On the coupling between hydrography and larval transport in Southwest Greenland waters

Ph.D. thesis
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This Ph.D. thesis can be found on: http://ocean.dmi.dk/staff/mhri/Docs/PhD.html

FrontPage: Southwest Greenland coast seen from the inner station on the Cape Desolation transect (about 61°N) onboard the Danish naval ship Agdlek in early June 2002 (Buch and Ribergaard, 2003). Small icebergs and growlers are seen in front of the coast of Greenland. Picture taken by Mads Hvid Ribergaard.

Abstract

This thesis investigates the coupling between climate, hydrography and recruitment variability of fishery resources off West Greenland with special emphasis on cod and shrimp recruitment. The main hypothesis investigated is, that the variability of these stocks at West Greenland is highly controlled by advection of water masses into the area during their pelagic stages. This hypothesis is investigated through three different studies:

1. A review of the past 50 years of climatic conditions off West Greenland is given based on hydrographic and atmospheric observations off West Greenland. Relationships between the past variations in fisheries resources, hydrographic conditions, and the large-scale climatic conditions, expressed by the North Atlantic Oscillation (NAO), are found. A reduction of the inflow of heat and salt to the West Greenland area during positive NAO phase from the late 1960s until the mid-1990s reduces the recruitment of cod larvae from Iceland. Mature cod migrated out of the area reducing the predator pressure on the shrimp stock. By-catches of cod by the shrimp fishery contributed to keeping the predation pressure low.
2. Shrimp larvae transport from larval release to settling at the bottom is studied, using a particle-tracking model forced by a high resolution ocean circulation model. Residual anticyclonic eddies are generated around the shelf banks north of 64°N which largely affect the retention times at the shelf banks. Similar, plankton distributions were related to the physical environment described by observations and the same ocean circulation model. Areas of permanent upwelling west of the shelf banks found in the model are suggested to increase the productivity of plankton.
3. Transport of cod eggs and larvae from the Southwest Icelandic spawning grounds to the Greenland waters is studied for the period 1948–2001 using model based particle tracking forced by a regional nested ocean model with horizontal resolution of 20–25 km. Surprisingly, no relation was found between the drift of recruits from Iceland to Greenland and the observed year class strength of West Greenland cod. However, we can not reject the hypothesis. A validation of the ocean model against observations reveals that the model is far from perfect and even lacking important events as the Great Salinity Anomaly in the late 1960s in Greenland waters. Results from the drift model suggest that the drift of cod larvae towards Greenland is very sensitive to the position of the spawning grounds.

On the coupling between hydrography and larval transport in Southwest Greenland waters

1 Processes responsible for major variations in the marine ecosystem

1.1 Productive areas – a general view

Most of the world's ocean can be regarded as biological deserts. The reason for this is in principle two-fold (Bakun, 1996):

1. Sunlight typically penetrates only a few tens of meters into the waters and therefore photosynthesis can only take place near the sea surface.
2. Almost everywhere the ocean is stably stratified and therefore nutrients tend to remain trapped in the deeper layers.

Plant cells use nitrate, phosphate and other nutrient components dissolved in the ocean surface layer when producing organic matter. This organic matter is later exported to deeper layers by various processes. The dead plant cells may sink when they die or other organism may consume the plant cells and expel the surplus organic matter as fast sinking fecal pellets. The organism may also migrate vertically and get eaten at depths. The net result is a transport of nutrients away from the illuminated surface waters towards deeper layers. Over shallow continental shelves and banks the organic matter can only sink to the depth of the sea bottom. This may be close or even within the photic zone.

Processes exists that enhance the biological productivity by adding nutrients to the photic zone. In particular the four eastern boundary currents of the subtropical gyres in the Pacific and the Atlantic Oceans are very productive, caused by upwelling of nutrient rich water. The Peruvian anchovy fishery was the most productive fishery in the world in the 1970s (Stensted, 2002) and in 1986 over 20 percent of the worlds landings were caught in the Chile-Peru Current system (Yáñez et al., 2001).

Bakun (1996) summarized a number of physical processes that result in an enrichment of the photic zone by nutrients:

- Coastal upwelling caused by an offshore (wind-driven) Ekman transport.
- Shelf-break upwelling. This can be generated by several mechanisms such as bottom friction, bottom Ekman transport, eddies, coastal trapped waves and (breaking) internal tides.
- Upwelling at the sea-ice edge caused by wind-driven Ekman transport.
- Mixing through a stationary thermocline separating the summer mixed layer from the deeper layers. The thermocline is build up by an equilibrium between the solar radiation from above that tends to deepen the thermocline, whereas mixing processes from below that tend to break it down resulting in an upward mixing of deeper waters towards the warm surface layer.
- Tidal mixing over shallow regions.
- Winter convection mixes the waters down to the depth of convection.
- Equatorial upwelling caused by poleward Ekman transports to the south and north forced by the trade winds.
- Ekman pumping generated by a wind stress curl.

- Eddies. Eddies formed between two different watermasses mixes these and thereby also the nutrient content. Moreover in a cyclonic eddy, deeper water is lifted towards the surface by upwelling.
- Coastal and river runoff adding nutrients from land.
- Mixing at fronts between two water masses where the denser water mass subducts under the lighter water mass. Fronts can range from local to large scale.
- Wind mixing.

These can be divided up into three main classes:
- Enrichment processes which make nutrients available to biological production.
- Concentration processes which enhance the availability of food for a predator through increasing the concentration of nutrients.
- Retention processes which help individual animals to be in an appropriate position suitable for them.

1.2 Large scale climate variability on decadal timescales affecting the marine ecosystem

Maybe the most remarkable variability in the oceans fishery is the altering shifts between sardine and anchovy regimes. These regime shifts have occurred more or less simultaneous in four of the ocean's coastal regions: Off Japan and in the California, Humboldt and Benguela current systems, with the Benguela system out of phase with the three Pacific systems (Lluch-Belda et al., 1992).

It is widely believed, that large scale climate variability influences the marine ecological processes through a variety of local factors such as temperature, wind, rain, sea-ice and ocean currents as well as a combination of these. The best known and the strongest climate variations on decadal timescale are the El Niño–Southern Oscillation (ENSO) and the North Atlantic Oscillation (NAO) (e.g. Stenseth et al., 2002; Stenseth et al., 2003).

Like in the Pacific, an El Niño like oscillation exists in the Atlantic Ocean, called the Benguela Niño. Similar, the North Pacific index (NP-index) is the Pacific pendant to the NAO. Its variations is commonly described by the Pacific Decadal Oscillation (PDO), which are defined using sea-surface temperature instead of sea level pressure, but both clearly capture much of the same climate fluctuations (Stenseth et al., 2003).

Here some examples are given on the marine responses on these natural climate fluctuations, except for the NAO, which is described in Chapter 2.

1.2.1 North Pacific index / Pacific Decadal Oscillation (PDO)

Catches of various Pacific salmon have shown similar variability with decadal timescales of 20–30 year. However, catches of Alaska salmon and West American salmon were out of phase (Figure 1). Alaska salmon is more abundant during warm conditions, less abundant during cold conditions, while salmon in the Northwest shows the opposite pattern (Mantua et al., 1997).

Both the Alaska and West American stocks are believed to be dependent on upwelling resulting in enhanced primary production and increased zooplankton biomass and on increased precipitation over land which increases the streamflows in the rivers (Mantua et al., 1997). However, the upwelling mechanism is quite different. West American salmon benefits from wind-driven coastal upwelling within the California Current, while the Alaska stock is affected by Ekman pumping and wind-generated mixing over the shelf area (Hare, 1996; Mantua et al., 1997). These authors showed that the decadal variations are closely connected to the Pacific Decadal Oscillation (PDO), which has similar variations as the North Pacific index (NP index) but on longer timescales.

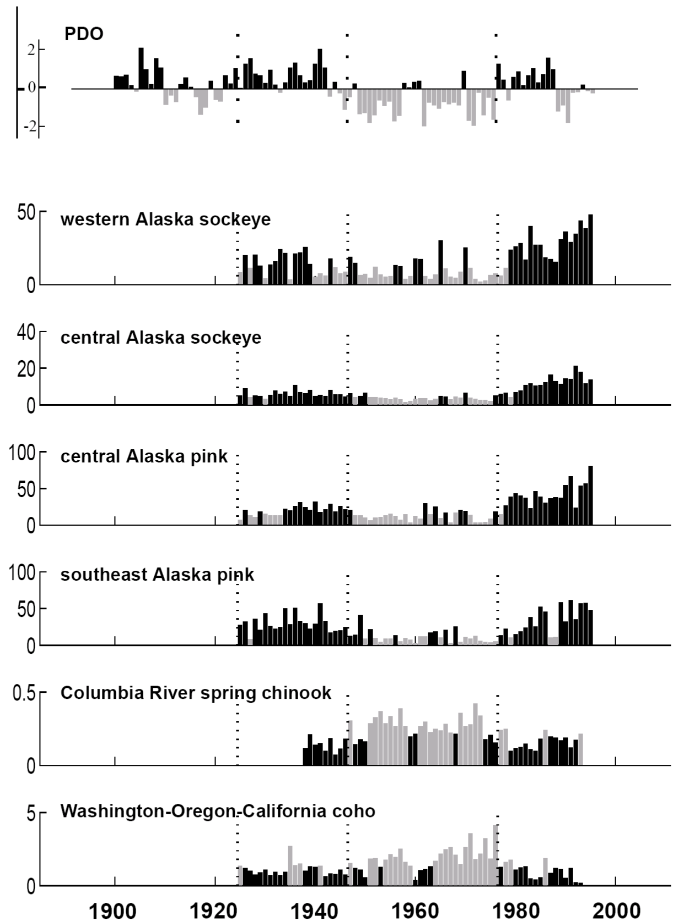


Figure 1. Top: Pacific Decadal Oscillation (PDO). Bottom: Selected Pacific salmon catch records. The black (grey) shadings denote values that are greater (lower) than the long-term median. Note, the shading has been reserved for the two lower figures. Dotted vertical lines are the PDO reversal times 1925, 1947, and 1977. From Mantua et al. (1997).

The NP index is defined as the area-weighted sea level pressure over the North Pacific region 30N–65N, 160E–140W (Trenberth and Hurrell, 1994). It describes the intensity of the Aleutian Low - the Pacific pendant to the Icelandic low in the Atlantic Ocean – which accounts for the largest portion of variability in the Pacific. In periods of deep Aleutian Lows, the cyclonic atmospheric circulation speeds up which affects both air temperature and precipitation and temperature and currents in the ocean. Air temperatures over the western part of the North Pacific experience anomal warm conditions with dominating winds from south caused by an increased northerly thermal advection, while the central and eastern sides are cooled by air masses from the north (Trenberth and Hurrell, 1994). Strictly, it is the gradient of the pressure that determines the strength of the geostrophic flow. However the variance of the Aleutian low is much higher than the variance of the surrounding higher pressure, and it is therefore a good proxy for the climate conditions.

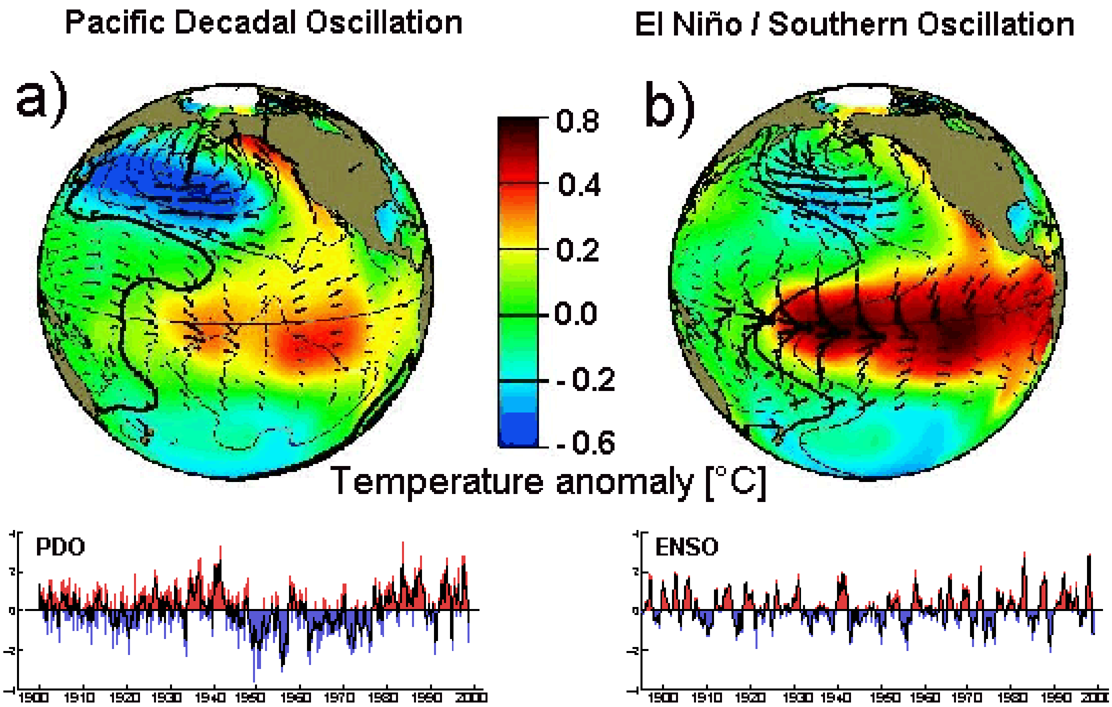


Figure 2. Top: Comparison of typical warm sea surface temperature in degrees (colours), sea level pressure (contours), and wind anomalies (vectors) for PDO (a) and ENSO (b). Bottom: Timeseries of PDO and ENSO for the period 1900–2000. Modified from: http://tao.atmos.washington.edu/pdo/graphics.html.

The variations in the North Pacific are sometimes described by the PDO, which are defined using sea-surface temperature instead of sea level pressure, but both clearly capture most of the same climate fluctuations (Stenseth et al., 2003). The PDO was defined by Hare (1996), as the leading principal component of monthly sea surface temperature variability in the Pacific poleward of 20°N. It describes long-term fluctuations of the ocean in the North Pacific (see Figure 2a).

A well described regime shift from cold to warm conditions in the Northeast Pacific was observed in 1976–1977, and a shift back to cold conditions in 1988–1989 (Trenberth and Hurrell, 1994; Hare, 1994; Mantua et al., 1997; Francis et al., 1998; McFarlane et al., 2000; Hare and Mantua, 2000; Mantua and Hare, 2002; Bogard and Lynn, 2003). Long timeseries of PDO and NP reveal additional two regime shifts in the past century (Mantua and Hare, 2002).

Mantua et al. (1997) suggested that during a warm PDO period, the subpolar gyre of Alaska is spun up as sketched in Figure 3. Thereby upwelling is enhanced and the primary production and the zooplankton biomass, which salmon feeds on, are increased. Further, precipitation is increased over Alaska resulting in higher streamflows which favour survival of the juvenile salmon. Contrary, during a warm PDO period, the California Current is characterized by increased stratification and a deepening of the thermocline, which makes upwelling less efficient in mixing nutrients to the surface layers as shown by Bogard and Lynn (2003). This has a negative effect on the Northwest salmon stock. Moreover, as sketched in Figure 3, the warm PDO period is characterized by decreased precipitation over East America resulting in lower streamflows which has a negative effect on the survival of juvenile salmon.

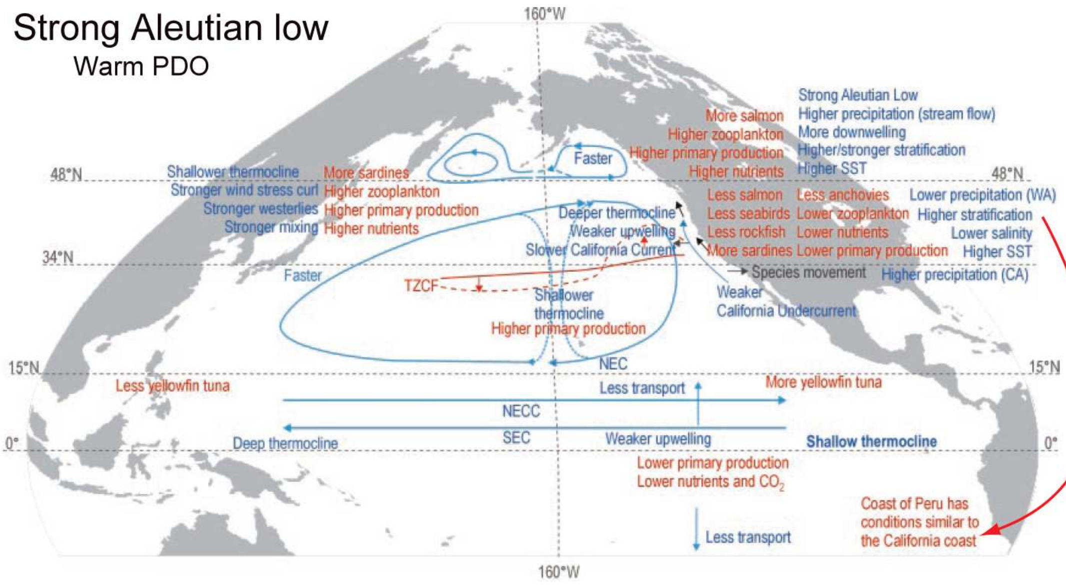


Figure 3. Schematic showing the general conditions in the Pacific during an intensification of the Aleutian low corresponding to high North Pacific index and a "warm" Pacific Decadal Oscillation. Physical processes are in blue, and red are biological and chemical changes. Modified from Chavez et al. (2003).

It has been speculated that the regime shift described by NP and PDO is actually connected to the frequency and intensity of El Niños in a given decade (Trenberth and Hurrell, 1994; Greene, 2002). The spatial patterns are much alike, though with different intensity, and the area of influence overlap (see Figure 2), but this hypothesis still needs to be tested.

Gu and Philander (1997) came up with a physical mechanism that may explain decadal variations simultaneous on both hemispheres in the tropics and subtropics, based on the theory of the ventilated thermocline (Luyten et al., 1983; Pedlosky, 1998). They suggest that anomal warm water subducted in the extratropics flows southwestward along isopycnals and are found at the surface at the equator several years later, caused by equatorial upwelling. Here it affects the climate similar to El Niño events (see Chapter 1.2.2). Eventually the warm anomaly moves back towards the subtropics in the surface layer to complete one circuit of the meridional overturning. The subduction acts on decadal timescales similar to the PDO. This theory was later modified by Kleeman et al. (1999), who suggested a dynamically changing ocean circulation, which speeds up or slows down the meridional circulation so that water masses with relative constant properties are subducted at different rates. Such decadal fluctuation of the meridional overturning was observed in the Pacific by McPhaden and Zhang (2002).

1.2.2 El Niño / Southern Oscillation (ENSO)

In the Humboldt Current along the South American coast a regime shift from a fishery on anchovy (cold) to a fishery on mainly sardine (warm) but also mackerel is observed (Bakun, 1996; Chavez, 2003; Ñiquen and Bouchon, 2004) as shown in Figure 4. It has been shown, that the abundance of pelagic fishes is mainly controlled by long term climate variability, even taking into account the fishing pressure (Yáñez et al., 2001).

Similar altering regime shifts are also seen in the California Current off North America (Lluch-Belda et al., 1992; Chavez, 2003). A 250-year sediment core taken in central Gulf of California reveals a series of regime shifts in the California Current fish population, with a clear tendency of the anchovy and sardine to fluctuate out of phase (Holmgren-Urba and Baumgartner, 1993). By comparing it to shorter timeseries outside the Gulf within the California Current, it seems that the regions undergo the same regime shifts with sardine preferring warm conditions and anchovy preferring cold conditions, but some migration between the regions are also seen on decadal timescales.

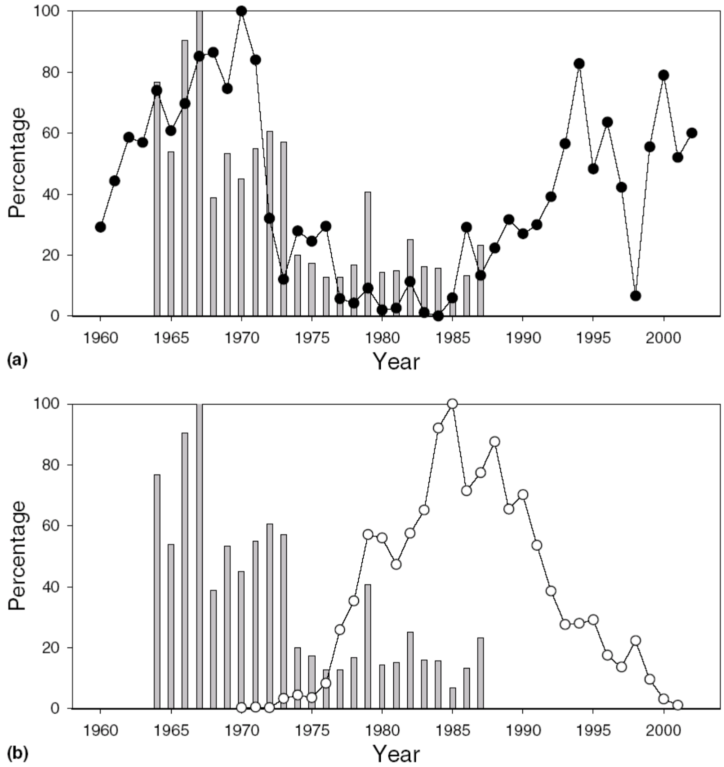


Figure 4. Annual variation of a) anchovy and b) sardine expressed as percentage. Columns are zooplankton volumes, also in percentage. For anchovy are used the northern and central Peruvian stock, whereas the northern and central Peruvian, and southern Peruvian and northern Chilean stock are used for sardine. Zooplankton volumes were calculated by summing all sub-regions for the different seasons in Peruvian waters. From Alheit and Ñiquen (2004).

