Ocean Fluxes: *Dickson et al.*, 2008) that has as its major objectives to measure and model variability of fluxes between the Arctic Ocean and the Atlantic Ocean and to understand the role of these fluxes in decadal climate variability. The recent results of *Serreze et al.* [2006] and *White et al.* [2007] refine some of the numbers in the accepted mean Arctic Ocean fresh water budget, but they did not fully resolve its seasonal, interannual and decadal variability, nor identify mechanisms responsible for changes.

The general approach to fresh water studies by the AOSB, ASOF and other investigators assumes *a priori* that the inflow and outflow of fresh water are balanced at any given time. Little attention has been given to the mechanisms of fresh water storage in the Arctic Ocean and its variability. Our approach to the FWC problem, formulated in *Proshutinsky et al.* [2002] (hereafter *P2002*), allows for imbalance between the numerous sources and sinks of fresh water in the Arctic Ocean. That is, we allow that the Arctic Ocean can accumulate a significant volume of fresh water during one climate regime and later release it to the Greenland, Iceland, and Norwegian Seas and NA during another climate state or circulation regime. This tenet has allowed us to formulate several hypotheses to explain salinity anomalies in the NA and decadal variability of the Arctic climate [*P2002*; *Dukhovskoy et al.*, 2004, 2006].

The largest freshwater storage area of the Arctic Ocean is located in the Beaufort Gyre: identified by a regional salinity minimum at depths of 5–400 m (Figure 1, see also

P2002; Proshutinsky *et al.*, 2009). Existence of the BG fresh water reservoir was inferred in *P2002* via an analysis of salinity fields, wind, and ice drift patterns. *P2002* hypothesized that in winter, the wind (a dynamic factor) drives the ice and ocean in a clockwise sense and the BG accumulates fresh water mechanically through a deformation of the salinity field (Ekman convergence and associated downwelling). In summer, winds are weaker (and may even reverse to be counterclockwise). The resultant anomaly in Ekman convergence releases fresh water, relaxes salinity gradients and reduces the BG FWC. *P2002* tested this mechanical hypothesis for fresh water accumulation and release by employing a relatively simple model where wind was the major driving force (the influences of sea ice and ocean thermodynamics were neglected). They further pointed out that thermodynamical processes may also be important - in winter, ice growth and subsequent salt release reduce the (liquid) FWC of the BG, and in summer, ice melt increases the FWC. The interplay between dynamic- and thermodynamic forcing is clearly complicated. Another factor likely to be important and not considered by *P2002* is the change in FWC associated with the advection of surface, Pacific, and Atlantic waters into the BG.

In this paper, we attempt to address questions and test hypotheses formulated in *P2002* associated with BG FWC variability at seasonal to decadal time scales. We investigate basin-scale mechanisms regulating FWC in the BG area using measurements from the 2000s obtained by the BG Exploration Project and by drifting Ice-Tethered Profilers (ITPs), historical observations of 1950s-1980s from the Joint U.S. – Russian Atlas of the Arctic Ocean [*EWG*, 1997, 1998], and data collected in the 1990s by different national and international expeditions. Section 2 describes the data sources and

our methodology of FWC estimation from the historical gridded and bottle data, and from CTD profiles (shipboard, moored profilers and profiles from drifting buoys). FWC variability on seasonal, interannual and decadal time scales is analyzed in Sections 3 and 4. Major findings are summarized in Section 5.

2. Study area, methods and data sources

To facilitate temporal comparisons of BG integral parameters, we define the Beaufort Gyre Region (BGR) in terms of the climatologic location of the gyre as bounded by 70.5° N to 80.5° N and 130° W to 170° W (Figure 1). FWC calculations were carried out for this region at grid points having total depths greater than 300 meters.

2.1 Fresh water content calculations

Fresh Water Content (liquid), FWCL (in meters) in the ocean is calculated as

$$\text{FWCL} = \int_{z_2}^{z_1} \frac{[S_{ref} - S(z)]}{S_{ref}} dz \quad (1),$$

where the z axis is defined as positive up with the surface $z=0$. The reference salinity, S_{ref} is taken as 34.8; $S(z)$ is the salinity of the water at depth z . We take z_2 as the depth level where $S(z)=S_{ref}$ while z_1 defines the upper level of the FWCL integrations. For total water column FWCL, $z_1=0$. Change in FWCL is thus a measure of how much liquid fresh water has accumulated or been lost from the ocean column bounded by the 34.8 isohaline and $z = z_1$. (See *A&C*, 1989; *Carmack*, 2000; *Carmack et al.*, 2008 for details on the significance of the reference salinity in the analysis of FWCL).

The FWC in sea ice (FWCI or solid) is formally given by equation (2)

$$FWCI = \int_{zice}^0 \frac{[Sref - Sice]}{Sref} dz + \int_0^{fb} \frac{[Sref - Sice]}{Sref} dz \quad (2).$$

Here, $zice$ is the draft of the sea ice that we measured using Upward-Looking Sonars (ULSs); fb is the freeboard of the sea ice. The freeboard can be measured by satellites but in this paper we calculate sea ice thickness (Hi , which is $zice+fb$) from the sea ice draft as $Hi=zice(\rho_w/\rho_i)$, where ρ_i and ρ_w are ice and water densities, respectively. Assuming that sea ice salinity ($Sice$) is vertically uniform in the BGR, $Sice=1.8$ according to *Perovich et al.*, [2009] the FWCI is calculated as

$$FWCI = 0.948Hi \quad (3).$$

For snow FWC calculations we follow the recommendations of *Sturm et al.* [2002] who analyzed the evolution and spatial distribution of the snow cover on the Arctic sea ice during the Surface Heat Budget of the Arctic Ocean (SHEBA) project. Specifically we assume that the snow has no salt and that the snow water equivalent, $SWE = 0.343Hsn$ where Hsn is snow thickness.

2.2 Interpolation technique

Optimal interpolation (OI) was used to construct spatial maps (on a 50-km square grid) of the freshwater distribution in summer for the decades of the 1990s and 2000s and each year from 2003 through 2007 using all available CTD stations and ITP profiles, between the last week of June and first week of October (inclusive). The decadal interpolations ignore interannual variability and combine all available data for the whole decade from the summer months. Sufficient spatial coverage of observations since 2003 allowed optimal interpolations to be constructed for each year up through 2007. The

optimal interpolation method improves on earlier methods used to calculate the same quantities [e.g. *Carmack et al.*, 2008], because it uses statistics from the data to minimize the mean square error of the interpolated estimates, and produces an objective estimate of the expected error. The OI method [e.g. *Gandin*, 1965; *Bretherton et al.*, 1975; *Reynolds and Smith*, 1993; and *Emery and Thomson*, 1997] is based on the assumptions that the statistics of the data being mapped are stationary, homogeneous, and isotropic. Values are calculated at grid points from the averages weighted by the distance from each observation. The weights are proportional to the covariability of the data which in our application was calculated from the observations and could be approximately represented by a Gaussian function with length scale of 300 km.

In addition to the raw data locations, desired grid locations, and covariance length scale, the method requires a starting reference background field and estimates of the error in the raw measurements (0.2 m is allowed in our OI generated maps) and variance of the observations for determining expected error estimates. The OI mapping procedure is applied to anomalies from the reference field and the result is added back to the reference to produce the final map. Specifying a realistic reference field is particularly important to the results away from the raw data locations, because at such sites, the OI technique returns estimates that approach the reference field values. Climatological estimates are often used as the reference field in geophysical applications, but could not be used here since preliminary analysis indicated that there was significant change in the freshwater distribution in the BGR in the 2000s from the best available climatology [*EWG*, 1997, 1998]. Furthermore, our region was only sparsely sampled in the decade preceding the first year of the BG program. Consequently, in this application it was necessary to

construct a best guess reference field estimate for the 1990s and for the first year of the BG program (2003) from the data themselves (supplemented with observations from 2002 to enhance spatial coverage). A polynomial surface fitting algorithm (as in *Carmack et al.* [2008]) was used to construct the base map for 2003 using raw data from 2002-2003. In subsequent years, the calculated OI map of FWC from the previous year was used as the base map.

A map of the multiyear mean FWC distribution during the Beaufort Gyre Observational System (BGOS, <http://www.whoi.edu/beaufortgyre>) time period was obtained as the average of the annual maps from 2003-2007; error for the 5 year time period was computed from the OI determined mean error for all years. Estimates of the interannual trend in FWC were derived from linear regressions at each grid cell of the same annual maps.

2.2 Data sources

2.2.1. Ocean and sea ice data from the Beaufort Gyre Observing System (2003-2007)

Beginning in August 2003, time series measurements of temperature, salinity, currents, geochemical tracers, sea ice draft, and sea level have been acquired in the BGR by a team of Woods Hole Oceanographic Institution investigators in collaboration with scientists from Fisheries and Oceans Canada, Institute of Ocean Sciences and the Japan Agency for Marine-Earth Science and Technology under the JWACS (Joint Western Arctic Climate Studies) and BGEP (Beaufort Gyre Exploration Project supported by the U.S. National Science Foundation) activities consisting of a series of BGOS expeditions

(Figure 2). Observations of ocean temperature and salinity (T&S) were obtained from moorings, drifting buoys, and shipboard instruments. Standard hydrographic sections consisting of fixed sites revisited annually (August-September) have allowed investigation of interannual variability of measured parameters in the BGR during summer. The mooring data allow estimation of sea level variability, sea ice draft and variations in the vertical distribution of FWC at three (and later four) locations throughout the year (Figure 2). In addition, eight Ice-Tethered Profilers (ITPs, Figure 2) measuring T&S daily at depths ranging between ~7 and 750 meters have returned year-round information from the BGR in real time via satellite (see *Krishfield et al.* [2008] and ITP project web site <http://www.who.edu/itp> for more details).

Time series of fresh water content at fixed location spanning the entire annual cycle are derived from salinity profiles acquired by BGOS bottom-anchored moored profiler instrumentation (McLane Moored Profiler, MMP). MMP data acquisition followed a burst sampling scheme consisting of two one-way profiles separated in time by 6 hours initiated every 48 hours. To avoid collision with ice keels, the tops of the moorings were positioned at or deeper than 45 m below the surface. On moorings A and C in 2003, the MMPs profiled between 95 m and 2000 m, while for all other moorings and years the MMPs profiled between 65m and 2000 m. Three of the BGOS moorings (A, B, and C) have been continuously maintained since 2003; mooring D was first installed in 2005. As documented by *Ostrom et al.* [2004], MMP salinity data are believed to have an uncertainty of less than 0.005; resultant errors in FWCL estimates (for depth intervals sampled by the MMPs) are less than 0.1 m.

As above, the high-accuracy shipboard salinity measurements result in FWCL uncertainties for individual profiles of less than 0.1 m. Our calculations of interpolation errors for the 2003-2007 hydrographic surveys estimate that the expected error varies from 0.1 m to 1.5 m depending on the distance a given grid point is from that year's observations.

Eight ITP instruments returned approximately 7200 T&S profiles from the BGR in the 2004-2007 time period (Figure 2). Typically, each system occupied two vertical profiles per day between 7 m and 750 m depth. After calibration, the accuracies of the ITP salinity data are believed to be comparable to those from the ship-based and bottom-moored instruments [Krishfield *et al.*, 2008]. In order to construct FWCL estimates integrated to $z=0$, a method was needed to extrapolate upward the MMP and ITP profiles. A scaling procedure was developed, described in the Appendix, to reconstruct FWC in the upper ocean. Because the ITPs typically sampled from 760m and up to 7m depth, uncertainty in the ITP full-water-column FWC estimates from vertical extrapolation is relatively small (no worse than 0.50 m in summer, decreasing to zero in winter when the ITPs sampled up into the surface mixed layer). In turn, uncertainty in the reconstructed mooring FWC estimates (0.50 m) was quantified by referencing the CTD and ITP data collected near the moorings (see the Appendix for details). Lastly, in order to differentiate temporal and spatial variability in the drifting ITP observations, we developed an FWC anomaly estimation procedure relative to a reference spatial map. The technical issues are also discussed in the Appendix.

The expected error of our sea ice FWC calculations derive from errors in the sea ice draft measurements ($\sim 0.10\text{m}$), errors associated with sea ice salinity (± 1) and errors

related to sea ice density $\pm 20 \text{ kg/m}^3$ [Perovich *et al.*, 2009]. Fortuitously, the errors in sea ice density and salinity compensate each other in the sea ice FWC calculations; we assume that the total error in FWCI is $\pm 0.10 \text{ m}$.

The expected FWC errors due to snow water equivalent calculations are comparable with errors in sea ice FWC computations. For the entire winter, the mean snow depth at the SHEBA site was 0.337 m, giving a mean snow water equivalent of 0.116 m [Sturm *et al.*, 2002]. This snow was accumulated until end of May and then started melting. Observations of snow depth in the BGR by the Ice Mass Balance Buoys [Richter-Menge *et al.*, 2006] show that the mean snow depth in 2003-2004 was 0.26 m with snow water equivalent of 0.089 m and the snow typically started melting in early June. Taking into account that sea ice draft changes recorded by ULSs include changes due to snow on the sea ice surface ($\sim 0.10 \text{ m}$ in water equivalent during seasonal cycle) we assume that the snow contribution to uncertainty in our FWC estimates is encompassed by the estimated error in sea ice thickness.

2.2.2. Ocean data from 1990s

Hydrographic T&S surveys from the 1990s and early 2000s used in this study were obtained from Fisheries and Oceans Canada, Institute of Ocean Sciences, the Japan Agency for Marine-Earth Science and Technology, and the National Snow and Ice Data Center and National Oceanographic Data Center, including data from the SCICEX (Scientific Ice Expeditions, 1993-1998 that occupied repeat XCTD transects) and SHEBA (1997-1998) expeditions.

