C PROJECT DESCRIPTION

C.1 RESULTS FROM PRIOR NSF SUPPORT

H.T. Rossby and D. Hebert: Dynamical Adjustment along Gulf Stream Trajectories (OCE 9314480, $356,000, 1 Nov 94 – 30 Oct 97)

In 1995, two cruises were undertaken to examine the dynamical adjustment of water parcels travelling from a Gulf Stream trough to crest. The two goals of this experiment were to determine how the different components of the potential vorticity (e.g., horizontal shear) change as the water parcel (represented by an isopycnal float) travels from a meander trough to a crest and how the potential vorticity changes when the water parcel is expelled from the Gulf Stream.

The experiment consisted of two components. The first was the deployment of a $f/h$ float that measured the separation between isopycnal and tracked acoustically from the ship (which we will call the Swallow float after the pioneering work of John Swallow for this technique). These floats are isopycnal floats that cycle between three density surfaces. The second component consisted of the deployment of a dozen standard $f/h$ RAFOS floats in the core of the Gulf Stream (these are used in the ballasting comparison in the main text).

With real-time directional and differential, and sometimes p-code, GPS navigation, the ADCP and CTD data were used to predict where an isopycnal float if deployed (i.e., a pseudo-float) would go. CTD and ADCP surveys were made around the predicted float location to map out the density and velocity fields, allowing us to estimate the potential vorticity terms. An analysis of the potential vorticity components and their variability as the floats (both Swallow and pseudo) are ejected from the Gulf Stream has been completed (Rajamony et al. 1998).

The CTD and ADCP data from these cruises have been submitted to the appropriate data centers.


This project consisted of a field program with Lagrangian floats (Rossby) and a suite of modeling studies (Rothstein), all designed to address the structure of the North Atlantic Current (NAC) and the exchange processes across it. One hundred acoustically tracked isopycnal RAFOS floats were deployed in and to either side of the western boundary NAC front in summer of 1993 and falls of 1993 and 1994 with over a 90% return rate. The floats were tracked for 10 months and were ballasted for either the 27.2 or 27.5 $\sigma_0$ surface. All floats cycled to density surfaces $\pm 0.1 \sigma$ units apart to measure stretching vorticity. The data consist of three-dimensional tracks that cover the western boundary portion of the NAC intensively for an approximate two-year period.

The path of the NAC between the Tail of the Grand Banks where the Gulf Stream turns north, and the Northwest Corner at 52°N consists of a series of steep (large amplitude, short wavelength) stationary meanders whose locations are clearly linked to topographic features such as the Southeast Newfoundland Rise, the Newfoundland Seamounts and Flemish Cap. The meanders vary in amplitude, but do not propagate (as they do in the Gulf Stream) due to the topographic control (Rossby 1996). The permanent meander structure observed in the float data emerges clearly in high resolution hydrographic analysis (Kearns and Rossby 1998, 1999). Five-year prognostic Princeton Ocean Model runs, using the Kearns and Rossby (1999) climatology, reproduce the pattern of stationary meanders described by the floats. No 44°N trough develops when the Newfoundland Seamounts are removed, reinforcing the view that the meandering results primarily from topography and not instability processes (Rowley 1999). Exchange of waters
from the NAC to surrounding waters occurs almost exclusively upstream of meander extrema on the convex side, and conversely to the NAC from the surrounding waters downstream of meander extrema. Exchange across the entire current occurs only very rarely (Rossby 1996; Dutkiewicz et al. 1999a).


We seek to identify the pathways of warm-water flow and the sites and mechanisms of cross-frontal exchange using a proven Lagrangian tool. For this purpose, we have deployed isopycnal RAFOS floats in the vicinity of the subpolar front to explore the structure of the front which separates the cold, fresh subpolar water to the north and the warm, saline subtropical water to the south, and to address the sites, mechanisms, and rate of exchange between the two gyres.

A total of over 80 floats have been deployed across the subpolar front at 37°30’W in the Fall of 1997 and the Summer of 1998. All of the floats obtain navigational information once a day, make a measure of dissolved oxygen once a day, and make measurements of temperature and pressure every four hours (Rossby et al. 1998). The floats are isopycnal and were ballasted for the 27.5 $\sigma_0$ surface. The bulk of the floats have 18 to 22 month missions and have surfaced/will surface from summer 1999 to early 2000.

While the RAFOS floats complete their missions in the North Atlantic, we have continued the analysis of the NAC float experiment (see above). The NAC RAFOS float experiment provided excellent temporal and spatial coverage on isopycnals in the Newfoundland Basin, resolving the spatially-dependent mean flow field.

