Buoyancy Flux at Ocean Weather Station Bravo

S. SATHIYAMOORTHY AND G. W. K. MOORE

Department of Physics, University of Toronto, Toronto, Ontario, Canada

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ABSTRACT

Deep water formation at high latitudes is believed to be the driving mechanism behind the ocean’s thermohaline circulation. The exchange of heat and water with the atmosphere causes the density of the surface waters to change, with subsequent downwelling and upwelling resulting as the system relaxes toward convective equilibrium. The characteristics of this atmosphere–ocean exchange are examined by studying the temporal variability of the buoyancy flux at OWS Bravo, a location where deep water formation is known to occur. The authors find that there is significant high-frequency variability in the buoyancy flux attributable to the passage of synoptic weather systems, variability that is masked by monthly analyses. At high latitudes, precipitation plays a significant role in the buoyancy flux. If it is ignored, the buoyancy loss is overestimated (positive coordinate is downward). Precipitation also causes the buoyancy flux to become positive during the passage of a cyclone. The timescale for this change in buoyancy flux is found to be similar to the timescale for the convective plumes in the ocean, suggesting a link between the two. In addition, a strong negative correlation is found to exist between the sensible heat flux at Bravo and the North Atlantic Oscillation.

1. Introduction

The circulation of the abyssal ocean is thought to be driven by the formation of deep water. Deep water is formed only in certain parts of the world’s ocean, including the Labrador Sea, the Nordic seas, and the Weddell Sea. Air–sea interaction is at the heart of the deep water formation process. An exchange of heat and mass with the atmosphere results in the upper layers of the ocean becoming dense enough to sink and displace less dense waters to the surface. Herein lies the atmosphere’s pivotal role: the air–sea coupling, both on short and long timescales, can determine the intensity and the frequency of the convection.

Very few data have been collected in the high-latitude regions where open-ocean convection occurs. Fortunately, historical datasets from ocean weather stations (OWSs) situated in various locations in the North Atlantic are available. These OWSs have now been decommissioned, but they have provided us with a relatively long record of surface meteorology (20 to 30 years worth) at a temporal resolution high enough to resolve individual weather systems (Diaz et al. 1987). In this study, we investigate what the data from OWS Bravo can tell us about the forcing that drives deep water formation in the Labrador Sea basin.

The circulation in the Labrador Sea is mainly cyclonic, with the West Greenland Current on the east, the Baffin Island Current along the north, the Labrador Current along the west, and the Atlantic Current to the south (see Fig. 1). This gyre circulation does two things: First, it raises the isopycnal surfaces, thereby weakening the stratification of the upper layers of the water. Second, it acts to isolate the water in the center, allowing the surface waters to be subjected to the preconditioning phase for longer.

Deep convective activity is generally understood to be a three stage process (Clarke and Gascard 1983). The first stage is a preconditioning phase that can last from weeks to months. The preconditioning in the Labrador Sea comes from the continual stream of cold dry arctic air flowing over the relatively warmer ocean. The second stage is a deep mixing phase, initiated by strong surface forcing, and the final stage is a spreading phase. The intense surface forcing can come from mesoscale cyclones [like polar lows that are known to form in the Labrador Sea (Moore et al. 1996; Rasmussen et al. 1996)], synoptic scale cyclones (Reed et al. 1988; Huo et al. 1996a,b; Moore et al. 2001), and cold air outbreaks from eastern Canada (Renfrew and Moore 1999; Pagowski and Moore 2001). The OWS Bravo data analysis gives insight into the temporal variability of this strong surface forcing.

Past studies of atmospheric and oceanographic fields in the North Atlantic have dealt with aggregated analyses, such as monthly means (Lazier 1980; Smith and Dobson 1984). The disadvantage of these studies is that variability on timescales shorter than a month is not
resolved. Not only are the high frequencies not resolved, but the aggregated values are much lower than those associated with the individual events. Gascard and Clarke (1983) observed a convective event in the Labrador Sea during March 1976. During the event, the observed wind speeds were on the order of 60 knots with air–sea temperature differences of 20°C (A. Clarke 1995, personal communication). This would translate to an instantaneous sensible heat flux of about 750 W m\(^{-2}\). In contrast, Smith and Dobson (1984) find monthly mean wintertime sensible heat fluxes around 100 W m\(^{-2}\). Our study shows that there is, in fact, high-frequency forcing present greatly in excess of monthly mean values and, furthermore, that these events may be the triggering mechanism that initiates rapid downwelling in the ocean.

