14

Inverted Circulations in the Arctic Mediterranean Seas

The Arctic Mediterranean seas play an important role in global climate through its effects on the role of the ocean in the local radiation balance and water exchange with lower latitudes. To understand these effects, the large-scale circulation pattern and water mass distributions should be first studied.

14.1 Geographic Features

The Arctic Mediterranean seas, comprised of the basins lying north of the Greenland–Scotland Ridge, of about $9.5 \times 10^6\,\mathrm{km}^2$, representing 2.6% of the total oceanic surface (Fig. 14.1). The two principal areas of this system are the Greenland–Iceland–Norwegian (GIN) Sea and the Arctic Ocean. They are connected by the Fram Strait, the wide (600 km), deep (sill depth about 2,600 m) passage between Greenland and Spitzbergen.

The Arctic Ocean is roughly divided into the Canadian basin (maximum depth about 3,800 m) and the Eurasian basin (maximum depth about 4,200 m) by the Lomonosov Ridge (sill depth about 1,400 m). The Canadian basin is further divided into the Canada and Makarov basins by the Alpha–Mendeleyev Ridge, and the Eurasian basin is divided into the Nansen and Amundsen basins by Nansen–Gakkel Ridge.

The continental shelf from Alaska to Greenland is relatively narrow, but from Spitzbergen eastward (the Barents, Kara, Laptev, East Siberian, and Chukchi seas) the shelf is broad, typically from 600 to 800 km. A number of submarine canyons indent the shelf, the largest being the Svataya Anna and Voronian canons in the Kara Sea.

The Arctic oceans have surface layers freshened by continental inputs (rivers and glacial melt water) and by sea ice melt water. Discussion on stream flow from major rivers entering the Arctic Ocean is given by a number of authors. The total annual stream flow into the Arctic, including the Arctic Archipelago, is about $3{,}500\,\mathrm{km}^3\,\mathrm{year}^{-1}$. An additional $1{,}500$–$2{,}000\,\mathrm{km}^3\,\mathrm{year}^{-1}$

Fig. 14.1. Bathymetry of the Arctic Ocean showing the Eurasian and Canadian basins separated by the Lomonosov ridge. The 200 m depth contour marks the edge of the continental shelf

enters as a freshwater fraction in the Bering Strait inflow (Coachman et al. 1975). Aagaard and Coachman (1975) compute the fresh-water residence time (defined as stored volume divided by inflow) to be roughly l0 years for the whole Arctic basin, with local values as low as 2 years applying to the southern Eurasian basin.

Significant annual and interannual variations occur in stream flow (Cattle 1985). The large Russian rivers Yenisei and Lena exhibit on average about a 40-fold change between low flows in winter and peak flows in June and July; seasonal variability for the Mackenzie River is much less, about fivefold. Interannual variability is from 5 to 20% of the mean annual flow, depending on the individual rivers.

Interannually, the total extent of sea ice in the Arctic varies by about 5° of latitude at all locations where the ice advance is not bounded by land (Walsh and Johnson 1979). At the winter maximum this amounts to about 30% variability for individual regions (Johnson 1980). On the other hand, Carsey (1982) examined ESMR data from the Arctic at the time of minimum ice extent and noted that while regional variations may be large, the total coverage varies interannually by about 2% only.

14.2 Thermohaline Features

Thermohaline features can be detected from the joint US-Russian EWG atlas (Arctic Climatology Project 1997, 1998). The EGW Atlas has gridded data distributed on local Cartesian coordinate. All the grid cells have the same size. This is different from the WOA dataset, which uses the spherical coordinate system. At the North Pole, the size of the grid cell is zero. Here, the (T, S) fields from the EGW Atlas at 50 m (subsurface level), 500 m (intermediate level), and 2,000 m (deep level) are presented to show the thermohaline characteristics.

14.2.1 Subsurface Level (50 m Depth)

Three main types of surface water are recognized from the EWG atlas. The first is Atlantic Water, which is carried into the system as a branch of the Norwegian—Atlantic Current. It can be defined as water with temperature (Fig. 14.2) above 3°C and salinity (Fig. 14.3) greater than 34.9 ppt. The second is the Polar Water, which is water that has been diluted by admixtures with fresh water; it is generally cold (temperatures below 0°C) and fresh (salinity below 34.4 ppt). It occupies the upper layers of the Arctic down to 200 m depth and makes up the surface outflow within the East Greenland Current and the Canadian Arctic Archipelago. The third is the Arctic Surface Water, which is found mainly in the gyres of the Greenland and Iceland seas. This water is warmer and more saline than Polar Water, but cooler (0–3°C) and fresher (34.4–34.9 ppt) than Arctic Water. However, this water is notably denser than

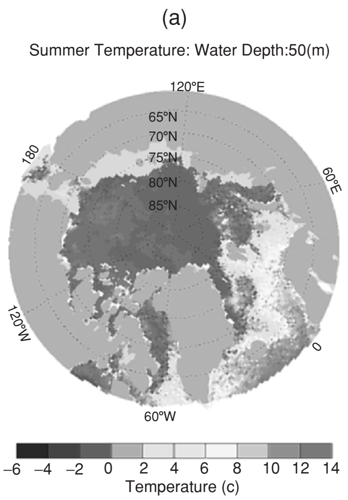

Fig. 14.2. Temperature of the Arctic Mediterranean Seas at 50 m depth from the Joint US-Russian EWG atlas: (**a**) summer, and (**b**) winter