C.2 INTRODUCTION

Mixing in the ocean has always been separated in two components: diapycnal and isopycnal. In the last few years, significant effort has been made to determine the diapycnal mixing in the ocean interior. For example, the North Atlantic Tracer Release Experiment (NATRE) confirmed that microstructure-based estimates of diapycnal eddy diffusivities on the order of $10^{-5} \, \text{m}^2 \, \text{s}^{-1}$ are typical for the interior of the ocean away from topography. The stirring and mixing of water along isopycnals is poorly understood and even the use of stirring and mixing depends on the horizontal scales that is being resolved. The effective isopycnal eddy diffusivity is believed to be scale-dependent (i.e., process-dependent). Garret (1983) suggested that the dispersal of a tracer on an isopycnal can be separated into three stages. During the first stage, the horizontal scale of the tracer patch is much smaller than that of the eddies that would stir (strain) the tracer. During this period, the growth of the patch can be modelled as a mixing (diffusive) process for which a horizontal eddy coefficient $K_s$ can be defined. Young et al. (1982) determined the shear dispersion eddy coefficient to be $K_s \propto \left( a/\omega \right)^{1/2} k_z$ where $a$ is the vertical shear and $\omega$ is the frequency of the vertical shear. Assuming the vertical mixing is due to near-inertial waves and using a $K_z = 10^{-5} \, \text{m}^2 \, \text{s}^{-1}$, $K_s = 10^{-2} \, \text{m}^2 \, \text{s}^{-1}$. The first stage of horizontal mixing continues until the scale on which the mesoscale eddy shear is as important as the smaller scale shear dispersion, which is $L = (K_s/\gamma)^{1/2}$ where $\gamma$ is the rms strain rate on the order of $10^{-6} \, \text{s}^{-1}$ (Garrett 1983). Thus, $L = 100 \, \text{m}$. The second stage of horizontal dispersion can be thought as a balance between the strain field of the eddies stretching out the dye lines, making them thinner, and the small-scale diffusion tending to make the streaks wider. The area covered by the dye patch grows exponentially. This growth continues until the scale of the tracer exceeds the mesoscale eddy size. The dispersion at this time, the third stage, can be modelled again as a diffusive process (e.g., a $K_e$ on the order of $1000 \, \text{m}^2 \, \text{s}^{-1}$ is usually found).

In May 1992, a patch of sulfur hexafluoride was released in the eastern North Atlantic in a region of low mesoscale eddy energy (Ledwell et al. 1998); the North Atlantic Tracer Release Experiment (NATRE). After the dye was injected on the isopycnal surface, it was sampled four times over a 30-month period. As well, ten SOFAR floats were deployed in the area (Sundermeyer and Price 1998). During the first stage of the horizontal dispersion of the tracer (which corresponds to stage 2 of the Garrett (1983) model), Ledwell et al. (1998) found that the eddy diffusivity was two orders of magnitude larger than that predicted for shear dispersion by internal waves. They suggested that processes at subinertial frequencies are responsible for dispersion at scales less than 10 km. It has been suggested that vortical modes (described below) are responsible for this dispersion (Polzin et al. 1999). One of the problems with the dye study to determine the horizontal dispersion is that the velocity (and strain) of the eddy field has to be estimated. From one survey to another, you cannot determine whether you are observing the same dye streaks or not. Ten SOFAR floats were deployed to assist in tracking the dye. However, the number of floats were too few to allow reliable estimates of the horizontal dispersion. Sundermeyer and Price (1998) used this data to provide a calibration of a 2-D quasi-geostrophic model which they used for their dispersion calculations.

Thus, the tracer release experiment has shown that the horizontal dispersion on scales less than 10 km is higher than predicted. The process responsible for this dispersion is unknown. It
has been speculated that vortical modes are responsible although not much is known about them (see below). The dynamics of dispersion on scales less than the mesoscale eddies appears to be in agreement with the ideas of Garrett (1983) but the coarse temporal sampling of the dispersing dye cannot confirm this; numerical models have been used (Sundermeyer and Price 1998). It is evident that a study with a significant number of isopycnal floats will be able to address the question of what processes and dynamics are responsible for horizontal dispersion on scales less than the mesoscale eddies. We propose to address this problem in this project.

Vortical mode Research

Forms of motion which exhibit nonzero amounts of potential vorticity have been called vortical modes (Müller 1984; Müller 1995). The best known forms of “vortical” modes are planetary geostrophic motion and eddies such as Meddies (Hebert et al. 1990). Müller (1984) suggested that even fluctuations with vertical wavelengths less than a few hundred meters and horizontal scales less than a few kilometers can carry potential vorticity anomalies. Determining whether the ocean fluctuations has potential vorticity (i.e., vortical modes) or not (i.e., internal waves) is difficult. Using data from a tri-mooring, Müller et al. (1988) showed that small-scale vortical modes exists with limited success. Eulerian measurements have the problem of Doppler smearing of near-inertial waves and fine structure on these small scales. An array of Lagrangian floats would reduce any Doppler smearing. Recently, using an expendable current profiler (XCP) survey, Kunze and Sanford (1993) found submesoscale vortical modes with vertical wavelengths of 50–400 m and horizontal wavelengths of 7–15 km. Lien and Müller (1992a) formulated consistency relationships for internal waves and vortical modes. Kunze (1993) and Polzin et al. (1999) have used these relationships to show the presence of small-scale vortical modes.

Polzin et al. (1999) has suggested that randomly oriented vortical modes could be responsible for the isopycnal dispersion of the tracer in the NATRE experiment on horizontal scales of 1 km to tens of kilometers. Vortical modes with horizontal scales of 1 km and velocities of 0.5 cm/s would provide the horizontal diffusivity \(K_H \approx \frac{ul}{5} = 5 \text{ m}^2 \text{ s}^{-1}\) needed to agree with the dye data.