In calculations of decadal-averaged FWC for the 1990s when CTD and XCTD data were used, there are instances where the available T&S profiles do not extend fully to the ocean surface. For example, the SCICEX XCTDs typically only sampled below 25 m depth. In these cases, the shallowest salinity value in a profile was extended uniformly to the surface. Given that the typical wintertime mixed layer depth in the BG is at least 25 m deep, this extrapolation results in a very small error in FWC (< 0.20 m). However in summer when mixed layers are thinner, this extrapolation may result in an underestimation of FWC by as much as 1.0 m (quantified using 1-m-resolution shipboard CTD data). The accuracy of calibrated CTD salinity data from this decade is estimated to be better than 0.01, which translates to a total FWC anomaly uncertainty of less than 0.10 m. However, transient surface variations and other uncertainties inherent in the data from the shallowest sampled layer more likely limit the overall accuracy of an FWC estimate from any single profile to 0.50 m. As indicated earlier, we use 0.20 m to quantify the accuracy of the measurements for the producing the optimal grids.

2.2.3 Ocean T&S climatology

Historical summer T&S gridded data for this region are taken from the Environmental Working Group Atlas [*EWG*, 1998]. These *EWG* gridded fields are used to calculate liquid FWC distributions for the decades from the 1950s through the 1980s. It should be noted that the historical data are most extensive during the 1970s [*EWG*, 1997, 1998], reflecting mostly winter conditions when Soviet aircraft could land on the ice. *EWG* Atlas data are based primarily on bottle data and are provided at standard levels after smoothing and interpolation. The accuracy of the salinity measurements from

the bottle data is estimated as 0.02. The errors in FWC associated with measurement accuracies and use of measurements at standard levels is less than 0.50 m. These errors were estimated by comparing FWC estimates based on CTD data with 1-m vertical resolution to FWCs calculated from standard *EWG* levels. The *EWG* atlas contains interpolated fields along with some information about interpolation errors which for salinity are reported to vary between 0.02 to 0.05. Assuming a worst-case scenario that these errors exist in all layers of the ocean, the maximum error in FWC due to interpolation will be between 0.20 to 0.45 m.

2.2.4 Atmosphere and satellite-based sea ice data

Complementing the *in situ* measurement programs, remote sensing projects and atmospheric reanalysis data are also utilized in the present analysis. Sea ice concentration data are taken from *Comiso* [1999, updated] and sea ice drift parameters from *Fowler* [2003, updated], both available from the National Snow and Ice Data Center (NSIDC). Sea level pressure (SLP) data from the National Center for Atmospheric Research and National Centers for Environmental Predictions (NCAR/NCEP) reanalysis are also used, along with geostrophic winds and surface wind stresses calculated from SLP to characterize the atmospheric forcing.

3. Beaufort Gyre Region fresh water content interannual and seasonal variability in 2000s

3.1. Observational results from moorings

Investigation of the interannual and seasonal variability in solid (ice) and liquid (ocean) FWC is important to understand the mechanisms responsible for the BG fresh water accumulation and release processes, their timing and scales of changes. Historical observations are limited in the BGR that is covered with sea ice year-round, making it impossible to reconstruct the record of interannual and seasonal variability of FWC there before the 2000s. But recently, high-temporal-resolution time series are available from the four BG moorings and ITPs.

Using the BGOS mooring data (MMP profiles) time series of liquid FWC, salinity and water density at moorings A-D for the period 2003-2007 were constructed (Figure 3). The data were smoothed using a 15-day running-mean procedure although there are still instances where eddy structures influence the FWC estimates (e.g. at mooring A in October 2005). At moorings B and C, the FWC exhibits a positive trend from fall 2003 through 2007 superimposed on seasonal and higher-frequency signals. At mooring A, the FWC has a positive trend for the period 2003-2005. Thereafter, FWC remained roughly constant or decreased (no MMP data were collected at this mooring during the 2006-2007 deployment). Mooring D data span only 2 years, limiting the ability to detect interannual trends. They do, however, indicate that for the 2005-2007 period there is an overall decrease in FWC below 65m at that location. Mooring D also shows a very well pronounced seasonal cycle with a peak in FWC in November-December and absolute minima in September-October. The other mooring records also document a FWCL seasonal maxima typically occurring sometime between October and January but the interannual variability is significant.

The FWCL variability reconstructed for the entire water column (see Appendix for details) shows much larger changes (Figure 4) with significant positive trends in FWCL at all moorings ranging from 64 cm/year at mooring D to 178 cm/year at mooring B.

The range of sea ice thickness changes (obtained from ULS records) varies among moorings. The seasonal magnitude of sea ice draft (Figure 4) is approximately 150 cm. The maximum ice draft/thickness in the BGR is observed between May and June and the minimum between late August and early November depending on the year and mooring location.

In order to analyze seasonal changes in the FWC we de-trended the time series of liquid and solid FWC components and constructed an average seasonal cycle for each mooring location (Figure 5, top and middle panels). In general, we observe two peaks in the liquid FWC anomaly during a seasonal cycle, one maximum in December – January and a second, smaller one (less pronounced at mooring A) is typically observed in July-August, depending on mooring location. We can also identify two minima in the liquid FWC anomalies: the first one is typically observed in April – May and the second in September – October. Note however that there is significant interannual variability in the liquid FWC anomalies during the 2000s (Figure 5).

The sea ice contributions to the liquid FWC variability are well pronounced during a seasonal cycle. Typically, the sea ice starts releasing liquid fresh water into the ocean in the middle of May, and the volume of fresh water from the ice reaches a maximum in the ocean in September-October. Notable however, there is more going on

in the BGR FWC seasonal cycle than simple phase change as sizable anomalies in total FWC (ice plus liquid) are observed (Figure 5 bottom panels).

3.2. Causes of seasonal FWC changes in 2003-2007

The seasonal cycle of liquid FWC revealed above is not fully consistent with the conventional understanding of the Arctic Ocean hydrological cycle [*EWG*, 1997, 1998; *Steele et al.*, 2001, *Serreze et al.*, 2006, see their Figure 11) in which maximum liquid FWC occurs in August and the minimum in April. This understanding was based on an idealized concept of the Arctic without sea ice drift, ocean currents, and mixing. In this simplified case, the seasonal cycle of sea ice growth and melt due to the seasonal changes in solar radiation results in ocean FWC variability influenced only by changes in air temperature (see Figure 1 in *Overland* [2009]). Our results indicate that there are two minima and two maxima in the mean seasonal cycle of BG liquid FWC and only the April minimum coincides with the “climatologic” view point.

Under pure thermodynamic forcing (i.e., without dynamic processes and mixing) the Arctic Ocean would quickly become a two-layer ocean with a completely fresh upper layer and a salty lower layer (see the description of this distillation process by *A&C*). In the real ocean, the seasonal variability in local liquid FWC (*FWCL*) is not only influenced by changes in sea ice seasonal transformations (*FWCI*) under the solar radiation seasonal cycle, but is also forced by changes in Ekman pumping (EP) (*FWCE*), ocean mixing (*FWCM*) and by changes in fresh water sources and sinks (*FWCSS*). The

latter includes precipitation, evaporation, river runoff and fresh water fluxes via ocean straits. We may then express $FWCL(t)$ (where t is time) as the sum:

$$FWCL(t) = FWCI(t) + FWCE(t) + FWCM(t) + FWCSS(t) \quad (4).$$

Based on (4) we speculate that the Arctic Ocean FWCL seasonal cycle should differ significantly from region to region, complicating assessments and intercomparisons. For example, in coastal regions influenced most directly by river runoff, the maximum FWCL is observed in the middle of July (following maximum river runoff in June) and the minimum is observed at the end of April (minimum river runoff and maximum land-fast ice thickness). Our observations in the BGR indicate that the seasonal fresh water anomalies associated with changes in sea ice thickness ($FWCI$) follow closely the seasonal anomalies of air temperature (Figure 5, middle panel) but do not explain fully the observed seasonal FWCL cycle (Figure 5, top panels). The influence of fresh water sources and sinks ($FWCSS$) in the formation of summer positive FWCL anomalies in the BGR is questionable because the maximum river runoff and the maximum inflow of the fresh water from the Bering Sea are observed in June – July [Woodgate *et al.*, 2005] and these signals cannot reach the central part of the Beaufort Gyre by August. The seasonal FWC anomalies in precipitation are mostly associated with snow (rain is very rare in the BGR). The information about changes in snow depth and snow water equivalent is already taken into account by our sea ice draft measurements as discussed above. Ocean mixing ($FWCM$) is an important factor which is able to preserve fresh water anomalies in the region. It depends on many oceanic internal and external parameters. The majority of them are associated with wind forcing (or sea ice stresses at

the ocean surface) and the ocean vertical stratification. For this study we assume that *FWCM* can be included in the term of EP which is also a function of wind stress and ocean stratification.

The EP forcing, which influences fresh water accumulation and release [*Hunkins and Whitehead*, 1992; *P2002*] in the BGR, was evaluated using time series of wind curl averaged over the BGR or in the vicinity of each mooring. The 2003-2007 NCAR/NCEP Reanalysis 1 wind curl estimates were de-trended and a mean seasonal cycle was constructed for each mooring location. The climatologic annual-mean wind stress curl over the BGR is negative (not shown, see *P2002*, *Yang and Comiso* [2007], *Yang* [2009]) and the EP is positive leading to the accumulation of FWC in the BG. In the typical seasonal cycle (Figure 5, bottom panel, blue lines) the wind stress curl anomaly usually reaches maximum in July – August (minimum EP) and minimum in December – January (maximum EP); and the cumulative wind stress curl (solid blue line) indicates that the accumulation of fresh water in the BGR usually occurs between November and June (negative wind stress curl anomaly). Starting in June – July and until October – November the wind stress curl does not support fresh water build up in the region and the BG can even release the fresh water because of the reduction or reversal of EP.

On the other hand, the role of EP could be estimated from equation (4). Knowing anomalies of liquid (*FWCL*) and solid (*FWCI*) FWCs and assuming that other factors (except EP) are already taken into account in *FWCE* or *FWCI* or are insignificant, the anomalies in EP can be expressed as a residual:

$$FWCE(t) = FWCL(t) - FWCI(t) \quad (5).$$

These anomalies in liquid *FWCL* due to EP were calculated for each mooring and are shown in Figure 5 (bottom panels, red line). One sees that these anomalies correlate well with the wind stress curl anomalies.

Based on these results we surmise that the second maximum in BG *FWC* observed in summer (June – August, depending on mooring) is a result of interplay between EP and fresh water changes due to sea ice seasonal transformations. The *FWCL* decreases from January to June – July due to reduction of EP and due to sea ice formation and salt release. At this time both factors work together towards a reduction of *FWCL*. In June-July, sea ice starts melting and *FWCL* increases while EP is weakening or relatively stable and does not play an important role. In August, EP reaches its minimum or even becomes negative (i.e. forces upwelling in the central BGR resulting in relaxation of horizontal density/salinity gradients, weakening the BG anticyclonic circulation). Sea ice at this time melts but the rate of melting is decreasing. As a result, the EP effect (*FWCE*) dominates changes in *FWCI* (due to ice melt) and *FWCL* in the region reaches its second minimum. In September, both factors contribute to the increase of *FWCL*: EP increases and sea ice continues melting but in October EP is the major influence on *FWCL* while sea ice starts growing reducing *FWCL*.

These simple considerations demonstrate that the departure of the BG *FWCL* seasonal cycle from an ideal ice/ocean state is related to the wind forcing. Respectively, the *FWCL* maximum observed in November – January (2-3 months later than expected from thermodynamic sea ice transformations) is forced by the atmospheric anticyclonic circulation cell over the BGR.

For a complete study of the seasonal variability of *FWCL*, it is important to analyze the potential for mixing between the upper and deep ocean layers as a mechanism that preserves fresh water in the BG. For example, Figure 6 demonstrates how the evolution of salinity and FWC in the BG can be influenced by: (i) upper ocean transformations due to ice melt and growth, (ii) EP, and (iii) mixing. While surface salinity is more influenced by sea ice transformation and reaches a maximum in mid-January due to ice growth and salt release, salinity at deeper levels is controlled more by EP and exhibits minimum values in mid-January.

In September and October the upper ocean layer is relatively fresh and the surface mixed layer quite thin (Figure 6). With time the surface layer deepens, entraining deeper layers and by November the mixed layer depth reaches 30 meters. Gradually, the salinity of the surface layer increases due to ice growth and salt release, mixing with deeper layers. In July, the upper ocean layer again receives fresh water from sea ice melt and the vertical ocean stratification increases. In addition, due to Ekman divergence salinity anomalies generated by winter EP have to relax. In an ideal ocean without mixing, the fresh water content of the area remains the same after completing the EP seasonal cycle. In the real ocean, we speculate that because of mixing and entrainment of deeper layers the ocean can accumulate fresh water or salt and preserve it for long periods. Such preserved anomalies are strongly related to the mean anticyclonic baroclinic BG circulation.