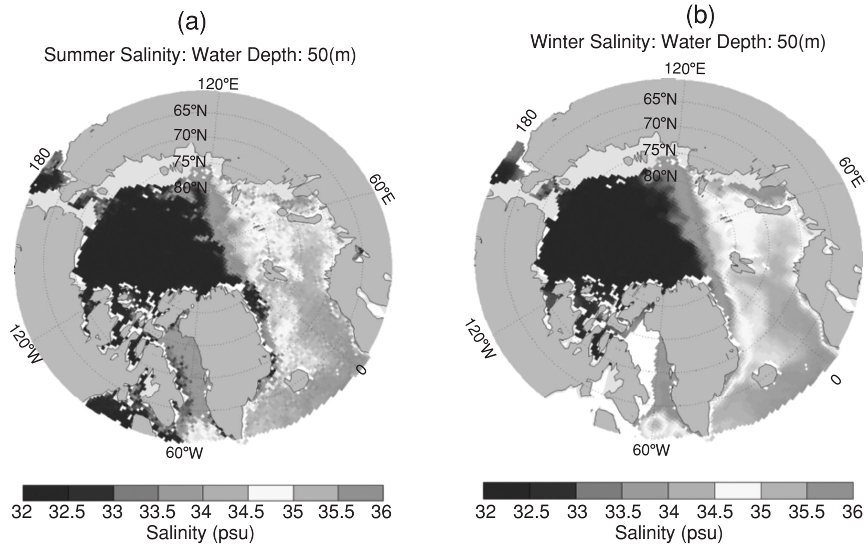

Fig. 14.3. Salinity of the Arctic Mediterranean Seas at 50 m depth from the Joint US-Russian EWG atlas: (**a**) summer, and (**b**) winter

either Polar Water or Arctic Water, indicating that it is not a simple mixture of the two and that large modifications due to air–sea exchanges occur locally.

In winter the surface layer tends to be uniform vertically in temperature and salinity. Ice melting in summer results in a pronounced salt stratification; however, this water remains near freezing except for areas that become completely ice free. The lower part of the surface layer contains the main halocline, the layer with low temperatures (less than $-1°C$) and salinities between about 30.4 and 34.4 ppt, which is arguably the most important feature of the Arctic Ocean. This cold, relatively saline layer is thought to be maintained by shelf drainage during winter (Aagaard et al. 1981; Melling and Lewis 1982). In the Eurasian basin the salinity increases rapidly with depth, reaching 34.9–35.0 ppt at about 200 m, while the temperature remains colder than $-1.5°C$ to 150 m and then increases with depth. In the Canadian basin the halocline is deeper, and the salinity increases more slowly with depth. Here, the temperature shows two minima, near the depths with salinities of 31.6 and 33.1 ppt, and a maximum near 32.4 ppt. The minimum near 31.6 ppt is possibly a remnant of winter cooling. The maximum near 32.4 ppt and the minimum near 33.1 ppt possibly reflect the inflow from the Pacific through Bering Strait. Hence, the Arctic halocline is not a uniform structure nor is it likely composed of single water mass. It likely derives from more than one region and is formed by more than one ventilation mechanism (Jones and Anderson 1986).

Figures 14.2 and 14.3 show the existence of thermohaline fronts or frontal zones in the GIN Sea. They are oriented more or less meridionally. It separates the warm and salty Atlantic Water in the Norwegian and West Spitzbergen currents from the cooler and fresher water of the Arctic domain (Dietrich 1969). In the Greenland Sea it follows the mid-At1antic Ridge (the Knipovich Ridge) and merges in the Fram Strait with the Polar front, owing to local recirculation. On the western side of the Arctic front a succession of cyclonic gyres in the Greenland Basin, the Boreas Basin, and the Fram Strait have been observed (Quadfasel et al. 1987; Quadfasel and Meincke 1987). In the transition different water masses interface and form frontal zones that not only separate water bodies with different hydrographic characteristics but also the regional biological systems. The GIN Sea is a key region in the advective–convective system that links the polar ice with the North Atlantic (van Aken et al. 1991).

The seas north of the Greenland–Scotland Ridge constitute a major heat sink in the global thermohaline circulation of the world ocean and therefore a crucial component of the earth climate (Aagaard et al. 1985). A large heat loss to the atmosphere, combined with sea ice production and melting, is responsible for the formation of deep and intermediate waters through winter convection, which, in some basins like the Greenland Sea, may reach down to the bottom.

The one-dimensional convection process is strongly affected by the horizontal dynamics of the convective basins including lateral exchanges of heat, salt, and ice at their boundaries. Concerning the Greenland Sea, important exchanges occur through its northern boundary, the Fram Strait, which is the only deep connection between the Arctic Ocean and the rest of the world ocean. A better knowledge of the dynamics of the strait is therefore crucial in the context of investigating the processes involved in the deep convection and their variability.

14.2.2 Intermediate Level (500 m Depth)

Three types of the Arctic intermediate water exist with different T/S characteristics (Swift and Aagaard 1981). The Arctic Surface Water usually lies above a temperature minimum at 75–150 m depth, a temperature maximum at 250 m, and a salinity maximum at about 400 m. To account for these features, the following distinction was made: lower Arctic Intermediate Water lies immediately above deep water, includes the temperature and salinity maxima, and has both temperature and salinity decreasing with depth; upper Arctic Intermediate Water lies in between the temperature minimum and the temperature maximum and has both temperature and salinity increasing with depth; and Polar Intermediate Water has to some extent lower salinities than the other two intermediate waters but is largely distinguished by its association with overlying Polar Surface Water.

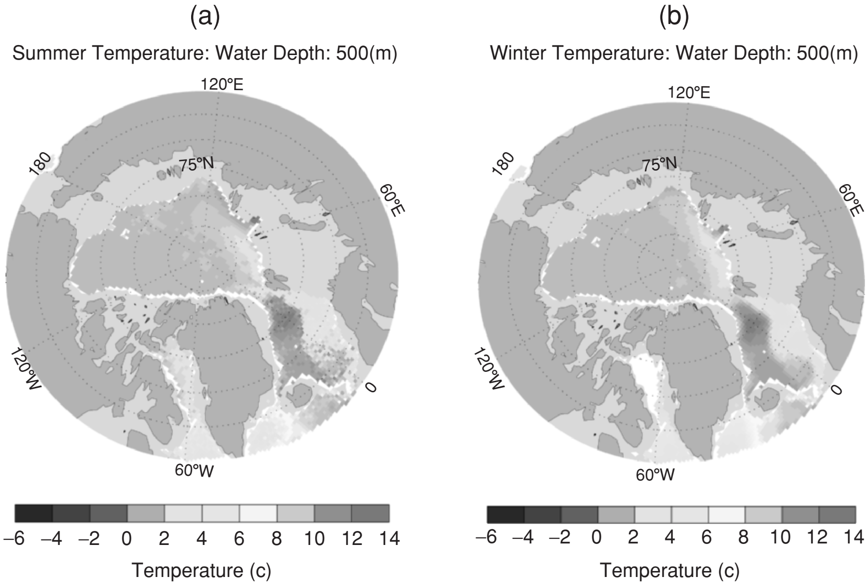

Fig. 14.4. Temperature of the Arctic Mediterranean Seas at 50 m depth from the Joint US-Russian EWG atlas: (**a**) summer, and (**b**) winter