Isopycnal Dispersion by Floats

Subsurface floats as well as surface drifters have been effectively used to study particle dispersion in the oceans. However, in most of previous studies, the data are too sparse in time and space to produce accurate statistics. That is, whether the limited observations can effectively decomposite the mean flow from the eddy fields as described below. There is also the question whether the floats follow neutrally buoyant Lagrangian particles in the ocean. Earlier SOFAR and RAFOS floats were designed as isobaric (e.g., Freeland et al. 1975; Riser and Rossby 1983; Rossby et al. 1983; Schmitz et al. 1988; Richardson et al. 1989; Armi et al. 1989; Richardson 1993), and these isobaric floats as well as (drogued) surface drifters are still widely used with the major aim to map out flow patterns. As the authors were aware, the isobaric floats and surface drifters are not strictly Lagrangian followers, since water parcels freely move along isopycnal surfaces (or neutral surfaces as coined by McDougall (1987) to emphasize that potential density surfaces are not true isopycnal surfaces). The discrepancy between isobaric and isopycnal surfaces manifests in dynamic front/baroclinic current regions (Zhang et al. 1999). For dispersion studies, the RAFOS Laboratory at the University of Rhode Island developed a new technique to make the floats essentially specific volume anomaly followers (Rossby et al. 1985; Rossby et al. 1994). When refer-
enced to local (regional) values, both specific volume anomaly and potential density surfaces are good approximations to an isopycnal surface (McDougall 1987; Zhang and Hogg 1992). More details of the properties of these floats will be presented later. These ‘isopycnal’ floats have been successfully applied in the Gulf Stream region (Bower and Rossby 1989; Song et al. 1995; Song and Rossby 1997; Rajamony et al. 2000) and the North Atlantic Current region (Rossby 1996; Carr et al. 1997; Prater and Rossby 1999; Carr and Rossby 2000; Dutkiewicz et al. 1999; Zhang et al. 1999).

In the early days of float observations, data coverage in space was usually quite sparse due to the small number of floats available for use. To calculate the statistics for the eddy field, often an ensemble mean over time as well as over the whole observation domain had to be used to obtain the background mean velocity (e.g., Freeland et al. 1975; Sundermeyer and Price 1998). This practice assumes that the mean flow is homogeneous and stationary over the entire observational region. The ignored mean current shears, when they exist, will bias the eddy statistics (Zhang et al. 1999; Fig. 1). To reduce this bias, later studies attempted to bin the data in large boxes to calculate the mean velocity (Riser and Rossby 1983; Krauss and Böning 1987; Spall et al. 1993; Poulain et al. 1996). Even those boxes are perhaps too large to resolve the mean current structures (e.g., Box 5 and 6 of Krauss and Böning in the North Atlantic Current region). In a recent study in the North Atlantic Current/Newfoundland Basin region (Rossby 1996), the mean structure of the North Atlantic Current was resolved by bining the observational data from about 100 isopycnal RAFOS floats over about a two-year period. The successful decomposition of the mean current structure from the total flow results in eddy statistics which are nearly isotropic (Zhang et al. 1999; Fig. 2). If the mean flow structure is unresolved (e.g. using a box bin size of 2 degrees or larger), the statistics for the eddy field are biased towards larger values and highly anisotropic (Fig. 1). Another example of successful mean-eddy decomposition is seen in the tropical Pacific (Bauer at al. 1998).

The lateral (either horizontal or isopycnal depending on the type of observations) eddy diffusivity can be compared to the (a) eddy velocity variance (or eddy kinetic energy, EKE) and (b) rms eddy velocity respectively for several studies using floats and drifters (Fig. 2). For all of the studies except the surface drifters and isopycnal floats, on the order of 10 floats were used in the study of lateral dispersion. Using Taylor’s (1921) dispersion theory, the slope in Fig. 2a represents the integral time scale and the slope in Fig. 2b the integral length scale. The data in Fig. 2a can be separated into four groups. The data with the largest time scale (11.5 days) are exclusively from observations in the subtropics (Rossby et al. 1983; Spall et al. 1993; Sundermeyer and Price 1998) and mostly from deep (> 500 m) isobaric floats. The large time scales are associated with the lowest eddy kinetic energy ($< 100 \times 10^{-4}$ m$^2$/s$^2$) and mostly low eddy diffusion rates in these regions/depths ($\leq 2 \times 10^3$ m$^2$/s) except for three points from Rossby et al. (1983). The data with an integral time scale of 2.3 days are mostly from the lower level isopycnal NAC floats (circles, Zhang et al. 1999), augmented by some data from the Nordic seas surface drifters (diamonds, Poulain et al., 1996). These moderate time scales are associated with the moderate EKE ($< 250 \times 10^{-4}$ m$^2$/s$^2$) and eddy diffusivity ($\leq 5 \times 10^3$ m$^2$/s). The data with the second shortest integral time scale (1.7 days) are mostly from the upper level isopycnal NAC floats (stars, Zhang et al. 1999) and the zonal component of the Krauss and Böning data in the Newfoundland Basin. In these regions, EKEs are high ($300 \times 10^{-4}$ m$^2$/s$^2 \leq EKE \leq 600 \times 10^{-4}$ m$^2$/s$^2$) with larger diffusion rates (can be as large as $10 \times 10^3$ m$^2$/s, but mostly of $5.5 \times 10^3$ m$^2$/s). The four data points around the smallest time scale...
of 1.2 days are all from the observations near the Northwest Corner: one pair from the upper level isopycnal NAC floats (stars, Zhang et al. 1999) and the other pair from Krauss and Böning (1987). This region has the largest EKE (of $600 - 700 \times 10^{-4}$ m$^2$/s$^2$) and with moderately large eddy diffusion rate (of $7 \times 10^3$ m$^2$/s). In general, the northern North Atlantic upper ocean has relatively constant integral time scales (1 to 3 days), even with a large range of EKEs and eddy diffusion rates ($50$ to $700 \times 10^{-4}$ m$^2$/s$^2$ and $1$ to $10 \times 10^3$ m$^2$/s respectively).

