New perspectives on eighteen-degree water formation in the North Atlantic

Terrence M. Joyce

Abstract In this report, eighteen-degree water (EDW) formation will be discussed, with emphasis on advances in understanding emerging within the past decade. In particular, a recently completed field study of EDW (CLIMODE) is suggesting that EDW formation within a given winter can have at least two different dominant physics and distinct locations: one type formed in the northern Sargasso Sea, largely away from the strong flows of the Gulf Stream where 1D physics may apply, and a second type formed along the southern flank of the Gulf Stream, in a region where the background vorticity of the flow and cross-frontal mixing play key roles in the convective formation process.

Keywords Mode water · Air–sea interaction · Gulf Stream

1 Introduction

The subtropical-mode water of the North Atlantic Ocean was first identified as having remarkably uniform temperature (17.9 ± 0.3°C) and salinity (36.50 ± 0.10 psu) properties by Worthington (1959), who named it eighteen-degree water (EDW), a name which has persisted over time. It is very evident in volumetric atlases from hydrographic data (Wright and Worthington 1970) and here using the modern gridded WOCE-era electronic atlas (Gouretski and Koltermann 2004, herein Fig. 1). One

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distinct "mode" of this 2D histogram of waters warmer than 7°C in the N. Atlantic is clearly the EDW identified by Worthington. Mode waters as a generic entity (Hanawa and Talley 2000) exist in all the major ocean basins and are pervasive water masses that can be detected throughout the year both locally, where they are formed, and in locations far from the formation site(s). Worthington identified the Sargasso Sea as the location where EDW could be readily found and the northern reaches of the Sargasso Sea, just south of the separated Gulf Stream, as the region of formation. Formation has generally been determined by where the mode water is seen to outcrop at the ocean surface in late winter. Worthington refined his ideas about EDW in a later study (Worthington 1976) where he associated the formation with the large mean negative oceanic heat flux to the atmosphere. At the time, information about ocean/atmosphere exchange was limited to rather coarse summaries from available shipboard data collected over time, and was clearly affected by limited sampling during periods of maximum exchange. These early summaries by Budyko (1963) and Bunker (1976) painted the first basin-wide picture of air–sea exchange. Worthington argued that the mean heat-loss region over the separated Gulf Stream (GS) was responsible for transforming subtropical waters into EDW, whereas Warren (1972) argued that the region where EDW was "formed", near 35N, 60W was one of zero net heat loss—leading him to conclude that EDW formation was the process by which the locally formed seasonal thermocline was removed during winter: something that occurs in the absence of strong advection and requiring no net heat loss when averaged over the year. Confounding these early ideas about EDW formation was the problem inherent in estimating heat exchange over the ocean, where short-wave and long-wave radiation measurements were few, and estimates of turbulent exchange

Volumetric T/S Census for warm water (>7°C) in N. Atlantic Ocean
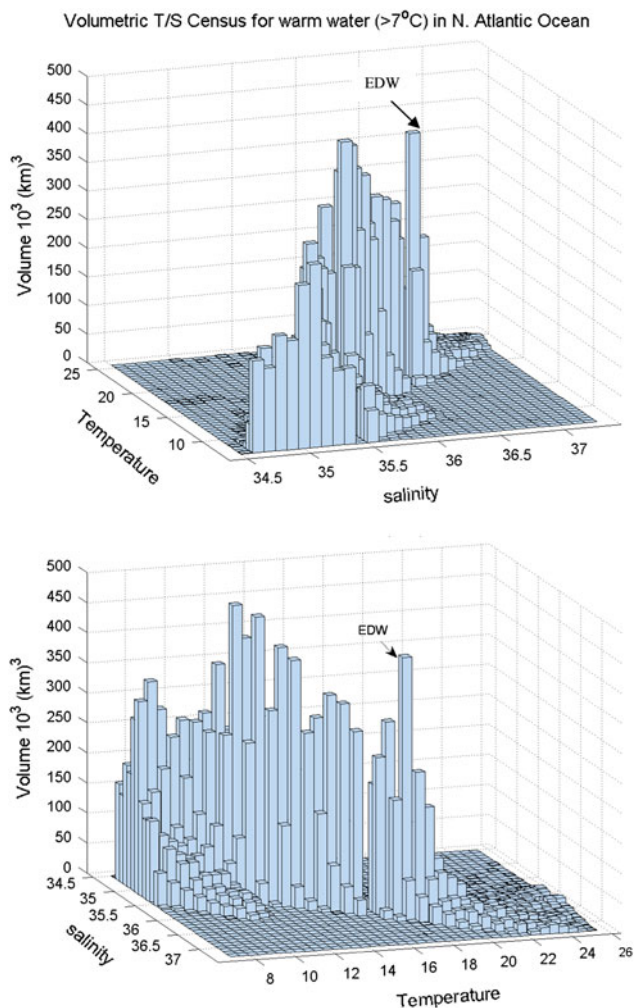
Fig. 1 Two different views of a 2D histogram showing the volume (in 10^3 km^3) of N. Atlantic waters having temperatures >7°C, based on the WOCE Global Hydrographic Climatology (Gouretski and Koltermann 2004). The total volume of EDW bounded by (36.4–36.6 psu, 17.5–18.5°C) is estimated to be 912 × 10^3 km^3

of sensible and evaporative heat were limited by both data and models of turbulent exchange over the ocean.