Because the density of the water is determined by both its salinity and temperature, it is the total buoyancy flux\(^1\) that one needs to look at. The main components of the buoyancy flux are the sensible and latent heat fluxes, evaporation, and precipitation. Precipitation has generally been ignored in the past, on claims that the sensible and latent heat fluxes were much larger terms in the buoyancy flux calculations or that little was known about its characteristics (Marshall and Schott 1999). However, at the sea surface temperature of the Labrador Sea, the precipitation can provide a positive buoyancy flux as large as the negative buoyancy flux provided by the heat fluxes (Moore et al. 2001).

Using the OWS Bravo dataset, we calculate the buoyancy flux for a period of 24 years. OWS Bravo’s location in the middle of the Labrador Sea (56°N, 51°W), enables us to obtain a sense of the temporal variability as well as the magnitude of the atmospheric forcing in the Labrador Sea from this analysis. In section 2, we discuss the dataset and describe how we calculated the buoyancy flux. Section 3 contains our results, and section 4 is a summary.

2. The data

OWS Bravo was located at 56°N, 51°W in the center of the Labrador Sea, and was active from 1945 to 1974. More than one ship served at the Bravo location, and given that the Labrador Sea is a harsh place to be in,

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\(^1\) The term buoyancy flux is a misnomer since there is no actual flux of buoyancy. Rather, there are fluxes of heat and moisture, and it is convenient to express the result of both by the fictitious buoyancy flux.
the fact that this dataset is available to the scientific community is itself a testament to the people that collected the data. Figure 1 shows a map of the Labrador Sea, with the Bravo location indicated.

Our analysis is based on the 24 years from 1949 to 1973, as that provided the most continuous length of surface records. Data points greater than three standard deviations from the long-term mean (less than 0.4% of the data) were removed and replaced by interpolation. Missing data were also interpolated to get evenly spaced 3-hourly time series for each field (except for precipitation, which was set to zero for missing data). On average, about 2% of the data points for each field had to be found by interpolation.

Both salinity and temperature (and also pressure, but here we are looking just at surface interactions) determine the density of seawater, and thus the total buoyancy flux of the surface waters is the appropriate quantity to study. The buoyancy flux can be written as the sum of a thermal forcing contribution and a saline forcing contribution (after Gill 1982):

\[
B = \frac{g\alpha}{c_w} (Q_s + Q_l + Q_b + Q_i + Q_p) + g\beta (P - E),
\]

where \( \alpha = \rho^{-1} \partial \rho / \partial T \) and \( \beta = \rho^{-1} \partial \rho / \partial S \) are the thermal and saline expansion coefficients, \( S \) is the surface salinity, \( T \) is the sea surface temperature, \( Q_s \) is the sensible heat flux, \( Q_l \) is the latent heat flux, \( Q_b \) is the outgoing longwave radiation, \( Q_i \) is the incoming solar radiation at the sea surface, \( Q_p \) is the heat flux contribution from the precipitation, \( E = Q_l / L_v \) is the evaporation, \( P \) is the precipitation, \( c_w \) is the specific heat capacity of water, and \( L_v \) is the latent heat of vaporization. Our positive coordinate is downward; a positive buoyancy flux implies that the buoyancy of the surface waters is increasing.

Both precipitation and evaporation enter twice into the buoyancy flux formulation, once as a thermal contribution and then again as a saline contribution. For precipitation, the thermal contribution comes from the melting of snow and the thermal equilibration of the precipitation and the sea surface, and the saline contribution comes from the addition of freshwater. For evaporation, the thermal contribution comes from the transfer of latent heat (in fact, evaporation is a rescaled form of the latent heat) and the saline contribution comes from the removal of freshwater.
For the sensible \( (Q_s) \) and latent \( (Q_l) \) heat fluxes, bulk formulæ from Smith and Dobson (1984) were used:

\[
Q_s = -
\rho C_p \left( T_s - T_{10} \right) U_{19.5}
\]

\[
Q_l = -L_v C_v \left( \rho_s - \rho_v \right) U_{19.5},
\]

where \( \rho \) is the density of air, \( C_p \) is the specific heat of air at constant pressure, \( C_v \) is the sensible heat flux coefficient, \( T_s \) is the sea surface temperature, \( T_{10} \) is the air temperature (at 10-m height), \( U_{19.5} \) is the wind speed (at 19.5-m height), \( C_v = 1.2C_i \) is the latent heat flux coefficient, \( L_v \) is the latent heat of vaporization, \( \rho_s \) is the density of water vapor at 10 m, and \( \rho_v \) is the density of water vapor at saturation at the sea surface temperature.

