Chapter 3: Ocean Currents and Eddies

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ABSTRACT : This chapter reviews the contribution of satellite altimetry to the observation and understanding of ocean currents and eddies. It focuses on mesoscale variability which is the dominant signal of the ocean circulation and is a crucial component of its dynamics. After a general introduction on mesoscale variability in section 1, section 2 discusses specific issues for altimeter data processing which are necessary for mesoscale studies including merging, sampling and velocity estimation. Section 3 deals with the description of ocean currents. We review first some of the main techniques used to estimate the absolute velocities and transports from altimeter data. We then look at each major current system separately, and review the considerable progress made by altimetric studies in each of the main western boundary currents, eastern boundary currents, open ocean currents, and in the semi-enclosed seas. Section 4 provides a global statistical description of mesoscale variability which is a unique contribution of satellite altimetry. Studies dealing with the geographical and seasonal to interannual variations in eddy intensity are analyzed. We also summarize the main findings on space and time scales of mesoscale variability and the relationship between spatial scales and the internal Rossby radius. We then proceed to the analysis of frequency and wavenumber spectra and the relationship with quasi-gesotrophic turbulence theories. Studies which compare these eddy statistics with eddy resolving model simulations are also summarized. Eddy dynamics (eddy/mean flow interaction, eddy transport and eddy diffusivity) are finally discussed.

1. INTRODUCTION

The large-scale ocean circulation can be described by large-scale gyres, with slow, diffuse equatorward currents at the eastern boundary, and fast, intense, poleward flows at the western boundary – the so-called western boundary currents. In the zonally unbounded Southern Ocean, the Antarctic Circumpolar Current (ACC) flows continuously around the globe and provides a major link for water property exchanges between the Atlantic, Indian and Pacific Oceans. Time scales of this large scale horizontal circulation are typically of a few years. Superimposed on (but also interacting with) this mainly wind-driven circulation, is the thermohaline vertical circulation which allows the exchange of heat between the equatorial and high latitude regions. Time scales of this vertical circulation (the famous, but oversimplified, conveyor belt representation) are typically of one to several hundred years. This characterization of the ocean circulation which results from theories of large scale wind-driven and thermohaline ocean circulation (see Pedlosky, 1996 for a review), while often illuminating, is, however, only a crude and first order approximation of the real ocean. The ocean is indeed a turbulent system and the ocean circulation is not a large scale and temporally stable phenomenon; it varies over an almost continuous frequency/wavenumber spectrum with space and time scales ranging from tens to thousands of kilometers and from days to years (e.g. Wunsch, 1981). An instantaneous view of the ocean circulation would thus reveal areas of intense and small scale ocean currents almost everywhere and would be dominated by mesoscale variability. In particular, western boundary currents and the ACC are areas of intense mesoscale variability. Open ocean currents which are part of the large scale gyre circulation are also often intense and narrow currents embedded with mesoscale eddies. At the eastern boundaries, superimposed on the broad equatorward flow are energetic currents and coastal upwelling currents, which can be highly variable in space and time.

The mesoscale variability is the dominant signal in the ocean circulation. There is not a precise definition of mesoscale variability but it usually refers to a sub-class of energetic motions with typical space and time scales of 50 to 500 km and 10 to 100 days. Phenomenological representations of this variability include eddies, vortices, fronts, narrow jets, meanders, rings, filaments and waves.

Over the past 30 years, a major effort has been made to observe and model eddy variability. Our understanding has progressed a lot and the eddy field has been visualized qualitatively (e.g. Holland and Lin, 1975; Robinson, 1982). Ocean eddies can be observed almost everywhere at mid and high latitudes, their energy generally exceeding the energy of the mean flow by an order of magnitude or more (e.g. Wyrtki et al., 1976; Richardson, 1983; Schmitz and Luyten, 1991). As in the atmosphere, baroclinic instability processes are thought to be the principal source of eddy energy in the global oceans. There is a distinct concentration of Eddy Kinetic Energy (EKE) along the mean frontal zones in the global oceans, which suggests that variability is primarily generated by instabilities of the mean flow. The baroclinic instability process depends on the presence of vertical shear in the mean flow, implying sloping density gradients, and an available store of potential energy. Under certain conditions, instabilities form which release this potential energy, and convert it to eddy potential and kinetic energies. If the mean flow has significant horizontal shear as in strong narrow jets, barotropic instabilities can also occur. In this case, there is a release of mean kinetic energy to EKE. In the real ocean, the mean flow has both vertical and horizontal shear so potentially both baroclinic and barotropic (mixed) instabilities can co-exist. Through these instability forcing mechanisms, eddy activity is thus maximum in regions of the major oceanic currents. This relationship between ocean currents and mesoscale variability explains, in particular, why the chapter includes a discussion both on ocean currents and eddies.

Mesoscale variability occurs everywhere in the ocean, not just in regions of strong currents. In regions of weak mesoscale activity, mesoscale variability may also be directly forced by fluctuating wind (Frankignoul and Müller, 1979; Müller and Frankignoul, 1981). There are also several examples of eddies which are directly forced by strong small scale wind stress curl (e.g. due to orographic effects). The interaction of currents with bottom topography is also an important (local) mechanism for eddy energy generation. Eddies downstream of small islands are thus often observed (e.g. Aristegui et al., 1994; Heywood et al., 1996). The highest eddy energy is actually also often found adjacent to major topographic obstacles which suggests the importance of topography in the stability of the mean flow. From these source regions, eddy energy can be redistributed throughout the oceanic gyre by radiation of Rossby waves (e.g. Pedlosky, 1977; Talley, 1983).

Mesoscale eddies or more exactly vortices (i.e. characterized by a persistent closed circulation) have their own dynamics which is dominated by non-linear effects (Nof, 1981; McWilliams and Flierl, 1979; McWilliams, 1985, Cushman-Roisin et al., 1990). In contrast with waves, they can transport mass (closed streamlines in the moving fluid) and properties such as heat, salt and chemical tracers including nutrients over long distances. They also have different propagation velocities. McWilliams and Flierl (1979), using a numerical quasi-gesotrophic model showed, for example, that a vortex moves westward at the greatest linear Rossby wave speed. Nof (1981) and Cushman-Roisin et al. (1990) have derived more general formulae for westward eddy motion. Anticyclones also have a tendency for southward propagation and cyclones for northward propagation (e.g. Cushman-Roisin, 1994). The relationship between vortices and waves is intricate and the distinction between eddy or wave signals may be difficult. To what extent is the mid-latitude westward propagation observed in altimeter data (see Chapter 2) due to Rossby waves or to eddies is thus still an open question. Eddies lose energy to smaller scales by dissipation but they also interact, merge and grow in the framework of quasigeostrophic turbulence (e.g. McWilliams, 1989). They can also feed energy back to the mean flow and drive deep circulation (Holland et al., 1982; Lozier, 1997). Their climate role in terms of heat and salt transport is, however, not yet well established. Eddies and more generally mesoscale variability form thus a crucial component of the ocean circulation dynamics, including the large scale and time mean ocean circulation.

Despite all this progress, our understanding of eddy and ocean current dynamics remains incomplete. We still need to quantify the energy sources and sinks of eddies, the eddy energy radiation mechanisms and to better rationalize their horizontal and vertical structures. Their contribution to the total heat transport and their interaction with the general circulation should also be better established : what part of the ocean circulation is eddy-driven ? What is the role of eddies in the ventilation of the thermocline ? Can we parametrize their effect on the general circulation ? Similarly, we need to better understand the ocean current dynamics, their seasonal/interannual variations and their forcing and instability mechanisms. A better understanding of these mesoscale phenomena requires an extensive, global observing system. In this respect, satellite altimetry has unique capabilities for producing a global and synoptic view of the ocean surface, even in remote areas. Spatial sampling along the satellite ground track is well suited for mesoscale studies while a temporal sampling period of 10-20 days should avoid aliasing of most of the mesoscale signal. This description is, of course, limited to the surface oceanic circulation. However, given the high vertical coherence of mesoscale variability, this information reflects more than just surface conditions. For a stratified rotating fluid, simple scaling arguments show that the ratio of horizontal scale/vertical scale should be of the order of N/f where N is the Brunt Vaïsalä frequency and f the Coriolis parameter (e.g. Wunsch, 1982). The influence of a mid-latitude eddy of 100 km radius should be seen to a depth of several hundred meters. Altimeter data, when assimilated in ocean models, can thus be a strong constraint for inferring the three-dimensional ocean mesoscale circulation (see Chapter 5). Complementary information on the vertical structure of the mesoscale variability (e.g. barotropic and baroclinic components) needs to be obtained, however, from in-situ data.

The mesoscale variability in high variability regions was one of the first ocean signals to be observed with Skylab, GEOS-3 and SEASAT altimetric missions (Fu, 1983a). SEASAT, in particular, despite its short duration provided first global description of mesoscale variability (Fu, 1983a). The US Navy's GEOSAT altimeter opened new horizons for the quantitative use of satellite altimetry. GEOSAT was particularly suitable for mesoscale observations due to its long duration — the satellite operated on a near-repeat orbit (17.05-day cycle) for almost three years (November 1986 through June 1989) — and its lower noise level (Douglas and Cheney, 1990; Le Traon, 1992). TOPEX/POSEIDON (hereafter T/P) and ERS-1/2 now provide an improved description of ocean currents and mesoscale variability with better accuracy and an improved space/time sampling (e.g. Fu and Cheney, 1995). Up to now, however, this enhanced capability has not been fully exploited mainly because most T/P data analysis has focused on the large scale signals (see Chapters 2 and 4). We now have more than 10 years of good mesoscale variability measurements from GEOSAT (1986-1989), ERS-1 and ERS-2 (1991-1999+) and T/P (1992-1999+), so we can learn more about the seasonal/interannual variations in mesoscale variability intensity.

Most of the discussion in this chapter centers on results obtained from GEOSAT, T/P and ERS-1/2 data, and shows how satellite altimetry contributes to the description of ocean currents and mesoscale variability. The chapter is organized as follows. Section 2 discusses the specific issues for altimeter data processing for mesoscale studies. Section 3 contains the description of ocean currents. Section 4 is focused on a global description of mesoscale eddies and of their dynamics. Section 5 provides the main conclusions and prospects for the future.

2. ALTIMETER DATA PROCESSING FOR MESOSCALE STUDIES

We proceed here with a brief description of altimeter data processing issues, focusing on the aspects which are relevant to mesoscale studies. These include signal extraction and measurement errors, mapping and merging, velocity estimation, and sampling issues.

2.1 OCEAN SIGNAL EXTRACTION

The altimetric observation of the sea surface topography S can be described by:

$$S = N + \eta + \varepsilon \qquad (1)$$

where N is the geoid, η the dynamic topography and ε the measurement errors. Present geoids are not generally accurate enough to estimate globally the absolute dynamic topography η except at very long wavelengths. The variable part of the dynamic topography η' ($\eta' = \eta - \langle\eta\rangle$) or Sea Level Anomaly (hereafter SLA) is, however, easily extracted since the geoid is stationary on the time scale of an altimetric mission. The most commonly used method is the repeat track method (collinear analysis). This method is suitable for satellites whose orbits repeat their ground tracks (to within ± 1 km) at regular intervals. For a given track, the variable part of the signal is obtained by removing a mean profile (e.g. over the mission duration), which contains the geoid and the quasi-permanent dynamic topography, from each profile.

Most mesoscale studies have used the repeat-track method to extract the SLA. The error variance induced by removing the mean is typically $\langle h'^2\rangle T_I /T$ where $\langle h'^2\rangle$ is SLA variance, T is duration of observation (e.g. one year) and T_I is decorrelation time of SLA. Assuming a decorrelation time T_I of 20 days, this amounts to a relative quadratic error between 5% (T/P) and 10% (ERS-1/2) for a one-year mean. Note also that the mean profile calculation procedure is complicated by the fact that altimeter data are generally gappy. This means that only partial profiles are used to estimate the mean profile to avoid a contamination in the mean sea level from the orbit errors. However, there are several ways to minimize this problem (e.g. Chelton et al., 1990). The sampling rate also induces aliasing of frequencies higher than the Nyquist frequency [(20 days)$^{-1}$ for T/P, (70 days)$^{-1}$ for ERS and (34 days)$^{-1}$

for GEOSAT].

Measurement errors ε can significantly affect the estimation of altimeter SLA (see Chapter 1). Most of these errors are of large scale although there are probably mesoscale components which are not well quantified. Spectral analysis of known error sources show that for wavelength shorter than 1000 km (mesoscale range), the oceanic signal dominates the errors (e.g. Fu, 1983b; Le Traon et al., 1990). The only exception concerns the short wavelengths (below 100 km) where altimeter instrumental noise and small scale ageostrophic variability (e.g. internal waves) starts to dominate the altimeter SLA spectrum. Cross-track geoid errors may also affect the small wavelengths but the effect is small and limited to very specific areas (e.g. fracture zones) (e.g. Minster et al., 1993).

The altimeter data processing for mesoscale studies should thus include a correction for long wavelength errors and a reduction of instrumental noise. A commonly used approach to correct the long wavelength errors is to approximate them by a first or second degree polynomial over a given arc length. Although the polynomial adjustment method induces non-negligible errors in the mesoscale signal (Le Traon et al., 1991), it is still necessary to use it for satellites or regions where the large scale errors are important (e.g. orbit error) and/or are not well known. Alternatively, to minimize the removal of the oceanic signal, one can use more complex methods such as global crossover minimization or inverse techniques. The instrumental and small scale noise can be reduced using appropriate low-pass filtering of along-track data (see also discussion in section 2.3).