These regime shifts can to a large extent be explained by fluctuating thermocline depths and thereby changes in the wind-driven upwelling of cold nutrient rich water. This is connected to eastward advection of warm water masses across the Pacific Ocean along the eastern coast. However, it is not the single El-Niño event that is responsible for the regime shift, but rather the persistent shifts between cold and warm phases on decadal timescales, which can be linked to the strength and frequency of El Niño events (Alheit and Ñiquen, 2004). A single event can abruptly affect the populations dramatically, but it seems that the anchovy is able to recover rather quickly a couple of years ahead. In contrast, the re-arrangement of the entire ecosystem during a regime shift takes several years and involves the whole food chain from phytoplankton to the top predators. The regime shifts on decadal timescales is likely connected to the PDO (Chapter 1.2.1), whereas large fluctuations on smaller timescales of a few years is associated with the El Niño phenomenon (Chavez et al., 1999; Chavez et al., 2003; Alheit and Ñiquen, 2004) like the 1997–1998 event. Trenberth and Hurrell (1994) and Greene (2002) speculated if PDO is actually connected to the frequency and intensity of El Niños in a given decade, but this remains to be tested.

El Niño events are large climate variations in the tropical Pacific Ocean, which occur about every 3 to 7 years with different strength (Trenberth, 1997). These events results in the strongest decadal oscillation in sea surface temperature observed on Earth. In the tropical Pacific Ocean, trade winds generally push and pile the surface waters westward. The surface water becomes gradually warmer as it is going westward because of its longer exposure to solar heating. Thermal expansion and wind stress results in a sea-level difference across the Pacific. El Niño is observed when the easterly trade winds weaken, allowing the warmer water of the western Pacific to migrate eastward as an equatorial Kelvin wave until it eventually reaches the South American coast and continues poleward in both hemispheres as a coastal Kelvin wave (e.g. Bakun, 1996).

In a normal year, there is a large pool of warm water in the western Pacific. Low air pressure dominates in the western Pacific over the warm water and high pressure dominates in the eastern Pacific over the colder water. During an El Niño event the pressure anomaly is reversed. This shift in surface pressure is called the El Niño Southern Oscillation (ENSO) and has a large impact on the precipitation distribution around the Pacific Basin. Figure 2b shows the typical sea surface temperature anomaly during an El Niño event. The southeast trade winds are weakening and the upwelling in the eastern Pacific is reduced. The cool nutrient-rich sea water normally found along the coast of Peru is replaced by warmer water depleted of nutrients which increase the sea-level but lower the depth of the thermocline. Consequently any upwelling that does occur is unable to tap into the rich nutrients found in deeper waters, resulting in a dramatic reduction and shift in the marine ecosystem (e.g. Barber and Chavez, 1983; Barber and Chavez, 1986; Ñiquen and Marilú, 2004).

During positive temperature anomalies the anchovy population migrates closer to the coast as the coastal upwelling is the most effective in bringing cold and nutrient rich water to the surface. Thereby the extent of their distribution and spawning grounds is reduced, which increases their catchability, increases cannibalism on eggs and larvae and further increases the spatial overlap with sardines, which feed on anchovy eggs. Moreover reduced upwelling leads to heavily reduced phyto- and zooplankton production which juvenile and adult anchovies feed on.

1.2.3 Benguela Niño – the Atlantic El Niño

Like in the Humboldt Current off South America, the ecosystem in the southern Benguela Current off South Africa has undergone regime shifts altering between anchovy-dominated and sardine-dominated states (Shannon et al., 2004), but out of phase (Lluch-Belda et al., 1992).

Sardine and anchovy feed on different zooplankton species which are dependent on the intensity of the wind-driven upwelling (Shannon et al., 2004; Cury and Sharron, 2004). The regime shifts in the southern Benguela ecosystem are most commonly believed to be connected with changes in upwelling. Additional the southern Benguela Current system is also highly affected by advection of warm water from the Indian Ocean. Therefore, in contrast to the boundary currents of America, advection of cold surface waters are not contributed from south (Shelton et al., 1985).

In the Atlantic Ocean a similar phenomenon as the El Niño exists called the Benguela Niño (Shannon et al., 1986; Boyer et al., 2000). The governing physics are believed to be basically the same, but the events are less frequent (Bakun, 1996; Boyer et al., 2000). During a Benguela Niño warm water crosses the Atlantic Ocean along the equator in an eastward direction until it reaches Africa. The thermocline is deepened, resulting in a decreased flux of cold and nutrient-rich water to the surface layers during upwelling. It has been speculated that the Benguela Niño events are coupled to the El Niño events in the Pacific, because they lag El Niño by approximately one year (Bakun, 1996).

As the African continent terminates at relative low latitude, it is influenced by warm water from southern origin from the Agulhas Current in the Indian Ocean (Shelton et al., 1985). Indeed, warm events in the southern Benguela Current are most commonly caused by this route from south, but it is the warm intrusions from the tropics associated with the Benguela Niño that tend to deepen the thermocline and suppress upwelling (Shelton et al., 1985).

1.3 Some other physical processes affecting the marine ecosystem

Besides the large scale climate variability described above, other physical processes can significantly alter fish populations. Here some examples are given.

1.3.1 Limited exchange of bottom water and low oxygen content

In fjords or basins with limited exchanges due to a shallow sill, the marine ecosystem is commonly dependent on the renewal of bottom water that prevents oxygen depletion. Mariager Fjord in Denmark is such an example. The fjord has a long shallow and narrow opening to the open water. The bottom water in the inner basin is normally anoxic, but in the upper layers normal marine life exists. However, in 1997 the fjord got anoxic in the whole water column and much of the life in the fjord died (Fallesen et al., 2000). The reason for this was an unusual warm, sunny and calm summer. Warming increased the oxygen consumption due to the decomposition of an earlier massive phytoplankton bloom and the surface waters were not ventilated with oxygen due to the calm weather. This was an unusual, but not unique weather situation.

Another example of a basin with limited exchanges is the Baltic Sea, which is stratified with a 50–60 m thick low-salinity surface layer and a deep relative saline layer. The survival of cod eggs, and thereby the cod stock, is significantly dependent on the oxygen content in this deep layer.

The principal mechanism influencing the replenishment of oxygen in the deep basins of the Baltic is the occasional inflow of saline oxygen rich waters originating from the North Atlantic through the Norwegian Trench. The inflow is mainly driven by local wind conditions over the inner Danish Waters (Matthäus and Franck, 1992). Due to the density of cod eggs, they concentrate in a narrow depth range within and below the permanent halocline (Wieland and Jarre-Teichmann, 1997). At this depth the eggs are frequently exposed to low oxygen concentration, which largely affects their survival (Nissling, 1994; Wieland et al., 1994). Variations in the frequency and magnitude of inflows events into the Baltic Sea influencing the salinity and oxygen content, are believed to be essential for variations in the Baltic cod stock (Köster et al. 2003; Nissling, 2004, and references therein), whereas temperature effects are of minor importance (Köster et al. 2003). Further, Köster et al. (2003) also found predation on cod eggs and prey availability for the cod larvae to be of significant importance for the mortality.

1.3.2 Mixing caused by eddies formed by baroclinic instability

The mortality of juveniles is often controlled by the amount of available food. Mixing onto shelf banks offers such conditions, whereas open waters are less optimal. An example from the Northeast American shelf illustrates how large eddies seriously decreases the survival of juveniles by an offshore transport of recruits away from the shelf banks.

Large eddies are generated at the front between the Gulf Stream and the shelf water along the Northeast American shelf. Occasionally these eddies reach the great fishing banks and entrain shelf water offshore. These events reduce the recruitment of fish stocks, because the fish larvae entrained are exported into a less suitable environment and some larvae do not survive the sudden contact with the much warmer water (Myers and Drinkwater, 1989; Drinkwater et al., 2000a).

1.3.3 Position of fronts and stability of the water column

Hydrographic conditions play a major role for the fisheries north of Iceland. There are three very different regimes: Atlantic, Arctic and Polar. The extent of these regimes depends on the position of the Polar and Arctic Fronts in the Nordic Seas (see Section 2.1.2).

Malmberg and Blindheim (1994) showed, that weak year-classes of cod north of Iceland are connected to Polar and Arctic regimes, whereas the year-classes fluctuates between high and low during Atlantic periods. Thereby temperature alone is not the only requirement for production of strong year-classes. Malmberg and Blindheim (1994) suggested that the survival rate cod larvae north of Iceland are also regulated by the feeding conditions in their nursery areas which vary with the stratification depending on the actual regime. During warm Atlantic periods the temperature decreases gradually with depth. Thereby the build up of the summer mixed layer will deepened relative slowly with a nutrients supply from below over relatively long period. These conditions favour the primary production. Contrary, in the Polar regime a shallow mixed layer will develop and a short and intense spring bloom will quickly use the nutrients. In the Arctic regime, the stratification is low and the formation of a surface mixed layer is slower than in the Atlantic regime. However the species composition may be different with a larger fraction of Arctic species, which are less favourable for cod larvae. Moreover Capelin, which adult cod feed on, is also negatively affected by the changing food composition in the Arctic regimes.

1.4 Human impacts on the marine ecosystem – the Black Sea as an example

The most obvious human factor affecting the marine resources is overfishing, but pollution and introduction of invaders may also have a large impact on the marine environment.

The Black Sea is a classical example of human impact. From being a productive area with a rich fishery, it is today dominated by jellyfish and the commercial fishery is almost absent (Bakun, 1996). This dramatic change was caused by a combination of overfishing, high nutrient loads which favoured planktonic species and introduction of a jellyfish invader to the ecosystem.

The Black Sea is a large basin with only one shallow and narrow opening to the Mediterranean Sea. Consequently the exchange of bottom water is limited and the bottom layer is anoxide (Bakun, 1996; Shiganova, 1998). The catchment area providing runoff to the Black Sea is extremely large. Some of the river systems drain regions that contain some of the biggest cities in Europe and they are highly polluted bringing exceeding nutrients and thereby increasing plankton production, which gives rise to eutrophication. Initially this increases the fishery, but overfishing and increased eutrophication made room for less desirable component in the ecosystem such as new zooplankton species with very low food value for higher trophic levels (Bakun, 1996). In the early 1980s a drastic change in the ecosystem was initiated. A jellyfish was introduced, possible within ballast water of ships (Shiganova, 1998). It reproduces very fast and it is not prey specific and suddenly it took over the phytoplankton and zooplankton. Much of the energy in the biosystem is accumulated into jellyfish but predators on these jellyfish are rare.

1.5 Summary

Climate variations strongly affect the marine ecosystem through several processes on different temporal and spatial scales all over the world.

The Pacific Decadal Oscillation (PDO) describes the multi-decadal fluctuations of the sea surface temperature in the Pacific Ocean. It influences the thermocline depth and thereby upwelling of nutrient rich water affecting the primary production. Altering between anchovy and sardine regimes off Chile, Peru and California, and changes in the salmon populations off West America and Alaska is driven by the PDO. Similar, El Niño events in the Pacific Ocean occur every 3–8 year. They affect upwelling in the eastern boundary currents much like the PDO do. They strongly affects the fish distribution with considerable consequences for the local fishery for periods of one to two year.

Other processes affecting the marine ecosystem are described. For example, it is shown how limited exchanges caused by a shallow sill can have a large effect on the ecosystem caused by a low oxygen bottom layer. In the Baltic Sea the replenishment of oxygen is mainly through inflow of oxygen rich waters, whereas oxygen is added by wind-mixing in the Mariager Fjord, Denmark.

Human activities also strongly affect the ecosystem both directly by (over)fishing and indirectly e.g. by increased nutrient loads affecting the plankton production or by introduction of new species. A combination of these three factors changed the Black Sea from a highly productive sea with a rich fishery to a sea totally dominated by a jellyfish species.

2 Hydrographic variability in the North Atlantic of importance for fishery resources

To understand the variability in fishery resources, it is essential to know both the general state and the natural variability of the hydrographic conditions. In the North Atlantic Ocean the dominating variability on decadal timescales is associated with the North Atlantic Oscillation (NAO). It will be described ending up with some examples of influences, but first the general circulation is briefly described.

2.1 Oceanographic conditions

The ocean currents around Greenland are part of the cyclonic sub-polar gyre circulation of the North Atlantic and the Arctic region. The bottom topography plays an important role for guiding the circulation and for the distributing the water masses.

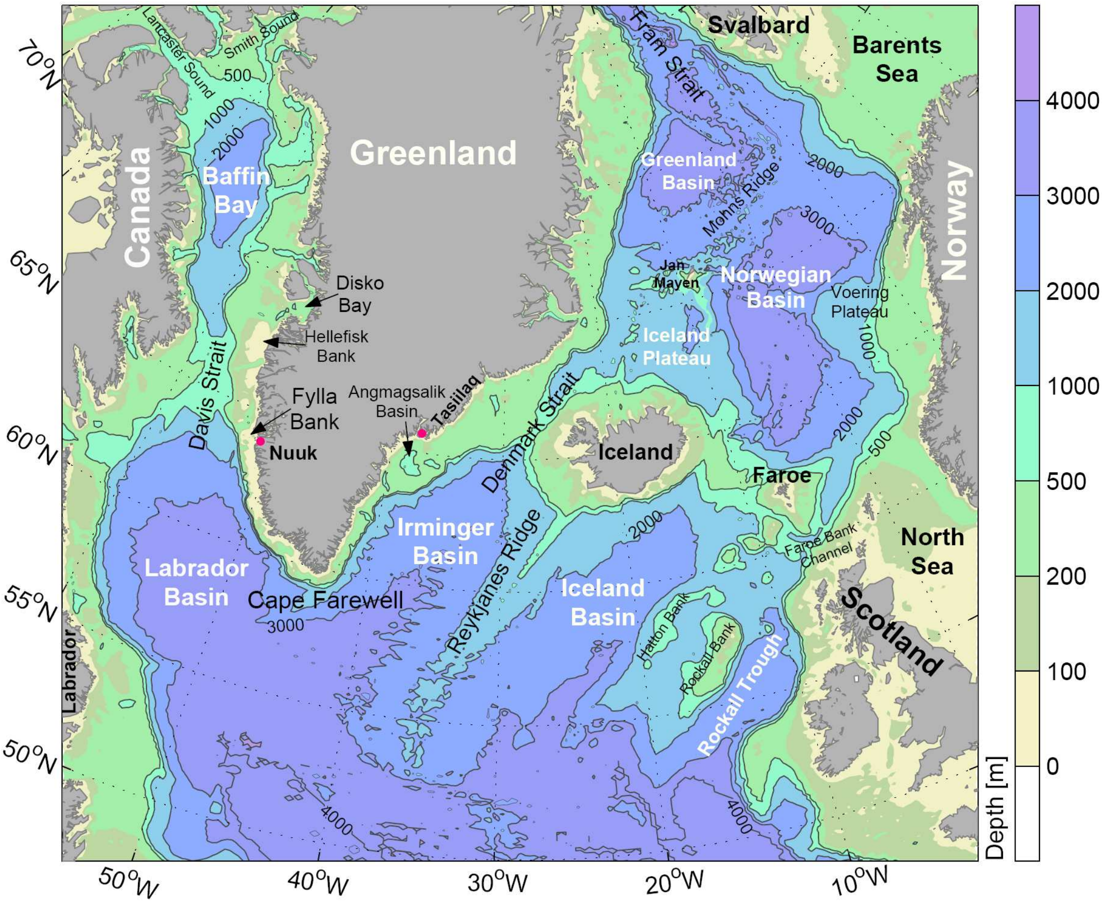


Figure 5. Bottom topography of the North Atlantic including the Nordic Seas. Bottom topography contours at 500, 1000, 2000, 3000 and 4000 m are shown as lines.

2.1.1 Bottom topography

The bottom topography is shown in Figure 5. The northern part of the Atlantic Ocean consists of the Labrador Basin, the Irminger Basin and the Iceland Basin which all are semi-closed basins which gradually increase their depth in the southward direction.

The ridge between Greenland and Scotland separates the North Atlantic from the Nordic Seas. The two most important routes of the overflow of dense water is the Denmark Strait with a sill depth of about 630 m and the 840 m deep Faroe Bank Channel. The maximum depth between Iceland and Faroe is about 480 m and the Wyville-Thomson Ridge west of Scotland is about 600 m deep. The Nordic Seas consists of three major basins: the Greenland, and Norwegian Basins and the Iceland Plateau. The Nordic Seas are connected to the Arctic Ocean through Fram Strait which is the primary opening and has a depth of more than 2600 m deep. In the Barents Sea another but shallow passage exists.

West of Greenland the Davis Strait makes a 670 m deep opening between the Labrador Basin and Baffin Bay. In the northern part of Baffin Bay the Canadian Archipelago makes a connection to the Arctic Ocean. The primary openings are Smith Sound and Lancaster Sound with depths of a few hundreds of meters.

Greenland is surrounded by a continental shelf with several banks separated by deep channels formed during the last ice age. The East Greenland shelf is generally wider and deeper than the Southwest Greenland shelf.

2.1.2 Surface water masses

The surface temperature and salinity of the region is shown in Figure 6. In principle only two watermasses are found in the North Atlantic. Warm and saline North Atlantic Water (NAW) entering from south and cold and low-saline Polar Water (PW) from the north. Due to Coriolis deflection, the eastern parts of the North Atlantic and the Nordic Seas are dominated by NAW and the western parts by PW.

The NAW stems from the Gulf Stream which is part of western boundary current of the subtropical gyre. It dominates most of the waters south of the Greenland-Scotland Ridge except for the western shelves and dominates the eastern part of the Nordic Seas. Contrary, PW dominates in the western part of the Nordic Seas, the Baffin Bay and on the western shelves in the North Atlantic Ocean. In the Nordic Seas, a third watermass called Arctic Surface Water (ASW) is found in between the NAW and PW. It is formed locally in the Greenland Sea and the Iceland Sea, as a mixture of PW and NAW from the boundaries and exposed to atmospheric exchanges. The front between NAW and ASW is called the Arctic Front and is easily detected in both temperature and salinity on the western side of the Norwegian Basin in Figure 6. The Polar Front between PW and the Arctic waters in the Nordic Seas is most easily observed in salinity and is found over the continental shelfbreak off East Greenland (Figure 6). These fronts can temporary vary a lot in position e.g. with major implication for the fisheries north of Iceland as described in Section 1.3

2.1.3 The mean surface circulation

The surface circulation generally guided by the bottom topography, which is easily seen in the various basins (Figure 7) and which is also evident from the surface temperature and salinity fields (Figure 6) in the Nordic Seas.

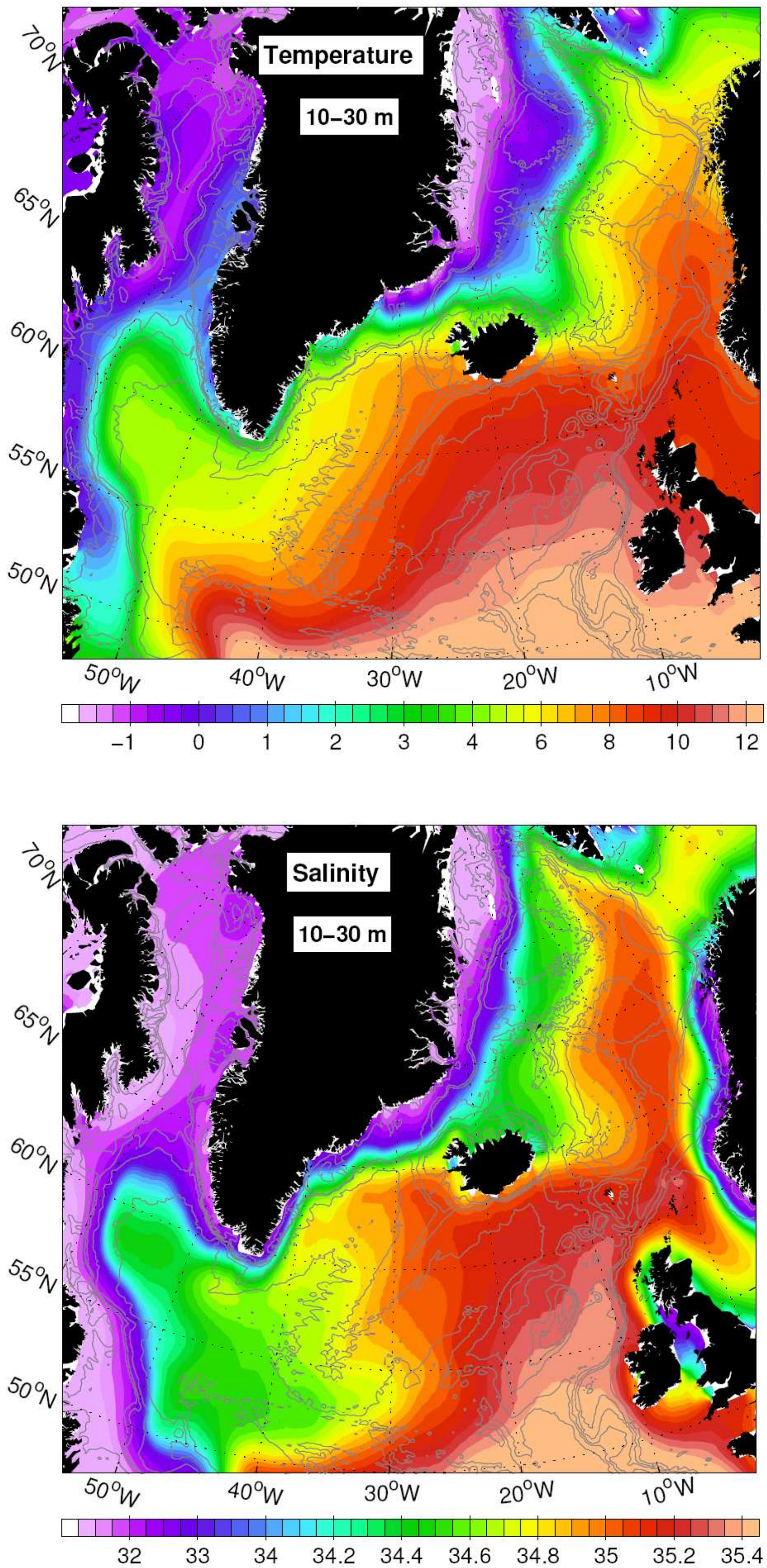


Figure 6. Annual mean surface (10–30 m) temperature (top) and salinity (bottom) from World Ocean Atlas 2001 (WOA2001). Note the change in scales at 4°C for temperature and 34 for salinity. Bottom contours at 500, 1000, 2000, 3000 and 4000 m are shown.

