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North Atlantic circulation and variability, reviewed for the CNLS conference

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Abstract

The circulation and water mass structure of the North Atlantic are reviewed, with emphasis on the large-scale overturning cell which produces North Atlantic Deep Water (NADW). Properties and transports for its major components (Nordic Seas Overflow Water, Labrador Sea Water, Mediterranean Water, Antarctic Intermediate Water and Antarctic Bottom Water) are reviewed. The transport estimates and properties of NADW coupled with the observed meridional heat transport in the Atlantic limit the temperature of northward flow which replenishes the NADW to the range 11–15°C. The high salinity of the North Atlantic compared with other ocean basins is important for its production of intermediate and deep waters; about one third of its higher evaporation compared with the North Pacific is due to the Mediterranean. The evaporation/precipitation balance for the North Atlantic is similar to the Indian and South Atlantic Oceans; the difference between the North and South Atlantic may be that high evaporation in the North Atlantic affects much greater depths through Mediterranean Water production.

Also described briefly is variability of water properties in the upper layers of the subtropical/subpolar North Atlantic, as linked to the North Atlantic Oscillation. The oceanographic time series at Bermuda is then used to show decadal variations in the properties of the Subtropical Mode Water, a thick layer which lies in the upper 500 m. Salinity of this layer and at the sea surface increases during periods when the North Atlantic westerlies weaken between Iceland and the Azores and shift southwestward. (The North Atlantic Oscillation index is low during these periods). Temperature at the surface and in this layer are slightly negatively correlated with salinity, decreasing when salinity increases. It is hypothesized that the salinity increases result from incursion of saline water from the eastern subtropical gyre forced by the southward migration of the westerlies, and that the small temperature decreases are due to increased convection in the Sargasso Sea, also resulting from the southward shift of the westerlies.

Keywords: North Atlantic circulation; Water mass production; North Atlantic Oscillation; Decadal ocean variability; Bermuda oceanographic station

1. Introduction

The North Atlantic is well known for its central role in the deep overturning circulation for the globe. Although the Antarctic makes an equivalent contribution to the deep and bottom waters of the global ocean, the formation of deep water in the North Atlantic has received more attention, perhaps because of the asymmetry in heat transport between ocean basins: the Atlantic transports heat northward from the Southern Ocean to Iceland (Bryan, 1962; Bennett, 1978; Hall and Bryden, 1982), whereas the Pacific heat transport is mostly poleward, as is the zonally averaged oceanic contribution (Oort and Vonder Haar,

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1976; Trenberth, 1979; Talley, 1984; Hsiung, 1985; Trenberth and Solomon, 1994). For reasons of location, the North Atlantic also has been studied more intensely over a longer period of time than any other basin.

Most of this work was presented as descriptive background for the large group of modeling and theoretical papers at the conference on nonlinear stability held in Los Alamos, NM in May, 1995. Thus much of this contribution, with the exception of the final section on decadal variability of ocean properties near Bermuda, is in the nature of a classroom presentation rather than being fundamentally new work. Reviewed briefly are the basic circulation of the North Atlantic, its role in the global overturning circulation, and observations of decadal and longer variability in the North Atlantic circulation and water properties along with their apparent connection to atmospheric forcing. The review draws heavily on the overview papers of Reid (1994), Schmitz and McCartney (1993), Schmitz (1995) and Dickson et al. (1996), to which the reader should look for more in-depth discussion.

Some new material is presented in the last section, showing strong decadal variations in the uppermost layer (Subtropical Mode Water) at a location near Bermuda; this work extends the time series shown in Talley and Raymer (1982), and complements recent work using the Bermuda time series by Joyce and Robbins (1996) an Levitus et al. (1995), both concentrating on the deeper ocean, and by Reverdin et al. (1996), who concentrated on subpolar time series.

2. Mean circulation of the North Atlantic

The reader is referred to Reid (1994) and Schmitz and McCartney (1993) for their comprehensive and complementary views of the circulation at all depths. The important horizontal (isopycnal) elements of the large-scale circulation are geostrophic except for the directly wind-driven Ekman layer at the surface and flow on the equator. The basic structure of the largest-scale circulation varies only slightly over long time scales, and is forced by the mean wind patterns and by buoyancy forcing. The winds cause Ekman transports whose convergences/divergences drive the interior Sverdrup transport and the near-equatorial circulation. The dominant buoyancy forcing is surface heating/cooling and evaporation/precipitation; thermohaline processes within the ocean, such as double diffusion, may also have a large-scale effect. The relative importance of buoyancy (thermohaline) forcing compared with wind forcing in creating the largest scale, dominant circulation in the upper 1000 m appears to be minimal, suggesting that thermohaline forcing is strongly important only in local regions and where the flow is weak (below the wind-driven circulation).

The near-surface subtropical circulation consists of an anticyclonic subtropical gyre with a deep-reaching western boundary current (Gulf Stream and North Atlantic Current), and a locally important, shallow eastern boundary current associated with upwelling (Canary Current). There are interesting differences even in analyses of the upper level circulation: Stommel et al. (1978) show a closed subtropical gyre near the surface relative to a deeper level, extending most of the way across the basin; Schmitz and McCartney's (1993) transport cartoon shows the gyre extending all the way across to Africa. On the other hand, Worthington (1976) shows the volumentrically important part of the gyre transport to be confined west of the mid-Atlantic Ridge. Reid's (1994) adjusted geostrophic surface gyre is closed only west of the mid-Atlantic Ridge, wih the eastern boundary flow leading into the tropics rather than connecting to the western boundary. Given the large amount of data available for surface circulation analyses, these varying interpretations suggest caution in adopting all details of any particular scheme for circulation at all depths.

The western part of the subtropical gyre circulation consists of two centers – one associated with the Gulf Stream and the other with the North Atlantic Current (NAC) east of Newfoundland (Worthington, 1976). They are connected through branching of the Gulf Stream extension (Clarke et al., 1980). The existence of both centers is dynamically consistent with the North Atlantic's mean wind pattern, in which Sverdrup transport streamlines which remain within the subtropical gyre can be found as far north as NAC western boundary current (Mayer and Weisberg, 1993). However due to the southwest–northeast tilt of the zero of Ekman pumping, the NAC western boundary current lies adjacent to an Ekman upwelling region. This marked Ekman upwelling region *within* the northwestern subtropical gyre, as also occurs in the North Pacific, probably affects the modification and subduction of subtropical surface waters.

The separated NAC extends eastward across the North Atlantic and bifurcates near the eastern boundary into southward flow which forms the northernmost part of the subtropical circulation and a northward flow which becomes the cyclonic subpolar circulation. The southward-turning branch carries a thick surface layer of 11–12°C water with it, which is "subducted" in the eastern part of the subtropical gyre (McCartney, 1992). Subduction is the process whereby the outcropping layers of the subtropical gyre slide or are capped beneath less dense waters as the circulation is driven southward by the Ekman convergence. Iselin (1939) first pointed out the relationship between the vertical profiles of properties through the subtropical thermocline and the winter surface properties to the north; Luyten et al.'s (1983) theory of wind-driven circulation with outcropping layers formalized the concept, which explains very well tracer and water mass properties in the eastern and southern parts of subtropical gyres (e.g. Sarmiento et al., 1982; Jenkins, 1987). Subduction in the eastern North Atlantic might well involve processes at the Azores Front (Joyce and Jenkins, 1993), a sharp water mass boundary which might be considered one of the eastward branches of the Gulf Stream extension.

The subpolar gyre has the East Greenland and Labrador Currents as its western boundary currents. Since the NAC loop extends fairly far northwards across the mouth of the Labrador Sea, the interaction between the Labrador Current and the NAC is of interest. Some portion of the deep-reaching Labrador Current flows southward under the NAC and then westward into the subtropical region; this is the deep western boundary current (DWBC) which carries the dense intermediate and bottom waters of the subpolar gyre to lower latitudes (thermohaline circulation). The subtropical and subpolar circulations are connected through branching of the NAC; the extent of the connection volumetrically, its variability, and the processes which control it are the subject of current observational programs and research. This region appears to be important in paleoceanographic studies, as the ice edge during glaciations appears to have moved as far south as the NAC and dense water production was confined to lower densities than in the current warm state. The NAC region, especially near the western boundary, appears to be important for understanding variability in the North Atlantic – anomalous SSTs propagate eastward from this region, which is a center of large amplitude of one of the dominant empirical orthogonal functions describing decadal variability (Kushnir, 1994; Deser and Blackmon, 1993).