3.3 Causes of interannual FWC changes in 2003-2007

The small positive trends observed in 2003-2007 in the sea ice draft (Figure 4) apparently are the result of sea ice redistribution under wind forcing and are not influenced by the thermal regime, because the 2003-2007 air temperature linear trend averaged over the BGR is positive (0.72 °C per year, not shown). Therefore, sea ice melt cannot contribute to the *FWCL* increase in the region at a rate of more than 100 cm/year (Figure 4, top panel). Geochemical water analysis in the BGR [Carmack *et al.*, 2008, Alkire *et al.*, 2008; Guay *et al.*, 2009] shows that the net effect of sea-ice formation and melting removes ~30% of the freshwater contributed to this region by other sources. Therefore the only candidate responsible for this trend is EP, if we follow our conclusions about the mechanisms of *FWCL* changes in the BGR. Indeed, in 2003-2007 there was a significant trend in wind stress curl (Figure 7) with an unprecedented (for 1948-2008 period, see analysis of decadal changes in Section 4) minimum of wind stress curl in 2007 when anticyclonic circulation dominated over all months including summer.

In order to analyze interannual changes in the BGR *FWCL* volume we have used results from BG summer observations at repeat CTD and XCTD stations in conjunction with all other available mooring and drifter data from the cruise times (Figure 2). The *FWCL* was calculated for all years employing eq. (1) for each T&S profile available in the region at this time. Then calculated FWCs were gridded using optimal analysis (Section 2.2). While during 2003-2004 *FWCL* in the BGR showed little change (Figure 8), in 2005 the total *FWCL* in the region began to increase, reaching 20,000 km³ in 2007. The maximum annual trend of *FWCL* accumulation was observed at the new center of the BG (which was first established in the 1990s, see Section 4) with a rate of approximately 1.5 m per year in the center and tapering away with increasing radius (Figure 9).

Preliminary data from the BGOS 2008 cruise indicate that the *FWCL* in the BG continued to rise in 2008 and reached 21,000 km³ – a historical maximum from all available years of observations. Compared to 1970s climatology (the pre-90s decade with the most extensive data coverage, (Figure 1) there has been a *FWCL* increase in the BG of approximately 5,000 km³. This is comparable with the volume of fresh water annually delivered to the Arctic Ocean by rivers and through Bering Strait (5700 km³ per year, *Serreze et al.*, [2006]).

Both the mooring (Figure 4) and ITP data (Figure 10) demonstrate that from 2003 to 2007 the absolute maximum of FWC was located in the southeast region of the BG (near the location of mooring D). The maximum rate of *FWCL* increase over this time period (178 cm/year, Figure 4) is observed at mooring B; somewhat smaller rates are observed at moorings A and C. Summer CTD data further indicate that this trend could be as high as 140-150 cm/year in places (Figure 9). The regional differences in trends can be explained by regional differences in the EP (atmospheric circulation). Figure 11 shows annual SLPs and geostrophic winds for 2003-2007 where centers of the atmospheric circulation (max/min EP) and their intensity can be easily identified. Based on Figure 11, the largest EP should be observed in 2004, 2005 and 2007 which is in agreement with wind stress curl data shown in Figure 7.

Moorings B and C have no overall *FWCL* increase between 2003 and 2004 or even a small reduction (Figure 4). This could be due to a reduced annual wind curl in 2003 (Figure 7; note also the monthly wind curl is positive in June, July, August, and October) with a low gradient SLP field (Figure 11). In 2004-2005, SLP gradients intensified and the absolute wind curl over the region (Figure 7) increased by

approximately 30% relative to the previous year and the region influenced by anticyclonic winds extended over the entire Canada Basin (Figure 11). In 2004, the center of circulation and maximum EP was very close to mooring B, accounting for the significant increase of FWC at this mooring by December of 2004 (Figure 4, Figure 3b). Moorings A and C also show accumulation of fresh water during this period but at a lower rate (Figure 3a, Figure 3c and Figure 4). In 2005-2006, the SLP gradients in the region relaxed, absolute wind curl decreased by approximately 35% and the fresh water accumulation at all moorings fell accordingly. Another jump in the FWC was observed in 2007 (Figure 4, Figure 3, Figure 7) when an anomalous anticyclonic circulation regime dominated the Arctic Basin for the entire year (Figure 11) and the annual wind curl reached its absolute minima in the 2000s (Figure 7).

4. Decadal shift

Based on our analysis of the decadal distributions of *FWCL* in the BGR (Figure 12), decadal distributions of SLP fields (Figure 13) and decadal changes of wind stress curl (Figure 14), we conclude that the anticyclonic atmospheric circulation dominated over the BG since at least the 1950s enabling fresh water accumulation in this region of the Arctic Ocean (via negative wind curl and consequently positive EP). The rate of fresh water accumulation changes from decade to decade (Figure 12) and the center of the FWC maximum also changes depending on the atmospheric conditions (e.g. location of the Arctic High, Figure 13). We speculate that fresh water accumulation/release can occur preferentially from different regions when the BG expands/contracts and surface waters have differing areal coverage of the negative wind curl. Indeed, the historical hydrographic data indicate that *FWCL* in the BGR did not change significantly from

decade to decade in the 1950-80s period (Figure 12) which is in agreement with decadal SLP fields and position of the Arctic High. Wind stress curl (Figure 14) also did not change significantly during the 1960-1980s. Note that the information about wind stress curl (Figure 14) and information about location of the Arctic High center have to be analyzed together to explain changes in *FWCL* and its center of maximum. For example, the wind stress curl values over the region in the 1960s and 1970s are practically identical but the center of the Arctic High shifted westward in the 1970s (Figure 13) and the center of the BG maximum of *FWCL* also shifted to the new position. It is also important to remember that the *FWCL* distribution for the 1970s is more accurate than for other pre-1990s decades because the hydrographic data coverage in the 1970s was relatively good (see *EWG* [1997, 1998]) and therefore the *FWCL* from the 1970s may be used as a climatologic benchmark characterizing all pre-1990s conditions.

The SLP (Figure 13) and *FWCL* (Figure 12) conditions in the 1990s differed significantly from the pre-1990s climatologic conditions: the maximum of the *FWCL* distribution shifted to the southeast and increased in magnitude. The spatially-integrated BG *FWCL* for the 1990's exceeded the climatologic mean for the 1950-1980 by approximately 1,400 km³; lateral gradients of *FWCL* increased as well. Thus, in the 1990s the fresh water reservoir of the 1950-1980s with its center at 76°N and 150°W “released” its fresh water which was relocated into a new center with coordinates 73.5°N and 143°W. We also conclude that the major cause of this relocation was the change in the atmospheric circulation, intensity and center of the maximum EP. While the EP in the BG “traditional” region shown in Figure 1 decreased, the center of EP shifted south- and east-ward (Figure 13) and reorganized the *FWCL* fields respectively. As a result the BG

freshwater reservoir contracted but intensified. Unfortunately, the hydrographic data from the 1990s are very sparse in the eastern half of the BGR so that we are unable to positively determine the BG *FWCL* maximum center precisely in the 1990s. But the information from the BG western half is good and indicates that there was a substantial decrease in *FWCL* in this region relative to pre-1990s (Figure 12), confirming at least partially our conclusion about *FWCL* center relocation in the 1990s.

In the 2000s, the “decadal” (based on 2000-2007 data) volume of liquid fresh water in the BG is comparable to what it was in the 1990s. This is because some data from 2000-2002 showed some reduction in the BGR *FWCL*. The *FWCL* increase in 2003-2007 is consistent with wind stress curl dynamics (Figure 14) and the SLP distribution for the 2000s (Figure 13). Except for 2000-2002, in the 2000s, the Arctic High was perfectly located over the BGR, centered over mooring B (where maximum trend in *FWCL* accumulation was observed) and the wind stress curl reached its historical minimum (maximum EP, Figure 14).

Summer time series of FWC in different vertical layers of the BG, including sea ice, shows that the maximum change occurred in the mixed layer (as expected) and that in the 1990s the FWC in the mixed layer increased by 2,400 km³ (Table) relative to the 1980s. The volumes of fresh water in summer and winter Pacific Water decreased. Volumetric changes of fresh water in the Atlantic Water layer for the same period are small relative to the accuracy of the estimates (150 km³). Small increase in *FWCL* in the mixed layer and reduction in the Pacific layer relative to 1990s were observed in 2000s (Table).

Our explanations for decadal variability of FWC in the BGR are speculative, but it is clear that the BG *FWCL* field in the 2000s is not in steady state despite the more than 10 years of prevailing anticyclonic atmospheric circulation (1997-2007). There are strong increasing trends in BG FWC in 2003-2007 observed at the mooring locations, along ITP trajectories and at the standard summer shipboard BGOS CTD station sites. We speculate and provide some observational evidence that there are three major causes of these changes, namely: (a) wind-generated EP which represents the mechanical part of the ocean fresh water redistribution, (b) mixing and changes of water stratification, and (c) sea ice transformations accompanied by fresh water and salt release. More quantitative analysis of these processes has to be done by means of numerical modeling: a subject for future studies.

5. Conclusions

Summarizing the major results of this study we conclude the following.

1. The Beaufort Gyre in the southern half of the Canada Basin is the largest fresh water reservoir of the Arctic Ocean. It contains approximately 20,000 km³ of liquid and approximately 800 km³ of solid (sea ice) fresh water (based on a mean ice thickness in the region of 2.0m). Decadal, interannual and seasonal changes in total FWC influence the extent of sea-ice cover, the surface albedo, the energy balance, the temperature and salinity structure of the water masses, and biological processes in the entire Arctic Ocean. Moreover, the intensity of the Atlantic

Ocean Meridional Overturning Circulation can be influenced significantly by fresh water fluxes from the Arctic Ocean. Our results demonstrate that the seasonal changes in BG total FWC range from 5% to 10% of its total volume depending on the intensity and sense of atmospheric circulation. Interannual changes based on the data from 2003-2008 can reach as much as 1000 km³ per year. The decadal change range is approximately 1000 km³ per decade prior to 2000, but has accelerated since 2003. This estimate of decadal-timescale variability is likely an a lower bound since it is based on climatologic fields for calendar decades (1950s, 1960s, 1970s, 1980s, 1990s, and 2000s) which do not coincide with the timing of climatic circulation regimes (cyclonic and anticyclonic, *P&J, P2002*). We speculate based on our calculation of wind stress curl (Figure 14, bottom panel) that the differences in *FWCL* between cyclonic and anticyclonic circulation regimes are significantly larger than the decadal changes recorded in the *EWG* [1997, 1998] atlas data.

2. The BG *FWCL* did not change significantly from decade to decade in the 1950-1980s period. However, the 1990s conditions differ significantly from the pre-1990s climatologic conditions: the center of the FWC maximum shifted to the southwest and appeared to contract in area. In spite of the areal reduction, the spatially integrated BG FWC increased by approximately 1,400 km³ relative to the earlier climatology, while lateral gradients of FWC increased. We speculate that under these conditions, the BG geostrophic circulation spun-up and the baroclinic part of the Transpolar Drift current intensified and shifted toward Canada, accelerating sea ice drift and redistributing sea ice in this region.

3. We speculate and provide some observational evidence that there are three major causes of the FWC changes observed in the BG at different time scales, namely:
 - (a) wind-generated EP which represents the mechanical part of the ocean fresh water redistribution, (b) mixing and changes of water stratification, and (c) sea ice transformations accompanied by the fresh water and salt release.
4. The major cause of the large *FWCL* in the BGR is the process of EP generated by the climatologic anticyclonic atmospheric circulation over the Canada Basin with a center in the Beaufort Gyre. Our observational data analysis confirms this hypothesis formulated in *P2002*. Mechanically, the seasonal variability of *FWCL* follows wind curl changes with a maximum in November – January and a minimum in June-August depending on changes in atmospheric circulation. The atmospheric and oceanic thermal regimes regulate seasonal transformations of *FWCL* due to the seasonal cycle of sea ice melt and growth. The seasonal *FWCL* in the Beaufort Gyre has two pronounced peaks shifted by approximately 6 months. The first (small) peak (June-August depending on mooring) is observed when the sea ice thickness reaches its minimum (maximum fresh water release from sea ice to the ocean) and when EP is very close to its minimum (maximum wind curl). The second maximum (large) is observed in November-January when wind curl reaches its minimum (maximum EP) and the salt flux from the growing sea ice has not reached its maximum.
5. The conventional practice for measuring the Arctic Ocean hydrography (sampling in fall when the Arctic is accessible by research icebreakers when sea ice coverage is minimum and by aircraft in spring when sea ice is sufficiently strong

6. The interannual changes in the BG were estimated based on 2003-2008 observations of the BG hydrography at standard sections repeated annually, and also using moorings and ITPs which provided year-round T&S high vertical resolution data. Both the mooring and ITP data demonstrate that the BG FWC field is in transition; the processes of the FWC accumulation and release are not in balance. There are strong FWC positive trends in the 2003-2007 pentade at the mooring sites, along ITP trajectories and at the standard BGOS summer shipboard CTD measurement sites. The maximum FWC trend of 178 cm/year for this period was observed at 78°N and 150°W while the FWC increased at a smaller rate at moorings A, C, and D. Preliminary analysis of our 2008 data shows that the FWC in the BGR is continuing to increase.

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APPENDIX: Reconstruction of mooring and ITP FWC data to take into account the upper ocean layers

In order to investigate the full scale FWC variability (including the upper layers of the ocean which were not covered by observations during the entire year) at each mooring location, we have reconstructed these data by a simple regression of FWC calculated from the shipboard CTDs, including those taken at each mooring site during deployment and recovery operations. In this procedure, we used CTD data to determine a ratio between FWC calculated for the entire water column (from surface to salinity 34.8 level) and for the column from 65m or 95m down to the 34.8 salinity level. These summer coefficients for 2003-2007 are relatively stable, both spatially and temporally (Figure A.1).