From southwest the North Atlantic Current (NAC), an extension of the warm and saline Gulf Stream, crosses the Atlantic Ocean. Part of it turns northwestward following the topography around the Reykjanes Ridge into the Irminger Basin as the Irminger Current (IC), which occupies the ocean area south of Iceland. Part of the IC passes Denmark Strait and continues north of Iceland along the coast, where it meets the cold and low-saline East Icelandic Current. The main part of the IC turns towards Greenland in the northern part of the Irminger Basin, where it flows southward along the east coast of Greenland as the Irminger Water branch of the East Greenland Current. Here it is located over the continental slope offshore the Polar Water component of the East Greenland Current (EGC). Part of this Irminger Water branch recirculates in the Irminger Basin while another part subducts under the Polar Water as it rounds Cape Farewell and continues northward as the West Greenland Current (WGC). Part of the water within the WGC crosses Davis Strait, while another part turns west towards Canada at about 64–67°N, and merges with the southward flowing Labrador Current.

The two main branches of the NAC enters the Nordic Seas (see Figure 7). One continues along the continental shelf passing west of Britain crosses the Faroe-Shetland channel and follows the continental shelf off Norway as the Norwegian Current. The other branch takes the northward route west of Hatton Bank. At the Voering Plateau the Norwegian Current splits into two components. One flows towards the Mohn Ridge between the Greenland Basin and the Norwegian Basin and follows the ridge to Fram Strait, whereas the other continues north along the Norwegian Coast. The latter splits into two as it passes the entrance to the Barents Sea between Norway and Svalbard. One component enters the Barents Sea and the other follows the continental slope into the Fram Strait within the West Spitsbergen Current. As it reaches Fram Strait it converges with colder and less saline Arctic Surface Water and continues as a subsurface current into the Arctic Ocean. Before entering the Arctic Ocean part of the NAC turns westward into the East Greenland Current. This water of Atlantic origin lies beneath the Polar Water from about 150–800 m.

Cold and low-salinity water is exported southward from the Arctic Ocean. The Arctic Ocean is supplied with fresh water from large rivers of which the Russian rivers are the most important (Aagaard and Carmack, 1989). The surplus of water leaves the Arctic Ocean through Fram Strait and through the Canadian Archipelago. The former is by far the most important making up about 75 % of the surface water outflow from the Arctic (Buch, 2002). It is transported southward on top of the East Greenland continental shelf to Cape Farewell within the EGC. On its way southward it branches two times following the bottom topography north of Jan Mayen and north of Iceland. More than half of the low-saline surface water transported through Fram Strait is entrained into the deep overflow waters within the Nordic Seas, whereas the rest continues southward through Denmark Strait as a surface current within the EGC.

Close to the coast the water is modified by run-off from Greenland and thereby increases its baroclinic momentum. This forms a fresh water jet named the East Greenland Coastal Current (Bacon et al., 2002). The run-off stems from melting of the Greenland ice sheet and from runoff of precipitation. The latter is especially important at southeast Greenland, where the low pressure systems are lifted on the steep and high mountains casing enhanced precipitation.

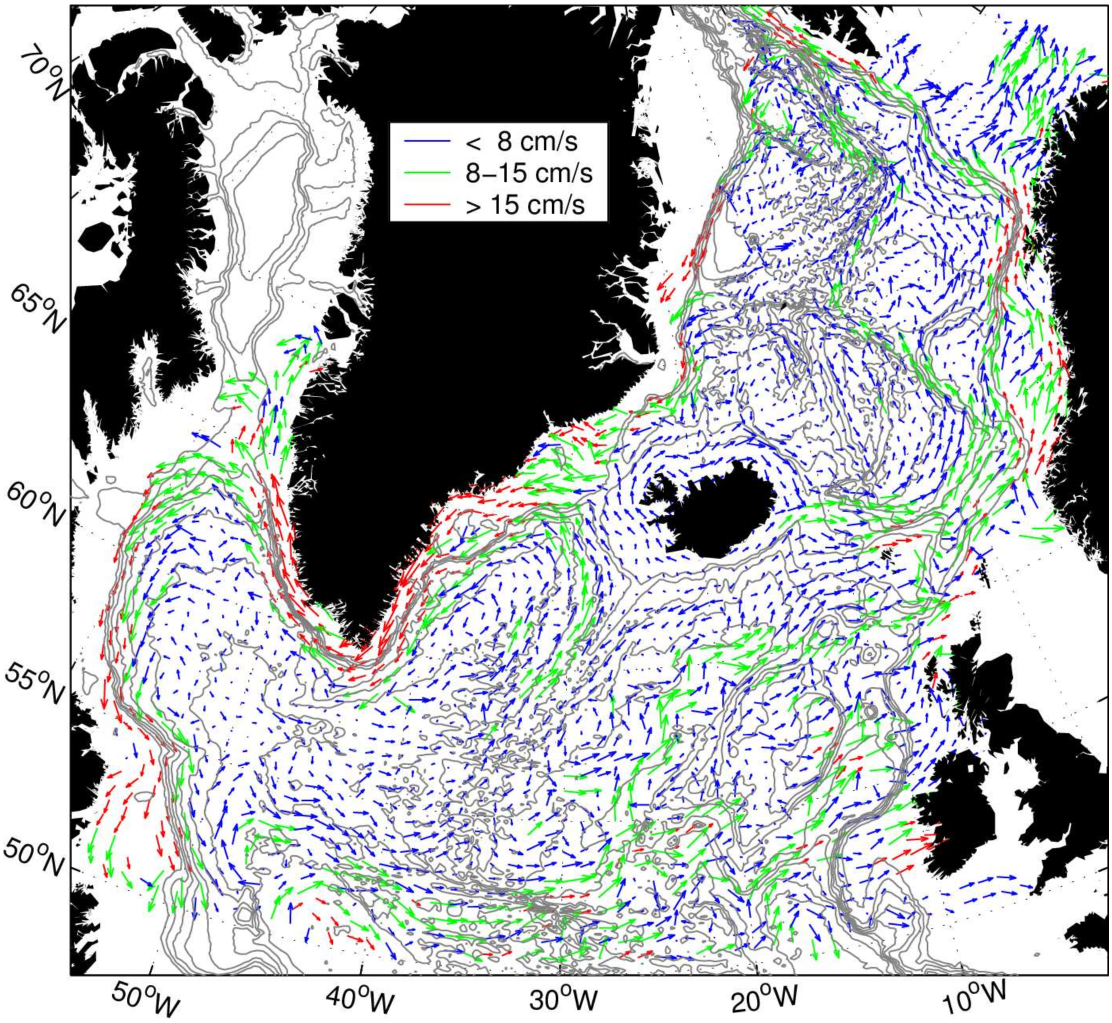


Figure 7. Quasi-Eulerian current vectors at 15 m depth derived from 18 day low-passed filtered drifter trajectories and averaged in overlapping 1° latitude × 2° longitude boxes. Note the different scales for low-velocity (blue), medium-velocity (green) and high-velocity currents (red arrows). Bottom topography at 500 m intervals is shown as grey lines. Reproduced after Jakobsen et al. (2003).

As the EGC rounds Cape Farewell it continues northward in the WGC on top of the West Greenland shelf. At about 64°N major parts of the WGC turns west and join the Labrador Current while the other part crosses the Davis Strait into the Baffin Bay.

The second major outflow from the Arctic Ocean enters Baffin Bay through the Canadian Archipelago. It flows southward on the western side as the Baffin Current, crosses Davis Strait and continues southward on the Labrador shelf as the Labrador Current. It meets the NAC at about 40–45°N. In Davis Strait a small branch of the Baffin Current turns eastward and joins the WGC.

The rim currents in the Labrador Basin and the Irminger Basin can be regarded as a large gyre. The water inside is relatively stagnant and the velocities are small. In the southern part of the gyre the Labrador Current meets the NAC and part of the Labrador Water is shifted north, then east and finally northeast heading towards the Reykjanes Ridge where it joins the Irminger Current. The barotropic component of the rim currents are strong; the surface currents derived from drifters (Figure 7) are much like the current at 700 m derived from floats (Figure 2a in Lavender et al., 2000).

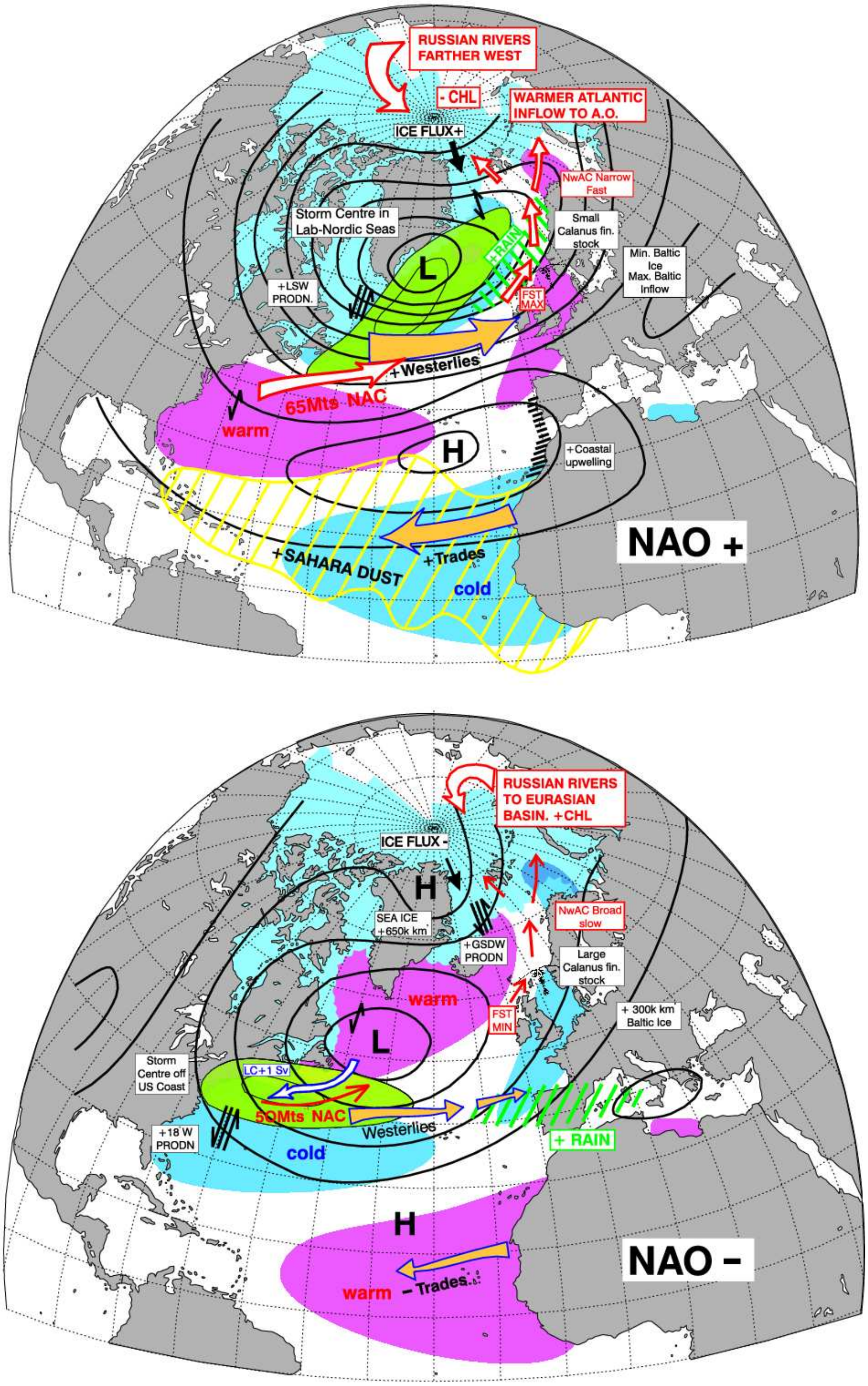


Figure 8. Schematic characteristic during high NAO index (upper) and low NAO index (lower). Modified from CLIVAR transparency D1 (courtesy of CEFAS, UK). http://www.clivar.org/publications/other_pubs/clivar_transp/d1_transp.htm

2.2 The North Atlantic Oscillation (NAO)

The North Atlantic Oscillation (NAO) is the most prominent mode of variability in the North Atlantic winter climate (Hurrell, 1995).

The NAO, which is associated with the strength of the westerlies across the Atlantic onto western Europe, describes the large-scale meridional fluctuation in atmospheric mass between the North Atlantic regions of the subtropical anticyclone high pressures near the Azores and the subpolar low pressure system near Iceland (van Loon and Rogers, 1978). It represents the most important anomaly pattern of the North Atlantic accounting for more than one-third of the total variance of the Sea Level Pressure (SLP) field during winter. It is widely used as a general indicator for changing winter climate conditions in Europe (Hurrell, 1995; Hurrell and van Loon, 1997). Because the signature of the NAO is strongly regional, a simple index of NAO is commonly defined as the difference between the SLP anomalies at the Azores and Iceland (van Loon and Rogers, 1978 and references therein). Storm activity in the subpolar North Atlantic is strongly related to the NAO wintertime index (Hurrell, 1995). A positive index is associated with a more northeasterly storm track (Jones, 1990).

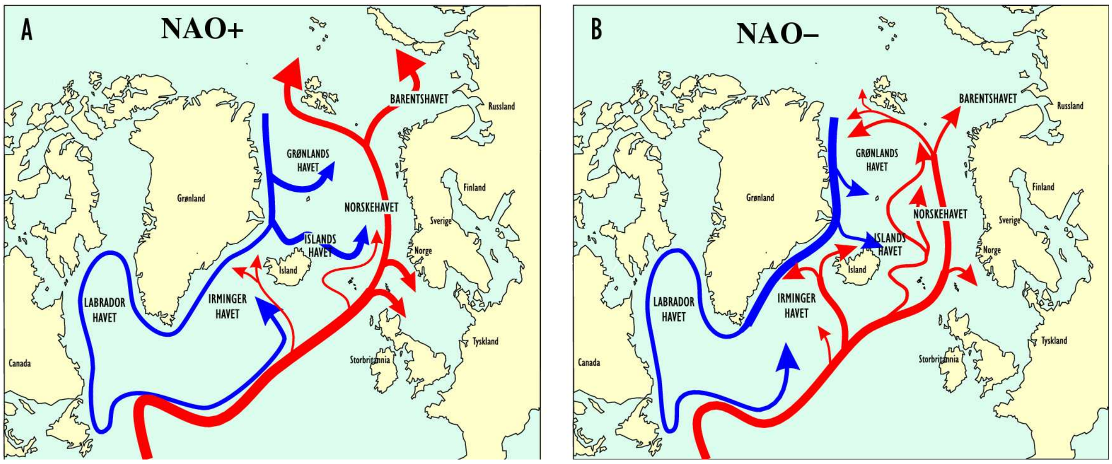


Figure 9. Schematic surface circulation during a) high NAO index, and b) low NAO index. From Blindheim et al. (2001).

The NAO is known to have a great impact on the ocean circulation and on the hydrographic conditions in the North Atlantic region including the Greenland Waters (e.g. Dickson et al., 1996; Dickson et al., 2000; Blindheim et al., 2000; Blindheim et al, 2001; Buch et al., 2004). Following Blindheim et al. (2000; 2001), the NAC is narrowed but strengthened during periods of high NAO. This is caused by the increased wind-stress as sketched on Figure 9a. The branches are reduced with exceptional reduction in the Irminger branch as the most noticeable. Contrary, the EGC becomes wider and the branches within the Nordic Seas are strengthening, leaving weaker transport through Denmark Strait. In periods of low NAO the NAC is widening and the branching is increased (Figure 9b). Noticeable is the increased Irminger Current. The branching of the EGC is decreased resulting in an increased transport through Denmark Strait.

The variability of the NAO index since 1865 is shown in Figure 10. Positive values of the index indicate stronger than average westerlies over the mid-latitudes associated with low-pressure anomalies over the region of the Icelandic Low and anomalous high pressures across the subtropical Atlantic.

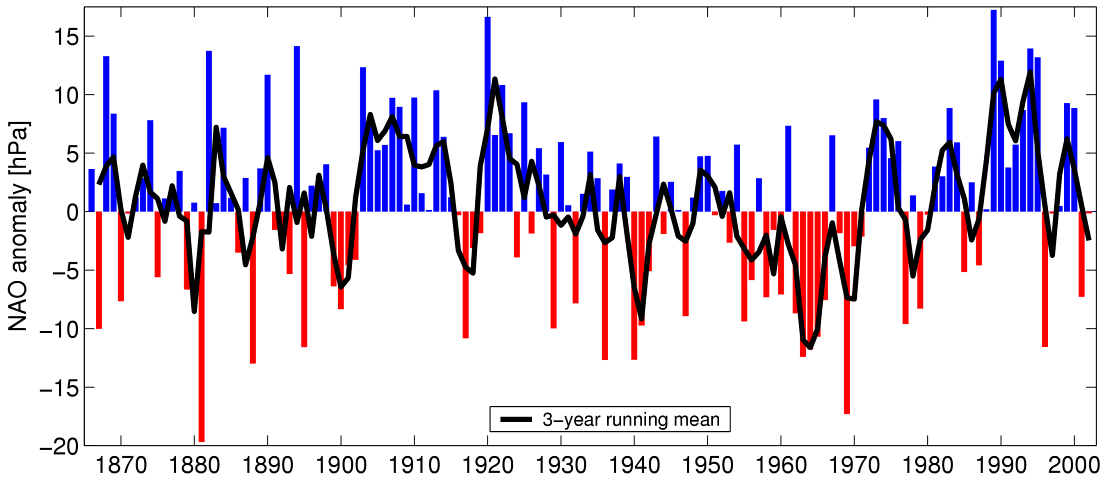


Figure 10. Time series of winter (December–March) index of the NAO from 1865–2003. The heavy solid line represents the meridional pressure gradient smoothed with a 3-year running mean filter. Note that values for both 2001/2002 and 2002/2003 were very close to zero. (Pressure data described in Buch et al., 2004). From Ribergaard and Buch (2004).

During phases of high NAO index, the westerlies are amplified. Due to the maritime influence, the temperature over northern Europe becomes higher than normal. Similar, continental air masses from northern Canada lower the mean temperatures over the Labrador Basin and at Southwest Greenland, as sketched in Figure 8a. Hurrell and van Loon (1997) showed that the temperature anomalies for the North Atlantic and surrounding land masses for the 1980–1994 period was strongly related to the persistent and exceptionally strong positive phase of the NAO. Similar, Hanna and Cappelen (2003) showed, that the cooling in the coastal southern Greenland region from about 1970 was significantly inversely related to an increased positive phase of NAO. Box (2002) even found the NAO to be highly significant related to the temperatures on the coast at western and southern Greenland for the period 1873–2001. This clearly demonstrates a strong correlation between the strength of the westerlies across the North Atlantic (the NAO index) and the air temperatures in Greenland and Europe. Further it shows that the air temperatures in Greenland and Europe are negatively correlated, a phenomenon referred to as a seesaw effect (e.g. van Loon and Rogers, 1978).

This is further illustrated in Figure 11 which shows a comparison of the air temperatures over Greenland from a low NAO period (1960–1969) to a high NAO period (1990–1999). Note that especially offshore of West Greenland, but also over the Irminger Basin, it was significantly warmer in the 1960s than in the 1990s. Contrary, northern Europe but also the Greenland Basin offshore northeast Greenland was significant cooler in the 1960s than the 1990s.

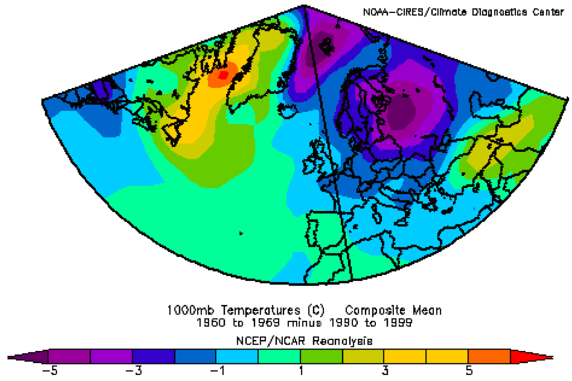


Figure 11. Difference in air temperatures at the 1000 hPa level between 1960–1969 and 1990–1999 calculated using NCEP/NCAR reanalysis data (www.cdc.noaa.gov).

2.3 Global ocean circulation, convection and NAO

Open-ocean deep convection plays an important role in maintaining the thermohaline circulation (THC), which governs the poleward transport of heat in the ocean. It is mainly observed in the Labrador Sea, the Greenland Sea, and the Weddell Sea in the Atlantic Ocean, as well as in the Mediterranean Sea (Marshall and Schott, 1999 and references therein). In the North Atlantic Ocean the poleward heat transport is associated with the NAC and the IC side branch and thereby their strengths are directly related to the variability in open-ocean deep convection. Fluctuations in the THC is on timescales of decades to centuries and have a large impact on especially the North Atlantic but also worldwide.

