Mesoscale subduction at the Almeria-Oran front. Part 1: ageostrophic flow

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Abstract

This paper presents a detailed diagnostic analysis of hydrographic and current meter data from three, rapidly repeated, fine-scale surveys of the Almeria-Oran front. Instability of the frontal boundary, between surface waters of Atlantic and Mediterranean origin, is shown to provide a mechanism for significant heat transfer from the surface layers to the deep ocean in winter. The data were collected during the second observational phase of the EU funded OMEGA project on RRS *Discovery* cruise 224 during December 1996. High resolution hydrographic measurements using the towed undulating CTD vehicle, SeaSoar,. traced the subduction of Mediterranean Surface Water across the Almeria-Oran front. This subduction is shown to result from a significant baroclinic component to the instability of the frontal jet. The Q-vector formulation of the omega equation is combined with a scale analysis to quantitatively diagnose vertical transport resulting from mesoscale ageostrophic circulation. The analyses are presented and discussed in the presence of satellite and airborne remotely sensed data; which provide the basis for a thorough and novel approach to the determination of observational error.

Keywords: OCEANIC FRONTS, MESOSCALE FEATURES, BAROCLINIC INSTABILITY, VERTICAL MOTION, DOWNWELLING, MEDITERRANEAN SEA, WESTERN MEDITERRANEAN, ALBORAN SEA, ALMERIA-ORAN FRONT, 2.5° W - 0.5° E, 35.0° N - 37.5° N.

1) Introduction

During the observational phase of OMEGA[1] a field experiment was carried out at the eastern end of the Alboran Sea to examine the impact of mesoscale motion on biological distributions. A total of 7 surveys were made of the Almeria-Oran front region on RRS *Discovery* cruise 224 (Allen et al., 1997a; Pugh et al.,1997) in December 1996 and January 1997 using the towed undulating CTD instrument, SeaSoar (Pollard et al., 1986; Allen et al., 1997b). Here we define the dimensions of oceanic mesoscale flow as, O(10-100 km) spatially, O(10 days) temporal period and Rossby number $\varepsilon < 1$.

The Alboran Sea fills a small and topographically complex region at the western end of the Mediterranean. Atlantic water flows into the Alboran Sea at the surface through the Strait of Gibraltar and generally forces two anticyclonic gyres, the Eastern and Western Alboran Gyres (**Figure 1**) (Arnone et al., 1990; Folkard et al., 1994; Tintoré et al., 1988). At the eastern side of the Eastern Alboran Gyre, an intensified front is formed between surface waters of recent Atlantic origin and those of the western Mediterranean Sea (Tintoré et al., 1988). Water leaving the Almeria-Oran frontal jet is either re-entrained into the Alboran Sea gyres or feeds the Algerian current, hugging the steep topography of the coast of Algeria (Arnone et al., 1990).

Both Alboran Sea gyres exhibit large variations in surface structure and may collapse entirely for periods of weeks to months (Heburn and LaViolette, 1990; Viudez et al., 1998). The sharp density gradient at the periphery of the gyres forms a front that is susceptible to instability at the mesoscale (Tintoré et al., 1991; Arnone et al., 1990). Atlantic waters eventually spread eastwards to the Eastern Mediterranean (Brankart and Brasseur, 1998; Font et al., 1998) and northwards into the central Western Mediterranean (Taupier-Letage and Millot, 1988; Millot et al., 1990). As they do so, they are modified by entrainment and mixing; in this paper we present an analysis of observations that illuminates the key dynamical processes at least in the region of the Almeria-Oran front. The repeated surveys presented here enable us to understand the evolution of the front and its relation to subduction. In the next section we describe the physical observations and compare water mass characteristics with previous studies. In sections 3 and 4 we present a dynamical interpretation of the data sets and

[1] Partly funded by the European Commission under MAST contract No. MAS3-CT95-00001

an omega equation (Hoskins et al. 1978) diagnostic analysis of mesoscale vertical flow and cross front transport. In section 5 we present a discussion of our findings. In an accompanying paper (Fielding et al., 2001 – this edition) biological data are presented and effect of physical motion on the upper ocean ecology of the Almeria-Oran front region is discussed.

2) Water mass characteristics and general description of the hydrography during RRS *Discovery* cr. 224 (OMEGA).