The data from surface drifters and the thermocline isopycnal floats can be separated into two length scales. The low variance regime (with $rms$ velocity $< 15 \times 10^{-2}$ m/s) has a smaller length scale of 20 km, and the larger variance regime (with $rms$ velocity $\geq 15 \times 10^{-2}$ m/s) has a larger length scale of 30 km. Again overall, the northern North Atlantic upper ocean has relatively constant integral length scales (20-30 km), even with a large range of EKEs and eddy diffusion rates.

Sorting the data points according to the type of float or drifter, we see that the data from the large number of the NAC isopycnal floats show a much tighter relationship to the Taylor dispersion theory; all the other data sets are more scattered. The data from the subtropical floats (Rossby et al. 1983; Spall et al. 1993; Sundermeyer and Price 1998), which are deep isobaric floats except the ones in Sundermeyer and Price (1998), show a large range of time and length scales (if lines are
Figure 2. Eddy diffusivity versus velocity variance (a) and versus rms velocity (b). The eddy diffusivities are either isobaric, isopycnal, or geopotential depending on the observations. SOFAR float results are represented by the solid dots (Rossby et al. 1983), triangles (Spall et al. 1993), and squares (Sundermeyer and Price 1998). Surface drifter results are represented by diamonds (Poulain et al. 1996) and crosses (Krauss and Bönig 1987). Isopycnal float results of Zhang et al. (1999) are represented by stars (σθ = 27.2) and circles (σθ = 27.5). Dark solid symbols are the zonal component; gray dotted symbols the meridional component. See text for discussions on the slope lines.

drawn passing the coordinate origins and the data points according to the Taylor theory), ranging from less than 20 km to 80 km in length scale. The data from the Nordic seas surface drifters (Poulain et al. 1996) show much peculiar behavior. Their data in the weak EKE subregions, i.e. the Iceland Plateau and the Norwegian Basin (the diamonds with EKE < 200 × 10^{-4} m^2/s^2 in Fig. 2a and with rms velocity < 12 × 10^{-2} m/s in Fig. 2b), show a time scale and a length scale which are consistent with our lower level isopycnal float observations. However, their data in the strong EKE subregions (also strong in MKE), i.e. the Norwegian North Atlantic Current, the Iceland-Faroes Front, and the Lofoten Basin, show large variations in eddy diffusivity but with about the same level of EKE and rms velocity (~ 200 cm^2/s^2 and 14 cm/s respectively). It is difficult to define a line passing through these data points and also through the coordinate origin. In other words these data points show large discrepancy from the Taylor dispersion theory. Note that these strong EKE subregions are also regions of strong dynamic fronts with strong mean velocity shears. As pointed out above, in these regions (1) there is strong discrepancy between the geopotential surface and isopycnal/neutral surface on which Lagrangian particles freely move; and (2) the eddy statistics are sensitive to how the data are binned in computing the mean velocity. In summary, isopycnal floats are most appropriate for dispersion and eddy statistics studies, especially when they are used in large numbers.

C.3 PROPOSED RESEARCH

The proposed work consists of two deployments, separated by 6 months, of a cluster of floats on two isopycnal surfaces. The isopycnal floats will be tracked acoustically over their 1 year mission. The goal is to determine the process(es) and dynamics responsible for isopycnal dispersion on horizontal scales from 1 km to more than 100 km. For this study, we wish to start in a region where the mean advective velocity is small compared to the effect of the isopycnal diffusivity. For
a mean velocity of 1 cm/s and eddy diffusivity of 1000 m²/s, horizontal diffusion dominates the mean advection for scales smaller than 100 km. The standard isopycnal RAFOS floats measure temperature and pressure. Recently, an oxygen sensor has been added to the float (Rossby et al. 1998). Thus, an optimal location would be one where both oxygen and temperature can be used as independent tracers.

Choice of Experiment Location

We have decided to investigate isopycnal stirring and mixing to the south of the North Atlantic Tracer Release Experiment (NATRE) since the presence of a low oxygen tongue that extends thousands of kilometers to the west of Africa (Fig. 3) provides an additional tracer to temperature that can be measured by the RAFOS floats. Like the NATRE site this region has a weak mean current and relatively low mesoscale eddy energy. The mean flow from $3^\circ \times 3^\circ$ historical hydrographic data (Siedler and Stramma 1983) is on the order of 1 cm/s. From transient-tracer fields, an eddy diffusivity of 1700 m²/s is estimated south of 29°N, compared to 2900 m²/s to the north (Thiele et al. 1986). For an advection–diffusive flux divergence balance over the $\beta$-triangle region centered at 27° N, Armi and Stommel (1983) found a horizontal eddy diffusivity $K_H$ of 500 m²/s. With advection and horizontal diffusion rates of these orders, the floats should remain well within range of our sound sources and survey region (see later) for more than 1 year. As seen in Fig. 3, the isotherms and $O_2$ isopleths are not co-linear. It seems likely that there must be significant mixing (either diapycnal or isopycnal) occurring over this region and these two tracers should help us determine these rates.