The convective activity in the Greenland and Labrador Seas has varied inversely on decadal timescales. The switch of the convective activity from the Greenland Sea to the Labrador Sea has been attributed to changes in the index of the North Atlantic Oscillation (Dickson et al., 1996). The different forcing over the Labrador and Greenland Sea during high and low NAO index is illustrated in Figure 11 which shows a comparison of the air temperatures over Greenland from a low NAO period (1960–1969) to a high NAO period (1990–1999).

During high phases of NAO the Labrador Sea is cooled in wintertime by cold air masses that stems from the northern Canada. The Labrador Sea is spun up by an increased windstress curl. Ekman divergence prevent the fresh surface water from the shelf to spread into the centre of the Labrador Sea, and it further generates a doming isopycnal structure resulting in reduced stratification in the interior. The dense water is brought closer to the sea surface and less buoyancy loss is needed for overturning.

Contrary to the Labrador Sea, the Greenland Sea is cooled during low phases of NAO when the local windstress curl is increased. The convection in the Greenland Sea is in principle similar to the Labrador Sea convection, but brine rejection from the formation of sea-ice is believed to be crucially important, as salt is needed to make the surface waters dense enough to overturn. Convection takes place in small scale plumes less than 1 km horizontally associated with eddies formed by baroclinic instability (Marshall and Schott, 1999).

Recently Pickart et al. (2003) showed that intermediate water masses taking part in the THC are occasionally formed in the southern Irminger Sea forced by the Greenland tip-jet (Doyle and Shapiro, 1999). Tip-jets are formed as low pressure systems pass just south of Greenland and strong winds from northeast are forced south of Greenland caused by its steep topography. Storm activity in the subpolar North Atlantic is strongly related to the NAO wintertime index (Hurrell, 1995). A positive index is associated with a more northeasterly storm track, and a greater number of low-pressure systems pass near Greenland (Jones, 1990). This indicates a positive relationship between the NAO and the number of tip-jet events in a winter, which is confirmed by Pickart et al. (2003). Whereas convection in the Labrador Sea is caused by broad-scale wind and buoyancy forces over most of the basin, convection in the Irminger Sea is on a relative small spatial scale. Buoyancy fluxes are strongly localized caused by a strong windstress curl formed east of the southern tip of Greenland during tip-jet events.

2.4 Ecosystem response on NAO

Changes in NAO affect the ecosystem in the North Atlantic sector on decadal timescales. The most obvious influence of NAO on the marine ecosystem is through temperature which affects the growth rates of the large majority of species (Ottersen et al., 2001).

However the ecosystem does not always respond directly to the NAO but rather indirectly through changes in the local physical and chemical characteristics, which are connected to the variability in the NAO (Ottersen et al., 2001; Drinkwater et al., 2003). The responses on NAO are very different between the eastern and western side of the North Atlantic Ocean, and they are generally out of phase.

2.4.1 The increased Barents Sea cod stock in relation to higher temperatures and copepods

Catches of Barents Sea cod increased in the 1980s and 1990s despite of high fishing pressure. This can be related to increased inflow of warm Atlantic Water during positive NAO phases (Ottersen and Stenseth, 2001).

Increasing temperature has a positive effect on the growth of cod in the Barents Sea (Planque and Frédou, 1999). Moreover, the transport of *Calanus finmarchicus* into the Barents Sea increases with increasing inflow of Atlantic Water, and hence more food is available for cod (Helle and Pennington, 1999; Ottersen and Stenseth, 2001; Ottersen et al., 2001 and references therein). Finally, the primary production is increased due to larger ice-free areas during high NAO phases, which has a positive effect on the survival of cod larvae.

2.4.2 The decline of the Canadian cod stock in relation to lower temperatures and overfishing

The landings of Canadian cod increased gradually from 100,000 tonnes to 250,000 tonnes in the period 1850–1950. During the late 1950s and early 1960s landings strongly increased to almost 800,000 tonnes as modern trawlers were introduced. In the 1970s the cod catches declined to 200,000 tonnes. A small increase in landings was followed by an almost total collapse in the early 1990s. Thereafter a moratorium was imposed on cod fishery.

Recently Drinkwater (2002) has reviewed the collapse of the Canadian cod stock in the early 1990s illustrating the direct and indirect consequences of a persistent temperature decrease from around 1970 to the mid-1990s caused by increased influence of cold continental air masses from the northern Canada during a period of high North Atlantic Oscillation index. Overfishing is believed to be the main reason for the collapse, but he argued that cold climate conditions also played an important role.

The heavy fishing pressure changes the demography of the stock, such that large and old cod was removed. This has two consequences: (1) fewer adult mature cod means fewer eggs leading to reduced recruitment. (2) As the egg mortality is high for first-time spawning cod (Makhotin and Solemdal, 2001), the recruitment decreases. As a result the cod stock is more dependent upon recruitment from first-time spawning cod than it would have been with a natural stock demography.

Temperature has a positive affect on the growth of the Canadian cod, which lives near its lower temperature limits (Brander, 1994; Brander 1995; Planque and Frédou, 1999). Krohn and Kerr (1997) found, that the growth rate appears to be determined early in their life cycle, such that a small cod at age-3 will remain small at older stages. Drinkwater (2002) argues that the weight defeat accounts for 30–50 % of the total biomass during the cold period from the late 1970s to the early 1990s, coupled to the positive NAO phase. This had two direct effects: (1) Since fishing quota is measured in weight, more cod is caught and (2) because larger cod is more valuable, fishermen dumped small cod overboard (Kulka, 1997) and thereby the fishing mortality is increased. Moreover, the generally poor conditions of the cod most likely led to decreased egg production and the egg mortality is increased in the cold environment. At the same time the cod moved southward.

The West Greenland cod stock also collapsed at about the same time as the Canadian cod stock (see Chapter 3.1). The fishery pressure was similar and West Greenland also experienced extremely cold conditions. However, the decline is believed to be a response to reduced drift of juvenile cod from Iceland to Greenland (Buch and Hansen, 1988; Hovgård and Buch, 1990; Buch et al., 2004; Stein and Borovkov, 2004; Stein, 2004). Indeed, Dickson et al. (1994) concluded, that it seems unlikely that the two known earlier cod periods at West Greenland in the 19th century could have occurred without a wind-induced strengthening of the Irminger Current which transported heat and cod larvae from Iceland to West Greenland.

Wieland and Hovgaard (2002) showed that during the 1950s and 1960s, when both the Canadian and West Greenland cod stocks were strong, large numbers of cod larvae crossed Davis Strait towards the Canadian side. These could have contributed to the Canadian cod stock. This drift of cod larvae was very low or absent during the cold period and the absence of recruits from Greenland might have amplified the decline of the Canadian cod stock.

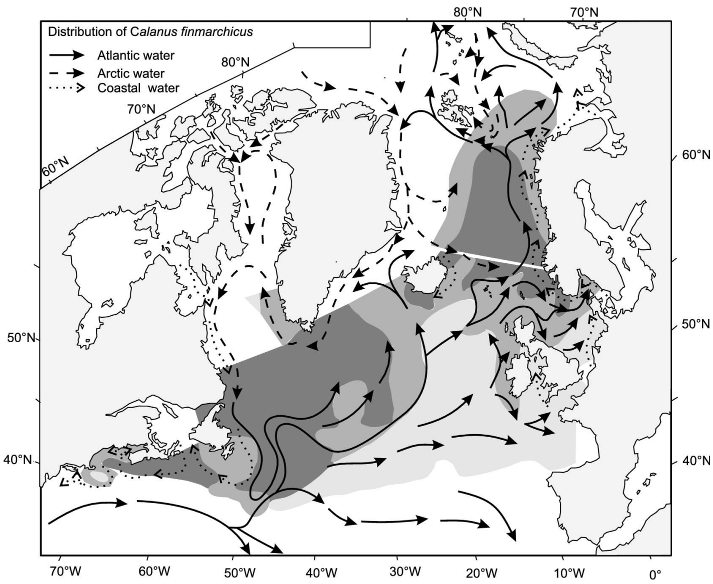


Figure 12 Distribution of *Calanus finmarchicus* in the North Atlantic including the Nordic Seas. Overlayed is the general surface circulation. From Sundby (2000).

2.4.3 Calanus finmarchicus

Zooplankton are passively floating animals that mostly feed on phytoplankton. They are important food for most fish larvae, juveniles and also some smaller adult fish, especially pelagic fish species like herring (*Clupea harengus*), sprats (*Sprattus sprattus*), sand eel (*Ammodytidae*) and many others.

Copepods in the North Atlantic and the Nordic Seas can be divided up into warm, sub-arctic and arctic species (e.g. Beugrand et al, 2002). In much of this area, the sub-arctic copepod *Calanus finmarchicus* dominates the zooplankton biomass and production in springtime (e.g. Greene et al., 2003). It is the dominant prey species for the early stages of cod larvae (Sundby, 2000). Its distribution is shown in Figure 12. In the North Atlantic it is mainly found in the Labrador and Irminger Basins to the west, whereas in the Nordic Seas it is found in the Norwegian Basin to the east. The North Sea area makes up its southern boundary distribution in the northeast Atlantic.

To understand the variability in *C. finmarchicus* it is necessary to understand its seasonal behaviour. During late spring and early summer it is most abundant in the surface layers after the spring bloom. In late summer it starts to descend to depths of 400–2000 m depth where it overwinters with reduced activity until it again emerges to the surface from late winter (Heath et al., 2000). Because it overwinters in deep water, advection is extremely important for transporting them to the shelf regions.

Temperature, and especially advection of copepods and warm water, turns out to be extremely important for the abundance of zooplankton in the North Atlantic (Backhaus et al., 1994; Greene and Pershing, 2000; Gaard and Hansen, 2000; Ottersen et al., 2001; Holliday and Reid, 2001; Beaugrand et al., 2002; Beaugrand, 2003; Greene et al., 2003; Beaugrand and Reid, 2003; Beaugrand, 2004).

A reorganisation of the distribution of the copepod in the North Atlantic has observed in the period from the 1960s to the 1990s (Beaugrand et al. 2002). The warm water species extended northward by more than 10° latitude in the Northeast Atlantic while the sub-arctic and arctic species have retreated (see Figure 13 and Figure 14a). Opposite trend is observed in the Northwest Atlantic (Beaugrand et al. 2002). This reorganization is closely linked to the NAO phase, which was generally positive and increasing for the period. The temperature increases and continental slope current strengthen in the Northeast Atlantic. Thereby warmer species are favoured by increasing temperature and increasing advection from south (Figure 13), which have replaced cold species copepods (Figure 14b). Some relate the period after the mid-1980s to increasing Northern Hemisphere temperature in addition to the NAO response (Reid et al., 2001; Beaugrand et al., 2002; Beaugrand and Reid, 2003; Reid et al., 2003; Beaugrand, 2004). They argued that a regime shift in the North Sea had taken place in the mid-1980s.

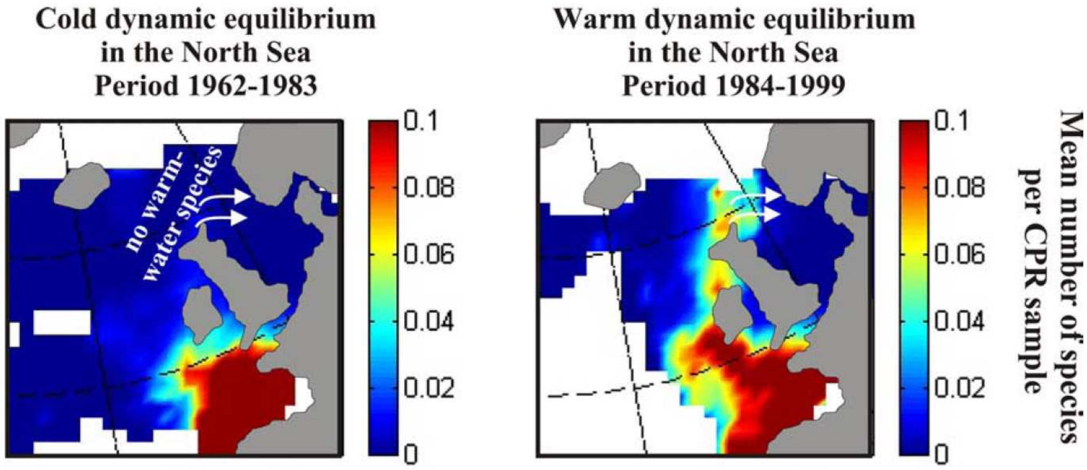


Figure 13. Mean number of warm-temperate copepod species in the northeast Atlantic Ocean for the period 1962–1983 and 1984–1999. White arrows indicate inflow into the North Sea. From Beaugrand (2004).

As cod larvae mainly feet on *C. finmarchicus*, the northward extension has affected the cod recruitment. Advection of *C. finmarchicus* into the Barents Sea is found to be positively related to NAO, favouring the conditions for juvenile cod (Helle and Pennington, 1999). For the North Sea, decreased input of *C. finmarchicus*, in addition to overfishing, have reduced the cod recruitment (Beaugrand et al., 2003).

Beaugrand and Reid (2003) found the abundance of salmon in the eastern North Atlantic to be significantly related to zooplankton abundance, whereas the horse mackerel fishery in the North Sea has increased since about 1988 (Reid et al., 2001). Beaugrand (2004) grouped fish recruitment in the North Sea into groups of flatfish and gadoid fish and made a principle component analysis. Flatfish recruitment made a marked improvement centred around 1980, while gadoid fish recruitment decreased after the mid-1980s.

Backhaus et al. (1994) suggested that the *C. finmarchicus* population in the North Sea is sustained by advection from the Faroe-Shetland channel. In spring *C. finmarchicus* ascents from the deep water into the NAC from where it is advected northward. The drift is highly sensitive to overwintering depth, timing and velocity of the ascent (Harms et al., 2000), as well as on the wind driven circulation (Backhaus et al., 1994; Heath et al., 1999). From the late 1960s to 2000 the overflow through the Faroe Bank channel has reduced by 15 percent due to reduced convection activity in the

Greenland Sea (Hansen et al., 2001). As *C. finmarchicus* overwinter in the Norwegian Sea Deep Water, Heath et al. (1999) argued, that this led to reduced supply of *C. finmarchicus* in the Faroe Bank Channel.

Like in the North Sea, *C. finmarchicus* on the southwestern Faroe shelf is also originating from the overflow waters in the Faroe Bank channel, but here the wind conditions are believed to play a larger role, as they have to be advected a long distance to reach the waters close to the Faroes (Gaard and Hansen, 2000). Opposite to the North Sea, which can be regarded as a marginal area for *C. finmarchicus*, import of *C. finmarchicus* onto the Norwegian shelf waters are less sensitive to the timing of the assent, as they overwinter in the Norwegian Sea Deep Water which is found all along the coast (Harms et al., 2000). The advection of *C. finmarchicus* into the Barents Sea is found to be positively related to NAO, favouring the conditions for juvenile cod (Helle and Pennington, 1999).

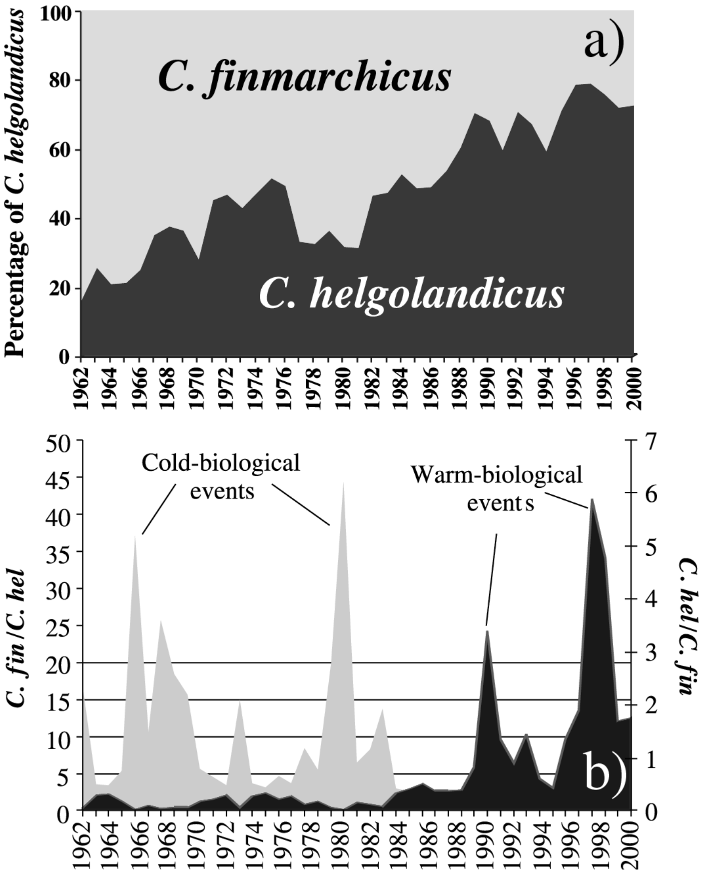


Figure 14 *Calanus finmarchicus* (cold species) versus *Calanus helolandicus* (warm species) in the North Sea for the period 1962–2000. a) Percentage of abundance. b) Ratios. Black is *C. helolandicus* to *C. finmarchicus* and grey is the reverse ratio. From Reid et al. (2003).

The abundance of copepods on the western shelves in the North Atlantic have varied opposite to the copepod abundance on the eastern shelves (Pedersen and Smidt, 2000; Greene and Pershing, 2000; Beaugrand et al, 2002; Beaugrand, 2004). The waters have gradually cooled and the cold water species tend to be more dominating. Greene and Pershing (2000) and Greene et al. (2003) suggested that advection of the shelf waters is of vital importance for the structure of copepods, but this remains to be tested.

2.5 Summary

The North Atlantic Oscillation (NAO) is the dominant variability on decadal timescales over the North Atlantic sector accounting for more than one-third of the total variations of sea level pressure field during winter. A positive index indicates a strengthening of the westerlies.

In periods of positive NAO phases, the air temperature over northern Europe becomes higher than normal due to increased westerlies and thereby increased maritime influence. Similar, the temperatures over the Labrador Basin and Southwest Greenland become lower than normal due to increased influence of continental air masses from northern Canada. Generally, the temperature anomalies over the eastern and western parts of the North Atlantic react opposite to NAO. The NAO has a strong impact on the ocean circulation. During high NAO phases, the North Atlantic Current is strengthened but narrowed, whereas the Irminger Current is reduced.

Changes in NAO affect the ecosystem in the North Atlantic sector. A reorganisation of the distribution of the copepod in the North Atlantic was observed during positive NAO phases. A strengthened continental slope current and warming in the Northeast Atlantic resulted in a northward extension of the warm-water species and a decrease in the southern extension of the cold water species. Opposite trends are seen in the Northwest Atlantic. As cod larvae mainly feed on *Calanus finmarchicus,* the northward extension in the Northeast Atlantic has affected the cod recruitment positively in the Barents Sea but negatively in the North Sea.

Temperature is positive related to the growth of cod in cold waters. Therefore a change to positive NAO has a positive effect on the Barents Sea cod but negative consequences for the Canadian cod.

3 Ecosystem variability in Southwest Greenland waters

Greenland economy is highly dependent on fishery. Dramatic changes in the fishery resources have resulted in not only changes in the fishing fleet, but also affected the Greenlandic economy which highly affects people socially. Today the Greenland economy, formerly being highly dependant on a rich cod fishery, is almost entirely dependent on the Greenland shrimp stock which by 1995 comprised 73 percent of Greenland's total export (Hamilton et al., 2000; Hamilton et al., 2003).

The area off South East and West Greenland are important fishing grounds. Ocean currents transporting water from the polar and temperate regions affect the marine productivity in the Greenland shelf areas, and changes in the North Atlantic circulation system therefore have major impact on the distribution of species and fisheries (Pedersen and Smidt, 2000; Pedersen and Rice, 2002; Buch et al., 2004).

A review of the past variability in both the fisheries and the hydrographical conditions off West Greenland and its possible relations to climate variability is given. Finally two drift studies are presented: 1) drift modelling of shrimp larvae and plankton at Southwest Greenland and 2) transport of cod larvae from Icelandic towards Greenland.

3.1 Variations in the commercial fishery off Southwest Greenland

The marine ecosystem off Southwest Greenland is characterised by relatively few dominant species, which interact strongly (Pedersen and Kanneworff, 1995; Rätz, 1999; Pedersen and Zeller, 2001).

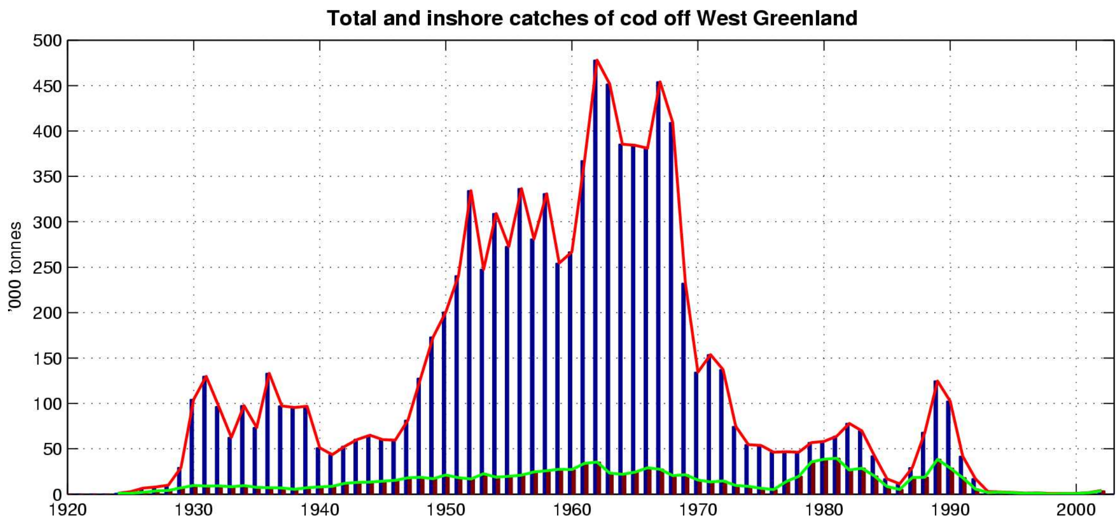


Figure 15. Total annual catches in 1000 tonnes of cod off West Greenland (blue bars and red line). The inshore catches are shown by red bars and the green line. From Ribergaard and Sandø (2004, in prep.).