RRS *Discovery* arrived in the Alboran Sea on the 2nd of December 1996. Prior to the cruise, Satellite IR images had been provided and processed by the Southampton Oceanography Centre (SOC) and the University of Pisa (UP) since the 1st October that year. The two Alboran Gyres had been clearly visible throughout October (Baldacci et al. 1998 and **http//radar.iet.unipi.it/OMEGA/ATLAS/atlas.html).** During November, the Western Alboran Gyre appeared to move eastwards to be replaced by a new Western Alboran Gyre and resulting in a period from the 19th November to the 24th of November during which three gyres appeared to exist (Viúdez et al., 1998). By the 28th November the 'central' and Eastern Alboran gyres had coalesced and the in-situ observations described here occurred during the existance of a more 'traditional' two Alboran gyre flow regime (**Figure 2**). A detailed discussion of the apparent three gyre flow is beyond the scope of this paper, however we mention it here to solicit further communication with fellow researchers who may have observed similar events during other periods of observation

RRS *Discovery* cr. 224 made two large scale surveys and three repeat fine scale surveys of the Almeria-Oran front during December 1996 (Allen et al. 1997a). .In addition a brief survey was also made of the head of the Algerian Current. Two further fine scale surveys of the Almeria-Oran front were made during January 1997 (Pugh et al. 1997) (**Figure 3**). The timetable and duration of these surveys was as follows:

16:15 GMT 2/12/96 – 17:15 GMT 5/12/96 Large Scale Survey 1 (LSS1)
17:40 GMT 6/12/96 – 04:30 GMT 10/12/96 Large Scale Survey 2 (LSS2)
18:30 GMT 11/12/96 – 10:00 GMT 15/12/96 Fine Scale Survey 1 (FSS1)

21:10 GMT 16/12/96 – 17:30 GMT 20/12/96 Fine Scale Survey 2 (FSS2)
19:50 GMT 21/12/96 – 22:15 GMT 24/12/96 Fine Scale Survey 3 (FSS3)
18:10 GMT 26/12/96 – 21:30 GMT 28/12/96 Algerian Current Survey (ACS)
Port call, Cartagena, Spain
18:00 GMT 30/12/96 – 14:00 GMT 2/1/97 Fine Scale Survey 4 (FSS4)
08:00 GMT 14/1/97 – 20:00 GMT 16/1/97 Fine Scale Survey 5 (FSS5).

Routine calibration and processing of SeaSoar CTD hydrographic data and RDI 150 kHz VM-ADCP current data are described in the relevant data reports, Allen et al. 1997b and 1997c, and at **http://www.soc.soton.ac.uk/GDD/omega/Disco/index.html**. It is worth noting here, however, that for the duration of the cruise an Ashtech 3D-GPS system was used to correct the ship's gyro heading errors and RACAL SkyFix differential GPS navigation had been purchased for the first leg of the cruise, i.e. until 29/12/96. The calibration and processing techniques are considered sufficient for accuracy in salinity to 0.01 and in absolute current velocity to order 1 cms^{-1}.

In **Figure 4**, we present a potential temperature versus salinity (θ/S) diagram for the SeaSoar CTD data collected during the second large scale survey. The lowest salinity waters, ~36.63, are surface waters of recent Atlantic origin and are referred to as Modified Atlantic Water (MAW) (Arnone et al., 1990; Sparnocchia et al., 1994). MAW, flowing as a strong jet along the Almeria-Oran front, splits to re-circulate into the Eastern Alboran Gyre and form the head of the Algerian Current (**Figure 5**). Mediterranean Surface Waters (MSW) (Arnone et al. 1990), salinity >37.5 temperature >15.5 °C, were found in the N. E. corner of the LSS2 area. This water appeared to be flowing slowly south westward along the Spanish coast but had not reached the area that was to be covered by the later fine scale surveys. The characteristic θ/S of MSW seems less well defined (**Figure 4**) than that of the MAW perhaps as a result of the the wide area and variability of its formation; predominantly old MAW that has remained at the surface in the Western Mediterranean (Benzohra and Millot, 1995; Gascard, 1978). Between these two surface water masses, a large area of lower temperature intermediate salinity water (< 15.5 °C and 36.7-37.5 psu) was observed at the surface north of the Almeria-Oran front (**Figures 5 and 6**).

Following Gascard and Richez (1985), we will refer to this as Atlantic-Mediterranean Interface Water (A-MIW). A-MIW is formed through mixing between MAW and intermediate Mediterranean waters either vertically or horizontally following the upwelling of the latter along the north coast of the Alboran Sea.

Below the surface waters, a temperature minimum layer (TML) exists to a depth of generally 250-300 m. This layer has a salinity around 38.2 and temperature below 13.5 °C, and is believed to be formed by winter cooling of MSW along the French coast (Gascard and Richez, 1985, Pinot and Ganachaud, 1999). A particular form of TML water was observed in an anticyclonic eddy at around 36.6° N, 0.9° W and will be the subject of a future study (J. T . Allen and D. A. smeed, SOC, personal comm.). This has the characteristics of Winter Intermediate Water (WIW) as described by Pinot and Ganachaud (1999), temperature < 13 °C and salinity 38.2. Below the TML there is a tight θ/S signature of Levantine Intermediate Water (LIW) that forms a distinct salinity maximum (**Figure 4**) (Sparnocchia et al., 1994; Gascard, 1978). SeaSoar data were only available to a depth of ~370 m and therefore it only just resolves the core of the LIW, ~38.5 psu (Gascard and Richez, 1985).