The low $O_2$ tongue is due to both isolation of the water (i.e. direct ventilation) and enhanced productivity and organic regeneration in the coastal upwelling off North Africa (Doney and Bullister 1992). The lowest $O_2$ values are found very near the coast, and clearly result from the enhanced biological activity at the surface due to upwelling and subsequent sinking of organic material and its regeneration. The nearly zonal flow from 10–20°N (Stramma and Siedler 1988) advects the low oxygen water westward from the coastal upwelling sites. Since the $O_2$ concentration of the water is increasing as it is advected westward and this surface is well below the euphotic zone, only isopycnal and diapycnal mixing of higher $O_2$ waters (Figs. 3-5) can be responsible for the observed $O_2$ minimum tongue. It has been suggested that ventilation occurs along the northwestward arm of the South Atlantic subtropical gyre followed by eastward advection of the water in the equatorial undercurrent. This higher $O_2$ mixes with the low $O_2$ water (Doney and Bullister 1992). The time scale of this ventilation is on the order of several decades. However, this description of the dynamics responsible for the low $O_2$ tongue assumes a mean westward zonal current which not been proven definitely.

Two relatively recent hydrographic sections (Tsuchiya et al. 1992; Friedrichs and Hall 1993) demonstrate that there must be some isopycnal mixing occurring. Along 11°N, temperature and $O_2$ change on isopycnal surfaces (Fig. 4). We also notice that the minimum of $O_2$ occurs at $\sigma_1 \simeq 31.5$ ($\sigma_0 \simeq 27.0$). A north-south section along 28°W shows that the low $O_2$ level extends only a few degrees north and south of its axis at about 10°N (Tsuchiya et al. 1992). These sections accord well with earlier depictions of the low-$O_2$ tongue in the climatological data (Fig. 3).

The north-south section also provides useful information on the dynamic height field across the low $O_2$–tongue. Relative to 2000 dbars, the north-south variation in dynamic height at 500 dbars...
is roughly 5 dyn. cm with the gradient such that the vertical shear is cyclonic around the $O_2$-tongue, that is, to the east south of the $O_2$ minimum and to the west north of the $O_2$–axis. Based on this individual transect, the axis of the $O_2$–minimum does not appear to be a locus of maximum advection in the sense of the classical core method, but possibly region of minimum advection. Stronger advection takes place both north and south of the low $O_2$ tongue itself. It is possible that the tongue is characterized by a recirculation around it. If so, the temperature field requires isopycnal or diapycnal mixing to be occurring. It may be that this circulation from the transect is due to a mesoscale eddy and another survey might show a different circulation and that the tongue is due to a mean westward advection (Siedler and Stramma 1983). In either case, the dynamics responsible for the structure of the $O_2$ is unknown. In addition to the goals of this dispersion study, we will be able to address these questions with the data collected.

**Vorticity/$O_2$ Measuring Floats**

A major component of this project is the use of RAFOS floats to measure the horizontal
Figure 4. Temperature and $O_2$ on $\sigma_1$ density surfaces for the east-west transect (Friedrichs and Hall 1993) shown as a dashed line in Figure 1.

Figure 5. Temperature and $O_2$ on $\sigma_1$ density surfaces for the north-south transect (Tsuchiya et al. 1992) shown as a dashed line in Figure 1.

dispersion on scales greater than 1 km and to measure the Lagrangian evolution of temperature and oxygen. In addition to obtaining the relative vorticity from the float trajectories, horizontal shear on 1 m scales will be obtained from each float. A critical aspect of deploying these floats is ensure that they are located on the “same” isopycnal surface. Before addressing this, we will briefly review the RAFOS float design and ballasting procedure.

The original RAFOS floats (and the ones commercially-available) are isobaric floats made of glass which has a low thermal expansion coefficient (Rossby et al. 1986). Thus, if the temperature on the isobar that the RAFOS is located changes, the float would not changes its volume (and density) much and would remain at the same pressure. Since the dominant large-scale motion of water parcels in the deep ocean is believed to be along isopycnals, Rossby et al. (1985) developed a float having approximately the same compressibility as seawater, through the use of a compressor, but retaining the low thermal expansion coefficient. Therefore, these floats should stay on isopycnal surfaces even if the temperature on an isopycnal surface changes. We refer to these floats as isopycnal RAFOS floats even though their characteristic is more like having a constant specific volume anomaly. Since the float does change its volume with temperature, the float is not strictly an isopycnal float. As well, the compressibility of the float is constant and matched to seawater at
a particular temperature, salinity and pressure. If an isopycnal parcel changes its temperature and depth, the compressibility of the parcel changes and the isopycnal RAFOS float’s motion would deviate from the isopycnal parcel. Therefore, we design the compresses and ballast the floats for a particular temperature and depth for each project. For example, in the NAC experiment, there are examples of floats changing depth some 500 meters with the temperature remaining constant to within 0.2°C. Given an ambient stratification of approximately 1°C/40 m, this means that the float is tracking the surface to approximately 10 m over a 500 meter depth change. This is quite impressive.