2.2 MAPPING AND MERGING OF MULTIPLE ALTIMETER MISSIONS

For mesoscale studies, there are two predominant approaches. The first approach uses along-track SLA data which provide a very good one-dimensional spatial sampling. Along-track altimeter data have been used extensively for computing mesoscale statistics. The second approach uses maps derived from one or preferably several altimeters. Maps can provide a synoptic view of mesoscale eddies allowing a much better visualization and understanding of eddy dynamics. However, there are particular processing and sampling issues which must be considered when dealing with maps. Sampling issues will be detailed in section 2.4 and we will focus here on the processing issues.

The mapping can be done using optimal interpolation methods which use an a priori knowledge of the space and time scales of the ocean signal. When data are mapped onto a regular space/time grid using such interpolation methods, the along-track long wavelength errors (or high-frequency ocean signals) can induce artificial cross-track gradient at smaller scales and thus spurious eddy signals (Le Traon et al., 1998). The effect is particularly important in low eddy energy regions and when several altimeter data sets are merged. Along-track long wavelength errors can also lead to other serious problems. A well-known example comes from GEOSAT for which errors in the tide correction were aliased to produce spurious Rossby wave like signals in the mapped data (Chelton and Schlax, 1994). To minimize these problems, the mapping method should either take into account an along-track long wavelength error (i.e. a correlated noise due to orbit, tidal or inverse barometer residual errors) or the long wavelength errors should be removed before the mapping (Le Traon et al., 1998).

The merging of multi-satellite altimeter data sets is generally necessary to map the mesoscale variability. This is not an easy task. To merge multi-satellite altimetric missions, it is first necessary to have homogeneous and inter-calibrated data sets. Homogeneous means that same geopotential model and reference systems for the orbit and same (as far as possible) instrumental and geophysical corrections should be used (e.g. same tidal models, same meteorological models, etc). Inter-calibrated means that relative biases and drifts must be corrected and also that the orbit error must be reduced. An effective methodology is to use the most precise mission (T/P, Jason-1) as a reference for the other satellites (Le Traon et al., 1995; Le Traon and Ogor, 1998). When altimetric data have been homogenized and inter-calibrated, the next step is to extract the SLA for the different missions. It is preferable that the SLAs from different missions are calculated relative to the same ocean mean using a common reference surface (e.g. either a very precise mean sea surface or mean profiles consistent between the different missions). The final step is to merge the SLAs from the different missions via a mapping or assimilation technique. Ducet and Le Traon (1999) provide a detailed analysis of the merging of T/P and ERS-1/2 over a five year period, in particular to quantify the contribution of merging for the description of the ocean mesoscale circulation.

2.3 SURFACE GEOSTROPHIC VELOCITY CALCULATIONS

Away from the equator, sea surface velocities can be calculated from the along-track sea surface slope using the geostrophic relation, i.e.,

$$f\, u = -\, g\, \partial\eta/\partial y \qquad (2)$$

where u is the surface velocity normal to the track (positive eastward), g is gravitational acceleration, f is the Coriolis parameter, y is the alongtrack direction and η is the SLA. That is, the ocean geostrophic currents are associated with a sea level slope across the current. The sea surface slope can be 1 m across the Gulf Stream, for example, and as such is easily detected by altimetry.

Note that this calculation gives only the cross-track component of the residual flow. In many altimetric studies, we assume that the flow is isotropic so that the variance of the cross-track component is taken as representative of the variance of the total surface velocity. This isotropic approximation is valid for a turbulent field, but fluctuations become anisotropic in the presence of mean shear, close to bathymetry, or at larger scales and in the equatorial regions (β effect) where the zonal fluctuations tend to dominate meridional fluctuations. Another point is that the direction of the cross-track component changes with the curvature of the ground track: the component measures more zonal variability near the equator while the part due to meridional flow becomes equal or larger at high latitudes.

The geostrophic relation is a good first order approximation for estimating the flow field, and is used extensively with hydrographic data as well as altimetry. However, approximately 10% of the flow field is not in geostrophic balance, including the frictional surface wind-driven Ekman transports which flow across isobars. Ageostrophic flow also occurs in frontal zones, in regions of meander formation, and associated with cyclostrophic effects in small eddies.

As the calculation of the derivative acts as a high pass filter, cross-track geostrophic velocities are much less sensitive to large scale signals and are thus more representative of the mesoscale signals. The computation of geostrophic velocity is sensitive to measurement noise and small scale ageostrophic ocean signals and needs a careful filtering of altimeter data (e.g. Zlotnicki et al., 1993; Morrow et al., 1994; Strub et al., 1997). Previous estimates obtained with GEOSAT data used filtering over at least 100 km to significantly reduce the contamination of noise. Even with the more accurate T/P data, the measurement noise influences the smaller spatial scales out to 50-70 km. As an illustration, Figure 1 gives the estimation of the impact of a SLA white noise of 3 cm rms on the error on the velocity field for different choice of Lanczos low-pass filtering cut-off wavelengths and for different latitude bands. This shows, for example, that in order to have velocity errors below 5 cm/s rms, it is necessary to filter the SLA wavelengths shorter than 200 km at 10°N and 80 km at 40°N. In altimetric studies, a satisfactory trade-off needs to be found between noise reduction and ocean signal removal, otherwise the filtered SLA data may miss a significant fraction of the eddy energy. This holds, in particular, at high latitudes, where eddies can have spatial correlation scales below 50 km. An analysis of the effect of different filter cut-off wavelengths on the estimation of cross-track velocity variance can be found in Stammer (1997).

2.3.1 ESTIMATING ORTHOGONAL VELOCITY COMPONENTS AT CROSSOVER POINTS

The magnitude and direction of the residual velocity can be resolved at crossover points by a simple geometric transformation. There, we have available two (generally non-orthogonal) components of geostrophic velocity from each of the ascending and descending passes ($V_{a'}$, $V_{d'}$), and a known angle (ϕ) between the ground-track and the north meridian, which varies as a function of latitude only. The residual velocity can be calculated in any orthogonal projection, e.g., in east / north components (u', v') following Parke et al. (1987) :

$$u' = \frac{V_{a'} + V_{d'}}{2\cos\phi} \tag{3}$$

$$v' = \frac{V_{a'} - V_{d'}}{2\sin\phi} \tag{4}$$

The errors associated with the geometric transformation in equations (3) and (4) can also be determined. If the residual sea surface slope measurements from the ascending and descending passes have a variance error of $\sigma_{s'}{}^2$, then using the law of propagation of variances and assuming $V_{a'}$ and $V_{d'}$ are independent, the variance of the errors in the u' and v' components will be:

$$\sigma_{u'}{}^2 = \frac{1}{2}\left(\frac{g\sigma_{s'}}{f\cos\phi}\right)^2 \tag{5}$$

$$\sigma_{v'}{}^2 = \frac{1}{2}\left(\frac{g\sigma_{s'}}{f \sin\phi}\right)^2 \quad (6)$$

If there were no measurement errors in the corrected altimeter data and no time lag between the ascending and descending passes, the geometrical transformation of equations (3) and (4) would resolve the orthogonal velocities perfectly. However, equations (5) and (6) show that the error in each velocity component depends on the estimated measurement error, $\sigma_{s'}$, i.e., how well the geostrophic sea surface slopes can be estimated from the filtered along-track residuals. The error in u' and v' will also vary with latitude, depending on three factors: the Coriolis parameter *f*, the crossover angle ϕ, and the time difference between the ascending and the descending passes (Morrow et al., 1994). For example, at low latitudes as the crossover angle, ϕ, decreases, the geometrical transformation tends to map more of the error into the northward velocity component. Particular care needs to be taken when applying this technique in low-eddy energy regions where measurement errors can dominate the ocean signal.

2.3.2 ESTIMATING ORTHOGONAL VELOCITY COMPONENTS FROM MAPPED DATA

The velocity anomaly field can also be derived from gridded fields of SLA or can be directly mapped from along-track SLA data (e.g. Le Traon and Dibarboure, 1999; Ducet and Le Traon, 1999). Mapping has the advantage of combining the information from ascending, descending and neighboring tracks. Even at crossover points, they can thus provide a slightly better estimation than the previous method because they take into account the sea level gradients between the tracks. They also allow a more rigorous treatment of measurement errors and are actually the only means to extract velocity from a combination of several altimeters. The drawback is that mapping errors are not homogeneous (see section 2.4) and this should be taken into account in the interpretation. The combination of T/P and ERS-1/2 yield, however, rather homogenous U and V mapping errors generally below 25% of the signal variance (Le Traon and Dibarboure, 1999). An analysis of velocity statistics derived from T/P and ERS-1/2 combined maps can be found in Ducet and Le Traon (1999).

2.3.3 REYNOLDS STRESSES AND VELOCITY VARIANCE ELLIPSES

Velocity variance statistics can be calculated from the time series of orthogonal velocity components, either at each crossover point or each grid point of the mapped data. From these time series we can compute the Reynolds stress terms, which include the north/east velocity variance terms ($\overline{u'^2}, \overline{v'^2}$) and the covariance term $\overline{u'v'}$. The magnitude and the direction of the eddy variability can be represented using variance ellipses or axes (see Morrow et al., 1994; after Preisendorfer, 1988). The direction, θ, of the axis of principal variability, measured anti-clockwise from east, is :

$$\tan\theta = \frac{\sigma_{11} - \overline{u'^2}}{\overline{u'v'}} \tag{7}$$

where the magnitude of the variance along the major axis and is given by

$$\sigma_{11} = \frac{1}{2}\left(\overline{u'^2} + \overline{v'^2} + \sqrt{\left(\overline{u'^2} - \overline{v'^2}\right)^2 + 4\left(\overline{u'v'}\right)^2}\right) \tag{8}$$

and along the minor axis by

$$\sigma_{22} = \left(\overline{u'^2} + \overline{v'^2}\right) - \sigma_{11} \tag{9}$$

Anisotropic flow is represented by an elongated ellipse, with the principal direction of the velocity variance aligned with the direction of the major axis. The orientation of the ellipse depends on the covariance term : the major and minor ellipse axes define the co-ordinate system in which u' and v' are uncorrelated. Ellipses with a major axis oriented in the northeast quadrant have a positive $\overline{u'v'}$; ellipses oriented towards the southeast quadrant have a negative $\overline{u'v'}$. As such, the direction of horizontal eddy momentum flux can be inferred from the ellipse orientation. Isotropic flow is represented by circular ellipses and zero covariance.

2.4 SAMPLING ISSUES

The along-track sampling of altimeters is quite well suited for resolving mesoscale eddies. A combination of several altimeter missions is required, however, for a 3D (x,y,t) mesoscale variability mapping. Le Traon and Dibarboure (1999) have quantified the mesoscale mapping capability when combining various existing or future altimeter missions in terms of SLA (Figure 2) and zonal (U) and meridional (V) velocities. Their main results which take only into account the different sampling characteristics for each satellite are as follows :

- The GEOSAT (or GEOSAT Follow On) 17-day orbit provides the best sea level and velocity mapping for the single-satellite case. The T/P+Jason-1 (interleaved T/P - Jason-1 tandem orbit scenario) provides the best mapping for the two-satellite case. There is only minor improvement, however, with respect to the T/P+ERS (or Jason-1+ENVISAT) scenario.

- There is a large improvement in sea level mapping when two satellites are included. For example, compared to T/P alone, the combination of T/P and ERS has a mean mapping error reduced by a factor of 4 and a standard deviation reduced by a factor of 5. Compared to ERS, the reduction is smaller but still a factor of more than 2. The improvement in sea level mapping is not as large when going from two to three, or three to four satellites.

- The velocity field mapping is more demanding in terms of sampling. The U and V mean mapping errors are two to four times larger than the SLA mapping error. The contribution from a third satellite is also more significant than for SLA. Only a combination of three satellites can actually provide a velocity field mapping error below 10% of the signal variance. Mapping of the meridional velocity is less accurate but by only 10% to 20% even at low latitudes. This suggests that the criterion of a satellite orbit with a rather low inclination for a better estimation of the velocity field is not really relevant, in particular for a multi-satellite configuration.

The main conclusions of that study are that at least two (and preferably three) missions are needed to map the mesoscale ocean circulation and that existing and future two-satellite configurations (T/P and ERS and later on Jason-1 and ENVISAT) will provide a rather good mapping of mesoscale variability. Note that these conclusions differ from those of Greenslade et al. (1997) who analyzed the resolution capability of multiple altimeter missions and concluded that the mesoscale signal cannot be mapped with an acceptable accuracy. This can be explained by the definition of resolution chosen by Greenslade et al. (1997). They required a very homogenous mapping error which cannot be achieved with two, three or even four satellites (in particular, the mesoscale signal will be always better estimated along the tracks). A resolution definition is, however, a fundamentally subjective choice and, although the mesocale mapping errors are not homogenous, Le Traon and Dibarboure (1999) argue that they remain sufficiently small relative to the signal (see Figure 2). All three satellite configurations they analyzed have, for example, a mapping error always below 10% of the signal variance.

3. OCEAN CURRENTS

In Chapter 2, the global, large-scale ocean circulation was described in detail. In this chapter, we will take another approach and look at each individual current system (Figure 3), and show how altimetry has improved our understanding of the dynamics of these ocean currents. Although all open ocean currents are essentially driven by external influences (principally wind stress, and heating/cooling), their responses are quite different. Not only are there regional differences in the atmospheric forcing, but the circulation in each ocean basin can be strongly controlled by local bathymetry and inter-ocean exchanges. Altimetry provides a regular sampling of all of these ocean currents, from the fast, energetic currents at the western boundaries and in the Antarctic Circumpolar region, to the slow, more quiescent flow in the center and east of most ocean basins.