The anemometer on a typical medium sized ship is at a height of 19.5 m, and the typical bridge deck level where temperature measurements are taken is at a height of 10 m (S. Smith 1996, personal communication). Since more than one ship served at a given OWS, we decided to use these generic heights, and did not attempt to reconstruct the exact heights on individual ships. For the sensible and latent heat flux coefficients, we used values given by Smith and Dobson (1984). The minus signs in the equations for \( Q_s \) and \( Q_l \) (and for \( Q_t \) below) establish the direction of heat exchange as going from the atmosphere to the ocean. Thus, a positive \( Q_s \), for instance, would act to warm up the ocean and hence increase its buoyancy.

For the shortwave \( (Q_s) \) and longwave \( (Q_l) \) radiative heat fluxes, empirical formulæ from Gill (1982) were used:

\[
Q_s = Q_{0s}(1 - \alpha_s)(1 - 0.7n_s)
\]

\[
Q_l = -0.985 \sigma T_s^4(0.39 - 0.5e_u^{1/2})(1 - 0.6n_v^2),
\]

where \( Q_{0s} \) is the net downward flux of solar radiation above the surface in cloudless conditions, \( \alpha_s \) is the surface albedo, the factors with \( n \) (fractional cloud cover) are cloud correction factors, and the term with \( e_u \) (vapor pressure of water) is a correction for back radiation.

The OWSs did not measure precipitation directly. Precipitation is a difficult quantity to measure. It is doubly hard on board a ship. On land, problems such as evaporation from the gauge, wind speed (as wind speed increases, gauge catch decreases exponentially), blowing snow, and splashing into or out of the gauge need to be addressed. On board a ship, there are added problems to deal with, including swaying (the roll and pitch of the vessel), ocean spray, and the aerodynamic effect of the ship itself (Legates and Willmott 1990). The sparse coverage over the ocean of observing stations and the difficulties involved in making such measurements might make satellite inference techniques attractive. However, the indirect relationship between what is observed by a satellite and the precipitation at the surface can result in large biases and errors that are difficult to quantify without ground truth (Xie and Arkin 1997). Unfortunately, the high-latitude oceans are regions where such ground truth is difficult to come by.

Rather than measure precipitation directly, the OWSs recorded present weather codes every three hours. A present weather code is a number from 00 to 99 that is a qualitative description of the current weather, subjectively assessed by the observer. Tucker (1961) developed a method of translating these qualitative codes to quantitative precipitation estimates by correlating the observed present weather codes to the measured precipitation at several coastal stations around the British Isles. He observed that there was a better correlation between his estimates and the actual precipitation for stations that were not affected by orography (which is the case for an OWS). He found about a 25% uncertainty in his estimates. Dorman and Bourke (1978) applied Tucker’s method to a wider geographical distribution of coastal stations and found a systematic error due to variations in air temperature (arising from the increased moisture capacity of the air at higher temperatures). We employ their correction to Tucker’s algorithm in our estimate of the precipitation at Bravo. While it may seem questionable to go from qualitative estimates to quantitative estimates (Tucker 1961), it should be remembered that this long time series of observations is the best we have available for that region. Furthermore, unlike most ships, the OWSs carried skilled observers. Therefore, it is likely that the present weather codes are consistent.

Precipitation contributes freshwater to and removes heat from the surface waters. It precipitates in one form or another about a third of the time at Bravo based on our data. This is confirmed by observations done on the 1997 cruise of the R/V *Knorr* (P. Guest 1997, personal communication). We assume that if the air temperature is equal to or below 0°C, then any precipitation is in snow form. According to our analysis, at Bravo a little less than 40% of the precipitation was in snow form. We furthermore assume that the temperature of the precipitation is 0°C. At Bravo’s latitude, the precipitation most likely forms in an ice/snow state and then possibly undergoes a phase transition as it falls into warmer temperatures. Whether it changes state to a raindrop or remains as snow as it falls, the precipitation temperature lags behind the air temperature by about a kilometer (R. List 1996, personal communication). Thus, it seems more reasonable to assume that the precipitation temperature is near the phase transition temperature rather than at the ambient air temperature at 10-m altitude. It is quite likely that the actual temperature of the precipitation is slightly different from 0°C; however, the amount we are potentially neglecting is very small. If we had a precipitation rate of 2 mm h\(^{-1}\) and the precipitation temperature was in reality −3°C, then the heat flux contribution coming from raising that temperature to 0°C is on the order of 5 W m\(^{-2}\) only. Moreover, the largest contributor by far to the precipitation heat flux
comes from the melting of snow into water, not from raising or lowering the precipitation temperature. When it is raining, the precipitation heat flux ($Q_p$) is given by

