Deep water distribution and transport in the Nordic seas from climatological hydrological data

HE Yan1,2*, ZHAO Jinping1, LIU Na2, WEI Zexun2, LIU Yahao3, LI Xiang3

1 Key Laboratory of Polar Oceanography and Global Ocean Change, Ocean University of China, Qingdao 266100, China
2 Laboratory of Marine Science and Numerical modeling. The First Institute of Oceanography, State Oceanic Administration, Qingdao 266061, China
3 Key Laboratory of Ocean Circulation and Waves, Institute of Oceanology, Chinese Academy of Sciences, Qingdao 266071, China

Received 10 April 2014; accepted 7 August 2014

©The Chinese Society of Oceanography and Springer-Verlag Berlin Heidelberg 2015

Abstract

Deep water in the Nordic seas is the major source of Atlantic deep water and its formation and transport play an important role in the heat and mass exchange between polar and the North Atlantic. A monthly hydrological climatology—Hydrobase II—is used to estimate the deep ocean circulation pattern and the deep water distribution in the Nordic seas. An improved P-vector method is applied in the geostrophic current calculation which introduces sea surface height gradient to solve the issue that a residual barotropic flow cannot be recognized by traditional method in regions where motionless level does not exist. The volume proportions, spatial distributions and seasonal variations of major water masses are examined and a comparison with other hydrological dataset is carried out. The variations and transports of deep water are investigated based on estimated circulation and water mass distributions. The seasonal variation of deep water volume in the Greenland Basin is around 22×10^3 km^3 whereas significantly weaker in the Lofoten and Norwegian Basins. Annual downstream transports of about 1.54×10^3 and 0.64×10^3 km^3 are reported between the Greenland/Lofoten and Lofoten/Norwegian Basins. The deep water transport among major basins is generally in the Greenland-Lofoten-Norwegian direction.

Key words: the Nordic seas, deep water, modified P-vector method, Hydrobase II


1 Introduction

The Nordic seas, also known as Greenland-Iceland-Norwegian (GIN) Sea, connect the Arctic and the North Atlantic. The deep circulation and transport of water masses in the Nordic seas are an important part of the global thermohaline circulation and strongly influence the Atlantic Meridional Overturning Circulation and the climate of adjacent areas (Meincke et al., 1997; Hansen and Østerhus, 2000).

The cyclonic flow patterns in the Iceland Plateau, the Norwegian Basin, the Lofoten Basin and the Greenland Basin were first reported by Helland-Hansen and Nansen (1909) and later studied with flow velocities derived from surface and mid-depth drifters by Poulain et al. (1996), Orvik and Niller (2002), and Voet et al. (2010). However, direct measurement of flow is expensive, and is hard to obtain the spatial distribution and temporal variation of the current field.

The geostrophic balance between horizontal pressure gradient and Coriolis force is true for circulation on scales of longer than a few days and over a hundred of kilometers. The vertical variances of flow are controlled by the spatial distributions of water temperature and salinity via thermal wind relation, with which one can calculate the flow velocities relative to a specific depth (usually the assumed level of no motion) with hydrographic data. However, the available bottom flow measurements indicate that these is no such a motionless level in the Nordic seas, meaning that some other constraints should be involved when retrieving the current fields from hydrographic data. Nøst and Isachsen (2003) developed a simplified diagnostic model to estimate the mean circulation of the Nordic seas. Along with climatological hydrographic data and wind stresses, a closed $f/H$, i.e., ratio between Coriolis frequency and water depth, contour and the assumption that the same contour is a streamline of the bottom geostrophic flow are introduced to drive the model. Hunegnaw et al. (2009) also gave an absolute flow field estimation with combined historical hydrographic data and geodetic data. In this paper, we use a modified P-vector method to compute the current field in the Nordic seas from Hydrobase II (HB-II hereafter) hydrographic data and satellite sea surface height (SSH).