The RAFOS float has been used to monitor other variables while following a water parcel. For example, Hitchcock et al. (1989) placed a solid-state fluorometer on a RAFOS float to monitor changes in chlorophyll $a$. Recently, oxygen sensors were incorporated into some RAFOS floats (Rossby et al. 1998).

Through ONR funding, we have developed a float to measure the very weak diapycnal velocities in the coastal ocean. This isopycnal COastal Ocean Lagrangian (COOL) float uses tilted vanes attached to the float body. As the water flows vertically past this float, a torque is generated and rotates the float. Using a compass to obtain the rotation rate, the vertical velocity can be easily determined (Hebert et al. 1997). During testing of the COOL float, a deployment was made with the vanes oriented vertically. Thus, the float acted as an isopycnal vorticity meter (Fig. 6). Other components within the RAFOS system needed to be removed to incorporated these additional sensors and because of storage space limits, mission length were limited to a few days. Thus, if we wish to use the isopycnal RAFOS platform with $O_2$ sensors and compasses, we need to update the microcomputer and electronics within the float.

As a matter of interest, assuming that the temperature fluctuations measured by the COOL float are due to vertical advection of the mean temperature gradient, we can see that the isopycnal float is isopycnal on short (buoyancy) time scales since the temperature-based displacements and velocities are much smaller than (and incoherent with) the pressure-based ones. The relative vorticity measured by the float can be large for long periods of time (Fig. 6). [The middle panel of Fig. 6 shows the rotation rate smoothed over 15 minutes while the lower panel shows the rotation based on changes over 64 s.] The rotation of the float is not coherent with the pressure fluctuations; thus, the rotation is not due to vertical motion of the float.

Isopycnal Ballasting

A key aspect of studying isopycnal dispersion using the RAFOS floats is to have the floats on the same isopycnal surface as close as possible. We need to make an estimate of the vertical scale of the motion over which the shear dispersion or stirring is occurring. Of course, we don’t know this value but we believe that it is on the order of tens of meters.

There are two errors in the ballasting of a set of RAFOS floats for a particular isopycnal surface – the absolute error between the mean density of the floats and the target density and the variability in the density of the different floats. It is the latter error which concerns us. It is not critical for us to ballast the floats to a particular density surface.

The ballasting procedure presently used at URI is very similar to that described by Swift and Riser (1994). The floats are weighed in air and then placed into water of a known density.
and temperature near the target temperature. Weights are added to the float to make it neutrally buoyant; thus, we know the volume of the float. Chains of known density of the float are hung from the bottom of the float to the bottom of the pressure vessel. As the vessel is pressurized, the float will lift some chain off the bottom to make it remain neutrally buoyant. (The fresh water in the tank is more compressible than the sea water we are trying to mimic.) We measure the vertical displacement of the float and calculate the additional weight that is floating. Normally, this measurement is done for 11 pressures. The compressibility of the float and the weight needed to make the float the target density at the target temperature is determined. Swift and Riser (1994) estimate the uncertainty in careful ballasting of isobaric RAFOS floats to be less than 50 db for floats deployed at 1000 db. This is the absolute error. For our project, we are only concerned with the relative error in density of the floats. This variability in density is mainly due to the variability and uncertainty in the calculated compressibility of the floats. For a set of floats, the range in compressibility is found to be $\pm 0.05 \times 10^{-6}$ dbar$^{-1}$. (The compressibility of seawater is
approximately \( 4.4 \times 10^{-6} \text{ dbar}^{-1} \). If the pressure of the target density surface is 100 m different from that the ballasting pressure, the floats would have a density variability of \( \pm 0.005 \text{ kg/m}^3 \) (less than 10 m given typical stratification in our area of interest).

Over the years the URI RAFOS float group has improved its isopycnal ballasting skills considerably. In a couple of recent experiments, pairs and triplets of float were deployed at the same time (Fig. 7). For most of these deployments, a CTD cast preceeded the float deployments. Some of the scatter in the pressure differences could be due to internal wave activity. The temperature and density differences could be related to the uncertainty in the temperature calibration. While the temperature measurement circuitry is calibrated, we depend upon the manufacturers guarantee that the thermistors are accurate to within \( \pm 0.1 \text{°C} \); at this time, we do not verify this. The scatter in the data is consistent with the scatter in the compressibility of the floats.

\[ \Delta \sigma_\theta \quad \Delta P \quad \Delta T / (\text{°C}) \]

![Figure 7. The magnitude of the temperature, pressure and \( \sigma_\theta \) differences between pairs and triplets of floats launched at the same time. The temperature and pressure differences are based on the first samples taken after deployment (usually between 3 and 6 h). The \( \sigma_\theta \) difference is based on the float temperature and CTD cast made at approximately the same location and time when available.](image)

The scatter in the ballasting is traceable to the technique for ballasting. A new technique, the use of a strain gauge, is in development that will significantly improve the float weight measurement in the pressure vessel. As well, the URI RAFOS float group is in the process of completing a new assembly, check-out and ballasting facility which will be the finest of its kind. It comprises a new stainless steel tank with external thermostated temperature control, automated pressure cycling, and quick open and close for rapid turn-around. We will be able to control the temperature of the pressure vessel; with the pressure temperature, the water tank is cooled only before and after groups of floats are ballasted. The automated pressure cycling and data acquisition system will allow a much better estimate of the compressibility of the float.