$$Q_p = -\rho_w P c_w (T - 0^\circ C),$$

which amounts to raising the temperature of the rainwater to the sea surface temperature. When it is snowing, the precipitation heat flux becomes

$$Q_p = -\rho_w P (l_f + c_w (T - 0^\circ C)),$$

which involves melting the snow (which is assumed to be at $0^\circ$C already), and then raising the temperature of the melted water to $T_s$. Here $\rho_w$ is the density of freshwater, $P_r$ is the precipitation rate, $c_w$ is the specific heat of water, and $l_f$ is the latent heat of freezing. At the Bravo location, the largest contribution comes from melting the snow—up to 250 W m$^{-2}$; the other terms sum to a maximum of 25 W m$^{-2}$. The 250 W m$^{-2}$ translates to about 2.5 mm h$^{-1}$ and is a maximum value from the winters; the mean is 20 W m$^{-2}$ and is equivalent to 0.2 mm h$^{-1}$.

There are several different global precipitation climatologies available, including those that are blends of some combination of gauge observations, satellite observations [infrared, outgoing longwave radiation, mi-
Fig. 5. Composites of different fields, ±120 h of a pressure minimum for (a) moisture, (b) heat fluxes, and (c) buoyancy flux. The different components are indicated in the figure legends.

crowave sounding unit, and microwave scattering and emission from the Special Sensor Microwave/Imager (SSM/I), and numerical model forecasts (Xie and Arkin 1997). In order to assess the reasonableness of our precipitation estimate at Bravo, we compared it to the Legates/MSU (Microwave Sounding Unit) precipitation climatology. This climatology was chosen because it contained global precipitation data over the World Ocean. It is a blend of two datasets and has a spatial resolution of 2.5° by 2.5°. Over land, the historical record of rain gauge measurements (Legates and Willmott 1990) is used, and over ocean, the 1979 to 1991 MSU precipitation estimates (Spencer 1993) are used.

The solid line in Fig. 2 shows the monthly mean precipitation from the Bravo (1949–1972) dataset. The dashed lines show the monthly mean precipitation from the Legates/MSU climatology. While the general form of the annual cycle is the same for both datasets, the Legates/MSU values are systematically lower. The difference is most pronounced during the winter months. Figure 2 also shows (with a dotted line) the UWM COADS (University of Wisconsin—Milwaukee/Comprehensive Ocean—Atmosphere Data Set) surface marine climatology (Woodruff 1987; Da Silva et al. 1995). This dataset parameterizes precipitation in a similar way to this study, using present weather codes. As expected, their climatology agrees well with our annual cycle and falls within one standard deviation for all months.

A few things can account for the difference between our dataset and the Legates/MSU climatology. First,
Spencer (1993) found that at high latitudes the satellite estimates needed an air mass temperature correction that resulted in a 25% reduction in the values. However, the MSU is calibrated with low elevation island and coastal rain gauges, most of which lie within 30° of the equator, with only a few stations as far away as 60° latitudinally (Spencer 1993). Hence, the applicability of this correction is debatable because there are few rain gauge data available at high latitudes (GHCC 1998). In fact, Spencer (1993) states that the best quantitative agreement between the gauges and the satellite measurements was found in the Tropics. As a Global Hydrology and Climate Center (GHCC 1998) report suggests, the MSU rainfall diagnoses might be biased low at high latitudes, especially in the winter months.

Second, MSU precipitation estimates are based on the radiometric warming of the brightness temperature caused by cloud water and rainwater. However, when ice is present, volume scattering off the ice can depress the brightness temperature (Spencer 1993), causing the precipitation to be underestimated. According to our results, approximately 40% of the precipitation at Bravo was in snow or ice form. In fact, the streamer clouds in Fig. 6c generally have precipitation in ice or snow form (P. Guest 1997, personal communication), and the MSU might underestimate the actual precipitation in those regions. Given these uncertainties in both the Legates/MSU dataset and the present weather code method of Tucker as modified by Dorman and Bourke, we assert that the two climatologies are broadly consistent.