The water masses in the Nordic seas are of distinct properties (Blindheim and Østerhus, 2005). The warm and saline Atlantic surface water (ASW) and Atlantic water (AW) encounters in the upper and mid-depth layers the cooler and less saline polar water (PW), while the deep water (DW), comprised of Greenland
Sea deep water (GSDW), Norwegian Sea deep water (NSDW), Arctic Ocean deep water (AODDW) and some other sub-classes, dominates the deep layer. The transformation and transport of water masses via isopycnal cabbeling (Gascard et al., 2002), ventilation (Smethie et al., 1986), basin-scale currents (Rudels et al., 1999), wind and sea ice (Meincke et al., 1992) and other effects (Hopkins, 1991) contribute to a significant deep water source to the Atlantic and potential driving factor to world ocean’s thermohaline circulation.

Piasek et al. (2008) carried on a census of water mass in the Nordic seas with several climatological, observational, and model-output datasets. They calculated the percentages in total water volume and seasonal variations of DW, AW, and other water masses. Here we processed the monthly HB-II data in a same way to compare HB-II with other climatological datasets, and to check the volume flux of DW among major basins: the Greenland, Lofoten, and Norwegian Basin to figure out the pathway and scale of deep-water transport in the Nordic seas.

2 Data and methods

2.1 Data

The gridded climatological data from HB-II provided by Woods Hole Oceanographic Institution are used in this study (Curry, 2001). The dataset is comprised of monthly objectively mapped fields of potential temperature and salinity at 85 standard depth layers. HB-II gridded products are integrated with isopycnal gridding and interpolation technology from many observations such as World Ocean Database 2001, WOCE Hydrographic Programme, NSIDC (Joint U.S./Russian Atlas of the Arctic Ocean), ICES, BarKode (Barents and Kara Seas Oceanographic Database) and some other data personally collected. The temporal coverage of these observations is from 1953 to 2002, and the spatial resolution is 0.25°×0.25°.

The SSH data used in the evaluation of geostrophic current fields are AVISO satellite altimeter products. The mean dynamic topography (MDT) during 1993–1999 period sea surface above geoid is taken as climatological SSH, and the climatological monthly mean SLA (sea level anomaly) data from Ssalto/Duacs multi-mission altimeter products is added on the SSH to estimate the monthly variation of the current field.

The SODA (Simple Ocean Data Assimilation) dataset is used in Section 2.3 to check the advantage of modified P-vector method. SODA is an oceanic reanalysis dataset consisting of gridded state variables for the global ocean, as well as several derived fields. In this study we use climatological monthly temperature, salinity, SSH, and current data. The climatological dataset has a uniform horizontal grid with 0.5°×0.5° horizontal resolution and 40 vertical layers.

2.2 P-vector method

The geostrophic currents are calculated from gridded HB-II data of temperature and salinity, which was largely done as per the P-vector inverse method (Chu, 1995) that is based on the mass conservation, geostrophic balance, and hydrostatic approximation. According to these preconditions, the conservation of potential density and potential vorticity can be obtained, meaning the geostrophic velocity should be perpendicular to both gradients of potential density and potential vorticity. The vector of velocity is defined like \( \mathbf{v}_P \), where \( \mathbf{P} \) is a unit vector, \( \gamma \) is a scalar, and its absolute value \( |\gamma| \) is the speed. The unit P-vector is given by

\[
\mathbf{P} = \frac{\nabla q \times \nabla \rho}{|\nabla q \times \nabla \rho|}
\]

where \( \rho \) is density and \( q \) is potential vorticity; \( \gamma \) can be computed by applying thermal wind relation between any two depth-layers.

The obvious advantage of this method is that the artificial defined motionless level is unnecessary, which is a puzzled question incorporating excess errors. No more special approximations are applied except for those mentioned. The geostrophic currents from the P-vector method is proved capable of reproducing the main features of circulation in open oceans and marginal seas (Chu, 1996). Although it is an over-determined system actually considering every two different layers can be used to calculate the scale of the vector, an optimization algorithm was developed to reach a final result with minimized errors (Chu, 1995).

However, there is a necessary requirement for this method. The isosurfaces of potential density and potential vorticity must not parallel to each other, which is a commonly necessary condition of inverse methods, e.g., the \( \beta \)-spiral method (Stommel and Schott, 1977). In this paper, we follow the calculation technique of Zhang et al. (2013). The P-vector method is not applied to the upper mixed layer, where the velocities are determined by the traditional dynamic calculation referring to a non-zero velocity calculated as shown below later. However, a new question emerges. What should be a reasonable depth for the upper limit of P-vector computation? It was not explained in that paper and the 800 MPa was set arbitrarily. For different thermohaline structures of the water in different regions, the upper limits should be sensitive to local circumstances. The Nordic seas are active regions of deep convection, and the mixed layer can be quite thick especially in winter. Therefore, we conducted a study on the impact of different upper limits in this study. In addition, a new technique of correcting the geostrophic currents from P-vector method is introduced.

