# 1 Introduction

This is a proposal to study 'Eighteen Degree Water' (EDW), the subtropical mode water of the North Atlantic. EDW is but one example of a pervasive tendency for mode waters to form adjacent to strong baroclinic fronts in all the world's oceans. EDW is a canonical example of a subtropical mode water, all of which are found in regions of significant air-sea exchange. EDW is created in the winter just south of the Gulf Stream, by convection in the presence of strong shear, with competing effects of vertical/lateral mixing and advection/stirring colluding to set its properties (e.g. Worthington 1959, 1976; Schroeder et al, 1959; Ebbesmeyer and Lindstrom, 1986). This proposal stems from two years of CLIVAR planning (with advice and support of both the Atlantic and US CLIVAR committees) to develop an experiment to attack a key process that is poorly understood and poorly represented in ocean climate models - i.e. the treatment of convection, eddy and mixing processes in setting properties of subtropical mode waters, the associated air-sea interaction, and the exchange of fluid between the mixed layer and the upper ocean.

EDW is in an historically well-observed region and yet major questions remain about the relative importance of the many processes that produce this major feature of the upper ocean's stratification. There is a considerable disconnect between the best available estimates of EDW formation and dissipation rates, with up to 15-20 Sv of formation estimated through indirect methods from air-sea fluxes (Speer and Tziperman 1992) and only about 5 Sv inferred to be injected seasonally into the subtropical gyre (Kwon and Riser 2004,a,b).

EDW formation includes a mix of processes: (1) cross-Gulf Stream fluxes, (2) transformation processes within the Gulf Stream, (3) impact of the recirculation region on stratification, (4) diapycnal mixing and subduction, and (5) buoyancy loss. A conclusive assessment of their relative roles has not been achieved because no comprehensive observations have been collected in the Gulf Stream region (or any other subtropical mode water region), through a full winter with the required eddy and finestructure spatial scales. Formation rates have thus been assessed only indirectly. CLIMODE proposes a full suite of winter measurements and year-round observations to conclusively assess the role of each of these processes, and permit an in situ estimate of the formation rate that can be compared with the indirect estimates.

Destruction of EDW starts with subduction and advection into the western subtropical gyre along with diapycnal mixing and fluxes. These are best assessed through seasonal inventories of the total amount and transport of EDW. Here there are at least observationally-based conclusions because of the larger spatial scales captured using hydrography and profiling floats. Because of the significant difference between the large formation and smaller dissipation rates, and because of demonstrated interannual variability, CLIMODE proposes a study of EDW dissipation at the same time as the formation study, in order to close the EDW budget.

EDW formation is important to ocean circulation and climate modeling insofar as it affects the Gulf Stream system circulation and associated large air-sea fluxes in the western subtropics. Models can capture mode water — see, e.g., Hazeleger and Drifhout, 1998; 1999; Marsh and New, 1996; Paiva and Chassignet 2002. However, uncertainties in the parameterization of the formation/dissipation mechanisms lead to systematic errors in modeled temperature, salinity and sea-surface temperature (SST) in these critical regions of air-sea exchange (Schopf et al 2004). The improved parameterizations of upper ocean mixing that will flow from CLIMODE will benefit many, if not all, classes of ocean models and the coupled climate models that are used in climate change research. We will work closely in this endeavor with the recently formed CLIVAR Climate Process Team (CPT) in upper ocean mixing.

In the following pages, we briefly review past work on EDW, and outline the relationship of CLIMODE to CLIVAR goals and the upper ocean mixing CPT. In section 2 we introduce the water mass transformation framework of Walin (1982), which has been used to guide the organization of our experiment, discuss the key hypotheses that motivate CLIMODE and briefly outline the proposed program elements. Section 3 presents in detail the proposed observational and modeling methods and the connection between them. Section 4 presents the synthesis of observations and models leading to improved estimates of EDW formation and dissipation and how they might be represented in models. Section 5 details programmatic information.

### 1.1 Background and relation to CLIVAR goals

Huge ocean to atmosphere annual-mean heat loss ($> 200\ Wm^{-2}$) occurs over the separated Gulf Stream in the North Atlantic. The region of most intense wintertime ocean heat loss corresponds to an area with relatively warm surface waters that are carried there by the Gulf Stream, Fig.1. Late winter SST's fall to approximately 18°C as water parcels move east under this cooling. The associated buoyancy loss from the ocean is believed to trigger convection to form what is known as EDW — Worthington (1959;1976) — the North Atlantic Subtropical Mode Water (STMW). The wedge of weakly stratified water spanning temperatures between about 17° and 19°C characteristic of a mode water are clearly evident in the Gulf Stream section shown in Fig.2.

It is believed that EDW is formed on the northern rim of the subtropical gyre in wintertime convection. The presence of EDW in the northwestern portion of the subtropical Atlantic is reflected in a substantial, 1000 km diameter 'bowl' of low potential vorticity (PV) fluid at depths of 200-500m found just to the south of the separated Gulf Stream (Lozier et al, 1995). This is the principal mode water of the subtropical gyre. Since its initial identification as a mode water having a temperature of around 18 °C, actual temperatures and salinities of STMW have varied somewhat over time. Variations in the hydrographic record at Bermuda (marked by the '×' in Fig.1, left) are clearly linked to interannual changes in air-sea exchange reflecting forcing by the North Atlantic Oscillation (NAO) — see Talley and Raymer, 1982; Talley, 1996; Joyce et al, 2000 and Kwon and Riser 2004b.

The region of EDW formation is particularly relevant to CLIVAR goals because, first, the annual mean ocean to atmosphere heat flux over the EDW formation region might be crucial for the maintenance of the Atlantic Storm track (Hoskins and Valdes, 1990). EDW, as being primarily the result of the associated ocean surface cooling, is a natural focus when thinking about processes able to sustain the annual mean heating rate of the atmosphere as large as 200 $\mathrm{Wm}^{-2}$. Second, EDW and the associated Gulf Stream recirculation and thermal structure is the key region where oceanic timescales can imprint on the atmosphere. Seasonal to interannual timescales are introduced by the thermal inertia of the ocean mixed layer/EDW layer system, whose evolution through the annual cycle is strongly connected to the re-emergence of SST anomalies from winter to winter (Alexander and Deser, 1995; Alexander and Penland, 1996; de Coëtlogon and Frankignoul, 2003). Such 'one dimensional' ocean - atmosphere interactions might introduce winter to winter memory to fluctuations in the storm-track and the jetstream. On longer timescales the intensity and path of the Gulf Stream affects air-sea exchange and mode water formation through interannual variations in low-frequency flow as well as lateral eddy heat fluxes — Czaja and Marshall (2001), Dong and Kelly (2004). How exactly such oceanic influences on climate work is a subject of great importance, controversy and subtlety - see the reviews by Robinson, 2000; Kushnir et al, 2002 as well as Marshall et al (2001) and Czaja et al. (2003).

CLIMODE should also be seen as making an important contribution to tying down the basin-scale air-sea heat budget and, by implication, quantifying the meridional transport of heat in the Atlantic basin. Finally CLIMODE provides a focussed observational context for the CPT in upper ocean mixing, a major thrust to improve climate models.

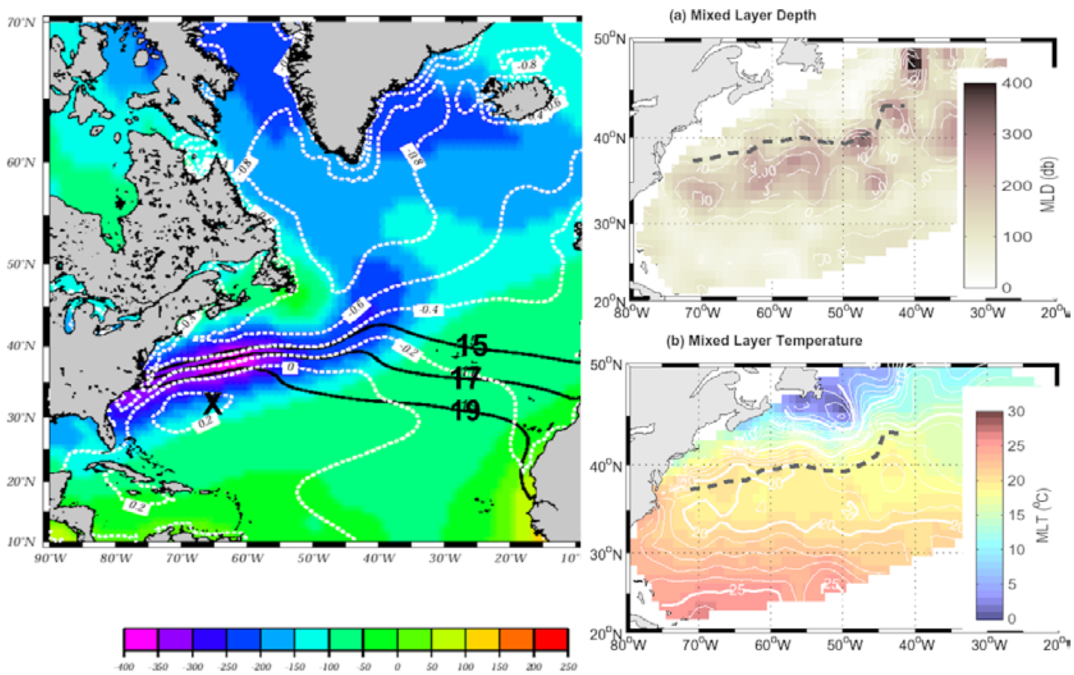


Figure 1: (Left) Wintertime net heat flux (colors - COADS), selected SST outcrops (black lines) and dynamic height field (dotted lines, provided by the ECCO data assimilation scheme using the MIT ocean model). The black cross marks Bermuda. (Right) Average winter (January-March) (top) mixed layer depth and (bottom) mixed layer temperature (CI$= 1^{\circ}C$) from profiling floats, 1997:2002. The dashed line is the average Gulf Stream position for the same period from the Topex/Poseidon altimeter. (Steve Riser, personal communication.)

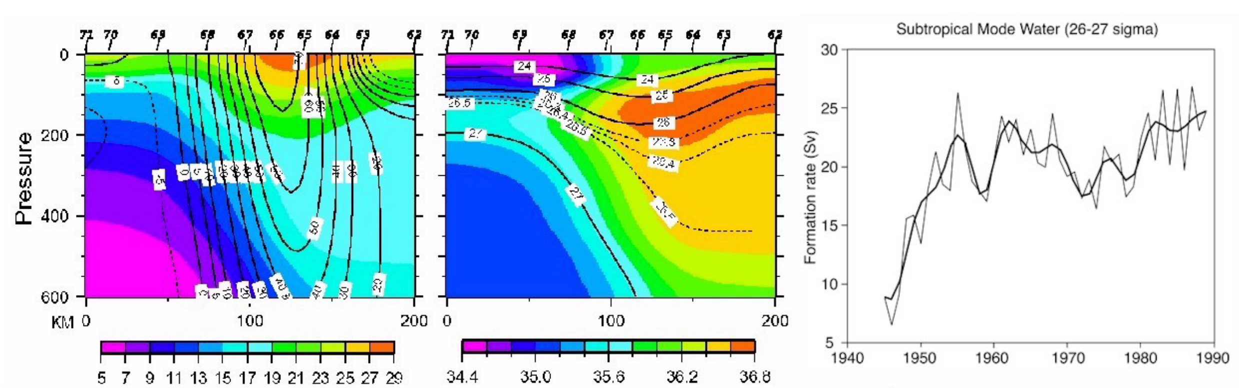


Figure 2: (left) Meridional section looking eastward across the Gulf Stream taken in the summer of 1997 using CTD & LADCP stations (station locations shown) along 66°W. Temperature and zonal velocity: contours in cm/s taken from Joyce et al, 2001a are shown, (middle) salinity and $\sigma_\theta$ contours. (right) Annual average transformation rate $\frac{\partial F}{\partial \sigma}$ (in Sv) by air-sea fluxes in the $\sigma = 26 - 27$ range (thin); smoothed over 3 years (thick). From Speer (2002).

## 2 CLIMODE: the CLIVAR Mode Water Dynamics Experiment

CLIMODE proposes a combined observational and modeling study to observe, diagnose and understand the physics of mode water formation and dissipation. In section 2.1 we begin by presenting the overarching theoretical framework within which the program has been organized, the hypotheses to be tested and the elements of the program which will be synthesized. In section 2.2 we discuss pertinent aspects of the treatment of mixing in the upper boundary layer of ocean models that is the focus of the CLIVAR CPT in upper ocean mixing. In section 2.3, we briefly describe the mix of observations and models that we propose to deploy in CLIMODE.

### 2.1 The Walin framework

EDW is formed very close to or within the Gulf Stream where surface heat loss is large. Based on air-sea flux integrations using Walin's (1982) framework, Speer and Tziperman (1992) estimated a formation rate of 15 to 20 Sv of EDW — see Fig.2 (right). The fundamental problem we are addressing is why this rate is so much larger than the EDW inferred seasonal changes based on profiling floats (e.g Kwon and Riser, 2004a) and implied by thermocline diapycnal mixing rates. Walin's (1982) framework includes all of the bulk upper ocean processes that respond to the air-sea flux, including subduction and diapycnal fluxes. Therefore we choose to discuss the proposed CLIMODE experiment in terms of Walin's precise integral statement of the problem (Fig.3). However, we emphasize that our experiment goes well beyond this framework, in explicitly treating the eddy, mixing, finestructure, volume flux pathways, and local air-sea exchange processes whose details Walin's framework only represents symbolically.

Walin considered the volume budget of an isopycnal layer outcropping at the sea surface, integrated across the ocean from one coast to the other, as sketched in Fig.3:right:

$$\underbrace{\frac{\partial V_\sigma}{\partial t}}_{\text{storage}} = \underbrace{-\frac{\partial A}{\partial \sigma}}_{\text{div of diapycnal vol flux}} + \underbrace{S}_{\text{subduction}} \qquad \text{`formation'} \tag{1}$$

where $V_\sigma$ is the volume of the layer of fluid within a chosen density ($\sigma$) class — for example the shaded region in Fig.3(right) — $S$ is the subduction out of the control volume and $A$ is the diapycnal volume flux through isopycnals. Thus $V_\sigma$ changes if fluid is subducted out of the control volume or if there is a divergence of the diapycnal volume flux $A$ through $\sigma$.

Walin showed that $A$ could be expressed precisely in terms of the diffusive fluxes, '$D$', acting across the surface of the control volume, and air-sea fluxes '$F$', thus (using Garrett et al's (1995) terminology):

$$\underbrace{A}_{\text{diap vol flux}} = \underbrace{F}_{\text{air-sea flux}} - \underbrace{\frac{\partial D}{\partial \sigma}}_{\text{diffusive flux}} \qquad \text{`transformation'} \tag{2}$$

(see Fig.3:right). Here $F = \frac{\partial \widehat{B}}{\partial \sigma}$ where $\widehat{B}$ is the integral of the air-sea buoyancy flux — defined by Eq.(6) — over outcrop windows. The Walin framework has been discussed and applied to water mass transformation by, for example, Speer and Tziperman (1992), Garrett and Tandon (1997); Marshall et al (1999) — see also the review by Large and Nurser (2001).