3. Results and discussion

Figure 3 shows the buoyancy flux for 1951, with the monthly means superimposed. The year was chosen arbitrarily. The figure illustrates that, by taking monthly means, much variability has been smoothed over. It is important to point out that the time series is not noisy because of random fluctuations. Rather, the peaks represent actual physical events, such as the passage of a mesoscale or synoptic-scale cyclone. That these short but intense flux events have been masked out in the monthly mean analyses of previous workers becomes more significant given that the North Atlantic is one of the few areas in the world where deep water is known to form. Furthermore, as we will show below, the timescale of this variability is similar to the convective timescale in the ocean.
FIG. 6. (Continued)
When the standard deviation is of the same order of magnitude as the mean, the mean does not characterize the time series well. Examining Fig. 3, we see that this is the case on a month-by-month basis. In fact, the average buoyancy flux for this particular year is \(-5.8 \times 10^{-9}\) N \(m^{-2}\) s\(^{-1}\), while the standard deviation is \(59.0 \times 10^{-9}\) N \(m^{-2}\) s\(^{-1}\), an order of magnitude larger. There is also a change in the character of the variability between summer and winter. The closely spaced spikes during the summer months are due to the diurnal nature of the shortwave radiation. During the winter months, the spikes are farther apart (typically 3–4 days) and are, as we shall see, due to the passage of storms. The large positive peaks are due mainly to precipitation events and occur throughout the year (during the summer, the shortwave fluxes contribute a positive diurnal flux as well). The large negative peaks can be attributed to the heat losses associated with the passage of storms that, as expected, occur mainly during the winter months.

The importance of precipitation as a contributor to the buoyancy flux in the Labrador Sea has been underestimated in the past. Precipitation is a source of positive buoyancy. As a low tracks past a point, such as the Bravo location, precipitation will occur ahead of the low, and heat fluxes will occur behind (this will be visible in our composites below). If the precipitation can make the surface waters sufficiently buoyant, then the subsequent negative buoyancy contribution coming from the heat fluxes will have to work that much harder to erode through the surface freshwater cap before reaching the weakly stratified waters below. However, this is of consequence only if the precipitation can contribute a sufficiently large positive buoyancy to the surface waters.

It turns out that this is the case in the northern latitudes. Due to the nonlinearity in the equation of state for seawater, at cold temperatures, changes in salinity are more efficient at changing the density of the seawater than changes in temperature (Moore et al. 2001). The thermal expansion coefficient \(\alpha\) is strongly dependent on the sea surface temperature, whereas the saline expansion coefficient \(\beta\) is not. Thus, when you have low temperatures, the relative contributions of the two terms in Eq. (1) change; the effect of the saline contribution is increased, while the effect of the thermal contribution is diminished.

Figure 4 shows the annual cycle of monthly mean values over the 24 years for different variables, with one standard deviation indicated by the error bars. The large standard deviations, especially during the winter months, show that there are considerable interannual variations. The upper panel shows the thermal forcing contribution [the first term in Eq. (1)], while the middle panel shows the saline forcing contribution (the second term). The vertical axes are the same, making for easy comparison of their relative importance. The thermal forcing term peaks in the summer, while the saline forcing peaks in the winter. During winter, the contributions
from the two components are of opposite sign but of comparable magnitude due to the temperature dependence of $\alpha$ mentioned above.

The third panel of Figure 4 shows the buoyancy flux with and without precipitation. It is clear that without precipitation the buoyancy loss from the ocean will be overestimated. The integrated buoyancy loss from the ocean over the year is about 3700 N m$^{-2}$; without the inclusion of precipitation, that number is about 7700 N m$^{-2}$, more than twice as large. Hence, if we do not include the effects of precipitation, there will be a systematic overestimation of the amount of buoyancy removed from the upper layers of the ocean. In addition, the restratification that takes place during the summer months will also be underestimated. Note also that the buoyancy flux drops from its peak values faster than it takes to rise. This leads to values of buoyancy flux in October that are not too far from wintertime (Dec, Jan, Feb) values. Thus the preconditioning may begin as early as October, well before the traditional winter months.

In order to see what a typical event looks like, we make use of composites. A composite will not tell us what any given event will be like, but it will give a sense of how the events behave in general. As we are interested in identifying a pattern associated with the passage of low pressure systems, we created composites of various variables by triggering on a pressure minimum and then looking at values out to 120 hours on either side of the pressure minimum. If one assumes uniform translation in space, then the time axis on the composite can be understood as indicating the spatial structure of the air–sea fluxes during the passage of a low pressure system. The composite can therefore also be interpreted as a snapshot of an event at a given instance in time. If two pressure minima were too close to each other, both were discarded on the assumption that one would contaminate the other in the compositing. All the events that passed our criteria of minimum pressure level and separation (409 events over 24 years) were included in the compositing, though the majority
of them occurred during the fall and winter. Figure 5 shows the composite for the moisture fields (mm h$^{-1}$) and the heat flux fields (W m$^{-2}$) and the buoyancy flux (N m$^{-2}$ s$^{-1}$). Zero hours on the x axis, marked by a vertical line, indicates when the low pressure passed Bravo’s location.