At Greenland there different cod stocks (Wieland and Hovgård, 2002). Offshore the Atlantic cod (*Gadus morhua*) predominates, whereas Greenland cod (*Gadus ogac*) is common in West Greenland but is found only in fjords and in coastal areas from about 60°N to 73°N. Inshore Atlantic cod shows almost no migration between different fjord systems, whereas offshore Atlantic cod shows alongshore migration (Storr-Paulsen et al., 2004). Focus is here limited to the Atlantic cod which is by far the most important for the fishery (Figure 15).

The fishery on Atlantic cod at West Greenland started around the mid-1920s (Figure 15) after a general warming of the Northern Hemisphere (Jensen, 1939; Dickson et al., 1994; Buch et al., 1994; Horsted, 2000), which replaced seal hunting as the most important resource for Greenlanders (Hamilton et al., 2003 and references therein).

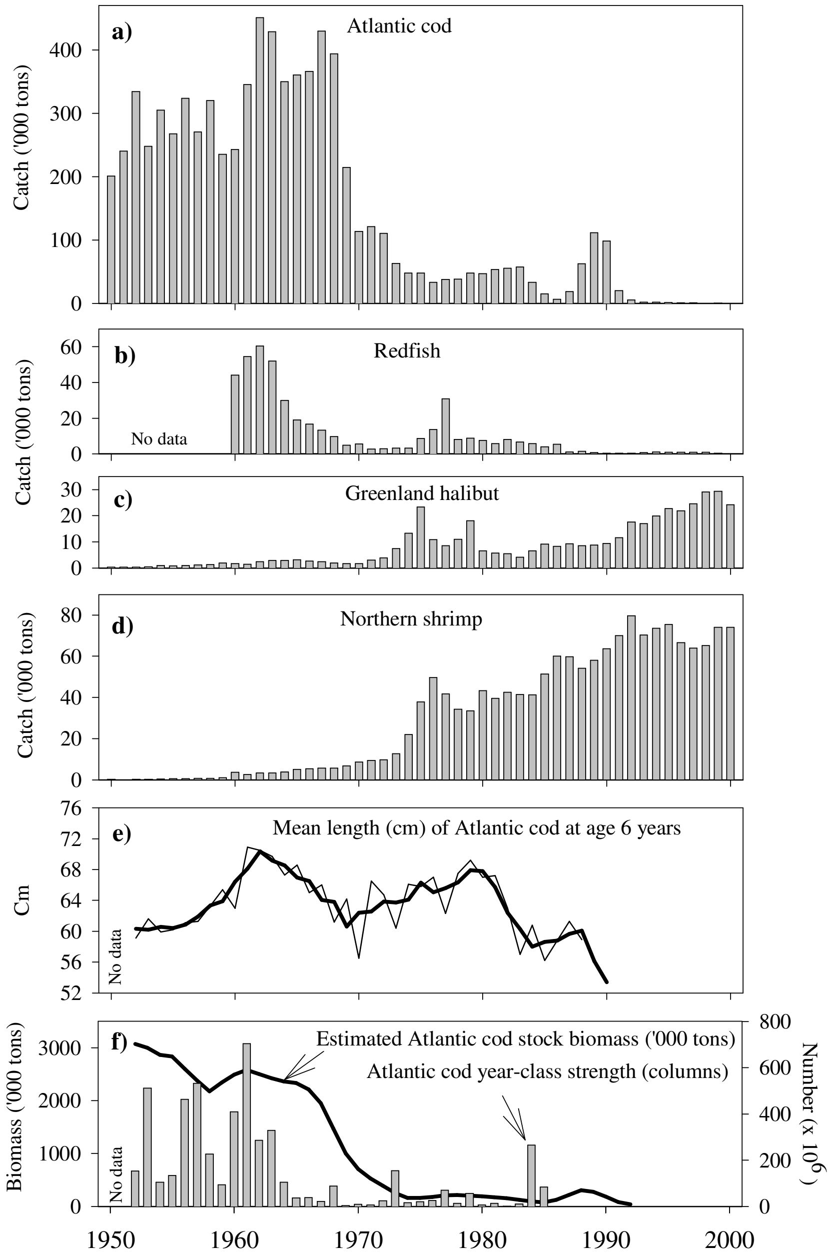


Figure 16. (a–d) Annual catches of four important commercial fish species off West Greenland (NAFO Subarea 1, inshore and offshore combined). (e) Mean length-at-age 6 of cod (heavy line is the 3-year running mean). (f) Estimated stock biomass and year-class strength of West Greenland cod. From Buch et al. (2004).

The catches of cod increased steadily only interrupted by the Second World War, where only the Portuguese were fishing at West Greenland. Afterwards it rapidly increased in the 1950s due to increased efforts and peaked in the 1960s with annual catches between 400,000 and 500,000 tonnes. A dramatic decline started in the late 1960s and afterwards catches of cod fluctuated on a much lower level with only two minor improvement due to strong year classes in 1973 and 1984/1985 (Figure 16f). After the early 1990s the cod fishery has been almost totally declined at Greenland. The dramatic decline in the late 1960s is also evident from the estimated cod biomass (Figure 16f). Catches of redfish has undergone a similar dramatic decline in the 1960s and the reported catches after the mid-1980s are at very low levels (Figure 16b).

The decline in the cod catches is believed, in large part, to be a response to climate variability. Cooling has caused cod to migrate southward during the 1980s and to a decrease in size-at-age (Hovgård and Buch, 1990; Riget and Engelstoft, 1998; Rätz et al., 1999; Horsted, 2000). The strong year-classes from 1961 to 1963 (Figure 16f) all showed up in large quantities at Iceland as age-7 cod (Schopka, 1991) indicating that the collapse in the late 1960s partly can be attributed to migration out of West Greenland. However, overfishing must also be taken into account. In the 1960s catches of cod peaked, while the total biomass decreased from about 1950 (Figure 16a,e).

The living conditions for the local cod stock at West Greenland worsened during the 1980s. Figure 16e shows the length of West Greenland cod at age-6 which is a measure of their condition. It increases during the 1950s, peaked around 1960, slowly declined in the 1960s but increased again slowly until the start of the 1980s. Hereafter their length decreased rapidly in the early 1980s and further decreased in the late 1980s. These decreases are likely linked to the cold conditions at West Greenland in the periods 1982–1984 and again from 1989–1994 (see Chapter 3.2) which marked the end for the cod fishery at West Greenland. The decrease centred around 1970 is due to the cooling associated with the "Great Salinity Anomaly" (Dickson et al., 1988, Belkin et al., 1998; Belkin, 2004).

At the time of the dramatic decline in the cod stock, catches of two other commercially important species, northern shrimp (*Pandalus borealis*), and Greenland halibut (*Reinhardtius hippoglossoides*) increased (Figure 16c,d). Overall, the catches of shrimp increased to more than 75,000 tonnes. In the early 1990s the catches decreased slightly as a result of management, but from the late 1990s it increased moderately. Even so the estimated shrimp biomass increased in the late 1990s indicating good living conditions for shrimps (Figure 17a). The positive reaction of shrimp may partly be explained by their larvae, which are less affected by cold temperature than cod. Moreover, predation pressure has been reduced by a declined cod stock (Koeller, 2000; Lilly et al., 2000) and by-catches has decreased the higher tropic level caused by the increased shrimp fishery (Kingsley et al., 1999; Pauli et al., 2001).

Since the late 1980s, the commercial fishing has moved southward (Figure 17b). This indicates that shrimp have migrated towards preferable temperature habitat due to the low temperature during the period 1989–1994 (see Chapter 3.2). From 1995 the average bottom temperature increased (Figure 17c), but they did not migrate back again. The present environmental conditions seem to be favourable for shrimp, as both the total biomass has increased as well as the recruitment in the late 1990s (Figure 17a).

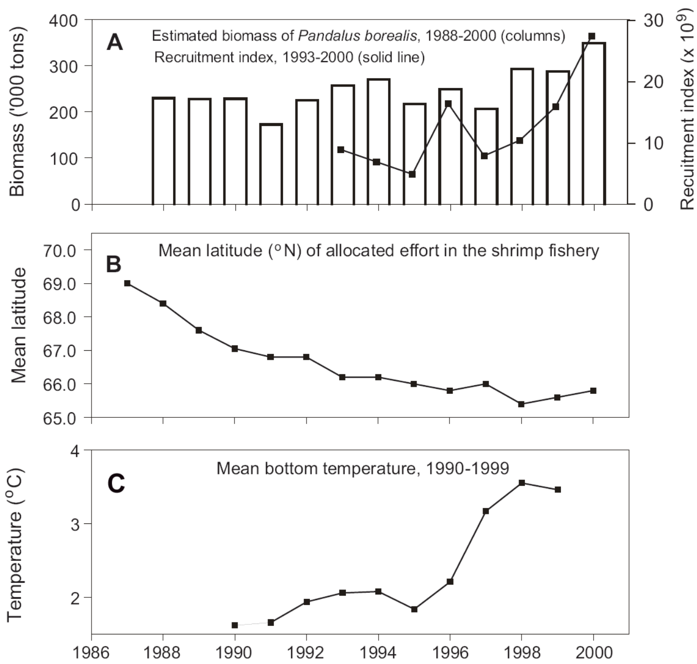


Figure 17. (a) Northern shrimp biomass indices and recruitment index from the annual shrimp survey. (b) Mean latitude of effort in the commercial shrimp fishery. (c) Mean bottom temperature during the survey. Modified from Buch et al. (2004).

3.2 Hydrographic conditions off Southwest Greenland

The surface waters around West Greenland and the Southeast Greenland are primarily composed of two very different water masses:

- Warm and saline Irminger Water (IW), a side branch of the North Atlantic Current.
- Cold and low-saline Polar Water (PW) originating from the Arctic Ocean.

These watermasses meet in the northern Irminger Basin and in Denmark Strait as seen in Figure 6. Here the main branch of the Irminger Current turns west towards East Greenland where it meets the PW, which is transported southward along East Greenland within the East Greenland Current (EGC) (Figure 7). It is the strength of these two currents that determines the hydrographic conditions around Southeast and West Greenland. In the Irminger Basin the two water masses flow side by side forming large meanders at the front where mixing takes place. PW is found at the surface on the continental shelf whereas IW is found over the continental slope and partly below the PW on the deeper parts of the continental shelf (Figure 18c).

As they round Cape Farewell the IW subducts under the PW (Figure 18b) forming the West Greenland Current (WGC). These watermasses gradually mix along West Greenland, but IW can be traced all along the coast up to the northern parts of Baffin Bay (Buch, 1990). At Cape Farewell IW is found as a 500–800 m thick layer on the continental slope with a core at about 200–300 m. The depth of the core gradually decreases from east to west as seen in Figure 18b, whereas the depth gradually increases from south to north to below 400 m in the northern Davis Strait and Baffin Bay.

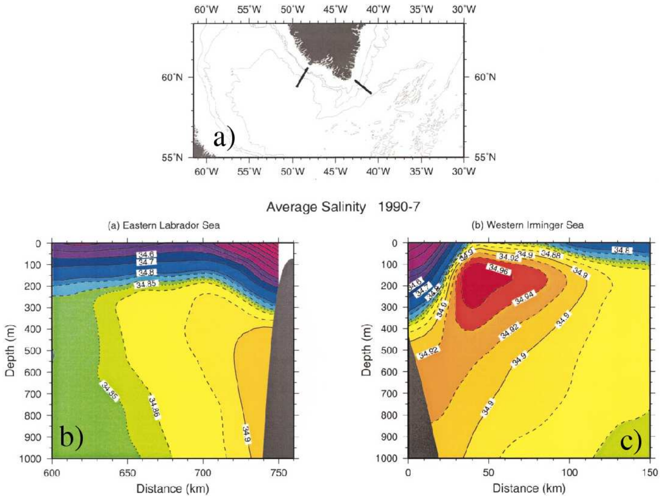


Figure 18. Mean upper-layer salinity sections for the period 1990–1997. a) Location of the two sections. Isobaths shown: 1000, 2000 and 3000 m. b) Eastern Labrador Basin. c) Western Irminger Basin. From Pickart et al. (2002).

On the fishing banks off West Greenland IW and PW dominates, as sketched in Figure 19. PW is continuously diluted by run-off from the numerous fjord systems. As the WGC reaches the latitude of Fylla Bank it branches. The main component turns westward and joins the Labrador Current on the Canadian side, while the other component continues north through Davis Strait.

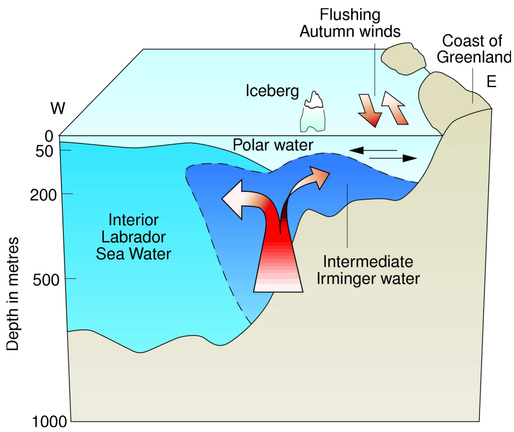


Figure 19. Sketch of the water masses off West Greenland in the Davis Strait region. From Valeur et al. (1997).

Off Southwest Greenland centred at about 61.5°N a source for of high eddy variability is found (Figure 20). These eddies are locally formed in the boundary current and propagates towards the interior of the Labrador Basin (Cuny et al., 2002; Lilly and Rhines, 2002; Prater, 2002; Pickart et al., 2002; Jakobsen et al., 2003). They mix water from the West Greenland shelf into the interior of the Labrador Basin. This has the effect, that the currents transporting IW and PW are extended much further offshore over deep water at West Greenland, contrary to the currents on the eastern side of Greenland, which are trapped to the shelfbreak (Figure 18; Pickart et al., 2002). The continental slope is particularly steep which is believed to be responsible to the high eddy kinetic energy in this specific place (Cuny et al., 2002; Buch, 2002). Baroclinic instabilities are formed at the front between the Irminger Water and the surrounding water masses and the intensity of the eddy activity is varying in time with the strength of the Irminger Current (Prater, 2002, Lilly and Rhines, 2002).

There is no literature available that describe the importance of these eddies for the transport of fish larvae and nutrient loads off West Greenland. But it seems natural, that an offshore transport would have a negative influence on pelagic fish larvae. This remains to be investigated in detail, e.g. by a coupled oceanographic, biologic and chemical field study.

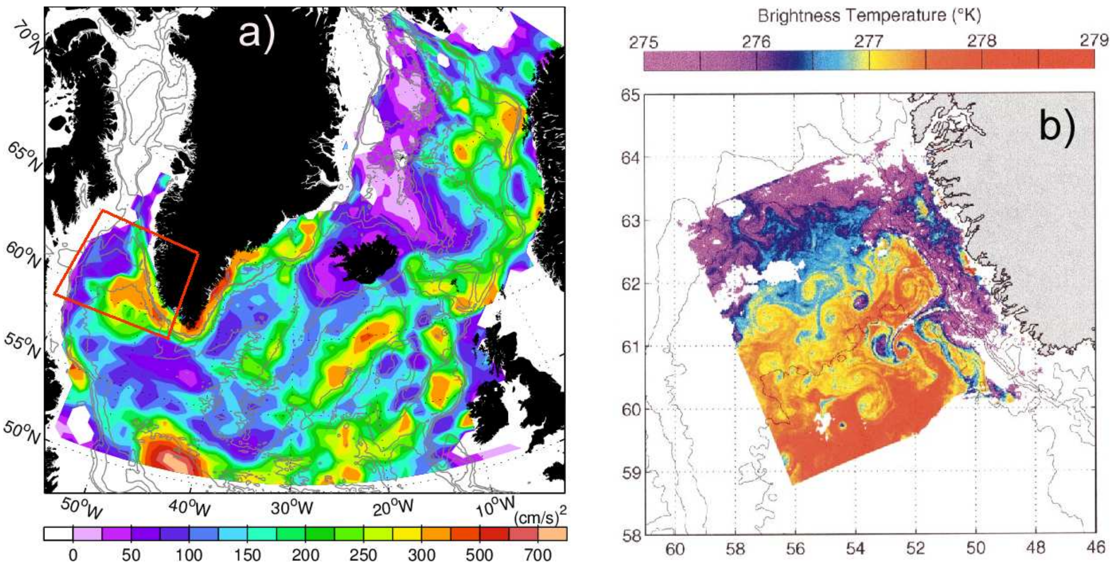


Figure 20. a) Distribution of eddy kinetic energy $0.5(\langle u'^2\rangle+\langle v'^2\rangle)$ calculated from high-pass filtered drifter data. Note the changing in scales at 300 cm^2/s^2. Bottom topography lines at 500, 1000, 2000, 3000 and 4000 m is shown. Reproduced after Jakobsen et al. (2003). b) Sea surface brightness temperature of the northern Labrador Basin marked by red square on a). Bottom topography lines at 500, 1000, 2000 and 3000 m is shown. From Prater (2002).

Over the shelf off West Greenland a totally different form of eddies is found (Figure 21) as described by Ribergaard et al. (2004). Here the formation of the permanent eddies is a result of the interaction between the complicated topography and the strong tidal currents in the area giving rise to a residual current around the banks. This can be explained by the concept of topographic steered flow. As a consequence of conservation of potential vorticity, the flow tends to follow isobaths with shallower depths to right. When the tidal flow is northward the current thus strengthen on the western side of a bank. Vice versa when the tide changes and flow is southward, the current strengthen on the east side of the same bank, and the mean flow is thus clockwise around the bank. The same argumentation gives counter-clockwise eddies around trenches. The most active area of permanent eddies is found between Fylla Bank (64°N) and Store Hellefisk Bank (68°N) where the tides are strongest.

At West Greenland the strongest tidal signal is located close to Nuuk at 64°N. The tides are primarily semidiurnal with large difference between neap and spring (1.5 m versus 4.6 m at Nuuk, Buch, 2002). Therefore the intensity of the anticlockwise circulation around the banks on the shelf will vary with a period of about fourteen days, as the strength of the barotropic tidal current and the tidal elevation is directly related.

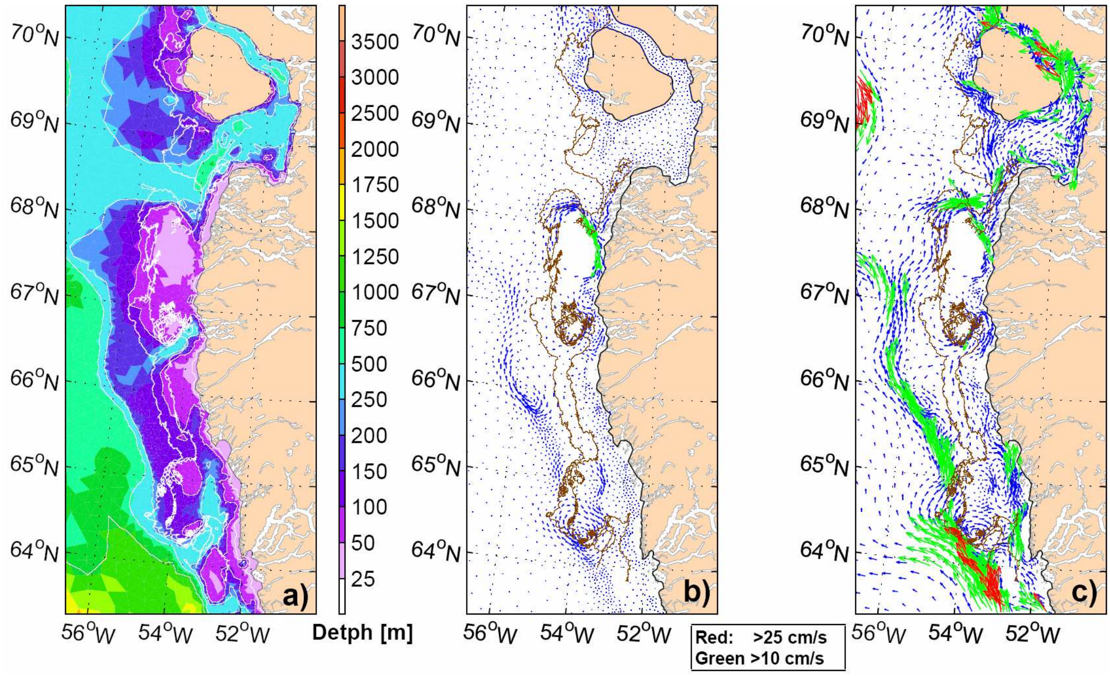


Figure 21. (a) Bathymetry, (b) Modelled barotropic mean currents (April to November) at West Greenland in 2000 and (c) the sum of the barotropic and baroclinic currents all at 50 m depth. Overlayed on the panels are trajectories of 2 WOCE-SVP drifters drogued at 30 m. From Ribergaard et al. (2004).

3.2.1 Atmospheric variability in Southeast and West Greenland

West and Southeast Greenland lies within the area which normally experiences warm conditions when the NAO index is negative (see Chapter 2.2). As can be seen from Figure 22 the annual mean air temperature in Nuuk shows large interannual variations of 1–2°C but sometimes nearly 5°C. The temperature was generally very low from the start of the timeserie in 1873 to the early 1920s. In the mid-1920s the mean annual temperatures suddenly increased by about 1°C at stay high at this level until the late 1960s. Then the temperature became slightly lower, but interrupted by three cold periods centred about 1970, 1982–1984 and 1989–1994. From the mid-1990s the temperature has remained fairly stable with slightly lower temperatures than during the warm period.