Within the Arctic Ocean the surface layer overlies the relatively warm and saline water of Atlantic origin, carried into the Arctic through the Fram Strait by the West Spitzbergen Current. When this water enters the Arctic basin, the initially high temperature and salinity of the water (over 3°C and 35 ppt in the Fram Strait) rapidly decrease as it is cooled by the atmosphere and by mixing with local waters. When the current encounters the ice margin northwest of Spitzbergen, melting further cools and dilutes this water until its T/S properties approximate those of water found within the Atlantic layer throughout the Arctic (Perkin and Lewis 1984; Aagaard et al. 1987).

The ice breeze is generated near the marginal ice zone by the thermal inhomogeneity between the ice and water surfaces. Such a local circulation causes ice drift and in turn impacts on the oceanic motion and thermohaline structure (Chu 1986, 1987a, b, c). The Arctic Intermediate Water was discussed by Swift and Aagaard (1981) and Aagaard et al. (1985), whose core is identifiable over the entire basin at depths between 200 and 800 m by a temperature maximum (Fig. 14.4) and relatively high salinity (Fig. 14.5).

14.2.3 Deep Level (2,000 m Depth)

The lower portion of the water column is occupied by varieties of cold (<0°C) deep water. Our basic understanding of the origin of the deep waters has undergone considerable revision in recent years (Aagaard 1982; Swift et al. 1983;

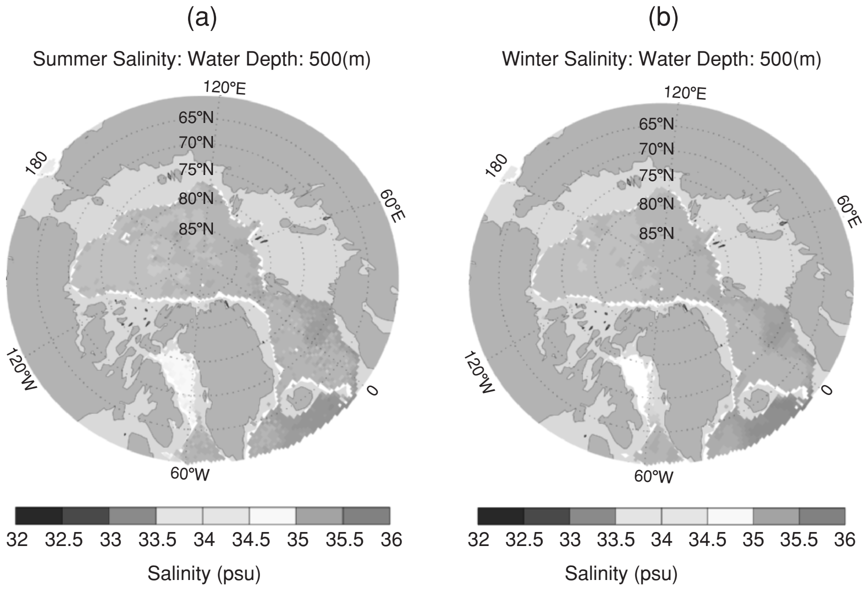

Fig. 14.5. Salinity of the Arctic Mediterranean Seas at 50 m depth from the Joint US-Russian EWG atlas: (**a**) summer and (**b**) winter

Aagaard et al. 1985; Rudels 1987; Swift and Koltermann 1988). Four basic varieties, each with distinctive T/S characteristics, are presently recognized. Greenland Sea Deep Water is the coldest (about $-1.2°$C) and freshest (<34.90 ppt) variety, and Canada Basin Deep Water is the warmest (about $-0.5°$C) and most saline (>34.95 ppt). In between are the Norwegian Sea Deep Water ($-0.9°$C, 34.92 ppt) and Eurasian Basin Deep Water ($-0.7°$C, 34.94 ppt) varieties.

14.3 Inverted Circulation in the Greenland–Iceland–Norwegian Sea

The different water masses encountered in the GIN Sea provide contrasting living conditions for zooplankton. The different faunas originating in the polar and Atlantic oceans are clearly distinguishable in the copepod population. Since the copepods cannot swim very well, their distribution primarily reflects the circulation pattern. From limited current meter observations and ecological measurements, general features of the GIN Sea circulation is presented in Fig. 14.6.

The GIN Sea is an important link between the Arctic Ocean and the North Atlantic, serving as a passageway for Atlantic waters streaming toward the Arctic waters, and Arctic waters passing in the East Greenland Current

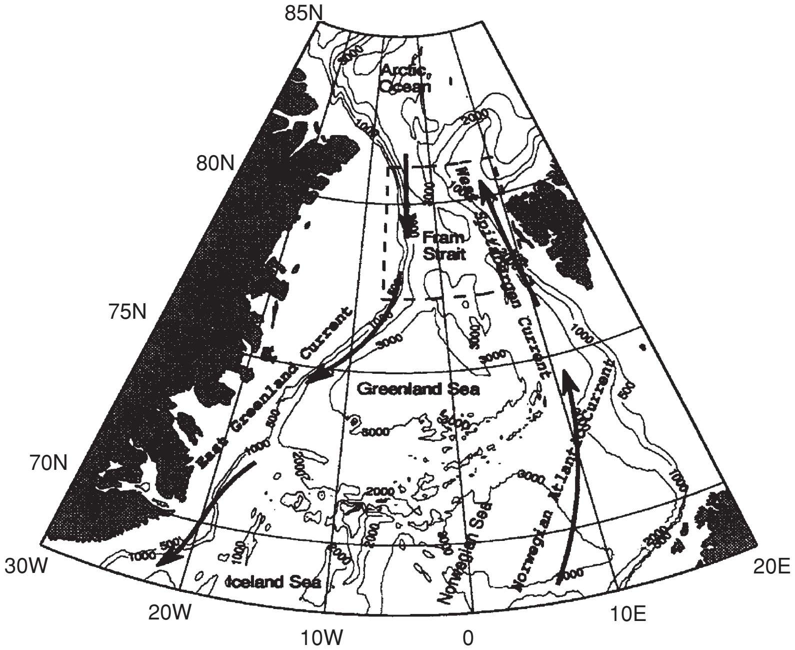

Fig. 14.6. Bathymetry of the GIN Sea and schematic circulation patterns

toward the Atlantic and the Labrador Sea. However, it also has its own important thermodynamic processes (particularly convection) that produce (or modify) deep waters that exit through the deeper parts of the Denmark Strait and the Faroe–Shetland Channel. To understand the role of the GIN Sea, the first step is to establish three-dimensional climatological velocity fields from the joint US-Russian EWG (T, S) data using the P-vector method.