The top panel of Figure 5 shows the precipitation rate and evaporation rate. It is interesting to note that there seems to be a continuous, though small, background precipitation rate. As expected, the precipitation peaks ahead of the low pressure (i.e., to the left of the zero line). The middle panel shows the sensible, latent, long-wave, and shortwave heat fluxes. The compositied curve for the precipitation heat flux was close to zero and
Fig. 11. Sensible heat flux for (a) Feb 1968 (NAO − year); (b) Jan 1972 (NAO + year).

hence has not been displayed. Note that, as before, a positive flux is directed into the ocean. The longwave and shortwave fluxes are largely a constant offset, though the shortwave flux shows a strong diurnal signal, and the magnitudes of both fields decrease slightly just before the low passes. This is due to the increase in cloud cover at that time. The sensible and latent heat fluxes are negative, meaning that the ocean is losing heat in the composited sense. Before the low passes, these two fluxes decrease in magnitude, but then rapidly increase (in magnitude) as the low passes. The increase in magnitude (or equivalently, the decrease in an absolute sense) is indicative of an increased flux from the ocean to the atmosphere.

Based on when the heat fluxes and buoyancy flux return to their background values, we can take −42 h to +69 h as the time period over which the main part of the weather system affects Bravo’s location (the hours are rounded to the nearest 3 h). Taking into account the 409 events that we used, this represents approximately 20% of the total available data, yet this subset contributed 35% of the total sensible heat removed, 30% of the total latent heat removed, 34% of the total rainfall, and 49% of the total buoyancy removed at Bravo’s location.

The composite clearly shows that at Bravo, during the passage of a low pressure system, an observer would first see the precipitation falling concurrently with a decrease in the magnitude of the fluxes. This would be followed by the minimum in the pressure, and finally the elevated (negative) heat fluxes, before the background values set in again. This is consistent with what would be expected to occur during the passage of an extratropical cyclone, as discussed by Moore et al. (2001). Figure 6 shows three instances in the life cycle of an extratropical cyclone over an approximate 24-h period. The Bravo location is indicated by the asterisk. In panel (a), the warm front is over Bravo and an observer would see the associated precipitation. In panel (b), the warm front has moved past, and the low pressure is now over Bravo. Finally, in panel (c), the cyclone has moved past, and the resulting wind field behind it is drawing cold dry air over the region, causing the streamers (elongated cloud structures) to form, indicative of the elevated heat fluxes occurring there.

Returning to Fig. 5c, the buoyancy flux composite is shown by the solid line, and the buoyancy flux composite without precipitation by the dashed line. There are several things to note here. First, if precipitation is not included, then the buoyancy flux is underestimated everywhere (i.e., the buoyancy loss from the ocean will be overestimated). The difference is most prominent just before the low passes (when the precipitation maximum occurs). It is interesting that the buoyancy flux changes sign as the low passes, indicative of the deposition of a freshwater cap that subsequent negative buoyancy fluxes have to penetrate in order for deep convection to occur. A summation under the curve from −42 h to +69 h found that without the precipitation contribution, the buoyancy loss over that time period was −15.5 N m⁻², an overestimate of 76% compared to the actual buoyancy loss of −8.8 N m⁻². Also note that the timescale for the change in buoyancy flux is about a day, which is of the same order as the timescale for the convective plumes occurring in the ocean.²

The peak values of the different variables in Fig. 5 are suppressed due to compositing. There are two reasons why this is so. The first is because we are averaging events with different maximum and minimum values. The other reason is because of the difference in phasing between the individual events. A histogram of the peak values of the actual events used in the compositing shows, for instance, that the precipitation rate ranges up to about 3 mm h⁻¹, whereas in the composite, it only reaches slightly higher than 0.5 mm h⁻¹. The maximum sensible and latent heat flux magnitudes of the actual events are also about two to three times larger than the composited values. As with the other variables, the maximum and minimum peaks of the buoyancy flux are also suppressed. Thus, the signal of the buoyancy flux changing sign as a low passes is, in fact, stronger than the

2 Convective plumes are the mixing agents that homogenize the fluid within the chimney. They have spatial scales of about one kilometer and temporal scales of about one day (Marshall and Schott 1999).
Fig. 12. Winter (DJF) autocorrelations for (a) sensible heat flux and (b) precipitation. The dashed lines indicate the 95% confidence level.

The diurnal and annual signals (and their subharmonics) are the most prominent features in the buoyancy flux power spectrum (Fig. 7). However, there is some suggestion of power around 12 days, which may be due to changes in the synoptic flow regime (Lab Sea Group 1998). There might also be some power around 10 yr, but we cannot associate much confidence to this since our time series is only 24 years long. However, there is power from about 2 to 5 yr. This may include influences from the North Atlantic Oscillation (NAO), the quasi-biannual oscillation, and the El Niño–Southern Oscillation.