To the south and west of the Almeria-Oran front, the A-MIW descends steeply below the deep surface mixed layer of MAW in the Eastern Alboran Gyre (**Figure 6**). This results in a large horizontal density gradient across the Almeria-Oran front and a current speed around 1 ms^{-1} in the frontal jet (**Figure 5**). In contrast, A-MIW and MSW have similar densities and the front between them has no strong jet associated with it. In the five consecutive fine scale surveys of the Almeria-Oran front (**Figure 2**), the presence of a θ/S signature of MSW is variable (**Figure 4**). This variability drew our attention to the presence of significant ageostrophic circulation and therefore we will spend some time discussing it here. The θ/S envelope for FSS1 shows no signature of MSW. By FSS2, there is a clear high salinity (~37.75) high temperature (~16.2 °C) signature of MSW. This becomes more pronounced, salinity >37.8 and temperature ~16.4 °C, during FSS3. During FSS4 and 5, the θ/S signature of MSW appears to mix away, becoming colder and more saline as it does so.

As discussed by Tintoré et al. (1988), there is a covergence at the northern end of the Almeria-Oran front between the rapid inflow of MAW and a net south-westward flow (10-20 cms^{-1}) along the Spanish coast (**Figure 5**). The convergence in the surface waters maintains a sharp density gradient at the front and provides a source of potential energy for a baroclinic component to the instability that we will demonstrate in this paper. During our observations, the net south-westward flow along the Spanish coast transported MSW (salinity > 37.5) ~100 km, from east of 0.5° W to the region covered by the fine scale surveys (**Figure 5**), between LSS2 and FSS2. This equates to a mean advection of ~15 cms^{-1}.

We can follow the MSW signature better by plotting salinity on the 27.9 σ_0 surface (**Figure 7**). Between FSS2 and FSS3, the high salinity signature of MSW, that had reached the convergence at the north end of the Almeria-Oran front, was advected along the front and, most importantly, down and across the front under the influence of ageostrophic flow. During FSS3, MSW was clearly observed at the southern end of legs *h*,*i* and *j* (**Figure 2 and Figure 7**) at depths of up to 150 m. This implies a mean vertical velocity of at least 25 mday^{-1}.

3) Instability of the Almeria-Oran front

During the quickly repeated fine scale surveys, FSS1,2 and 3, the position and shape of the Almeria-Oran front can be seen to change significantly in the SeaSoar CTD data sets (**Figures 5 and 7**). The slope of the density surfaces across the front also changed with time (Snaith et al., 1997), and position along the front (**Figure 8**) indicating the existence of cross front and vertical circulation (Pollard and Regier, 1992). Along leg e, the front moved south and steepened between FSS1 and FSS2. The front then moved back northwards so that during FSS3 the front was less steep and further north than observed in FSS1. Further down stream, along leg j, the front moved north and became less steep between FSS1 and FSS2. The front then moved back southwards and steepened between FSS2 and FSS3; ending up in virtually the same position as it had been observed in FSS1. These observations are consistent with baroclinic instability and the propagation of wavelike meanders along the front (Hoskins and Bretherton, 1972; Munk et al., 2000).

Killworth et al. (1984), considered a two layer model of a front similar to that found in our observations. Their model was set in a semi-infinite domain with an upper layer that vanished at some average position in the y axis direction forming a surface front (**Figure 9**). This configuration was shown to be unstable to small perturbations whatever the potential vorticity distribution. A detailed analysis of the model was presented both in the original paper and by Allen et al. (1994), therefore we chose not to repeat this discussion here. Following Killworth et al. (1984), the wavelength of the fastest growing mode of instability is given as

$$\lambda_f = 5.5 R_o (r-1)^{1/4}, \tag{1}$$

with a phase speed of

$$c_r = \frac{0.113}{(r-1)^{1/2}}, \tag{2}$$

where the Rossby radius is given by

$$R_o = (g'H)^{1/2} f^{-1}, \tag{3}$$

g' is the reduced gravity $g \Delta\rho / \rho_0$, f is the local coriolis parameter, H is the asymptotic thickness of the upper layer, rH is the total water depth (**Figure 9**) and the phase speed of instability has a real component, c_r. For the Almeria-Oran front we can take the values $f = 8.6 \times 10^{-5}\ rad\,s^{-1}$, $g' \approx 0.02\ ms^{-2}$, $H \approx 150\ m$ (**Figure 8**) and $R_o \approx 20\ km$. Previous authors have commented on the quiescent nature of the deeper layers of LIW and Mediterranean Deep Water (MDW) (Tintoré et al. 1991, Viúdez et al. 1996b) and have considered sensible levels of no motion at around 200 m, later on in this paper we will reinforce this view with our observations. This would lead us to consider perhaps a value of $r = 1.3$ and thus $\lambda_f \approx 80\ km$ and $c_r \approx 20\ cms^{-1}$ for the dominant mode of instability.