In an eddying ocean the various terms may be broken down thus (see Tandon and Zahariev, 2001):

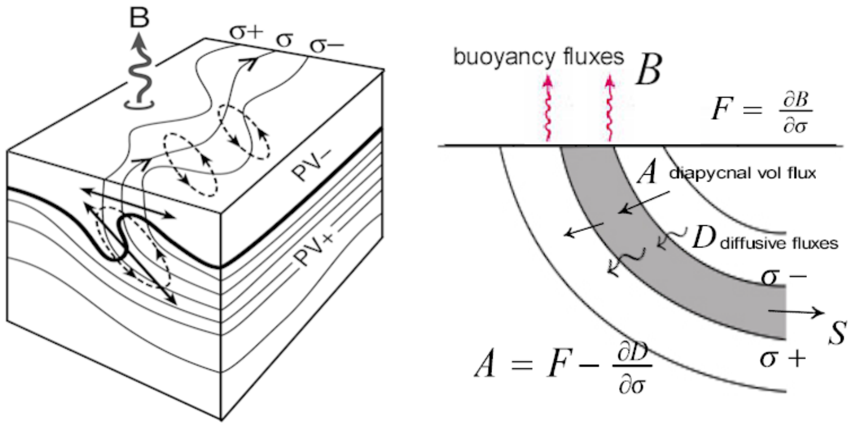


Figure 3: (Left) Schematic showing the interaction of a mixed layer (low PV) and the stratified interior (high PV) in a strong frontal region with outcropping isopycnal surfaces, $\sigma$, undergoing buoyancy loss, $B$. Eddies forming along the front play a central role in controlling horizontal fluxes through the mixed layer and two-way quasi-adiabatic exchange between the mixed layer and the interior. (Right) Application of the formalism due to Walin (1982): lateral diapycnal volume flux, $A$, whose divergence drives subduction, is related to 'diffusive' fluxes, $D$, acting across the boundary of the shaded control volume (which includes small-scale and diapycnal eddy fluxes) and air-sea buoyancy fluxes acting across the upper surface, $F = \frac{\partial B}{\partial \sigma}$, through Eq.(2).

$$\begin{aligned} A &\longrightarrow A_{mean} + A_{eddy} \\ D &\longrightarrow D_{eddy} + D_{int} \\ F &\longrightarrow F_{air-sea} + F_{ml} \end{aligned} \tag{3}$$

where $F_{air-sea}$ is related to the air-sea buoyancy flux, $F_{ml}$ represents entrainment into the mixed layer at its base and $A_{mean}$, $A_{eddy}$ are mean and eddy contributions to the diapycnal volume flux. We draw attention to (i) the partition of $D$ into horizontal eddy mixing terms, $D_{eddy} = \iint \overline{\mathbf{v}'\sigma'}.\widehat{\mathbf{n}}d\mathcal{A}_{rea}$, and interior diapycnal mixing terms, $D_{int}$ and to (ii) the partition of $A$ into $A_{mean}$ and $A_{eddy}$ : $A_{eddy}$ is related to the 'bolus flux' ($\overline{\mathbf{v}'h'}$) and eddy-induced subduction through Eq.(1). Note that eddies appear both in $A$ and $D$.

### 2.1.1 Estimates of formation and dissipation rates

The annual formation rate of EDW (in the density range 26-27$\sigma$) implied by climatological air-sea buoyancy fluxes alone (i.e obtained by neglecting mixing in Eq.2, thus setting $A = F_{air-sea}$, and computing the difference $A_{27} - A_{26}$) is some 15 to 20Sv — see Speer and Tziperman (1992); Marshall et al (1999) and Fig.2 (right). If the volume of EDW is to remain steady over long timescales some 20Sv must be dissipated.

It is, of course, generally accepted that some EDW is subducted into the thermocline, implying a compensating loss of EDW in the interior. Depending on how the control volume is defined, loss of EDW volume could be caused by diapycnal mixing processes, eddies and/or mean flow. We discount southward mean export of EDW to the rest of the world ocean because, if anything, EDW is a density class of net import into the North Atlantic across the equator. The mixing-induced EDW volume loss in the interior may be written as

$$\frac{\partial D_{int}}{\partial \sigma} = \frac{\partial}{\partial \sigma}\left[K_z \frac{\partial \sigma}{\partial z} \times Area_{xy}\right] \sim w^* \times \ Area_{xy} \sim 1.5 \ Sv \quad (4)$$

where we have used the common notation for diapycnal velocity, $w^*$, and used a turbulent eddy diffusivity $K_z$. Here $Area_{xy}$ denotes the size of the control surface bounding EDW. Assuming a constant diapycnal diffusivity of order $10^{-5}\,\mathrm{m}^2\,\mathrm{s}^{-1}$ (Ledwell et al., 1993), a 200 m vertical scale for the pycnocline, and a control surface area extending over the North Atlantic to roughly $45°N$ ($3 \times 10^{13}\,\mathrm{m}^2$) we obtain a diapycnal transport of less than 2 Sv, an order of magnitude less than the transformation rate inferred from climatological outcrops and air-sea buoyancy fluxes. We admit that it may be possible that the turbulent diffusivity in the upper ocean above the EDW may be enhanced above the background value of $10^{-5}\,\mathrm{m}^2\,\mathrm{s}^{-1}$. But as the corresponding area with enhanced diffusivity would be small, we must deem it unlikely that mixing can support an internal diapycnal transport out of the EDW layer of order 20 Sv. We do not believe, then, that interior diapycnal mixing supports significant loss of EDW. It is much more likely that the formation rates based on $F_{air-sea}$ alone are considerable over-estimates.

An independent measure of EDW formation rates was recently given by Kwon and Riser (2004a) using WOCE ACCE float data by monitoring the seasonal cycle of low PV waters. They find a seasonal increase in the volume of their EDW from 2.4 to $4.7 \times 10^{14}$ m$^3$ between late fall (NDJ) and late winter (FMA), with about 10% uncertainty. (Using climatological data over the past 40 years, Kwon and Riser, 2004b, report a somewhat smaller annual cycle of 2.8 to $3.9 \times 10^{14}$ m$^3$). From the autumn to spring volume difference, the implied annual EDW production rate is 7.3 (float) or 3.5 (climatology) Sv. Kwon and Riser also computed subduction rates [integrated over the winter outcrops of the 17 °C and 19 °C isotherms from the American coastline to 40W] based on the kinematic definition of subduction using float-derived geostrophic velocities (Kwon and Riser, 2004c), mixed-layer observations and wind stress data. The resultant 3.9 Sv subduction rate is about half their inferred rate from the seasonal volume change for this same period, although Kwon and Riser argue that their subduction rate is probably underestimated due to the Gulf Stream being overly smoothed in their mapped fields. We note that comparison of these estimated formation rates is made difficult by the different control volumes and geographic domains employed. Nevertheless, the factor of $\sim 4$ discrepancy in formation rates between the $\sim$ 20Sv based on Eq.(2) with $D = 0$, and the $\sim$ 5 Sv based on Eq.(1), hint that the volume of EDW actually subducted into the thermocline may be very much less than that inferred from climatological SST's and air-sea fluxes.

One possible shortcoming of the discussion above is the neglect of $F_{ml}$ which represents water mass transformations due to mixed layer deepening. Garrett and Tandon (1997) show that density redistribution in the vertical by turbulent mixing achieves diapycnal advection across an isopycnal surface. Thus changes in mixed-layer depth can move density in and out of a density class. However, Tandon and Zahariev (2001) shows that $F_{ml}$ is perhaps 10% of the surface contribution and so the effect seems unlikely to significantly lower the estimate of net formation.

### 2.1.2 The hypotheses to be tested in CLIMODE

Two possibilities for the disconnect between estimates of mode water formation and dissipation rates are: (1) neglect of lateral eddy processes acting in the mixed layer in and near the Gulf Stream which provide lateral mixing and (2) estimates of air-sea fluxes are incorrect. We treat these in turn.

1. **Partial balance of air-sea buoyancy loss by lateral eddy processes.** Weller et al (2004a) report that air-sea flux and in situ Eulerian measurements of upper ocean stratification cannot be reconciled in the 'Subduction Experiment' without invoking lateral eddy effects. Speer (2002) argues that horizontal — i.e. diapycnal — mixing by geostrophic eddies may be significant based on a straightforward scale analysis of the vertical and horizontal diffusion of heat in the upper ocean. An estimate of the diapycnal eddy term can be made analogous to Eq.(4) thus:

$$\frac{\partial D_{eddy}}{\partial \sigma} = \frac{\partial}{\partial \sigma}\left[K_H \frac{\partial \sigma}{\partial y} \times Area_{xz}\right] \sim v^* \times Area_{xz} \sim 6 \text{ Sv}$$

if one assumes eddy heat fluxes occur along the 1500 km region of EDW formation, lateral length scales of 50 km and a representative $K_H = 1000\,\mathrm{m^2\,s^{-1}}$to represent eddy mixing. If a Laplacian diffusion term is computed over the mixed layer using the Levitus climatology and converted to a transformation by integrating over isopycnal areas, the resulting $D_{eddy}$ term is as large as $F$ if $K_H = 1000\,\mathrm{m^2\,s^{-1}}$. Lateral eddy fluxes thus emerge as a candidate to abate our dilemma. Evidence from models — see, e.g, Fig.6 — suggest that $K_H$ is surface intensified and can be significantly larger than $1000\,\mathrm{m^2\,s^{-1}}$. Finally, Marshall et al (1999)'s diagnosis of a coarse resolution ($1^\circ \times 1^\circ$) global ocean model driven by climatological fluxes found that in the range 26-27$\sigma$, $F$ is not a good predictor of $A$: mixing is significant everywhere in the upper boundary layer of the ocean model. Moreover, within the mixed layer and seasonal thermocline, $D$ generally opposes tendencies and attenuates the influence of $F$.

   Dynamically, we hypothesize that the onset of late winter convection, when combined with Gulf Stream heat transport, intensifies the meridional slopes of near surface isopycnals, resulting in an explosion of baroclinic instability. The heat flux so generated is envisioned as balancing much of the air-sea heat exchange. Evidences of this are seen in the remarkable SST maps prepared by Kathie Kelly in Fig.4. As the winter proceeds the entire region south of the Gulf Stream has eddying fluid in the range 18-19 °C encroaching into warmer waters. We believe the lateral heat and bolus flux associated with these eddies play a central role in the mixed-layer budgets, ameliorating the buoyancy loss at the surface. In the deep convection literature — see Marshall and Schott (1999) — this process is called restratification. Indeed CLIMODE can be thought of as addressing the 'flip-side' of deep convection: convection on the rim of an anticyclonic gyre in strongly sheared, eddying flow, injecting low PV into the interior.

   In an eddying ocean we hypothesize that the key terms in the Walin balance are:

$$\underbrace{A_{mean} + A_{eddy}}_{\text{partial cancellation}} = \underbrace{F_{air-sea} - \frac{\partial D_{eddy}}{\partial \sigma}}_{\text{partial cancellation}} \qquad (5)$$

   Note that eddy terms appear on both sides of the equation. The diapycnal volume flux in a turbulent ocean has a contribution from the bolus transport $A_{eddy}$ which, we believe, partially, and perhaps even largely cancels $A_{mean}$ in the vicinity of the GS front. This is certainly true in other strong frontal regions such as the Antarctic Circumpolar Current. For example, the vanishing of the 'Deacon Cell' in the Southern Ocean is a case in point where equatorward Ekman transport is largely balanced by poleward eddy-induced transport – see, e.g., the diagnostic study of Karsten and Marshall (2002).

2. **Uncertainties in evaluation of the formation rate, $F$.** The statistics of the air-sea buoyancy flux are likely to be highly variable in space and time making computations of $F$

based on climatologies somewhat problematical. Moreover $F$ involves integrating buoyancy fluxes across outcrop windows which are also time-dependent — see Fig.4. It is thus possible that the total 15-20 Sv formation number (based on climatologies) is itself not representative of the true formation rate.

Hence it is seen there are two possible ameliorating influences which can remedy the apparent imbalance between EDW production and dissipation. These are (1) $D_{eddy}/A_{eddy}$ in the mixed layer, which, hitherto, have only been subject to rather coarse estimation, and (2) the potential inaccuracy of the estimation of transformation $F$ using climatological data. We hypothesize that it is a combination of these errors that resolves the apparent imbalance between EDW formation and dispersal. We suspect the former is more important, but CLIMODE is designed to address both processes.

### 2.2 Upper-ocean mixing and its parameterization in models

Our conceptual model of the mixed layer of the ocean remains stubbornly $1-$dimensional: i.e. air-sea and entrainment fluxes are typically assumed to dominate lateral mixing, $D$, in setting $A$. However, as the ocean surface is approached, eddy fluxes must develop a diapycnal component because density is maintained vertically homogeneous by strong surface boundary layer mixing and yet the eddying motions are constrained to be horizontal by the upper boundary, as sketched in Fig.3(left). We can call this transition layer between the mixed layer and the adiabatic interior in which isopycnals are intermittently in contact with the turbulent mixed layer, the 'surface diabatic zone'. As argued above, this zone is likely to play a key role in mode water formation and dissipation.

The evidence that lateral eddy fluxes play an important role in the dynamics of the upper ocean has only recently come to the attention of the modeling community. Eddy parameterization schemes have hitherto been framed for the ocean interior and must be modified at the ocean surface. In the ocean interior, eddies result from the adiabatic release of potential energy and do not produce any diapycnal transport (Gent and McWilliams, 1990). Presently ocean models use various tapering schemes to turn off the adiabatic eddy flux schemes at the surface (Danabasoglu and McWilliams,1995; Large et al, 1997; Gerdes et al 1999), but the $D_{eddy}$ term in Eq.(3) is often set to zero! on the erroneous assumption that eddies can only play an adiabatic role. Furthermore, the choice of tapering scheme has a strong impact on the rate and characteristics of water masses subducted into the ocean interior. As a result, atmosphere-ocean interaction in coupled models is greatly affected by inadequate parameterizations. Indeed, as highlighted in Schopf et al (2004), this is one of the most critical parameterizations that compromises the present generation of ocean and coupled climate models.

It was in recognition of the importance of near-surface mixing in climate models that the Climate Process Team EMILIE (Eddy MIxed-Layer Interactions) was recently funded to foster our understanding of the effect of transient eddy motions in the upper ocean and to develop parameterizations of these effects for IPCC-class climate models. CLIMODE's focus on the cycle of mode-water formation provides a specific context in which the general issues of upper-ocean mixing can be addressed.

### 2.3 The elements of CLIMODE

CLIMODE has been constructed around a two-year period of field measurements with particular emphasis on the late-winter/early-spring periods: time when EDW 'formation' is highest (just prior to restratification). Observations will be collected at high spatial resolution over the top 500 m of the ocean to capture the processes associated with mode water formation in the context of the meandering front. Simultaneously, we will measure the evolving marine boundary layer above and document the air-sea fluxes that drive the two fluids. On longer time scale, the subsequent capping and initial injection of the mode water into the subtropical thermocline will also be observed, as well as its eventual dispersal.