2.3 Modified P-vector method

In open ocean area where the P-vector method was first applied, the horizontal pressure gradient generated by sea level fluctuation in the surface layer is progressively cancelled out by baroclinic pressure gradient induced by deformation of isopycnal water column. Finally, till a certain depth layer, these two pressure gradient cancel each other. There would be no geostrophic current existing beneath that, and the layer is known as motionless level. In this scene, the field of horizontal pressure gradient could be fully reconstructed from the distribution of density, so the geostrophic current fields for whole water column can be calculated accurately from hydrographic data via inverse methods based on thermal wind relation such as P-vector method.

While in some margin areas, where stratification is weak or the density distribution is strongly affected by other factors, the baroclinic adjustment may not be able to cancel out completely the barotropic pressure gradient generated by sea level fluctuation. The field of horizontal pressure gradient cannot be fully reconstructed from the distribution of density. There would be part of barotropic geostrophic current existing form surface to bottom that inverse method cannot recognize. To solve this problem, we calculated surface barotropic currents, \( U_{\text{SSH},0} \), from horizontal gradient of sea level fluctuation, and geostrophic currents at surface layer from P-vector method, \( U_{\text{P-vector},0} \).

and then derived the difference \( \Delta U \) between them,
\[ \Delta U = U_{ssh, \Delta z} - U_{p-vector, \Delta z} \]

\( \Delta U \) is the aforesaid current portion which P-vector calculation cannot recognize. This barotropic portion of current exists uniformly from surface to bottom. To evaluate the geostrophic currents, we should add \( \Delta U \) on P-vector output fields for every depth layer,

\[ U_{\text{corr}, z} = U_{p-vector, z} + \Delta U, \quad z \text{ for any depth}, \]

where \( U_{\text{corr}} \) is the final estimation of geostrophic currents. It's worth noting that when this modified method was applied to the area where horizontal pressure gradient could be fully reconstructed from density field and original P-vector method works, the \( \Delta U \) decreases to zero and the modified P-vector method generates same results as original one. It should be mentioned that the nature of modified P-vector method is still an inverse method. It can only be utilized where the geostrophic balance is dominant, otherwise the result will be nonsensical.

An examination was taken to verify the performance of modified P-vector method. In this kind of examination, a complete dataset including SSH, temperature and salinity data, and corresponding current velocity fields is needed as reference system. Current velocities are calculated with different methods from SSH and hydrographic data in this reference dataset, and then compared with the referenced velocities to check the accuracies of the methods. The temporal resolution of dataset should be monthly or longer to cancel small-scale processes and to highlight large-scale geostrophic momentum.

The climatological monthly SODA dataset is used in our study. We calculated the current fields with P-vector method, modified P-vector method, and dynamic calculation, and then compared them with SODA current field. A total set of eight experiments was carried out with different upper limit depths for P-vector ranging from about 46 m (Layer 5 in SODA) to 2125 m (Layer 27 in SODA). The spatial-averaged root-mean-square errors of flow directions and magnitudes between calculations and SODA values are shown in Fig. 1. The original P-vector performs the worst in these three methods, while modified P-vector gives the best results. The modified P-vector method brings progress in both direction and magnitude aspects. New method deduces the errors of flow direction from more than 50° to about 20°. It is noticed that in some area the P-vector method gives almost opposite of the SODA flow directions, especially under a shallow upper limit, while new method makes them corrected. The new method also reduces the errors in flow magnitudes obviously, which is much more remarkable for shallow upper limit experiments. Compared to the results of dynamic calculation, the new method shows a notable progress in flow directions but slight progress in flow magnitudes. It should be mentioned that although the experiment with deepest upper limit gives minimum error, an excessively deep upper limit may shrink the computational domain extremely, and then cause a waste of data and loss of current field features. A compromise between accuracy and representativeness should be taken into account when choosing the upper limit depth.