Walin's (1982) ideas about direct estimation of water mass formation using air–sea exchange data were a major boost to the field, yet application to EDW formation (Speer and Tzipermann 1992; Maze et al. 2009) yielded greatly different estimates of EDW formation. This has prompted a concerted effort to collect field observations and conduct modeling studies to better understand the different physical processes in the ocean associated with water mass formation (advection, subduction, mixing) and better parameterize turbulent exchanges with the atmosphere in the northern Sargasso Sea: this experiment was called CLI-MODE (CLIvar MOde water Dynamics Experiment; The Climode Group 2009). In this update, an attempt will be made to summarize what we have learnt about EDW since

the review of Hanawa and Talley (2000), including the recently published results from CLIMODE. However, much of the CLIMODE analysis has yet to appear in print and it is expected that this review would be quite different if written a few years from now. It is hoped that enough of the recent thinking will show where advances are to be expected and will serve as a current assessment for EDW, to be contrasted with major contemporary efforts (reported elsewhere in this collection) on mode waters found in other ocean basins.

2 EDW definitions

To begin, it is instructive to see how one might define EDW in order to then estimate how much of it there is and where it is formed. Worthington (1976, Table 6) noted that although there were approximately 2512 × 10^3 km^3 of EDW (waters with temperatures between 17 and 19°C) in the N. Atlantic, if one looked in the North American Basin (west of the mid-Atlantic Ridge), the volume was reduced to 1763 × 10^3 km^3. Siedler et al. (1987) have since shown that a substantial amount of mode water with a temperature of ca. 18°C is formed every winter in the North African Basin, near the island of Madeira. This type of subtropical mode water was called Madeira mode water. Unlike its counterpart in the Sargasso Sea, by the end of summer heating, there is little evidence of a homogeneous ther-mostad, or region of reduced vertical temperature gradient: it has vanished into the seasonal thermocline (Siedler et al. 1987; Fig. 9). If one confines oneself to the region west of the Mid-Atlantic Ridge one can limit the contribution of this eastern basin water mass from the census.

The EDW volume can further be refined by estimating the anomaly resulting from the reduced temperature gradient of EDW imbedded in the permanent thermocline. Worthington (1976; Table 8) did this for the western North Atlantic by subtracting the volume of 16–17 and 19–20°C water from that of 17–19°C water in his census and found that the EDW volume for the western N. Atlantic was further reduced by approximately 50% to 889 × 10^3 km^3: thus for one author alone, published estimates of EDW volume differ by a factor of 3, depending on its definition. The last EDW volume estimate is quite close to one based on T/S criteria presented below. Others have tried looking at anomalies of EDW by limiting it to those waters having a temperature gradient less than some limiting value (Kwon and Riser 2004) or some limiting planetary poten-tial vorticity appropriate to the thermocline. Here we use both temperature and salinity constraints and find, on the basis of Fig. 1, the volume of EDW between 17.5 and 18.5°C with bounding salinities of 36.4 and 36.6 psu is 912 × 10^3 km^3; so addition of a salinity criterion and

restricting the temperature definition reduces the overall N. Atlantic numbers substantially. As a check on the new climatological data, the total volume was calculated with the WOCE-era atlas without salinity constraints and estimated for 7–19°C water to be 2237×10^3 km^3, within the error bars of the Forget et al. (2011) figure of 2360×10^3 km^3, and slightly less than Worthington's value given above. Clearly, a restricted temperature definition and use of salinity to define the EDW volume (this work) makes a substantial difference. Limiting the spatial domain (western basin) and requiring the local temperature gradient to be less than some reference value are the two principle issues that need further resolution in the literature in the definition of EDW volume. Resulting EDW volume estimates are clearly dependent on the definitions and thus arises the potential for conflicting claims on volumes, formation rates, and ventilation timescales, as we will see later.