Ocean currents are also highly variable. Shallow surface currents can be associated with seasonal changes in the mixed layer. The sea surface slope associated with these currents can also increase or decrease as the volume transport changes seasonally. The current's mean position also changes, influenced by meanders, the formation and separation of rings and eddies, and by the interaction with westward propagating waves and eddies. Altimetry allows us to monitor this variability, a task which is nearly impossible with in-situ hydrographic and moored current meter data alone. However, to estimate absolute velocities requires an accurate estimate of the marine geoid, or an independent estimate of the mean ocean surface, and thus the mean current. Transport estimates also require a knowledge of the vertical structure or subsurface measurements.

In the following section, we review some of the main techniques used to estimate the absolute velocities and transports of the major current systems from altimeter data. We will then look at each major current system separately, and review the considerable progress made by altimetric studies in the main western boundary currents, eastern boundary currents, open ocean currents (such as the Circumpolar and Agulhas Currents), and in the semi-enclosed seas. Studies on near-equatorial currents and eddies will be addressed separately in Chapter 4, Tropical Circulation.

3.1 ESTIMATING THE ABSOLUTE VELOCITIES AND TRANSPORTS FROM ALTIMETRY

Accessing the time-mean component of the flow is necessary for transport calculations, but a difficult task with altimetry given the large geoid errors at small spatial scales of O(100-200 km). One technique that has been developed to estimate the surface transport is to combine altimeter residual heights with a simple analytical model of the mean surface velocity profile (Kelly and Gille, 1990; Kelly, 1991; Tai, 1990). The analytical surface velocity model they use has a Gaussian shape which describes an isolated jet; the shape is based on previous studies of the Gulf Stream with subsurface moorings and ADCP data. The amplitude, width and position of the jet are allowed to vary as shown in Figure 4, and are adjusted to fit the altimetric residual heights data, using a least-squares technique. Kelly and Gille (1990) analyzed a single pass, and found maximum surface velocities between 1.2 and 2.0 m/s. Surface transports were also estimated by integrating the velocities along each track from 34 N to 40 N, across the Gulf Stream axis. Comparisons with simultaneous ADCP velocity measurements were favorable (Joyce et al., 1990), suggesting that the technique could provide a means for monitoring the surface Gulf Stream transport from altimetry alone.

Qiu et al. (1991) have refined the Kelly and Gille (1990) method. They estimate the two-dimensional mean surface height by combining synthetic profiles along ascending and descending tracks through an inverse method. Their method also accommodates possible modifications of the Gaussian shape such as recirculation gyres. This method was applied to 2.5 years of GEOSAT data in the Kuroshio Extension. The mean sea surface height agrees well with the climatological mean obtained from hydrographic data. This type of technique worked well in the Gulf Stream and the Kuroshio Extension because the horizontal mean velocity profile is well known from in-situ data, the mean velocity is strong so the signal-to-noise ratio is high, but also because the current existed as a single intense jet throughout the two-year observation period. However, this method is difficult to apply in other regions where the mean jet is weak, too stable in time, bifurcates, or has seasonal reversals.

An alternative is to estimate the mean height profile using climatological hydrographic data. The advantage is that the mean profile is available anywhere in the ocean (with varying degrees of accuracy). A disadvantage is that the mean is often weak and the oceanic fronts quite diffuse, especially in sparsely sampled regions where a large degree of spatial smoothing is necessary. Another problem is the uncertainty in specifying the reference level. For the ACC and most western boundary currents, this problem can be serious because they have a large barotropic transport. Despite these problems, composite maps of altimetric anomalies with a hydrographic mean have been used to derive surface geostrophic velocities in the Gulf Stream Extension (Willebrand et al., 1990) and Kuroshio (Ichikawa et al., 1995). These mapped velocities were significantly correlated with available drifter velocities and independent hydrographic sections, although the mapped velocities were generally weaker than drifter velocities. This is mainly due to the objective analysis mapping which removed energy on scales less than 100 km. Another similar technique is to use a model mean dynamic topography.

Some investigators have also attempted to estimate the marine geoid itself as accurately as possible. In the Gulf Stream, Rapp and Wang (1994) and Rapp and Smith (1994) applied a gravimetric geoid model combined with simple models of the meandering jet to estimate absolute Gulf Stream parameters (locations, width, velocity and height jump across the current). The gravimetric geoid undulations were calculated using a 360 potential coefficient model, land, ship and altimeter-derived gravity anomalies, and bathymetric data. The spatial resolution of the potential model is of order 1 degree (100 km), which can be improved locally by the available in-situ data. Hwang (1996) has also estimated the mean baroclinic transports from a combination of a marine geoid model, climatology and altimetry in the upstream region of the Kuroshio. This application may underestimate the transport estimates since the geoid and hydrographic mean do not include the smaller spatial scales and the calculation does not include the barotropic component of the flow. In both these examples, removing the geoid estimate does not really improve the transport calculations, since we are in regions where there is sufficient hydrographic data to estimate the mean flow, and the simple model of a meandering jet already works well. However, the geoid approach may work better in regions where there is little current meandering, and less dense hydrographic data. We note though, that in sparse gravity data regions, the errors in the geoid at the mesoscale will still tend to obscure the mean oceanic component. In the future, gravimetric missions such as CHAMP and especially GRACE and GOCE will provide a geoid improvement by one or two orders of magnitude. This should allow an estimation of a mean dynamic topography with accuracy of about 1-2 cm rms for scales larger than 100-200 km.

A final technique to estimate absolute velocities and transports is to combine the altimeter SLAs with simultaneous in-situ data to estimate a "synthetic geoid" or more exactly a mean dynamic topography. The technique proceeds as follows. In-situ data can provide estimates of the absolute dynamic topography η (although the barotropic part maybe more difficult to estimate) and satellite altimetry gives η'; the combination of the two estimates can thus yield the mean dynamic topography $\langle\eta\rangle$. This may be a very powerful methodology but it requires a large number of simultaneous data. This technique has been applied in the Gulf Stream (Mitchell et al., 1990; Glenn et al., 1991; Porter et al., 1992, Howden et al., 1999), the Kuroshio (Imawaki and Uchida, 1997) and the Azores Current (Hernandez, 1998).

The next logical step for monitoring the transport of ocean currents is to assimilate all available altimetric and in-situ data into high-resolution ocean models. This will be discussed more completely in Chapter 5.

3.2 WESTERN BOUNDARY CURRENTS

The western boundary currents are a crucial element in the ocean circulation system. They play an important role in the poleward transport of heat, and provide a boundary between the less dense surface water of the subtropical gyres and the denser surface water of the subpolar gyres. The western boundary currents are also regions of large vorticity dissipation, which act to balance the total vorticity input into the subtropical gyres by the wind field.

The western boundary current in the North Atlantic is the Gulf Stream; its partner in the North Pacific is the Kuroshio (Figure 3). Both currents are typically 100 km wide and have surface velocities of 1-2 m/s. They are associated with a large signal in mean transport and in variability. In the southern hemisphere, the western boundary currents are less intense but more complicated. They are strongly influenced by interactions with the Antarctic Circumpolar Current in the Atlantic (for the Brazil-Malvinas Confluence zone) and the Indian Oceans (Agulhas/Mozambique Currents). In the South Pacific, the lack of intensity in the East Australian Current is perhaps due to the existence of the Indonesian Throughflow at the western boundary at 10°S, and complicated by the presence of New Zealand further east.

Historically, the western boundary currents have attracted the largest number of hydrographic measurements in each ocean basin, but even so, their variability and true mean structure were not well defined. Now, with nearly a decade of altimetric measurements available, we have made considerable progress in understanding the dynamics of western boundary currents. The wealth of altimetric studies in the western boundary current regions exists for a number of reasons. Not only are the dynamics interesting, but the signal-to-noise ratio of variability is high so that even the early generation altimeters with larger errors (GEOS3, SEASAT and GEOSAT) could detect coherent signals. The western boundary currents also have the largest density of in-situ measurements to validate the altimetric signals, so new techniques are often validated in the Gulf Stream and Kuroshio regions before they are applied in more remote ocean areas.

The variability in these intense western boundary currents tends to be dominated by meanders of the current, by the separation and reabsorption of eddies, by the interaction of the mean jet with bathymetry and by the interaction of the eastward current with westward propagating Rossby waves. These combined mechanisms generate a very complicated surface height signal. Early descriptive studies concentrated on the statistics of the variability, and validating the observed variations with concurrent in-situ data. The more recent studies have analysed a combination of altimetry, in-situ data, other satellite data and models, and as such have made much progress in understanding the physical processes governing the observed variability.

3.2.1 THE GULF STREAM

The Gulf Stream, the intense western boundary current of the North Atlantic Ocean, is perhaps the best-known current of the world. The Gulf Stream is fed to the south by the Florida Current, flows north along the North American continental slope with an increasing transport reaching 70-100 Sv, and then separates around 35°N and continues eastward as an open-ocean inertial jet. The Gulf Stream transport increases downstream due to the continual inflow of Sargasso Sea recirculation water, reaching a maximum transport of 90-150 Sv near 65°W. The Gulf Stream Extension is defined east of the Newfoundland Rise at 50° W, where the flow separates into 3 parts: the northern branch eventually becomes the North Atlantic Current, the eastward branch feeds into the Azores Current, and a southward branch forms part of the Sargasso Sea recirculation. Current speeds reach a maximum of 1-2 ms^{-1} at the surface and decrease rapidly with depth. Instabilities of the flow form meanders which can eventually separate into eddies or rings. Approximately 8-11 warm-core rings pinch off poleward of the main flow, and 5-8 cold-core rings form equatorward of the current per year (Hummon and Rossby, 1998). These rings generally drift westward with speeds of a few kilometers per day. The Gulf Stream is very energetic and at any time up to 15% of the Sargasso Sea is covered by cold-core rings originating from Gulf Stream instabilities.

The seasonal variations of the Gulf Stream have been the subject of numerous altimetric studies. An analysis of climatological data showed maximum dynamic height differences across the current occurs in November (Zlotnicki, 1991), in contrast to the maximum volume transport which occurs in spring (Sato and Rossby, 1995). Various studies based on GEOSAT data during the period November 1986 to December 1988 suggested that maximum sea level slopes (Zlotnicki, 1991) and surface transports (Kelly and Gille, 1990) occurred in late autumn (September/October) somewhat earlier than in the climatological data. The annual cycle of sea level differences across the Gulf Stream during this period explained 60% of the observed variance, with an amplitude of 9 cm. Kelly and Gille (1990) noted that the maximum surface transports occurred in late autumn when the Gulf Stream was north of its mean position. Minimum surface transports occurred in the late spring, with the Gulf Stream south of its mean position. T/P data during the period 1992-1997 confirm that the maximum surface transport and most northerly position occurs in autumn, with a minimum in meandering in summer (Wang and Koblinsky, 1995; Kelly et al., 1999).

The cause for these annual variations is most likely the upper ocean seasonal heating cycle. Vazquez et al. (1990) found the annual signal to be dominated by meanders. However, if the SSH fluctuations due to meandering are removed, most of the seasonal transport variations can be explained by seasonal heating differences across the Gulf Stream (Figure 5). Hydrographic data confirm that the seasonal variations in sea level come from the upper 250 m (Kelly et al., 1999). This surface seasonal signal, with a fall maximum, obscures the seasonal variation in total volume transport, which has a spring maximum and is dominated by the larger volume transport below 250 m. Although there are some phase differences between the steric effect and the observed SSH, Kelly et al. (1999) suggest that these minor differences may be due to seasonal variations in advection.

The idea of a seasonal modulation in the Gulf Stream meanders has been refuted by Lee and Cornillon (1995), who examined the Gulf Stream path from a much longer series of AVHRR data and found a dominant 9-month cycle instead. T/P data also reveal a dominant 9-month signal propagating southwestward from the Gulf Stream jet (Rogel, 1995) which appears to be related to incoming westward propagating Rossby waves of that frequency interacting with the jet.

Kelly (1991) described in detail the Gulf Stream meandering and structure, using the technique of a simple analytical model to estimate the mean surface velocity profile. Two different flow regimes, separated by a transition region coinciding with the New England Seamount Chain, are found west of 64°W and east of 58°W. East of 58°W, increased meandering makes the mean current twice as wide as west of 64°W. Eulerian time scales also drop by a factor of 3 and peak velocities and surface transport drop by 25%. Tai (1990) has used a similar method on an ensemble of ascending tracks in the Gulf Stream and Kuroshio extensions. Surface transport is found to increase from $90x10^3\ m^2s^{-1}$ (after leaving the coast) to reach a maximum of $130x10^3\ m^2s^{-1}$ at 63°W for the Gulf Stream and at 150°E for the Kuroshio. These results compare well with surface transport deduced from drifter data (Richardson, 1985). The variability maximum coincides with the mean position of the jet.

Time series of the Gulf Stream transport have also been calculated near 71°W from T/P altimetry, a high resolution geoid, and an empirical model for the Gulf Stream vertical structure by Howden et al. (1998). The vertical structure is a function of main thermocline depth, which is then calibrated to the sea surface topography using historical AXBT data and reprocessed GEOSAT data. The transport is corrected for seasonal steric height changes. Altimetric sea surface changes across the Gulf Stream of about 1.2 m are comparable to in-situ measurements from a weekly-repeated XBT and ADCP line between Newark, New Jersey and Bermuda. These results suggest that altimetry, combined with a tight relation for the vertical density structure, can be used to monitor changes in the Gulf Stream position and transport.