The cyclonic gyre is complicated by the Greenland–Iceland–Faroe sills; if they formed a solid boundary, the flow would simply be cyclonic past the southern end of Iceland, but instead part of the transport enters the Norwegian Sea in the east. The warm water which enters the Norwegian Sea is primarily Subpolar Mode Water (McCartney and Talley, 1982), which by its great thickness (>400 m) must have some convective forcing in winter since the most vigorous wind stirring cannot mix a layer this deep. (This is notable because the buoyancy loss in the subpolar region is actually weak compared with the large buoyancy losses near the Gulf Stream and in the Norwegian Sea (Schmitt et al., 1986).) The inflowing waters are transformed in the Nordic Seas (also known as the Greenland–Iceland–Norwegian Seas) and farther north in the Arctic (Swift and Koltermann, 1988; Mauritzen, 1993). The transformed waters are joined by a small amount (order $1 \times 10^6\ m^3\ s^{-1}$) of Pacific water which enters the Arctic through Bering Strait and a small amount of runoff in the Arctic. They return in the shallow, fresh East Greenland Current and the dense sill overflows through the Denmark Strait and gaps in the Iceland–Faroe Ridge and Faroe–Shetland Channel (e.g. Lee and Ellett, 1965; Worthington and Wright, 1970); the sill depths permit only the intermediate and shallower waters of the Nordic Seas to be exported southwards.

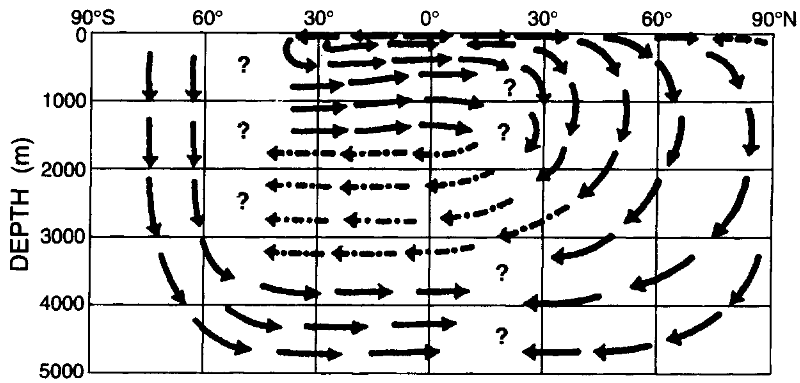

Fig. 1. Cartoon of the overturning circulation reproduced from Merz and Wüst (1992), based on concepts due to Buchanan (1885).

3. Water mass structures and the overturning circulation

The North Atlantic is complex in terms of water mass structures because of the number of locally produced waters with widely differing characteristics. The basic layering was mapped and understood before the time of Wüst (1935). Buchanan (1877, 1885) prepared the first vertical sections of salinity, including a meridian in the Atlantic, showing the northward-extending Antarctic Intermediate Water layer splitting the higher salinity North Atlantic Deep Water and evaporative subtropical surface water. Merz and Wüst (1922) enunciated the basic layer conception of the Atlantic Ocean. They presented several cartoons of the vertical overturning circulation; their favorite (reproduced here as Fig. 1) followed Buchanan's (1885) ideas, and looks remarkably like what we might sketch today. The biggest unknown in their favorite overturning cell is the location of upwelling; because of the salinity layering revealed by Buchanan, they discarded Lenz's (1847) and Schott's (1902) concept of dominantly equatorial upwelling from the deep to the upper ocean. Ideas swung over to uniform ocean-wide upwelling as the simplest model for the overturning circulation (Stommel and Arons, 1960; Munk, 1966), but attention has returned to determining the spatial distribution of upwelling.

A much more complete picture of the water mass sources for the North Atlantic dates from the 1970s and is described by Reid (1994) and Schmitz and McCartney (1993), among others. The basic structures are apparent from any long meridional section of salinity, oxygen or nutrients; a representative salinity section is shown in Fig. 3 (location in Fig. 2). Principal inflow from the South Atlantic occurs in two layers: the Circumpolar Deep Water (or Antarctic Bottom Water) at the ocean bottom, and from the sea surface to about 1500 m encompassing the thermocline and Antarctic Intermediate Wate layers. Outflow to the South Atlantic occurs principally in the so-called North Atlantic Deep Water (NADW) between about 1500 and 4000 m.

Modification of the inflowing southern waters occurs along their path through the North Atlantic, and includes air–sea interaction and mixing with the outflowing components of NADW. One clearly identifiable part of this process is the passage of surface waters around the North Atlantic Current system and then cyclonically around the subpolar gyre; the thick winter surface layers involved in this part of the modification are referred to as Subpolar Mode Waters (SPMW) (McCartney and Talley, 1982). The outflowing components are produced through major transformation in three locations: the Mediterranean, the Labrador, and the Arctic/Nordic Seas (Wright and Worthington, 1970; Worthington and Wright, 1970).

Each of these sources provides a identifiable characteristic for NADW; its properties would be altered by deletion or major change in any one. Only in the subtropical South Atlantic and farther south is NADW

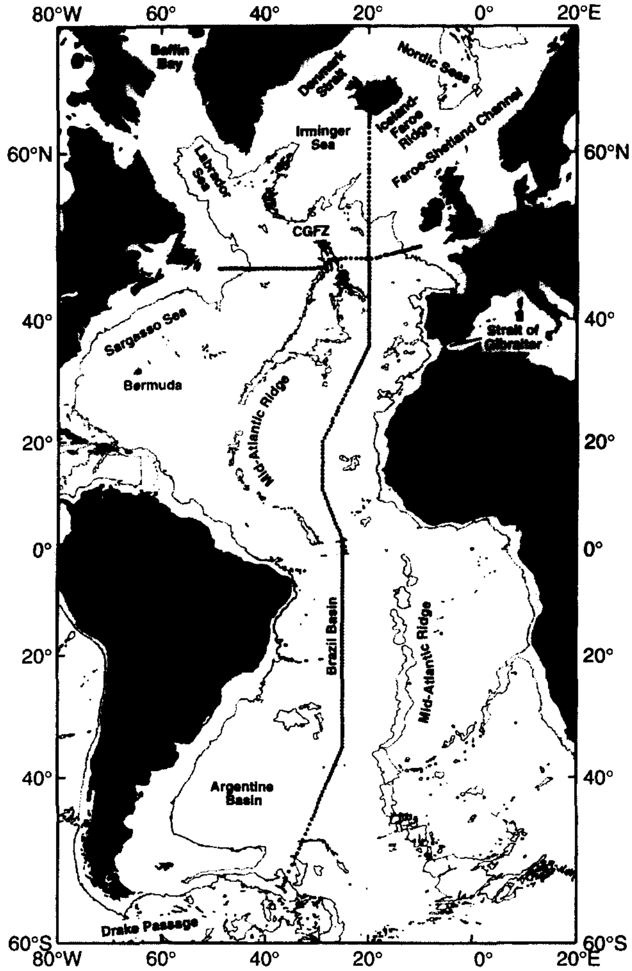

Fig. 2. Geographical locations mentioned in the text, with the 3000 m isobath. The section used in Fig. 3 is the long meridional section at 20–25°W. The section used in Fig. 4 is the zonal section at 48°N. The oceanographic time series station near Bermuda is shown with a +.

apparent as a single undifferentiated water mass. In the tropical South Atlantic there remain clear vestiges of NADWs varied source waters: Wüst's (1935) Upper, Middle and Lower NADW result from Mediterranean, Labrador Sea and Nordic Seas Overflow Waters, respectively. Mediterranean Water (MW) is high in salinity but low in oxygen. Labrador Sea Water (LSW) is relatively fresh and high in oxygen due to local renewal. The Nordic Seas overflows (NSOW) are dense, cold, and high in oxygen due to intermediate water renewal in the Greenland Sea (Swift et al., 1980) and entrainment of LSW as they overflow.

Circulations and transformation rates for each of the sources, in terms of properties in the layers or on the isopycnals which best characterize them, can be found in Reid (1994), Schmitz and McCartney (1993), and Schmitz (1995). In the northwestern North Atlantic, where the intermediate and abyssal depths are dominated by LSW and NSOW, the ventilation time scale is on the order 15–20 years and variations in the deepest waters' properties are easily observed (Swift, 1984), as reviewed in Section 4.

The easily described meridional circulation implicit in the preceding sections and in Fig. 1 belie the complicated and lengthy horizontal circulation taken