With the (T, S) and corresponding velocity data, it is possible to estimate, through budget studies, the relative importance of the roles of the GIN Sea. In particular, a careful census is made of the different water masses entering and leaving, and their production in the basin due to interior mixing and atmospheric fluxes. The following subsections describe the major current characteristics inverted from the hydrographic data.

The seas north of the Greenland–Scotland Ridge constitute a major heat sink in the global thermohaline circulation of the world ocean and therefore a crucial component of the earth climate (Aagaard et al. 1985). A large heat loss to the atmosphere, combined with sea ice production and melting, is responsible for the formation of deep and intermediate waters through winter convection, which, in some basins like the Greenland Sea, may reach down to the bottom.

The one-dimensional convection process is strongly affected by the horizontal dynamics of the convective basins including lateral exchanges of heat, salt, and ice at their boundaries. Concerning the Greenland Sea, important exchanges occur through its northern boundary, the Fram Strait, which is the only deep connection between the Arctic Ocean and the rest of the world ocean. A better knowledge of the dynamics of the strait is therefore crucial in the context of investigating the processes involved in the deep convection and their variability.

14.3.1 Circulation Patterns

The inverted subsurface (50 m) velocity field shows the following pattern. Gyres are formed over each of the sub-basins in the GIN Sea (Fig. 14.7). In the northward flowing Norwegian Atlantic Current north of about 66°N commonly two maximum salinity cores are observed (e.g., Dietrich 1969; Hopkins 1991). Hopkins (1991) assumed that this bifurcation is caused by the bathymetric control of the barotropic flow when the current passes the Voring Plateau. The western branch veers northeastward on encountering the Hohn Ridge, converging with the Jan Mayen Current until it rejoins the eastern branch of the Norwegian Atlantic Current and becomes the West Spitzbergen Current. On its way north the temperature and salinity of the Norwegian Atlantic Water decreases mainly as a result of air–sea interaction, not by mixing with the underlying cooler and fresher water (Hopkins 1991).

Upon entering the Arctic basin, the West Spitzbergen Current encounters the southward-moving pack ice carried by the Transpolar Drift. Here, sensible heat carried by the West Spitzbergen Current is used to melt the ice, resulting in a cooled and freshened upper layer (Quadfasel et al. 1987; Untersteiner 1988; Moore and Wallace 1988). This point of encounter is also a zone of maximum density at the surface. Before reaching this point, the surface cooling has increased the density of the surface layer; later, melting begins to decrease the surface density.

The southward flowing East Greenland Current extends to around 2,500 km along the Greenland coast. In addition to transporting surface, intermediate, and deep water, the East Greenland Current exports sea ice from the Arctic, removing between 4,000 and $5,000\,\mathrm{km}^3$ of ice each year, an amount approximately equal to the freshwater inflow (Wadhams 1983; Vinje and Finnekasa 1986). There are two branches off the mainstream of the EGC: a relatively minor one at about 77°N, which strikes southeast along the Greenland Fracture Zone, and a larger one called the Jan Mayen Current that branches eastward at about 73°N to form the southern margin of the Greenland Gyre (Fig. 14.7). This pattern is more evident in summer (Fig. 14.7a) than in winter (Fig. 14.7b).

Vertical cross sections of v-velocity (north–south) component along 75, 70, and 65°N for summer (Fig. 14.8) and winter (Fig. 14.9) clearly show the existence of the two basic currents with baroclinic characteristics: northward flowing Norwegian Atlantic Current (see 65°N cross section) and becoming the West Spitzbergen Current (see 70, 75°N cross sections) in the eastern