A decade of altimetric studies have provided a better statistical description of the Gulf Stream transport, its seasonal and mesoscale variability, and the importance of surface heating in generating its annual cycle. With longer time series, we will also be able to investigate interannual changes, its interaction with the subpolar and subtropical gyres, and eventually learn more on its role for interdecadal variations such as for the North Atlantic Oscillation.

3.2.2 KUROSHIO

The Kuroshio Current system is the strong western boundary current of the North Pacific subtropical gyre. The characteristic feature of the Kuroshio is that it has several quasi-stationary paths, controlled partly by topographic features. Upstream, the Kuroshio passes east of Taiwan, and then flows along the eastern boundary of the East China Sea, where its path undergoes strong seasonal meanders. Around 30°N, the Tsushima Current separates from the Kuroshio to the west of Japan. The main branch passes eastward through the Tokara Strait, south of Kyushu, and undergoes large meanders before reaching the Izu Ridge south of Honshu. The Kuroshio separates from the Japan coast near 35°N and becomes the Kuroshio Extension, where it again undergoes large meanders before passing over the Shatsky Rise at 157°E and the Emperor Seamounts at 170°E. The transport of the Kuroshio also increases along its path from about 60 Sv near 135°E (Imawaki and Uchida, 1997) to up to 130 Sv near 145°W (Wijffels et al., 1998). The Kuroshio Extension is also very energetic; in periods where the current follows a stable path, around 5 rings form each year, increasing to 10 rings per year during transition periods.

The seasonal cycle in the Kuroshio and its Extension is more variable than for the Gulf Stream, with a large interannual modulation. Zlotnicki (1991) analyzed sea surface differences across the Kuroshio Extension for the first two years of GEOSAT (1986-1988) and found a maximum peak in September/October, similar to the Gulf Stream, although he noted that the annual cycle in climatological dynamic height differences was negligible. Qiu et al. (1991) found annual variations of the Kuroshio Extension in agreement with Zlotnicki's results. They also observed that the maximum surface height difference in the Kuroshio Extension lagged by 2 months the maximum sea level difference across Tokara Strait. The Kuroshio Extension annual cycle calculated by Wang and Koblinsky (1995) from T/P data in 1993 also shows a maximum in October. As in the Gulf Stream, the maximum surface transport occurs when the current axis is north of its mean position. Annual variations are found to be significant in the upstream region (141°E to 153°E). On average, they explain only 15% of the total variance whereas interannual variations explain 23%. Qiu (1992) found that the seasonal transport variations were closely related to the seasonal change in intensity of the southern recirculation gyre. However, other authors suggest that the principal cause of this seasonal cycle is differential heating over the Kuroshio Extension (Zlotnicki, 1991; Wang and Koblinsky, 1995).

Wang et al. (1998) have separated the annual and intra-annual sea level variability into separate spatial scales. They find the large-scale annual variability (> 2000 km) is essentially a standing oscillation, related to seasonal heating. The shorter-scale annual and intra-annual variability (800 to 1500 km) is found to have strong, westward phase propagation and is associated with large bathymetric features. In particular maximum low-frequency wave activity occurs west of the Shatsky Rise (~160° E).

Pronounced interannual fluctuations are evident in the Kuroshio Extension region. Qiu (1992) suggests that interannual variations during the GEOSAT period were possibly influenced by the intensification of the subtropical wind gyre with the 1986-1987 ENSO event. During the first 2 years of T/P, Qiu (1995) also noted significant variations in EKE levels in the Kuroshio Extension and its recirculation gyre (Figure 6). Significant interannual changes in EKE were also observed by Adamec (1998). Qiu (1995) and Adamec (1998) analyzed those changes in terms of eddy-mean flow interaction (see discussion in section 4.6).

Mean velocities and transports have been estimated in the Kuroshio region using a number of different methods. The technique of combining a simple Gaussian model of the mean jet with the variable flow characteristics derived from altimetry has been applied to map the mean sea surface and variations in the Kuroshio Extension from 2.5 years of GEOSAT data (Qiu et al., 1991), and including the associated recirculation gyres (Qiu, 1992; Qiu, 1995). Using the derived absolute surface velocities, the authors investigated quasi-stationary meanders, and the relative eddy and mean kinetic energies. The ratio of EKE to mean kinetic energy is high with a nearly constant value of 1.5 to 2.0 along the Kuroshio Extension path. Propagation of mesoscale fluctuations is generally westward except for the upstream region of the Kuroshio Extension.

Recently, the ASUKA Group[1] undertook an ambitious program to monitor the absolute volume and heat transports of the Kuroshio and its recirculation south of Shikoku, Japan. ASUKA carried out repeat hydrographic surveys under a T/P satellite groundtrack near 135°E during 1993-1995 (Imawaki and Uchida, 1997); the intensive survey also included 9 current meter moorings with 33 current meters, 2 upward-looking ADCPs and 10 inverted echo sounders (Figure 7a). The volume transport in the top 1000 m showed a very tight relation with the surface dynamic topography (Figure 7c). This means that variations in the sea surface slopes across the Kuroshio measured by altimetry can be used to monitor the variations in total transport. For absolute transport estimates, a mean sea surface profile was derived from the along-track hydrographic data, and combined with the 10-day altimetric anomalies. The average volume transport was estimated to be 63 ± 13 Sv, with no apparent seasonal cycle although interannual and intraseasonal cycles were very strong (Figure 7b). The latter were probably influenced by fluctuations in local stationary eddies. With such an accurate alongtrack estimate of the mean profile, the surface geostrophic velocities compared

[1] Affiliated Surveys of the Kuroshio off Cape Ashizuri

extremely well with surface drifters (Uchida et al., 1998), and highlighted the drifter sampling bias as they tend to converge in the high-velocity core of the Kuroshio.

These altimeter studies of the Kuroshio Current System have improved our understanding of seasonal heating variations on the seasonal transport cycle, and the important role of topographic forcing in generating meanders, Rossby waves, and other low-frequency variations. With longer time series, altimetry has also revealed a strong interannual signal, which may be directly influenced by tropical Pacific variations, but may also have a direct effect on the dynamics of the subpolar gyre. The Kuroshio appears an important conduit between the subtropical and subpolar ocean regions, and future altimetric studies will certainly examine its role in modulating the circulation and climate of the North Pacific.

3.2.3 BRAZIL-MALVINAS CONFLUENCE REGION

The two western boundary currents of the South Atlantic are the Brazil Current and the Malvinas Current. In this region, the Brazil Current flows poleward along the continental slope of South America. The Malvinas Current originates as a branch of the Antarctic Circumpolar Current. Between 35°S and 39°S, the two current systems, which have strongly contrasting water types, converge in the Brazil/Malvinas Confluence region and form a strong frontal structure. The Confluence region is associated with an intense and complex mesoscale variability field characterized by rings, eddies and filaments. Several studies have used GEOSAT data to analyze the mesoscale variability of the Brazil/Malvinas Confluence area (Provost and Le Traon, 1993; Forbes et al., 1993; Goni et al., 1996). The mesoscale field is highly inhomogeneous with low sea level variability and EKE in the Malvinas Current (< 8 cm and 150 cm^2s^{-2} respectively), intermediate values in the Brazil Current (16 cm and 800 cm^2s^{-2}) and high values in the Confluence region (30 cm and 1700 cm^2s^{-2}). Provost and Le Traon (1993) showed that the mesoscale variability has a marked anisotropy with larger meridional velocity variance and is dominated by relatively large spatial scales and low frequency fluctuations. Contrary to most western boundary and open ocean currents, they found little energy at the annual frequency while there is a significant semi-annual signal (Figure 8), probably indirectly related to atmospheric forcing. Note that semi-annual Rossby waves were also observed in the interior of the South Atlantic subtropical gyre (Le Traon and Minster, 1993). More recently, Vivier and Provost (1999) showed that T/P data can be used to monitor the volume transport of the Malvinas Current. The transport is estimated from T/P data using a priori statistical information on the vertical structure of the flow. T/P transport compares favorably (correlation of 0.8) with estimates from current meter data. Transport time series derived from 3 years of T/P data show salient periodicities between 50 days and 80 days and at the semi-annual period. Little energy is again found at the annual period. They suggest that the semi-annual variability may be remotely forced by winds at Drake Passage. The semi-annual periodicity of the Malvinas Current may explain the importance of the semi-annual signal in the Brazil/Malvinas Confluence region.

3.2.4 EAST-AUSTRALIAN CURRENT AND EAST AUCKLAND CURRENT

The western boundary current of the South Pacific, the East Australian Current (EAC), flows south along the Australian coast, then separates near 32°S to form the Tasman Front, re-attaches to the shelf break near North Cape, New Zealand, then continues southeast alongshore as the East Auckland Current. Although the mean flow associated with the EAC is weaker than other western boundary currents, the eddy field is quite energetic. Using a combination of T/P altimeter data and AVHRR sea surface temperature data, Ridgway et al. (1998) observed that the eddies, both cyclonic and anticyclonic, are advected southward in the EAC about 3 times per year. These eddies appear phase locked with westward propagating Rossby waves generated in the eastern Pacific. They suggest that the eddies develop as instabilities when the Rossby waves interact with the eastward mean flow in a region of sharply deepening bathymetry. Wilkin and Morrow (1994) calculated eddy kinetic energy and Reynolds stresses at GEOSAT crossover points in the Tasman Sea region. They found high EKE associated with instabilities of the EAC extended from the coast to the Lord How Rise at 160°E. Velocity variance ellipses from the altimeter and drifter data showed distinct anisotropic structure (Figure 9), with a strong alongshore component at the coast, with offshore Reynolds stresses converging around the first semi-permanent meander of the Tasman Front. Farther offshore, the ellipses are orientated north-south, consistent with meridional velocities associated with the incoming Rossby waves.

The East Auckland Current has been investigated with a combination of T/P altimetric data, broad-scale XBT data, two repeating high resolution XBT transects, and neutrally buoyant floats by Roemmich and Sutton (1998). The mean transport of the East Auckland Current is about 9 Sv, with an additional 10 Sv in the recirculation of three permanent warm-core eddies. Their study further compares the extensive time-series data with hydrographic climatology, and is one of the first studies to quantify the error in the estimated climatological mean field, in particular from the significant mesoscale contribution.

3.3 EASTERN BOUNDARY CURRENTS

The eastern boundary currents are generally characterized as diffuse, equatorward flows which extend from the eastern boundary into the ocean interior, as part of the return flow of the wind-forced gyres. Superimposed on this broad equatorward flow are narrow coastal currents and often upwelling currents, driven either by the wind stress curl or the equatorward wind stress which exists along the eastern boundaries. These coastal currents are noted for their equatorward surface currents and poleward undercurrents, but can be highly variable, even reversing under certain conditions. Altimetric studies of the broad eastern boundary currents are rather limited, essentially for the same reasons why western boundary current studies are so numerous. From an altimetric perspective, they have lower signal-to-noise ratios, especially for the earlier missions; they have fewer in-situ measurements for validation; and at first order, their variability has less influence on the overall dynamics than for western boundary currents. The narrow coastal boundary currents have also been difficult to measure from altimetry, since they are sampled by few data points, are subject to various coastal errors (tidal errors, non-isostatic pressure responses, altimeter or radiometer data dropouts close to land), and the wind-forced currents can have a strong ageostrophic component which is not seen by altimetry.

Despite these early measurement difficulties, the new generation of more precise altimeters such as T/P have allowed us to monitor the variability associated with these eastern boundary currents, and in some cases measure their velocities and transports. This is not only important for understanding the dynamical processes at the eastern boundaries, but also because perturbations and instabilities at the eastern boundary can then propagate westward as Rossby waves, and so influence the ocean interior at later periods. In the following section, we will consider each eastern boundary region separately, and look at how altimetry has aided our understanding of eastern boundary current variability.

3.3.1 CALIFORNIA CURRENT

The California Current System on the eastern boundary of the North Pacific is one of the few eastern boundary regions to be extensively studied with altimetry. This region is one strong seasonal currents forced by the large wind stress curl variability. The mainly equatorward current system also reverses seasonally, with periods of poleward flow over the shelf from

October to March (also known as the Davidson Current). A combination of GEOSAT altimeter data and satellite AVHRR sea surface temperature data have been used to describe the large-scale summer circulation (Strub and James, 1995). The circulation is characterized by a large-scale equatorial jet and temperature front that lies close to the Oregon coast (20-50 km), and extends farther offshore from the Californian coast in a convoluted, meandering jet. Eddies associated with this jet may persist for 3-6 months, and play an active role in offshore transport. A comparison between T/P altimetry and sub-surface current meter data shows that the altimeter resolves horizontal scales of 50-80 km in the alongtrack direction (Strub et al., 1997). The rms difference between the altimeter and current meters was around 7-8 cm/s, and mostly due to strong and persistent small-scale variability in the currents. Velocity variances, eddy kinetic energy levels and the major axis of variance ellipses were in good agreement, and demonstrated the anisotropic nature of the variability. Most of the high eddy energy within 500 km of the coast is associated with the the seasonal jet which develops each spring and moves offshore, rather than deep ocean eddies moving onshore.

A comparison of EKE measured by different instruments has also been made in the Californian Current region (Kelly et al., 1998). Surface drifters, T/P altimetry data, and moored current data all show similar seasonal cycles in EKE, with maximum values in late summer/autumn, a month or so earlier than for the western boundary currents. The velocity fluctuations were typically 240-370 km wavelength, and lasted several months. In the northern section (36° to 41°N), the maximum EKE migrates zonally on a seasonal time scale, with maximum EKE associated with increased equatorward flow. A simple model was used to show that the seasonal variations in EKE amplitude were related to wind stress curl forcing, but this did not explain the offshore movement of the front.