Talley and Raymer (1982) first pointed out that there were spatial variations of EDW within a given winter and that there was substantial year-to-year variability in the thickness and other properties at Bermuda. Joyce et al. (2000) showed that interannual variations of EDW PV, as observed at Bermuda, were well-correlated with the NAO (with no lag/lead), suggesting that changes in air–sea forcing as reflected in major climate indices were important to newly formed EDW observed at Bermuda. Because EDW is not locally formed near Bermuda in most winters, its time of arrival there has been estimated as a few to

several months after formation by using oxygen (Jenkins 1982) or PV (Talley and Raymer 1982; Phillips and Joyce 2007). We will return later to the issue of spatial variations of EDW formed in a given winter with some new data drawn from CLIMODE. Because newly formed EDW is highly saturated with dissolved oxygen, Jenkins and Goldman (1985) were able to study biological processes affecting EDW oxygen consumption after subduction by calibrating change against "true" age determined from tritium/helium-3.

As was shown by Schmitz (1996; Fig. 1-21) the southern re-circulation gyre of the GS occupies the region bounded by a box with lat/lon corners at (40N, 50W) and (30N, 55W) to the east and (35N, 75W) and (25N, 75W) to the west. This roughly agrees with the float-based circulation on the 26.5 kg m^{-3} surface by Kwon and Riser (2004) reproduced here (Fig. 2). The float trajectories were used to map the geostrophic pressure on an isopycnal associated with EDW. Data for this circulation map are drawn from a period from 1998 to 2002. The 60 cm contour (Fig. 2) roughly delineates the southern GS re-circulation gyre identified by Schmitz (1996; Fig. 1-21). Clearly the mean circulation is defining or being defined by the presence of EDW, which is largely contained within this region, on the basis of anomaly calculations using a reference vertical temperature gradient. The WOCE mean climatology atlas was used to examine subtle water mass variations of EDW in the Sargasso Sea (Fig. 2, right panel). One can see that saltier (and also warmer—not shown)
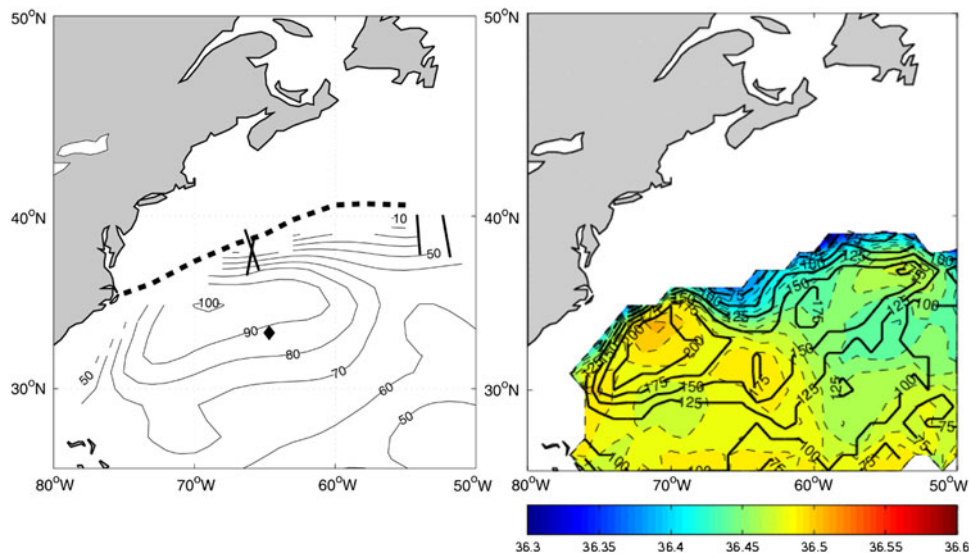

Fig. 2 *Left panel* from Kwon and Riser (2004), the geostrophic pressure on the $\sigma_\theta = 26.5$ kg m^{-3} potential density surface (in cm), based on mean float velocities, is plotted with hydrographic lines from the CLIMODE winter cruise of the Knorr in 2007 (*thick black lines*), the north wall of the Gulf Stream defined by the 200 m temperature $= 15$°C from 1955–2008 (*heavy dashed line*), and the location of

Bermuda (*black diamond*). *Right panel* from the WOCE mean climatology, the thickness of the 17.5–18.5°C temperature surface (*solid black contours*) has been plotted for all values exceeding 50 m. The color represents the average salinity (psu) in this layer, and the contour interval (*dashed lines*) is 0.01. The thickness and salinity reveal a climatological pattern of EDW variation in the Sargasso Sea

varieties of EDW are found within the recirculation center, whereas fresher (and colder) varieties surround them, with a clear tendency for fresher values along the northern boundary in proximity to the Gulf Stream. Because this representation includes all seasons, the wintertime properties are obscured, especially the thickness, which is reduced because of restratification.