The climatological vertical-averaged (918–3125 m, SODA Layers 22–31) flow patterns from the three calculations and SODA data are shown in Fig. 2 to give an intuitionistic comparison. The upper limit in P-vector calculation is about 918 m (SODA Layer 22) here. Original P-vector method (Fig. 2a) and dynamic calculation (Fig. 2b) underestimate obviously the magnitudes of velocity, especially for those in the Lofoten and Norwegian Basins, while the modified P-vector method (Fig. 2c) reproduces the SODA current field (Fig. 2d) much better with only slight overestimation.

It should be stated that the SODA is oceanic reanalysis dataset, so its current patterns, even the climatological patterns, are not entirely geostrophic flows, and of course slightly differ from those calculated from hydrographic data, especially on continental slope and in nearshore areas.

The modified P-vector method proves to perform well in this region. Next, we will use this method and field observations (AVISO SSH and HB-II hydrological data) to evaluate the deep circulations in the Nordic seas.

3 Results

3.1 Deep circulation patterns

In order to give an “observed” deep circulation pattern, the data used in this calculation are all from observations, i.e., HB-II hydrological data, AVISO remote sensed MDT and climatological monthly SLA. Due to the seasonal existence of ice cover around northwest continental shelf area, the satellite remote sensed SSH at that region are mostly extrapolated from adjacent area and so of low quality. In fact, strong geostrophic currents

![Fig.1](image-url). Spatial-averaged errors of flow directions (a) and magnitudes (b) between calculations and SODA values.
are generated from the horizontal gradient of AVISO MDT here, which is in contrary to the observations and theoretical results that there is a strong southerly East Greenland Current flowing along the coast of Greenland and carrying cold Arctic water into the Atlantic. Here we checked the quality-control parameter of MDT data, picked out the areas with relative low quality and turned off the current calculation in these areas. P-vector calculation was carried out to the same area where barotropic geostrophic current was figured. The upper limit was set to 950 m, Layer 25 in HB-II.

Previous studies revealed that the circulations in the Nordic seas generally follow the closed f/H contours, and the flow patterns are similar from top to bottom due to weak stratification. Our calculations give identical results. The barotropic geostrophic part generated from sea surface gradient is much stronger than the baroclinic part contributed by density distribution. The flow patterns are almost the same from upper limit to bottom. Flows steered by topography on edges of basins are stronger and clearer than those inside the basins. The major currents along the margins of the Lofoten and the Norwegian Basins are cyclonic, which extent as far as to the eastern edge of the Greenland Basin. Some sub-basin eddies exist inside the basins. No result is provided owing to SSH issues. The flow pattern near northwest continental shelf is questionable out of the same reason.

Figure 3 illustrates the vertical averaged deep current patterns (950 m to bottom) in summer (Fig. 3a) and winter (Fig. 3b). The basin scale circulation is cyclonic all year round with slight seasonal variation of strength. The circulation strengthens in winter, which is more obvious for the currents along Jan
Mayen Ridge and the eastern boundary.

The current patterns in Fig. 3 seem much more fragmentary than those from SODA. It may attribute to that HB-II and AVISO data have much higher resolutions and then may contain more small- and meso-scale structures than SODA. An average of multiple years data could reduce some short period structures and retain the dominant ones, and that is why we use climatological monthly data to reconstruct the flow patterns. However, although the HB-II data are all from observations, the spatial and temporal distributions of observations are of course not uniform, and the flow patterns are by no means 100% real. Despite all of this, these flow patterns derived from high-resolution observations give a more detailed and authentic description than before.

3.2 Water mass distribution

The statistical analysis of water properties was made based on the well known T-S (potential temperature-salinity) relation. Figure 3 is the climatological T-S diagram for the Nordic seas. The climatological values are derived by averaging the climatological monthly data. In order to facilitate the comparison with the results of Piacsek et al. (2008, P08 hereafter), we took the census in the same region they chose 60°–80°N, 20°W–20°E. We counted the numbers of data point falling in each 0.01×0.01°C square and showed them in Fig. 4 with distinct gray levels. The primary decomposition of T-S domain into seven major classes by P08 is introduced to this study, and this T-S division is also shown in Fig. 4.