In mass production the mean compressibility of the floats used in the NAC study were found to be within \( 0.5 \pm 1\% \) of target. We consider this to be very good, but it can be improved further. By measuring the compressibility of the glass pipes and compressees separately, we can match them and thereby reduce the scatter significantly. The pipe compressibility has to be measured in a pressure vessel. The compressees can be determined by measuring the spring constant for each
spring. We think this gives us enough freedom to compensate for variations in glass pipe stiffness. (Actually, this effort may not be worthwhile for the proposed study since the expected pressure variations are so minor and we ballast the floats at the working pressure anyway.)

Outline of Research Program

In the first year of this project, the first task will be to upgrade the microprocessor on the RAFOS float to allow control of more sensors, collect more data, process data if necessary, and take advantage of the upgraded ARGOS system. The limited speed, programming memory (16-32K; the lower value is for commercially-available float and the higher value is for modified URI floats), data storage (15-26K) of the 6805 microcomputer in the RAFOS float has made the programming of the system (done in FORTH) sometimes difficult when new sensors are added (only 14 control lines are available) or non-standard RAFOS missions planned. For example, the ARGOS transmitter is driven by code written in both Assembly and Forth and the detection of the arrival time of the acoustic signal from the sound sources is written in machine language. If a different ARGOS transmitter or acoustic signal is used, machine-level programming needs to be done to ensure proper timing in the circuits.

The basic time-tested hardware and electronics of the RAFOS float would remain the same. The only physical change would the replacement of the microprocessor and its circuit board. We would need to re-write all present software — ARGOS transmitter, hydrophone receiver correlator, data acquisition (which we have to do to measure both $O_2$ and compass angle) and standard mission code, most likely in C.

After completion of these tasks, we would start construction of 100 RAFOS floats, all with $O_2$ and vorticity (vanes/compass) sensors. As well, a float launcher based on a modified CTD rosette will be constructed. Sound sources would be purchased and preparation of the mooring by the WHOI Rigging Shop would start.

In the second year, we would continue the construction of the floats and ballast them. We propose to do two deployments of floats separated by 6 months. The mission length would be 1 year (or maybe a bit longer). For each deployment, 25 floats would be deployed as a group on two density surfaces. All floats would have $O_2$ sensors and vertical vanes/compasses. The floats will be deployed in a group using a rosette-like launching device similar to that used by Zenk et al. (2000) for “parking” RAFOS floats on the bottom.

The first deployment of floats would be planned for May 2002. First, the four sound sources would be deployed as shown in Figure 3. The sound sources will ‘pong’ four times a day. While only two sound sources are required for tracking a float, a third source is used to remove a possible ambiguity in the triangulation. A fourth source is used for redundancy in case one of the other sources failed. Also, it is likely that we can obtain better tracking with all four sources available. The absolute accuracy of the float locations will be on the order of 1 km. However, we may be able to resolve the relative distances between floats to a smaller scale. This requires careful synchronization of the float clocks just prior to launch. As well, since the floats are launched at the same location, any time offset can be determined using the first arrival time. For the dispersion calculations, we are only interested in relative position so any drift in sound sources or floats can be minimized. The multiple fixes per day (about 12 fixes per inertial period) will also add in the tracking and its accuracy.
A high resolution CTD survey of a 400 km × 400 km region in the center of ensonified region would be done (CTDs spaced every 50 km; if available, a lowered ADCP will also be used). This survey will allow us to determine the mesoscale eddy field in which the floats are deployed. The vertical (along-isopycnal) velocity can be determined from the quasi-geostrophic omega equation (Lindstrom and Watts 1994; Shearman et al. 1999).

At the center of the survey region, we would do the rapid expendible current profiler (XCP) survey (a 5-10 km radius circle) for determining the vortical mode structure (if present) before the float deployments. The cross pattern used by Kunze (1993) is not optimal for determining the vortical mode structure; a circular sampling scheme is much better (Kunze, pers. comm. 2000). To properly resolve the horizontal wavenumber structure, 9 or more current sensors are required (Lien and Müller 1992). We propose to use 15 XCPs per survey to allow for failed XCP profiles. Dr. Kunze will loan us the XCP data acquisition equipment and assist in its setup on the ship. After this XCP survey, 50 RAFOS floats would be deployed (25 floats at each depth). A second XCP survey would be conducted approximately 24 hours after the first. (The inertial period at this latitude is approximately 3 days.) The two surveys will help us to separate the inertial motion from the subinertial motion. (Dr. Kunze will lend us the equipment to record the XCP data, assist in the set-up of the equipment and processing of the data.)

The second deployment of floats would occur in November 2002. The cruise for this deployment will be the same as the May 2002 cruise except that there will be no mooring work. For approximately 6 months, we will have 50 floats at each target density. The floats from the first deployment would provide the larger scale flow field for the dispersion of the second group of floats from a point.

During May 2003, the first set of floats will surface and start transmitting their data. The second set of float will surface during November 2003 and start transmitting their data.