Since 2004, significant amount of T&S data have become available from Ice-Tethered Profilers (ITPs: <http://www.whoi.edu/itp>). ITP profiles cover depths between 7 and 760 m, capturing the surface mixed layer in winter but missing the thin, very fresh waters just below the ice that can occur in summer. In order to better reconstruct the seasonal cycle of FWC change at moorings, we have made calculations based on ITP data to investigate how the FWC ratio changes during the course of a seasonal cycle. The FWC in the upper 8m (shallower than the top of the ITP profile) were calculated assuming that the salinity at 8 m extends to the ocean surface. This approach works very well for at least 9 months of the year when the upper mixed layer extends below 8m. During ice melt season (June-August) this procedure underestimates the amount of FWC in the water column by approximately 0.15-0.20m. This is the difference between FWC calculated by multiplying the FWC below 8m by a ratio coefficient between FWC

calculated for the entire water column (from the surface to salinity 34.8 level) and FWC calculated for the column from 8m down to the 34.8 salinity level using summer CTD data. This ratio is very stable and is around 1.1 and results in the underestimation of FWC from ITPs by approximately 10% of the total FWC in the water column in summer (not shown).

ITP FWC anomalies for Figure 10 are calculated relative to a reference surface based on the 2003 to 2006 summer BGOS CTD stations. The ITP FWC is computed as described in the text above. Unlike the other maps presented in this paper, a polynomial surface was fit to the CTD station data to produce the gridded surface at 50 km spacing (as in *Carmack et al.* [2008]). The anomalies at each time are the ITP FWC value subtracted from the corresponding grid value.

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Table Fresh water content ($\times 10^3 \text{ km}^3$) in the Beaufort Gyre Region relative to salinity 34.8.

Decade/Layer	1970s	1980s	1990s	2000s
<i>Mixed Layer ($S < 31$)</i>	4.2	4.9	7.3	7.5
<i>Summer Pacific ($31 < S < 33$)</i>	8.5	8.0	7.1	6.6
<i>Winter Pacific ($33 < S < 34$)</i>	2.7	2.7	2.2	2.1
<i>Deeper Layers ($S > 34$)</i>	1.1	1.1	1.1	1.1
Total	16.5	16.6	17.7	17.3

Figure Captions

Figure 1 Climatology of fresh water content in the Arctic Basin (shown in colors). Solid lines depict mean 1950-1980 salinity at 50m. Fresh water content is calculated relative to salinity 34.8 based on 1950-1980 data from EWG [1998] averaged for all decades. The Beaufort Gyre Region (BGR) is bounded by thick dashed blue lines.

Figure 2 The composition of the Beaufort Gyre Observing System (BGOS) by year. Large yellow circles show mooring locations (A, B, C, and D indicated); small white circles show locations of CTD stations and white crosses depict locations of XCTD castings; colored lines depict ITP trajectories; Background colors show bathymetry from the International Bathymetric Chart of the Arctic Ocean (IBCAO, Jacobson et al. [2000]). Vectors show mean 1979-2000 ice drift pattern.

Figure 3a Top panel: 15-day running mean FWC (in m) time series below 65m (solid line) and 95 m (dotted line) from MMP at BGOS mooring A (75°N, 150°W) from summer 2003 to summer 2006. Bottom panel: 15-day running mean seawater salinity (solid lines) and density (dashed lines) contours versus depth.

Figure 3b Same as in Figure 3a but for mooring B (78°N, 150°W)

Figure 3c Same as in Figure 3a but for mooring C (77°N, 140°W)

Figure 3d Same as in Figure 3a but for mooring D (74°N, 140°W)

Figure 4 FWCL time series (in m) from MMPs using derived FWCL coefficients (top panel), and time series of sea ice draft (in m) from ULSs (bottom panel) from BGOS moorings for 2003-2007. FWCL trend lines in the top panel (dotted) are calculated from the corrected FWC MMP data. Ice draft trends are also shown in the bottom panel as dashed black lines. Note that the ice draft data are offset from one another by 2 meters.

Figure 5a Mean annual seasonal cycles (red thick lines) of liquid (upper panels) and solid (middle panels) FWC component anomalies (in meters) at mooring A (left panels) and mooring B (right panels). Thin colored lines show seasonal changes for particular years (2003 – deep blue, 2004-2005 – light blue, 2005-2006 – light green, 2007 – dark

green). Middle panel: Dotted and thick blue lines show anomalies of monthly air temperature ($^{\circ}\text{C}$) and sum of degree days ($^{\circ}\text{C}$) averaged for 2003-2007 over the BGR. Note that the air temperature and sum of degree days are scaled with a factor of 0.025 to be compared with FWC values. Bottom panel shows anomalies of FWC due to Ekman pumping (red). Blue dotted and solid lines depict seasonal anomalies of wind stress curl and sums of wind stress curl (similar to sums of degree days) averaged over a 300x300 km box centered at each mooring site.

Figure 5b Same as in Figure 5a but for moorings C (left panel) and mooring D (right panel).

Figure 6 ITP-5 60-day running mean time series of water salinity at levels 10 db through 100 db with 5 db interval. Thick lines show 25 db (25 m) and 50 db (50 m) levels. In September and October the upper ocean layer is fresh. With time this layer mixes with deeper layers and by November the mixed layer depth reaches 30 meters and, gradually its salinity increases due to mixing with deeper layers and due to ice growth, convection and Ekman pumping. In July, the upper ocean layer gets fresh water from ice melting and the vertical ocean stratification increases.

Figure 7 Wind curl time series over the BGR: annual (top), monthly (middle) and integrated in time (bottom).

Figure 8 Late summer FWCL distribution characteristics. Colors and contours show BGR FWCL (in m) relative to salinity 34.8. Numbers in bottom left corners of each panel show regional summer FWC volume (in km³) for each year. Black dots show locations of CTD data used for the analysis. Grey colored pixels depict areas where FWCL error of interpolation exceeds 2m.

Figure 9 FWCL trend (right panel, meters per year) for 2003-2007 in the BGR based on results of optimal interpolation. Left panel shows standard deviations (m).

Figure 10 FWCL content anomalies (in m) calculated based on ITP data relative to 2003-2006 mean FWCL distribution for the region (numerals in inset). Colored lines show different ITP FWCL time series. The black solid line depicts the overall linear trend calculated using all ITP data (193 cm/year). The inset shows ITP trajectories and the mean gridded values.

Figure 11 Annual SLP (hPa, solid contours and colors) and geostrophic wind direction (vectors) in 2003-2007. A, B, C, and D depict mooring locations and black lines bound the BG Region. Bottom white panel shows SLP and wind direction averaged for 2003-

2007. It is expected that largest Ekman pumping should be observed in 2004, 2005 and 2007 which is in agreement with Figure 7.

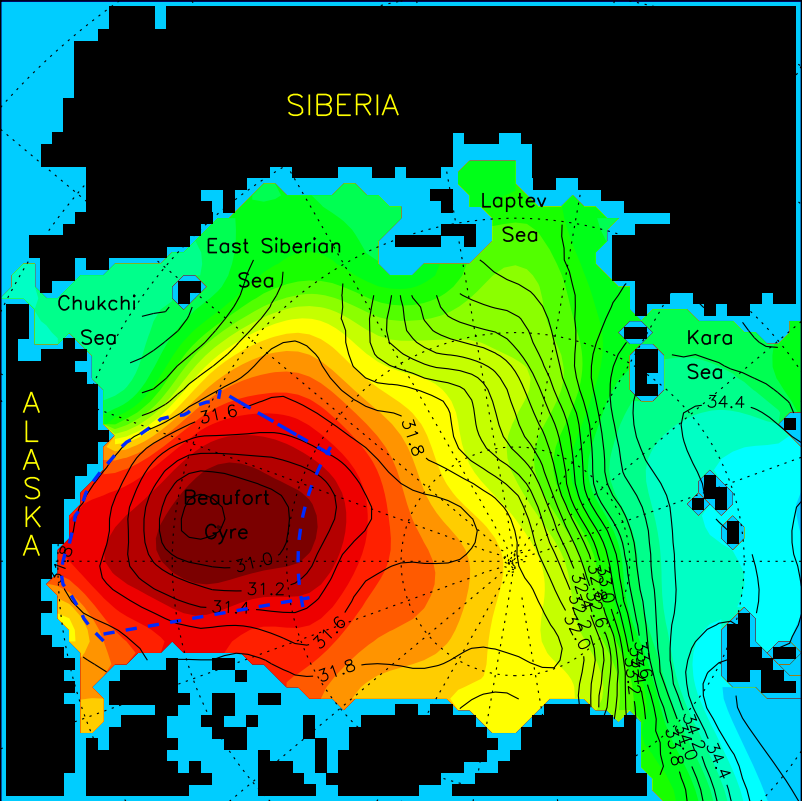
Figure 12 Contour plots of decadal summer FWCL (in m) in the BG relative to salinity 34.8. The summer FWCL volume estimates for the BG region (in 1000 km³) for each decade are indicated in the bottom left corners of each panel. Data for 1950-1980 are from summer EWG gridded fields. The 1990 and 2000 data are optimally interpolated fields from observations acquired during the summer period for each decade. Grey shading indicates expected error in the interpolation is greater than 2 m. Black dots show locations of BGOS mooring locations for reference.

Figure 13 Annual SLP (hPa, solid contours and colors) and geostrophic wind direction (vectors) in 1950s through 2000s. A, B, C, and D depict mooring locations and black lines bound the BG Region. In 1990s, the Arctic High shifted southeast and this explains relocation of the center of FWCL accumulation observed in 1990s (Figure 12).

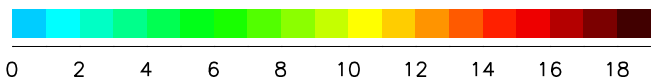
Figure 14 Wind stress curl ($\times 10^4 \text{ gcm}^{-2}\text{s}^{-2}$) averaged for decades (top) and circulation regimes (bottom) over the BG Region. Bottom panel: periods of anticyclonic circulation regimes are shown as blue bars and cyclonic regimes as red ones.

Figure A.1 FWCL coefficients (ratio of FWCL from sea surface to the level of salinity 34.8 and FWCL from 65 m to the level of salinity 34.8) calculated from summer CTD surveys for each year from 2003-2007. Right bottom figure shows the 2003-2007 mean.

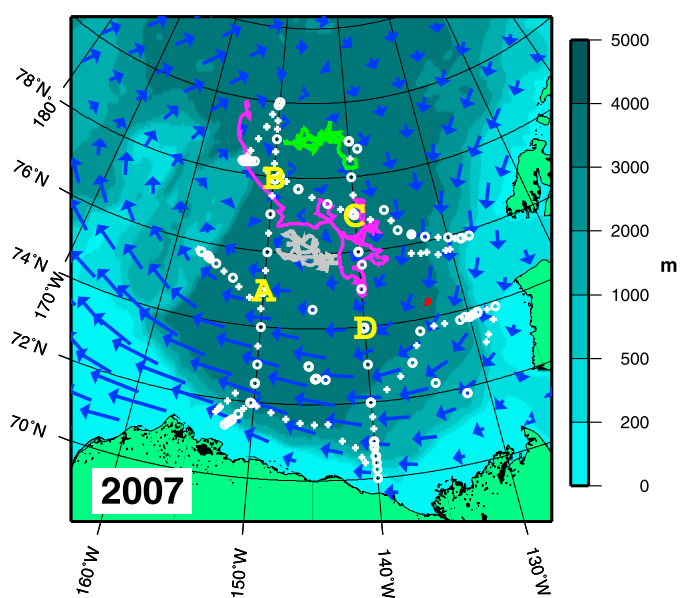
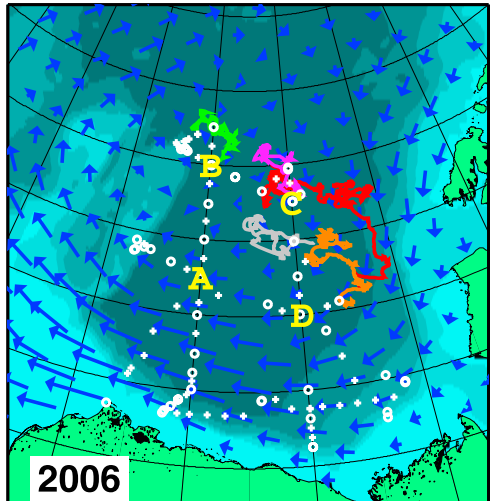
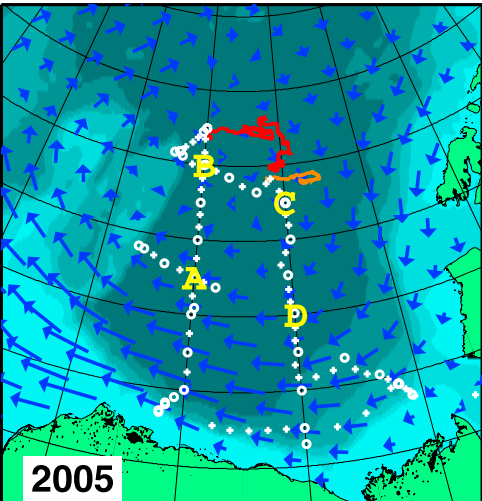
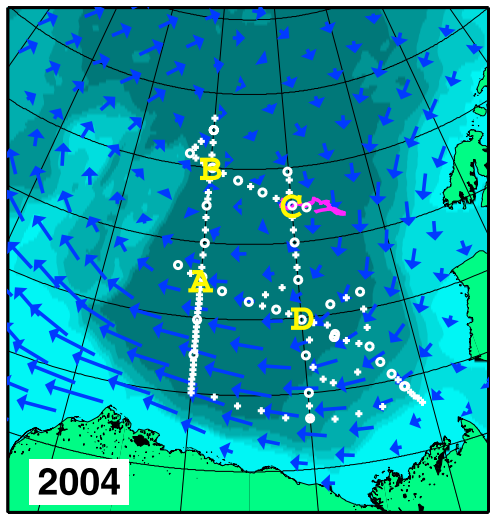
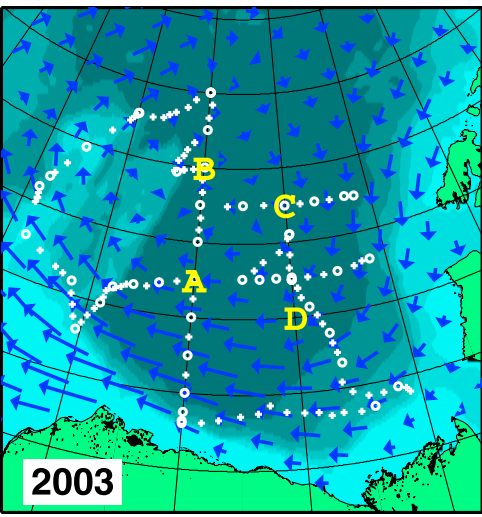
Open blue circles depict locations of BGOS moorings A (75°N , 150°W), B (78°N , 150°W), C (77°N , 140°W) and D (74°N , 140°W).