Through study of the processes contributing to each term in Eqs.(2) and (1), a grand synthesis of the observations to quantify the production of new EDW becomes an attainable CLIMODE goal as described in Section 4. With this in mind, we strongly hope that the various elements of the proposed program can go forward together. Kwon and Riser's (2004) finding that interannual variability of EDW is comparable to its annual cycle suggests that in the absence of a fully validated model (such a model being a program goal), a Walin-style synthesis of data from different years would be problematic. On the other hand, prior studies of air-sea and lateral ocean fluxes (e.g. Dong and Kelly, 2004) and stratification changes (e.g. Kwon and Riser, 2004ab) will provide context for the proposed CLIMODE observations within the record of interannual EDW variability.

| CLIMODE | Air-sea fluxes $F$ | Eddies and mixing $D$ | Subduction, dispersal $A, V$ |
|---|---|---|---|
| Obs. | direct air-sea fluxes | ocean $\mu-$structure profiles | Lagr.& Eul. obs of stratification and bolus flux |
| | moored atmos. boundary layer measurements | fine-scale GS frontal surveys | EDW volume observations |
| | remote sensing of SST, winds sea level anom. | Lagr. obs of surface, upper ocean velocity/T/S | |
| Models | regional atmos. model | regional ocean model | regional ocean model |
| | Process | Process/CPT | Process/CPT |

A variety of measurements and modeling activities are proposed under CLIMODE; section 3 presents these in detail. Fig.4 and the table above provide an overview. For the two-year observation period, moorings (one surface, two subsurface) will be maintained in the EDW transformation region surrounded by an array of profiling floats (some acoustically tracked, others ARGO-style). Continuous remote sensing of the ocean surface properties (SST, winds, sea level anomalies) in the region will also be carried out, in conjunction with an array of surface drifting buoys. The surface mooring (addressing atmospheric forcing, $F$) will be positioned in the region of climatological maximum ocean heat loss, providing accurate local surface fluxes to quantify air-sea exchange at that point and anchor improved flux fields synthesized from in situ, remotely-sensed, and regional and global atmospheric model data and fields. The subsurface moorings (to quantify stratification variations and the subduction processes, the $A$ and $V$ terms) will be sited where maximum mixed layer depths are seen (as shown in Fig.1 right). Floats (acoustically tracked and ARGO-style) will further document subduction and injection of EDW into the subtropical gyre. The moorings will be serviced on October cruises during which some hydrographic sampling will be done, complementing the float observations. In addition, the Hydrostation S and BATS sampling programs off Bermuda and the Oleander VOS programs will continue their long-term observations.

The two winter cruises designed to measure upper ocean mixing processes, $D$, will follow a similar general plan. A compound, drifting spar buoy will be repeatedly deployed on the southern

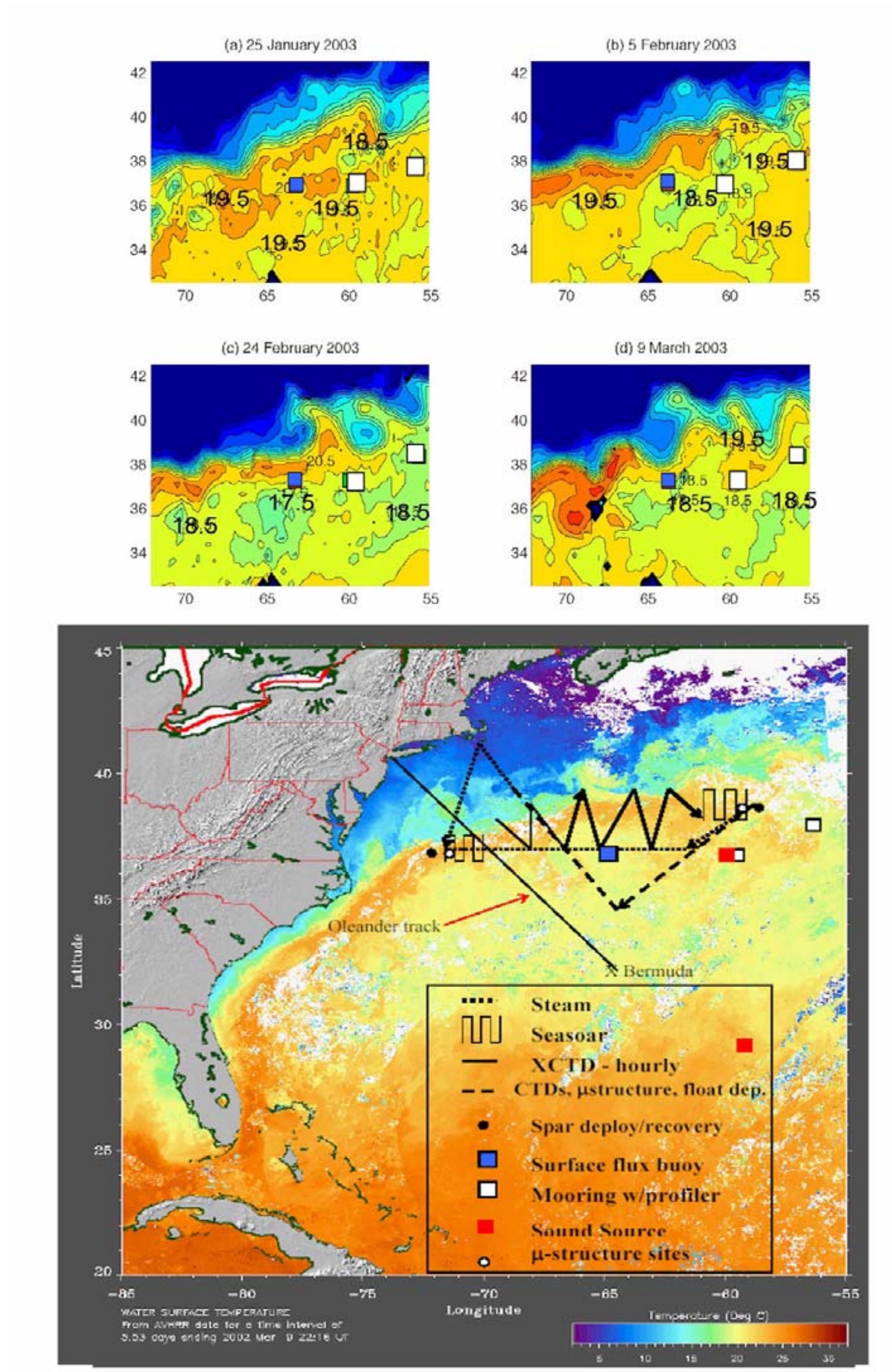


Figure 4: (top) Wintertime SST from the AMSR-E microwave sensor, courtesy of Remote Sensing Systems; contour interval 1 degree. Positions of the surface (blue) and subsurface (white) moorings are indicated. Note the warm core of the Gulf Stream and the irregular opening of the EDW ventilation window (classically between about 17.5 and 18.5 C). Bias errors of up to 0.5C may be present in these newly available data; Kelly will work with the data originators during CLIMODE to produce more accurate fields. (bottom) Schematic of CLIMODE fieldwork. Shown are nominal beginning and ending locations for the spar drifts, the SeaSoar and XCTD survey patterns, a subset of microstructure sampling sites, and two hydrographic section lines. Positions of the surface and subsurface moorings and two of the sound sources are also indicated. (The other two sound sources will be placed along 50W.)

flank of the Gulf Stream (GS) south of Woods Hole and tracked downstream as local ship surveys of the upper ocean and atmospheric boundary layer are carried out spanning the GS front. Ocean finestructure and direct air-sea flux measurements will be made from ship and compound spar; microstructure observations and Radiosondes, floats and drifters will also be released. After transiting the EDW transformation window, the spar will be recovered and the ship will head back west for redeployment, standing off the surface buoy for a time while enroute to inter-compare flux measurements. After the second spar drift, short hydrographic sections will be run towards the center of the Sargasso Sea and then back across the GS during the return to port. An extended observation period of 3 years, beyond that of the focused period, is also planned using ARGO floats and remote sensing to provide large-scale data on EDW distribution and properties.

In parallel with the observational program, modelling and theoretical studies will be carried out. The challenge of the modeling component of CLIMODE is to devise and implement an approach that has strong contact points with the observational program, is directed at testing the hypotheses that underlie the program discussed above, and, at the same time, permits transfer of understanding to the large-scale models used in climate research. A data-assimilative regional atmospheric model will be used to produce high-resolution maps of air-sea buoyancy flux. A combination of regional and process ocean models will be used to address the phenomenology of EDW formation and dissipation, ultimately leading to a better understanding of the leading order dynamics. In this regard we are fortunate in having a strong common interest with the CPT-EMILIE in upper ocean mixing. CLIMODE will test the success of this CPT and ground it in an observational context. In conjunction with that program we will be able to explore the whole range of scales with a hierarchy of numerical models of increasing complexity. Specific modeling plans are set out in Section 3.4 and summarized in the table above.

# 3 Proposed research

The rationale for the mix of observations and models described in this section is discussed in Section 2.3. The synthesis of the observations and models described here to test the hypotheses that drive CLIMODE are discussed in Section 4.

## 3.1 Air-sea fluxes: $F$ (Edson, Weller, Kelly, Samelson; and Skyllingstad)

Air-sea exchange is responsible for the buoyancy loss that causes mode waters to form. Specifically, determination of transformation $F$ requires accurate estimates of the sensible, latent, and radiative heat fluxes to compute the net heat flux, $Q_{net}$ (e.g., Weller and Anderson, 1996). The net heat flux is then combined with estimates of the net freshwater flux to compute the buoyancy flux

$$B(\sigma, t) = \alpha Q_{net} + c(P - E) \tag{6}$$

where $P - E$ (precipitation - evaporation) represents the net freshwater flux and $\alpha$ and $c$ represent a number of numerical constants and mean state variables (c.f., Garrett et al, 1995; Marshall et al., 1999) . Finally, the variation of the buoyancy flux integrated across isopycnals at the ocean surface defines $F = \partial\widehat{B}/\partial\sigma_{surface}$ and is the gradient required to compute transformation rates.

Obtaining accurate air-sea heat and freshwater fluxes and precise estimates of their gradients are crucial if we are to better understand and model the formation of EDW and evolution of upper ocean structure in the EDW formation region. However, there are large uncertainties in air-sea fluxes obtained from climatological data, atmospheric models, or remotely sensed data (e.g. Moore and Renfrew, 2002) such as those used to produce Fig.1. Errors in these fields stem from sparse sampling in a region of strong spatial gradients in the ocean, inaccuracies in atmospheric boundary layer properties and structure, especially during cold air outbreaks, and from the choice of flux algorithms. In CLIMODE for the first time we will accurately quantify the magnitude of the fluxes, their temporal and spatial variability, and their role in EDW formation. This requires long time series, spatial sampling (both remote and in situ), numerical modeling, and efforts to improve the parameterization of air-sea exchanges in bulk aerodynamic formulae. Therefore to reduce these errors and provide accurate estimates of the transformation and formation rates in the EDW region we propose to:

— Directly measure the spatial gradient of $B(\sigma, t)$ across the EDW outcrop to estimate F from direct covariance (DC) and radiative flux packages. These packages will be deployed on a research vessel and drifting spar during two winter cruises in the EDW formation region.

— Obtain a two-year time series of DC fluxes to improve and provide bulk aerodynamic (BA) fluxes from instrumentation deployed on a 3-m discus buoy.

— Implement a state-of-the-art regional mesoscale model over the EDW region to estimate and integrate the transformation rate over synoptic time scales.

— Use the continuous measurements from the discus buoy to provide continuous estimates of $B(\sigma, t)$ to ground truth the remotely sensed products, to evaluate the numerical modeling results.

— Use all of the direct observations of the heat fluxes and marine atmospheric boundary layer structure to improve analyzed air-sea flux products from remote sensing and BA parameterizations used in atmospheric models, targeting provision of accurate fields of surface forcing during the winter cruise case studies and, more generally, much improved regional air-sea flux fields over the two years of field work.

— Use the combination of data assimilative numerical models and remote sensed products to determine spatial gradients in $F$ and in particular to span outcrop EDW windows (inferred from remote sensing and in-situ measurements) to improve our estimates of $F = dB/d\sigma$.

### 3.1.1 In situ Spatial Estimation of $B(\sigma, t)$: (Edson, Weller)

A DC flux system (DCFS) and radiative flux package will be deployed aboard the wintertime survey ship. This package will directly measure fluxes of momentum, sensible heat and latent heat fluxes when the relative wind directions are favorable during transects (roughly half the time). The DCFS combines sonic anemometers, sonic thermometers and infrared hygrometers with accelerometers and angular rate sensors to motion correct and compute the fluxes using the DC method (Edson et al.,1998). The package will also directly measure the downwelling short and longwave radiative heat fluxes and precipitation. A downward looking radiometer will continuously measure the skin temperature to provide estimates of the upwelling longwave radiative heat fluxes. In combination, these measurements provide all of the terms required to estimate $B(\sigma, t)$. Additionally, a GPS rawinsonde system will be operated to launch radiosondes providing measurements of atmospheric boundary layer structure at 6-12 hour intervals.

A heavily instrumented drifting Air-Sea Interaction Spar (ASIS) buoy will be deployed from the survey ship to directly measure momentum, heat, moisture, and radiative fluxes; direction wave spectra; and near surface current and TS profiles to complete the FILIS profile from 8 m to the surface — see Fig.5. The air-sea flux package will be a slightly modified version of the DCFS used aboard the survey ship to provide additional estimates of $B(\sigma, t)$. The ASIS is an ideal platform for near surface layer turbulence measurements on both sides of the air-sea interface. It has been designed to reduce both the flow-distortion and platform motion. As a result, it permits higher quality flux estimates than is possible from research vessels and moorings. These design characteristics also allow high quality measurements of the directional wave field and near surface currents.

The ASIS will be built during the first year of the project at WHOI. It will consist of a mast — see Fig.5 — containing the DCFS, radiometers, mean meteorological instrumentation, a downward-looking ADCP, an upward looking high-resolution ADCP, a wave-wire array, and CTDs. The wave-wire array will provide estimates of the directional wave spectra. The CTDs and upward/downward looking ADCPs will extend the FILIS profile measurements from 8 m to the near surface as shown in Fig.5. As described in Section 3.2, FILIS will be tethered to ASIS (Fig.5) as they drift within the Gulf Stream towards the EDW formation site and the surface mooring. Each deployment will last approximately 10 days and the two deployments will be made during each cruise.

The ship and spar-based flux systems will provide estimates of $B(\sigma, t)$ as they zigzag across or drift through the EDW formation region. The estimates of $B(\sigma, t)$ can then be temporally averaged over, say, each 10 day deployment period to provide spatial maps of $B(\sigma)$, which then provide estimates of $F$. We will collaborate with the atmospheric modelers (see Section 3.1.4) to map out the variability in the marine atmospheric boundary layer (MABL) and its impact on $B(\sigma, t)$ by combining the modeling studies with the in situ and remotely sensed data as described in following sections.