On the southeast coast of Greenland, the temperature at Tasiilaq showed a similar history (Figure 22). However, the interannual variations are in general much lower around 0.5°C. The regime shifts in temperature are much more evident in this timeserie. The cold period at the start of the timeserie came to an end between 1925 and 1926 whereas the warm period was interrupted by a decrease in temperature between 1964 and 1966. The following cold period was more evident at Tasiilaq than at Nuuk. A return to warmer climate conditions is seen from 1996. After 2000 the temperatures reached similar high values as in the previous warm period.

These temperature variations are closely linked to the NAO index as described in Chapter 2.2 and shown in Figure 10. At about 1925 the NAO index shifted from a generally positive phase to generally negative phases, which culminated in the mid-1950s and the 1960s. In the same period both Nuuk and Tasiilaq experienced warm conditions. High NAO values in the early 1970s, the early 1980s and the early 1990s are clearly reflected in temperature drops. However, after the mid-1990s the temperatures were fairly high despite of high NAO values. This was due to a displacement of the Icelandic low-pressure cell towards the northeast (ICES, 2000, 2001, 2002). In 2003 the low-pressure cell was displaced towards southwest and the high pressure-cell towards northeast resulting in a northwestward wind anomaly over the Iceland and Irminger Basins (Ribergaard and Buch, 2004). Therefore it should be stressed, that using NAO as an indicator for the climate at West Greenland should be done with caution, as the positions of the high-pressure and especially of the low-pressure cell varies over time.

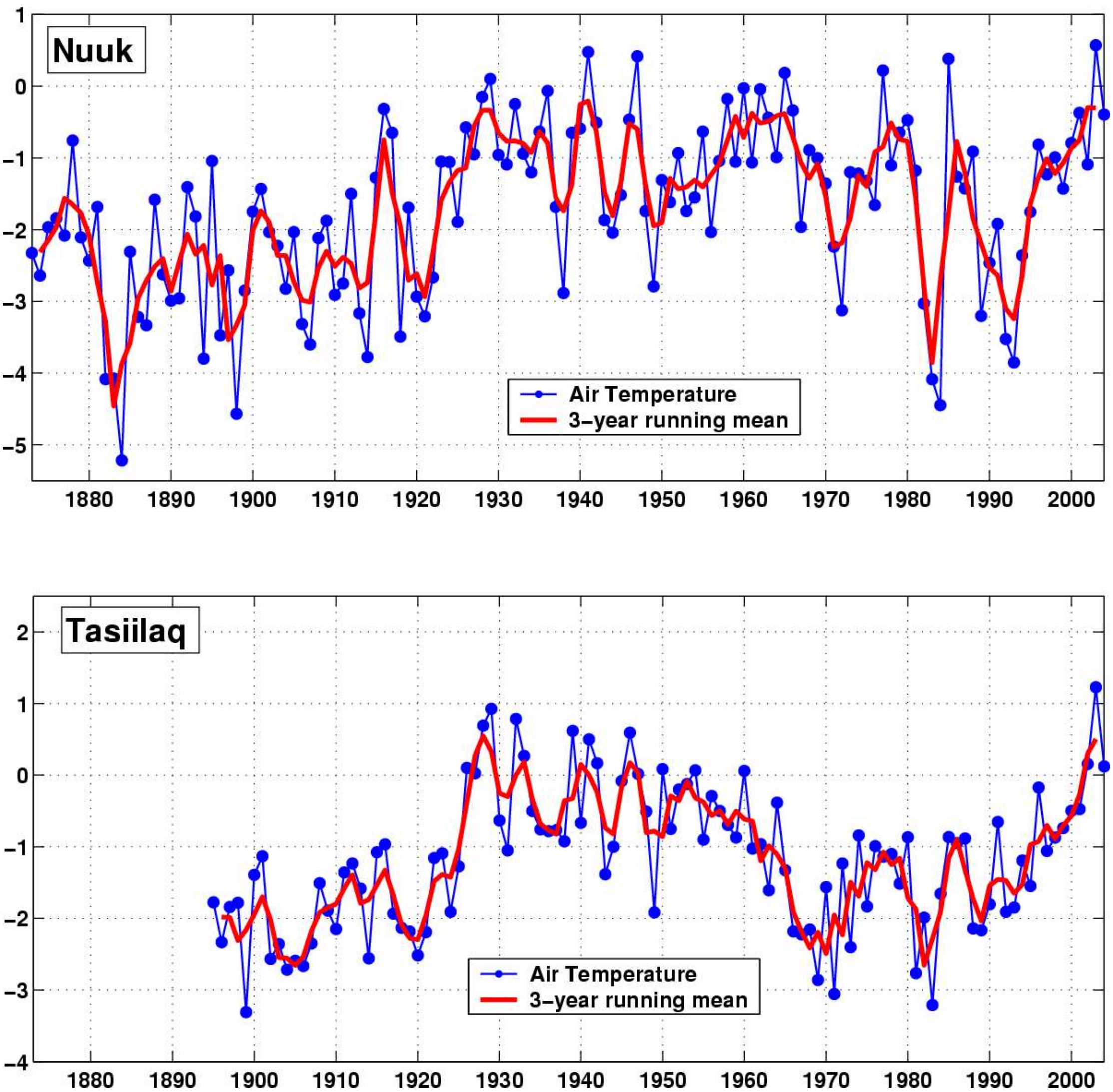
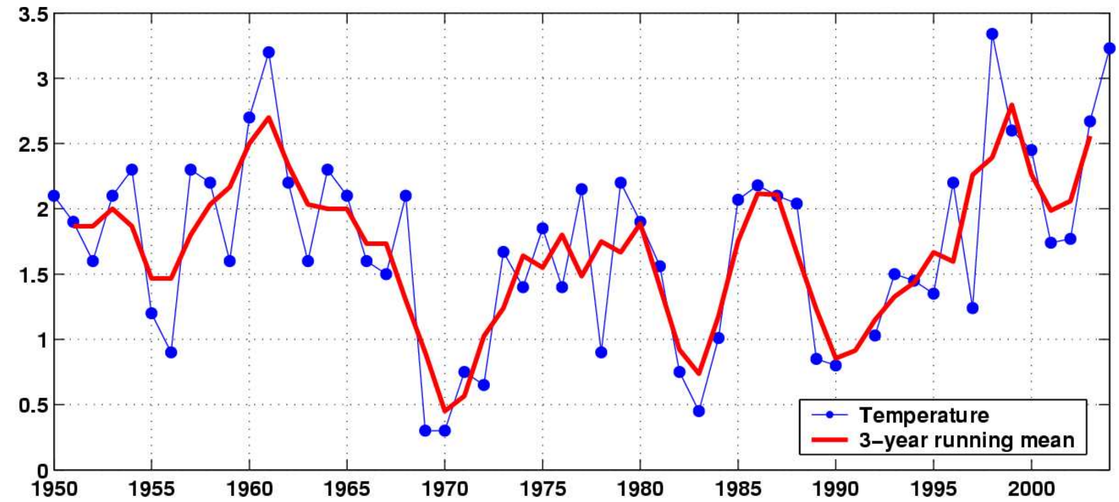


Figure 22. Annual mean air temperature southwest (Nuuk) and southeast (Tasiilaq) of Greenland for the period 1873–2004. See Figure 5 for location.

3.2.2 Hydrographic variability off Southwest Greenland

Changes in the ocean climate off West Greenland generally follow those of the air temperatures. Exceptions are years with great salinity anomalies i.e. years with extraordinary inflow of Polar Water (Dickson et al., 1988, Belkin et al., 1998; Belkin, 2004).



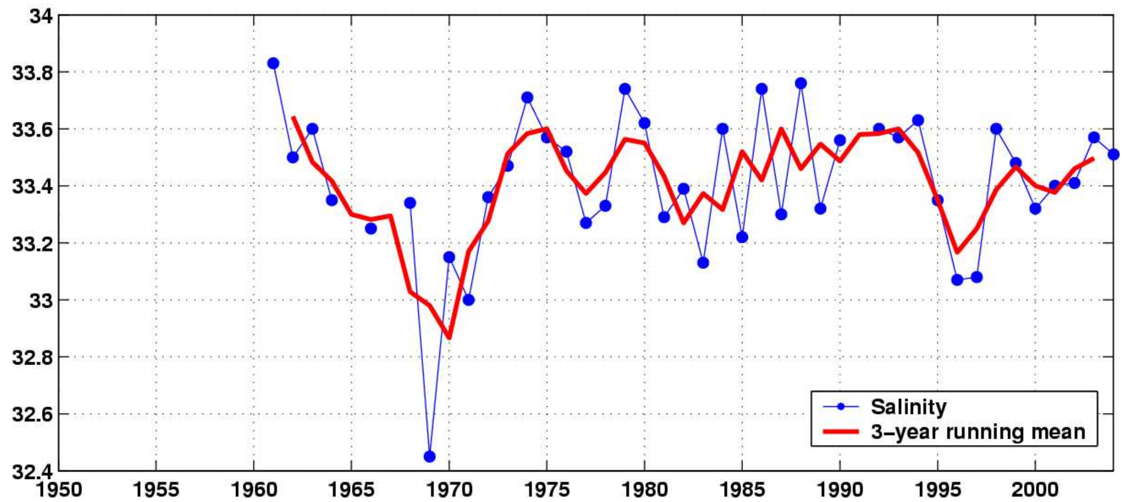
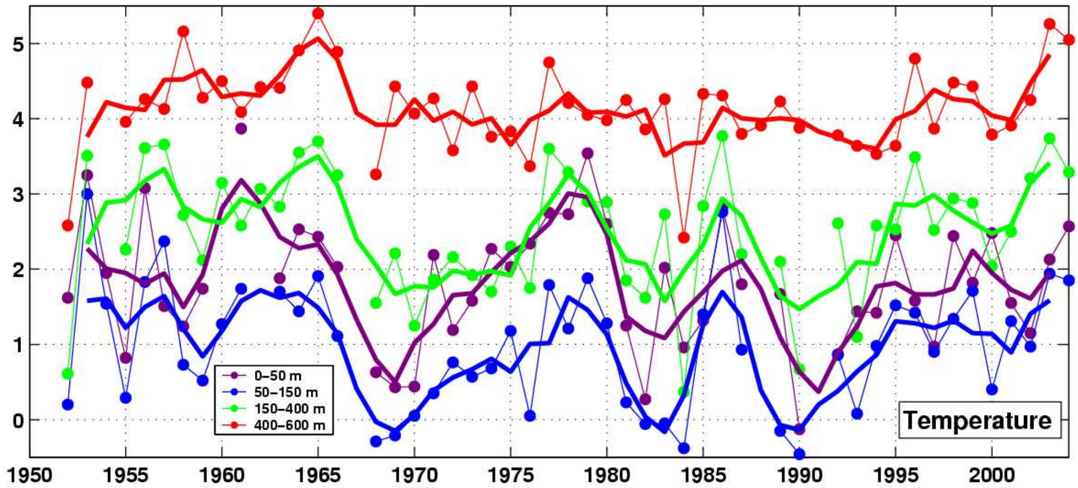


Figure 23. Mean surface (0–40 m) temperature (top) and salinity (bottom) on top of Fylla Bank st. 2. for the period 1950–2004.

The mid-June mean (0–40 m) temperature and salinity on top of Fylla Bank st. 2 is shown in Figure 23. They show large interannual variations, sometimes by more than 1°C in temperature and 0.5 in salinity. The 3-year running mean better reflects the large scale climate variations, because it smoothes out the high-frequently variations. Warm conditions were observed in the period 1950–1968. Around 1970 the temperature drops to the coldest period experienced due to an anomalous high inflow of Polar Water, which was closely linked to the "Great Salinity Anomaly" (Dickson et al., 1988, Belkin et al., 1998) and is clearly reflected in the salinity. At the same time, a shift to cold atmospheric conditions was observed reflecting the shift from negative to positive NAO values. In the early 1980s and early 1990s additional two cold periods were observed both in the atmosphere and the ocean reflecting the local cooling due to high NAO conditions. During the late 1990s higher ocean temperatures are observed despite of high NAO values, which was caused by a northeastward displacement in the low pressure cell. In 2003 a displacement of the low-pressure cell towards southwest and the high-pressure cell towards northeast resulted in anomaly southwesterlies winds over the Iceland and Irminger Basins causing the East Greenland Current through Denmark Strait to weaken and the Irminger Current to strengthen (Ribergaard and Buch, 2004). In 1997 low ocean temperature was observed despite quite warm atmospheric conditions. Low salinities suggest that this was due to a high inflow of Polar Water.



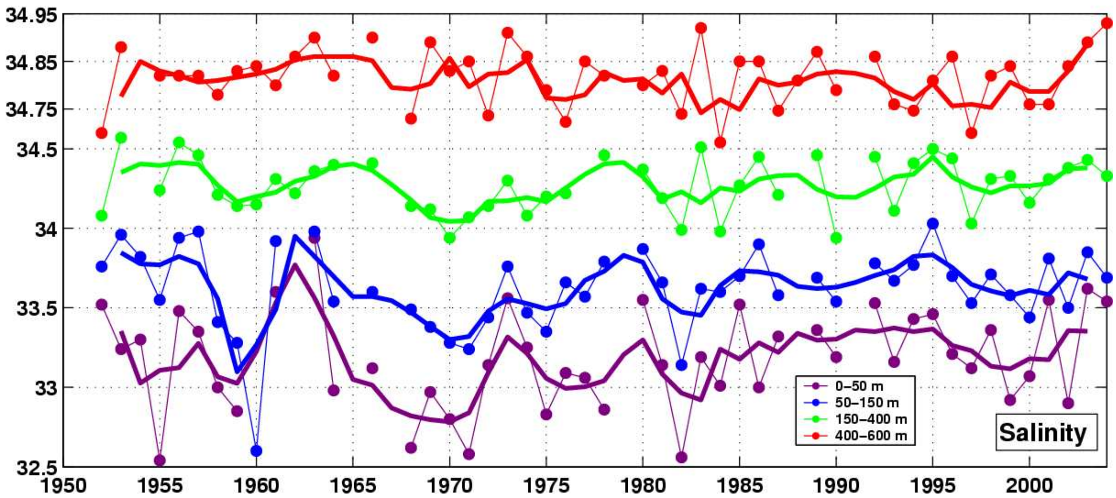


Figure 24. Mean temperature (top) and salinity (bottom) for the period 1950–2004 in four different depth intervals west of Fylla Bank (st. 4) over the continental slope. Note the change in scales at 34.75 for salinity.

Similar ocean temperature history is seen at the continental slope west of Fylla Bank in the upper 400 m (Fylla Bank st. 4, Figure 24). The core of the IW is found in the depth interval from 400–600 m. A clear decrease in the salinity and temperature in the end of the 1960s indicate a reduction of the inflow of IW to the area as expected for a change from negative to positive NAO values (see Section 2.2). Increasing temperatures are measured from the second half of the 1990s, but the salinities did not increase until the end of the 1990s. The same trend are also seen at a section off Cape Farewell (Buch et al., 2004), but here the salinity did increase form the mid-1990s. Similar variations were reported by Mortensen and Valdimarsson (1999) south of Iceland indicating increased strength of the Irminger Current since the mid-1990s. This relative long period of stable warm conditions may be beneficial for species like the Atlantic cod though the temperatures are slightly lower than in the warm 1960s.

3.2.3 Sea-ice and its variability

Sea-ice is important in Greenland waters. At West Greenland two different types of sea-ice dominate.

West-ice is first-year ice formed in Baffin Bay and Davis Strait. It normally does not affect the Southwest Greenland waters, because the inflow of warm IW within the West Greenland Current peaks in late autumn and early winter and limits its extent to the northern Hellefisk Bank (Buch, 1990; see Figure 5 for location). However, during cold periods its distribution is extended further south. The low temperatures in the around 1970, the early 1980s and the early 1990s are clearly reflected in the west-ice coverage (Figure 25).

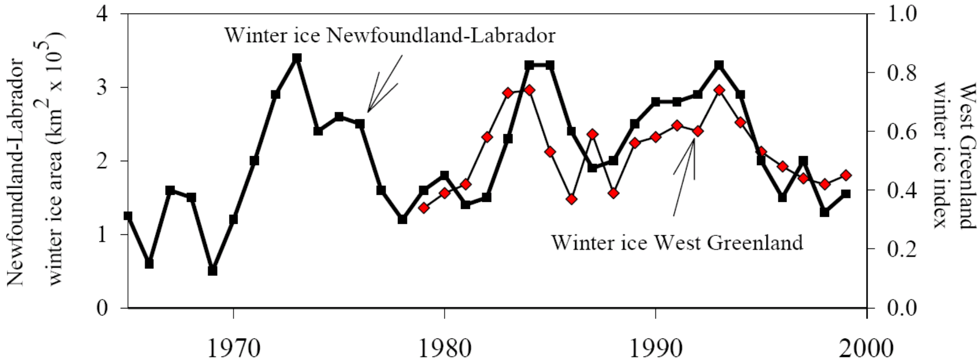


Figure 25. Area of winter (January–March) ice cover (km^2) off Newfoundland-Labrador, 1963–1999 from Drinkwater *et al.* (2000b) and area index of winter (January–February) ice cover "West-ice" off West Greenland (62°N–70°N) from Leif Toudal Pedersen, Danish Technical University. From Buch et al. (2004).

Multi-year ice ("storis") dominates in the southern part of West Greenland and is present 8–9 month a year in the Julianehaab Bright from Cape Farewell to about 61°N. It is originating from the Arctic Ocean and is carried to Southwest Greenland by the East Greenland Current. The amount of multi-year ice entering West Greenland waters shows great interannual variability as well as variations on multi-decadal timescales (Schmidt and Hansen, 2003). The mechanism behind the multidecadal variations is poorly understood, but may be linked to the circulation within the Arctic Ocean. The interannual variability is controlled by several factors such as outflow of sea-ice from the Arctic Ocean, formation of sea-ice along the northeastern coast of Greenland and in the Greenland Sea, and wind conditions over the Greenland and Irminger Basins and over the Iceland Plateau. Extreme amount of multi-year ice entered West Greenland waters in 1968–1970, 1982, 1984, 1989, 1990 and 1993 (Keld Q. Hansen, Danish Meteorological Institute, Greenland Ice Service, personal communication).

3.3 Relations between climate, hydrography and fishery resources off West Greenland

The shift in the community structure and landing composition of fish in Greenland in the second half of the 20th century took place at the same time as large climatic changes were observed in the West Greenland area. It therefore seems reasonable, that the observed changes in the living resources are largely driven by changes in the hydrography. Pedersen and Rice (2001) suggested that recruitment of fish and shellfish at West Greenland is related to water temperature, stability of watermasses and drift of fish larvae in the surface currents. These mechanisms are all related to the inflow of waters from outside West Greenland which in turn is related to the NAO. The individual strengths of the East Greenland and Irminger Currents have a strong effect on the hydrographic environment of the shelf areas in Southeast and West Greenland.

The ocean transport of salt and heat towards West Greenland is believed to have decreased drastically after 1970 (see Section 3.2.2 and Buch et al, 2004). This statement is not based on direct transport observations but on watermass analysis. This reduction seems to have had a negative effect on the recruitment success of the West Greenland cod and redfish stock, and a positive effect on the production of northern shrimp and Greenland halibut.

3.3.1 Goodbye cod, hello shrimp

Already in the 1950s, Hermann (1953) suggested a positive relationship between the cod year-classes and the temperature conditions off West Greenland. This was further investigated by Hermann et al. (1965) who found the relationship to be significantly linear correlated for temperatures above 1°C measured at the top of Fylla Bank in June. For low temperatures, the cod year-classes were at a constant low level. Brander (1994, 1995) found that temperature is responsible for most of the observed differences in growth between several cod stocks in the North Atlantic.

Calanus finmarchicus is the most important food source for cod larvae in West Greenland waters (Bainbridge and McKay, 1968). As the abundance of *C. finmarchicus* in West Greenland waters decreased from the warm to the cold period (Pedersen and Smidt, 2000), this may explain part of the relationship between temperature and cod-year classes.

The temperature decrease after the late 1960s resulted in poor conditions for cod at West Greenland, seen as a reduced growth, and the cod was migrating southward out of the area. By analysing the structure of the cod stock Buch and Hansen (1988) concluded, that the actual collapse in cod fishery in the late 1960s was caused by a general failure in recruitment. The intensive fishery in the 1960s reduced the cod stock substantially, but it would have been stabilized on approximately 180,000 tonnes if the recruitment were kept at the 1952–1963 level. In reality, the recruitment failed and the fishing mortality was high, which led to a doubling of the rate of decline of the total spawning stock biomass. Theoretical considerations on transport time combined with German and Icelandic surveys indicate, that the recruitment of cod at West Greenland is highly dependent on the drift of cod larvae and young fish from Iceland and less dependent on the local spawning at West Greenland. It was therefore suggested, that it is the changes in the oceanographic conditions in the Irminger Sea that drive the large fluctuations in the West Greenland cod recruitment rather than the local temperature.

Hansen and Buch (1986) reanalysed the recruitment data from West Greenland and found a linear temperature dependence between the year class size at age-3 (YC) of the local offshore cod stock recruitment off Greenland and the temperature on top of Fylla Bank station 2 (reproduced and updated in Figure 26). In some years this relation however fails as the YC was by far too high. They relate years of extreme YC to drift of cod larvae from Iceland to Greenland linked to increased influx of warm Irminger Water. Assuming that year-classes which distribute mainly off Southeast and Southwest Greenland originate from Iceland, they found the year classes, 1956, 1961, 1963 and 1973 to be Icelandic recruits. They argued that the year-classes 1953, 1957, 1960 and 1962 are of Icelandic origin. In some years, including 1973 and 1984, Astthorsson et al. (1994) found from 0-group survey that a considerable fraction of the Icelandic cod larvae drifted across the Denmark Strait.