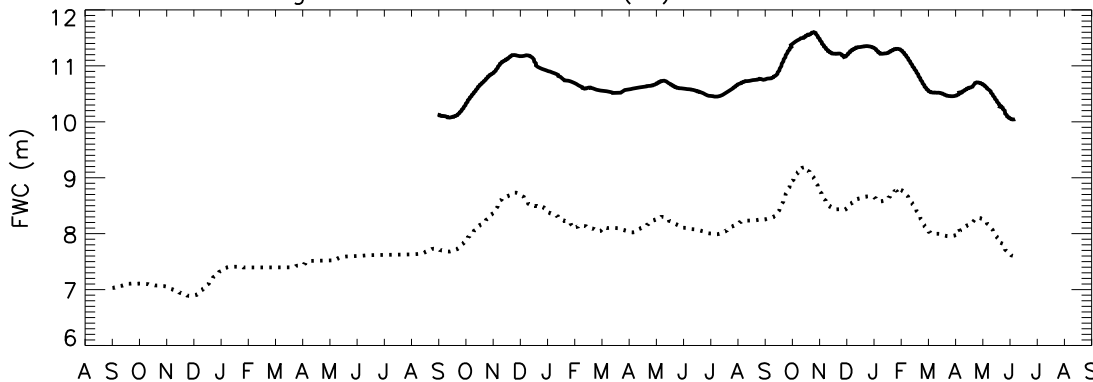
Freshwater content, meters



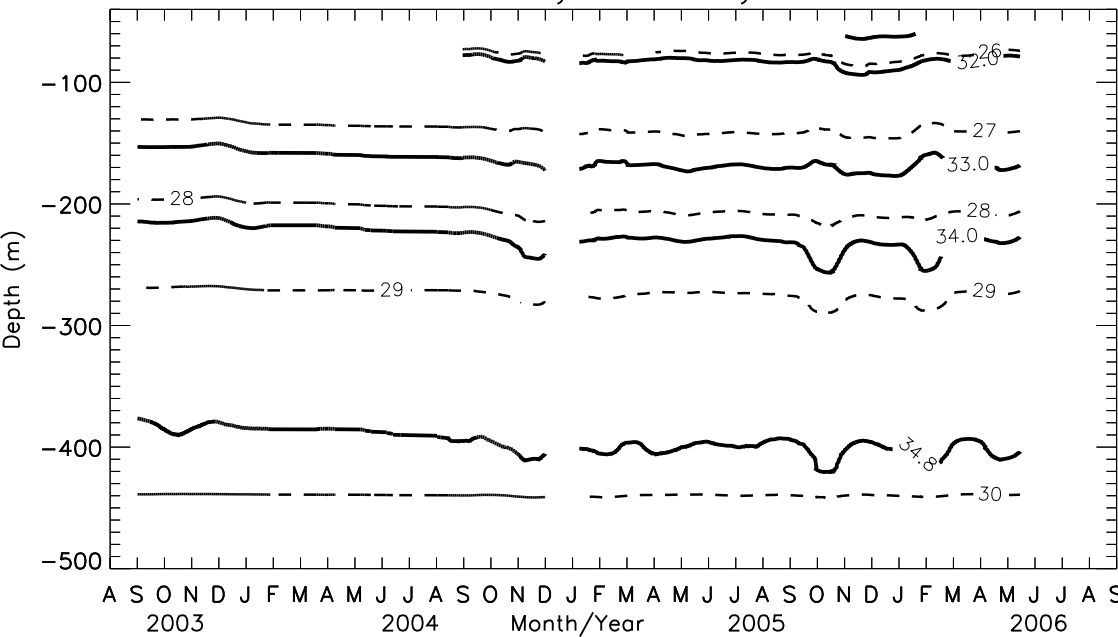
Beaufort Gyre Observing System



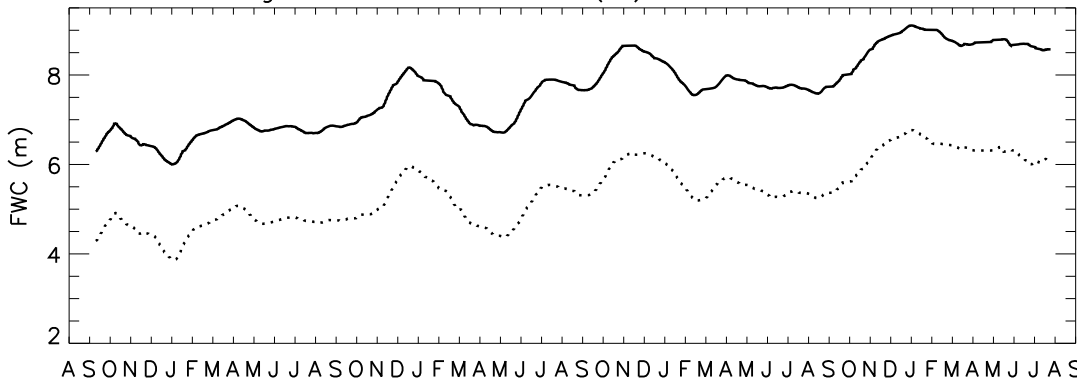
Mooring A: 2003–2006 FWC (m) below 65 and 95 m



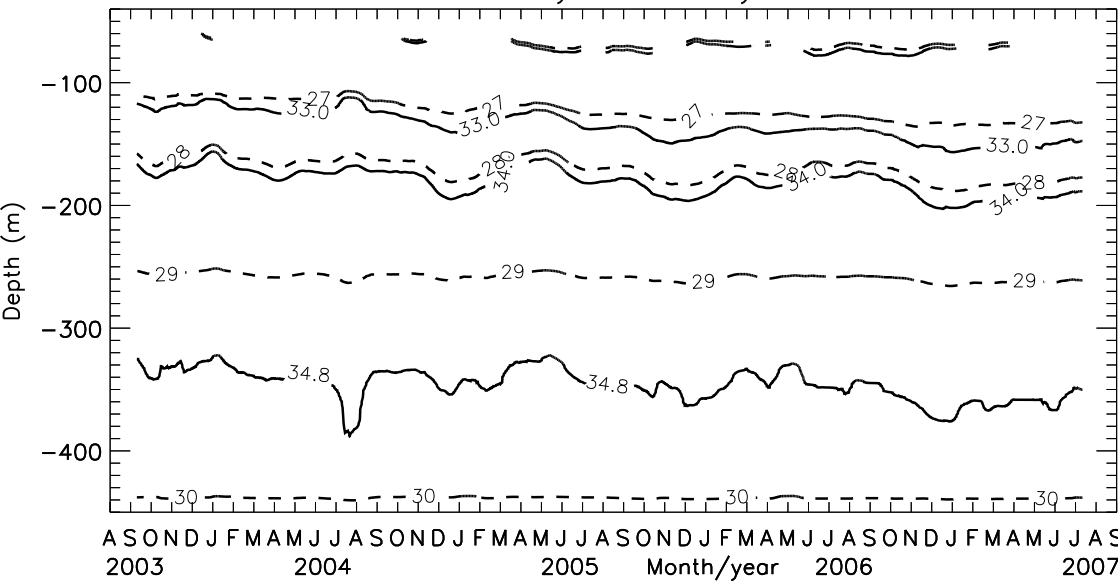
Salinity and Density



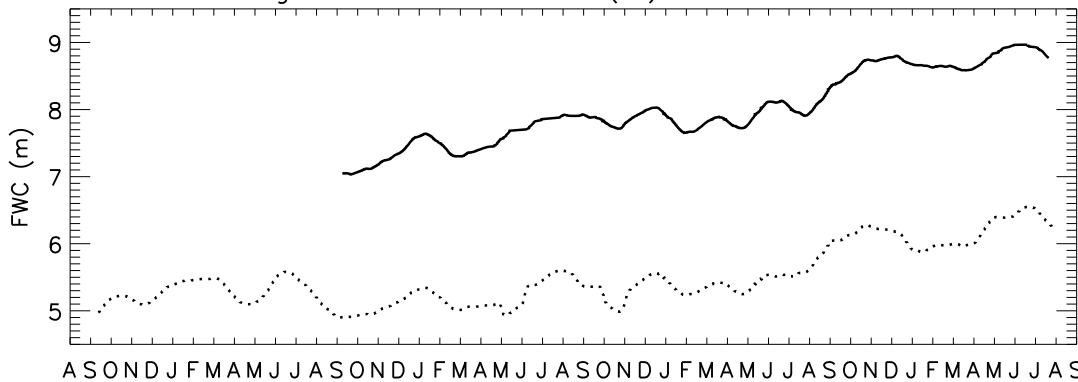
Mooring B: 2003–2007 FWC (m) below 65m and 95 m



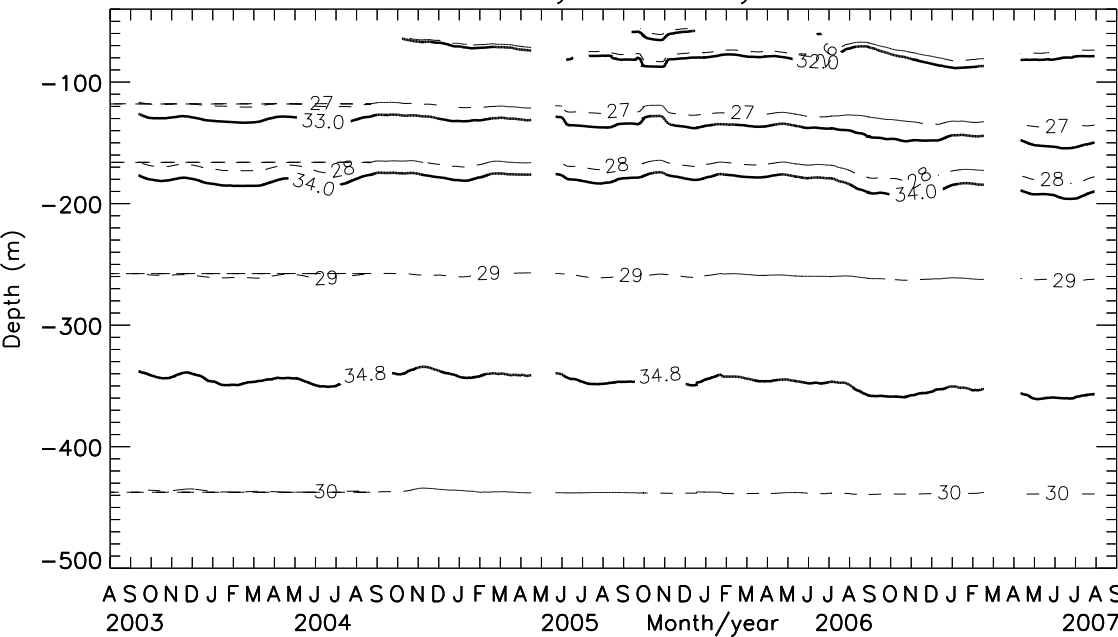
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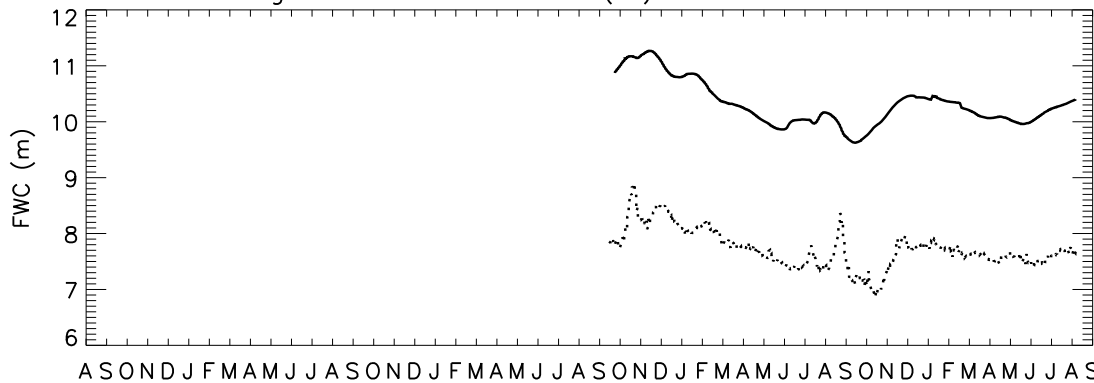
Mooring C: 2003–2007 FWC (m) below 65 and 95 m