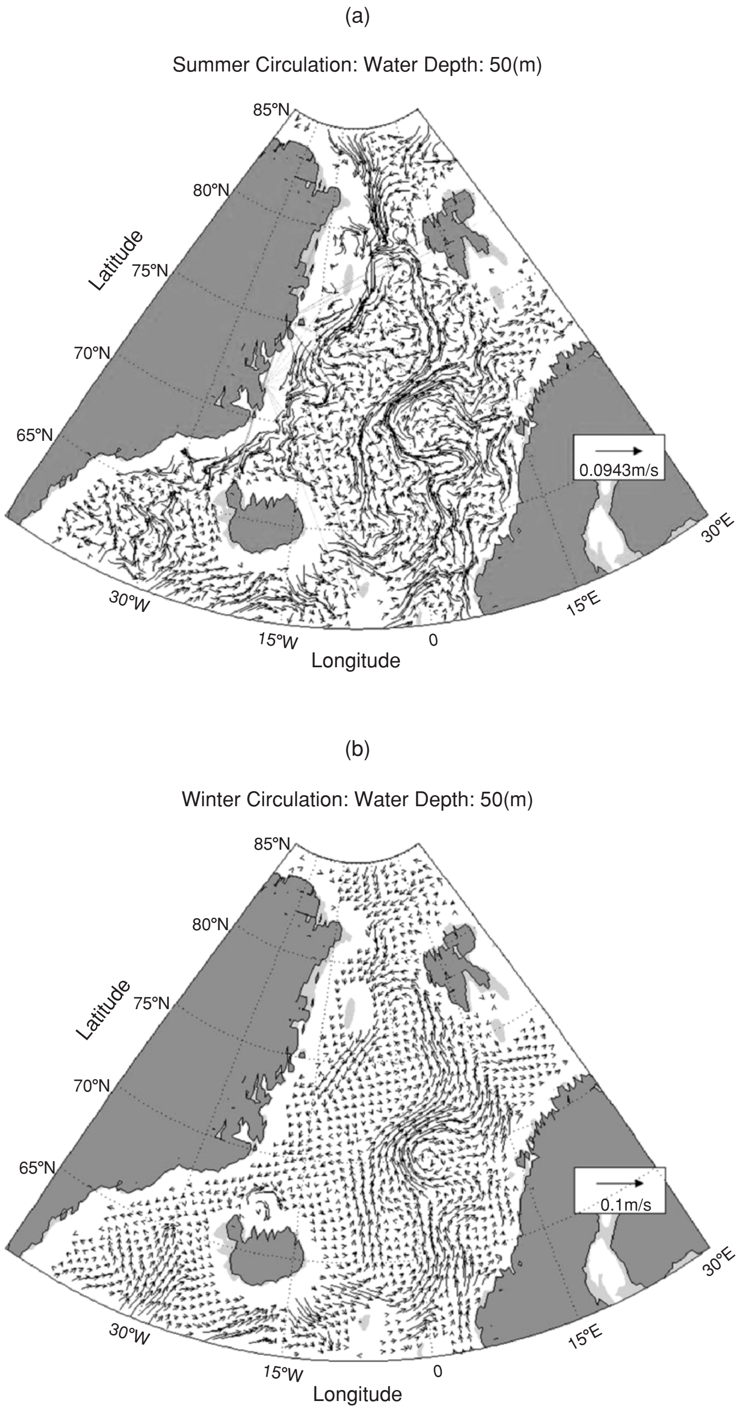

Fig. 14.7. Inverted subsurface (50 m depth) velocity vectors in the GIN Sea: (**a**) summer and (**b**) winter

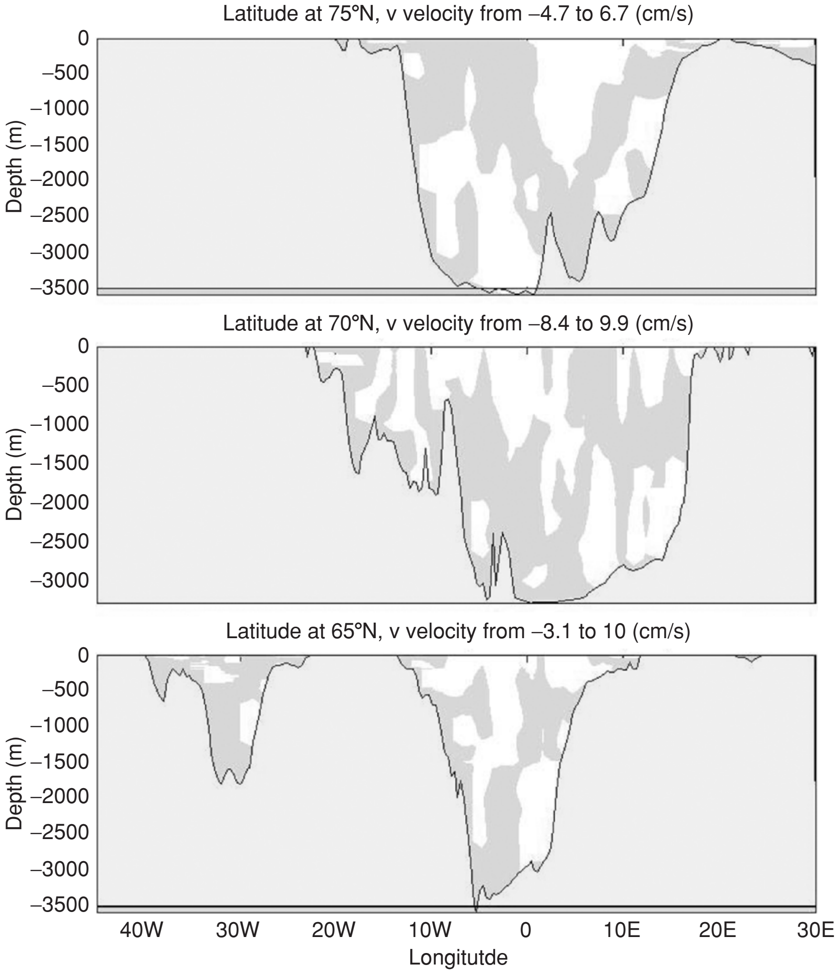

Fig. 14.8. Vertical cross sections of v-component (north–south) in summer along 75, 70, and 65°N. Here, *unshaded* areas refer to positive v (i.e., northward motion) and *shaded* areas refer to negative v (i.e., southward motion). The maximum value of the northward motion is $6.7\,\mathrm{cm\,s^{-1}}$ at 75°N, $9.9\,\mathrm{cm\,s^{-1}}$ at 70°N, and $10\,\mathrm{cm\,s^{-1}}$ at 65°N. The maximum value of the southward motion is $4.7\,\mathrm{cm\,s^{-1}}$ at 75°N, $8.4.\,\mathrm{cm\,s^{-1}}$ at 70°N, and $3.1\,\mathrm{cm\,s^{-1}}$ at 65°N

GIN Sea and southward flowing East Greenland Current in the western GIN Sea.