Perhaps more significantly, Killworth et al. (1984) derived an equation for the growth of the zonally averaged total perturbation energy. This separated the relative importance of horizontal and vertical shear processes in the growth of instability. Allen et al. (1994) simplified their ratio of barotropic to baroclinic contribution as

$$\frac{g'H}{f^2 L_o^{\,2}} = \frac{R_o^{\,2}}{L_o^{\,2}} \tag{4}$$

where L_o is the width of the front and geostrophic balance is assumed. For our observations of the Almeria-Oran front, $L_o \approx 45\ km$ (**Figures 5 and 8**) and therefore only 20% of the total perturbation energy would be estimated to result from barotropic, horizontal shear, processes.

Returning to our observations, the ADCP current vectors show a coherent circulation over the scales of the surveys (**Figures 5 and 7**) suggesting only small contamination by tidal flows and inertial motions even near the surface. The wind field, recorded by the meteorological team on board (Allen et al., 1997a), was generally light to moderate (rarely above 15 ms^{-1}) and of short fetch (less than 24 hours) until the end of FSS3 (**Figure 10**). Plotting geostrophic shear profiles for leg e of FSS1 (**Figure 11**), calculated from the density gradients between SeaSoar CTD profiles, indicates significant vertical shear in the Almeria-Oran frontal jet associated with the sharp change in stratification across the pycnocline between the surface and intermediate water masses. VM-ADCP velocity profiles (**Figure 11**) are consistent with the geostrophic shear profiles but indicate the existence of an ageostrophic flow concordant with the position and spatial scale of the front. The vertical shear in the jet is such as to eliminate any coherent residual signature of the surface currents at depths greater than 150 m (Allen et al., 1997c). A streamfunction fitted to the VM-ADCP velocities at 198 m, following Allen (1995) and Pollard and Regier (1992), produces a dynamic height anomaly field an order of magnitude lower than that at the surface.

In the following analyses of the SeaSoar hydrographic data we have used the VM-ADCP data to provide a reference dynamic height anomaly field at 198 m depth. However, as suggested above, this does not give results that are significantly different from those obtained by assuming a level of no motion at 198 m (not shown). For this data set we have chosen not to employ the method of Rudnick (1996) or Naveira Garabato et al. (2000), where VM-ADCP data at all depth levels are used in the form of streamfunctions, and our justification is as follows. We have discussed that above 150 m our data indicate the existence of a large horizontal ageostrophic flow coherent with the scale of the front, perhaps 15-25 cms^{-1} in magnitude (**Figure 11**). Indeed Viudez et al. (2000) and Gomis et al. (2001) suggest that much of this ageostrophic flow may be horizontally non-divergent and not balanced by the vertical circulation that we are trying to diagnose. The diagnosis of mesoscale vertical motion using the omega equation, either under QG balance as we will present in section 4 or semi-geostrophic balance as introduced by Hoskins (1975), requires the geostrophic velocity field to be well known and not significantly contaminated by ageostrophic flow whether or not the latter is in balance with the vertical circulation.

Processed SeaSoar and VM-ADCP data were mapped to a regular grid using a computationally inexpensive anisotropic Gaussian filter as described in Allen et al. (1995) and (2000). At each depth level, the filtered values Φ_k at the grid coordinates (x_k, y_k) were given by

$$\Phi_k = \frac{\sum_{i=1}^{n} \phi_i e^{-\left(d_{ik}^2\right)}}{\sum_{i=1}^{n} e^{-\left(d_{ik}^2\right)}} \tag{5}$$

where ϕ_i are the n observations at each depth level,

$$d_{ik}^2 = \frac{1}{L^2}\left[\frac{(x_i - x_k)^2}{a^2} + \frac{(y_i - y_k)^2}{b^2}\right], \tag{6}$$

$a \times L$ and $b \times L$ are the horizontal length scales and the summation is over observations which fall within a search ellipse defined by the values of *a* and *b*. The grid had 71 points in each of the x and y directions and depth levels every 8 m from

5 m to 405 m. The axes of the grid were rotated such that the positive x direction was 66° clockwise from north on a mercator projection of reference 36.25° latitude. The north (south) corner of the grid had geographical co-ordinates 1.0941° W, 37.0900° N (1.8601° W, 35.4825° N). The anisotropic filter axes were rotated relative to the grid to lie approximately along and across the FSS track legs, i.e. with the filter positive x axis direction 5° clockwise from north. The along track and across track filter length scales were selected to be 8 km and 25 km respectively, to reflect the Nyquist minimum resolvable wavelength for observations with along track and across track sample spacing of ~4 km and ~12 km respectively.