The zonal migration in EKE is probably related to annual Rossby waves. White et al. (1990) first studied these waves using GEOSAT altimetry. They found that altimetric sea level residuals showed local maxima at three key locations off the Californian Coast : at Point Eugenia at 27°N, SW of Point Conception at 32°N, and between Monterey and Cape Mendicino (37° to 40°N). These local maxima were mainly associated with annual Rossby waves (53% variance), which originated near the coast, with wavelength scales of 400-800 km, periods of 6-12 months, and with westward phase speeds at 2-5 cm/s, faster at lower latitudes which were consistent with Rossby wave theory. In a later study, Pares-Sierra et al. (1993) compared two numerical models to test whether these waves were generated by a coastal response to wind forcing or baroclinic instabilities of the California Current. Only the wind-forced model simulated correctly the distribution of eddy variance, suggesting that the dominant source of this mesoscale eddy activity is the wind forcing adjacent to the coast, modified by both Rossby and Kelvin wave dynamics.

These studies have concentrated on the offshore dynamics. The narrow coastal currents are more difficult to access with altimetry, and statistical gridding techniques often give unrealistic results when extrapolated towards the data-sparse coastal region. An alternate method is to combine offshore altimeric sea level measurements with coastal tide gauge data. This has been done for the California Current region using T/P and ERS-1 altimeter data (Strub and James, 1997) and a “successive correction” method, which results in a smoother field in data sparse regions and a more detailed field around the data. The tide gauge data are filtered to remove both tides and coastal trapped waves, using a 20-day filter. Including the tide gauge data lowers the sea level along the coast, and reveals an alongshore jet close to the coast which is not sampled by altimetry. Satellite SST data confirm that the jet flows along the outer edges of colder upwelled water, which is an expected pattern for this region of coastal upwelling. However, several cold filaments with horizontal scales of 50 km are apparent in the SST data which are not resolved by the gridded altimetric/tide gauge data. This reaffirms that features with scales less than 100 km are not adequately resolved by altimetry, even when combining T/P and ERS-1 data.

3.3.2 ALASKA CURRENT AND GULF OF ALASKA

North of the California Current system is the Alaska Current, which forms the eastern component of the Pacific subpolar gyre. The current is constrained to the shelf region by a strong cross-shelf pressure gradient which is maintained by the freshwater input from Alaska’s rivers which reduces the coastal density. Most energetic eddy activity is confined to the shelf/slope and boundary current region, with little eddy energy in the gyre interior (Lagerloef et al., 1994). The coastal mesoscale eddy activity is around 100 km wavelength (Matthews et al., 1992), the most notable being the Sitka eddy around 57°N which has an approximate annual cycle interior (Lagerloef et al., 1994). Two-dimensional spectral analysis across the Gulf of Alaska reveals both westward and eastward propagating features at this latitude, with mainly annual period and longer wavelength of 1000 km. Using a longer time series of nearly 4 years of GEOSAT data, Bhaskaran et al. (1993) also identified significant interannual variability, which was closely linked to baroclinic variations detected by hydrographic data.

3.3.3 THE PERU/CHILE CURRENT

The Peru/Chile current is the eastern boundary current of the South Pacific subtropical gyre. The equatorward current transports enough cooler water from higher latitudes to lower the SST along the South America by several degrees from the zonal average. Superimposed is a vigorous upwelling circulation, which lowers the coastal SST by another 2°-4°C. Strub and James (1997) used maps of T/P and ERS-1 altimetry to study the nearshore circulation off Chile. They described the nearshore equatorward jet which was aligned along the SST front between February/March 1993. The jet appeared to break up into a series of eddies in April, which widened the region of cooler coastal water by 100 km. Longer-term monitoring of this coastal jet was difficult due to the poor spatial resolution (100 km) of the gridded altimeter data. Strub et al. (1995) have also monitored the Peru/Chile Countercurrent, which flows from 10°S to 35°-40°S, and lies approximately 100-300 km offshore. Although the temporal mean current was not available in the altimetric observations, they assumed the mean countercurrent was poleward, based on historical observations. The altimeter variability allowed them to deduce maximum velocities in spring and minimum in autumn.

3.3.4 LEEUWIN CURRENT

The dynamics of the eastern boundary current in the southern Indian Ocean are quite unusual, since it flows poleward against the prevailing equatorward winds. The current is forced by the large alongshore pressure gradient along the West Australian coast; which is five times greater than for other eastern boundary regions. This large pressure gradient is maintained by the lower density water which enters to the north from the Indonesian Throughflow. The alongshore pressure gradient drives an eastward geostrophic flow towards the coast; when geostrophy breaks down at the coast the flow accelerates poleward down the pressure gradient, and overcomes the equatorward wind-driven flow. Morrow and Birol (1999) have used T/P altimetry to monitor the variability of this alongshore pressure gradient; the mean pressure gradient is derived from climatological data. The altimeter data confirms previous studies that the alongshore pressure gradient is maximum in May, when the coastal boundary current, the Leeuwin Current, is strongest (Figure 10). However, the 3 year time series also reveals a semi-annual signal, with a secondary peak in November which is not evident in the climatology, although present for this period in XBT data. There is also significant interannual variability in the pressure gradient, and also in the structure of instabilities associated with the Leeuwin Current. Away from the coast, T/P data reveals a band of energetic Rossby waves between 20°-35°S which propagate westward across the basin, originating near the eastern boundary, with near semi-annual periods of 100-200 days. Birol and Morrow (1999) used a simple vorticity model forced by ERS-1 scatterometer winds, and an eastern boundary condition derived from XBT data to determine that these semi-annual waves are not locally wind forced, but appear remotely forced from the north.

3.3.5 BENGUELA CURRENT

The Benguela Current is the broad equatorward flow in the southeast Atlantic Ocean. As in other eastern boundary currents, the flow is dominated by geostrophic eddies, and is associated with a strong upwelling system. This current system is also instrumental in the Indian/Atlantic exchange of mass, heat and salt, since eddies from the Agulhas region (see below) are often entrained by the Benguela Current and transported north. Altimeter data has been combined with in-situ data to monitor the transport in the Benguela Current system by Garzoli et al. (1997). From the in-situ data, they derive a correlation function between sea surface elevation and thermocline depth, indicating a dominant baroclinic flow. With a combination of these correlations and a linear stratification geostrophic relation, they can calculate baroclinic transports. During the 3 years from 1993-1995, the mean northward baroclinic transport is between 12 and 15 Sv, with fluctuations up to 25 Sv due to high velocities associated with ring shear. The interannual variability in transport was small, although the source regions changed from year to year. In 1993/1994, the main contribution comes from the South Atlantic, with equal partitions of Indian Ocean and tropical Atlantic contributions. In 1995, the main contribution was from the Indian Ocean, for reasons yet unknown.

3.3.6 CANARY CURRENT

The easternmost branch of the Azores Current (see below) feeds the Canary Current, an essential part of the North Atlantic eastern boundary current system. Due to the seasonal shift of the trade winds coastal upwelling north of 25°N is strongest in summer and fall. Filaments of cold water from the upwelling regime often extend several hundred kilometers offshore. Most of them develop near the capes at the African coast and can be traced in satellite images of sea surface temperature or pigments (Hernandez-Guerra and Nykjaer, 1997). An analysis of more than 5 years of T/P and ERS-1/2 data in the Canary basin (Hernandez and Le Traon, in preparation) clearly show the seasonal variation of the Canary Current and the associated upwelling. There is a remarkably strong variability south of the Canary Islands (7-9 cm rms, 150-200 cm^2/s^2), which is related to the eddy field generated by instabilities of the Canary Current as it crosses the archipelago, and by the wind stress curl forcing downstream from the Islands (Aristegui et al., 1994). The signal has a very clear seasonal modulation with a maximum EKE observed in summer and fall.

3.4 OPEN OCEAN CURRENTS

3.4.1 ANTARCTIC CIRCUMPOLAR CURRENT

The Antarctic Circumpolar Current (ACC) is unique in many respects. It is the only ocean current to flow continuously around the globe in the zonally unbounded Southern Ocean, and it provides a major link for water property exchanges between the Atlantic, Indian and Pacific Oceans. It is a region of large heat and momentum exchanges between the ocean and the atmosphere. It is also a region of high eddy variability, and ocean eddies may play a role in the southward oceanic heat transport and in transporting momentum in the circumpolar region. Before the advent of satellite altimetry, there had been few long-term observations of the ACC and its variability due to the difficulty and expense in making hydrographic or current-meter measurements in this remote and inhospitable environment. Altimetry allows us to monitor the surface signature of these remote currents and their variability.

One of the early findings from altimetry was that a vigorous eddy field existed along the path of the ACC, with large geographical variations (Zlotnicki et al., 1989). This had never been fully quantified from sporadic hydrographic sections or the limited current meter moorings, and eddy-resolving models were rather limited at this time. The spatial sampling of GEOSAT was particularly useful for Southern Ocean mesoscale studies. Regions of high mesoscale variability were significantly correlated with the mean circulation, as determined from historical hydrographic data (Chelton et al., 1990). This indicated the importance of hydrodynamic instabilities in the dynamics of the ACC. The geographical distribution of both the mean currents and their variability was also shown to be strongly controlled by bathymetry (Sandwell and Zhang, 1989; Chelton et al., 1990). Gille and Kelly (1996) analyzed the spatial and temporal characteristics of the mesoscale variability, and also concluded that local instability mechanisms were more important for the ACC dynamics than basin-scale processes.

The dynamical significance of this mesoscale variability was quantified by Morrow et al. (1994) who calculated velocity variance ellipses and Reynolds stresses at GEOSAT crossover points in the Southern Ocean. They demonstrated that the eddy variability was often anisotropic, in particular close to bathymetric features and in the vicinity of the mean current. This anisotropy is associated with horizontal eddy momentum fluxes. Calculations made along the path of the ACC showed that eddies tend to converge momentum into the mean current, and on average tend to accelerate the frontal jets. The GEOSAT data reveal a surprisingly complex geographical distribution of this eddy momentum flux convergence (and in some regions divergence), which was confirmed by comparisons with numerical models (Wilkin and Morrow, 1994).

Part of this mesoscale eddy variability is associated with propagating Rossby waves, with wavelength of about 300 km and periods of 4 to 12 months (Hughes, 1995). These waves are advected eastward in the mean axis of the ACC, and westward elsewhere, and are evident in both T/P data and the Fine Resolution Antarctic Model (FRAM), as shown in Figure 11. Morrow et al. (1994) estimated the mean axis of the ACC to coincide with the meridional maximum of $\overline{v'v'}$, which they interpreted as the mean position of the time-varying meanders. The finer-resolution FRAM model supports this, but also shows that the meanders are often wavelike with a steady eastward propagation. The characteristic patterns of eddy momentum flux convergence form a arrowhead pointing eastward (Morrow et al., 1994), this is also the hallmark of Rossby waves travelling along a waveguide formed by the eastward jet (Hughes, 1995).

The dynamics of the eddy-mean interactions near steep topographic slopes in the Southern Ocean was evaluated by Witter and Chelton (1999) using GEOSAT altimetry and a wind-forced quasi-geostrophic channel model. They found that largest eddy energies occur downstream of zonal modulations of bottom topography, in particular when topographic steering forces the flow into regions of reduced ambient potential vorticity. The type of instability processes are also modified. With zonally uniform topography, baroclinic instability and recycling of eddy energy occurs, whereas in regions with strong zonal modulations in topography, mixed baroclinic-barotropic instabilities occur with an associated strong downward transfer of eddy energy. This is an important mechanism in the transfer of wind-input momentum from the surface to deeper levels, where it can be balanced by form drag over topographic ridges. Thus the GEOSAT observations showed enhanced variability near steep topographic slopes in the ACC; the model results detail the vertical structure and suggest that these high eddy energy regions make a large contribution to the dynamics of the ACC.

The large-scale sea level variability was examined by Chelton et al. (1990) in the Southern Ocean using 2 years of GEOSAT data, smoothed to a resolution of 12° longitude by 6° latitude by 9 days. The variability was dominated by the seasonal cycle, with a zonally coherent annual component and a semi-annual component that showed large amplitude and phase changes over the 3 ocean bases. Interannual variability of the seasonal cycle was also strongly regional. Park and Gambéroni (1995) considered the large-scale circulation in the southern Indian Ocean sector of the ACC, using 18 months of T/P data. During this period and in this location, they find no significant interannual variations, and find that the variability is strongly constrained by the current-topography interactions.

Monitoring the transport of the ACC with altimetry is a tricky task. The frontal regions carry up to 75% of the transport due to their enhanced horizontal density gradients and associated geostrophic currents. However, because these fronts are so narrow and variable they are also difficult to sample with altimetry, and mapped altimetry data tends to smooth the fronts and reduce their gradients and associated currents. The ACC fronts also have a large barotropic component which is measured by altimetry, but difficult to estimate from hydrographic data. Despite these difficulties, a number of altimetric studies have estimated the mean circulation and transport of the ACC. Gille (1994) applied a model of a meandering Gaussian jet to estimate the mean sea surface across the Subantarctic Front and the Polar Front (Figure 12). The meandering jet model explains between 40% and 70% of the height variance along the jet axes; the remaining variance represents rings and eddies that separate from the jets. The fronts are substantially steered by topography, the with mean jet width of 44 km, and meanders of around 75 km either side of their mean position. The latitude of the ACC axis varies from 40°S to 60°S, and although the flow retains its sharp frontal structure, the frontal width varies with latitude due to the changing Rossby radius. The average height difference across the Subantarctic Front was 0.7 m with a 0.6 m drop across the Polar Front. In the southern Indian Ocean sector of the ACC, Park and Gameroni (1995) calculated the mean circulation using a combination of altimetry and a geoid model. Their mean altimetric circulation indicated the presence of an anticyclonic subtropical gyre north of the ACC, and two cyclonic subpolar gyres south of the current, either side of the Kerguelen Plateau. These gyre structures were in good agreement with a fine resolution numerical model, but were not evident in the climatological data, probably due to the strong barotropic component of these currents.