### 3.1.2 In situ Continuous Measurements of $B(\sigma, t)$: (Weller, Edson)

To obtain a continuous, two-year, in-situ time series of surface meteorology (wind speed and direction, air and sea surface temperature, relative humidity, incoming shortwave and longwave radiation, precipitation, barometric pressure), direct and indirect estimates of the air-sea fluxes, and upper ocean temperature, salinity, and velocity, a surface mooring will be deployed at 37°N, 64°W (as marked in Fig.4) in October 2005, recovered and redeployed in October 2006, and recovered in October 2007. The surface buoy will be instrumented with two redundant IMET (Improved Meteorological) systems sampling once per minute, a package to sense motion of the buoy, a rugged 2-axis sonic anemometer for further redundancy in measuring mean wind, and a 3-axis sonic anemometer for direct estimation of momentum and buoyancy fluxes using the approach described by Edson et al.(1998). The mooring line will have 15 SeaBird SBE 39 temperature recorders, 4 SeaBird SBE 37 temperature/conductivity, and 6 Nortek Aquadopp acoustic Doppler current meters distributed across the upper 300 m, to record the evolution of temperature, salinity, and horizontal velocity in the upper ocean over two annual cycles.

Challenges to be addressed include mooring and instrument survival. A preliminary mooring design study has been conducted which indicated that a surface mooring with this instrument payload of small, light instrument will survive even in the face of Gulf Stream velocities (which lead to high static loads) and strong surface waves (which can lead to high cyclic loads). In the first year, a comprehensive mooring design study will be conducted to ensure the ability to stay on station in the face of strong currents and to tune the dynamic response of the mooring so that resonant responses are at frequencies away from those of the surface waves and swell.

We will deploy a sensor package to provide a record of buoy tilts, translational motion (e.g., heave), and angular velocities. These data will be combined with a 3-axis sonic anemometer-thermometer to make direct estimates of turbulent momentum and buoyancy fluxes. The system will be a lower-power version of the DCFS deployed on the survey ship and ASIS and will complement these measurements by providing a reference time series from a fixed location. The collocation of the DC fluxes with the vertical structure measurements will reduce the errors associated with bulk aerodynamic (BA) formulae. This will allow us to assess at that point the timing of maximum cooling, the timing of deepest mixed layer depth, the extent to which local 1-D processes govern mixed layer depth and upper ocean heat content. Ultimately, the range of conditions we expect to encounter over the 2-year deployment will provide an unprecedented data set for the evaluation and improvement of BA formulae, as well as a benchmark for model evaluation in this region.

During the two years the buoys are deployed, hourly-averaged surface meteorological data will be telemetered via satellite. This data will be withheld and not assimilated by numerical weather prediction centers so as to provide an independent data set. It will, however, be freely shared with other investigators and colleagues at the modeling centers to stimulate discussion of the performance of atmospheric models and to guide choice of forcing fields for ocean models. After recovery and post-calibration, the one-minute IMET time series and 7.5 minute-sampled oceanographic time series will be quality controlled. The buoy motion package and the sonic anemometers will provide blocks of data, likely to be 15-30 minutes every 2-4 hours (depending on the power budget on the buoy). Analysis and work to address our goals will begin after recovery and processing of the first year of data and continue after the recovery and processing of the second year of data.

#### 3.1.3 Remotely Sensed Estimates of Regional Variability in $B(\sigma, t)$: (Kelly, Weller)

In the analysis of data and models, remotely sensed measurements will be used in conjunction with in situ measurements of velocity, surface characteristics, air-sea fluxes, and heat content. Satellite observations will include JASON plus the TOPEX/Poseidon altimeter (1.5° spacing between subtracks), SST from both AVHRR and from the AMSR (cloud piercing) microwave sensor (see Fig.4), and winds from the QuikSCAT scatterometer (daily winds with a resolution of 0.5°). The SST data, in conjunction with field measurements and models, will be used to locate outcrops of EDW during the course of the experiment (see Fig.4). The remotely sensed fields and atmospheric model output will be combined with the in situ data to generate regional flux products for our investigations. For example, following Yu et al. (2004), improved regional fields of latent and sensible heat flux will be derived from a combination of numerical weather model and remote sensing to produced daily maps at 1° spatial resolution. These will be validated against the buoy data. Additionally, surface radiation fields from the ISSCP (Zhang et al., 2000) and Langley's GEWEX Surface Radiation Budget Project (Paul Stackhouse, personal communication) will be compared to the in situ data and used to develop regional fields of the radiative heat fluxes. By combining the latent, sensible, and radiative fluxes with model other estimates of precipitation (e.g., the majority of EDW outcrop region is generally south of 40°N where TRMM products are readily available), we will produce regional maps $B(\sigma, t)$ and estimates of $F = dB/d\sigma$.

#### 3.1.4 Regional Model Estimates of $B(\sigma, t)$ and Flux Fields: (Samelson, Skyllingstad)

CLIMODE investigators will take advantage of several different operational and regional atmospheric models and model products in order to quantify air-sea fluxes and their spatio-temporal variability, study air-sea interaction processes, and place in situ flux measurements in a broader context. In addition to the large-scale global model products discussed above, corresponding fields from the 12-km operational NCEP Eta mesoscale forecast model will be archived and analyzed. Mesoscale atmospheric modeling systems available for high- resolution regional simulations will include the Naval Research Laboratory's COAMPS (Hodur, 1997) and the MM5/WRF family of community research models. Samelson and Skyllingstad have experience with COAMPS in the Oregon and North Carolina coastal zones (Perlin et al., 2004; Skyllingstad et al., 2004), and Samelson is currently using MM5 to estimate small-scale winds west of Greenland (http://www-hce.coas.oregonstate.edu/ cmet/nares/).

The main advantage of regional mesoscale models over operational models is our ability to adjust their treatment of the sea-surface temperature field, the planetary boundary layer, and surface flux parameterizations, based on comparisons with the in situ and remotely sensed data. For example, Renfrew et al. (2002) have shown that different parameterization schemes lead to substantial mesoscale variations in air-sea fluxes. In some cases, these effects may be strong enough to affect synoptic flow patterns. Our efforts will focus on flexible approaches to addressing this problem, using mesoscale models in combination with the proposed observations. For example, to examine the variability in the MABL across the EDW formation region and its impact on air-sea exchange, case studies will be run nested in global models or reanalyses. The role of the mesoscale variability will be investigated by comparing the resulting model fields with independent rawinsonde data and DC fluxes measured from the mooring, ship, and ASIS.

We propose the following set of computations and analyses to study and quantify air-sea fluxes and provide estimates of $B(\sigma, t)$ using regional atmospheric models: (1) We will analyze the NCEP Eta flux fields and compare to CLIMODE observations and the large-scale model flux fields; (2) We will recompute surface flux fields from the NCEP Eta surface atmospheric fields, using a variety of flux parameterizations and SST fields improved where possible by detailed observational analysis during the CLIMODE intensive observational periods; (3) We will implement a regional mesoscale model over the region using lateral boundary conditions from a global operational model, archive and analyze the resulting air- sea flux and surface fields, and compare these to the NCEP Eta products developed in (1) and (2) above; (4) Finally, in addition to providing flux fields to CLIMODE ocean modelers described in section 3.4, we hope in the latter years of the project to pursue high-resolution coupled ocean-atmosphere simulations of EDW formation during the CLIMODE observational period.

### 3.1.5 Results from prior NSF support

**Surface Forcing and Upper Ocean Response in the Western Equatorial Pacific Warm Pool**, (ATM95-25844) $940,000 (February 15, 1996 to January 31, 2000) R. Weller, S. Anderson and A. Plueddemann Co-PIs. During TOGA–COARE a surface mooring was deployed in the center of the Intensive Flux Array from October 1992 to March 1993. The data from this mooring became a focal point for the field program and returned the first accurate and complete time series of the air–sea fluxes in the western Pacific warm pool (Weller and Anderson, 1996). We also produced high quality air-sea flux maps (Zhang et al., 2000) and followed up with a review of the state of the art of air-sea flux observations (Weller et al., 2004b).

**An Investigation of Gas Transfer Coefficients using Direct Estimates of $CO_2$ Flux**, (OCE-9711218), $414,000 (September 1, 1997 to August 31, 2000) W. McGillis and J. Edson, Co-PIs. The goal of this project was to perform direct covariance measurements of air-sea $CO_2$ flux during GasEx98. GasEx98 was performed in May/June 1998 in the North Atlantic during a time when the area is a significant $CO_2$ sink. Publications acknowledging support from this grant are Fairall et al. (2000), McGillis et al. (2001a,b).

**Analysis of Interannual Variations in the Upper Ocean Heat Transport and Storage in the Western North Atlantic**, OCE-0095688, Kathryn A. Kelly, Susan Hautala, University of Washington. (15 February 2001 – 28 February 2004, $434,890). Satellite observations and XBT data were used to examine interannual variations in heat storage, lateral heat flux convergence, and variations in mode water volume. Dong and Kelly, 2004; Dong, Hautala, and Kelly, in prep.

**Large Eddy Simulation of the Upper Ocean response to Inertially Resonant Winds**, (OCE-9711862), $182,000 (August 15, 1997 to July 31, 2000) E. Skyllingstad, B. Smyth and G. Crawford, Co-PIs. An LES model was used to examine how changes in wind direction affect upper ocean shear and the production of turbulence:Skyllingstad et al. 2000, Smyth et al. 2002.

**Coupled Dynamics of Large- and Meso-Scale Ocean Circulation**, (OCE-9415512 OCE-

981684), $325,000 (July 1, 1995 to June 30, 1998 and April 1, 1998 to March 31, 2000). R.M. Samelson and G.K. Vallis, Co-PI's. The primary goal of this project was to investigate the coupled dynamics of the large-scale and meso-scale ocean circulation: Samelson and Vallis, 1997a, 1997b; Samelson, 1998; Samelson et al., 1998, 2000a, 2000b; Samelson, 1999a; Samelson, 1999b; Samelson, 1999c.

## 3.2 Eddies and mixing: the 'D' term (Joyce, Gregg, Toole and Lumpkin)

A central component of the CLIMODE observational program involves winter upper ocean sampling in conjunction with direct air-sea flux measurements. The specific goals of this effort are to document and quantify the processes responsible for cross-frontal exchange and mixing sketched schematically in Fig.3 and visible in Fig.4. The specific hypothesis that we wish to explore here is that diapycnal processes are significant in the Walin framework and act to diminish the formation rate from that inferred solely from air-sea buoyancy flux.

We plan detailed field observations in February/March of project years 2 and in January/early February of year 3. These periods span the seasonal peak in cumulative ocean buoyancy loss by air-sea exchange and are when EDW formation is "most active" (Kwon and Riser, 2004a). Cross-frontal exchange and mixing have been observed least often during these times (late winter) and environmental conditions (strong baroclinic current). Consequently we have the least confidence in their model representations.

During each month-long cruise, a compound spar buoy (supporting atmospheric boundary layer sensors — see Section 3.1 —- and water column profiling to 500 m depth — see below) will be repeatedly deployed in the Gulf Stream (GS) within a larger-scale array of surface drifters. At the beginning and end of each drift, a high-resolution SeaSoar survey will be made of the GS front. And while the Lagrangian systems are carried downstream through the EDW formation region, repeated cross sections of the GS front will be made using loose-tethered microstructure instruments and expendable profilers. The spar system will be deployed on the southern flank of the GS where there is weak vertical shear (e.g. near station 64 in Fig.2). It will log and relay GPS positions to the research vessel in real time to guide the shipboard sampling. Based on observed GS speeds and past drifter/model data analyses (Garraffo; personal communication), we believe the system will remain in the Stream and transit the study area in about 10 days. Then the spar system will be recovered and redeployed back to the west for a second drift. In total we expect to recover approximately 36 drift days of winter data from the compound spar during the two cruises. In light of the seasonal climatology and frequency of atmospheric storms, the planned sampling should capture the full range of winter atmospheric forcing conditions. Should the spar system become entrained by a ring formation event or otherwise move away from the EDW ventilation window, we will recover and redeploy it. The surface drifter data will span the period from late fall through to at least the springtime capping of winter ventilated waters each year.

### 3.2.1 Synoptic Surveys of the upper ocean (Joyce)

Air-sea cooling over a strong baroclinic current will lead to a rapid breakdown of streamwise flow producing cross-frontal fluxes. These are expected to be modulated by the meandering Gulf Stream with preferential sites for exchange associated with meander phases — see Fig.3(left). To document the cross-frontal exchange, SeaSoar surveys that resolve the meanders and which can measure mixed layer T/S structure on intermediate scales will be carried out. The surveys will consist of 5 cross-front sections of 100 km length separated along-front by $\sim$ 25 km. They will provide visualization of the subduction process as illustrated in a related study in the Sea of Japan by Craig Lee (personal comm.). There is considerable T/S change on the density surfaces undergoing convection ($\sigma_\theta$ = 25.4 - 26.5 kg m$^{-3}$) across the Gulf Stream — see Fig.2. These range up to 4 °C, 1pss increase from north to south across the front. This "spiciness" signal provides an important and useful "tracer" for examination of cross-frontal exchange and the vertical transition from diabatic mixing to adiabatic stirring moving across the 'surface diabatic zone' — see Section 2.2. SeaSoar can routinely profile from the surface to 400m depth with approximately 1-2 km resolution along the tow path. In addition, the vessel will be equipped with ADCP/Ashtech GPS systems for measuring absolute upper ocean velocity and host an atmospheric boundary layer measurement program (see Section 3.1).

Between the SeaSoar surveys, the downstream evolution of the cross-frontal structure will be sampled using expendable instrumentation and microstructure profilers (see below), roughly following the displacement of the compound spar. Our plan calls for a series of cross-front sections in a zigzag pattern over the ca. 2.5 days between SeaSoar surveys. We will sample the temperature and salinity field with eXpendable CTDs (XCTD) and the velocity field with shipboard ADCPs. At an average XCTD deployment rate of one per hour during these periods, each survey requires 60 probes that will form between 7 and 9 cross-front sections. Each winter cruise thus requires 120 XCTDs.

### 3.2.2 Microstructure (Gregg)

The role of turbulent mixing processes in the wintertime formation of EDW will be quantified through study of ocean microstructure data. Note that we need to know $K_z$ to better than a factor of 3: for example if $K_z$ reaches levels of $3\times10^{-5}\,\mathrm{m^2\,s^{-1}}$, over large areas, then our estimate in Eq.(4) can reach sizeable values of $5-6$ Sv. Bursts of microstructure profiles will be collected during each of the winter cruises using two Deep Advanced Microstructure Profilers (AMPS, capable of sampling to 1100 m), supplemented by two Modular Microstructure Profilers (MMPs) which carry more sensors but are limited to 300 m. The bursts will be taken as often as permitted by SeaSoar operations at the beginnings and ends of the spar drifts, and regularly during the intervening period (the main microstructure sampling time). Several bursts per day of 1-hour duration are planned forming both time-series about the drifting spar and sections across the GS front with distance downstream. To better resolve shear in the vertical, Gregg will install his 150 kHz broadband ADCP on Knorr and operate it in parallel with the ships 75 kHz broadband which penetrates deeper but with lower vertical resolution. (Simultaneous operation of these two ADCP′s was carried out by Gregg on a recent cruise to the Black Sea with no indication of cross-contamination of the acoustic signals.)