Geostrophic relative vorticity,

$$\varsigma_g = \left(\frac{\partial U}{\partial y} - \frac{\partial V}{\partial x}\right), \tag{7}$$

was calculated from the geostrophic velocities,

$$(U, V) = \frac{1}{\rho f}\left(-\frac{\partial p'}{\partial y}, \frac{\partial p'}{\partial x}\right), \tag{8}$$

where p' is the dynamic height, from the SeaSoar CTD data, referenced to a streamfunction fitted to the VM-ADCP velocities at 198 m (discussed above). In **Figure 12** maps of geostrophic relative vorticity at a depth of 53 m are plotted for FSS1-3. These maps suggest that the regions of high cyclonic vorticity propagate along the front. The direction of subduction of MSW discussed in section 2 and the frontal model of Killworth et al. (1984), discussed at the beginning of this section, suggest that we can assume the direction of propagation of these vorticity anomalies is the same as that of the mean flow. Therefore, by inspection of **Figure 12**, we estimate that the wavelength and phase speed of these features are ~ 90 km and order 10 cms^{-1} respectively.

Satellite images (**Figure 13**) indicate mesoscale instability at a wavelength of 40-50 km and a phase speed of order 20 cms^{-1}. A detailed sea surface temperature survey, carried out by the UK Meteorological Office Research Flight on the 14th December

1996 using an IR radiometer on board their Lockheed Hercules C130 aircraft (**Figure 14**) (Allen et al., 1997a), also supports a characteristic instability wavelength considerably less than that reported above for the in-situ FSS observations, in this case perhaps 50-60 km. The apparent inconsistency here between in-situ and remote observations is in fact easily explained by considering the synopticity (Allen et al., 2000) of the FSS in relation to the evolution of instability at the Almeria-Oran Front. In a box style survey pattern of equally spaced parallel legs like those discussed by Allen et al. 2000, then there is an effective velocity of the research vessel along the front, v_f, given by,

$$v_f = v_s \frac{S}{(S + T_l)}, \tag{9}$$

where v_s is the speed of the ship, T_l is the length of each cross front leg of the ship's track and S is the track leg separation. For FSS1-3, T_l is not a constant and therefore we have to define a mean velocity along the front for the research vessel,

$$\bar{v}_f = v_s \frac{S}{(S + \bar{T}_l)}, \tag{10}$$

where the over-bar denotes a mean value over the duration of each survey. Taking $v_s = 4.5\ \mathrm{ms}^{-1}$ (~ 9 knots), $S = 12$ km and $\bar{T}_l \approx 90$ km then $\bar{v}_f \approx 0.5\ \mathrm{ms}^{-1}$. We can now define a non-dimensional synopticity parameter for our surveys,

$$\Gamma = \frac{v_f}{c_r} \approx 2, \tag{11}$$

where we recall that c_r is the real component of the characteristic phase speed for the instability. Following the Doppler shift like analogy of Allen et al. (2000), then the apparent wavelength, λ_a, from the in-situ FSS1-3 may be corrected by

$$\lambda = \lambda_a \left(1 - \Gamma^{-1}\right) = 45\, km\,. \tag{12}$$

And therefore the in-situ and remote observations are quite consistent.

These results highlight the difficulty of obtaining synoptic observations of mesoscale features. Furthermore, we have only considered the real component of the phase speed of instability. The instabilities may have a finite growth rate that we are unable to determine from these observations. However, **Figure 8** indicates that the extent of horizontal movement of the front is similar everywhere along the front. Indeed we know that the growth of instabilities may be limited by larger scale deformation fields (Spall, 1997); in this case perhaps the confluence of Atlantic and Mediterranean surface waters at the northern end of the front.

4) Vertical velocities

Assuming tidal velocities and inertial motions are small, then for any suitable depth level we can consider a Rossby number, ε, defined as

$$\varepsilon = \left| \frac{\zeta}{\zeta_g} - 1 \right| \tag{13}$$

where ζ_g is given in (7) and ζ takes a similar form replacing the geostrophic velocities with those measured by the VM-ADCP. Perceiving that ε is too noisy to map directly, in **Figure 12** we present maps of ζ for FSS1-3 at a depth of 54 m. Comparing with those of ζ_g also in **Figure 12** and discussed earlier, we estimate that $\varepsilon \leq 0.3$ provides a sensible bound for the Rossby number of the flow. It is therefore reasonable to suggest that quasi-geostrophic (QG) balance is sufficient to quantitatively examine the ageostrophic flow at least to leading order.