As Fig. 4 shows, most of water samplings in the Nordic seas are classified into DW (T<0°C, 34.85<S<35.2), LAIW (lower Arctic Intermediate water, 0°C<T<3°C, S>34.9) and AW (Atlantic Water, 3°C<T<10°C, S>34.9), few into UAIW (upper Arctic Intermediate Water, T<2°C, 34.7<S<34.9, outside DW domain), ASW (Arctic surface water, 34.4<S<34.9, outside PIW and UAIW domains) and PW (S<34.4), and only precious few into PIW (Polar intermediate water, T<–0.5°C, 34.4<S<34.7). Considering there are more than 300 thousands of data points in total, the dark area located in DW and LAIW regions along 1.028 kg/m³ iso-potential-density line could contains more than half of the total water volume.

We also calculated the distribution (percentages) of these water masses in the complete Nordic seas, just the same as what P08 did. Besides, the monthly variations of the distributions were investigated as well. Figure 5 shows the results for the seven major water masses. As seen in Fig. 4, the three main water masses: DW, LAIW, and AW take more than 90% in total, and DW alone takes more than 60%. Except for PIW, the temporal variations of all water masses are small compared to their own magnitudes.

A comparison between HB-II and other climatologies was carried out and the results are shown in Table 1. The HB-II wa-
water mass distributions are in better agreement with WOA01 than with the other 2 datasets, especially in fall. The HB-II/WW01 ratio is 15.8/13.3 and 16.1/15.2 in spring and fall, respectively for AW, and 67.0/66.3 and 67.2/65.9 for DW, while the ratio lower to 11.0/15.1 and 11.0/12.6 for the transitional LAIW mass, meaning that the mixing between DW and AW therefore the generation of LAIW are more obvious in WOA01 than in HB-II. We convert the seasonal changes in percentage to those in volumes by multiplying the percentage changes with the total water volumes. It is shown in HB-II that the spring to summer increase \( \Delta V \) of AW is about 8.9×10³ km³, the spring to fall increase is about 16.8×10³ km³. For \( \Delta V \) of DW the values become 18.0 and 9.0×10³ km³, respectively. For \( \Delta V \) of ASW the values become 16.2×10³ and 36.2×10³ km³, respectively. The increases of AW and DW during spring to fall correspond to decreasing \( \Delta V \) of −50.7×10³ km³ for spring to summer and −63.1×10³ km³ for spring to fall. This result indicates that the mixing between AW/ASW and DW and thus the generating of AW, and DW, and 67.0/66.3 and 67.2/65.9 for DW, while the ratio lower to 11.0/15.1 and 11.0/12.6 for the transitional LAIW mass, meaning that the mixing between DW and AW therefore the generation of LAIW are more obvious in WOA01 than in HB-II. We convert the seasonal changes in percentage to those in volumes by multiplying the percentage changes with the total water volumes. It is shown in HB-II that the spring to summer increase \( \Delta V \) of AW is about 8.9×10³ km³, the spring to fall increase is about 16.8×10³ km³. For \( \Delta V \) of DW the values become 18.0 and 9.0×10³ km³, respectively. For \( \Delta V \) of ASW the values become 16.2×10³ and 36.2×10³ km³, respectively. The increases of AW and DW during spring to fall correspond to decreasing \( \Delta V \) of −50.7×10³ km³ for spring to summer and −63.1×10³ km³ for spring to fall. This result indicates that the mixing between AW/ASW and DW and thus the generating of LAIW are only strong in winter half year, but in summer half year, this process is much weaker. Two vertical sections are picked up to illustrate the mean spatial distribution of water masses. Section-W lies along 4°W, crossing over the Norwegian Basin, Jan Mayen Fracture Zone and central Greenland Basin (Fig. 6). Section-E lies along 4°E, passing through the Voring Plateau, the Lofoten Basin, the Mohn Ridge and eastern Greenland Basin (Fig. 7). The potential density is around 1 028 kg/m³ in the Nordic seas. On both sections, it increases rapidly from as low as 1 027 at surface to 1 028 at about 500 m depth, and then increases very slowly to about 1 028.1 at bottom. Water of low density occupies the upper layer adjacent to the Fram Strait and in the Lofoten and Norwegian Basins while isopycnals shift up in the Greenland Basins. The low density water in upper Lofoten and Norwegian Basins are sorted to warm, saline AW, and those adjacent to the Fram Strait are sorted to PW and LAIW. The less salty PW extends from the Fram Strait to the entire Greenland Basin at surface layer, with UPIW lying below it. The area below 500 m depth is almost occupied by DW. The warm and fresh ASW appears on south end of Section-E where Norwegian Coastal Current flows by.