To supplement the microstructure sampling, diapycnal mixing rates will be inferred from finescale shear and strain measurements using techniques that are now standard. The velocity measurements will come from the vessel-mounted ADCP's. During the Knorr's reoccupation of the A22 hydrographic section along 66W in November 2003, good ADCP data was logged down to around 750 m through the Sargasso Sea (J. Hummon, personal communication). We therefore anticipate velocity observations at 10 m scale in the upper 2-300 m from the higher-frequency ADCP and 25 m scale down to 600-700 m from the low-frequency ADCP. The microstructure bursts will define where diapycnal mixing occurs across the GS front with distance east (following the spar buoy), quantify the mixing intensity (e.g. estimate the diapycnal diffusivity) and document the physical mechanisms supporting the mixing. The latter is particularly key to the parameterization effort. We will also take some microstructure drops as close as possible to the time and location of SeaSoar profiles to determine how well turbulent dissipation rates can be estimated using density overturns observed by the towed body.

EDW is usually assumed to be ventilated in strongly convecting surface mixed layers on the gyre

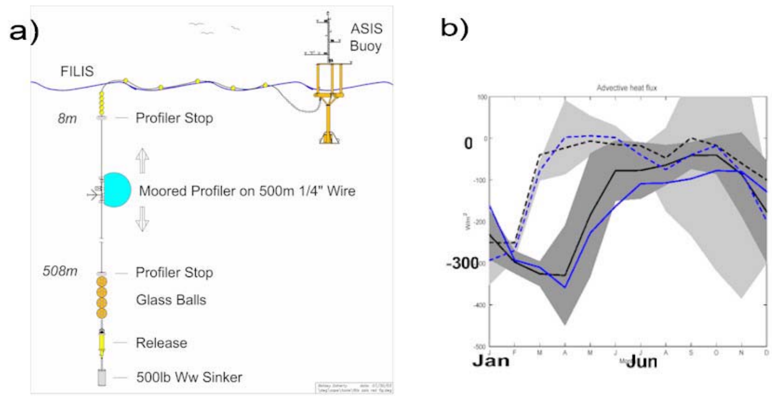


Figure 5: (a) Schematic of ASIS/FILIS compound spar. (b) Advection estimate from simulated drifters in the MICOM model over an annual cycle. Total (mean plus eddy) advective heat flux in the formation region, calculated from the model's Eulerian output (blue) and from all 3014 simulated drifters which sampled the region (black). Same but for eddy fluxes (dashed lines). Shading: rms error for an array of 200 drifters released in the formation region during the winter months.

side of the Gulf Stream front, but the observational evidence is inconclusive. Our observational plan allows for the possibility that intense mixing in sheared, stratified water below the surface may also be involved. The Gulf Stream in winter has the same elements as found in the upper thermocline on the equator in the central Pacific. There, 'deep cycle' mixing during nighttime convection extends far in to the stratified shear zone above the undercurrent core (Moum and Caldwell, 1985). Shear in the Gulf Stream is on average a little weaker than in the undercurrent, but during winter convection it can be much stronger, particularly during cold air outbreaks. If 'deep mixing′occurs, the most likely site is in the strongly sheared region at 80-140 m depth at 80 -140 km on the horizontal axis of Fig.2.

### 3.2.3 Drifting Compound Spar (Toole)

The proposed drifting spar buoy system will highlight departures from one-dimensional vertical balance during the development of the 18 °C thermostad. The system represents the marriage of two proven instruments: ASIS (Air-Sea Interaction Spar, Graber et al., 2000) and FILIS (the FInescale Lagrangian Instrument System: a standard Moored Profiler - Doherty et al., 1999; Toole et al. 1999 - deployed on a vertical tether suspended below the surface) — see Fig.5 (left). Sensors on ASIS will directly measure the air-sea exchange of buoyancy and momentum (see section 3.1) while the FILIS will repeatedly sample the oceans temperature, salinity and velocity profiles at 1m vertical resolution from approximately 10m to 500m depth at better than hourly interval during each drift. The resultant data set will be interpreted through conventional mixed layer models (in a Lagrangian framework), with residuals possibly manifesting lateral eddy flux divergence.

The FILIS configuration of the Moored Profiler (MP) was tested off Bermuda in November 2001. The distributed buoyancy (Fig.5) acts to minimize vertical heaving of the profiling wire. The arrangement was found to work well during the test deployment; tether heave was limited to $< 1\ m$ peak to peak for 12 seconds swells with near total suppression of wind wave motions; the resultant wire heave was well within the operating capability of the MP. The scalar data logged by the MP were of good quality and clearly documented diurnal stratification changes at the surface.

Individual velocity profile data exhibited 'noise' in the upper 50-75m due to surface wave motions but below the surface mixed layer the wave signal was much reduced and persistent finescale velocity features were observed in the thermocline on successive profiles.

During each drift, the SeaSoar, XCTD and microstructure surveys across the GS (with concurrent direct atmospheric flux measurements from the vessel) together with repeated bursts of microstructure profiles at the spar will link the ASIS/FILIS observations to the evolving structure of the Gulf Stream front.

### 3.2.4 Surface drifters (Lumpkin)

Drifter observations will be used to quantify eddy fluxes of mass and heat in the CLIMODE study region, and the role of these fluxes in modifying the transformation/formation of EDW. Analysis of these observations will produce a gridded field of near-surface total and Ekman-removed velocities in the study region which will be used in heat budget calculation and included in the CLIMODE data archive for the oceanographic community. The proposed drifter deployment strategy was tested using simulated drifters in a $1/12°$ version of MICOM (see Fig.5 b). The magnitude and spatial distribution of the eddy heat fluxes derived directly from the Eulerian model output compare well with those calculated from the simulated drifter observations using the residual method.

Satellite-tracked surface drifting buoys ('drifters') provide in situ, concurrent measurements of sea surface temperature and near-surface horizontal velocity. Satellite fixes of drifter position will be obtained at a rate of 6-8 per day, with subsequent quality-control and interpolation performed at NOAA/AOML's Drifting Buoy Data Assembly Center (DAC). A total of 200 drifters will be deployed in the study region. The deployment strategy is designed for coarse sampling throughout the region during the autumn and early winter months of EDW formation preconditioning, and dense spatio-temporal coverage during the late late winter cruises coinciding with deep convection and EDW formation. Deployment will be conducted as follows: during the 2005 and 2006 October cruises, 20 drifters will be deployed south of the Gulf Stream at $\approx 5°$ spacing throughout the recirculation region. An additional 20 will be deployed in January 2006 from the AX10 (New York to Puerto Rico) VOS platform, within and south of the Gulf Stream; subsequent trajectories of these drifters will aid in final planning of the February cruise. And on each late winter cruise (February 2006, January 2007) 70 drifters will be deployed before and during the first spar drift and synoptic XCTD survey. This drifter array will span the study region, with denser batches placed in particular features of interest such as meanders. Any additional drifters which enter the study region will also be included in the analysis. Of the 200 drifters to be deployed in CLIMODE, 60 will be funded by NOAA as part of its Global Drifter Program, at no cost to this proposal. NOAA will also cover the transmission and processing costs for these 60 drifters, and for all drifters after they have been in water for three months. All drifter observations will be made publicly available in real time on the GTS and in quality-controlled, regularly interpolated format by the DAC.

### 3.2.5 Results from prior NSF support

**Decadal variability of Subtropical Mode Water, Gulf Stream separation, and large-scale atmospheric fields in the N. Atlantic Ocean from 1954-1999.** (OCE-9818465) \$330,000 (March 1, 1999 to August 31, 2002) T. M. Joyce Principal Investigator. The variability of EDW at Bermuda was examined in terms of atmospheric forcing, the NAO, and GS path (Joyce et al., 2000), and more recently using Topex/Poseidon data (Frankignoul et al., 2002). Wintertime climate variability going back to the late $1800'$s was examined for the Eastern US and Sargasso Sea (Joyce, 2002) and a WOCE hydrographic section was combined with tracer and LADCP data using an inverse model to obtain absolute zonal transport across the section (Joyce et al., 2001a).

**Mixing under the ITCZ during EPIC 2001**, (OCE-0002903) $814,158 (August 1, 2000 to July 31, 2003) M. Gregg and J. Miller, Co-Principal Investigators. We observed mixing under the Intertropical Convergence Zone and the ocean's response to several hurricanes spinning up above us.

**Field Experience with a Moored Profiling Instrument**, (OCE-9617072) $233,468 (January 1, 1997 to December 31, 1999) J. M. Toole and D. E. Frye, Co-Principal Investigators. Prototype testing and ongoing development of the Moored Profiler, a device that can repeatedly profile the water and resolve the ocean′s temperature, salinity and velocity at high vertical resolution over long time: see Doherty,etal. (1999); Toole, et al. (1999); Morrison, et al. (2000, 2001); Worrilow, et al. (2002).

**Southern Ocean Transport** OCE-0117618, October 2001-September 2004, $414,000. R. Lumpkin and K. Speer (FSU), with A. Orsi (TAMU). We compute isopycnal transports and diapycnal transformation within the Southern Ocean using WOCE and other hydrographic sections in the framework of a box inverse model, with a focus on resolving the meridional overturning circulation. See Lumpkin et al (2002) and Lumpkin and Speer (2003).

### 3.3 Subduction, evolution and dispersal: the 'A' term (Fratantoni, Sloyan, Talley)

In contrast to the gyre-scale, heavily-smoothed view of EDW evolution presented by Kwon and Riser (reviewed in Section 2.1.1) we propose to describe in detail the process of EDW formation and evolution using a combination of moorings, floats, and shipboard observations. Measurements will be focused on a region of active EDW formation just south of the Gulf Stream and will resolve time and space scales compatible with those of the winter observations. This will enable evaluation of the terms hypothesized to be dominant in the Walin balance (Eq. 5) using synoptic, mesoscale-resolving observations.

The specific goals of this effort are to:

1. Quantify the EDW formation rate through direct observation of the subsurface velocity and density fields in both Eulerian and Lagrangian frameworks

2. Directly observe the Lagrangian evolution of temperature, salinity, and potential vorticity of recently subducted EDW and identify the mesoscale processes contributing to its horizontal redistribution and vertical modification

3. Measure the time varying volume and hydrographic properties of EDW in a domain adjacent to the formation region and, at lower resolution, over the entire subtropical gyre.

Three linked approaches will be applied: (1) High-vertical-resolution moored observations of velocity and density structure in the EDW formation region, (2) Detailed Lagrangian measurements of EDW subduction, evolution and dispersal and (3) Shipboard and ARGO float observations of regional hydrographic structure. These approaches are highly complementary: the moored time series will be used to determine the mean properties and thickness of EDW at formation, while variability from this mean both at the mooring and from float data will indicate the importance of lateral eddy fluxes in the subsequent evolution and dispersal of EDW (processes unresolved by previous low-resolution float studies and climatological analyses). The combined MP and float data will provide an independent estimate of subduction rate that can be compared with the diapycnal volume flux (or transformation) rate derived from the air-sea and diffusive flux terms of Eqs.2

and 1. Hydrographic measurements collected during mooring deployment and recovery cruises and continuously via profiling floats will provide larger scale context and enable estimation of seasonal and interannual changes in EDW volume.

The proposed moorings and floats will significantly extend our ability to observe and quantify the time-varying process of EDW formation beyond that possible during the two wintertime surveys. We will collect highly resolved measurements of the subsurface velocity and stratification over two complete annual cycles. While the most intensive phase of these observations will be coincident with wintertime measurements of surface buoyancy flux and mixing processes, the Eulerian and Lagrangian measurements will persist through the remainder of the year capturing in unique detail the subduction process, its initiation, and its eventual termination. Together the proposed measurements directly address the formation rate of EDW, as described in Section 4.2. The float observations will also permit computation of the lateral induction term ($\mathbf{v} \cdot \nabla h$) required for direct computation of subduction in the manner of Marshall et al (1993), Qiu and Huang (1995) and Kwon and Riser (2004a).

#### 3.3.1 Subsurface moorings equipped with Moored Profiler (Sloyan)

Two moored profiler (MP; Doherty et al. 1999) moorings will be maintained for a two-year period near 37°N, 59.6°W and 38°N, 56°W. The MP's will straddle a region extending from the southern edge of the Gulf stream to 500 km south where the 18°C outcrop diverges from the Gulf Stream in late winter and the deepest winter mixed layers are observed (Figs. 1 and 4). The moorings will be recovered and re-deployed after 12 months. The MPs, in addition to estimating the EDW subduction rate S, will also yield a time series of the seasonal property evolution of EDW water, including winter deepening of the mixed layer and subsequent surface restratification in spring. The MPs are similar to that which will be deployed on the drifting compound spar (ASIS/FILIS described in Section 3.2) except that they will profile at fixed locations. The maximum observed winter mixed layer depth at the mooring site is approximately 300 m, and the upper boundary of the EDW usually lies at depths greater than 200 m, apart from when the EDW outcrops during winter. A profiling interval between 50 m and 600 m will sample the bulk of the surface mixed layer and into the underlying stratified waters in winter, and span the full subsurface EDW layer during the remaining seasons. Temperature and salinity recorders will log 30 minute averages of temperature and salinity above (30 m) and below (650 m) the MP profile depth. In addition, Hydrostation S time-series data will be used, along with the floats and hydrography, to describe the dispersal of EDW within the northwestern Atlantic.

The MP will be able to profile five times per day for 6.5 months between October and mid-May and twice per day between late-May and late September, yielding a diurnal cycle of mixed layer deepening during the cooling season and more limited sampling during the restratification season. Continuous temporal sampling will come from temperature and conductivity recorders placed at the upper and lower limits of the MP profiling interval. These data will also aid in calibrating the MP conductivity data.

The MP mooring will provide a time series of velocity, temperature and salinity data through two complete seasonal cycles at higher temporal resolution than possible from the profiling floats (see Section 3.3.3 below). Active subduction is accompanied by velocity turning with depth, which is thought to be largest during the strongest cooling period (Schott & Stommel, 1978; Bingham, 1992; Spall, 1992). Thus we might expect significant changes in the velocity profile over the seasonal cycle, with maximum velocity turning in late winter/early spring as newly formed EDW is actively subducted into the ocean interior. If the subduction mechanism is primarily eddy-driven, the MP will also observe the associated fluctuations in velocity, temperature and salinity, quantified as eddy fluxes. Fine structure from cross-frontal processes associated locally with subduction might also be observed. Overturning scales will be estimated from the MP's CTD to investigate diapycnal mixing rates. Ability to infer dissipation rates from overturning scales with MP data was recently demonstrated using a data set off Hawaii (personal communication, Matthew Alford, 2003).

#### 3.3.2 Acoustically-tracked floats (Fratantoni)

Acoustically-tracked subsurface floats will be used to directly measure the time-varying velocity and stratification in the region of EDW formation, to quantify the rate of EDW subduction, and to investigate the roles of large-scale advection, eddy stirring, and inertial recirculation on the horizontal redistribution and vertical modification of EDW. The floats will be targeted specifically at the EDW layer and most will be deployed in a region of active mode water formation. The floats will be acoustically tracked with a temporal resolution of 8 hours (compared to 10 days for a non-acoustic ARGO float) enabling resolution of convoluted dispersal pathways and eddy fluxes of momentum, thickness and potential vorticity both near the formation region and over the entire low-PV EDW reservoir in the western subtropical gyre.