Following previous observational studies (Leach 1987, Tintoré et al. 1991, Pollard and Regier 1992, Fiekas et al. 1994, Viúdez et al. 1996a, Allen and Smeed 1996, Rudnick 1996 and Shearman et al. 1999) and modelling studies (Strass 1994, Pinot et al. 1996 and Allen et al. 2000) we have used the QG form of the omega equation (Hoskins et al., 1978) to diagnose the vertical velocity field for each FSS. On an β-

plane, and for timescales at which the effects of diffusion can be considered negligible, this equation can be written as

$$f^2 \frac{\partial^2 w}{\partial z^2} + N^2\left(\frac{\partial^2}{\partial x^2} + \frac{\partial^2}{\partial y^2}\right) w = \nabla_h . \mathbf{Q} + \frac{\beta}{\rho_0} \frac{\partial^2 P}{\partial x \partial z}, \tag{14}$$

where $\mathbf{Q} = \left[2f\left(\frac{\partial V}{\partial x}\frac{\partial U}{\partial z} + \frac{\partial V}{\partial y}\frac{\partial V}{\partial z}\right), -2f\left(\frac{\partial U}{\partial x}\frac{\partial U}{\partial z} + \frac{\partial U}{\partial y}\frac{\partial V}{\partial z}\right)\right]$, and may be easily derived from the QG momentum equations as given in a number of the references above and therefore not repeated here. The symbol usage is conventional, f is the coriolis parameter, $(U, V, 0)$ are the geostrophic components of the velocity field, w is the vertical component of the total velocity field, N is the Brunt Väisälä frequency and (x, y, z) are the usual right handed axis set with z positive upwards. The β term, where $\beta = \frac{f \cot \phi_0}{R}$ for a mean latitude ϕ_0 and earth's radius R, is generally not significant in forcing the solution for w. By assuming some boundary conditions for w at every point around the data set equation (14) may be solved for a given geostrophic velocity field provided the variability in all three dimensions is properly resolved by the spacing of the observations.

After breaking (14) down into a set of simultaneous finite difference equations, solutions for w were obtained using a NAG library routine based around Stone's Strongly Implicit Procedure. We sought solutions for the boundary condition $w = 0$ everywhere around our observational domain. Allen and Smeed (1996) suggested that solutions for w showed little sensitivity to lateral boundary conditions. However, more recent studies (Gomis et al., 2001) have shown that there may be much greater sensitivity to lateral boundaries in strongly advective regions. Therefore, we have chosen to ignore values of w where solutions to (14) for lateral boundary conditions $\partial w / \partial \mathbf{n} = 0$,where $\mathbf{n}$ is a unit vector perpendicular to the boundary, vary from w by more than 30% of RMS w for any particular depth level. This is approximately equivalent to a 10% tolerance on the peak values of w. Allen and Smeed (1996) discussed how solutions to (14) can be sensitive to the bottom boundary condition: however, we have shown that at the Almeria-Oran front there is a sharp contrast between the frontal dynamics in the upper 150 m of the water column and the quiescence of the deeper temperature minimum layer (TML) and LIW. Thus we expect our boundary condition of $w = 0$ at the bottom of our dataset, 397 m, to be consistent with the observations.

In **Figure 15**, we have mapped w at a depth of 77 m. This represents a sensible compromise for all three surveys, FSS1-3, where the solutions to (14) derive maximum vertical velocities at water depths between 50-100 m. This is consistent with the change in slope of the frontal interface shown in **Figure 8**. The horizontal pattern of vertical motion does not change significantly with depth. As expected (Gill, 1982) we see downward vertical velocities on the upstream side of positive (cyclonic) vorticity maxima (**Figure 12**). The magnitude of the derived vertical motion is only ~10 mday^{-1} $\left(\approx 1.0 \times 10^{-4} ms^{-1}\right)$ and therefore less than half of that expected in section 2 by the cross front transport of MSW. However, we have shown in section 3 that the error in assuming synopticity for each FSS1-3 results in an overestimation, λ_a, of the wavelength of instability, λ, such that from (11) and (12)

$$\frac{\lambda_a}{\lambda} \approx 2 . \tag{15}$$

Allen et al. (2000) showed that to first order in a quasi-synoptic survey, the apparent magnitude of w, $|w_a|$, varied as the inverse of the apparent wavelength; i.e.

$$\frac{|w_a|}{|w|} \approx \frac{\lambda}{\lambda_a} = \left(1 - \Gamma^{-1}\right) \approx 0.5 . \tag{16}$$

Thus both our diagnoses and our direct observations are consistent with mesoscale vertical motion of around 20-25 mday^{-1}. Vertical velocities of this magnitude are also consistent with primitive equation modelling studies of upwelling and downwelling associated with baroclinic instability (Samelson, 1993; Pinot et al., 1996 and Nurser and Zhang, 2000).