As a choke point, Drake Passage has been the site of numerous hydrographic campaigns, including extensive current meter measurements, so the ACC transport and its variability is reasonable well known here. Challenor et al (1996) have estimated the surface geostrophic currents in Drake Passage from a combination of altimetry, hydrography and current data. Their method depends on having in-situ density and current data along an altimeter groundtrack to specify the absolute surface current for one altimeter pass. This in-situ surface current is then used as a reference for the altimetric data, providing a time series of absolute surface current from altimetry. Their technique finds two main jets at 56.6°S and 58.5°S, which remain consistently in position. Woodworth el al. (1996) compared altimetry with bottom pressure measurements (for the barotropic component) and FRAM model sections. Although sea level and bottom pressure measurements were in good agreement in the northern part of Drake Passage, the FRAM results showed that they were virtually uncorrelated with transport fluctuations. In the southern section which is more important for transport studies, sea level and bottom pressure show significant rms differences, explained by a larger baroclinic variability. Their conclusion is that altimetry will need to be combined with extensive in-situ measurements to properly understand the dynamics of Drake Passage.

These Southern Ocean altimetric studies have highlighted the large regional differences in eddy energy and fluxes, the existence of seasonal and interannual variations, and the very strong dependence on bottom topography. However dynamical studies have also shown the difficulty in establishing a strong relationship between the altimetric surface heights, and the sub-surface velocities in this region of weak vertical density gradients and a strong barotropic component. Future altimetric studies will almost certainly rely on a combination of satellite surface measurements with in-situ data and numerical models.

3.4.2 AGULHAS CURRENT AND RETROFLECTION

The Agulhas Current system south of Africa is actually the strongest western boundary current in the Southern Hemisphere, fed to the north by the Mozambique Current, the East Madagascar Current and from recirculation in the southwest corner of the Indian Ocean. As for most western boundary currents, the Agulhas increases its transport along its path from around 70 Sv near 31°S to 100-135 Sv near 35°S (Tomczak and Godfrey, 1994). The reason why it is included in this section on open ocean currents, is that as it moves southwest into the Atlantic, the current retroflects and turns eastward in the direction of the prevailing winds and eastward flowing Antarctic Circumpolar Current. So the Agulhas Current System, including the Retroflection, is closer to an open-ocean current than a western boundary current.

The Agulhas Retroflection is one of the most highly variable oceanic regions, with mean variance levels measured by GEOSAT altimetry of around 550 cm^2, compared with levels around 350 cm^2 for the Gulf Stream (Zlotnicki et al., 1989). To the west, the high variability is mainly due to eddy shedding; the Retroflection moves westward until its western part pinches off and forms an eddy and the loop retreats to its most eastern position. Further east, the high eddy variability is due to instability processes as the current interacts with topography and the eastward flowing Circumpolar Current.

Seasonal variations in the Agulhas current and mesoscale activity have been reported by a number of studies, which used up to two years of altimeter data (Zlotnicki et al, 1989; Quartly and Srokosz, 1993). There is no clear seasonal signal either in climatological data nor the FRAM model (Quartly and Srokosz, 1993). Given the short duration of these time series, and the fact that intermittent ring shedding and Retroflection meanders may bias the seasonal results, it is difficult to quantify any true seasonal cycle –a much longer time series analysis may be necessary.

The separation of Agulhas rings from the Retroflection region, and their westward propagation into the South Atlantic has been extensively studied using altimetry. These rings may be one of the important mechanisms for the inter-ocean transport of mass, heat and salt between the Indian and Atlantic Oceans. However, given their limited spatial extent and sporadic detachment they have been difficult to measure using hydrographic sections, and so their effect on the total mass, heat and salt budgets was largely uncertain. Altimetry has allowed us to monitor the separation and propagation of these eddies away from the Retroflection region, and build a catalogue of their surface characteristics. Gordon and Haxby (1990) describe one such eddy using a combination of GEOSAT altimetry and in-situ data. Dynamic height calculated from both data sets was equivalent, furthermore this eddy had distinct signs of Indian Ocean stratification.. Most studies find that between 5 and 6 rings are shed from the Retroflection region each year, whether during the GEOSAT period (Gordon and Haxby, 1990; Feron et al., 1992; Byrne et al., 1995) or T/P period (Grundlingh, 1995; Goni et al., 1997). Eddy shedding can be detected from altimetry by sharp changes in the stability of the sea level patterns (Feron et al., 1992). An analysis of 3 years of T/P by Goni et al. (1997) found that ring shedding was neither continuous nor periodic, and for long periods there were no ring formations. The eddies detected were mostly anticyclonic, and drifted west-northwest across the South Atlantic with speeds of 3-7 cms^{-1} (Grundlingh, 1995), with a slowed translation across strong topographic slopes. The residence time of the eddies in the South Atlantic was around 3-4 years, and their amplitude decayed slowly with an e-folding distance of O (1700-3000 km) (Byrne et al., 1995).

A combination of altimetry and concurrent hydrographic sections have been applied to estimate the annual mean property fluxes from the Indian to the Atlantic Ocean that are associated with these Agulhas eddies. In terms of water mass transfer, estimates range from a minimum of 5 Sv (Byrne et al., 1995) to a maximum of 15 Sv (Gordon and Haxby, 1990); van Ballegooyen et al. (1994) estimate a transport of 6.3 Sv of water warmer than 10°C, and 7.3 Sv warmer than 8°C. The total upper layer transport across 32°S in the south Atlantic has been estimated at 10 Sv northward (Schmitz, 1995), so the contribution from Agulhas eddies is a substantial component of the total inter-ocean exchange. The energy flux associated with these eddies is the order of 10^{17} J (Byrne et al. 1995: GEOSAT; Goni et al., 1997: T/P). Annual heat fluxes estimates vary between 3.8 x 10^{13} W (Grundlingh, 1995: T/P) and 4.5 x 10^{13} W (van Ballegooyen et al, 1994: GEOSAT). Annual salt fluxes are estimated at 78 x 10^{12} kg per year (van Ballegooyen et al, 1994).

Interannual changes in the behavior of these Agulhas eddies has also been investigated. Witter and Gordon (1999) used an empirical orthogonal function analysis of T/P data and found a transition in basin-scale sea level in the eastern South Atlantic, from higher sea level and enhanced gyre-scale circulation in 1993/1994, to lower sea level and sluggish circulation in 1996. The Agulhas eddy trajectory also changed; eddies dispersed over a broad region in 1993/1994, and remained in a narrow corridor in 1996 (Figure 13). This suggests that in different years, the injection of mass, heat and salt from Agulhas eddies may occur in different regions of the subtropical gyre, and be partially controlled by interannual variations in the large-scale circulation.

The mechanisms which cause ring-shedding in the Retroflection region have also been investigated using altimetry. One of these mechanisms is the Natal Pulse, an occasional large meander which forms in the Agulhas Current off the African coast. An analysis of GEOSAT, ERS-1 and T/P data has been used to monitor the Natal Pulses from Durban to the Agulhas Bank (van Leeuwin and de Ruijter, 1999), and after a time lag of around 100 days an Agulhas ring is shed. Rings also appear to be shed by a combination of Natal Pulses and the arrival of Rossby waves in the Agulhas Return Current. Possible triggering mechanisms for these Natal Pulses have been investigated from altimetry and cross-shelf velocity measurements (de Ruijter et al., 1999). The intermittent formation of these pulses is argued to be related to barotropic instabilities of the Agulhas Current. Occasionally, the correct conditions for these instability processes occur in the Natal Bight, particularly when the intensity of the incoming jet exceeds the average. The combined velocity structure and altimetry observations show that the increased sharpness is related to offshore anomalies, upstream and eastward of the Natal Bight.

3.4.3 AZORES CURRENT

The Azores Current is part of the North Atlantic subtropical gyre recirculation. West of the Mid-Atlantic Ridge, it is the extension of the southeast branch of the Gulf Stream. It crosses the ridge near the Oceanographer and Hayes fracture zones between 32°N and 37°N and then flows across the Canary Basin at about 34°N (Gould, 1985). In the eastern basin, it splits into three main southward branches. These branches, which vary seasonally and interannually, are found just east of the Mid-Atlantic Ridge, in the central basin near 23°W and near the coast of West Africa feeding the Canary Current, respectively (Stramma and Siedler, 1988; Klein and Siedler, 1989; Siedler and Onken, 1996). The mean transport of the Azores current is about 10 Sv in the eastern basin while it is about 30 Sv west of the ridge (Gould, 1985; Klein and Siedler, 1989). As observed by Pingree and Sinha (1998), however, the instantaneous transport in the Azores front in the eastern basin (near 32°W) can reach up to 30 Sv but the net transport is much smaller due to north and south recirculations. The Azores current is related to a thermohaline front with strong meanders which are embedded in a field of mesoscale eddies. It is poorly reproduced in most model simulations and its associated mesoscale variability largely underestimated. This makes the analysis of altimeter data particularly useful.

The Azores Current/Front mesoscale variability was first analyzed with GEOSAT data by Tokmakian and Challenor (1993) and Le Traon and De Mey (1994). They showed a tongue of high variability along 34°N associated with the Azores Front with rms SLA larger than 8 cm and EKE between 100 and 200 cm^2s^{-2}. Dominant space and time scales are greater than 300 km and 100 days (Le Traon and De Mey, 1994). Propagation velocities were westward and they also noted the wavelike structure of the mesoscale field which they attributed to Rossby waves generated near the eastern coast. Cippolini et al. (1997) also found westward propogation both in T/P and ATSR (SST) data with Rossby wave characteristics consistent with the theory of Killworth et al. (1997). An analysis of Reynolds stresses and eddy mean flow interaction (see section 4.6.1) from GEOSAT data by Le Traon and De Mey (1994) showed that the eddy field tends to accelerate the mean flow and its recirculations.

Hernandez et al. (1995) have analyzed T/P and ERS-1/2 data in the Azores front region east of the Mid-Atlantic ridge together with Semaphore in-situ measurements (Eymard et al., 1996; Tychensky et al., 1998). There is very good agreement between hydrographic and drifter measurements and altimetry (Figure 14). Both altimetry and in-situ data reveal the development of large Azores front meanders from July 1993 to November 1993. In situ data show that this development is due to the interaction of the Azores Front with Mediterranean lenses (Meddies) coming from the east (Tychensky et al., 1998). Meddies are often observed in the Canary basin and an early study with GEOSAT data (Stammer et al., 1991) showed that their surface signature could be detected by altimetry. In the Azores Frontal region just west of the Mid-Atlantic Ridge, a large cyclonic eddy associated with a sea level depression of 40 cm (a "Storm") was observed both with ERS-1/2 and hydrographic data (Pingree and Sinha, 1998). The in-situ data showed that the eddy signature penetrated down to 4000 m. The eddy was likely associated with a cyclonic meander of the Azores Front. Pingree and Sinha (1997) conclude that the observed westward propagating features in the Canary basin are cyclonic and anticyclonic eddies rather than Rossby waves. From their field data, the eddy separation was ~500 km and the eddies move at a mean westward speed of 3 km/day (in the eastern basin). Cromwell et al. (1996) have applied a synthetic geoid technique using ERS-1 and hydrographic data on the northern side of the Azores Front near 28°W, in order to analyze absolute velocities for the ERS-1 3-day repeat periods in spring 1992 and spring 1994. They found a persistent westward flow at 35°N flow with velocities of typically 25 cms^{-1}, which is probably related to recirculation north of the Azores Front. The Azores Front recirculation has also been observed from hydrographic data and is probably eddy-driven through rectification mechanisms (Alvès and Colin de Verdière, 1999).

Most of these results are confirmed by a recent analysis of more than 5 years of T/P, ERS-1 and ERS-2 data in the Canary basin (Hernandez and Le Traon, in preparation). Marked seasonal to interannual variations exist, both in the relative position of the current and the sea level variability. However, the seasonal behavior of the Azores Current and recirculation is less clear since the signal is dominated by meoscale variability and Rossby wave signals. In agreement with GEOSAT results (Le Traon and De Mey, 1994), altimeter data show the existence of Rossby wave signals with wavelengths larger than 500 km. They correspond to Azores Front meanders which propagate upstream (westward) (see Cushman-Roisin, 1994).

While altimetry has provided a much better visualization of the Azores front, the mechanisms responsible for its mesoscale variability are not yet completely understood. Baroclinic instability is probably one important mechanism but Rossby waves, Meddies, Mediterranenan outflow, wind forcing and bathymetry have also to be taken into account to explain the dynamics of the region.