A total of 90 floats will be deployed to detail the export and recirculation pathways of subducted EDW and reveal the extent to which EDW is re-exposed to the atmosphere in subsequent winters. Thirty will be isopycnal RAFOS floats (Bower and Rossby, 1989) yielding temperature, pressure, and velocity on a single isopycnal within the EDW layer on missions of 18-24 months duration. The remaining 60 will be acoustically-tracked bobbing floats ("bobbers") measuring the velocity, temperature, salinity and thickness of the EDW layer. The bobbers will continually profile while drifting within the EDW layer (defined by locally-measured upper and lower bounding isotherms and encompassing depths from 0 to 700 m). Bobber float missions will be approximately 1 year.

Three distinct mechanisms will be used to deploy floats:

— 1. Approximately two-thirds of the floats will be launched during the two wintertime survey cruises in regions of active mode water formation. EDW formation is expected to be most prevalent between 45°W— 65°W along the southern edge of the Gulf Stream as found by, for example, Kwon and Riser (2004a) — see Fig. 2 (right). The precise location of intensive wintertime float deployments will be based on observations of deep mixed layers (e.g. from SeaSoar) and satellite SST analyses (such as Fig.4: top). We plan to deploy relatively dense arrays of approximately 30 floats (20 bobbers, 10 RAFOS) each winter over a roughly 500 km x 150 km patch abutting the southern edge of the Gulf Stream. The floats will tag water in the deep mixed layer and follow it as it is subducted into the interior with an initial resolution of 50 km, comparable to the local baroclinic deformation radius

— 2. In order to deploy floats over a broader time window than possible during the two wintertime cruises, an additional six bobbers will be deployed each winter, at a rate of one every other week, from the M/V Oleander on its transit from New Jersey to Bermuda.

— 3. The remainder (nominally 8 bobbers and 10 RAFOS) will be deployed serially from a SALP (Submerged Autonomous Launch Platform; see Results from prior support) installed on the southwest MP mooring. These will be released weekly during January-March 2006 and 2007.

The serial release of floats both upstream of the subduction region (Oleander) and near an anticipated site of significant mode water formation (SALP) will provide a unique record of temporal variability in the location of subduction and the pathways followed by recently formed EDW. The floats released from the SALP will tag subducted water measured in detail by the MP mooring.

Following subduction, we expect most of the EDW floats to recirculate within the western subtropical gyre and have devised an acoustic tracking array accordingly. One source will be collocated with the proposed MP mooring. The others will be deployed on stand-alone moorings in the central subtropical gyre. All of the sources will be deployed in October 2005 and recovered at the conclusion of the experiment.

#### 3.3.3 Shipboard hydrography and ARGO float analysis (Talley)

Gyre-scale context for the intensive EDW experiment will be provided by ARGO and ARGO-equivalent profiling floats and limited shipboard CTD sampling. Estimates of EDW volume mapped coarsely throughout the western part of the gyre will be analyzed together with the subducted volume estimates from the moored profiler moorings. Variations in upper ocean circulation, thermocline and oxygen structure will be mapped monthly. This large-scale mapping in conjunction with the very intensive Gulf Stream process study at the core of CLIMODE will allow new interpretation of the more extensive, historical WOCE float data set. Our profiling ARGO floats will also include oxygen sensors, unlike the WOCE floats and the acoustically-tracked floats, giving additional information on formation processes, rates and advection (Jenkins, 1982). Winter near-surface oxygen undersaturation is an indicator of convective mixing (Talley et al 2003).

The official ARGO program provides for only 6 floats in the entire region, parked at and profiling from 1800-2000 dbar every 10 days. We will add 12 floats, parked at 500 dbar and profiling from 1800 dbar to the surface every 5 days. Our floats, combined with the existing ARGO floats, will yield 80 to 90 profiles per month and will extend the WOCE ACCE float study of 1997-2002 (Kwon and Riser, 2004a,c). That study had 53 functioning floats, with about 40 to 170 profiles per month. Reliance on ARGO alone would provide only 18 profiles per month and is not sufficient to map EDW mixed layer evolution and dispersal, given its patchiness (Talley and Raymer, 1982; Kwon and Riser, 2004a).

Eight of our profiling floats will be deployed between 30°N and the Gulf Stream Extension, with one in the slope water north of the Gulf Stream and three south of 30°N. To deploy in the interior of the recirculation, the October 2005 cruise track detours to near Bermuda. CTDO casts to 2000 dbar for float profile calibration will be conducted at all profiling and acoustic float launch sites and at mooring sites. Salinity samples will be collected for CTD calibration with oxygen samples for coarse watermass mapping, CTDO casts will be made on the four cruises that include float deployments and/or mooring operations.

#### 3.3.4 Results from Prior NSF Support

Sloyan: No prior NSF support.

**Direct observation of North Brazil Current Ring structure, evolution, and core water dispersal.** OCE-9729765. $572,428 (4/1/98-3/31/02; D. Fratantoni and P. Richardson, co-PI's). The first comprehensive study of the structure and behavior of North Brazil Current (NBC) rings. (Fratantoni and Glickson, 2002).

**Red Sea Outflow Experiment (REDSOX): Lagrangian observations of outflow pathways and lateral dispersal.** OCE-9818464. $1,117,911 (3/1/00–2/28/04; A. S. Bower and D. M. Fratantoni, co-PI's.). A total of 50 acoustically-tracked RAFOS floats were launched in the western Gulf of Aden to observe the Red Sea outflow (Bower et al., 2002).

**Submerged Autonomous Launch Platform (SALP)**. OCE-0136255, $ 626,497 (2/1/02-2/29/05; Fratantoni, D. Frye, J. Valdes). SALP is an intelligent float launch device deployed in-line on a subsurface mooring. A SALP prototype is presently installed south of Bermuda for a one-year sea trial. See http://glider.whoi.edu/dmf for details.

**Pacific circulation based on WOCE observations; Pacific WOCE Hydrographic programme atlas**. OCE-9712209. $410,000 (11/1/1997-9/30/2002) L. D. Talley. WOCE data analysis and Pacific WHP Atlas (in progress: http://gyre.ucsd.edu/whp_atlas). Talley, 1999 a,b, McCartney et al., 2000, Hanawa and Talley, 2001, Talley et al., 2001; Talley and Yun, 2001; Yun and Talley, 2003, Talley (2003)

### 3.4 Proposed Ocean Modeling Studies (Dewar, Ferrari and Marshall)

We propose to examine EDW formation and its impact on the overall dynamics of the ventilated thermocline in a sequence of hydrostatic and non-hydrostatic modeling studies using the MITgcm. The long-range objectives of this work includes assimilation in nonlinear environments and the use of extreme resolutions in non-hydrostatic calculations.

The CLIMODE modeling components are designed to address several objectives. First, by their nature, the observations will provide detailed, but incomplete samplings of the turbulent, time-dependent EDW production. Models are needed to draw together the various observations into a single, dynamically coherent framework. This is the primary objective of the regional/Gulf Stream modeling described below. Activities there are to result, in later proposal years, in an assimilative regional model. We also propose a sequence of more idealized, hydrostatic and non-hydrostatic experiments to provide detailed examinations of convection along the energetic rim of a basin recirculation. Given short convective timescales, such events should not be sensitive to the idealizations, and thus a numerical framework for the interpretation of field data will result. Other applications will address dynamical consequences of EDW formation and variability. With regards to these broader issues, mode water generation represents a potential vorticity input into the subsurface, adiabatic interior that the background mesoscale turbulence must contend with. In so dispersing this water along isopycnals, the larger scale flow will be shaped. Ultimately the large-scale flow will set the basic framework in which new EDW is formed and, from the intrinsic variability of the circulation, low frequency variability ocean-atmosphere heat exchange will result. Last, our modeling strategy for CLIMODE bridges seamlessly to the ongoing CPT-EMILIE program in ocean surface layer parameterization. Specific CPT modeling plans (already funded) are also briefly reviewed.

**What Questions Will be Addressed?** The modeling studies are designed to address: a) the two hypotheses set out in Section 2.1.2 — i.e. the importance of ocean eddy processes in balancing atmospherically implied subduction and the impact of atmosphere/ocean synoptic activity on formation b) the larger-scale influence of low potential vorticity on the dynamics of the North Atlantic circulation.

#### 3.4.1 Specific Modeling Plans

**1. Characterization of lateral stirring by geostrophic eddies (Ferrari and Marshall)**

A central goal of CLIMODE is to quantify the role of the lateral transport of geostrophic eddies at the surface of the ocean acting directly across outcropping isopycnals. As argued in Section 2.1.2, this process is likely to play an order one role in Eq.(2) through $D_{eddy}$.

We propose to infer $D_{eddy}$ by studying the evolution of an idealized tracer driven by the velocity field derived from satellite altimetry. The theoretical basis of the approach is set out in Nakamura (1996) and Winters and d'Asaro (1996) — it is closely related to the Walin formulation — and has been applied to diagnose tracer transport in the stratosphere (Haynes and Shuckburgh; 2000) and in the southern ocean (Marshall et al; 2004). The attractiveness of the approach is that, by adopting a tracer-based area coordinate, tracer transport can be rigorously phrased as a diffusion problem in which $D_{eddy} = -K_{eff} \int |\nabla b| \, d\mathcal{A}$ in Eq.(5) where

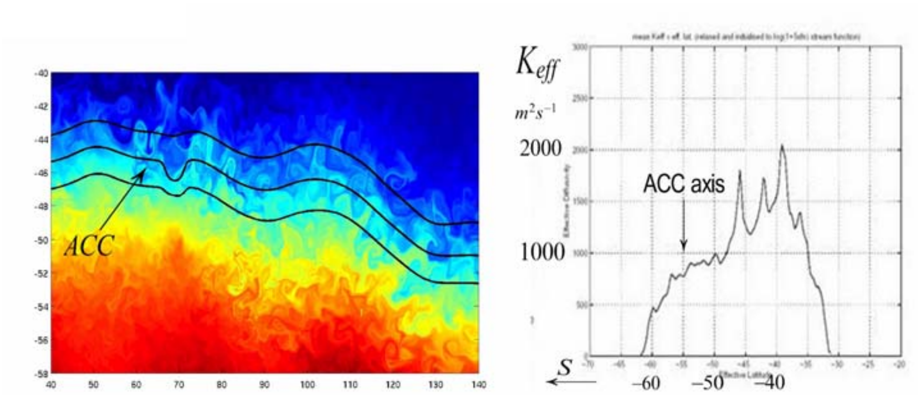


Figure 6: (left) Instantaneous tracer distribution in the southern ocean obtained by advecting an initially monotonic tracer distribution by altimetric observations. The calculation was done globally — only a 10000×2000 km region is shown (right) The $K_{eff}$ diagnosed using Eq.( 7) plotted as a function of equivalent latitude.

$$K_{eff} = k\frac{L^2}{L_{eq}^2} \tag{7}$$

where $k$ is the (numerical) microscale diffusivity used in the forward integration, $L$ is the length of the tracer contour strained out by the eddy field and $L_{eq}$ is its unstrained (coarse-grained, equivalent to $\overline{b}$) length. The ratio $\frac{K_{eff}}{k}$, a sort of Cox number, can be interpreted in terms of the degree to which tracer contours are lengthened by the straining field, as measured by $\frac{L^2}{L_{eq}^2}$. In turbulent flows, like the ocean, $K_{eff}$ is independent of the magnitude of $k$, because it is the stirring of tracers by the large scale velocity field that controls the tracer gradients on which the microscale diffusion acts.

Fig.6 shows our estimates of $K_{eff}$ in the southern ocean obtained as described above: — note that $K_{eff}$ is large on the equatorward flank of the ACC and diminishes as the axis of the ACC is approached where (we surmise) mixing is inhibited by the large PV gradients associated with the surface expression of the ACC. As discussed in Karsten and Marshall (2002) $K$'s of this magnitude suggest that eddies play an O(1) role in the dynamics of the ACC.

In CLIMODE we propose to:

— implement an advection-diffusion model driven by altimetric observations over the Gulf Stream and its recirculation to infer $K_{eff}$. We anticipate that $K_{eff}$ will be large ($> 1000\,\mathrm{m^2\,s^{-1}}$) to the south and decrease as the GS front is approached. Our estimates of $K_{eff}$ will be compared to those inferred from CLIMODE field observations (see Section 4.2.1) and employed to estimate the importance of $D_{eddy}$ in the Walin synthesis.

— generalize the Nakamura approach for a non-passive tracer such as SST in which damping terms due to air-sea interaction come in to play. This will enable us to understand and quantify how lateral eddy stirring at the sea surface can affect SST.

— apply the $K_{eff}$ calculation level by level to the regional eddy-resolving model (see 2. below) to compare with CLIMODE observations and make use of in the synthesis.

**2. A regional model of the Gulf Sream and its recirculation (Marshall and Ferrari)**

We propose to set up a regional model of the North Atlantic with full thermodynamics of the kind described in Paiva et al. (1999), Smith et al. (2000), Oschlies (2002) and Chassignet et al (2003). The MITgcm will be used. This model will ultimately become the vehicle of a data assimilation effort in years 4 and 5 of the project. In setting up this regional model we assume a global ECCO (Stammer et al, 2002) assimilation product is available (also employing MITgcm) which will be used to supply open boundary conditions. The 'interior region' will have a horizontal resolution of $\frac{1}{12} \times \frac{1}{12}^{\circ}$ covering the focus area of the field program. The regional model has 40 vertical levels in all, with high vertical resolution near the surface ($\lesssim 10m$). As in the global model it employs the KPP mixed layer parameterization of (Large et al, 1994).

Our strategy will be to:

— run the regional model in forward mode driven by analyzed winds and air-sea fluxes using initial and lateral boundary conditions provided by the global ECCO model. The model will be compared to historical/climatological data to see if broad aspects of the numerical solution — GS separation, mixed-layer depth, mode water formation, recirculation etc — are plausible. Preliminary study using a regional $\frac{1}{6}^{\circ}$ model, shows that the model can produce mode water but it is slightly too cold and, due to GS separation problems, shifted somewhat far north. We believe that our $\frac{1}{12}^{\circ}$ model will have sufficient resolution and can be tuned to significantly improve the fidelity of the solution following the methods used by Chassignet et al with the MICOM model.

— compute $K_{eff}(z)$ at each level in the model applying the techniques described in 1. above.

— compute terms in Eq.(2) and decompose them into mean and eddy contributions to $D$ and $A$. This has already been done in great detail for a non-eddy resolving version of this model in Marshall et al (1999). In the presence of eddies, how important are the eddy contributions relative to the mean in Eq.(2)? Is there indeed a compensation as suggested by the hypothesis, Eq.(5)?

— make use of the model to place the observations collected in CLIMODE in a larger-scale context – see synthesis section. Does the model capture any of the phenomenology seen in the observations?