The NAO, an oscillation in the atmospheric mass field
over the North Atlantic, possibly exerts the strongest influence of these, primarily because of its proximity to the Labrador Sea. The NAO index (Rogers 1984) gives a convenient way of describing the magnitude and phase of the oscillation. Positive values of the index are characterized by an intense Icelandic low and a strong ridge over the Azores, while negative values imply anomalies of the opposite sign. In the literature, there is much focus on the NAO as a seasonal, primarily winter, phenomenon (van Loon and Rogers 1978; Rogers 1984; Hurrell 1995; Lin and Derome 1998; Kwok and Rothrock 1999). We have, however, chosen to look at higher-frequency variability in the mass field over the North Atlantic, specifically on a monthly timescale. Using the Rogers (1984) method, we have calculated the monthly mean NAO index from the surface fields in the NCEP reanalysis (Kalnay et al. 1996).

Figure 8 shows scatterplots of the December NAO index values versus the January and February values in the top and bottom panels respectively. Although there are winters in which there is a positive correlation, the general picture is one of little or no correlation in the index between months in the same winter. The coefficients of determination \( r^2 \) are \( 1 \times 10^{-5} \) and \( 2 \times 10^{-2} \), respectively, for the top and bottom panels. Hence, if the Icelandic low is anomalously low in December, it does not guarantee that the following February it will also be anomalously low. The slight correlation we see between December and February is, in fact, a negative correlation. A similar situation exists for adjacent
months within a given winter. Granted, a month is still an arbitrary separation point, but nonetheless, this figure shows the usefulness of looking at shorter timescale NAO index values. The winter of 1997 provides an excellent example of the dramatic changes in the mass field over the North Atlantic that can occur within a given winter. In this instance, January was a month with a low NAO index, while February was a month with a very high index (Lab Sea Group 1998).

Figure 9 shows the correlation scattergraph between the NAO index and the sensible heat flux at OWS Bravo. The upper panel shows the scattergraph for all the months. Since we have 24 years of data, there are 288 points. The coefficient of determination for the upper panel is 0.11. The bottom panel has a similar plot, but with the data restricted just to the winter months. The solid line indicates a straight line fit in a least squares sense. In the upper panel, the data points for the nonwinter months are clustered mainly around 0 W m\(^{-2}\) but at all values of the NAO index. However, the NAO index is normalized by the standard deviation, the index gives equal weight to the summer and winter months; otherwise, the NAO pattern is most prominent during the northern wintertime. Hence, when we consider the correlation between the NAO index and the sensible heat flux for the winter months, we find a stronger correlation.

It should be remembered that the NAO is not the oscillation of a “low” that one would find on any given day. Instead, the two patterns of the NAO are the result of averaging the passage of several storms over a course of time. The two states of the NAO are illustrated in Fig. 10 using monthly averaged surface data from the NCEP reanalysis. The location of OWS Bravo is marked with an asterisk. Arrows are drawn to indicate wind speed and direction. The upper panel shows the monthly mean sea level pressure for February 1968, when the NAO was in its negative phase. The lower panel shows the same for January 1972, when the NAO was in its positive phase. In the positive phase, the pattern of airflow is conducive to drawing a large amount of cold dry air from the Canadian Arctic and over the Labrador Sea. When this cold dry air crosses over the ice edge and becomes exposed to the relatively warmer ocean waters, large heat fluxes occur. During the negative phase of the NAO, however, the airflow pattern hinders this flow over the Labrador Sea.

Hence, one would expect larger and more frequent fluxes during a positive NAO year compared to a negative NAO year. Indeed, this is the case. Figure 11 shows the sensible heat flux at Bravo for the same two months as in Fig. 10. In the upper panel, for the negative NAO month, the heat flux is relatively modest, with only one event below −200 W m\(^{-2}\). The mean for this month was −58 W m\(^{-2}\) and the standard deviation was 55 W m\(^{-2}\). In contrast, for the positive NAO month (lower panel) there are several events close to or below −200 W m\(^{-2}\), with the largest event reaching below −300 W m\(^{-2}\). The mean for this month (−133 W m\(^{-2}\)) is more than twice that for the negative NAO month, and the standard deviation about the monthly mean is also higher at 74 W m\(^{-2}\). Thus, during the positive NAO month the heat fluxes are larger in magnitude, and they occur with greater frequency.