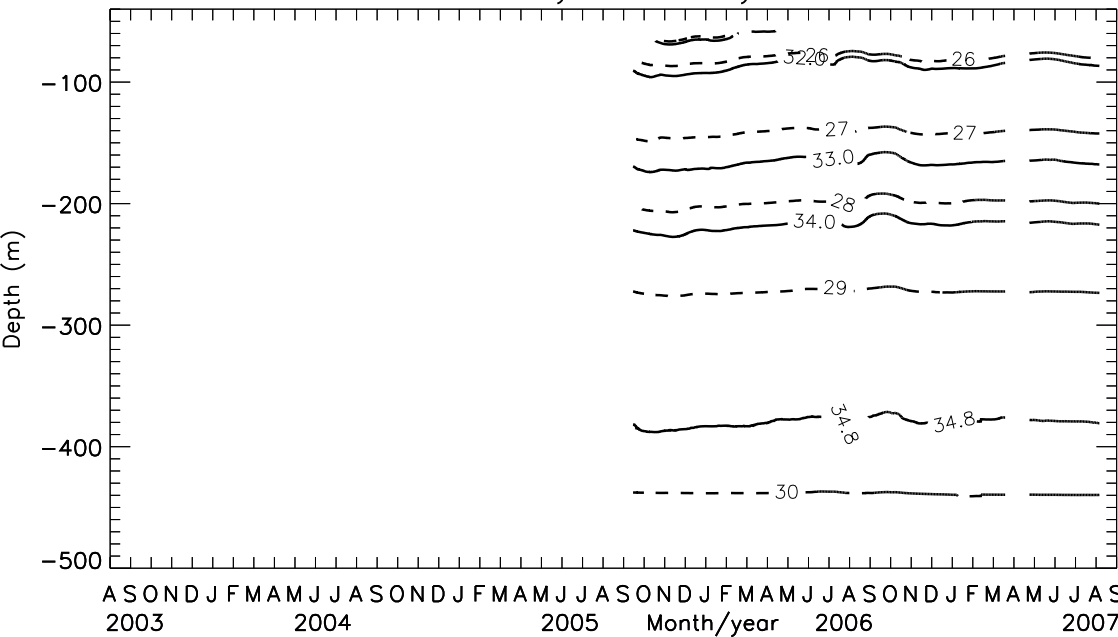
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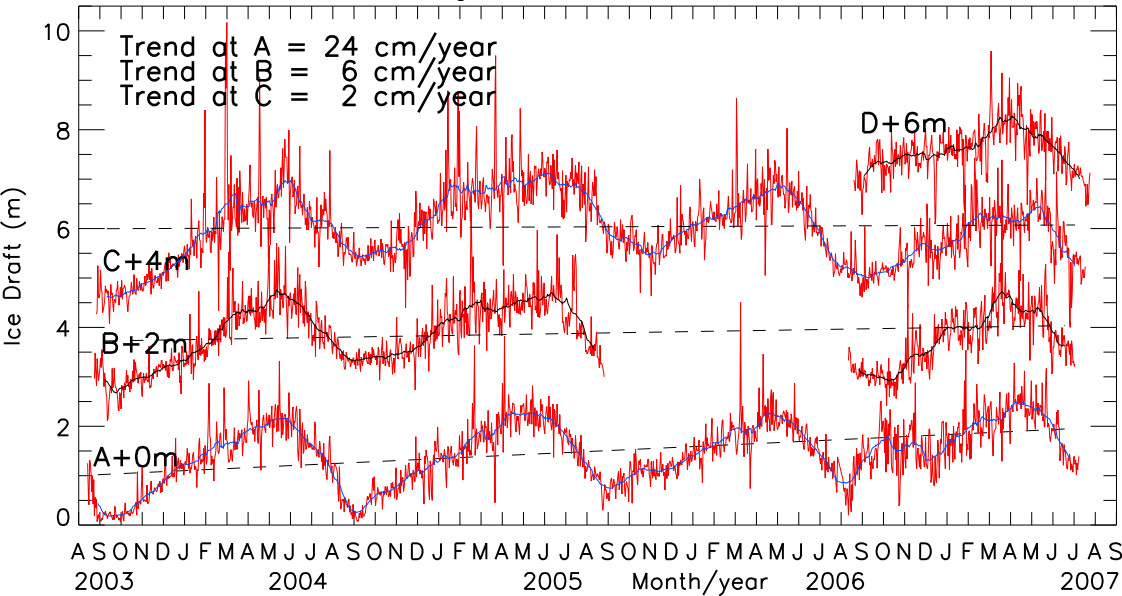
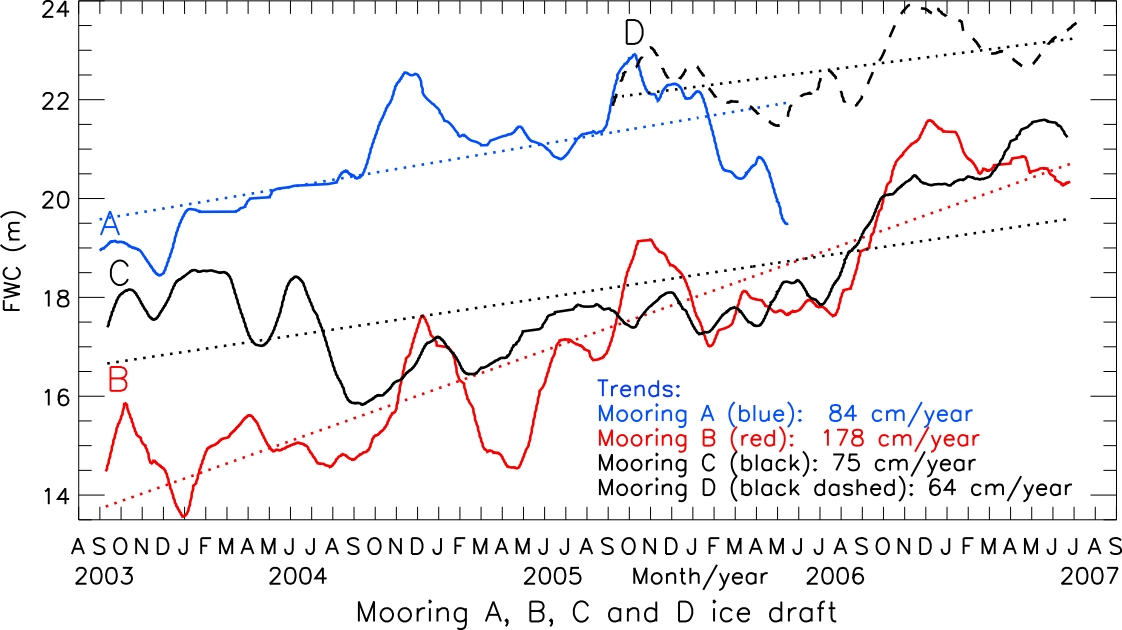


Mooring D: 2003–2007 FWC (m) below 65 m and 95 m

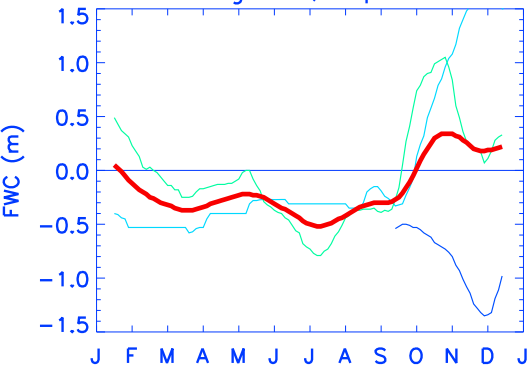


Salinity and Density

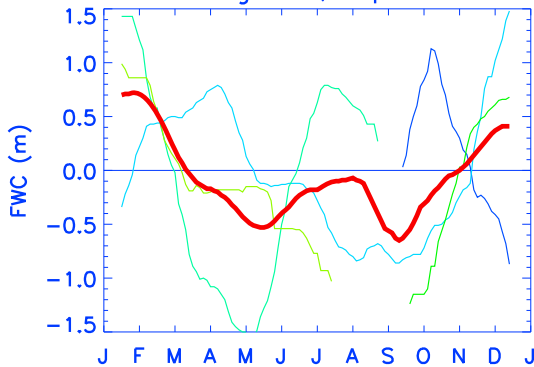




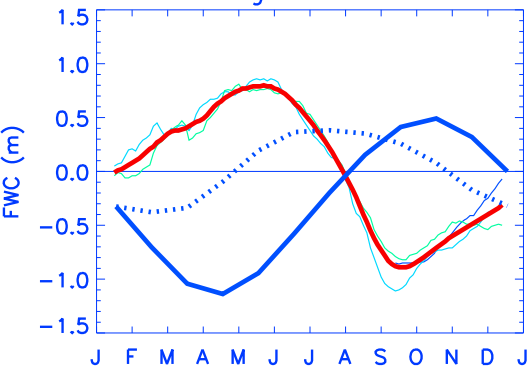
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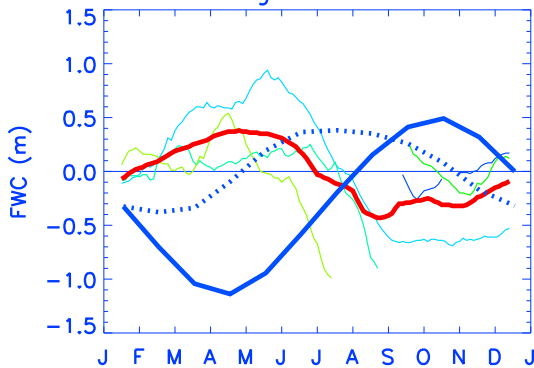
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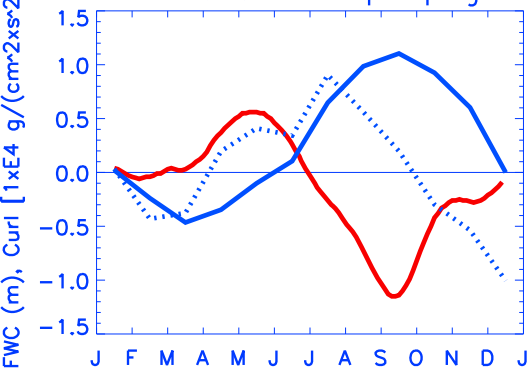
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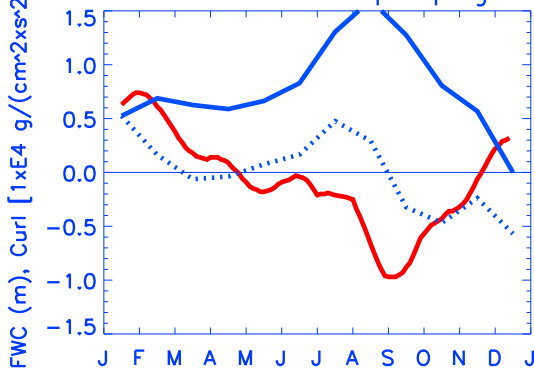
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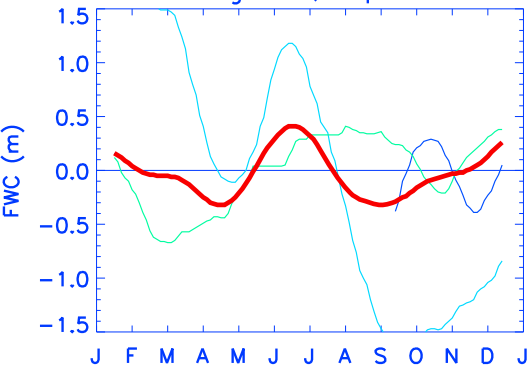
Moor. A: Ekman pumping



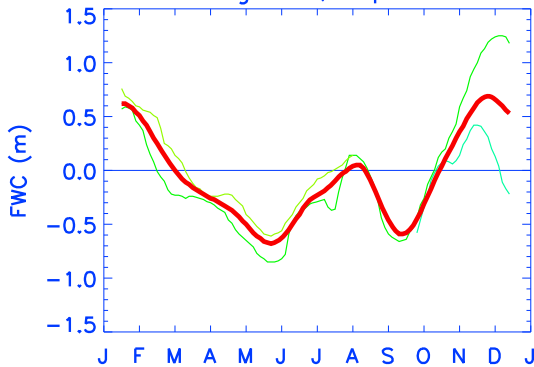
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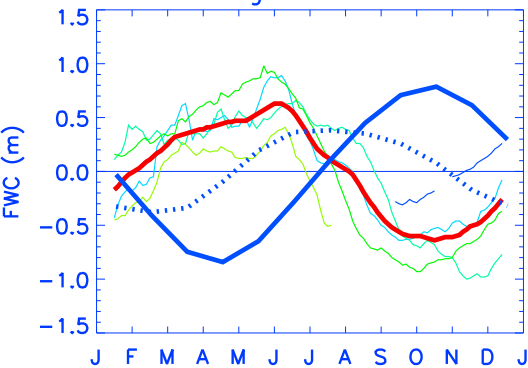
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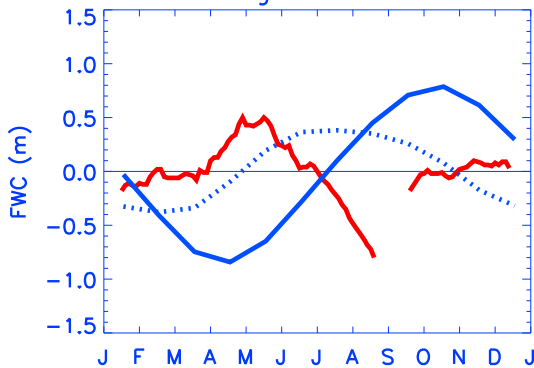
Mooring - D, Liquid FWC