3.4.4 ICELAND-FAEROE FRONT AND NORWEGIAN ATLANTIC CURRENT

The Iceland-Faeroe Front region, lying between Iceland and the Faraoe Islands, is a permanent but highly variable frontal region, separating the warm, salty North Atlantic water from the colder, fresher Arctic waters coming from the Greenland and Norwegian Seas. Although this is a dynamically important region, there are limited in-situ observations which do not resolve the large spatial and temporal variability of the front, and the frequent cloud cover limits satellite SST measurements. Altimetry provides a means for monitoring the position and strength of the front, which has cross-frontal slopes of 10-35 cm, current speeds of 25-50 cm/s, and meanders with radii of curvature of around 25 km (Robinson et al., 1989, Pistek and Johnson, 1992a). The 3-day repeat ERS-1 data shows that the variability in the frontal zone is on time scales of days (Tokmakian, 1994), although the time scales are longer away from the main jet. This suggests that significant frontal variability may be missed by longer repeat missions, and indicates that combinations of satellite missions (ERS-1 and T/P, for example) are necessary for better monitoring of surface frontal characteristics. Altimetric studies have also been used to monitor the occurrence of cold cross-frontal jets, related to the formation of cold eddies south of the frontal boundary (Scott and McDowall, 1990).

The transport of the Norwegian Atlantic Current has also been determined from a combination of altimetry and along-track and climatological CTD data by Pistek and Johnson (1992b). Mean transports of volume, heat, and salt were calculated as 2.9 Sv, 8 x 10^{11} Kcal/s, and 1 x 10^{8} kg/s. Samuel et al. (1994) used a different technique of combining GEOSAT residuals with a mean sea surface determined from an isopycnal ocean model, but found similar mean transports with a significant seasonal variation of 1.8 Sv, with maximum transports in winter (Feb-Mar) and minimum in summer (Jul-Aug).

3.5 SEMI-ENCLOSED SEAS

A major problem for studies of semi-enclosed seas is that the standard altimetric corrections for the tides and atmospheric pressure response have larger errors in the coastal or semi-enclosed seas than in the open ocean. These local tidal and inverse-barometer correction errors are often reduced by applying specific regional corrections.

3.5.1 MEDITERRANEAN SEA

The Mediterranean Sea circulation is characterized by a complex combination of mesoscale, sub-basin and basin scale signals (Millot, 1991; Robinson et al., 1991) (Figure 15). In the western basin, the Atlantic Water first flows eastward along the Spanish coast, veers south across the Alboran sea, and forms the unstable Algerian current whose instabilities generate intense mesoscale activity in the form of anticyclonic eddies and meanders (e.g. Millot, 1991). Other components of the western circulation are the cyclonic gyres in the Tyrrhenian sea and in the Balearic basin and the Liguro-Provençal current. The Atlantic Water then enters the eastern basin via the Sicily Straits, meanders in the Ionian basin and forms the Mid-Mediteranean jet. This current is described as being persistent and weak. On both sides of the jet, sub-basin gyres can be identified, such as the rather stable cyclonic Rhodes and Ionian gyres to the north, and the seemingly transient anticyclonic Mersa-Matruh and Shikmona gyres to the south (e.g. Robinson et al., 1991). The seasonal and interannual variability of the circulation is thought to be mainly due to wind stress forcing, although heat flux variations and changes in inflow/outflow at Gibraltar are also important mechanisms.

Satellite altimetry has provided an important contribution to analyzing the Mediterranean Sea circulation variability. Use of satellite altimetry in the Mediterranean Sea is particularly difficult, however, because the signal is small and the geometry (narrowness, straits, islands) of the basin is complex. The latter makes the orbit error reduction more difficult and means that the specific problems in coastal areas (radiometer correction, tides) need to be addressed. Larnicol et al. (1995) analysed T/P data in the entire Mediterranean Sea, and showed that the surface circulation was more complex in the eastern basin than in the western basin. The eastern basin is composed of sub-basin-scale gyres, such as the so-called Mersa-Matruh and Shikmona gyres, which do not have an obvious recurrence period. They observed a winter intensification of the large-scale cyclonic circulation in the western and in the Ionian basins. Several mesoscale structures, such as the Alboran gyres (see also Vazquez et al., 1996) and the Ierepetra gyre, also showed a clear seasonal cycle with a maximum in summer. The good qualitative and quantitative agreement of the results with in-situ data from the Mediterranean illustrated the improved accuracy of T/P over its predecessors (see, for example, Fuda et al., 1998). Iudicone et al. (1998) also analyzed two years of T/P data and AVHRR data and confirmed the conclusions of Larnicol et al. (1995). They also suggested that the variations in mesoscale variability intensity between 1993 and 1994 were related to wind forced inter-annual variations. Algerian eddies were analyzed in several studies (Ayoub, 1997; Ayoub et al., 1998; Iudicone et al., 1998; Bouzinac et al., 1998). These studies showed the development of Algerian eddies, their eastward propagation and their detachment from the coast and propagation to the northwest. Bouzinac et al. (1998) deduced from a complex EOF analysis of combined T/P and ERS-1 maps (from Ayoub et al., 1998) that Algerian eddies may detach from the coast at two locations (4°E and 8°E). Nardelli et al. (1999) have performed a detailed analysis of the Sicily channel using T/P and ERS-2 along-track data, AVHRR data and in-situ measurements. They found a good match between altimeter and hydrographic data.

Ayoub (1997) and Ayoub et al. (1998) extended the Larnicol et al. (1995) work using ERS-1 and ERS-2 data and T/P over 5 years. SLA maps were systematically calculated from T/P and ERS-1/2 separately and T/P and ERS-1/2 combined. They showed that the improved sampling given by the combination of T/P and ERS-1/2 is vital for monitoring mesoscale signals in the Mediterranean. Over 5 years, the 3-month variations of combined SLA maps show the main characteristics of the Mediterranean circulation (Alboran gyres; Algerian eddies; Ionian, Ierepetra, Mersa-Matruh and Shikmona gyres, etc.) and the seasonal variations in these features (strengthening of cyclonic circulation in winter, strengthening of anticyclonic Alboran and Ierepetra gyres in summer and fall) (Figure 16). T/P and ERS-1/2 reveal some of the particularly strong signals very well : Alboran gyres east of Gibraltar and Ierepetra gyre south-east of Crete. The strong seasonal signal from the Ierepetra gyre (its diameter is about 100-200 km which is much larger than the local internal Rossby radius, hence the gyre appellation) is probably linked to direct forcing by strong Etesian winds, which interact with the Cretan topography (orographic effect). There is also a large interannual variability, in particular in the Levantine basin. The interannual variability in the Levantine might be related to a change in the Etesian winds and the switching from a state with a well developed Ierepetra gyre to a state with a large anticyclonic system in the central Levantine basin (with the development of the Mersa-Matruh and Shikmona anticyclonic gyres) (Ayoub, 1997).

These results provide, for the first time, a global view of the intra-seasonal, seasonal and interannual variations of the circulation in the Mediterranean sea. This is crucial for model validation and for a better understanding of the Mediterranean sea circulation (e.g. Pinardi and Navarra, 1993).

3.5.2 YELLOW AND EAST CHINA SEAS

The Yellow and East China Seas form a vast expanse of continental shelf bounded by the Chinese mainland, and limited offshore by the Kuroshio with Kyushu to the north, and Taiwan to the south. The circulation in this region is strongly determined by the proximity to the Kuroshio, and by the monsoon winds which are northerly in winter and southeasterly in summer. Altimetry has been used to describe the seasonal variations in surface circulation by Yanagi et al. (1997), who apply a regional tidal model to correct the T/P residuals. For the Yellow Sea, they find a cyclonic circulation develops in summer, with an anticyclonic downwelling circulation in winter. In the East China Sea, the wintertime circulation is cyclonic. The sea surface response to regional wind forcing in the Yellow and East China Seas has also been studied at intra-seasonal time-scales by Jacobs (1998). The residual tidal variations are removed by a least-squares fit of the 8 main tidal frequencies at each point along the ground-track. Although the instantaneous wind stress sets up local ageostrophic Ekman transports, this explains only a small fraction of the sea level response. Most of the sea level variability is remotely forced, and an extended EOF analysis of the wind field reveals the origins. The principal wind mode is significantly related to variations and intrusions of the Kuroshio along the shelf break. Another mode is due to northerly wind bursts in the winter, which produces large drops in sea surface height in the Bohai Bay and northern Yellow Sea regions. The third and fourth modes are associated with typhoon passages in the autumn, which are related to positive SSH anomalies moving from the northern Yellow Sea to the Chinese coast.

3.5.3 GULF OF MEXICO

In the Gulf of Mexico, GEOSAT altimetry has been used to investigate the shedding of eddies from the Loop Current. Johnson et al. (1992) tracked two major rings shed from the Loop Current, which drifted southwestward across the Gulf; they also monitored the build-up of the Loop Current during 1985/1986. Jacobs and Leben (1990) found eddies shed from the Loop Current with period of approximately 10.5 months, consistent with results from numerical models and satellite infra-red data. More recently, various US institutions have used a combination of T/P and ERS-1/ERS-2 data in near-real time, to monitor Loop Current eddies and other circulation variability in the Gulf of Mexico. As an example, Biggs et al. (1996) describe the cleavage of an anti-cyclonic Loop Current eddy, Eddy Triton, by a deep water cyclonic circulation. The altimeter data allowed them to track both pieces: the major part drifted northwest to the "eddy graveyard" in the northwest corner of the Gulf; the minor part drifted southwest to the continental margin, turned north and later coalesced with the major fragment, and was ultimately entrained into another eddy. This level of detailed surveillance is important in this region for oil exploration, for fisheries and tracking marine mammals.

3.5.4 ARABIAN SEA AND BAY OF BENGAL

The ocean variability in the semi-enclosed seas of the northern Indian Ocean is strongly controlled by the monsoons. The western part of each basin is highly energetic, with root mean square values of 15-17 cm in the Somali Current, and 13-15 cm in the western part of the Bay of Bengal (Jensen et al., 1997). Altimetry has been used to monitor the three dominant gyres off the coast of Somalia, i.e., the Great Whirl, the Southern Gyre and the Socotra Eddy (Subrahmanyam et al., 1997). The strength of these eddies varies significantly with the strength of the monsoon. Propagating signals with annual and semi-annual periods have also been examined from altimetry using complex principal component analyses. In the eastern Arabian Sea, Rossby waves which radiate from the west coast of India are associated with the passage of Kelvin waves along the coastline (Subrahmanyam et al., 1997). Altimetry has also been used to study the Arabian Sea Laccadive High – an anticyclonic circulating feature that forms off the southwest coast of India during the northeast monsoon. Bruce et al. (1998) find that the Laccadive High consists of multiple eddies, and is forced by local and remote seasonal monsoon forcing, as well as being influenced by an intra-seasonal signal that originates in the Bay of Bengal. Finally, Subrahmanyam et al. (1995) noted the existence of the poleward East Indian Coastal Current (EICC) along the east coast of India, using T/P data. Although the current is not detected in climatological hydrographic data, the T/P analysis suggests the current persists for the entire year, though it changes direction from north (Jan - Aug) to south (Sep - Dec), and is dominated by eddies in Jan-Mar. When fully developed in April, the current extends some 800 km, and the speed of the residual current is around 20-30 cm/s.

4. MESOSCALE EDDIES

The first part of the chapter was focused on the description of the main ocean currents and of their associated mesoscale variability (meandering, eddy shedding) as observed by satellite altimetry. The second part deals with a global statistical description of the mesoscale variability. The global space-time sampling of satellite altimetry is very well suited for statistically describing mesoscale phenomena. Such a description can reveal geographical and temporal variations in mesoscale eddy statistics. This information can then be used to interpret the eddy structures, identify sources of eddy energy, and analyze energy transfer from these sources. This can help us to understand eddy dynamics, particularly the mechanisms which generate and dissipate eddies. Global statistical descriptions are also a means of testing and validating eddy-resolving models. They are also necessary for inverse modeling and altimeter data assimilation studies.

In the first section, we start with a description of the rms sea level variability, cross-track geostrophic velocity variance and EKE (section 4.1). We then review in section 4.2 the analysis of seasonal to interannual variations in eddy intensity. Section 4.3 summarizes the findings on space and time scales of mesoscale variability and their relationship with the internal Rossby radius while section 4.4 deals with SLA frequency/wavenumber spectra and their relationship with quasi-geostrophic turbulence theory. Section 4.5 focusses on the comparison of these eddy statistics with eddy-resolving model simulations. Eddy dynamics (eddy-mean flow interaction and eddy transport) are finally discussed in section 4.6.

4.1 GLOBAL STATISTICAL DESCRIPTION

The simplest description of mesoscale variability is that obtained by the global mapping of the rms of SLA. While the first estimates derived from SEASAT revealed only one tenth of the total mesoscale energy (because of the short - 3-month - duration of the repeat mission), estimates obtained from GEOSAT were quite accurate (e.g. Koblinsky, 1988; Sandwell and Zhang, 1989; Shum et al., 1990). T/P and ERS-1/2 values are comparable to those from GEOSAT with some important differences. For example, T/P results contain more of the large-scale signal (which was generally removed from GEOSAT by the orbit error adjustment) and also the longer time scale signal (depending on the duration of the data set).

T/P also has a reduced noise level and improved altimetric corrections. The satellites also have different repeat period and do not alias or observe the same frequencies (see also section 2.1). Using the 3-day repetitive ERS-1 data over a 3-month period, Minster and Gennero (1995) have estimated that between 5 and 10% of the energy in the Gulf Stream and Kuroshio regions was at periods shorter than 20 days and 34 days respectively and are thus aliased in T/P and GEOSAT sampling. Wunsch and Stammer (1995) have estimated the contribution of mesoscale variability to the total energy. In low-eddy energy regions, the contribution of the large scale and long time scale signals (e.g. steric signals) can be as high as half of the total energy.