The current systems are stronger in summer than in winter. For example, the maximum speeds of the northward flowing Norwegian Atlantic Current and West Spitzbergen Current vary from ($6.7\,\mathrm{cm\,s^{-1}}$ at 75°N, $9.9\,\mathrm{cm\,s^{-1}}$ at 70°N, and $10\,\mathrm{cm\,s^{-1}}$ at 65°N) in summer to ($4.8\,\mathrm{cm\,s^{-1}}$ at 75°N, $6.6\,\mathrm{cm\,s^{-1}}$

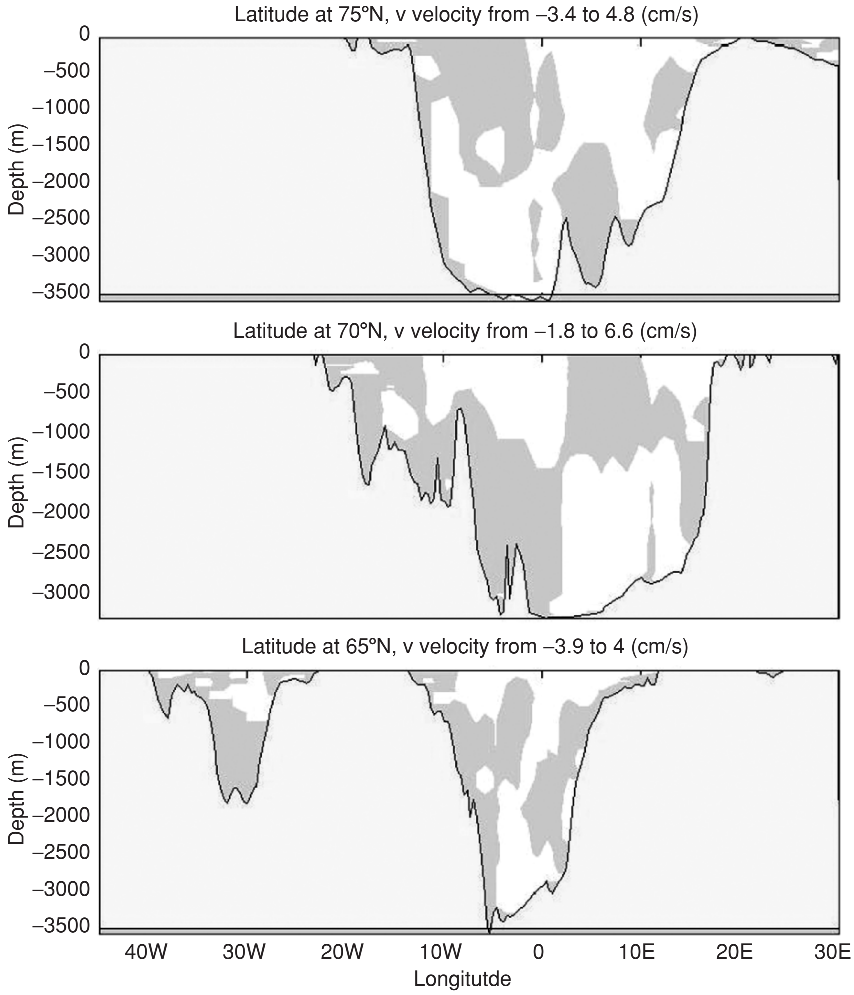

Fig. 14.9. Vertical cross sections of v-component (north–south) in winter along 75, 70, and 65°N. Here, *unshaded* areas refer to positive v (northward motion) and *shaded* areas refer to negative v (southward motion). The maximum value of the northward motion is 4.8 cm s^{-1} at 75°N, 6.6 cm s^{-1} at 70°N, and 4 cm s^{-1} at 65°N. The maximum value of the southward motion is 3.4 cm s^{-1} at 75°N, 1.8 cm s^{-1} at 70°N, and 3.9 cm s^{-1} at 65°N

at 70°N, and 4 cm s^{-1} at 65°N) in winter. The maximum speeds of the southward flowing East Greenland Current vary from (4.7 cm s^{-1} at 75°N, 8.4 cm s^{-1} at 70°N, and 3.1 cm s^{-1} at 65°N) in summer to (3.4 cm s^{-1} at 75°N, 1.8 cm s^{-1} at 70°N, and 3.9 cm s^{-1} at 65°N) in winter. Southward overflow in Denmark Strait is also identified from the vertical cross section of v-velocity along 65°N in summer (Fig. 14.8) and in winter (Fig. 14.9).

14.3.2 Fram Strait Exchange

The detailed flow pattern in the strait, however, is more complex and numerous recirculations with large spatial and temporal variability (Hopkins 1991) make reliable estimates of the transports through the strait more difficult.

Previous estimates of the transports through the Fram Strait based on hydrographic measurements mostly rely on the baroclinic component of the geostrophic currents (e.g., Timofeyev 1962). Direct current measurements are extremely sparse and contaminated by the mesoscale activity (Foldvik et al. 1988). Lagrangian observations can give insight into the circulation (Gascard et al. 1995) but are unable to provide transport estimates. Numerical models, which include the northern Greenland Sea, either have a highly coarse resolution in view of the complex bottom topography of the strait (e.g., Gerdes and Schauer 1997) or, considering a restricted domain, use the Fram Strait as an open boundary so that currents or transports cannot be reasonably predicted in the strait (e.g., Legutki 1991).

The P-vector inverse method combines hydrographic information with relevant constraints and offers an alternative approach. The inverted surface velocity shows the flow pattern through the Fram Strait. Two main currents exchange water between the Arctic and the world ocean (Fig. 14.10). On the eastern side of the strait, the northward West Spitzbergen Current carries relatively warm and salty waters of Atlantic origin above relatively cold and fresh deep waters that is formed in the Greenland and the Norwegian Seas. On the western side, sea ice and cold and fresh surface water are exported from the Arctic Ocean in the East Greenland Current above relatively warm and salty deep waters.

Recirculation of Atlantic water is also detected from the inverted velocity field (Fig. 14.10). Much of it joins the southward-moving East Greenland Current to flow back into the Greenland and Iceland seas. The continuation of the West Spitzbergen Current into the Arctic is also complex, as the current appears to split into two or more branches. This is consistent with earlier studies by Perkin and Lewis (1984), Aagaard et al. (1987), and Quadfasel et al. (1987).

Upon entering the Arctic basin, the West Spitzbergen Current encounters the southward-moving pack ice carried by the Transpolar Drift. Here, sensible heat carried by the West Spitzbergen Current is used to melt the ice, resulting in a cooled and freshened upper layer (Quadfasel et al. 1987; Untersteiner 1988; Moore and Wallace 1988). This point of encounter is also a zone of maximum density at the surface. Prior to reaching this point, surface cooling has increased the density of the surface layer; afterward, melting begins to decrease the surface density.