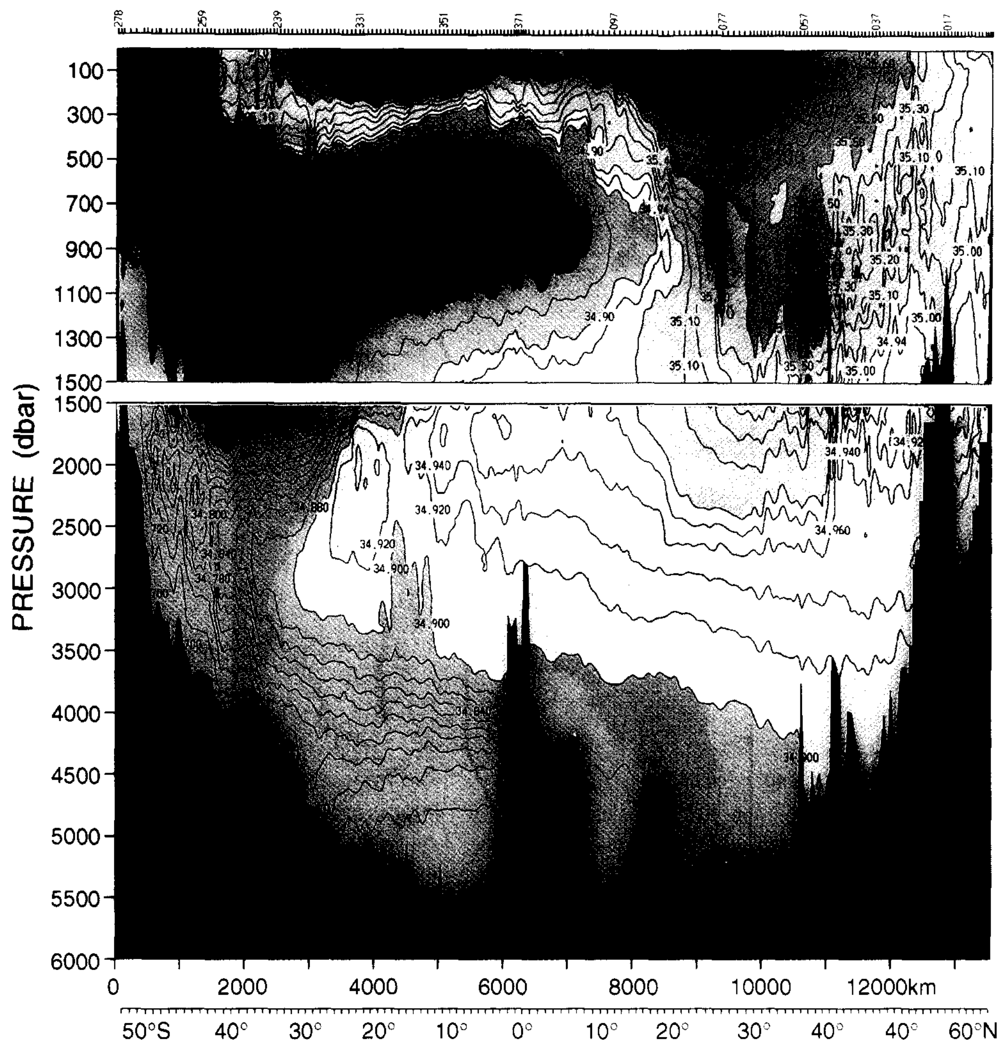

Fig. 3. Meridional section of salinity along the section shown in Fig. 2 (Tsuchiya et al., 1992; Tsuchiya et al., 1994). This section is similar to any other meridional section of salinity through the Atlantic in terms of the large scale features: near-surface – high salinity surface waters in the subtropics, low salinity surface waters at higher latitudes; 500–1500 dbar – northward extending low-salinity Antarctic Intermediate Water, high salinity Mediterranean Water centered at 30–35°N and "Meddies" around 25°N; 1500–2500 dbar – low salinity Labrador Sea Water in the north; 1550–4000 dbar – high salinity NADW extending southward in the South Atlantic; 4000 dbar-bottom – lower salinity Antarctic Bottom Water (Lower Circumpolar Deep Water).

by water parcels as they move from far south to far north and back again. The complexity of the lateral circulation at each depth is apparent in Reid's (1994) smoothed circulation analyses. A large dynamical barrier to easy north–south communication occurs in the tropics, where the large-scale potential vorticity (angular momentum) field is primarily dominated by the meridional variation of the Coriolis parameter, i.e. the β-effect (e.g. Keffer, 1984; Talley, 1996). In the tropics, broadly defined as 15–20° on each side of the equator, the potential vorticity distribution at intermediate and deep levels is essentially zonal. Northward or southward motion by water parcels thus becomes an exercise in changing potential vorticity, which requires a source such as friction or deep stress. From maps of deep and intermediate properties on isopycnals (Wüst, 1935; Reid, 1994; Suga and Talley, 1995) it is clear that most meridional transport in the tropics occurs in the western boundary layer, and that otherwise the tropical circulation is essentially zonal.

The following sections provide more specific description of the source waters for NADW, followed by a review of water mass production rates.

3.1. South Atlantic source waters for NADW

The waters which flow into the North Atlantic and whose mass transport is balanced by the outflow of NADW are the intermediate/surface waters and the bottom water. Some controversy has arisen about the source of the intermediate/surface waters themselves – whether they are (1) largely warm waters which enter the South Atlantic from the Indian Ocean south of Africa (Gordon, 1986) or (2) composed more of colder surface waters which enter the South Atlantic from the Pacific through Drake Passage (e.g. Rintoul, 1991). The former view is encapsulated in an NADW "conveyor belt" cartoon (Broecker, 1991) which is popularly used to convey the idea of a global overturning circulation. Schmitz's more global view (1995) on the other hand includes Drake Passage transport, which Broecker (1991) acknowledges is also a factor.

Gordon's (1986) argument for the warm water route is based on the total heat transport and the volume transport of NADW and warm waters across 30°S. The crux of his argument is the large northward heat transport, on the order of $0.5–1.0 \times 10^{15}$ W by almost all calculations, for a large part of the North and South Atlantic. With an NADW production rate and southward transport on the order of 15–20 Sv (1 Sverdrup $=10^6\,\mathrm{m^3\,s^{-1}}$), as described in Section 3.5, the northward flow must occur at a temperature of 11–15°C at almost all latitudes in the Atlantic. It is not possible to reduce the temperature of the northward flow below this range unless the northward heat transport is badly overestimated or the NADW production rate severely underestimated. Details of the gyre circulations at any latitude create the 4°C range in temperature of the northward flow. For instance, Gordon's inclusion of the southward-flowing warm Brazil Current does increase the required temperature of the northward flow from about 13–15°C, but the total range of northward flow temperatures is actually not very sensitive to the assumed Brazil Current flow.

Gordon (1986) extends his argument about the temperature of the northward flow upstream (southward) in the South Atlantic by hypothesizing that the warm return water must come from the Agulhas Current in the Indian Ocean rather than from the colder waters of Drake Passage. Rintoul (1991) argues on the other hand that Drake Passage water can be warmed in the southwestern Atlantic after its emergence from Drake Passage, which would permit it to be a significant source of the northward warm water flow.

The bottom of the South Atlantic upper layer which flows into the North Atlantic is the Antarctic Intermediate Water (AAIW). AAIW originates on the northern side of the Antarctic Circumpolar Current; its specific origin for the South Atlantic is the Falkland/Brazil Current confluence in the southwest Atlantic, with a component of older recirculated AAIW coming also from the Agulhas Current (Talley, 1996). The Upper Circumpolar Water, which lies below the AAIW in the subtropical South Atlantic, also originates in the Antarctic circumpolar region, but it is truncated at the subtropical/tropical boundary in the South Atlantic and does not reach the North Atlantic (Tsuchiya et al., 1994). The AAIWs principal signature is a salinity minimum, identifiable to about 20–25°N, north of which it is obliterated by the high salinity from the MW (Wüst, 1935; Fig. 3). Traces of AAIW, identifiable as a silica maximum, are apparent well up into the North Atlantic's subpolar gyre (Fig. 7(f) in Reid, 1944) despite large dilution with North Atlantic waters (Tsuchiya, 1989).

The other northward flow into the North Atlantic is of Antarctic Bottom Water (AABW), lying below the NADW. It reaches the equator from the south in a deep western boundary current. At the equator, it attempts to flow eastward but is restricted by the Mid-Atlantic Ridge; it "crosses over" from the western boundary and flows northward along the western flank of the ridge into the western North Atlantic. AABW is apparent west of the Mid-Atlantic Ridge up to about 40°N. Upwelling out of the AABW layer occurs during its northward transit, as is clear from the bottom temperature distribution (Mantyla and Reid, 1983). The northward intrusion of AABW and loss of cold water is also graphically depicted in the potential temperature surface maps of Worthington and Wright (1970). Hogg et al. (1982) use estimates of AABW transport through the connecting channel between the Argentina and Brazil Basins and the loss of all cold water in the Brazil Basin to compute upwelling and diapycnal mixing rates. Luyten et al. (1993) also use the loss of AABW and also of Denmark Strait Overflow Water upward into the NADW layer to estimate diapycnal mixing rates. Transport estimates for the northward flow of AABW across 4°N and into the North Atlantic are 3–5 Sv (McCartney and Curry, 1993; Luyten et al., 1993). Schmitz and McCartney (1993) summarize transport and upwelling of the AABW with its eventual complete loss at around 40°N. An important aspect of their transport cartoons are the recirculations of the bottom and deep waters, as described in McCartney (1993); these increase the apparent residence times and permit dilution of characteristics.

3.2. Mediterranean Water

The Mediterranean Water (MW) originates as the very saline outflow through the Strait of Gibraltar. Its signature is apparent over a large portion of the North Atlantic's subtropical gyre, both horizontally and vertically (e.g. Wüst, 1935; Reid, 1994). The flow through the strait is two-layered with mean inflow at the surface and outflow below. Both layers are strongly affected by the tides which form an internal bore (Armi and Farmer, 1988).