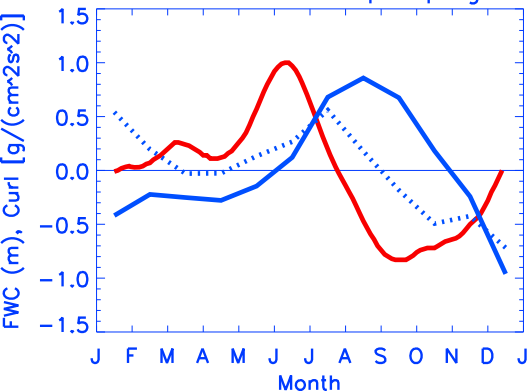
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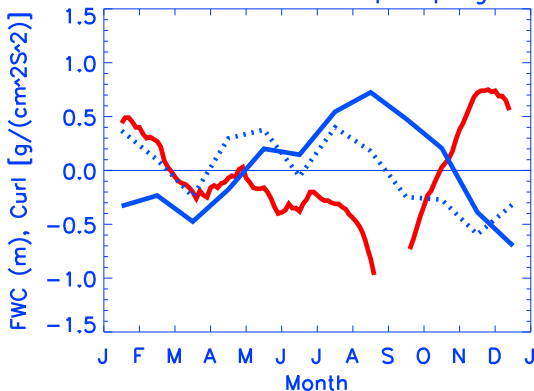
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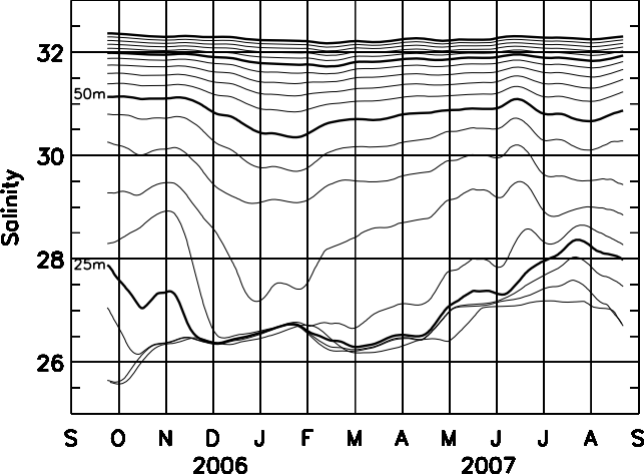


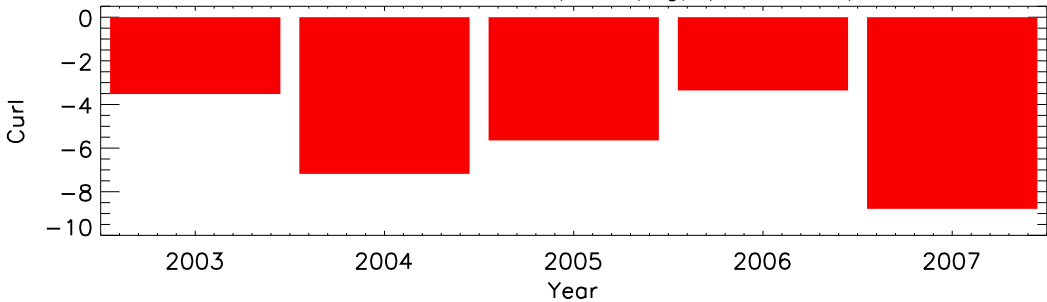
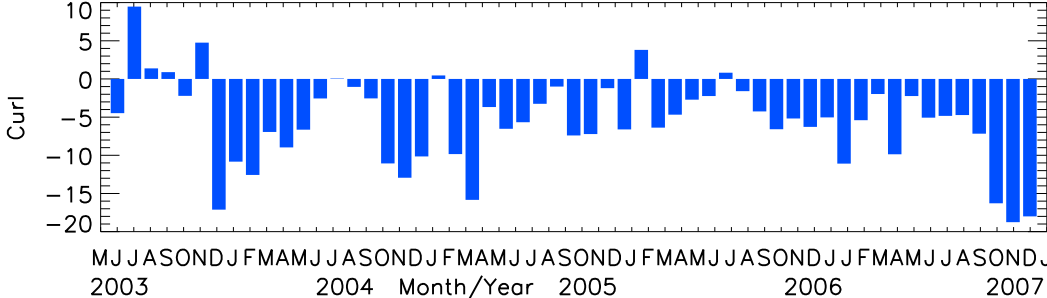
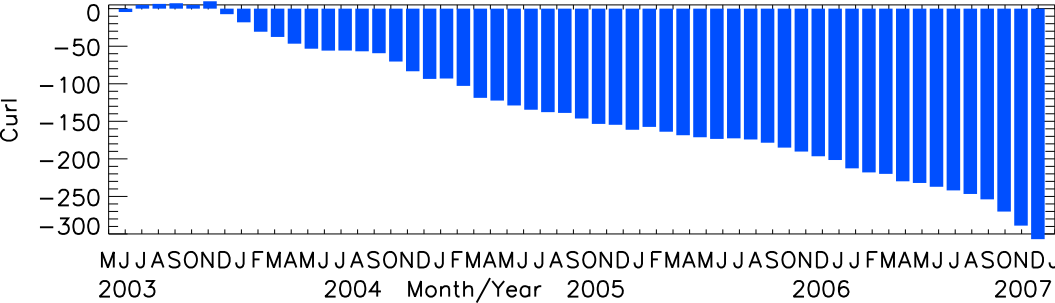
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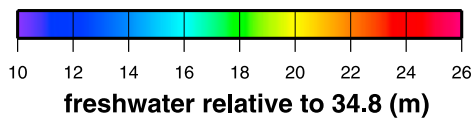
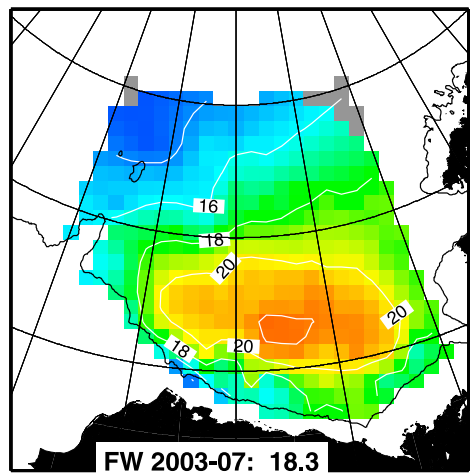
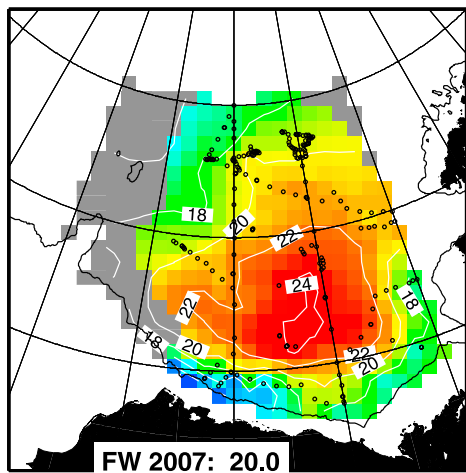
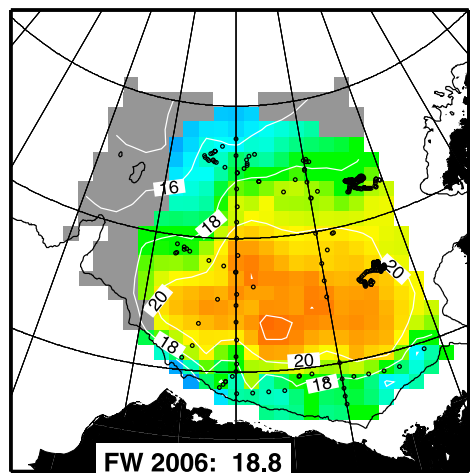
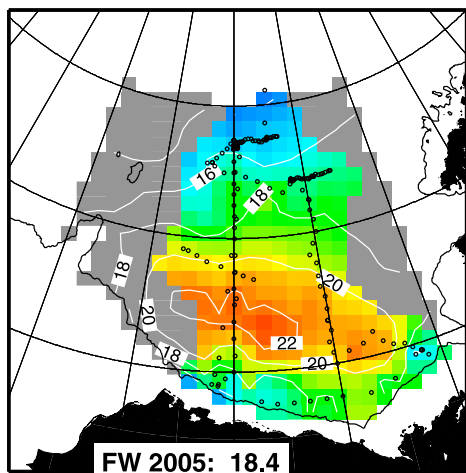
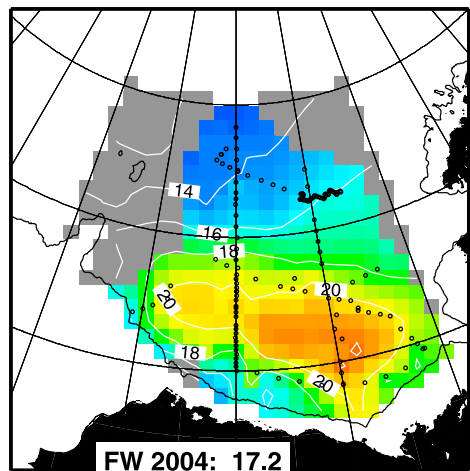
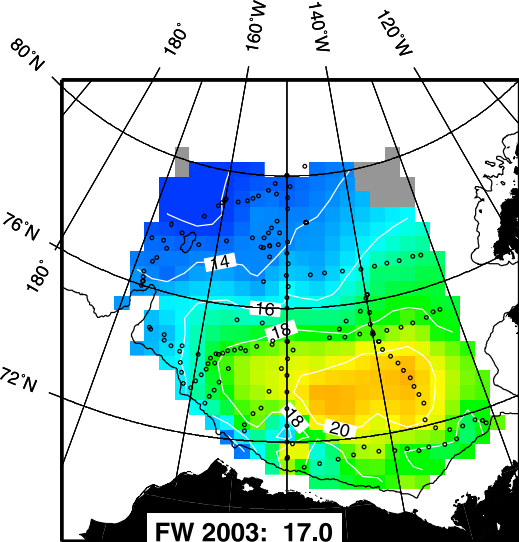


Moor. D: Ekman pumping

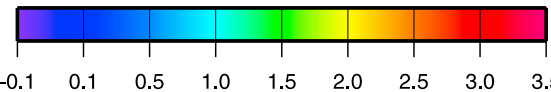
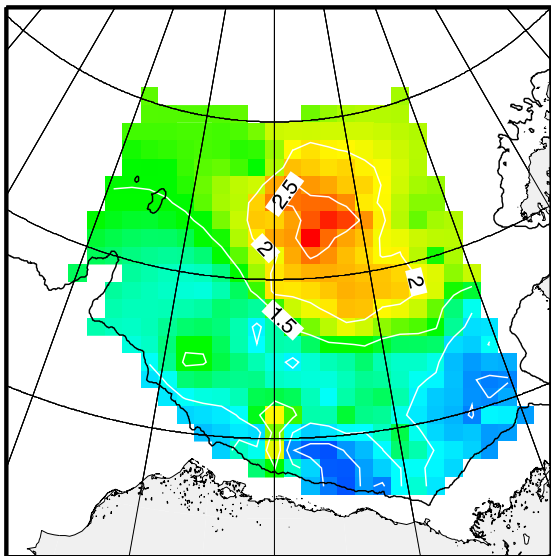




Annual wind curl, ($\times 1E4$) $g/(cm^2xs^2)$ Monthly wind curl, ($1 \times E4$) $g/(cm^2xc^2)$ Cumulatively added monthly wind curl, ($\times 1E4$) $g/(cm^2xs^2)$ 

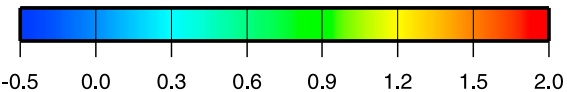
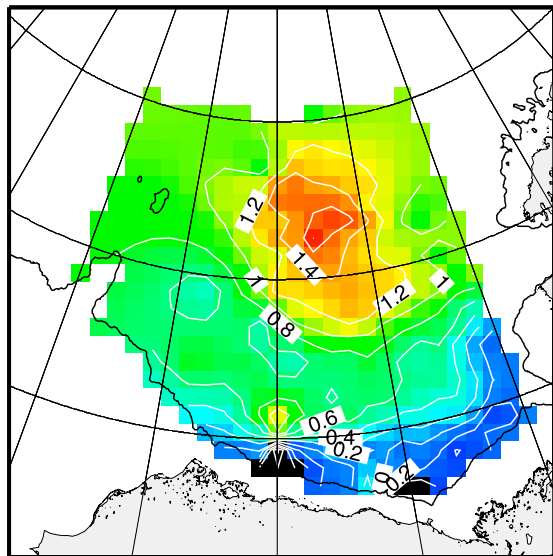