3 Annual EDW budgets and formation

EDW formation budgets using the "Walin" method have been nicely exemplified by Forget et al. (2011) for the 3-year period 2004–2006, inclusive. Various data sets (e.g. SST, SSH, Argo floats) have been assimilated in a 1° × 1° lat/lon resolution numerical model that has 16-month, overlapping annual cycles for the above period. Estimates for air–sea exchange have been examined with the inverse machinery and "adjusted" when necessary to reduce model/data misfits. The resulting annual EDW cycle has been compared with more traditional, float-based estimates of EDW volumes during the same period. The region considered is the entire N. Atlantic as reflected in surface areas and volumes in various temperature classes with 1°C resolution; there was no attempt to stratify the calculation using salinity or surface salinity forcing. The amount of EDW formed annually during the winter months is estimated to be 8.6 ± 1.8 Svy (1 Svy = 31.5×10^3 km³), where they have been able to account for the effects of subgridscale mixing in the model, which actually acts to increase EDW formation during winter, although reducing EDW formation during the remainder of the year. A previous estimate of the amount of EDW produced in winter (Kwon and Riser 2004) of 3.5 ± 0.5 Svy was based on increases in EDW volume during winter from hydrographic data alone. Although the error bars of the two different estimates do not overlap, they are remarkably close considering that the EDW definition criteria chosen by Kwon and Riser (2004) focus on the western N. Atlantic region and satisfy a constraint that the vertical temperature gradient be less than some maximum value, which we saw earlier gave a factor of 3 reduction in EDW volume estimated by a single author (Worthington 1976). Yet both cited works define EDW formation on the basis of volume increases in winter—so how volume is defined is inherent to estimates of annual formation. Whereas the Kwon and Riser (2004) calculation is independent of air–sea forcing, relying only on changes in EDW volume based on hydrography, the Forget et al. (2011) calculation is not. So we need to address the sensitivity of "formation" rates to the input air–sea fluxes.

Wintertime air–sea heat loss over the region of EDW formation can be o(300 W m^{-2}) with uncertainties of

o(10%). Joyce et al. (2011) have shown that the maximum of winter heat and water loss in the N. Atlantic is located exactly over the location of the meandering, separated GS, with the warm core of the GS defining the region of largest ocean–atmosphere exchange in winter. During the period of CLIMODE (Feb/Mar 2007, Fig. 3) the winter mean SST field indicates that after traversing the region from 75W to 55W, the warm surface core water of the GS eventually disappears. If one were to choose the 17.5 and 18.5°C isotherms for Walin-type formation analysis, it would be clear (Fig. 3) that the domain would be multiply connected and that 10% errors in the surface heat flux together with the sensitivity in flux products to spatially dependent air–sea temperature and humidity differences could lead to substantial differences in water mass formation calculations. Forget et al. (2011) show that, depending on the flux product used, EDW volume increases during winter range from 5 to 13 Svy based on the Walin-type calculation alone. Because their EDW calculation was constrained by ocean data, air–sea exchange forcing is adjusted to give a better model/data fit and is, therefore, given a reality check that the unconstrained range of estimates lacks. The annual EDW formation rate of ca. 13 Svy by Speer and Tzipermann (1992; Table 1) is at an extreme limit of this Walin-type calculation, although the authors made an effort to do the formation calculation in density rather than temperature space, taking into account freshwater forcing. Presumably a constrained estimate of this calculation using Argo float salinities and newer flux estimates might be obtained in the near future. For the present, it would therefore seem that the range of EDW formation of 3.5–8.6 Svy between the two previously cited estimates of formation based on
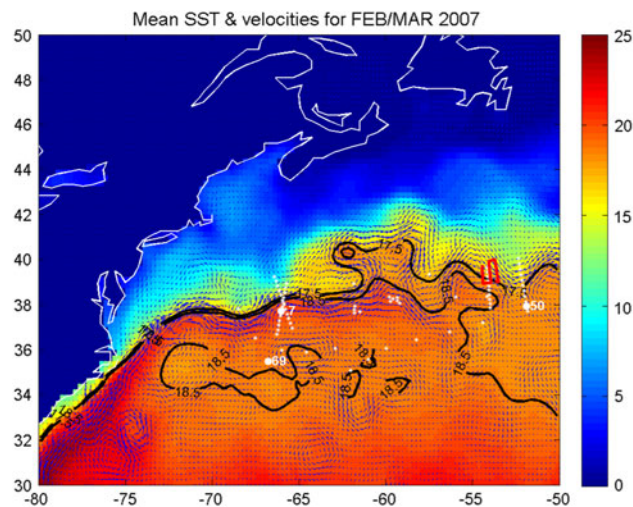

Fig. 3 The February/March 2007 mean SST and SSH results with CLIMODE CTD stations (3 selected for Fig. 4). The EDW outcrop region can be seen by the area between the 17.5 and 18.5°C SST contours. The *red* ladder pattern near 40°N, 54°W is from the Seasoar track of Knorr 188, part of which is shown in Fig. 5

temperature must be because of the definition of EDW. Forget et al. (2011) estimate that the principle agents for dissipation of EDW are air–sea exchange in the warm months, leading to formation of a surface thermocline, and cross-frontal mixing of EDW with higher PV waters, principally found to the north of the Gulf Stream. One final comment related to air–sea exchanges, the CLIMODE program offered some unique extensions for improving the parameterization of fluxes using bulk formulae. With high winds and large air–sea temperature differences, and with clear problems with the algorithms at wind speeds above 15 m s^{-1} (The Climode Group 2009), the expected changes to bulk formulae should improve future estimation of turbulent air–sea fluxes for both heat and water.