14.4 Inverted Circulation in the Arctic Ocean

The inverted annual mean surface velocity vector field in the Arctic Ocean shows two major characteristics (Fig. 14.11). The first is the Transpolar Drift,

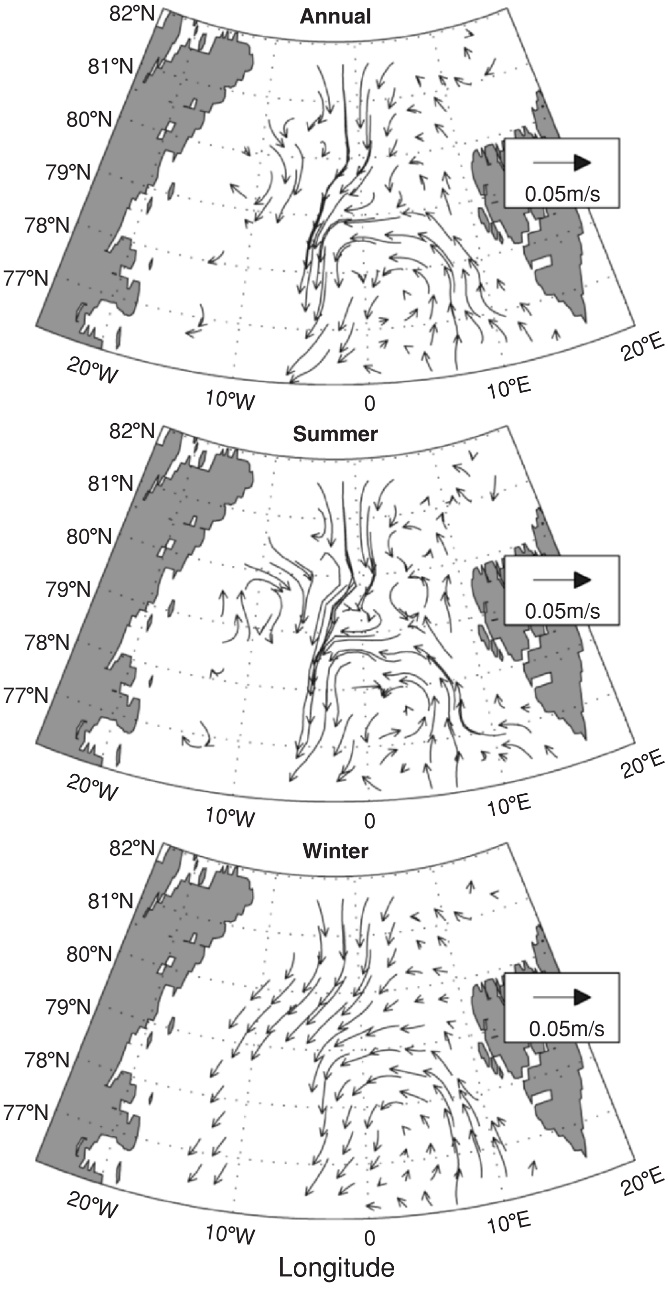

Fig. 14.10. Inverted surface velocity vectors in the Fram Strait from annual mean, summer, and winter EWG (T, S) data. Grid points without inverted velocity shows the necessary conditions for the P-vector method that are not satisfied. The recirculation of Atlantic Water through the West Spitzbergen Current is also identified

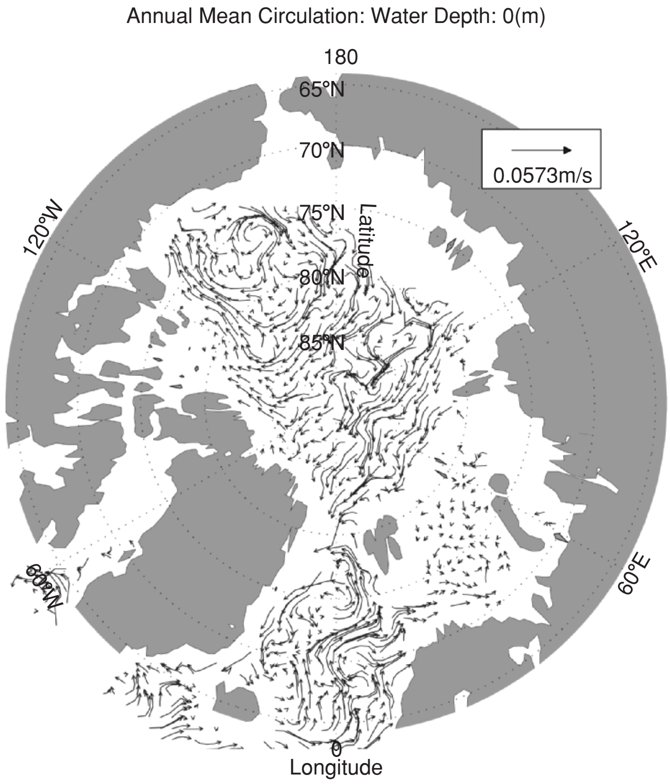

Fig. 14.11. Inverted surface velocity vectors in the Arctic Ocean from annual mean EWG (T, S) data. Grid points without inverted velocity shows the necessary conditions for the P-vector method that are not satisfied. The recirculation of Atlantic Water through the West Spitzbergen Current is also identified

in which the surface waters of Eurasian basin move across the basin toward the North Pole and then toward the Fram Strait; the second is the anticyclonic flow around the Beaufort Gyre in the Canadian basin. These two features are consistent with earlier studies (Coachman and Aagaard 1974; Gorschkov 1983). The maximum current speed is around $5\,\mathrm{cm\,s^{-1}}$. Mean current speeds are slow in the central ocean, about $2\,\mathrm{cm\,s^{-1}}$, but increase as water exits the basin as part of the East Greenland Current, which is consistent with Wadhams et al. (1979) as well as Vinje and Finnekasa (1986).

As the West Spitzbergen Current passes through the Fram Strait and subsides, it appears to branch. North of 79°N, where the 200-m and deeper isobaths diverge, the current splits into two main cores. The western or offshore branch follows the western flank of the Yermak Plateau. North of 80°N a portion of this flow again splits off to contribute to the recirculation within the East Greenland Current. The eastern or in-shore branch of the West Spitzbergen Current follows the shelf break around Spitzbergen and into the Arctic Ocean. During its transit it is cooled and freshened by mixing with overlying waters, transforming the original Atlantic water into Arctic intermediate water. The inverted velocity is smaller (Fig. 14.12, 14.13) than the directly measured value by Aagaard (1989), who noted eastward flow that increased with depth to typical speeds of 0.2–0.3 $\mathrm{m\,s^{-1}}$.