The water column in the Greenland Basin is colder but slightly fresher than adjacent waters on both sides. T-S sectional patterns imply that the cold water might be generated locally by surface cooling and vertical mixing. The cold, fresh water goes deep by vertical stirring, then densifies during mixing with adjacent salty water, and finally forms densest deep water. The local cooling effect could make the Greenland Basin a cool source for the whole Nordic seas and the cold water should expand down to adjacent areas along isopycnals.

DW in the Nordic seas can be classified into some sub-masses according to P08, among which three major masses are GSDW (Greenland Sea deep water, 34.88≤S≤34.92, −1.3°C≤T≤−1.1°C), NSDW (Norwegian Sea deep water, 34.90≤S≤34.92, −1.1°C≤T≤−0.5°C), and AODW (Arctic Ocean deep water, 34.92≤S≤34.94, −1.1°C≤T≤−0.5°C). Indicated from the sectional patterns, the coldest water cumulated at the bottom of the Greenland Basin forms GSDW, while the mid-depth cold water transports along isopycnal over the Mohn Ridge and forms a slight fresh tongue in the mid-depth layer of the Norwegian Sea. This cold water mass mixes with deep salty water located in the Norwegian Sea and then forms NSDW and more salty AODW near bottom. The distribution of DW coincides with conclusions draw from previous research (Swift and Koltermann, 1986; Blindheim, 1990; Aagaard et al., 1991)

### 3.3 Deep water transport

The Nordic seas are important source region of Atlantic deep water. An investigation of deep water variation and transport in this region may help deepen understanding of deep circulations in Atlantic. We investigated the volume variations of DW in the three major basins: Greenland, Lofoten, and Norwegian. The monthly DW volume variations are shown in Fig. 8.

The DW volume in the Greenland Basin undertakes a remarkable seasonal variation of 22×10³ km³, which is in the same order as the result of Carmack and Aggard (1973). The volume increases dramatically during March and June and then slowly decreases in the other months. It might due to the strong

### Table 1. Water mass distribution (%) in the Nordic seas for four climatologies. Here the data for GDEM93, MODAS01 and WOA01 are derived from Table 4 of P08. HB-II values of March to May are averages for spring, and September to November for fall

<table>
<thead>
<tr>
<th>Water mass</th>
<th>GDEM93 (0.5°)</th>
<th>MODAS01 (0.125°)</th>
<th>WOA01 (0.25°)</th>
<th>HB-II (0.25°)</th>
<th>GDEM93 (0.5°)</th>
<th>MODAS01 (0.125°)</th>
<th>WOA01 (0.25°)</th>
<th>HB-II (0.25°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>PW</td>
<td>1.4</td>
<td>1.1</td>
<td>1.0</td>
<td>0.8</td>
<td>1.2</td>
<td>1.1</td>
<td>1.7</td>
<td>1.2</td>
</tr>
<tr>
<td>PIW</td>
<td>0.3</td>
<td>0.2</td>
<td>0.5</td>
<td>0.3</td>
<td>0.5</td>
<td>0.1</td>
<td>0.2</td>
<td>0.0</td>
</tr>
<tr>
<td>AW</td>
<td>17.7</td>
<td>18.8</td>
<td>13.3</td>
<td>15.8</td>
<td>18.5</td>
<td>19.0</td>
<td>15.2</td>
<td>16.1</td>
</tr>
<tr>
<td>ASW</td>
<td>1.1</td>
<td>1.1</td>
<td>2.2</td>
<td>2.7</td>
<td>1.6</td>
<td>2.5</td>
<td>3.1</td>
<td>3.4</td>
</tr>
<tr>
<td>UAIW</td>
<td>3.4</td>
<td>2.0</td>
<td>1.6</td>
<td>2.2</td>
<td>2.8</td>
<td>2.6</td>
<td>1.2</td>
<td>1.0</td>
</tr>
<tr>
<td>LAIW</td>
<td>14.8</td>
<td>12.3</td>
<td>15.1</td>
<td>11.0</td>
<td>13.2</td>
<td>12.6</td>
<td>12.6</td>
<td>11.0</td>
</tr>
<tr>
<td>DW</td>
<td>61.3</td>
<td>64.7</td>
<td>66.3</td>
<td>67.0</td>
<td>62.2</td>
<td>62.3</td>
<td>65.9</td>
<td>67.2</td>
</tr>
</tbody>
</table>
cooling effect of ice-cover melting and relative low surface air temperature in spring, both of which promote the formation of cold water. However, the cases for the Lofoten and Norwegian Basins are dissimilar to that for Greenland. The DW volumes in these two basins keep fluctuating all year round without any long term increase/decrease term, implying more complicated control dynamics.