4 Multiple types of EDW formed annually

Examination of the outcrop window for EDW (Fig. 3) during CLIMODE indicates two possible scenarios for EDW formation. The first is in the region from 73–63 W at a latitude near 35N. Here in the northern Sargasso Sea and well south of the GS flow, EDW can be seen at the surface. The vertical characteristics of the water are portrayed (Fig. 4) at 3 selected CTD stations from the Knorr cruise in 2007. Station 7, taken in the "upstream" region within the warm surface core of the GS, shows a strong stratification

near the surface and evidence of tropical salinity maximum water (Worthington 1976; Wüst 1936) with elevated salinity at a pressure of ca. 200 dbars. Below this layer, a region of reduced stratification, "old EDW" can be seen with oxygen saturation values below 90%. In the northern Sargasso Sea (Station 69), a homogeneous water mass, "new EDW", can be seen between 100 and 400 dbar pressures with elevated oxygen content, a temperature slightly above 18°C, and a salinity of approximately 36.6 psu. A warmer, less dense surface layer has capped the EDW, probably because of restratification in the presence of lighter surrounding waters. This is also suggested by the lack of an outcrop of the EDW layer near the location of the station (Fig. 3). The salinity of this EDW looks approximately like that of the vertically mixed version of the incoming waters, exemplified by station 7. In contrast, the downstream station near the GS (station 50) is located in the other EDW outcrop zone which stays close to the GS until approximately 55 W where it erupts from the front and fills the region to the south. This EDW is colder and fresher. Joyce et al. (2011) argue that formation of EDW within the frontal region entrains cooler, fresher waters across the GS front from the north and results in a lower salinity, and a slightly colder version of EDW. Both types of EDW are highly saturated with oxygen and are different types of EDW formed in the same year, as was first noted by Talley and Raymer (1982), who speculated

Fig. 4 Selected CTD stations from CLIMODE in 2007. Station 7 (*red curves*) is from the upstream inflow region of the GS just south of the strongest zonal flow. Stations 50 (*black line*) and 69 (*blue dotted line*) are from the regions of newly ventilated EDW along the GS front (50) and in the northern Sargasso Sea (69). The four panels show temperature, salinity, potential density, and percentage oxygen saturation for the stations. Station locations are indicated in *white* on Fig. 3
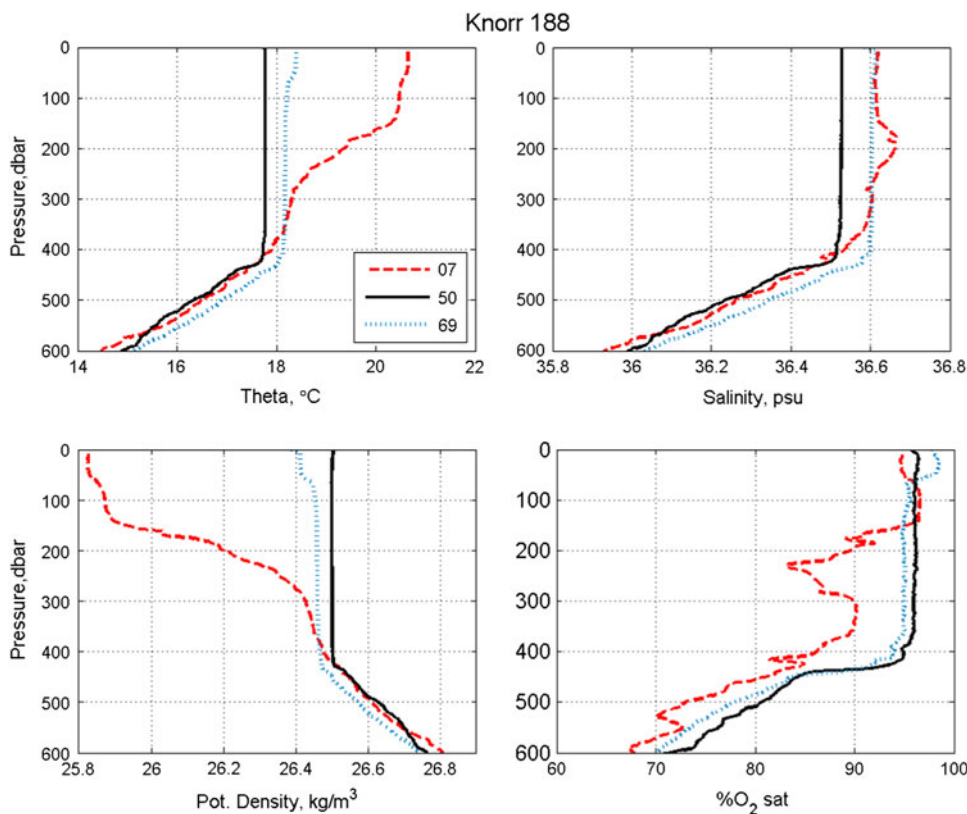
that EDW variations occurred west to east as the EDW further circulated under the strong surface cooling and thus became progressively cooler. According to Joyce et al. (2011), the high-heat loss region of the GS front is also one in which evaporation exceeds precipitation, making air–sea exchange a candidate for increasing, not decreasing, EDW salinity in the region. By invoking the need for lateral mixing across the GS front ($\kappa \sim 100$ m^2 s^{-1}), Joyce et al. (2011) were able to satisfy both the downstream advective budgets for salinity and heat for the EDW formed in the GS, clearly indicating that EDW formed along the GS would necessarily be fresher (and colder) than that found forming further to the south and west. The cooling and freshening of the northern version of EDW is affected by cross-frontal exchange and this is reflected in the mean property changes in the hydrography (Fig. 2, right panel). Forget et al. (2011) did not include salinity in their budget calculations, but found that cross-frontal mixing in winter could actually enhance EDW production. With salinity, it would have been clear that this new component of EDW formation introduced lower salinity water into the region and affected mean EDW properties of temperature and salinity in the northern Sargasso Sea.