In years 4 and 5 of the project we will begin to use this model in data-assimilative mode using the adjoint methodologies developed in the ECCO project and the analyzed air-sea fluxes and winds provided by the '$F$' group — see Section 4.1. State estimation in the presence of strong eddies can be problematic. Lea et al (2000, 2002) and Köhl and Willebrand (2002; 2003) show examples of fluid systems in which the local derivatives required by the adjoint method can fail to exist or become inaccurate. Gebbie, Heimbach and Wunsch at MIT have shown however, that in a realistic, full eddy-resolving model of the eastern North Atlantic, no particular problems were encountered in practice. Some of this straightforward behavior probably lies with having an adequate set of observations (provided by the 'Subduction Experiment') to constrain otherwise unstable model components. CLIMODE will provide as good a data set to constrain the model, but the western North Atlantic is more energetic and includes a powerful meandering jet. Our strategy is to first attempt the assimilation with the adjoint method. Only in case of failure will other methods be employed.

This regional model, together with coarser-grained versions of it, will also serve as a vehicle in which to test out the new parameterization schemes coming out of CPT-Emilie, in which we are heavily involved.

**3. Process Hydrostatic and Non-hydrostatic Modeling (Dewar)**

The MIT model will be deployed in a very high resolution, non-hydrostatic configuration in order to study the effects of seasonality and jets on mode water formation and subduction. Three differing forcing scenarios will be considered; namely, F1 steady forcing, F2 climatological seasonal forcing and F3 synoptic seasonal forcing. This sequence addresses several questions, with perhaps the most prominent being the role of variability in "tuning" the EDW. The arguments rationalizing this sequence are as follows. Many steady (or nearly so) numerical North Atlantic models produce an EDW-like water mass. Thus, it can be reasoned that "EDWs" are a property of the steady general circulation. Thus, scenario F1 above is the simplest realization of this type of "EDW" subject to non-hydrostatic dynamics. Such a calculation provides a foundation for examination of later calculations, and gives us a first look at the important ocean eddy processes in "EDW" formation. The next step in complexity is F2, this being also the level at which the Speer and Tziperman (1992) analysis was conducted with observations. There are examples of "EDW" studies at coarser resolution and subject to seasonal forcing where "EDWs" have been generated (often with characteristics inconsistent with observed behavior). Scenario F2 will address fundamental questions about the influence of variable forcing on STMW. In addition, interactions between forcing and ocean variability that modify EDW eddy subduction mechanics will be studied by comparisons of annual averaged F2 results with F1. Finally, analysis of the most realistic F3 case will address the impact of atmospheric synoptic variability on EDW formation by performing pure Lagrangian Walin analyses of F3 for comparison with climatological analyses of F2 and F3. Long-time mean F3 EDW will be compared and contrasted with F1 EDW. Finally, causes of these differences will be examined using the full, time dependent results from the EDW formation zones.

A second set of numerical experiments will consist of off-line tracer release studies using archived model fields. One suite will employ passive tracers released upstream of the separated jet, in order to locate regions of mode water recruitment and to describe the buoyancy history of that fluid. Comparisons between steady and variably forced experiments will clarify the role of the seasonal cycle. We also plan tracer release experiments in mode water formation zones in order to address the long term fate of mode water production. Last, we will generate a computational analog to Speer and Tziperman (1992) by computing Walin analyses for mode water and neighboring potential densities. The goals of these calculations are to determine where and how net formation is balanced by loss and contrast the mode waters with non-mode waters.

The model configuration will involve a rectangular 1000 to 1500 km square basin. High resolutions, O(100m) in the horizontal and from O(5m) in the vertical, will be employed to resolve the zone of the jet and the near surface boundary layer. A 1.5km deep basin should prove adequate to capture the relevant dynamics, as mode waters and wind-driven flows do not penetrate deeper than 800m. The model will be initialized by interpolating from coarser resolution results run to thermodynamic equilibrium. Fine resolution model adjustment should be advective on a recirculation time scale of about 200 days. Thus the model will be run for 2-3 years prior to archiving results.

### 3.4.2 Coordination with CPT (Dewar, Ferrari and Marshall)

The goal of the CPT-EMILIE initiative is to put the interaction of eddies and mixed layers on a solid theoretical foundation and to derive parameterizations for $A_{eddy}$, $D_{eddy}$, and $F_{ml}$. The numerical experiments planned as part of the CPT have great relevance for the present proposal, but the CPT group is not funded to test their results in the specific context of CLIMODE. Our model calculations have been designed to bridge the gap between the CPT initiative and the observational field program.

The non-hydrostatic model calculations complement the MIT CPT group lead by Ferrari, who is studying a non-hydrostatic jet in a channel, by introducing basin geometry of a basin, synoptic variability in the wind and heat flux, essential elements for the study of EDW formation. We plan to cooperate closely with this CPT effort. We also complement the UCLA CPT effort, lead by McWilliams, where plans to embed a regional, fine scale, O(1km), model in the western boundary and separated jet region of a coarse grid North Atlantic model are progressing. Plans are to employ a hydrostatic model with KPP to compute unresolved physics. Our regional modeling contrasts with this effort in resolution, model design (the MIT model vs ROMS), and in the use of reanalysis forcing. Our non-hydrostatic modeling differs in dynamics, explicit boundary layer computations, its implementation in an idealized basin and in the various forcing scenarios. Again, all phases of the CLIMODE modeling effort plan close cooperation with the UCLA group.

Two other CPT groups, led by Vallis and Hallberg, plan to run idealized experiments to shed light on the interaction between turbulent mixing and ageostrophic circulations driven by eddy straining. The results of these calculations will be used to interpret the analysis of secondary circulations from SeaSoar profiles.

Finally, CPT groups led by Rudnick and Speer are analyzing existing Seasoar, moored and hydrographic observations to find what statistics best quantify the eddy effects in the upper ocean. We plan to cooperate with these groups to find-tune our sampling strategies and optimize measurement of the relevant statistics.

### 3.4.3 Results from Prior NSF support

**Self-Propagating Eddies in Realistic Basins**, NSF Award OCE-0220884: August 1, 2002 to July. 31, 2005, G. Sutyrin and W. Dewar, Co-Principal Investigators. This grant funds research in vortex-topography interaction and tracer transport. In addition, it has supported climate, modeling and thermocline studies. Four publications have been submitted or are in press that acknowledge the grant, and one graduate student is supported by it.

**Studies of Circumpolar Currents**: OCE-0221369; 01/02 – 31/05. PI: John Marshall. This is a study to develop and apply simple theoretical models of the Antarctic Circumpolar Current to further our understanding of this central component of the global ocean circulation: — see, e.g. Karsten, R. and J. Marshall, (2002); Marshall and Radko (2003).

**Collaborative Research: Interaction of eddies with mixed layers:** OCE-0336839; 9/03–8/06 ($308,156). PI: Raffaele Ferrari, Co-PI: John Marshall. A collaborative study among ten different institutions to tackle the problem of eddy parameterizations in the ocean mixed-layer with theoretical, numerical, and observational tools.

# 4 Synthesis

The CLIMODE team will work collaboratively to improve our understanding of mode water formation, dispersal and dissipation. Several cross-cutting themes will be explored, organized again by the Walin framework with observational and theoretical/modeling approaches inter-twined. We group the discussion of these efforts as follows: (1) formation (2) mixing processes (3) subduction and dispersal (4) Walin synthesis (5) parameterizations in models.

## 4.1 Formation: $F$

CLIMODE will produce synoptic maps of air-sea buoyancy flux, $B_{air-sea}$, and outcrop windows, $\sigma_{surface}(t)$, and estimate the magnitude of the entrainment flux at the base of the mixed layer, $F_{ml}$.

**Air-sea buoyancy flux:** $B_{air-sea}$: We plan to combine in situ and remotely sensed data within the atmospheric mesoscale model to yield 2-D maps of $B$ over the experimental period. The in situ team (Weller and Edson) will work to combine the direct covariance and bulk meteorological measurements to provide local time series of the surface forcing, and together with the remote sensing (Kelly et al) and atmospheric model teams (Samelson, Skyllingstad, and Jiang), produce spatially-resolved flux fields over the EDW ventilation window. The atmospheric modelers will also utilize these measurements in a series of model runs with and without the in situ data to help evaluate the model's ability to internally predict the boundary layer response. As such, both the measurements and model output will be used to investigate the role of MABL processes on air-sea interaction in the region. The COAMPS model will be used as a tool for all testing and evaluation of newly developed surface flux parameterizations. The starting point here will be the TOGA COARE bulk algorithm (Fairall et al, 1996) which, although initially developed in the low-wind Pacific warm pool, has now been refined and extended to higher wind speeds as described in Fairall et al. (2003). We expect that the range of winds, sea states and atmospheric stability that characterize the EDW region will significantly expand the parameter space for bulk-formula validation.

We will also collaborate with Peter Taylor and the air-sea flux group at Southampton Oceanography Centre, with whom we have had a 10-year collaboration. They have identified the CLIMODE region as one where large bias in present flux estimates is a concern, both locally and because it precludes closure of the heat budget over the basin (Josey et al., 1999). We will also carry out comparisons of operational model products with our in situ measurements in collaboration with Dick Reynolds (NCDC), Hua-Lu Pan (NCEP), and Anton Beljaars (ECMWF). Such interactions promise to be very fruitful in CLIMODE, particularly because of the shipboard boundary layer work and rawinsonde measurements.

**Outcrop windows:** $\sigma_{surface}(t)$**:** Kelly, Lumpkin and Joyce will use remote sensing (SST) and in situ observations (from drifters, SeaSoar and floats) to produce synoptic maps of the outcrop windows, $\sigma_{surface}$ which will be used in the computation of transformation discussed below. These synoptic fields will be compared to outcrop positions based on smoothed representations of the observations typical of the climatologies used previously to estimate EDW formation.

**Entrainment at mixed-layer base:** $F_{ml}$ .

Garrett and Tandon (1997) derived formulas to evaluate diapycnal advection due to mixed layer time dependence and entrainment fluxes $F_{ml}$ . The formulas require air-sea buoyancy flux, mixed-layer depth, and horizontal gradients of temperature and salinity in the mixed layer. Ferrari and Joyce will use $B_{air-sea}$, SeaSoar profiles, and drifter data to obtain estimates of $F_{ml}$ . Tandon and Zahariev (2001) found that $F_{ml}$ is only 10% of $F_{air-sea}$ in the North Atlantic, but their calculation was based on a one-dimensional mixed-layer model with very poor horizontal resolution. CLIMODE will be an ideal data set for this calculation because air-sea fluxes will be available together with high-resolution vertical and horizontal profiles of the mixed layer.

**Computation of formation rates**

The analyzed fields of $B_{air-sea}$ and $\sigma_{surface}$ will be combined by Edson and Kelly to infer $F = \frac{\partial B}{\partial \sigma}_{surface}$. This requires that we integrate $B_{air-sea}$ over outcrop windows (and then differentiate). It will be key here to compare the magnitude of $F$ obtained by taking coarse-grained and fine-grained approaches: i.e.

1. climatology — combine monthly-mean outcrops and monthly-mean $B_{air-sea}$ to yield $\overline{F}$.
2. synoptic air sea fluxes and synoptic outcrops — these will yield $F$.

How will $\overline{F}$ compare to $F$ and the climatology shown in Fig.2? How does $\overline{A}$ compare to $A$ (from Eq.2)? Of course the difference between them is (largely) the $\frac{\partial D_{eddy}}{\partial \sigma}$ and $A_{eddy}$ terms which gives us a nice contact point with other estimates of $D_{eddy}$ and $A_{eddy}$ discussed below.

## 4.2 Mixing processes: $D$

### 4.2.1 Near-surface eddy diffusivity: $K_{eff}$

Estimates of surface eddy diffusivities are particularly important because current parameterization schemes in ocean climate models are phrased in terms of diffusivities and the product $-K_{eff}\nabla\sigma$ will provide an estimate of the magnitude of the $D_{eddy}$ term in Eq.(3) in a form that can be easily implemented in ocean models — see Section 4.5. We have a number of strategies to estimate the magnitude of lateral mixing processes in the upper ocean. Lumpkin will take the lead in estimating Lagrangian statistics, both single-particle scales and ensemble dispersion, from the drifter trajectories. These statistics are typically averaged in relatively large spatial bins (e.g., Lumpkin et al., 2002) due to sparsity of data; for the dense winter arrays deployed in CLIMODE, averages within winter-mean isopycnal outcrops will be calculable at much higher spatial resolution. In addition, because the drifters measure temperature, the diabatic dispersion with respect to the winter-mean isotherms will be calculated from the rate of growth of the Lagrangian temperature variance of drifter ensembles (c.f. Sparling et al., 1997). This dispersion depends upon the statistics of the lateral eddying motion about the winter-mean state, and upon the spatial structure of winter heat loss to the atmosphere, the former process associated with the diffusive $D$ term in the Walin framework. The planned resolution will permit calculating diabatic dispersion for drifter ensembles in various dispersion regimes, characterized from high to low frequency by inertial oscillations, frontal meanders, rings, etc. encountered within the study domain. This will be done in collaboration with Dewar where analogous calculations will be made in the context of the non-hydrostatic modeling presented in Section 3.4.

Marshall and Ferrari will take the lead in computing $K_{eff}$ from satellite altimetry using the tracer advection technique described in Section 3.4. The ship-based sampling about the GS front by Joyce will enable us to estimate $L^2$ in Eq.(7) along a segment of the EDW outcrop and so obtain a check on $K_{eff}$. Comparisons of $L^2$ for active (buoyancy) and passive (spice and fluorescence) tracers will be instructive.

#### 4.2.2 Delineation of mixed layer and the surface diabatic zone (SDZ)

XCTD and SeaSoar data will be principal CLIMODE data showing the spatial structure and depth change of the mixed layer, which is central to the heat and freshwater balance of the water column. Joyce will analyze these data with Kelly and Lumpkin to investigate the surface heat balance of the GS and mixed layer. Variability of the cross-stream flow from the SeaSoar surveys also will be used to estimate property fluxes across the front — see Section 4.2.4. Thermohaline spiciness, analyzed as a tracer, will contribute here, both in the data analysis and in parallel calculations of model output. Modelled and observed variability of the mixed layer will be compared. A complementary, long-term view will be provided by Weller and Sloyan's analysis of the surface mooring and profiling mooring data.

#### 4.2.3 Computation of $D_{eddy}$ and $dD_{eddy}/d\sigma$

From observations of mixed-layer depth and the above inferences of $K_{eff}$, Marshall and Ferrari will compute $D_{eddy} = -K_{eff}\ \nabla\sigma$ integrated across the depth of the boundary layer and go on and estimate the $dD_{eddy}/d\sigma$ term in the Walin transformation equation. Note that the combined use of the modeling approach set out in Section 3.4, reality-checked with in situ inferences (see section 4.2.4 below), make this a sensible calculation to attempt. How large is $dD_{eddy}/d\sigma$ compared to $F$ deduced in Section 4.1? Can we understand the distribution in terms of the inference of $dD_{eddy}/d\sigma$ based on the difference between the coarse-grained and fine-grained computation carried out using air-sea fluxes described in Section 4.1?