Although the major circulations are almost topographically constrained inside the basins, there are still relative weak overflows existing on edges of basins, transporting deep water among the basins. Based on our results of deep circulations, we calculated the overflows of DW across the Mohn Ridge and the Jan Mayen Fracture Zone (Fig. 9), which are basin boundaries for Greenland-Lofoten and Lofoten-Norwegian, respectively. The DW transport is weak and generally in the Greenland-Lofoten-Norwegian direction with some short-term inverse flows. The annual mean transports are about 0.049×10^6 and 0.020×10^6 m^3/s respectively at the two boundaries, carrying about 1.54×10^3 and 0.64×10^3 km^3 DW downstream per-year. This result, together with the water mass distribution patterns, indicates that the Greenland Basin is a major origin of DW and the Greenland-Lofoten-Norwegian direction should be a pathway for DW.

4 Conclusion

The deep circulation and water mass distribution in the Nordic seas are investigated with a climatological monthly hydrological dataset HB-II in this paper. The HB-II dataset has a res-
olution of 0.25°×0.25°, so the flow patterns and the water mass distributions derived from this dataset are more detailed and authentic than before.

The major currents are generally constrained inside several deep basins and flowing cyclonically along isobaths. The currents strengthen slightly in winter while the flow patterns remain similar all year round. The flow patterns do not vary much from surface to bottom, showing strong barotropic nature. In the calculations of flow fields, the P-vector method was employed and a remarkable modification was made to the method by introducing sea surface heights into the calculation to cover the remaining portion of barotropic pressure gradient that cannot be cancelled out by baroclinic adjustment of density fields. The modified P-vector method proves to perform better than original P-vector method and dynamic calculation, which could provide a good example about geostrophic current estimation in area without motionless level.

The water in the Nordic seas can be classified into several distinct water masses. Among them, DW, LAIW, and AW together take more than 90% of the volumes, while DW alone takes more than 60%. The volumes of water masses do not undertake much seasonal variations. DW occupies most of the area lower than 500 m, whose upper extreme is even shallower in the Greenland Basin. AW takes upper layer in the Norwegian Sea, while ASW exists in the southeast coastal surface area nearby Scandinavian Peninsula. The PW and U/LAIW are generally distributed in the entrance area of Arctic water around the Fram Strait.

Fig. 7. Potential density, potential temperature, and salinity patterns on Section-E.
The DW volume in the Greenland Basin undertakes a distinct seasonal variation of $22 \times 10^3$ km$^3$, which increases dramatically during March and June and then decreases slowly during the other months. The seasonal variations of DW in the other two basins are weaker and more complicated. The transports of DW among basins over the Mohn Ridge and Jan Mayen Fracture Zone are weak and generally in the Greenland-Lofoten-Norwegian direction with some short-term inverse flows. The annual mean transports are about $0.049 \times 10^6$ and $0.020 \times 10^6$ m$^3$/s respectively at the two boundaries, carrying about $1.54 \times 10^3$ and $0.64 \times 10^3$ km$^3$ DW downstream per-year. The transport calculations and the water mass distribution patterns indicate that the Greenland Basin is a major origin of DW and it transports over basin rims form Greenland to Norwegian Sea.

**Acknowledgements**

MDT_CNES-CLS09 was produced by CLS Space Oceanography Division, while the SLA altimeter products were produced by Ssalto/Duacs. They are distributed by AVISO, with support from CNES (http://www.aviso.oceanobs.com/).

**References**