It would seem that the two different types of EDW observed in 2007 are the end points of the disparate formation hypotheses of Worthington (1959, 1976) and Warren (1972): frontal formation under the strong air–sea cooling region (Worthington) and formation in the northern Sargasso Sea away from strong advection and cooling (Warren). Thus, the mean properties of EDW are saltier (Fig. 2, right panel) and warmer within the center of the southern recirculation gyre, where Warren argued for one non-advective endpoint for EDW formation. Worthington's (1972) circulation schematic for northward surface flow into the formation region and southward spreading of new EDW away from the region are reasonable if interpreted in density space rather than depth space: lighter waters advect northward and eastward in the GS and are cooled and (slightly) freshened. They become detached from the GS and spread southward to the east of 60W where the southern re-circulation gyre detrains fluid southward from the GS (Fig. 2) and spreads this fresher and cooler version of EDW around the edges of the recirculation gyre. How much of the EDW formation occurs in the two regions remains to be determined. Joyce et al. (2011) estimated that ca. 1.8–2.9 Svy were formed in the GS flow during 2007, which amounts to either 47–82% of the total according to Kwon and Riser (2004), or 21–34% if one uses the Forget et al. (2011) estimate for annual EDW formation. Other CLIMODE data sets, for example Argo floats having salinity sensors, should be able to address this point also.

The formation of EDW within the GS represents a new paradigm for mode water formation, one in which the strong lateral and vertical shear of the flow alters convective processes and allows for symmetric and inertial instabilities (Joyce et al. 2009). Regions of negative potential vorticity are still formed by cooling, but along front wind stress is also an important contributor to the forcing as denser fluid is advected by winds equatorward across the front (Thomas 2005). The forced, slantwise convection can have lateral scales of only a few km and is not easily observed except in sub-mesoscale resolving surveys using devices such as Seasoar (Fig. 5) during CLIMODE. The entire region between 30 and 110 km along the undulating track is filled with new EDW, but closer inspection (see Joyce et al. 2009) reveals some of the distinctive fine structure of the formation: note the subtle changes in oxygen and fluorescence along the track as one passes through waters recently near the surface (elevated values of these tracers) and waters likely to be upwelling from below (lower tracer concentrations). The entire ventilation process is different from 1D convection and is poorly parameterized in models not resolving the sub-mesoscale, and poorly sampled in most ocean observations.

5 Some concluding thoughts and challenges

Were the formation rates and volumes of EDW well established, one could take the ratio of the two to estimate the ventilation rate for EDW: how much of it is ventilated in a typical year. An independent estimate of this was attempted by Jenkins (1982) using tritium/helium-3 dating at Bermuda: the EDW estimate is difficult to extract from his figures because the age change is large with potential density, and the potential density resolution, limited by discrete sampling bottles, is not good for low potential densities (for example, for EDW): an upper limit is 8.5 years (for $\sigma_\theta = 26.5$) from Jenkins (1982; Table 1) and 3 years (Jenkins and Goldman 1985; Fig. 9, depth ca. 230 m), which defines a range nearly identical to that between Kwon and Riser's (2004) estimate of 3.6 years and the Forget et al. (2011) value of 8.7 years. Thus, independent tracer ages do not seem to offer a way out of this conundrum at present. Although we have identified at least two types of EDW formed within a given year, we have not yet ascertained which of the two types is dominant in the annual budget.