#### 4.2.4 Measurements of the synoptic eddy field and fluxes

A variety of approaches will be made to estimate the lateral eddy fluxes. We anticipate that cross-frontal flows from the cold to warm side of the GS will advect cooler, fresher water and visa versa for northward flows. Any rectification, due to vigorous vertical mixing, will result in a net property flux across the GS. Shipboard, float, drifter and remotely-sensed data will be used to explore this process in the ocean while a parallel analysis is carried out with model data. Eddy fluxes $\overline{v'T'}$ and $\overline{v'S'}$ will be computed from mooring data by Sloyan and Weller. These "direct" flux estimates will contain both a "skew"-component which contributes to $A_{eddy}$ and a diabatic component which contributes to $dD_{eddy}/d\sigma$. We plan to separate the two components following the decomposition along and across mean density surfaces proposed by Ferrari and Plumb (2003). These estimates will be compared with those inferred from eddy diffusivities and property gradient fields. By sorting the flux estimates over time and space, we hope to constrain our estimates of $A_{eddy}$ and $dD_{eddy}/d\sigma$ based on $K_{eff}$.

Analysis of the drifter observations by Lumpkin in collaboration with Kelly will focus upon estimating eddy fluxes in the mixed layer. Eddy heat fluxes will be calculated in two methods. The first follows Swenson and Hansen (1999), in which drifter estimates of temperature, velocity and Lagrangian time rate of change in temperature are averaged in space and combined to derive the magnitude of the eddy heat fluxes as a residual. The second method, previously applied in the Kuroshio Extension region by Niller et al. (2003), involves a synthesis of drifter, wind and altimetry observations to calculate time series of fluctuation velocity and temperature. The relative strengths of these data are exploited in this approach: dense, eddy-resolving, in-situ, but inhomogeneous drifter observations of total near-surface currents; temporally homogeneous, spatially regular, but relatively coarse geostrophic velocity anomaly estimates from altimetry. Ekman-removed currents (Ralph and Niiler, 1999) will be derived from the drifter trajectories and wind, and before synthesis will be compared to concurrent velocities calculated independently from Geosat Follow-On (GFO) and JASON-1. Correlations will be examined as a function of smoothing scale and of parameters in the Ekman model. Results will quantify the significance of non-Ekman ageostrophic velocities, and will guide the subsequent synthesis used to produce the spatially and temporally-interpolated fluctuation time series. At synoptic scales, these fluctuation velocities combine with fluctuations in mixed layer depth to produce eddy advective mass fluxes. During the period of peak EDW formation, these fluxes will be calculable directly from the dense suite of drifter and subsurface observations focused along the Gulf Stream front.

Toole and Edson will analyze the ASIS/FILIS data and Weller will investigate the surface mooring data to infer eddy fluxes and their divergence. They will use the ocean data from the mooring to identify the response to local forcing, examine how well 1-D budgets and models perform (e.g. Price et al., 1986), quantify the role of local wind driven mixing and convective deepening and point out, by the lack of closure, the importance of lateral advection. In contrast with the mooring, the effects of mean lateral advection will be minimized by the Lagrangian nature of the ASIS/FILIS. Thus, differences between the air-sea heat and buoyancy exchanges and the changes in ocean storage following the spar system will manifest eddy flux divergences, e.g. $\nabla\cdot\overline{v'T'}$.

All of the above will be compared to analyses of the regional, eddy-resolving model by Marshall, carried out in the same manner.

#### 4.2.5 Diapycnal mixing: $D_{int}$

Gregg will take the lead analyzing the microstructure data. After quality control, the microstructure data will be used to estimate turbulent dissipation rates and in turn, turbulent diapycnal eddy diffusivities. The microstructure data will be partitioned in space/time to form a composite time series about the drifting spar, and repeated cross-front sections. The finescale CTD and velocity data that will be collected in conjunction with the microstructure sampling will be used to investigate the relationships between scales of variability. In the stratified ocean interior a strong relationship exists between the characteristics of the internal wave field and the rate of turbulent energy dissipation (Gregg, 1989; Polzin et al., 1995). We will examine if these relationships are also valid in the sheared Gulf Stream region during strong buoyancy forcing. If finestructure and microstructure are found to be related in this environment, we will be able to infer longer, more complete timeseries of diapycnal mixing rates from the Moored Profiler data from FILIS and the two subsurface moorings (the latter efforts to be in collaboration with Toole and Sloyan).

These new microstructure observations, when analyzed in conjunction with past measurements from the Sargasso Sea and recent tracer dispersion studies, will allow us to estimate diapycnal volume transports — $w^*$ in Eq.(4) — in terms of inferred diapycnal diffusivities and large-scale maps of the EDW bounding isopycnals. This will lead Gregg to estimates of $D_{int}$.

## 4.3 Subduction and dispersal: $A$ and $V$

Wind stress estimates from a variety of sources will be used by Kelly and Lumpkin to derive Ekman transports about the EDW ventilation window. Most important will be the scatterometer fields and surface drifter data and the analyzed winds from atmospheric models (Samelson), the latter having incorporated our buoy and ship data. We expect substantial divergence/convergence arising from wind stress changes across the GS. The surface stress will vary on the scale of the flow because it is affected by the ocean flow (Cornillon and Park, 2001, Kelly, et al., 2001), and by changes in the stability of the atmospheric boundary layer (e.g. Chelton et al., 2001). Such gradients can perturb what might be considered a uniform Ekman flow into one with substantial (20 $\mathrm{m\,d^{-1}}$) vertical motion, of the same order of magnitude as that generated adiabatically due to meanders (Bower and Rossby, 1989). Moreover, the vorticity of the surface geostrophic flow (Niiler, 1969) needs to be considered as it is both large and has spatial structure associated with the flow. We note that vertical velocity at the base of the surface Ekman layer within strong geostrophic flows is given by $\nabla \times \left(\frac{\tau}{(f+v_x-u_y)}\right)$. In addition to Ekman calculations, estimates of ageostrophic flow will be made by Joyce using SeaSoar and ADCP data, and a combination of drifter and altimeter observations. The vertical structure of the Ekman and geostrophic flows in the upper ocean will be examined by Weller using the 2 year time series from the surface flux mooring.

We will interpret $A_{eddy}$ through the concept of bolus fluxes ($\overline{\mathbf{v}'h'}$), the correlation between horizontal velocity and layer thickness anomalies ($h' = \frac{b'}{b_z}$). Information on this term will be provided by Fratantoni, Sloyan and Talley through combined analysis of bobber float, MP data, ARGO float and hydrographic surveys. We will focus the analysis on the northeastern area of our experimental domain where newly ventilated EDW enters the subtropical gyre. Here MP data and bobber float data will provide much of the information. This will be compared to estimates of $A_{eddy}$ from models using $K_{eff}$ and, in eddy-resolving models, from $\overline{\mathbf{v}'h'}$.

The remaining term, $A_{mean}$, will be estimated by Talley and collaborators through a combination of hydrography (from profiling floats, ships, and moored profilers), altimetry and direct velocity measurements (floats and ships). One control volume we will use for this analysis is defined by the Oleander VOS line between New Jersey and Bermuda and the pair of moored profiler moorings extending northeast from the island. The transport divergence/convergence by density classes calculated from the Oleander and MP/float layer transport estimates will provide an estimate of mean density transformation and formation of EDW south of the Gulf Stream.

**Subduction rates: float estimates**

Following Kwon and Riser (2004a), Frantantoni, Sloyan and Talley will infer the annual subduction rate of EDW by computing $V$ and $dV/dt$ on seasonal time scale from ARGO and bobbing floats and available hydrography allowing us to connect $S$, $A$ and $F$ together through Eq.(1 &

2). Whereas Kwon and Riser used a stratification index to define the spatial bounds of the EDW, we will also use isopycnals (isotherms), and explore alternative mapping approaches (e.g. estimating layer thickness from raw profile data and then map). We will begin with a reanalysis of the publically available data including float and hydrographic data obtained, and transition to the CLIMODE data when it comes on line.

Float (bobber and augmented ARGO) profiles will provide monthly maps of mixed layer depth, EDW volume, temperature, salinity, oxygen, and thermocline structure. The cruises will provide some transect information. CTD profiles from ongoing programs (e.g. Bermuda time series, CLIVAR repeat hydrography) will also be incorporated. These products will provide the overall thermocline shape and circulation in which the EDW is embedded. We will also work with historical float and hydrographic data (e.g. Talley and Raymer, 1982; Talley, 1996; Joyce and Robbins, 1996; Lozier et al.,1995; Owens 1991; Molinari et al 1997; Kwon and Riser, 2004a,b), to provide the climate context for the intensive field phase (North Atlantic Oscillation phase, mixed layer properties and vigor of EDW formation, strength of recirculation and dispersal of EDW).

Subduction estimates from the hydrography, float and MP data are based on the kinematic definition of subduction. This estimate of subduction will be reconciled with transformation and formation estimates from the winter survey and flux mooring, and modeling efforts. Through our combined efforts we will address the discrepancies between the subduction rates suggested by the different techniques. Marshall and Dewar will diagnose their regional and idealized models in the same framework, as described in detail in Marshall et al (1999). The floats launched at the MP mooring site will tag EDW that subducts into the recirculation gyre. The floats (ARGO and bobber) recirculating in the gyre will document its thickness and property evolution within the recirculation gyre.

Finally, SSH has been shown in this region to be a proxy for upper ocean heat content and to be negatively correlated with EDW volume (Y-O Kwon and S. Dong, personal communication). Thus Kelly will focus on understanding the interannual modulations of the mode water processes that can be inferred from remotely sensed and ongoing in situ measurements. These analyses will be undertaken in the light of flux estimates, velocity measurements, and dispersion estimates.

**Subduction rates: SeaSoar estimates**

Cross-frontal vertical and horizontal motions will be constructed from SeaSoar data, as in previous studies of meandering flows such as the Gulf Stream (Lindstrom et al, 1997), Kuroshio Extension (Joyce et al., 2001b), and Azores fronts (Rudnick, 1996). The latter was based on the quasi-geostrophic $\omega$-equation, used in meteorology to diagnose vertical motion in fronts. However, the complication of strong cooling and potentially strong Ekman convergence/divergence will require additional attention here. Conventional application requires an estimate of the geostrophic flow: we can estimate these from the SeaSoar and the deeper ADCP data together with lateral vorticity/density advection estimated from each spatial survey. As noted above the system will not be adiabatic: density will be changing due to buoyancy loss at the surface and sub-surface mixing. Surface estimates of this buoyancy flux will be available from shipboard sensors (Edson) and turbulence measurements (Gregg) will be made to enable an estimation of the diabatic processes occurring in the water column. Our data-based effort to exploit the $\omega$-equation dovetails nicely with related modeling work being carried out under the CPT lead by Vallis and Hallburg — see Section 3.4.

### 4.4 Walin synthesis (all participants)

An overall CLIMODE group goal is to combine our estimates of the various terms in Eqs.(2) and (1) based on observations and diagnosed from the eddy-resolving regional model to judge if the

Walin balances close within our measurement uncertainties. More scientifically interesting than the actual sums will be discovering which terms are dominant in the balances. Past results seemingly do not close, with formation rates based on climatological air-sea fluxes and outcrop positions greatly exceeding seasonal EDW volume changes and residence times. While some narrowing of the discrepancy is expected with our improved estimates of air-sea transformation (both through improved BA flux formulas and analysis of synoptic fields), we hypothesize that lateral eddy fluxes will be determined to play a significant role in the budget. CLIMODE will investigate the mechanisms responsible for those lateral fluxes and will test with theoretical and modeling tools the implications of these processes for the dynamics of the North Atlantic circulation.

Ultimately we would hope to constrain our regional eddy-resolving model with CLIMODE observations and compute terms in the Walin balance from this constrained model driven by air-sea fluxes from the regional atmospheric model.

## 4.5 Parameterization in climate models

The success of models and their parameterization schemes will be examined using observations made in CLIMODE. Because of the importance of mode water formation to the large scale three-dimensional circulation of the ocean, and because of the large export of heat carried north from the tropics into the atmosphere in this region, it is important to improve the performance of oceanic and atmospheric climate models. This work will fall into two areas. First, the discrepancies in the Walin budgets and between ocean model results and ocean observations will be used to examine the success of parameterizations of eddies. By the time of the synthesis phase of CLIMODE the CLIVAR CPT group focusing on the interaction of eddies with mixed layers will have made progress at transitioning improved eddy parameterizations schemes to ocean models. The CLIMODE group has members also involved in the CPT: Ferrari, Marshall and Dewar plan in the synthesis phase to follow up on the CPT by focusing on the interaction of eddies with mixed layer to document the sensitivity of air-sea coupling, mode water formation and mixed layer-thermocline interaction to the 'tapering' profiles used in current eddy parameterization schemes. We expect the CLIMODE field program to become a reference data set for testing mixed-layer models, because it will provide accurate estimates of surface fluxes, high-resolution horizontal T-S gradients, and diapycnal turbulence. Extant data sets used to validate mixed-layer models do not carry information about the horizontal structure on eddy scales. We will also collaborate the scientists of the national modeling centers (Gent and Large at NCAR and Griffies and Hallberg at GFDL) who will have been the recipients of the improved parameterizations passed on by the CPT. We expect the CLIMODE observations to become the reference data set for testing these numerical schemes, because there are no other comprehensive measurements of mixed-layer evolution that address lateral, as well as vertical processes.

In a similar way, we (Edson, Weller, Samuelson, and Skyllingstad, in particular) will work with atmospheric modeling centers (ECMWF, NCEP, NCAR) to examine the success of atmospheric models at realistically simulating the air-sea exchanges and atmospheric boundary layer variability and structure over the CLIMODE region. This will begin during the experiment with the sharing of the withheld, near-real time data from the 2-year flux buoy and the radiosondes from the ship. It will continue after the fieldwork with diagnosis of the errors in the models and interaction about how improvements to surface flux and boundary layer parameterizations can be made and whether or not these improvements can successfully be incorporated in research and operational models.

# 5 Management of CLIMODE

## 5.1 Program Coordination

Terry Joyce (WHOI) and John Marshall (MIT) will have overall responsibility for the CLIMODE program, continuing their preparatory work of organizing investigator meetings, developing and presenting CLIMODE to the Atlantic CLIVAR panel and US CLIVAR steering committee and organizing the planning and writing of this proposal. Joyce will oversee the observational program; Marshall will coordinate the theory and modeling activities as well as the interaction of CLIMODE with the CLIVAR Ocean-Mixing CPT. Towards this end, CLIMODE participants will regularly meet with the ocean mixing CPT investigators to help insure that our results translate into model improvements. With PI's distributed across the US, we have scheduled a series of CLIMODE meetings to facilitate program coordination and early dissemination of results. Program and cruise planning meetings will be held in fall 2004 (MIT) and 2005 (WHOI). Late in 2006 we will arrange to meet at a national meeting (e.g. AGU) to refine plans and present early results. We have targeted the Ocean Sciences meeting in early 2008 to present major individual results. Scientific synthesis will become the main focus in the final two years of the program, organized around an PI meeting in Spring 2008 designed to foster collaborative studies. Results of collaborative work will be presented at an international meeting (e.g. 2009 EGS), with the goal of stimulating international participation in further synthesis work. Finally, CLIMODE participants will regularly meet with the ocean mixing CPT to help insure that our results translate into model improvements.

## 5.2 Data Management

To facilitate the early exchange of CLIMODE data we will develop a web-based data archive. Data sets and accompanying meta-data will be made available to investigators as soon as possible after their acquisition, and to the general community by the start of project year 5. Terry Joyce, with the assistance of Jane Dunworth (WHOI Senior Information Systems Assistant), will oversee the data base and organizing access.

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