We will have three different estimates of the horizontal structure of the relative vorticity field. The individual floats will measure the relative vorticity on a horizontal scale of approximately 1 m. The trajectories of the floats will provide an estimate of the velocity field on the two target density surfaces. Middleton and Garrett (1986) describe a method to examine for a preferred sense of rotation using drifters. The horizontal scale of this vorticity field will range from approximately 1 km to the areal extent of the floats which will increase with time. The XCP surveys will provide two submesoscale maps on the two density surfaces as well as other depths down to 1500 m.

Shear dispersion estimates will be made for a wide range of spatial and temporal scales. The trajectories of the floats will be used to determine the dynamics of the process(es) responsible for the observed dispersion. The temperature and $O_2$ time series will aid in the analysis. As well, we should have a sufficient number of floats to be able to make estimates of the along-isopycnal fluxes of temperature and oxygen.

Finally, the ensemble of data should allow us to address the dynamics of the low $O_2$ tongue. Is it due to a slow westward advection of low oxygen water generated near the coast of Africa mixing isopycnally and diapycnally with higher $O_2$ water? Or, is the circulation around the tongue which is a region of stagnant water?
C.4 SUMMARY

In this proposal we develop a program to investigate more closely the processes for the maintenance of these conspicuous hydrographic features. The project will focus on the mechanisms of isopycnal stirring and mixing, but in order to these into their larger scale context the mean flow in the O2 tongue in the vicinity of the float cluster will be estimated from a wider distribution of some floats.
D REFERENCES CITED


Rossby, T., D. Dorson and J. Fontaine. 1986: The RAFOS system, *J. Atmos. Oceanic Techn.*, 3,


F BUDGET JUSTIFICATION

A. Senior Personnel

Hebert, who has a 8-month state supported position, will be responsible for the overall project and his main interest lies in the dispersion dynamics on the submesoscale scales and the vortical modes. Using his previous experience on dispersion by isopycnal floats, Zhang will examine the data for dispersion dynamics and eddy dynamics. Rossby, who has a 12-month state supported position, will investigate the dynamics of the low $O_2$ tongue. While we all have specific interests, we will work together on all aspects of the project.

B. Other Personnel

James Fontaine is responsible for the construction of the RAFOS floats and operating the ballating facility. He will make the modifications to the RAFOS floats for the new microprocessor.

The acoustic tracking and initial processing of the RAFOS data is done by Sandy Fontana. She also assists in the collection and processing of the CTD and ADCP data collected during the cruises. Sea pay (B.6) is included for her.

Stipend and tuition support is requested for a graduate student.

Clerical person will assist in preparation and submission of camera ready copy of manuscript for publications. The preparation will include, but not be limited to, the typing, layout, verification of references, and appropriateness of figure and table captions. Manuscripts submitted in camera ready form results in significant savings in publication costs.

D. Permanent Equipment

We proposed to purchase 4 Webb Sound Sources for this project. Based on mooring designs for both a recoverable and expendible mooring, the expense of the additional hardware and personnel time (we are subcontracting this work to the WHOI Rigging Shop) makes it more cost effective to make the moorings expendible. This analysis does not include the additional expense of ship time. With support from NSF, Rossby and Miller are designing a low cost sound source. The goal is to make an expendible, easy-to-deploy, sound source mooring. During this summer, a prototype of this mooring will be tested. If all tests go well, we will used this sound source instead of the Webb source.

Based on large quantities of RAFOS floats constructed, the cost of a RAFOS float with an updated microprocessor, $O_2$ sensor, compass and vertical vanes is $3,500 per float.

E. Travel

In Years 2 and 3, foreign travel to and from the Cape Verde Islands for 8 people is required to participate in the cruises. In Year 2, two WHOI personnel will participate also; their travel costs are included in WHOI subcontract.
Funds are requested for two person to attend a national meeting, such as the Fall AGU meeting, to present results from this work in Years 3, 4 and 5.

G. Other Direct Costs

*Materials & supplies:* Standard laboratory and cruise supplies for the cruises are requested.

*Publication costs:* Costs are based on a 20 page JPO paper.

*Subcontact:* A subcontract to the WHOI Rigging Shop for the purchase and fabrication of four expendable moorings (Year 1) and their deployment (Year 2).

*Other:*

30 expendible current profilers (XCPs) will be purchase in Years 2 and 3. Dr. Eric Kunze will provide the equipment to record the data. One of his technicians will assist in the setup of the equipment (their travel expenses are included in E). Dr Kunze will also assist in the post-processing of the data.

Based on past experience, the costs of operating and maintaining the ballasting tank facility is $50 per float.

In the first year, prototype boards and supplies are required to interface the new microprocessor with the present RAFOS float hardware.

Funds are also requested for to cover a portion of the maintenance contracts and software licenses on the PIs workstations.

The present RAFOS system is written in FORTH. With the new microprocessor, we will use a new operating system such as C. The cost of converting and testing the new software is based on experience at URI with the Persistor microprocessor.

We will construct a rosette-like float launcher similar to that developed by Zenk et al. (2000), ARGOS user fees for receiving and accessing the float data is included in Years 2 and 3.

Funds are requested to cover copying charges related to the scientific/technical aspects of this project, communication costs related to this project, excluding local line access (Scientific/technical use only).

In Years 2 and 3, shipping costs of transporting the URI equipment to and from the Cape Verde Islands is needed. In addition, funds to cover agent, custom and port handling fees for the URI and WHOI equipment in the Cape Verde Islands is required.