The issue of the fate of EDW after formation has not been addressed yet with CLIMODE data. One might expect some revised mean circulation update on that offered in Fig. 2, left panel, and some assessment of the role that eddies play in the spreading and homogenization of EDW from the formation region. One can also expect some future

Fig. 5 EDW formation within the GS is seen on this Seasoar track (*blue line*, *upper left*) with the north on the left hand side of the other panels which show temperature (°C, *upper right*), salinity (psu, *middle left*), σ_θ (kg m^{-3}, *middle right*) with the Seasoar track shown as *thin black lines*, dissolved oxygen (µmol kg^{-1}, *lower left*), and chlorophyll fluorescence (uncalibrated volts, *lower right*)
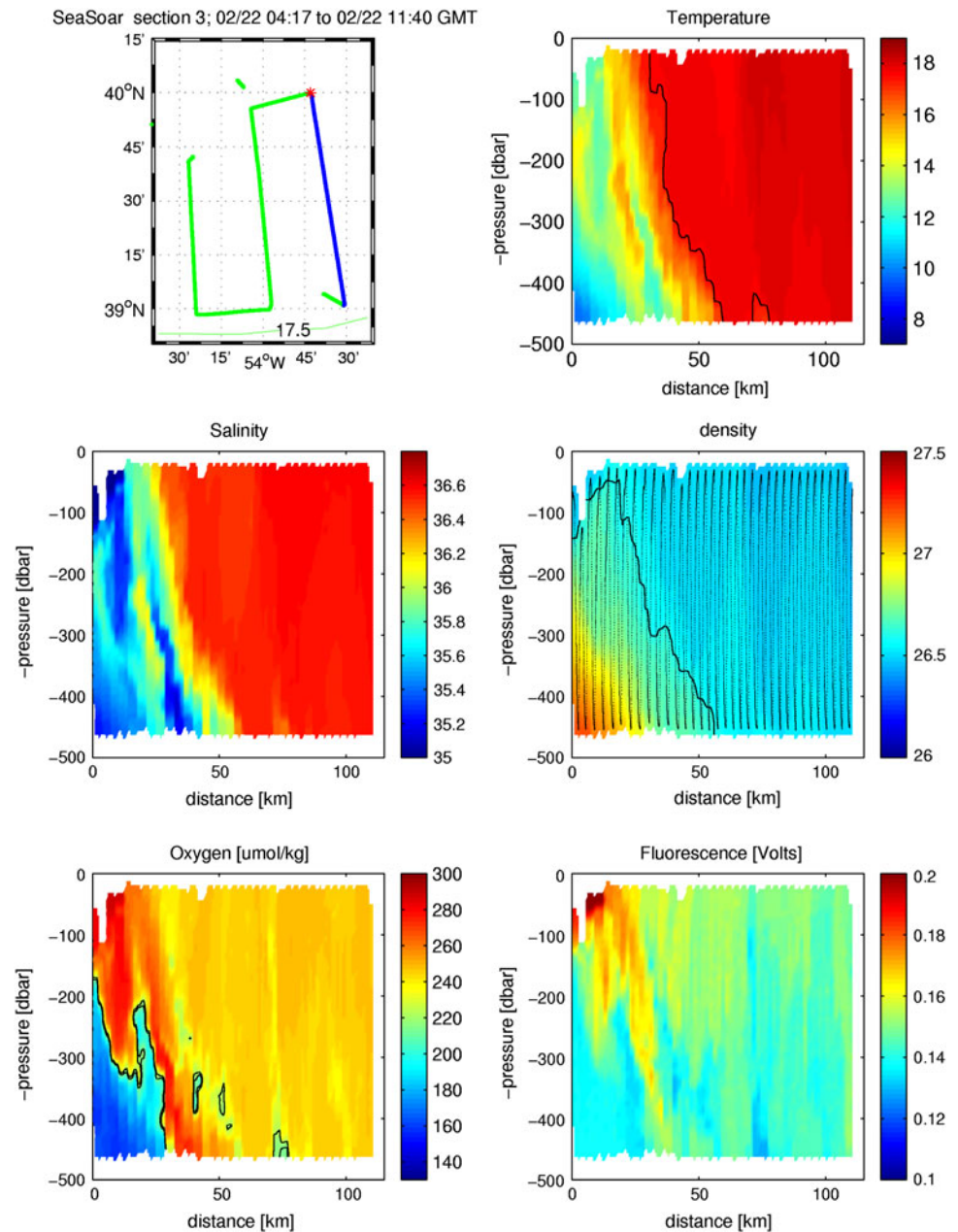
assessment of how representative the CLIMODE period is compared with past winters. Joyce et al. (2011) noted that air–sea exchange during the winter of 2007 was larger over the region than the mean of the previous 19 years. This has not been translated into EDW formation or circulation at this stage and thus represents another aspect of EDW dynamics that may yet emerge from our recent field program. Finally, we should expect that comparative dynamics will emerge for EDW and its close kin, North Pacific subtropical mode water, found immediately to the south of the separated Kuroshio.