The MW properties at Gibraltar are around 13°C, 38.45 psu, and 29.07σ_θ (Ochoa and Bray, 1991; Price and Baringer, 1994). Armi and Farmer (1988) show even more extreme properties with salinity near 39 psu and σ_θ near 29.4 at the sill. Strong entrainment upon outflow rapidly lowers the salinity and density. The density of MW at its sill source is actually greather than that of the NSOW which flows over the sill in Denmark Strait, between Greenland and Iceland. This "Denmark Strait Overflow Water" (DSOW) has properties 0°C, 34.9 psu, 28.03σ_θ, and hence has lower density than MW at is sill source. Yet the DSOW reaches the bottom of the North Atlantic and becomes a deep water while MW remains an intermediate water. Price and Baringer (1994) model both overflows and attribute the lower end density of MW to the greater contrast in density between the MW and the surrounding water, leading to more rapid entrainment compared with the DSOW; thus the DSOW can plunge to a much greater depth. The difference in compressibility of the two source waters might also be a factor: although MW has greater potential density relative to the sea surface, its potential density relative to 4000 dbar is 45.99σ_4, compared with 46.23σ_4 for the DSOW source.

MW flowing out of the Strait of Gibraltar first turns northward and then joins the anticyclonic circulation of the subtropical gyre (Reid, 1994). In the north, MW mixes predominantly with LSW. Meddies (eddies of concentrated MW) spin off toward the south from the Strait of Gibraltar and are often found fairly far to the southwest (Armi and Stommel, 1983; Arhan and King, 1995; typical example at 25°N in Fig. 3) from whence they can be advected westward across the Atlantic. The high salinity of the MW gives the NADW its characteristic salinity maximum in the tropical Atlantic, where this core is known as Upper NADW (Wüst, 1935). The Upper NADW disappears south of about 20°S in the South Atlantic, apparently truncated by the southern sources of water in this density range – the AAIW and Upper Circumpolar Water (Tsuchiya et al., 1994).

NADW influence is evident as a salinity maximum in the southern South Atlantic and Indian Oceans and in the Pacific Ocean not because of the MW per se, but because all water of North Atlantic origin is relatively saline compared with waters of northern Pacific or Antarctic origin (Reid and Lynn, 1971; Reid, 1994). We can roughly estimate the importance of the Mediterranean to the overall higher salinity of the North Atlantic compared with other oceans using Mediterranean exchange values (Ochoa and Bray, 1991) and freshwater transports (Wijffels et al., 1992). In the Mediterranean Sea, approximately 0.05 Sv is lost through evaporation based on Ochoa and Bray's

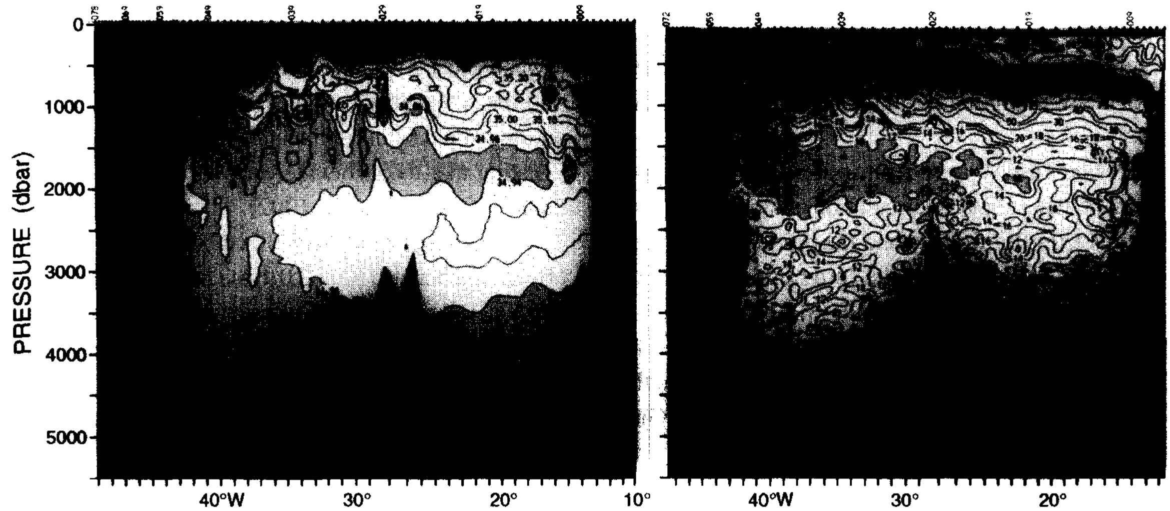

Fig. 4. (a) Salinity (psu) along a vertical section at 48°N (positions shown in Fig. 2). The low salinity layer extending across between 1500 and 2000 dbar is the Labrador Sea Water. The high salinity intrusions in the eastern portion are remnants of Mediterranean Water. (b) Isopycnic potential vorticity $(f/\rho)\partial\rho/\partial z(\times 10^{-14}\,\mathrm{cm}^{-1}\,\mathrm{s}^{-1})$ at approximately 48°N. The low potential vorticity layer centered between 1500 and 2000 dbar is the Labrador Sea Water. The low potential vorticity layer near the surface is the Subpolar Mode Water. The very low potential vorticity water in the bottom of the eastern basin is the Northeast Atlantic Deep Water which is primarily of southern origin (Tsuchiya et al., 1992).

transport and salinities of the in and outflowing waters at the Strait of Gibraltar. This compares with a net evaporative loss of about 0.6 Sv in the subtropical North Atlantic (Wijffels et al). Thus within the North Atlantic itself, the Mediterranean contributes only about 10% of the net freshwater loss. Also from Wijffels et al., the net subtropical freshwater loss in the North Pacific is about 0.45 Sv. Thus when the two northern hemisphere oceans are compared, the Mediterranean accounts for about one third of the difference in evaporative loss.

Wijffels et al. (1992) also show that the net evaporative losses in the subtropical South Atlantic, Indian, and South Pacific are about 0.6, 0.7 and 0.3 Sv, respectively. The total evaporation/precipitation losses for the Atlantic, Indian and Pacific are 0.9, 0.6 and −0.5 Sv, respectively. Consider the South Atlantic, where evaporative losses may be equivalent to the North Atlantic's – the principal differences between the two basins are that the evaporative loss in the South Atlantic is applied only to the surface layer, which is moving northward on average; surface water is converted to deep water only in the North Atlantic. In the North Atlantic, a portion of the evaporative loss is applied directly to the deep water through the Mediterranean's production of dense, very saline water. Ultimately all of the evaporative loss in the Atlantic is communicated to depth through the other parts of NADW formation. Even though some portions of the incipient NADW, e.g. LSW and NSOW, are fresher than surface or MW, they are all quite saline compared with the South Atlantic's Circumpolar Water; this is directly due to the net evaporation at the Atlantic's sea surface (Gordon and Piola, 1983), which includes the relatively small contribution from the Mediterranean.

3.3. Labrador Sea Water and subpolar Mode Water

LSW is formed by convection in the Labrador Sea and is not governed by a sill control. It is recognized at 1500–2000 m depth by its low salinity, high oxygen and low potential vorticity (e.g. Figs. 3 and 4). It has been demonstrated that its formation varies decadally (Lazier, 1980; Talley and McCartney, 1982; Lazier, 1995). Its properties vary with each passing winter, depending on the surface forcing over the Labrador

Sea, the properties of the entering SPMW (which integrates surface forcing changes over several years in the subpolar region), and the properties of the reservoir of LSW recirculating in the region from previous years.

LSW and NSOW arise from waters from the upper portion of the subpolar circulation plus fresher contributions from Baffin Bay and the Arctic. McCartney and Talley (1982) show the progression of thick surface layers which make up the SPMW, a principal source of the LSW and the warm water inflow to the Norwegian Sea. Also contributing to new LSW are the last vestiges of AAIW which flows north in the Gulf Stream system (Tsuchiya, 1989), and low salinity surface waters from Baffin bay (Lazier and Wright, 1993) and the East Greenland Current (Worthington, 1970). Previously formed LSW recirculates in the Labrador Sea and subpolar region and preconditions the formation of LSW for the following years (Lazier, 1995).

LSW flowing out of the Labrador Sea can be detected through its low salinity, high oxygen and low potential vorticity. In Fig. 4(b) it is clearly seen as the low potential vorticity layer centered at 1700 dbar. It fills the subpolar North Atlantic. Its traces are easily discernible far down along the western boundary (Talley and McCartney, 1982), whence it flows as the middle layer of the deep western boundary current. The upper part of the deep western boundary current is made up of subarctic intermediate water (e.g. Worthington, 1976) and does not cross the Gulf Stream with the LSW and NSOW (Pickart and Smethie, 1993). Even in the central and eastern subtropical North Atlantic, the LSW signature is apparent as an oxygen maximum beneath the MW; here it is known as Middle NADW rather than LSW, due to its gradual admixture with MW (see Tsuchiya et al., 1992). In the tropical South Atlantic, the high oxygen core of the Middle NADW, originating as LSW, is quite apparent, as stated above.

In the eastern North Atlantic the LSW mixes with MW, since the two water masses are adjacent to each other horizontally and vertically. On the very largest scale, MW density decreases away from its source and LSW increases in density away from its source, suggesting a very large scale influence of double diffusive mixing (warm, salty MW overlying cold, fresh LSW) (Talley and McCartney, 1982). A mixing region between the LSW and MW appears to occur along the northern boundary of the saline MW tongue, where strong zonal fronts delineate a region of intrusions suggesting that the two water masses are mixing actively there (Tsuchiya et al., 1992).