To investigate the intraannual variations in more detail, we looked at autocorrelations and power spectra for only the winter months (DJF). Twenty-three winters were available, so we were able to average the results for the 23 winters, thereby reducing the variance. In addition, a 1-day moving average was performed on the data before the autocorrelations were taken in order to remove the diurnal signal. Figure 12 shows the autocorrelations for the sensible heat flux and the precipitation. The dashed lines show the 95% confidence levels. The sensible heat flux has a gradual drop-off, lasting 5 to 6 days. There are signs of an anticorrelation between 10 and 20 days, before the heat flux reverts back to being uncorrelated about 25 days later. In contrast, the precipitation has a sharp drop-off lasting a day or two, before becoming uncorrelated. This indicates that if there are strong (or weak) fluxes, then that will continue for a few subsequent days, whereas, if it rains on a given day, then it is as likely or unlikely to rain the following day.

The power spectra for the wintertime buoyancy flux (Fig. 13a) has the expected diurnal signal and its subharmonics, but it shows a peak in the vicinity of 12 days, similar to that visible in Fig. 7. This is likely due to changes in the flow regime in the North Atlantic. The power in the 30-day and lower frequency range is due to lingering remnants of the annual signal after we detrended the winter subseries. The power between 12 and 1 days is from cyclonic activity.

The lower panel in Figure 13 has the autocorrelation for the wintertime buoyancy flux. As in the case of the sensible heat flux, the autocorrelation graph drops off slowly, and it flattens from about 5 to 10 days. A flattening in an autocorrelation graph indicates persistence. The anticorrelation between 10 and 20 days establishes the phase of the signal (change in flow regime) that was visible in the power spectrum.

4. Summary

In summary, high frequency variability is present in the buoyancy flux, and is masked when one does aggregated analyses as previous investigators have done. It is important to not disregard these high frequency fluctuations, because they are not random noise. Instead, they are events caused by the passage of weather systems, the nature of the variability of which may play an important role in deep water formation.

For instance, how would a realistic temporally varying forcing affect the densification of the upper layers of the ocean as compared to a constant time-independent
forcing (as is generally used in deep convection models)? Our analysis shows that the forcing is definitely not constant; the mean buoyancy flux over a winter at Bravo is about $-3.7 \times 10^{-3}$ N m$^{-2}$ s$^{-1}$, whereas the actual buoyancy flux ranges from about $-1.7 \times 10^{-4}$ to $1.6 \times 10^{-4}$ N m$^{-2}$ s$^{-1}$, a range of $3.3 \times 10^{-4}$ N m$^{-2}$ s$^{-1}$. Hence, it may be important to consider a time-dependent forcing field when driving deep convection models.

As the composites showed, the timescale for the change in the buoyancy flux (order of a day) was of the same order as the timescale for the convective plumes in the ocean. Although there is no a priori reason for this to be the case, the fact that it is suggests that changes in buoyancy due to atmospheric processes may modulate convective activity in the ocean.

Due to the nonlinearity in the equation of state for water, the thermal expansion coefficient $\alpha$ is strongly dependent on the sea surface temperature, while the saline expansion coefficient $\beta$ is not. This has the consequence that at cold temperatures, the effect of the thermal forcing contribution is lessened with respect to the saline forcing contribution, thereby allowing precipitation to play a non-insignificant role in the buoyancy flux. If precipitation is not included in the calculations, the amount of buoyancy flux removed is overestimated. Additionally, the positive buoyancy contributed by the precipitation allows the buoyancy flux to become temporarily positive during cyclone events, indicating the formation of a freshwater cap that the negative buoyancy contributions from the heat fluxes have to erode through before making the water dense enough to sink. Further, the annual cycle in the fields indicates that large buoyancy fluxes can occur in the autumn months as well and not just in the winter months.

Spectral analysis of the buoyancy flux shows that power exists between two and five years. This is likely due to modulation by the North Atlantic Oscillation. A strong negative correlation exists between the sensible heat flux and the NAO index. In other words, when the NAO is in its negative phase (with the climatological low situated over Iceland), the wind field is such that it enhances the flow of cold dry air from the Canadian Arctic and over the warmer Labrador Sea, resulting in large heat fluxes. This correlation would explain the observed link between the NAO and deep ocean convection in the Labrador Sea (Lab Sea Group 1998).

There is also evidence of a cycle between 10 and 20 days resulting from changes in the flow regime. The autocorrelations establish the phase of the signal as negative; that is, if the fluxes are positive today, then 10 to 20 days later they will be negative, and vice versa. Weather systems were in the vicinity of Bravo for about 20% of the time, yet in total they contributed almost 50% of the buoyancy removed from Bravo’s location. This further illustrates the importance of the high frequency forcing for deep convection.

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