References

Budyko MI (1963) Atlas of heat balance of the world. Glav. Geofiz. Obo, Moscow (in Russian). Translation by I.A. Donehoo, Weather Bureau, WB/T106, Washington, DC

Bunker AF (1976) Computations of surface energy flux and annual air–sea interaction cycles of the North Atlantic Ocean. Mon Weather Rev 104:1122–1140

Forget G, Maze G, Buckley M, Marshall J (2011) Estimated seasonal cycle of North Atlantic eighteen degree water volume. J Phys Oceanogr 41:269–286. doi:10.1175/2010JPO4257.1

Gouretski VV, Koltermann KP (2004) WOCE global hydrographic climatology. Berichte des Bundesamtes für Seeschifffahrt und Hydrographie, 35/2004

Hanawa H, Talley L (2000) Mode waters. In: Siedler J, Church J, Gould J (eds) Ocean circulation and climate. Academic Press, New York, pp 373–386

Jenkins WJ (1982) Tritium and 3He in the Sargasso Sea. J Mar Res 40(Supplement):533–569

Jenkins WJ, Goldman JC (1985) Seasonal cycling and primary production in the Sargasso Sea. J Mar Res 43:465–491

Joyce TM, Deser C, Spall MA et al (2000) On the relation between decadal variability of subtropical mode water and the North Atlantic oscillation. J Clim 13:2550–2569

Joyce TM, Thomas LN, Bahr F (2009) Wintertime observations of subtropical mode water formation within the Gulf Stream. Geophys Res Lett 36:L02607. doi:10.1029/2008GL035918

Joyce TM, Thomas LN, Dewar WK, Girton JB (2011) Eighteen degree water formation within the Gulf Stream: a new paradigm arising from CLIMODE. Deep Sea Res II (submitted)

Kwon Y-O, Riser SC (2004) North Atlantic subtropical mode water: a history of ocean-atmosphere interaction 1961–2000. Geophys Res Lett 31:L19307. doi:10.1029/2004GL021116

Maze G, Forget G, Buckley M, Marshall J, Cerovecki I (2009) Using transformation and formation maps to study the role of air–sea heat fluxes in North Atlantic eighteen degree water formation. J Phys Oceanogr 39:1818–1835

Phillips HE, Joyce TM (2007) Bermuda's tale of two time series: hydrostation S and BATS. J Phys Oceanogr 37:554–571

Schmitz WJ Jr (1996) On the world ocean circulation, vol I. Some global features/North Atlantic circulation. WHOI Technical Report WHOI-96-03

Siedler G, Kuhl A, Zenk W (1987) The Madeira mode water. J Phys Oceanogr 17:1561–1570

Speer K, Tzipermann E (1992) Rates of water mass formation in the North Atlantic Ocean. J Phys Oceanogr 22:93–104

Talley LD, Raymer ME (1982) Eighteen degree water variability. J Mar Res 40(Supplement):757–775

The Climode Group (2009) Observing the cycle of convection and restratification over the Gulf Stream system and the subtropical gyre of the North Atlantic ocean: preliminary results from the CLIMODE field campaign. Bull Am Meteorol Soc 90:1337–1350

Thomas LN (2005) Destruction of potential vorticity by winds. J Phys Oceanogr 35:2457–2466

Walin G (1982) On the relation between sea-surface heat flow and thermal circulation in the ocean. Tellus 34:187–195

Warren BA et al (1972) Insensitivity of subtropical mode water characteristics to meteorological fluctuations. Deep Sea Res 19:1–19

Worthington LV (1959) The 18° water in the Sargasso Sea. Deep Sea Res 5:297–305

Worthington LV (1972) Negative oceanic heat flux as a cause of water-mass formation. J Phys Oceanogr 2:205–211

Worthington LV (1976) On the North Atlantic circulation. Johns Hopkins oceanographic studies. The Johns Hopkins University Press, Baltimore

Wright WR, Worthington LV (1970) The water masses of the North Atlantic Ocean: a volumetric census of temperature and salinity. Ser Atlas Mar Environ 19:8 (7 plates)

Wüst G (1936) Oberflächensalzgehalt, Verdunstung, und Niederschlag auf dem Weltmeere. Lünderkundliche Forschung. Festschrift Norbert Krebs, Stuttgart, pp 347–359