Although our surveys FSS1-3 provide quasi-synoptic snapshots of the propagation of mesoscale instabilities along the Almeria-Oran front. They are asynoptic with respect to the advective flow along the front, $\bar{v}_f < U$ (**Figures 7 and 11**). Therefore we cannot directly derive the vertical transports associated with the diagnosed vertical velocities by correlating the w fields with temperature or any other observed water parameter. However, as a scale analysis, if we take a scale for w of 20 mday^{-1} $\left(\approx 2.0\times10^{-4}ms^{-1}\right)$, $\lambda/2 \approx 20\,km$ and assume circular geometry with radius $\lambda/4$ then we can achieve the periodic subduction of MSW at a rate of ~$6\times10^4\ m^3s^{-1}$ vertically. Taking a mean temperature difference of 1.5 °C between MSW and A-MIW, a density of $1027.9\,kgm^{-3}$ and a specific heat capacity of $3900\,Jkg^{-1}K^{-1}$, then this implies a periodic vertical heat transport of ~$1.15\times10^3\ Wm^{-2}$. In **Figure 16**, we present contoured temperature and salinity sections for leg h of FSS3 which clearly shows the subducted warm MSW temperature anomalies in the thermocline between the MAW and TML/LIW.

Josey et al. (1998, 1999) indicate that in winter, the surface waters of the W. Mediterranean loose heat to the atmosphere at a rate of ~$1.0\times10^2\ Wm^{-2}$. Of course this climatology represents a basin scale average, but our scaling arguments suggest that upper ocean mesoscale vertical motion, associated with unstable fronts and eddies may provide a local mechanism for significant heat loss from the surface layers to the deep ocean. Baldacci et al (2001) determined regions of upwelling in the Alboran Sea, by analysis of AVHRR sea surface temperature data, in areas where Gomis et al. (2001) have derived vertical velocities similar in magnitude to those presented here. Through an analysis of CTD surveys of the western Alboran Sea and a coastal upwelling model, Sarhan et al. (2000) computed $\sim 3\times10^7\ m^2$ annual vertical flux per unit length parallel to the coast for each of two mechanisms, the departure of the Atlantic inflow jet from the Spanish coast and the wind driven offshore transport. Both mechanisms are shown to exist for approximately 50% of the year and therefore over our length scale of 20 km the more instantaneous vertical transport of $\sim 4\times10^4\ m^3s^{-1}$ is remarkably consistent with our subduction rate for MSW.

5) Summary and conclusions

During the second component of the observational phase of the EU OMEGA project on RRS Discovery cruise 224, five repeated fine-scale surveys were made in the Alboran Sea in the region of the Almeria-Oran front. The multidisciplinary surveys included bioacoustic data from both the shipboard 150 kHz ADCP and a SIMRAD EK500 echosounder (at 38, 120 and 200 kHz), and hydrographic data from SeaSoar. In addition, high resolution biological net samples were taken across the front with a Longhurst-Hardy Plankton Recorder (LHPR). In this paper we have presented a detailed analysis of the hydrographic data for the first three fine scale surveys (FSS) which were rapidly repeated during December 1996. A complementary paper (Fielding et al., 2001) compares these analyses with the concurrent observations of biological distributions to look at the interdisciplinary influence of mesoscale physical flows.

The variability in the position and shape of the Almeria-Oran front and the strongly sheared velocity field were indicators of mesoscale frontal instability and the presence of significant ageostrophic flow. The analysis of temperature and salinity on density surfaces has shown Mediterranean Surface Water (MSW) advecting westwards along the Spanish coast until it reaches the Almeria-Oran front. At which point, some of this water is entrained into the frontal jet and is drawn down along the front at a subduction rate estimated at 25 mday^{-1}. Horizontal current shear across the front is skewed to the north-east, the cyclonic side of the front. In the pycnocline, layer thickness changes in response to the generation of cyclonic relative vorticity. From the rapid repetition of the surveys we infer the advection of vorticity and vertical motion. Solving the quasi-geostrophic omega equation, we have calculated vertical velocities. These derived vertical velocities, ~10 mday^{-1} are smaller than expected from the observed subduction of MSW. However, recent studies of the effects of asynopticity on dynamical data analyses (Allen et al. 2000) enable us to explain the differences in the apparent magnitude of w.

Regarding the synopticity of our observations, we have addressed this through an error analysis based on information from concurrent airborne and satellite observations that give a more synoptic view of sea surface temperature. These observations indicate a real component of the phase speed of instabilities at around 20 cms^{-1}. Each leg of our fine scale surveys were repeated with a five day interval. And therefore for a wavelength of 50-100 km the repeat period of our surveys is a little outside the Nyquist limit for properly resolving the propagation of instabilities. Thus we cannot interpolate between surveys to obtain a more synoptic survey as described in Gomis et al. (2001). We intend to use the method of Rixen et al. 2001 to relocate our observations relative to some derived mean flow in a future analysis. However, this technique is novel and complex, it is properly the subject of a future manuscript and beyond the scope of the work presented here.