The properties of LSW have been shown to change by as much as 1.0°C and 0.1 psu over the years (Lazier, 1980; Talley and McCartney, 1982; Dickson et al., 1996) with some extraordinary recent examples showing a decrease in temperature well below the norm for this century (Lazier, 1995). LSW formation is apparently strongly controlled by surface salinity in the Labrador Sea, with convection nearly ceasing when the region is capped by low salinity (Lazier, 1980, 1995; Talley and McCartney, 1982; Dickson et al., 1988). In recent years LSW formation has been so vigorous that it is actually punching into the saline NSOW layer below and is now increasing in salinity as well as becoming colder (Lazier, 1995; Dickson et al., 1996). Cunningham and Haine (1995) demonstrate that LSW properties vary significantly in the eastern North Atlantic, and attribute the property variations to changes in the LSW source.

3.4. Nordic Seas Overflow Water

The Nordic Seas Overflow Water (NSOW) is composed of overflow water from three locations: the Denmark Straits west of Iceland, the Iceland–Faroe Ridge east of Iceland, and the Faroe–Shetland Channel farthest to the east. The latter two overflows are often considered together as the Iceland–Scotland Overflow Water. The overflow east of Iceland is much saltier than the overflow west of Iceland, probably due to entrainment of more saline surface and intermediate waters. The overflows have been carefully documented in recent years based on extensive current meter work in the northern North Atlantic (Dickson and Brown, 1994; Schmitz and McCartney, 1993). Monitoring of these overflows is ongoing.

The overflow waters probably originate as intermediate waters north of the Greenland–Scotland ridges (Swift et al., 1980), although an alternative hypothesis

Table 1
Events associated with recent cycles of the NAO

	1963–1969	1972–1976	1977–1979	1981–1984	1985–1987	Reference
NAO	Low	High	Low	High	Low	Hurrell (1995), Dickson et al. (1996)
Westerlies	Weak	Strong	Weak			Cayan (1992), Kushnir (1994)
SST subtropics	Low	High	Low	High		Dickson et al. (1996), this paper
SSS subtropics	High	Low	High	Low	High	(this paper)
SST Newfoundland	High	Low	High			Dickson et al. (1996)
SSS Newfoundland	High	Low	High			Reverdin et al. (1996)
Davis Strait ice volume (−2 years)	Low	High	Low			Deser and Blackmon (1993)
STMW convection (+1 to 3 years)	Strong	Weak	Strong	Weak	Strong	Dickson et al. (1996). this paper
LSW convection	Weak	Strong	Weak			Lazier (1980)
Greenland Sea convection	Strong	Weak	Strong			Dickson et al. (1996)

of greater influence from higher latitudes in the Arctic has recently been offered by Mauritzen (1993). The intermediate waters are produced by relatively deep convection at well-defined locations in the Greenland Sea (Morawitz, 1995; Clarke et al. 1990). The source waters of the Greenland Sea convection are ultimately the warm, saline SPMWs which flow into the Nordic Seas, predominantly near the eastern boundary. Deep convection in the Greenland Sea is time-dependent (Schlosser et al., 1991; Meincke et al., 1992), and is most vigorous when the North Atlantic Oscillation (winter pressure difference between Iceland and the Azores) is low (Dickson et al., 1996), cf. Table 1. The properties of the convected water also vary markedly with time.

Dickson and Brown (1994) had the first adequate date to begin to study the variability of the overflow in Denmark Strait. They observed no seasonal variation and suggest that the evidence is that interannual variability may be weak. This suggests that the overflow rate is controlled by hydraulics at the sills rather than by the production rate of intermediate and deep waters upstream in the Greenland Sea, although production changes likely affect the overflow water properties. Swift (1984) found an 0.15°C decrease and 0.02 psu decrease in deep water properties between 1972 and 1981, which he attributed to variations in surface properties in the Greenland Sea and northern North Atlantic.

3.5. NADW production rates

Water mass production rates are usually quoted for each component of NADW in the northern North Atlantic and then for the transport of NADW as a whole as it moves southward into the subtropical gyre, across the equator, and then into the South Atlantic and beyond. Production rates for the input components are based on the amount of water which enters a given temperature/salinity class each winter, but are usually calculated as the transport of those properties slightly downstream and moving away from the source.

The most recent calculations of production rates for each of the NADW components are presented in Schmitz and McCartney's (1993) and Dickson and Brown's (1994) work. It is generally accepted that NADW transport into the South Atlantic is of order 15–25 Sv (e.g. Roemmich, 1983; Schmitz and McCartney, 1993). This transport is estimated by defining the appropriate NADW layer and calculating its southward transport using geostrophy, employing either a reference level or an inverse method.

Estimating single values for water mass production rates only makes sense if there is a clearly defined, localized process. The LSW production rate is usually assumed to be that of winter convection into the dominant temperature/salinity for LSW for that year. The sources of MW and NSOW are sill overflows for which it is usual to include the entrained water in the production estimates. It is expected that the overflow transport variability might be small even in the face of large seasonal, interannual and decadal variability in surface forcing since they are governed by hydraulic control of a large upstream reservoir rather than by the particular characteristics of the reservoir water. The temperature/salinity properties of the overflows however can and do vary to some extent.

The production rate of LSW has been estimated at 4–9 Sv (Clarke and Gascard, 1983; McCartney and Talley, 1984). Schmitz and McCartney's (1993) summary uses a production rate of 7 Sv, of which 4 Sv escape to the south under the North Atlantic Current and 3 Sv are entrained into the deeper water; they show a recirculation of about 14 Sv superimposed on this. The production rate is variable, including a complete shutdown in the 1960s (Lazier, 1980; Talley and McCartney, 1982).

The denser portion of the NADW (below the LSW) is formed from Nordic Seas overflows, entrainment of LSW and upwelling of AABW. Schmitz and McCartney (1993) show 6 Sv of NSOW, joined by 3 Sv of LSW and 3 Sv of AABW, for a total of 12 Sv escaping southward under the North Atlantic Current. Dickson and Brown (1994) used direct current measurements of the Denmark Strait Overflow, flow through the Charlie Gibbs Fracture Zone, and Saunders' (1990) Faroe Bank measurements to estimate a Nardic Seas input to the NADW of 5.6 Sv, plus an entrainment of 3.1 Sv of LSW. Price and Baringer (1994) modeled the plunging, entraining plumes in the Denmark Strait and Iceland–Shetland ridge to predict the density and transport of the entraining plumes along their paths. They assumed source water transports of 2.6 and 2.7 Sv at Denmark Strait and Iceland–Scotland, respectively, resulting in model transports for the completely developed plumes of 4.1 and 5.7 Sv, respectively. Their results thus match the observations quite well.

Ochoa and Bray (1991) used hydrographic data near the Strait of Gibraltar to estimate a MW outflow of 0.7 Sv. This rapidly increases to about 3 Sv downstream as the plume descends and entrains surrounding water (Zenk and Armi, 1990; Price and Baringer, 1994).

The rate at which southern source waters (surface/AAIW and AABW) contribute to NADW is more difficult to define. The contributions could be estimated as the northward flow of the waters into the subtropical North Atlantic since all of it vanishes to the north, but any other point to the south or north could be chosen.

Near the equator, a net southward flow of 17 Sv of NADW was shown by Schmitz and McCartney (1993), balanced by an inflow of 13 Sv of surface/intermediate waters above the NADW and 4 Sv of bottom water below the NADW. In the subtropical gyre, they show 3 Sv upwelling from AABW into NADW, divided into 1 Sv west of the Mid-Atlantic Ridge and 2 Sv east of the ridge. This picture of two layers in and one layer out of the North Atlantic may be overly simplistic since there is southern source Circumpolar Water at the same density as the NADW which affects properties in the tropical Atlantic (Reid, 1994). The Circumpolar Water originates in the Antarctic Circumpolar Current, and is a mixture of deep waters from all basins, including the North Atlantic, Indian, Pacific, and Antarctic. However the zonally integrated flow over a large range of depths is southward (Roemmich, 1983), and dyed by the high salinity/high oxygen of the North Atlantic source (Wüst, 1935; Reid, 1994). The southward decay of the strong salinity/oxygen signal is due to isopycnal mixing with the Circumpolar Water and diapycnal mixing with the intermediate and bottom waters; the relative importance of isopycnic vs. diapycnic mixing has not been quantified.

4. Variations in water properties in the subpolar and subtropical North Atlantic

Dickson et al. (1996) have recently documented long-term changes in water properties in the Nordic Seas, the northern North Atlantic, and the western subtropical gyre and shown how they relate to long-term changes in atmospheric forcing. Earlier, Dickson et al. (1986) had shown that a patch of anomalously low salinity had arisen from the Arctic in the late 1960s and propagated cyclonically around the subpolar North Atlantic until the late 1970s. The low salinity affected water mass properties and convection although it does not seem to have shut off convection in the Greenland Sea. When it was present in the Labrador Sea, convection was greatly diminished (Lazier, 1980).