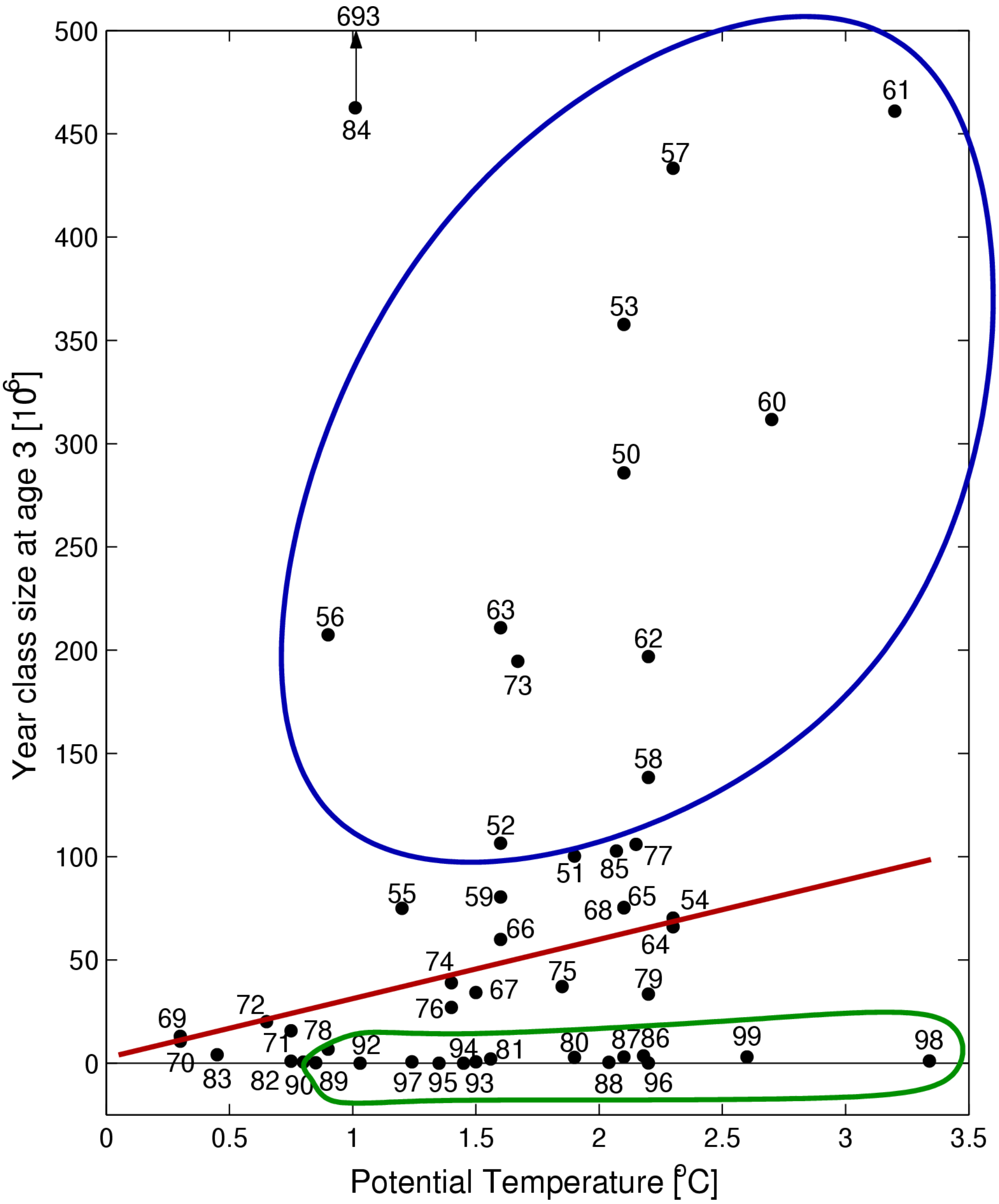


Figure 26. Year class size of West Greenland cod at age-3 (in tonnes) as function of the mid-June temperature on top of Fylla Bank (0–40m) for the period 1950–1999. Note, the 1984 year class is outside the y-axis. See text for explanation of the red line and the blue and green ellipses. From Ribergaard and Sandø (2004a).

Catches of haddock show similar variations as catches of cod at West Greenland, but as haddock do not spawn successfully at West Greenland, they must have drifted to

Greenland from Iceland as larvae. It is therefore a useful tracer for this drift. By using catches of haddock Hovgård and Messtorff (1987) and later Dickson and Brander (1993) also suggested the collapse in the West Greenland cod stock mainly to be caused by a less frequent inflow of young cod from Iceland since the early 1960s.

Following the ideas of Hansen and Buch (1986) Figure 26 can be interpreted as follows:

- The red straight line shows a linear dependence with year-class strength (YC) increasing with temperature. The years close to this line are believed to reflect recruitment from offshore West Greenland cod stock being closely related to the marine environment reflected by the temperature in the area.
- The years falling inside the blue ellipse have an extremely high YC, well above the linear temperature-YC dependence line (red). These strong YC are believed to be a result of drift of Icelandic cod larvae and eggs from the Icelandic spawning grounds to Greenland. All the years falling into this group are from the 1950s and early 1960s as well as 1973 and 1984. The cod landings (Figure 16a) were record high in the 1950s and especially in the 1960s - exactly the period where the drift from Iceland gave strong YC at Greenland. It is known that the strong YC in 1973 and 1984 are a result of drift of Icelandic cod and the effect of these strong year classes are seen few years later in the cod landings.
- The years falling inside the green ellipse are all years from the mid-1980s and the 1990s when the Atlantic cod stock at West Greenland was extremely low or almost non existing since 1992. Therefore the local recruitment was very low, as there were nearly no local spawning cod left. In the same period the drift of recruits from Iceland was low or absent.

Buch et al. (2004) relate the strength of the North Atlantic Oscillation (NAO) to the climate variability at West Greenland in the second half of the 20th century, reflected in both hydrographic data, sea-ice data, atmospheric data and biological data. They argue that the surface currents were altered during high NAO phase, resulting in a weakened Irminger Current and thereby decreasing heat transport to the West Greenland Fishing Banks in line with the ideas of Blindheim et al. (2001). At high NAO index, cold air masses from Canada and the Arctic flow over the Davis Strait and the Labrador Sea resulting in local cooling at West Greenland. Moreover, the weakened Irminger Current results in a decreased drift of cod larvae from the Icelandic Spawning grounds to West Greenland. All together this explains the major part of the dramatic collapse in the Greenland cod stock in the late 1960s, but other mechanisms such as overfishing must be taken into account.

Rätz (1999) suggested the failure of recruitments of cod in the late 1990s, when the temperatures got higher, could be due to significant by-catches in the expanding shrimp fishery combined with over-fishing on the spawning cod stocks. However, recently Stein (2004) reported high abundance of cod and haddock around East and West Greenland during the 2003 autumn surveys which is estimated to be similar in year-class as the 1984 year-class (Manfred Stein, personal communication). This could be a return of the Atlantic cod at West Greenland, but management of the fishery resources are needed in order to reduce the by-catching and prevent overfishing. However, history has shown that a shift in the climate conditions can occur within a short period, which dramatic consequences for the ecosystem. The physics behind regime shifts are poorly understood and needs further investigations in the future.

Northern shrimp larvae are less affected by the low temperature. They prefer relative low temperature in the range 1–6°C (Shumway et al., 1985) which may partly explain the positive relation of the West Greenland shrimp stock biomass to the colder conditions after the 1960s. However, the increase in the West Greenland shrimp stock biomass can most likely not be attributed solely to the changes in climate. The almost disappearance of the cod stock have reduced the predator pressure (Koeller, 2000; Lilly et al., 2000) and by-catches of the shrimp fishery contribute to keeping the predator pressure low (Kingsley et al., 1999; Pauli et al., 2001).

The relation between the climate changes and the observed shift in the marine ecosystem is, however, based mainly on the use of ocean temperatures as a proxy for the climate change. Only few scientific investigations to understand the ecological and physical processes behind changes in the marine ecosystem have been done.

3.4 Modelling the drift of shrimp larvae and plankton off Southwest Greenland

Variability in year class strength of northern shrimp (*Pandalus borealis*) are to a large extent determined during their pelagic larval stages, mainly driven by changes in the environmental conditions (Anderson, 2000; Pedersen et al., 2002), but the processes behind the observed recruitment variability in West Greenland is not well understood. As the currents on the Southwest Greenland shelf are strong, it is expected that advection during their pelagic stages are extremely important both directly and indirectly through advection of food, temperature and through predator/prey interactions (Pedersen et al., 2002).

The two years, 1999 and 2000, showed marked differences in the year class strength of the West Greenland shrimp population as indicated by shrimp size distributions (Figure 27) with the 1999 year-class being much larger than the 2000 year-classes. Pedersen et al. (2002) suggested that differences in the wind conditions may have been a key factor for larval shrimp food production and survival.

Motivated by this hypothesis Ribergaard et al. (2004) modelled the currents in 1999 and 2000 using a barotropic 3-dimensional hydrodynamic finite element ocean model forced only by tides and realistic windfields from an operational atmospheric model for Greenland (DMI-HIRLAM). High spatial resolution was given on the shelf and over the shelfbreak, whereas coarser resolution was used in the interior Labrador Basin. A barotropic model was used in order to separate the influence of winds from advection of water masses. Afterwards a diagnostic baroclinic field was added based on observations from 2000.

The current fields were used as forcing for a particle drift model simulating the drift of shrimp larvae from larval release to settling. The particle model was developed to work on the same finite element grid as the ocean model in order to get the full value of unstructured grid (Ribergaard et al., 2004; Ribergaard and Sandø, 2004a). Random walk diffusion was included (see e.g. Hunter et al., 1992; Visser, 1997; Spangol et al., 2002; Ribergaard et al., 2004a). Four different release areas were defined with progressively later time of hatching from south to north. The hatching followed a binominal distribution with duration of one month each. Particles were kept at fixed depths: 30, 50, 80 m.

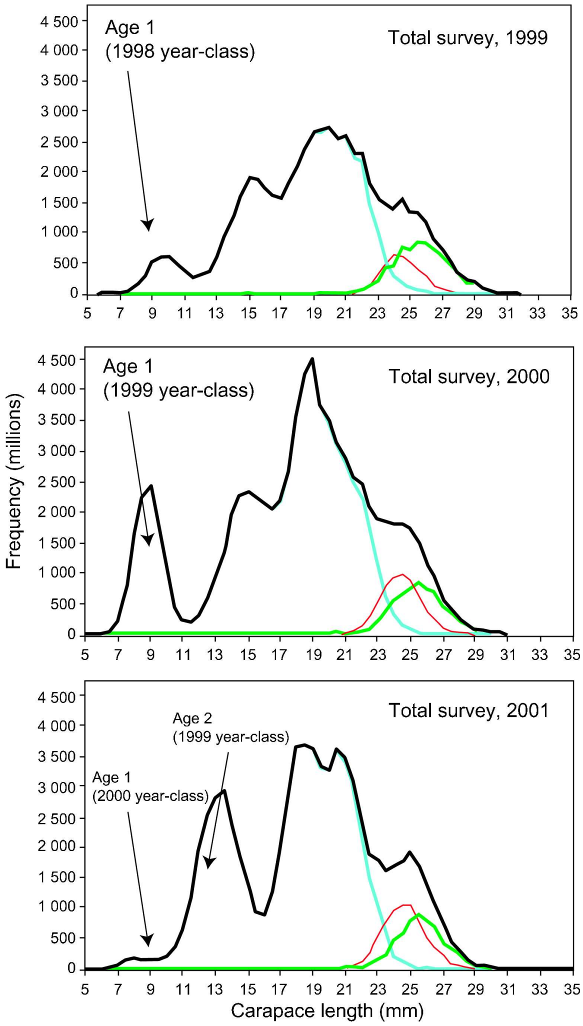


Figure 27. Length distribution of West Greenland shrimp for the surveys during the period 1999–2001. Black line: total, blue line: males, red line: primiparous females, and green line: multiparous females. From Pedersen et al. (2002).

Off West Greenland Ribergaard et al. (2004) found strong residual anticyclonic circulation around the shelf banks north of 64°N basically caused by conserving of vorticity, i.e. topographic steered flow (Figure 21). They are formed in areas dominated by strong tidal currents and steep topography. Trajectories of two surface drifters drogued at 30 m depths confirms their existence.

The permanent eddies had a large influence on the particle tracking. Particles tended to have long retention times over these shelf banks even with high diffusion coefficient. Relating the particles to shrimp larvae, this has important consequences for their food availability. Moreover, larvae-predators will have easier access to their food, as shrimp concentrates on the shelf banks. Only few particles drifted towards the Canadian shelf, which was a bit surprising, but this finding is certainly related to the large retention over the shelf banks.

Initially a particle model was used with regular grid but same random diffusion applied and interpolated the current onto this grid. This solution turned out to give completely different results than the finite element drift model, which fully benefits on the increased resolution on the shelf. Due to the much coarser resolution on the continental shelf, the particles did not follow the depth contours as strict as in the finite element model, and the permanent eddies were weakened due to interpolation. As a result the retention times over the fishing banks was significantly reduced in the model. This clearly demonstrates the importance of including all the information gained by a fine resolution model.

The model setup was further used for estimating the potential advection of plankton across the shelf banks and compared visually to observations (Pedersen et al., 2005). Upwelling was favoured west of the shelf banks for northerly winds resulting in offshore Ekman transport, which probably increases productivity and carbon cycling over the shelf banks. The circulation around the banks seems to create retention areas entrapping plankton for periods. However, the observational dataset is much too coarse temporally, but also spatial, to verify these findings.

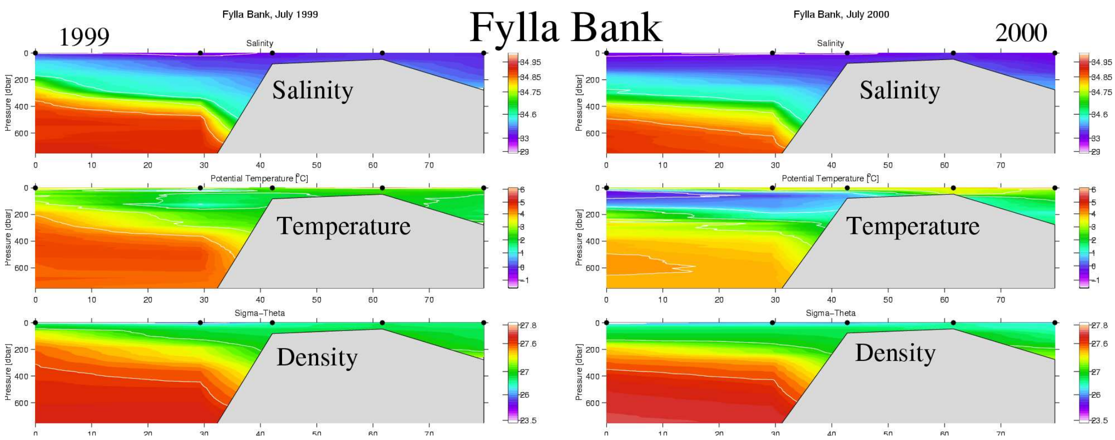


Figure 28. Salinity, temperature and density on the Fylla Bank transect (0–600 dbar) in early July 1999 (left) and early July 2000 (right). Same colormap has been used for both years.

The results from the particle tracking was almost identical for 1999 and 2000 indicating that the direct advection was of minor influence for the observed differences in year-class strength. Hydrographic conditions were however different between 1999 and 2000 (Figure 28) though small on decadal timescales (see Figure 23 and Figure 24). West of Fylla Bank the salinity and temperature in the IW was higher in 1999 compared to 2000 and the interface between this water and the PW above was closer to the surface. Sharper front in 1999 compared to 2000 between the IW and the PW and a smooth and hot core of PW indicates, that the IW component was dominating, and the PW component was weak in 1999 compared to 2000. Note, the mean (0–40 m) temperature on top of Fylla Bank st.2 were similar for the two years (Figure 23), because the temperature on the upper few meters was high in 1999. Below this upper surface layer the PW temperature was much higher in 1999 (e.g. compare 50–150 m in Figure 24). Similar hydrographic differences were observed all along Southwest Greenland on five additional sections. Altogether, the conditions was warmer and because the interface was closer to the surface, wind driven upwelling could easier reach the IW in 1999 than in 2000 enhancing the nutrient loads to the surface waters.

For investigating these differences in further details a prognostic 3-dimensional setup was needed. However, the model reveals serious problems in baroclinic mode. The deep water and the shelf water mixes along the seafloor, a well known problem in models using terrain following coordinates named the pressure gradient error (Kliem and Pietrzak, 1999). Several attempts were made to reduce this problem like introducing z-coordinates in the surface layers, a new and better advection scheme (Kliem, 2004) and several of other fixes. However, the model did not succeed a long run in baroclinic mode.

It is speculated that this is a general problem in 3-dimensional finite element models, and may not be related to the pressure gradient error. Until now no successful integration of more than a couple of months has been reported. In finite element modelling, the continuity equation is reduced to a wave equation for the surface elevation by using the cross differentiating the shallow water momentum equations in the x and y direction (Lunch et al., 1996). By this trick a first order differential equation has become a second order one. This could give additional solutions for the vertical profile. This certainly needs further investigations in the future.

3.5 Modelling the drift of cod larvae and eggs from Iceland to Greenland

The hypothesis, originally suggested by Hansen and Buch (1986) and presented in Chapter 3.3.1, that the strong year classes of offshore West Greenland cod mainly are recruits from the Icelandic spawning grounds, was tested by Ribergaard and Sandø (2004a) using particle tracking for the period 1948–2001. Random walk diffusion was added and taken proportional to fluctuation velocities calculated from a surface drifter set from the 1990s analysed by Jakobsen et al. (2003). Surface currents were taken from a regional nested ocean model of the North Atlantic, forced by NCAR/NCEP data and with a resolution of 20–25 km in the Denmark Strait region.

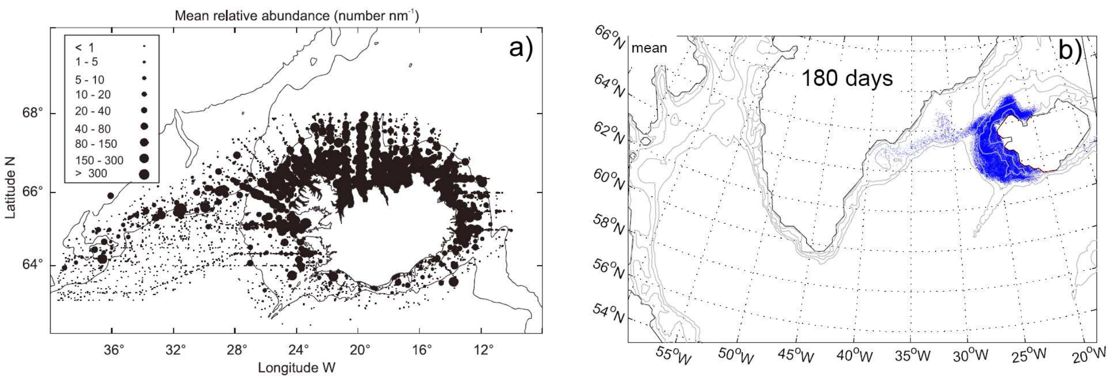


Figure 29. a) Mean (1970–1998) relative abundance of pelagic juvenile cod from Begg and Marteinsdottir, 2000a. b) Modelled mean distribution after 180 days (particles 120–180 days underway) using monthly mean currents for the period 1948–2001. Depth contours are 500 m (left) and 100, 200, 500 and 1000 m (right). From Ribergaard and Sandø (2004a).

Particles were released off southwest Iceland during a 2 month period as a binomial distribution centred around May, 1st. The spawning site used for particle tracking was chosen close to the 230 m contour forming a line from south to north. The fraction ending up north of Iceland, at Greenland and the fraction staying on the shelf southwest of Iceland after 180 days was calculated for each individual year for the period 1948–2001. The fraction ending up at Greenland (Figure 30) was compared with the observed year class strength at West Greenland (Figure 26) assuming that exceptional strong year-classes was a result of recruitment from Iceland. It turned out that the modelled drift towards Greenland was far from similar to the observations, assuming the recruitment hypothesis is correct, and consequently they could not confirm the hypothesis originally given by Hansen and Buch (1986).

Large interannual variations in the modelled drift motivated Ribergaard and Sandø (2004a) to group the individual years into years with strong and weak year-classes at West Greenland. From these they constructed mean surface current fields representative for the two regimes on a monthly basis. They initiated the particles from several locations on the Southwest Icelandic shelf. They found slightly more particles drifting to Greenland using the high year-classes currents compared to the low year-classes, but these differences were not significant. Similar, they followed the hypothesis of Buch et al. (2004), that the NAO index during the period was representative for the strength of the Irminger Current and thereby the transport towards Greenland. They constructed surface currents representative for high and low NAO years, but they did hardly find any differences. However, the particles that end up at Greenland cross Denmark Strait at about 66°N, which is similar to what is observed from the 0-group survey around Iceland.

Form all experiments it became obvious, that the final position of the particles was highly dependent on the initial position, despite of random walk diffusion and a 180 day long integration forward in time. Particles close to the Icelandic coast seem to be trapped close to the coast. Similar difficulties in transporting particles away from the Icelandic coast are found in a similar particle tracking experiments using currents from an ocean model with much higher spatial resolution (Ingo Harms, personal communication). According to e.g. Marteinsdottir et al. (2000), the main spawning ground at Southwest Greenland is found very close to the coast. However, this is mainly based on trawl information started in the early 1970s. Based on the particle tracking Ribergaard and Sandø (2004a) suggested that the main spawning grounds could have changed position in the past, but this remains to be investigated in detail using observation.