Following a scaling argument we have estimated the heat transport associated with the mesoscale subduction. Traditionally, baroclinic instability is responsible for upwelling warm, light water and subducting cold, dense water as available potential energy is converted to kinetic energy and released across the front. During the winter months the surface waters of the western W. Mediterranean are made up of MSW that has been heated and evaporated during the preceeding summer and MAW inflowing from the Strait of Gibraltar and passing around the gyre system of the Alboran Sea. The MSW is denser and subducted by the baroclinic instability at the Almeria-Oran front, but it is also warmer; the high density arising from its high salinity. Therefore these instabilities may result in a net heat loss from the surface layers to the deep ocean. Further more this heat loss could locally exceed that lost to the atmosphere during the winter.

Acknowledgements.

We thank the Master, crew and scientists on board RRS *Discovery* cruise 224 for both a comprehensive data set and a splendid Christmas. We also thank the EU commission for their support under Framework 4, contract no MAS3-CT95-0001 (OMEGA), and the DERA under the NERC/MOD Joint Grant Scheme project TOES. Many thanks also to the editor and the reviewers for their comments and suggestions.

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Figures:

Figure 1: Cartoon of the two gyre surface circulation of the Alboran Sea, following Arnone et al. (1990) and others.

Figure 2: NOAA-14 AVHRR images provided by the Natural Environment Research Council through the Southampton Oceanography Centre and processed at the University of Pisa (Baldacci et al., 1998).

Figure 3. Cruise tracks for Large Scale Surveys 1 and 2 (LSS1 and LSS2), Fine Scale Surveys 1-5 (FSS1-5) and a survey of the head of the Algerian Current (ACS).

Figure 4.

(a) Potential temperature as a function of salinity for all the SeaSoar data collected during the second large scale survey. Lines of constant density are also shown at intervals of 0.1 kg/m^3. Note the relatively fresh surface modified Atlantic waters (MAW) in the eastern Alboran gyre, the warm salty Mediterranean surface waters (MSW), the Levantine water (LIW) and the temperature minimum layer (TML).

(b) Envelopes of potential temperature as a function of salinity for SeaSoar LSS2 and FSS1-5 of the Almeria-Oran front.

Figure 5. Salinity (coloured dots) and VM-ADCP derived current velocity vectors at 14 m depth for LSS2 and FSS1-3.

Figure 6. Composite indicating the three dimensional structure of the upper 350 m of the water column during LSS2. The dotted cruise tracks are coloured to show the horizontal salinity distribution at depths of 13 m and 157 m. The vertical contoured section shows temperature along leg PD of the cruise track (Figure 2), the vertical axis has been removed for clarity but the dotted tracks for leg PD on the two horizontal slices are shown in the plane of the vertical section. Only the horizontal planes are annotated. The solid black cartoon lines on the horizontal planes indicate the boundaries between water types observed at the time of LSS2.

Figure 7. Salinity (coloured dots) on the density surface σ_0=27.9 and VM-ADCP derived current velocity vectors at 54 m depth for FSS2 (a) and FSS3 (b), pressure also shown (c) for FSS3.

Figure 8: (a) contoured density sections for leg e of FSS1-3 (1-3, top-bottom); (b) density interval sigma-t = 27.4-27.6 for leg e of FSS1-3 overplotted on the same diagram and; (c) density interval sigma-t = 27.4-27.6 for leg j of FSS1-3, overplotted on the same diagram.

Figure 9: The frontal model configuration used by Killworth et al. (1984).

Figure 10: Wind speed and direction, averaged over 1 hour intervals, for the period Julian day 345-364 during RRS Discovery cruise 224.

Figure 11: Geostrophic velocity profiles relative to no motion at 198 m (solid lines) and ADCP velocity relative to 194-202 m bin (dotted lines) for leg e of FSS1. Each pair of profiles is offset by 50 cms^{-1}.

Figure 12: Maps of geostrophic relative vorticity (a) and VM-ADCP relative vorticity (b) at a depth of ~50 m are plotted for FSS1-3 (top-bottom).

Figure 13: AVHRR SST images for 02:31 on the 26th December 1996 (top) and 02:21 on the 27th December 1996 (bottom). Dotted red curves have been added to indicate the growth and propagation of an instability on the Almeria-Oran front. (Images courtesy of the Remote Sensing Data Analysis Service, Plymouth, http://www.npm.ac.uk/rsdas/)

Figure 14. Sea surface temperature as observed by an IR radiometer on board the UK Meteorological Office Research Flight's Lockheed Hercules C130 aircraft during the 14th December 1996.

Figure 15: Maps of vertical velocity at a depth of 77 m are plotted for FSS1-3 (top-bottom).

Figure 16: Contoured temperature and salinity sections for leg h of FSS3; which clearly show subducted MSW temperature and salinity anomalies south of the front in the thermocline between MAW and TML/LIW.