In a more recent study, Dickson et al. (1996) show that the "great salinity anomaly" was a century event. They show that times of extreme surface flux and convection in the Greenland Sea occur simultaneously with extreme surface fluxes in the western subtropical gyre (subtropical mode water formation area – Sargasso Sea), and that these are out of phase with periods of extreme convection in the Labrador Sea. Their review of various time series suggests that strongest convection periods for the Greenland Sea/western subtropical gyre occurred in 1880s and 1960s. Meanwhile the Labrador Sea is now in a strong convection period, even as Greenland Sea deep convection appears to be curtailed.

The extreme events in the 1880s and 1960s were apparently a long-term modulation of quasi-decadal oscillations observed in water properties in the subpolar and western subtropical Atlantic, possibly associated with the North Atlantic Oscillation (Dickson et al. 1996). Numerous investigators are currently piecing together the variations in water properties from the surface to the bottom in the subtropical and subpolar Atlantic; Dickson et al. (1996) review many of the most recent papers on the Greenland, Labrador and Sargasso Seas. Earlier work by Swift (1984), Lazier (1980), Talley and McCartney (1982), and Roemmich and Wunsch (1984) documented significant changes in deep and intermediate water properties from 24°N up to the Greenland–Scotland ridge and into the Greenland Sea. Talley and Raymer (1982) showed how the upper ocean properties had varied in the western Sargasso Sea.

Recent examples of variability are Parilla et al.'s (1994) reoccupation of a subtropical zonal section showing continuing warming since Roemmich and Wunsch (1984). Levitus et al. (1995) showed strong decadal oscillations in temperature and salinity in the upper layers in the western subtropical gyre and the central subarctic gyre, with the decadal changes extending down to 1750 m at Bermuda superimposed on a warming trend there; although there was no decadal oscillation at intermediate depth in the subarctic gyre, there was a warming in the early 1970s and a cooling/freshening thereafter. Joyce and Robbins (1996) show a long-term warming trend at intermediate depths extending starting at least as early as the 1930s at Bermuda. Reverdin et al. (1996) show that the strong decadal oscillations near the sea surface extend over a large region of the subarctic and western subtropical Atlantic, and that they are correlated with the NAO.

5. Relation of water property variations to the North Atlantic Oscillation (Table 1)

The North Atlantic Oscillation (NAO) is traditionally defined as the winter (December–February) surface atmospheric pressure difference between Iceland and the Azores (van Loon and Rogers, 1978; Rogers, 1984). When the NAO is low, the pressure difference is small and the westerlies between Iceland and the Azores are weak, and vice versa. Empirical orthogonal function (EOF) analysis shows it to be the dominant mode of surface pressure variability (Barnston and Livezey, 1987; Cayan, 1992). Hurrell (1995) shows the NAO for the past 130 years, with a strong correlation with precipitation/temperature in Europe. The surface pressure difference between Iceland and the Azores shows a quasi-decadal variation in strength, with much longer periods also apparent – there were especially low values in the 1880s and 1960s (Dickson et al., 1996). A decadal oscillation is most apparent over the past 40 years; prior to that the record has less of a dominant frequency.

When the NAO is low, the strength of the westerlies between Iceland and the Azores is reduced and the subtropical storm tracks retract to the east coast of the US. Dickson et al. (1996) argue that the southwestward shift of the storm tracks increases convection in the STMW formation area of the Sargasso Sea and reduces convection in the Labrador Sea. Along with weakened westerlies between Iceland and the Azores, a weak NAO creates northerlies in the Greenland Sea which increase the convection and decrease the surface temperature there.

Deser and Blackmon (1993) show a strong relationship between the NAO and sea ice presence in the Labrador Current. They hypothesize that weak westerlies during low NAO conditions create warm SSTs off Newfoundland which further sustain the weak winds. When the winds are strong, cold SST appears and strengthens the winds. Events of cold SST might be initiated by periods of large sea ice buildup in the Labrador Current – they observed that large sea ice volume precedes cold SST off Newfoundland by about two years. (The melting sea ice would cap the region with low salinity water which could then become colder than usual.)

Kushnir (1994) on the other hand suggests that there is a feedback in which warm SST creates strong winds which cool the region east of Newfoundland, resulting in weak winds which warm the region east of Newfoundland. Both Cayan (1992) and Kushnir (1994) find that the atmosphere leads the ocean on seasonal and interannual time scales in the extratropics, but suggest that feedback between the ocean and atmosphere may be important on interdecadal time scales.

Cayan (1992) uses surface pressure EOFs, winds and fluxes to build a scenario for the NAO, which is his dominant mode. He shows that low NAO coincides with a southward shift of the strong westerly winds and increased latent heat loss in the western subtropical gyre (and decreased latent heat loss in the labrador/Irminger Seas). Hansen and Bezdek (1996) took a different view of the variations in SST, looking at the actual progression of temperature anomalies around the subpolar and subtropical gyres, rather than relying on EOF analysis which tends to lead to a viewpoint of a stationary pattern.

6. Decadal variations in Subtropical Mode Water properties at Bermuda

In the western subtropical North Atlantic, the near surface water column is dominated by a thick layer at approximately 18°C, referred to as Subtropical Mode Water (STMW) or 18°C Water. It is thickest and outcrops just south of the Gulf Stream, from whence it is advected eastward and southward to fill the western subtropical gyre (Worthington, 1976). Its presence is principally due to the warm bowl caused by the Gulf Stream's baroclinicity, but its vertical homogeneity and small changes in properties in time and space result from vertical convection in the high buoyancy loss area in and adjacent to the Gulf Stream. Talley and Raymer (1982) showed that along the southern side of the Gulf Stream, STMW is colder to the east, which was attributed to heat loss in this region. They also documented variations in the properties of STMW using the hydrographic time series begun in 1954 near 32°10′N, 64°30′W (just off Bermuda) with sampling every two weeks to a month. The time series available in 1980 showed periods of strong and weak STMW presence, based on thickness (isopycnic potential vorticity) of the water column. They also showed variations in density of the core of the STMW as defined by the minimum in isopycnic potential vorticity. The observed changes were not easily related to changes in local surface heat flux.

6.1. *Updated STMW property variations*

Using the Bermuda time series from 1954 to 1995, it is now possible to see that the STMW properties are related to the NAO, which has had three "oscillations" since 1954. The oceanographic data consist of temperature, salinity and oxygen at discrete depths measured every two to four weeks. No data were collected in 1979. Data prior to 1991 are available from the National Oceanographic Data Center and easily accessible online through our "hydrosearch" database at SIO; the most recent years were acquired from Dr. A. Michaels at the Bermuda Biological Station. Salinity quality improved starting in 1959. The upward shift in oxygen around 1959 could be attributed to a change in convective activity, but the size of the shift suggests that sample analysis methods may also have been a source of the change. Data were treated as in Talley and Raymer (1982). They were interpolated to isopycnals using an Akima cubic spline. Isopycnic potential vorticity $(f/\rho)\partial\rho/\partial z$, where f is