OI std

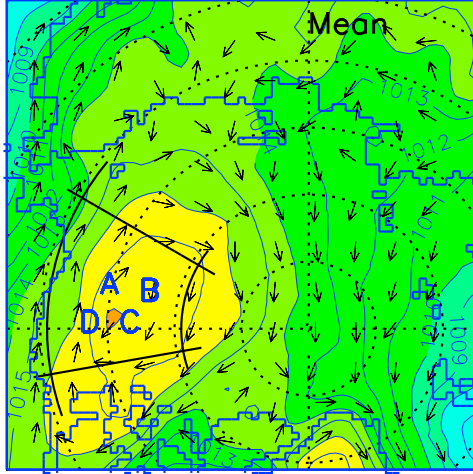
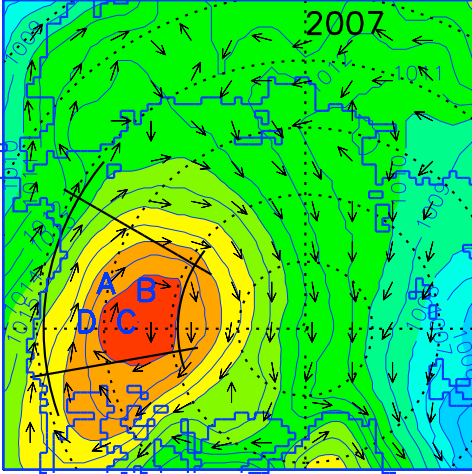
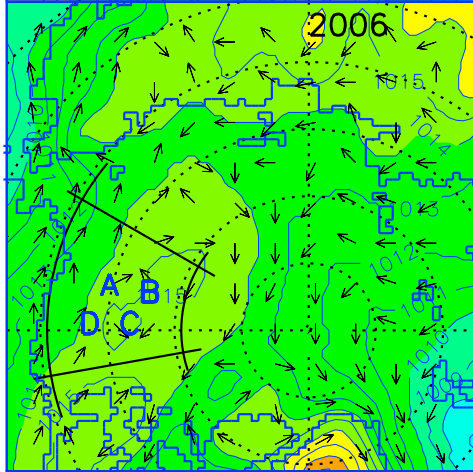
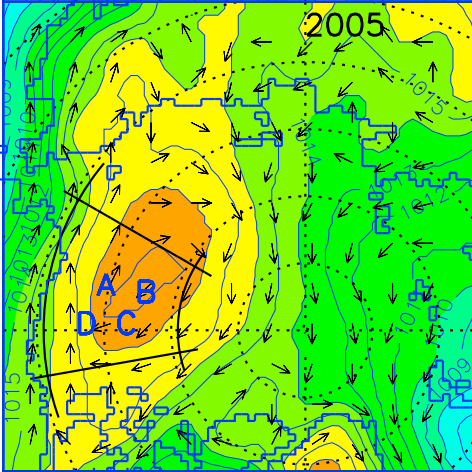
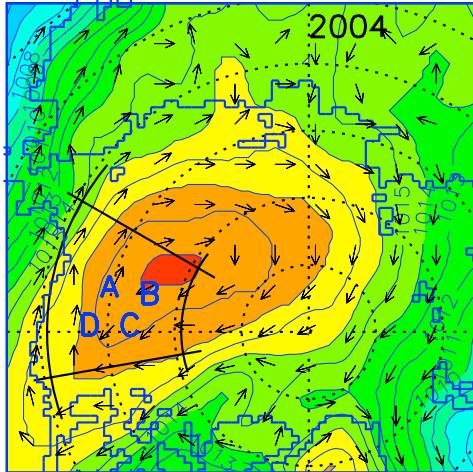
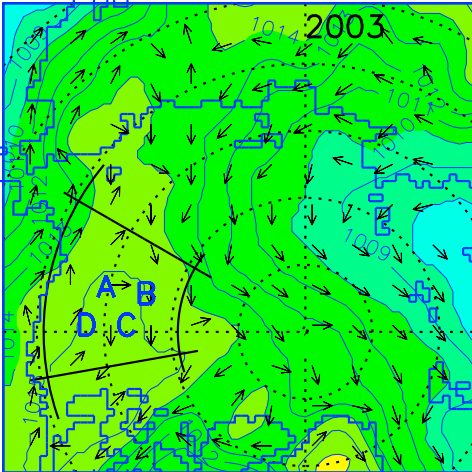


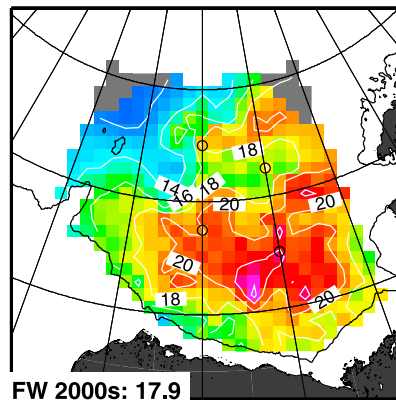
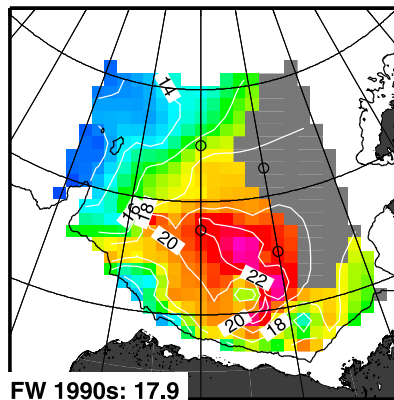
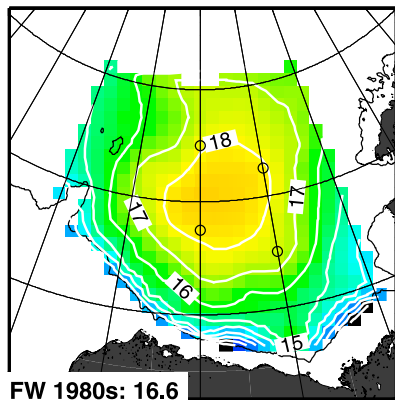
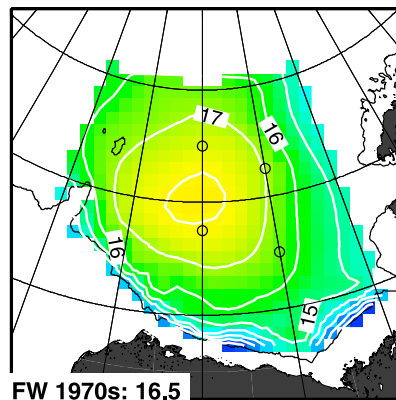
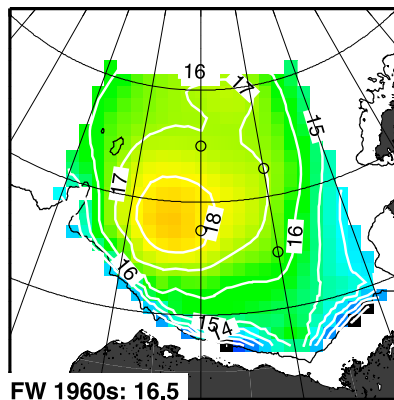
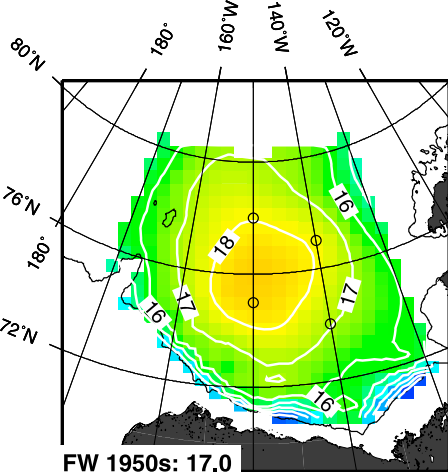
m

OI trend



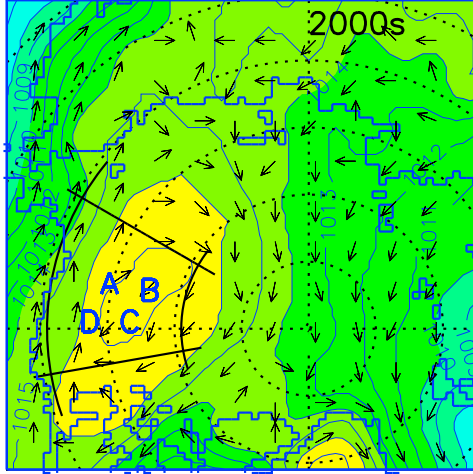
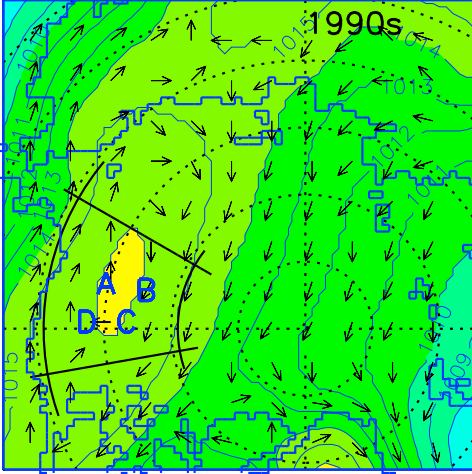
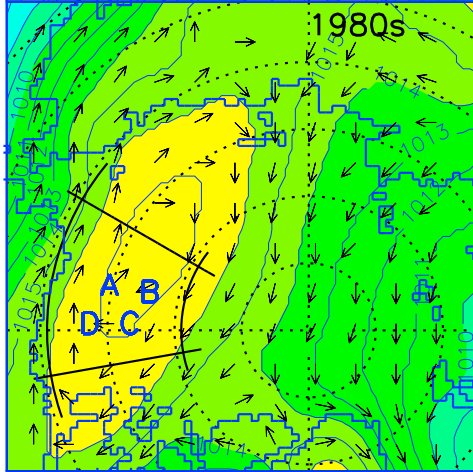
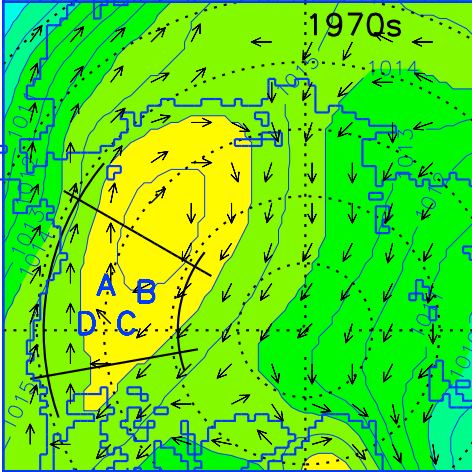
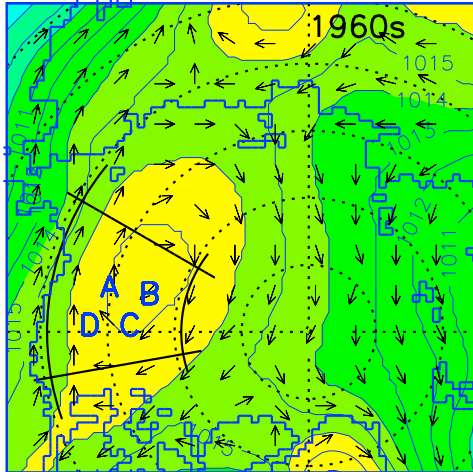
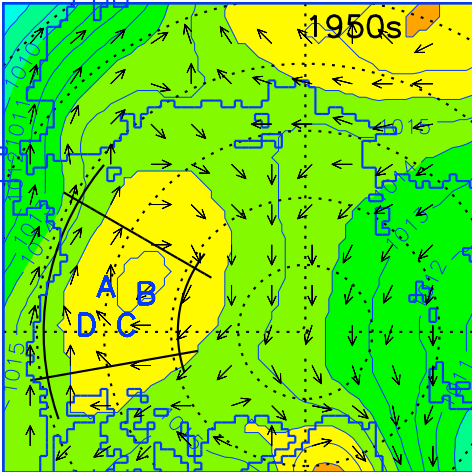
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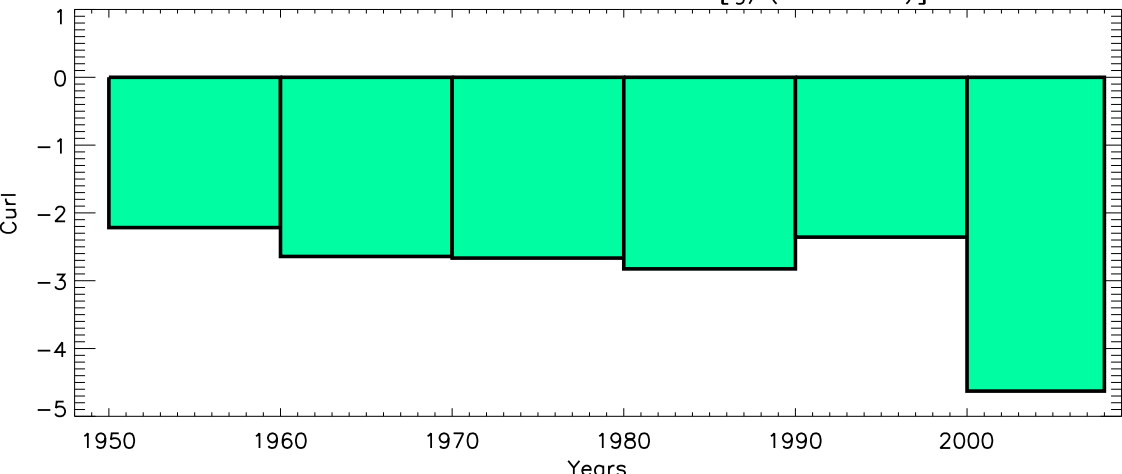




10 12 14 16 18 20 22 24 26

freshwater relative to 34.8 (m)



Decades: wind stress curl $\times 10^4$ [$\text{g}/(\text{cm}^2 \times \text{s}^2)$]Circulation regimes: wind stress curl $\times 10^4$ [$\text{g}/(\text{cm}^2 \times \text{s}^2)$]