Even though Ribergaard and Sandø (2004a) could not confirm the drift hypothesis by Hansen and Buch (1986), they could neither reject it because:

- The quality of the ocean model in the region is questionable (see Section 3.5.1).
- The coarse NCEP/NCAR might be too coarse to describe the conditions in the Denmark Strait region, where the Polar Front is sharp. Only a small displacement is needed to shift a particle from e.g. the Irminger Current to the East Greenland Current (see Section 3.5.2).
- Related, daily mean winds reduces the peak winds, which could happen to be important for the cross-frontal transport in Denmark Strait.
- Related, weekly current fields also reduces the high frequency variability, that might be important in the exchanges between the IC and the EGC.
- If spawning grounds have changed position in the past, this could explain part of the drift.
- The abundance could have changed considerable in the past.

For a future drift study, Ribergaard and Sandø (2004a) recommended higher spatial and temporal resolution in both the ocean model, but also in the atmospheric model in order to describe the cross-frontal transport in the Denmark Strait correctly.

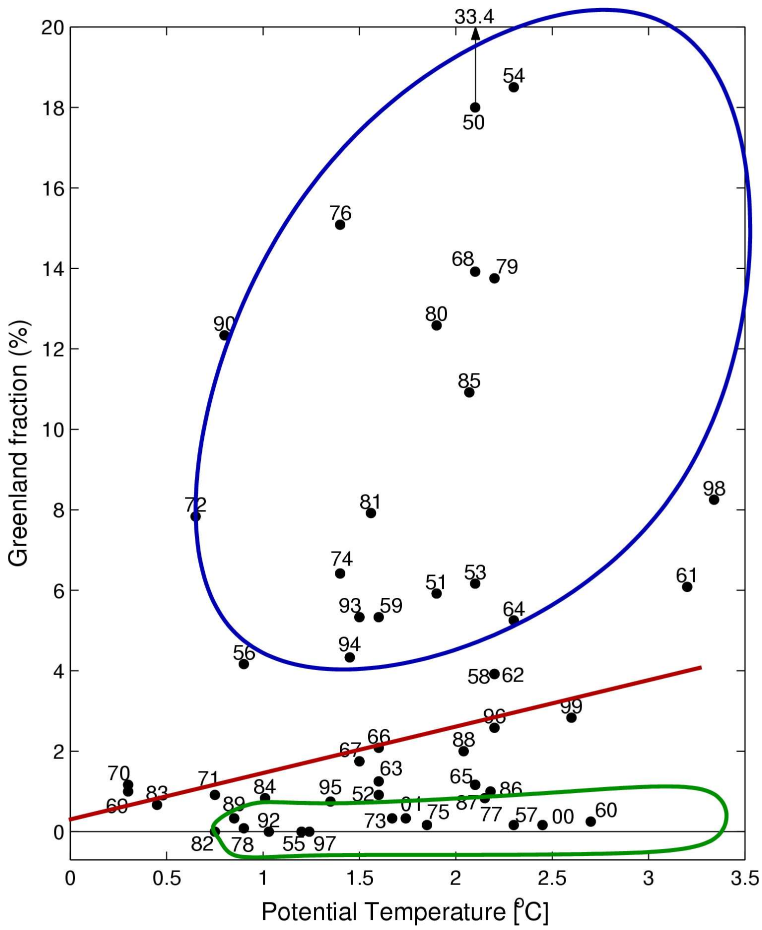


Figure 30. Pseudo year-class strength calculated as percentage of particles ending up at Greenland as function of the mid-June temperature on top of Fylla Bank (0–40m) for the period 1950–2001. Note, the 1950 year-class is outside the y-axis. Data from Ribergaard and Sandø (2004a).

3.5.1 Validation of the ocean model in Southeast and West Greenland waters

The ocean model used has proven its ability of reproducing the seasonal and decadal temperature variability in the Faroe-Shetland inflow waters observed in the past 50 years. However, Ribergaard and Sandø (2004b) found it necessary to validate the model performance for the Northwest Atlantic region with special focus on the surface currents in the Irminger Basin, the Denmark Strait and Southwest Greenland. The quality of the particle trajectories is highly limited by the quality of the ocean model delivering the current field. Therefore an understanding of the strengths and weaknesses of the calculated current field is a prerequisite for a proper interpretation of the particle transport.

The ocean model did not perform as well on the western side of the North Atlantic as on its eastern side. They found the surface currents to be generally too weak compared to the currents derived from drifters from the 1990s by Jakobsen et al. (2003). This was most evident in the Polar Water branch of the East and West

Greenland Currents, which was significant too low and almost absent off West Greenland. This could in part be related to the extent of multi-year-ice (storis) is much too low in the model, which directly affects the temperatures at South East and West Greenland and indirectly changes the baroclinic currents. As soon as the sea-ice has gone, the temperature in the surface layer is increased. Further, melting of sea-ice will normally keep the salinities low, but as it has gone, the salinities increase gradually by mixing with sounding waters.

In order to examine the annual variations in temperature and salinity, a harmonic analysis was preformed. It turned out, that the annual variations in salinities in the shelf are much too low in the model by about 0.5 compared to the ¼ degree World Ocean Atlas 2001 (WOA2001) climatology, whereas the annual variation of temperature is to high with 1–2°C or even more on the Southeast Greenland shelf caused by the lack of multi-year ice.

The salinity variations at Fylla Bank are poorly represented in the model, while the temperature variations are much better reproduced. In the model, there is no sign of the "Great Salinity Anomaly" in the late 1960s, but the two following cold periods in the 1980s and 1990s are easily seen. Thereby Ribergaard and Sandø (2004b) concluded that the air-sea exchanges appear to be well represented in the model whereas the advection processes are less well represented in the Irminger Basin / Labrador Sea area.

3.5.2 Statistics of the drift towards Greenland based on surface drifters

Surface drifters was used for estimating the transport of cod larvae from Iceland to Greenland (Ribergaard and Sandø, 2004a). They only considered drifters that at some instant are found on the Southwest Icelandic shelf (for exact criterion see Ribergaard and Sandø, 2004). They grouped each drifter into four groups depending on their final position (Figure 31 and Table 1) and compared their results to statistics of the 0-group surveys around Iceland for the period 1970–1992 (Figure 5 in Astthorsson et al., 1994). Astthorsson et al. (1994) also included the fraction staying at west Iceland, which was skipped in the drifter statistics.

Table 1. Statistics of final position of surface drifters. The colours in brackets refers to the colour used in Figure 31. Note, both the initial position and time is different for all the drifters and the statistics is only based on 81 drifters.

Final position	**Percentage**
Polar Water branch of East Greenland Current (blue)	20–25
Irminger Water branch of East Greenland Current (red)	10–15
North of Iceland within the North Icelandic Irminger Current (green)	50–60
South of Iceland (magenta)	5–10

Both statistics found slightly more than half with final positions north of Iceland, but Astthorsson et al. (1994) only found about 10 percent at Greenland, whereas Ribergaard and Sandø (2004a) found 35–40 percent. However, of these 20–25 percent ends up in the cold Polar Water component of the East Greenland Current. However, the cod larvae that end in Polar Water have less chance of survival, caused by the low temperature and thereby slow growth, compared to the cod larvae drifting within the warm Irminger Water. Therefore our estimate derived from the drifters is likely representative for the drift of cod larvae in a normal year.

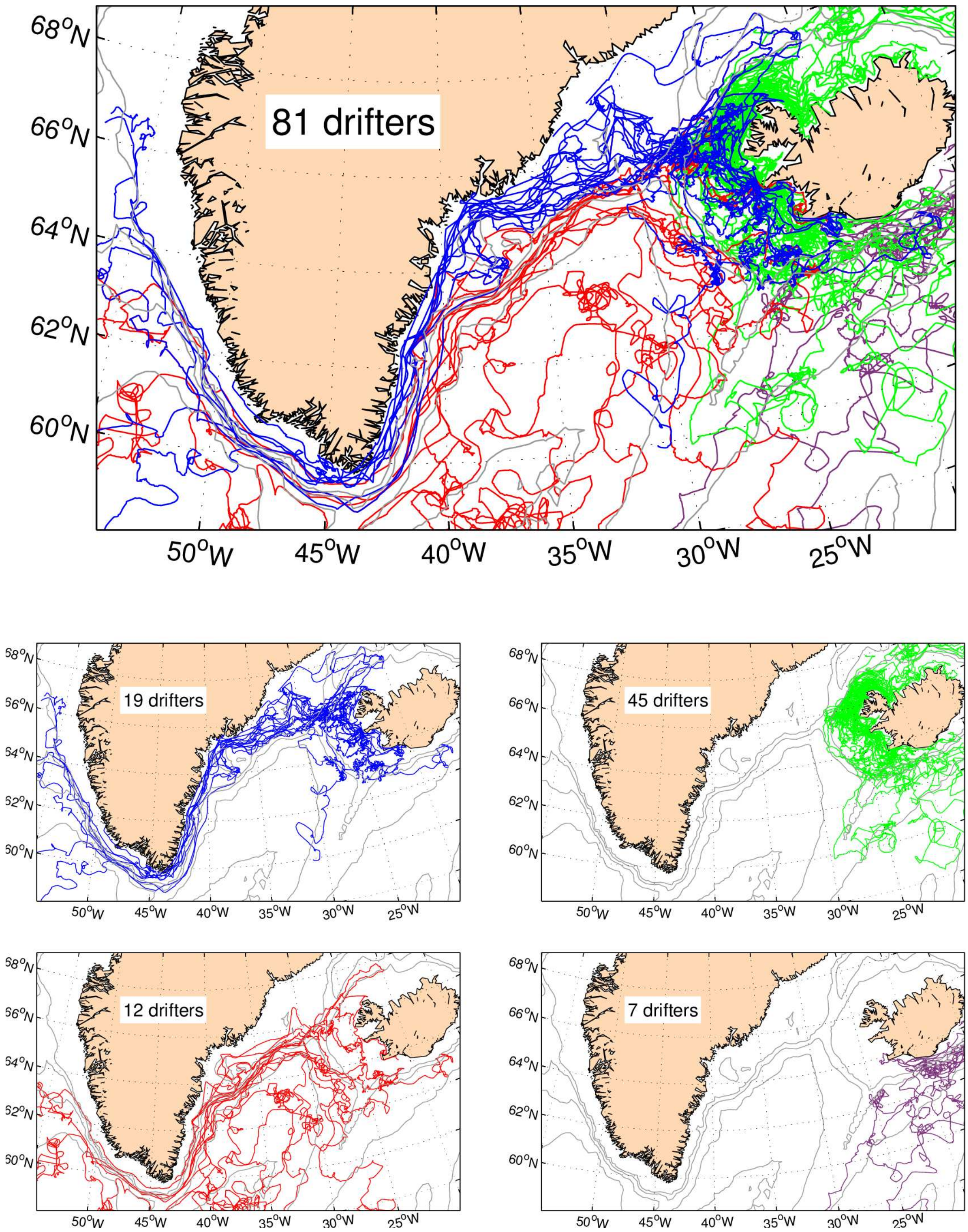


Figure 31. Low pass filtered trajectories of surface drifters drogued at 15 m in the 1990s. Only drifters passing close to the southwest Iceland shelf are considered defined in Ribergaard and Sandø (2004a). In top all drifters are displayed with different colours according to their final position and below the four groups are printed alone. Blue: Cross the Denmark Strait and continue in the East Greenland Current closest to the Greenland coast. Red: Cross the Denmark Strait and continue in the Irminger Water located at about the 500–1000 m depth contour. Green: End north of Iceland. Magenta: flowing eastward south of Iceland. Depth contours shown: 500, 1000, 2000, 3000 m. From Ribergaard and Sandø (2004a).

In the Denmark Strait the Polar front between the Irminger Water and the Polar Water is very sharp. From the drifter set (Figure 31) and the mean distribution of 0-group larvae (Figure 29a) it is evident, that the transport from the Southwest Icelandic shelf to Greenland mainly takes place within the Denmark Strait at about 66°N. Some physical factors can cause such a displacement (see also Ribergaard and Sandø, 2004a):

- Instability in the frontal system forming intense meanders and eddies. The eddies act as small scale mixing agents (on the order of 10 km).
- Ekman transport forced by changing wind conditions over the Denmark Strait. The area is dominated by low-pressure systems which, after deepening off the east coast of North America, move into the Iceland region. This introduces highly variable wind conditions in the region.

3.6 Summary

The hydrographic conditions off West Greenland are related to the large-scale climate conditions expressed by the North Atlantic Oscillation (NAO). A shift in NAO phase in the late 1960s from mainly negative to positive values resulted in decreased inflow of warm saline water to the area. Extreme cold conditions in the early 1980s and early 1990s was related to direct atmospheric cooling during high NAO values.

The West Greenland economy, formerly being highly dependent on a rich cod fishery, is today almost entirely dependent on the Greenland shrimp stock. A positive relationship is found between recruitment of cod and water temperature as a proxy for the hydrographic condition. The decreased inflow of warm saline water since the late 1960s indicated reduced inflow of cod larvae from the Icelandic spawning grounds and extreme cold conditions resulted in migration out of the area. The shrimp reacts positive to the lower temperatures, but the reduction of the cod stock also decreased the predation pressure and by-catches of cod keep the pressure low.

Particle tracking of the drift of cod larvae and eggs from Iceland to Greenland did not confirm that strong year-classes of West Greenland cod are recruited from Iceland. The model did not display the right distribution of salinity, temperature and sea-ice in the Western Atlantic, which in turn gives an incorrect circulation, making the above result inconclusive. A strong dependence on the initial position of the final position suggests that the Icelandic spawning grounds have changed positions over time.

An analysis of surface drifters suggests that the drift of cod larvae and eggs from Icelandic to Greenland waters occurs in the Denmark Strait near 66° N. This drift seems to be very dependent on wind conditions and eddies formed in the frontal system. In order to simulate this transport correct, higher spatial and temporal resolution is needed in both the ocean and the atmospheric model.

High resolution ocean modelling revealed clockwise circulation around the main fishing banks from Fylla Bank to Hellefisk Bank (see locations in Figure 5). These increases the retention times of fish larvae and plankton on the banks and influences the food supply and makes predation on fish larvae easier.

4 Final discussion and outlook

Changes in the climate system strongly affect the ecosystem in many ways all over the world. However, the mechanisms behind major regime shifts in the climate system on decadal timescales are generally poorly understood. In the Pacific Ocean, the physics associated with the El Niño/Southern Oscillation (ENSO) are quite well understood, whereas the physical processes behind the multidecadal variations, reflected by the Pacific Decadal Oscillation (PDO) remains to be explained and verified in detail. Both factors affect the ecosystem, but whereas El Niño events result in large variations in the fishery, the PDO lead to shift in the whole community structure and landing composition of fish seen e.g. by the altering anchovy and sardine regimes in the Pacific.

In the North Atlantic, annual to decadal variability is associated with the North Atlantic Oscillation (NAO). However, regime shifts on multidecadal timescales have been observed which totally changes the ecosystem. For example, the West Greenland cod arrived in the mid-1920s, the stock then suddenly collapsed in the late 1960s, and it is probably returning in these years caused by a shift in climate since the mid/late 1990s. These changes are driven by long periods of generally high or low phases of NAO, but it is unknown why the climate system behaves like it does.

Unlike El Niño events, which can be forecasted about half a year ahead using sea surface temperatures, changes in the NAO are so far impossible to predict. However, the NAO is a useful indicator when analysing the North Atlantic climate back in time, but it should be used with care and always in connection with a physical mechanism, connecting the NAO to the phenomena described.

In this study it is found, that the hydrographic conditions in the Southwest Greenland waters are generally well described by changes in the NAO. There are however exceptions in the late 1990s. Strong local cooling and reduced inflow of warm Irminger Water is related to positive NAO values, whereas negative values of NAO result in increased inflow of warm saline Irminger Water. These changes have affected the marine ecosystem. The relation between the climate changes and the observed shifts in the marine ecosystem are, however, based mainly on the use of ocean temperatures as a proxy for the climate change. Scientific investigations to understand the ecological and physical processes behind changes in the marine ecosystem need to be done.

In this study the drift of shrimp larvae, plankton and cod larvae is investigated. The drift simulations of shrimp larvae turned out to be too simplified to explain differences in year class strength between 1999 and 2000. Currents from a barotropic model were used, but it is argued, that a baroclinic model is needed in order to address the differences correctly. The barotropic model reveals relative strong and permanent clockwise circulations around the shelf banks north of 64°N which are very important fishing grounds at West Greenland. It has potential effect on the ecosystem caused by long retention times in the shelf banks.

It is hypothesised, that strong year classes of cod at West Greenland are a result of recruitment from Iceland. This drift of cod larvae and eggs from Iceland to Greenland was investigated using model based particle tracking, but the hypothesis could not be confirmed. Verification of the model has shown that the model did not display the right distribution of salinity, temperature and sea-ice in the Western Atlantic, which again gives an incorrect circulation. Thereby no conclusive results could be obtained. However, results from the particle tracking revealed, that the initial position of the particles was extremely important for the final position. Almost no particles initiated at the known spawning locations drift to Greenland waters. It was therefore suggested, that the position of the spawning locations have changed over time. This hypothesis needs further investigations using observational data.

Surface drifting buoys has been used to describe the drift from Iceland towards Greenland. It was found, that the drift from Icelandic to Greenland waters takes place in a very limited area within the Denmark Strait in the frontal zone between Irminger Water and Polar Water. Such a cross frontal drift is highly sensitive to the wind conditions, which is dominated by low-pressure systems frequently passing by. It was argued that in order to simulate this cross frontal transport correct, higher spatial and temporal resolution is needed in both the ocean and the atmospheric model.

4.1 Future development

The individual strengths of the East Greenland and Irminger Currents strongly affect the hydrographic environment of the shelf areas in Southwest Greenland which again affects the marine productivity in the shelf areas. The link between ecosystem variability and changes in the hydrography is mainly based on ocean temperatures. Almost no direct current measurements have been performed that could determine the strength and variability in the transport of Irminger Water and Polar Water to the area. At Cape Farewell, which makes the entrance to West Greenland, these two watermasses are well defined and separated.

- It is therefore recommended to get direct current, temperature and salinity observations at Cape Farewell in order to quantity the transport of Polar Water and Irminger Water to the West Greenland shelf areas.

For understanding the variability in the transport of cod larvae from Iceland to Greenland and its connection to changes in the hydrographic conditions, long timeseries are absolutely necessary. Additional, they are needed for validation of ocean models and particle tracking models.

- It is therefore recommended to continue the hydrographic standard section at West Greenland, the Icelandic 0-group survey and the German autumn survey at Southeast Greenland in order to get a better understanding of variations in the larval transport of cod from Iceland to Greenland and its coupling to hydrography.

Particle tracking, in combination with high resolution ocean modelling of the West Greenland shelf showed, that small scale circulation around shelf banks affect the retention time considerable. However, the barotropic setup was unable to explain the difference in the year class strength of shrimp between 1999 and 2000 which is believed to be caused by different hydrographic conditions and therefore requires a baroclinic simulation. The transport of cod larvae from Iceland to Greenland was estimated using a baroclinic ocean model with fairly coarse resolution of about 20 km. However, drifter trajectories of surface drifters shows that the drift from Icelandic to Greenland waters is very localized at about 66°N within the Denmark Strait at the front between the Polar Water and the Irminger Water. Individual low-pressure systems are suggested to be important for this cross-frontal drift. Hence, the spatial and temporal scales used in the ocean model, and in the atmospheric model, were much to coarse for reproducing the correct circulation in the Denmark Strait.

- It is therefore recommended to increase the spatial and temporal resolution in both the ocean model and the atmospheric model in order to describe the cross-frontal transport in the Denmark Strait correctly.

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Co-author statements

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The manuscript is based on work done in connection with Mads Hvid Ribergaard's Ph.D. project. The work has grown out of discussions with Pedersen and Simonsen. All Figures and Tables were preformed by Pedersen except Figure 6–8 by Ribergaard. Ribergaard made the numerical modelling part. The manuscript text was initiated by Pedersen and was finalized in collaboration.

Søren A. Pedersen

Søren Anker Pedersen

Claus Stenberg Simonsen

Mads Hvid Ribergaard

Manuscript: Ribergaard, M.H., Pedersen, S.A., Aadlandsvik, B., and Kliem, N., 2004. Modelling the ocean currents on the West Greenland shelf with special emphasis on northern shrimp recruitment. *Continental Shelf Research,* **24** (13–14), 1505–1519. (*Included in Appendix*)

The manuscript is based on work done in connection with Mads Hvid Ribergaard's Ph.D. project. The work has grown out of discussions with Pedersen, Aadlandsvik and Kliem. All Figures and Tables were preformed by Ribergaard, except Figure 2 by Pedersen (hatching curves inlay by Ribergaard). All the numerical modelling and the data handling were done by Ribergaard. The manuscript text was initiated by Ribergaard, later the biological part by Pedersen and it was finalized in collaboration.

Søren Anker Pedersen

Bjørn Aadlandsvik

Nicolai Kliem

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Manuscript: Ribergaard, M.H. and Sandø, A.B., 2004. Modelling the transport of cod eggs and larvae from Iceland to Greenland waters for the period 1948–2001. *In preparation*. (*Included in Appendix*)

The manuscript is based on work done in connection with Mads Hvid Ribergaard's Ph.D. project. The ocean model was setup and run by Sandø. All drift simulation was done by Ribergaard including development of the drift model. All text and figures are made by Ribergaard, except the model description by Sandø.

Anne Britt Sandø

Anne-Britt Sandø

Mads Hvid Ribergaard

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The manuscript is based on work done in connection with Mads Hvid Ribergaard's Ph.D. project. The ocean model was setup and run by Sandø. All validations work was done by Ribergaard. All text and figures are made by Ribergaard, except the model description by Sandø.

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