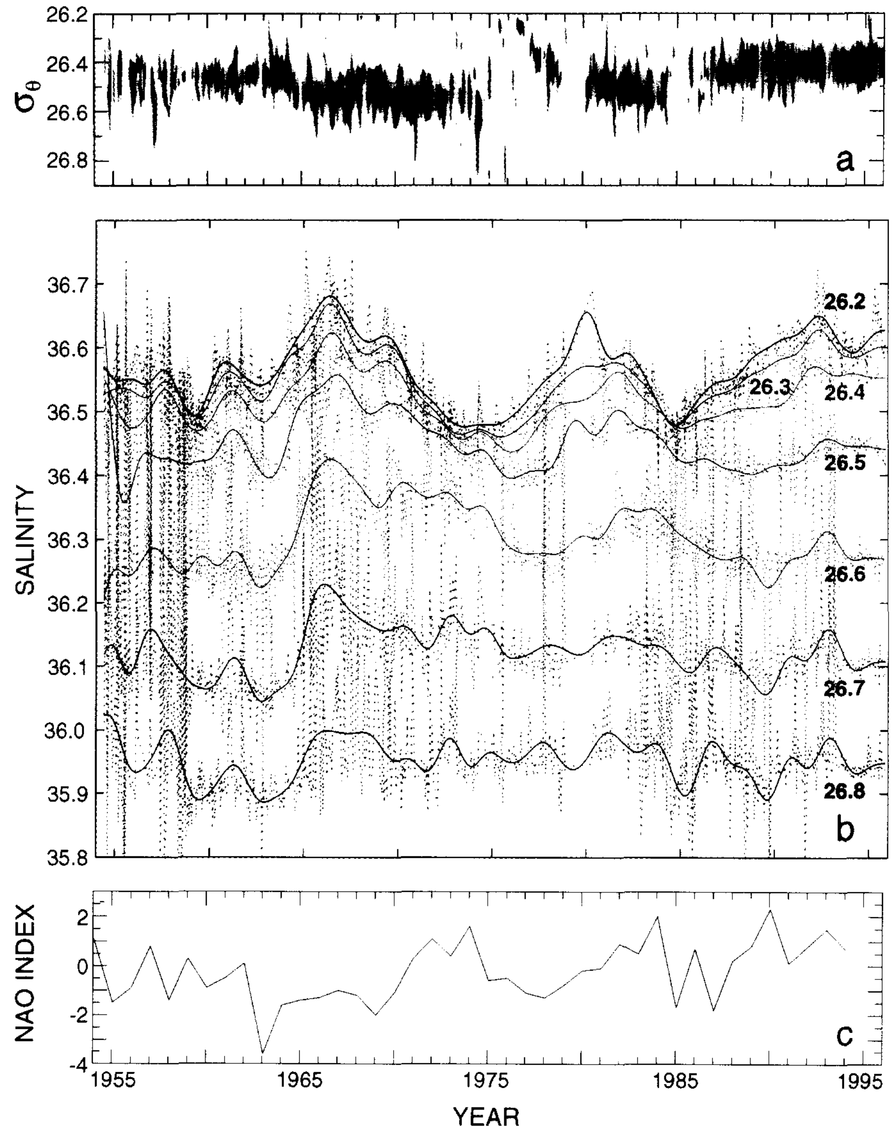

Fig. 5. (a) Isopycnic potential vorticity $(f/\rho)\partial\rho/\rho z(\times 10^{-12}\,\mathrm{cm}^{-1}\,\mathrm{s}^{-1})$ as a function of time and potential density σ_θ at the time series station located at 32°10′N, 64°30′W near Bermuda. Blue is less than 0.7, green is 0.7–1.0, and yellow is 1.0–1.3 $(\times 10^{-12}\,\mathrm{cm}^{-1}\,\mathrm{s}^{-1})$. (b) Salinity on isopycnals 26.2–26.8 σ_θ, spanning the STMW layer. Solid curves are low-passed with a cutoff period of two years. Dashed curves are the actual data. Because these are isopycnals, the cycles in salinity mirror those of potential temperature. (c) North Atlantic Oscillation index (atmospheric pressure at Iceland minus atmospheric pressure at the Azores, December–February average), courtesy of D. Cayan.

the Coriolis parameter, ρ is density, and z the vertical coordinate, was computed from the Brunt–Vaisala frequency. Isopycnic salinity and potential vorticity are shown in Fig. 5 along with the NAO index, computed and provided by D. Cayan, and similar to that of Hurrell (1995). The STMW core is defined as the minimum in isopycnic potential vorticity at densities lower than $26.9\sigma_\theta$. STMW properties are shown in Fig. 7. The mean and standard deviation of each property on selected isopycnals surrounding STMW and at the STMW potential vorticity minimum core are given in Table 2, using data taken after 1959 because

Table 2
Average STMW and isopycnal properties near Bermuda (1960–1991)

	STMW core	$26.3\sigma_\theta$	$26.4\sigma_\theta$	$26.5\sigma_\theta$	$26.6\sigma_\theta$
Depth (m)	287 ± 96	180 ± 55	253 ± 67	362 ± 74	485 ± 53
$\overline{\theta}(^\circ C)$	17.88 ± 0.63	18.61 ± 0.16	18.12 ± 0.13	17.57 ± 0.19	16.64 ± 0.23
$\overline{S}$(psu)	$36.50 \pm .09$	36.54 ± 0.05	36.51 ± 0.04	36.46 ± 0.06	36.30 ± 0.07
$\overline{O_2}$(%sat.)	91 ± 5	92 ± 3	90 ± 3	87 ± 3	82 ± 4
$\overline{\sigma}_\theta$	26.45 ± 0.12				

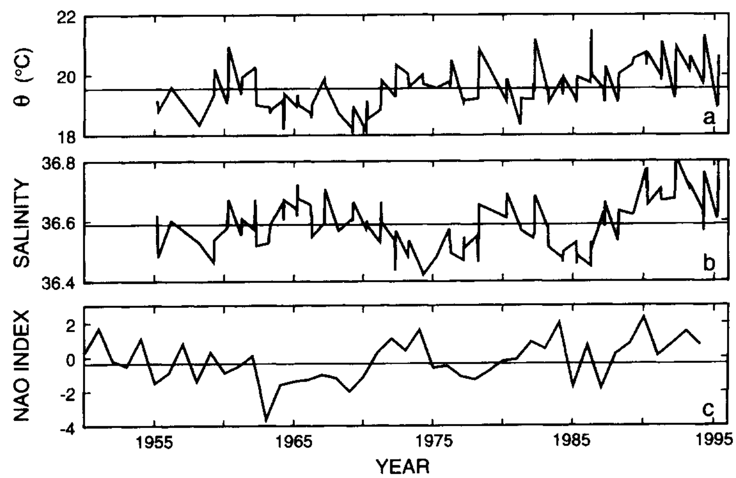

Fig. 6. (a) Surface temperature (°C) and (b) surface salinity (psu) in March/April at the Bermuda station (32°10′N, 64°30′W). (c) NAO index from Cayan, as in Fig. 5. The lines show the average values for 1954–1991.

of the relatively larger scatter in salinity and oxygen prior to that year.

At Bermuda, late winter surface salinity and isopycnic salinities/temperatures from all months were high at three times: the mid 1960s, around 1980 and after 1990. There were three low periods: around 1958, 1973 and 1985 (Figs. 5 and 6). Surface temperature shows a long-term warming of about 1°C over 40 years; its decadal variability is noisy and not well correlated with surface salinity, as shown by Joyce and Robbins (1996) and Reverdin et al. (1996), but late winter surface temperature does reach a minimum in the late 1960s, in the latter part of the long high salinity period. Thus there is a suggestion of a slight negative correlation between temperature and salinity although it is unlikely to be significant. The peak to trough variations in salinity and potential temperature on the near-surface isopycnals were 0.2 psu and 0.6°C.

Salinity (and hence potential temperature) on isopycnals (Fig. 5(b)) shows a time lag with increasing density during the cooling (freshening) part of the cycle. The increase in salnity around 1965 occurred on all isopycnals shown, with almost no time lag with depth. The salinity increase in the late 1970s however did show a lag with increasing density. During periods of increased surface and isopycnic salinity, penetration of high oxygen saturation to greater density also occurs (not shown). If one looks at the lagged freshening alone, the associated vertical velocity is approximately 350 m in 3 years, or about $4 \times 10^{-4}\,\mathrm{cm\,s^{-1}}$.

Subtropical Mode Water core characteristics at Bermuda reflect the near-surface and isopycnic behaviors. Fig. 5(a) prepared for Dickson et al. (1996) for their North Atlantic convection study shows the STMW core density and strength using isopycnic potential vorticity $(f/\rho)\partial\rho/\partial z$ (ignoring relative vorticity) as a function of time and potential density, σ_θ. Lowest potential vorticity indicates the thickest layer. (Computer contouring in Fig. 5(a) results in

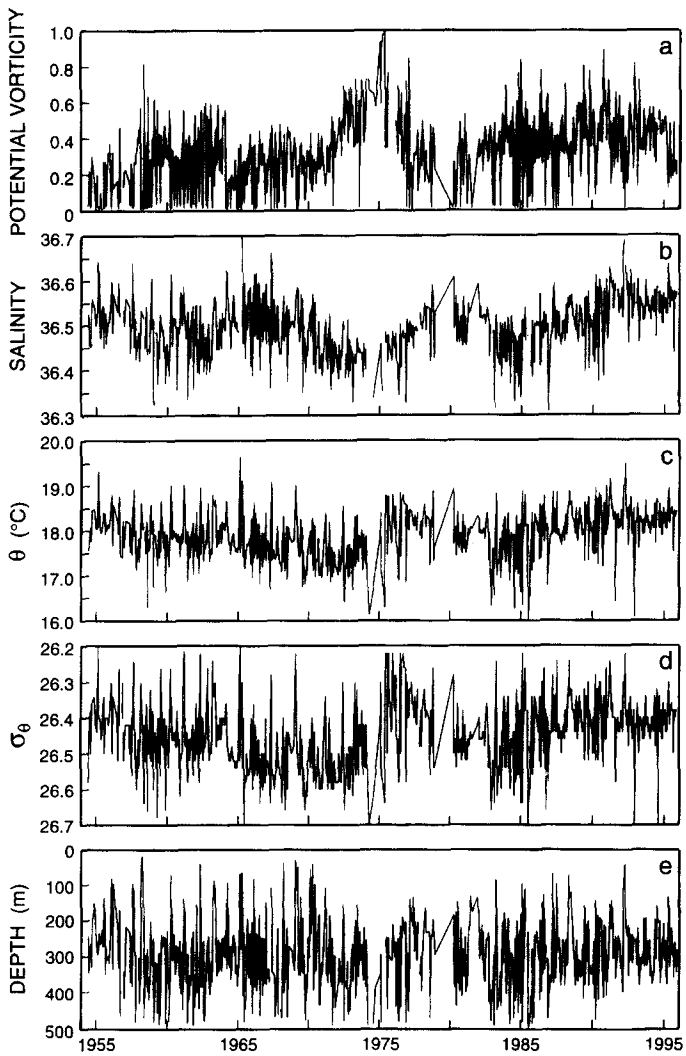

Fig. 7. (a) Isopycnic potential vorticity $(f/\rho)\partial\rho/\partial z(\times 10^{-12}\ \mathrm{cm}^{-1}\ \mathrm{s}^{-1}$; (b) salinity (psu), (c) potential temperature (°C); (d) potential density ($\mathrm{kg\,m}^3$); (e) depth (m) at the potential vorticity minimum lying between 26.0 and 27.2 σ_θ, characterizing Subtropical Mode Water at the Bermuda station (32°10′N, 64°30′W).

poorer resolution of the annual cycle than in the Talley and Raymer (1982) presentation, which was hand contoured.) The core of STMW, as measured by its isopycnic thickness or potential vorticity, was strong in three periods: 1956–1957, the latter 1960s, and the early 1980s, all coinciding with low NAO. The protracted low NAO in the late 1960s was matched by an especially strong STMW core. As noted in Talley and Raymer (1982), the STMW was exceptionally weak and of low density and high temperature from 1975 to 1978; extension of the time series through 1995 shows that this weak STMW period was anomalous compared with the whole record. Other weak STMW periods were the late 1950s and around 1985. All weak STMW periods coincide with low surface/isopycnic salinity and high NAO.