Although the southern Beaufort Sea is generally thought of as an area of westward (clockwise) water and ice motion (Fig. 14.11), the average subsurface motion above the continental slope is in the opposite direction (Fig. 14.13). Aagaard (1984) called this flow the Beaufort Undercurrent. The presence of the undercurrent is indicated by a subsurface maximum in temperature caused by the eastward flow of water originating in the Bering Sea. Aagaard (1984) described the Beaufort Undercurrent as being a topographically steered eastward flow extending seaward of the 50 m isobath out to the base of the continental slope. Speeds are of the order 0.1 $\mathrm{m\,s^{-1}}$ and increase with depth down

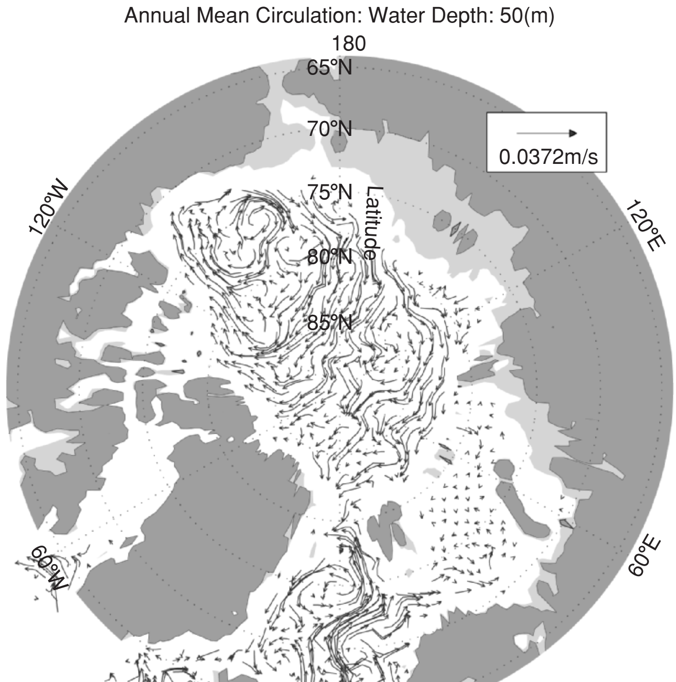

Fig. 14.12. Inverted velocity vectors at 50 m depth in the Arctic Ocean from annual mean EWG (T, S) data. Grid points without inverted velocity shows the necessary conditions for the P-vector method that are not satisfied. The recirculation of Atlantic Water through the West Spitzbergen Current is also identified

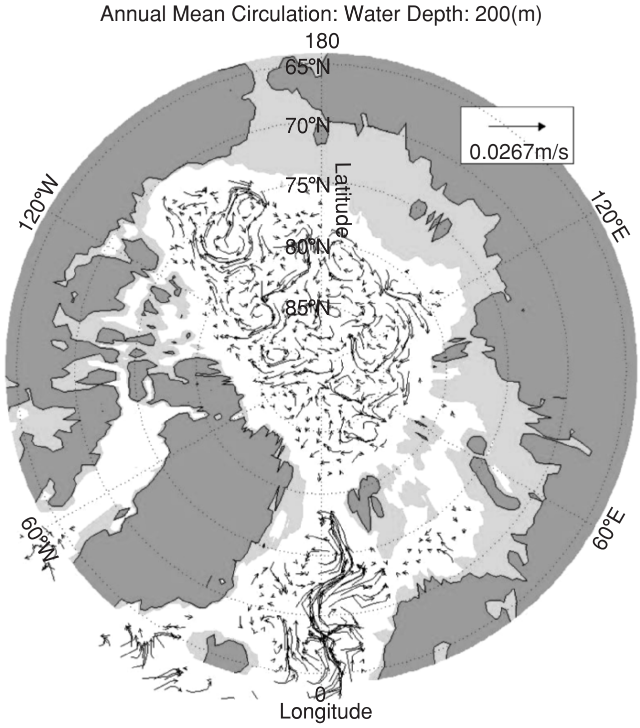

Fig. 14.13. Inverted velocity vectors at 200 m depth in the Arctic Ocean from annual mean EWG (T, S) data. Grid points without inverted velocity shows the necessary conditions for the P-vector method that are not satisfied. The recirculation of Atlantic Water through the West Spits Bergen Current is also identified

to about 150 m. Transports are of the order 1 Sv. The current is probably part of the large-scale circulation of the Canada basin and thus not locally driven.

A possible forcing mechanism for undercurrents in the Arctic has been discussed by Holloway (1987). He argues that the interaction of eddies with along-shore variations in topography, together with coastally trapped planetary wave propagation, result in a systematic forcing that acts on the mean flow. Applied to the Arctic, this mechanism predicts an eastward (cyclonic) flow following the basin margins of similar magnitude to the Beaufort Undercurrent.

Questions and Exercises

(1) Why is the GIN Sea circulation important for the climatic and ecological systems?
(2) Download the US-Russian EWG and WOA (T, S) data for the Arctic Ocean. Compare the differences between the two datasets.

(3) The WOA dataset uses the spherical coordinate system. Do you think it is good for the Arctic Ocean? Why?
(4) What coordinate system does the US-Russian EWG dataset use for the Arctic Ocean?
(5) What are the major characteristics of the GIN Sea thermohaline structures?
(6) What are the major features of the northward flowing West Spitzbergen Current and southward flowing East Greenland Current? What are the roles of these currents in global ocean circulation and in turn in the climatic system?
(7) What are the major features of the Fram Strait recirculation? Discuss its effect on the Arctic circulation.
(8) What are the major features (including seasonal variation) of the Arctic circulation?
(9) Select an area with your interest. Download the data of the volume transport stream function, absolute velocity at z-coordinate, and absolute velocity at isopycnal coordinate from the DVD-ROM. Discuss the major characteristics of the circulation in the area including the seasonal variation, causes of these features, and the effect of the circulation on the global system.