Salinity at the STMW core (Fig. 7(b)) mirrors the strong decadal oscillation of isopycnic salinity/temperature. Potential temperature, potential density and potential vorticity were less oscillatory, responding only to possibly stronger forcing events. Nevertheless, since the strong events (which are taken to be the times of high salinity on isopycnals and at the sea surface) were approximately decadal, these other properties had some decadal variability as well. Potential temperature of the STMW core decreased from 18°C to 17.2°C from 1954 to 1974 with one slightly larger decrease around 1965. Between 1974 and 1975, it jumped to almost 19°C and settled in at 18°–18.5°C from 1980 to 1990, with a low in the mid-1980s corresponding to low NAO. Thus lower values or decreases occurred at the end of low NAO cycles, with one large increase at the onset of new convection in 1975. Potential density followed a similar pattern with the core at about $26.45\sigma_\theta$ from 1956 to 1964, followed by a jump to $26.55\sigma_\theta$ which lasted until 1975, with a sudden decrease to $26.3\sigma_\theta$ and slow increase back to $26.45\sigma_\theta$ by 1980. There was another higher density period around 1985, and then a low density of $26.4\sigma_\theta$ through 1995. Pressure of the STMW core remained around 300 dbar throughout the 35-year record, with a shoaling to about 250 dbar between 1975 and 1982, and increased depth to almost 400 m during the low salinity periods around 1960 and 1985; the mean pressure was 291 ± 41 dbar for the whole record (Table 3). The overall patterns (more saline, colder, higher density when the NAO was low) suggest that STMW was being formed vigorously during the low phases of the NAO compared with the high NAO periods.

The changes observed at Bermuda are typical of a much larger region of the western subtropical gyre. Molinari et al. (1996, in preparation) relate the temperature anomalies from the surface to 400 m to the NAO in most of the subtropical Atlantic west of Bermuda. Based on their figures, the Bermuda time series is typical of the region south of 34°N and east of about 76°W, that is, south and east of the Gulf Stream and its intense recirculation. But throughout the western North Atlantic, they find variations correlated with the NAO. They show that the ocean properties lag the atmospheric changes, with increasing lag with depth. They also show no apparent change in Gulf Stream intensity or position as a function of the NAO.

Changes in properties at greater depths at Bermuda have been examined by Levitus et al. (1995) and Joyce and Robbins (1996). The trends they report were mentioned in Section 4. Levitus et al.'s salinities at 1750 m show two big increases which are in phase with the two big increases in the STMW layer: centered around 1965 and centered around 1985.

Reverdin et al. (1996) have shown that Bermuda near-surface salinities are well correlated with upper ocean salinity at Flemish Cap (47°N, 47°W) and Ocean Weather Station C (52.8°N, 35°W), both of which are well north of the Gulf Stream, with no apparent phase lag. At the two higher latitude sites surface temperature is positively correlated with salinity, increasing (warming) when salinity increases. The Bermuda time series differs in having little or slightly negative correlation between surface temperature and salinity.

The NAO (Figs. 5(c) and 6(c)) is especially low when the upper ocean salinity is high at all three sites. Visually there appears to be a time lag, with the NAO leading the salinity, perhaps by one to two years. Cayan (1992), Kushnir (1994), Dickson et al. (1996) and Reverdin et al. (1996) remark that a low NAO is accompanied by weak westerlies at the higher latitudes (45–60°N) and strengthened westerlies at lower latitudes (25–40°N). Dickson et al. (1996) also note that the winter storm tracks associated with the westerlies retreat southward and westward when the NAO is low, hence enhancing convection in the subtropical gyre offshore of the Gulf Stream due to the local increase in storminess. As remarked above, Dickson et al. (1996) have shown that the NAO and its consequences reached a century low in the 1960s.

An hypothesis for the observed variations at Bermuda is described in the following paragraphs: that there is some increase in local (Sargasso Sea) convection due to increased storminess and latent heat loss, plus increased transport of saline waters and decreased transport of fresher waters into the western subtropical gyre when the NAO is low.

Cayan (1992) shows increased ocean surface heat loss in the western subtropical gyre when the NAO is low. (Since Bermuda lies near a node of his Eastern Atlantic pattern, most variation in surface winds and heat loss is due to the NAO.) He shows that changes in heat loss are primarily due to latent heat flux, associated with evaporation. Increased storminess and latent heat loss would increase the density and salinity of the winter surface layer and STMW, and decrease their temperature. The winter surface layer depth would increase. Isopycnic oxygen saturation would also increase due to penetration of high surface oxygens to greater densities. We see each of these effects in the Bermuda time series.

However, local forcing cannot be the entire answer. It is easy to demonstrate that the observed changes in salinity at Bermuda are too large for local forcing, and there also appears to be a time lag with the NAO leading (Fig. 6). Cayan (1992) shows a 30 W m^{-2} increase in latent heat loss in this region during the low NAO periods compared with high. If such a heat loss were spread over a layer 250 m thick (wintertime mixed layer thickness), it would result in a temperature decrease of 0.08°C and a salinity increase of 0.004 psu over one month. The observed variation in temperature at the core of the STMW is about 0.6°C, and so would require an increased heat loss over 7– 8 winter months (which might be spread over several years), whereas the observed salinity change is about 0.2 psu, which requires about 50 months of increased latent heat loss.

Secondly, variations in salinity are much more pronounced and oscillatory compared with temperature, density, potential vorticity, and oxygen at the surface, on isopycnals, in the STMW core, and even at much greater depth (e.g. Levitus et al., 1995). There is also a clear phase lag with depth during the isopycnic cooling/freshening part of the cycle. This suggests an influx of higher salinity water throughout the water column when the NAO is low, and a relaxation to subduction properties from the northern subtropical gyre when the NAO is high. The near-surface high salinity source would be the subducted high salinity subtropical underwater (Worthington, 1976), and the deeper source would be the Mediterranean tongue water to the east (e.g. Reid, 1994), with a decrease in the amount of higher latitude, fresher water entering either from the North Atlantic Current or directly across the Gulf Stream. Sarmiento et al. (1982) mapped salinity on near-surface isopycnals in the early 1970s; this corresponds to a time of high NAO and low salinity on the Bermuda isopycnals. Their mapped salinities near Bermuda match those shown in Fig. 5(b), and indicate that the advective source of more saline water is subduction from the sea surface east of the mid-Atlantic Ridge, whereas the source of fresher water would be the Gulf Stream and slope water region to its north. Thus the simultaneous shift to higher salinity in the upper 1750 m at Bermuda in the 1960s suggests that eastern subtropical water had a much increased influence in the Sargasso Sea in that period. The apparent time lag between the NAO and the salinity at Bermuda also suggests an advective response.

This scenario fits with that of Reverdin et al. (1996) for changes in the subpolar Atlantic. They concluded that reduced cooling (higher surface temperature) is due to weakened westerlies there during low NAO periods. The westerlies do not simply decrease in strength during low NAO, rather the strong westerlies shift southwestward out of the Iceland–Azores region and into the Sargasso Sea and Gulf Stream area. The increased salinity at the subpolar sites during low NAO could be accomplished by an increase in Gulf Stream transport, consistent with this southwestward shift of the strong westerlies (e.g. Dickson et al., 1996). This shift of the westerlies also increases convection in the Sargasso Sea, producing stronger STMW which is also slightly colder. It appears also to cause increased advection of the saline waters from the eastern subtropical gyre into the Sargasso Sea, resulting in salinity increase through the STMW following the lowest NAO events.

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Acronyms

AABW	Antarctic Bottom Water
AAIW	Antarctic Intermediate Water
DSOW	Denmark Strait Overflow Water
DWBC	Deep Western Boundary Current
LSW	Labrador Sea Water
MW	Mediterranean Water
NAC	North Atlantic Current
NADW	North Atlantic Deep Water
NAO	North Atlantic Oscillation
NSOW	Nordic Seas Overflow Waters
SPMW	Subpolar Mode Water
STMW	Subtropical Mode Water
SSS	Sea surface salinity
SST	Sea surface temperature

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