Global statistics from T/P can be found in Wunsch and Stammer (1995) while ERS-1/2 results can be found in Le Traon and Ogor (1998). Figure 17 shows the rms SLA derived from almost 5 years of T/P and ERS-1/2 combined maps (Ducet and Le Traon, 1999). The rms sea level variability in western boundary currents (Gulf Stream, Kuroshio, Brazil/Malvinas Confluence, Agulhas) and the Antarctic Circumpolar Current (ACC) is higher than 30 cm rms and can reach up to 50 cm rms. The maximum eddy variability is actually observed in the Agulhas region, followed by the Gulf Stream, Kuroshio and Confluence regions and finally by the ACC. In very low eddy energy regions, i.e. in regions with no or weak mean currents, the mesoscale signal after instrumental noise filtering is typically 3 cm rms and may be influenced by other small scale signals (e.g. internal waves). ERS-1/2 results are in excellent agreement with T/P results when orbit error has been corrected (Le Traon and Ogor, 1998). In high latitudes areas (which are not sampled by T/P), the signal is generally weak (5-8 cm rms).

As the calculation of derivative acts as a high pass filter, cross-track geostrophic velocities are much less sensitive to large scale signals and are thus more representative of the mesoscale signals. The computation of velocity is more sensitive, however, to measurement noise and need a careful filtering of sea level data (see section 2.3). The map of cross-track geostrophic variance is equivalent to EKE [EKE = ½ ($\langle u'^2\rangle + \langle v'^2\rangle$)] if the field is isotropic. An analysis of global SLA maps obtained from the combination of T/P and ERS-1/2 shows that, to a first order, this approximation is valid outside the tropics (Ducet and Le Traon, 1999). Shum et al. (1990) produced a global map of cross-track geostrophic velocity variance using GEOSAT data. More than 65% of the ocean is shown to have EKE values less than 300 cm^2s^{-2}. The maximum EKE exceeds 2000 cm^2s^{-2} for most of the western boundary currents but reaches only 500 cm^2s^{-2} in the ACC. Sandwell and Zhang (1989) have produced a global map of the variance of dynamic topography slope, which translates directly into variance of geostrophic velocity outside equatorial areas. They established a correlation between the intensity of the variability and the ocean depth: areas of highest variability are in deep basins (> 4 km). In the ACC, there is a close relationship between the geographical distribution of mesoscale variability and the strength of the mean circulation. This is not surprising since these narrow currents are likely to be baroclinically and barotropically unstable. There is also strong topographic control of the eddy field (and the mean field) consistent with numerical simulations (e.g. Treguier and Mc Williams, 1990). Most of these areas are also characterized by intense mean currents (western boundary currents and ACC) which are the main source of eddy energy through instability.

Ducet and Le Traon (1999) provides a more recent estimation of the EKE based on the combination of five years of T/P and ERS-1/2. They show that the maximum levels of EKE can reach values of up to 4500 cm^2/s^2 in the western boundary currents. Because of the higher resolution provided by the combination of T/P and ERS-1/2 and a low background noise variance (about 15 cm^2/s^2), their estimation reveals many details which cannot be accessed with along-track data only. Stammer (1997) analyzed the correlation between the T/P derived EKE (assuming isotropy) and the mean kinetic energy (MKE) (0/1000 dbar geostrophic current) as derived from Levitus historical data. As expected, there is good correspondence between T/P EKE and MKE maxima as the currents are the main sources of eddies. There are a few noteworthy exceptions in the Agulhas retroflection, the East and West Australian currents and the Brazil/Malvinas Confluence regions. In the ACC, rather high mesoscale variability areas are also found in regions of abrupt changes of bottom topography which do not appear to be associated with strong mean currents. This may imply that in these areas, baroclinic instability (due to vertical shear) is not likely to occur and that other mechanisms are to be sought (e.g. barotropic instability, mean current/bottom topography interaction). Finally, it should be noted that altimeter estimates generally agree well with in-situ measurements (in particular drifting buoys). Differences can almost always be explained by differences in sampling and/or differences in measurement content (ageostrophic signals, differences in mean signal removal – time mean or space/time mean for drifting buoy) (Le Traon et al., 1990; Hernandez et al., 1995; Stammer, 1997).

4.2 SEASONAL VARIATIONS OF MESOSCALE VARIABILITY INTENSITY

Several mechanisms could explain seasonal or interannual variations in mesoscale variability intensity. In regions of intense currents (e.g. western boundary currents), where baroclinic/barotropic instabilities of the mean currents are the main forcing mechanisms, a seasonal/interannual intensification of the current may induce a growth of the instability rates. A change in the current position could also imply a similar shift of the eddy field. These changes could be indirectly related to wind forcing which is the main forcing of the ocean circulation. Strass et al. (1992) also suggest that the development of the seasonal pycnocline during the heating season can generate strong potential vorticity gradients favoring baroclinic instability. In such a case, heat flux forcing would be indirectly responsible of an enhanced mesoscale variability during the summer season. In regions of low eddy energy, fluctuating wind forcing could be a possible direct source of mesoscale variability (e.g. Frankignoul and Müller, 1979; Willebrand et al., 1980; Müller and Frankignoul, 1981). At high and intra-seasonal frequencies, the wind forcing generates predominantly barotropic fluctuations in the form of forced waves and Rossby waves and has been invoked for explaining the coherence between wind stress curl fluctuations and deep ocean mooring sites in the North Atlantic and North Pacific (Willebrand et al., 1980; Samelson, 1990; Luther, 1990) and the increase in deep eddy energy in winter (Dickson et al., 1992; Koblinsky et al., 1989). As emphasized by Lippert and Müller (1995) and Müller (1997) and in qualitative agreement with observations, local and non-local coherence between ocean response and wind forcing is expected because the ocean response integrates forcing from different places. Recent model simulations show also high frequency large scale barotropic signals in high latitude regions consistent with T/P observations (Fu and Davidson, 1995). Except in a few localized regions, the barotropic signals are small, however, and will not induce an important change in the surface mesoscale signals as seen by an altimeter which are dominated by baroclinic signals (Wunsch, 1997). Wind fluctuations also generate baroclinic signals (e.g. Frankignoul and Müller, 1979; Müller and Frankignoul, 1981). Wind is thus one of the forcing mechanisms of the low-frequency (mainly seasonal to interannual) baroclinic Rossby waves which are ubiquitous in altimetric data (see Chapter 2). In the mesoscale band, the ocean response to stochastic wind fluctuations can also be baroclinic when non-linearities and bottom topography are taken into account (Treguier and Hua, 1987; Treguier and Hua, 1988). Wind forcing is thus a possible candidate for direct surface mesoscale signal generation.

Several attempts have been made to analyze seasonal variations of mesoscale intensity with GEOSAT data (Fu et al., 1988; Zlotnicki et al., 1989). Zlotnicki et al. (1989) showed that some regions (North-East Atlantic and North-East Pacific) had higher mesoscale energy during winter, when wind intensity is stronger. This suggests that wind forcing may be a source of eddy energy there. Small seasonal variations were also found in the western boundary currents where a maximum of eddy energy was generally found during summer and fall. Stammer and Böning (1996) reexamined results from GEOSAT and concluded that the maximum of mesoscale energy was in fall rather than in winter thus before the maximum of wind stress curl energy supporting the thermodynamic mechanism proposed by Strass et al. (1992). Most of these results were based on only two years of GEOSAT data only and should be interpreted with caution. In addition, to analyze the mesoscale variability variations, it would be preferable to analyze the EKE rather than the rms sea level variability as the latter includes more large-scale signals.

Garnier and Schopp (1999) have analyzed the effect of wind on the mesoscale activity in the North Atlantic using two years of T/P data and ERS-1 scatterometer winds. Their study suggests that the wind plays an important role in the time evolution of the mesoscale variability. The mesoscale variability associated with the Gulf Stream and the North Atlantic Drift increases when the wind induces an intensified eastward Sverdrupian velocity. In this case the baroclinic instability mechanism could be responsible for the mesoscale variability intensification. Seasonal and interannual variations in eddy energy were analyzed by Stammer and Wunsch (1999) using 4 years of T/P data. Over most of the subtropical region and along major mean fronts, the zonal variations are weak and often negligible. Western boundary current extensions show an annual cycle with a maximum occurring in late summer/early fall. A pronounced annual cycle in eddy energy was also apparent in the eastern North Pacific, and the northern and eastern North Atlantic. In these regions, Stammer and Wunsch (1999) found a significant correlation between EKE and wind stress forcing at seasonal and interannual scales (Figure 18). Note also that seasonal modulations of the eddy energy are also clearly observed in other regions (e.g. California Current, Canary Current, Mediterranean Sea) (see previous section). In the California Current, Kelly et al. (1998) found, in particular, maximum EKE values along fronts which coincided with regions of maximum Ekman pumping (due to wind stress curl) suggesting an indirect relationship between the eddies, presumably formed by baroclinic instability, and winds.

Qiu (1999) has made a detailed analysis of the North Pacific Subtropical Countercurrent (STCC) using more than 5 years of T/P data. The mean EKE of the STCC is about 330 cm^2s^{-2} and reaches half the energy of the Kuroshio Extension. The unique characteristics of this current system is that it has a very clear EKE seasonal cycle with a maximum of EKE in April/May and a minimum in December /January (Figure 19). Using a 2 ½ layer reduced-gravity model, representing the vertically-sheared STCC/North Equatorial Current (NEC) system, Qiu (1999) convincingly shows that the seasonal modulation of the STCC eddy field is related to seasonal variations in the intensity of baroclinic instability. The seasonal cooling/heating of the upper thermocline modifies the vertical velocity shear of the STCC/NEC and the density difference between the STCC and NEC layers. As a result, the spring-time condition is considerably more favorable for baroclinic instability than the fall-time condition. The theoretically predicted e-folding time scale of the instability is 60 days and matches the time lag between the EKE maximum and the maximum shear of the STCC/NEC (Figure 20).

4.3 SPACE AND TIME SCALES OF MESOSCALE VARIABILITY

Space and time scales can be derived from altimeter covariance functions which are mathematically equivalent to frequency and wavenumber spectra (through a Fourier transform). We will focus here on the main space and time scales as defined by the zero crossing of the correlation function while the next section on frequency/wavenumber spectra will allow a more detailed discussion of the frequency and wavenumber content of altimeter signals.

Results obtained with GEOSAT in the Atlantic (Le Traon et al., 1990; Le Traon, 1991, Stammer and Böning, 1992) showed a clear latitudinal variation of the space scales, decreasing towards the poles. Although a relationship with internal Rossby radius (IRR) was observed, the spatial scales seen by altimetry varied by a factor of 2 from the equator to the pole, while the Rossby radius varied by more than a factor 4 (Le Traon, 1993). Time scales were shorter in high variability areas while longer time scales were observed above the Mid-Atlantic Ridge and in low eddy energy regions. They were generally not proportional to space scales (Le Traon, 1991). Bottom topography appeared to play an important role in the temporal coherence of mesoscale structures. Stammer (1997) recently repeated the calculation using T/P data. Since T/P SLA data also contain the large scale and low frequency signals (which were in large part removed during the GEOSAT orbit error removal), T/P space and time scales are much larger than for GEOSAT. If the large scale signal is removed, however, or if the analysis is performed on geostrophic velocities, the scales are much more representative of the mesoscale variability. Time and space scales are then consistent with GEOSAT values. Globally, they vary between 5 and 20 days and between 60 and 200 km respectively. A relationship with the IRR similar to GEOSAT results is noted by Stammer (1997) (Figures 21 and 22). Spatial scales in the high latitudes seen by T/P (or GEOSAT) are too large, however, compared to the IRR. Because of low stratification and larger Coriolis parameter, the ocean is expected to be more barotropic and baroclinic instability may not be the main eddy generation mechanism. A relationship with IRR would be valid only if one could separate barotropic and baroclinic signals in altimeter SLA measurements. While baroclinic signals dominate SLA signals (Wunsch, 1997), they may not be negligible at high latitudes. This was confirmed by an analysis of CME model simulations (Stammer, 1997).

The relationship of spatial scales with IRR certainly points to the baroclinic instability as the main generation mechanism for mesoscale variability. Indeed, in baroclinic instability, the most unstable perturbation has the IRR scale (e.g. Gill, 1982). However, the dependence is not expected to follow a simple law. It depends, in particular, on the further evolution of ocean eddies through turbulent cascade which may not exist everywhere (see discussion below). Different relationships with IRR can also be observed in case of other forcing mechanisms. For wind forcing, the energy input is on larger wavelengths than the IRR and it is the ratio of the largest forced wavelength to the IRR that determines the space scales of the ocean response (Treguier and Hua, 1987).

The relationship with IRR is more difficult to estimate from in-situ data because of the lack of adequate space/time sampling. Paillet (1999) in an analysis of scales of vortices from XBT data reports no significant relationship in the eastern North Atlantic. He argues that in the baroclinic instability mechanism the most unstable perturbation is not necessarily the one that is excited and that grows. Vortices are also long lived structures and may be observed far away from their initial formation region. Finally, vortices are known to coalesce with neighbors (quasigeostrophic turbulence – see below) but the turbulence may not be well developed in quiet regions because it requires a large number of eddies. There, eddies would keep their initial size while in regions where turbulence is active they would grow. Paillet (1999) points to this as a possible explanation for rather large eddies in the North Atlantic

Current compared to regions more to the South. On the contrary, Krauss et al. (1987) report a good relationship between eddy spatial scales and the IRR from drifting buoy data.

4.4 FREQUENCY/WAVENUMBER SPECTRAL ANALYSIS

4.4.1 FREQUENCY SPECTRA

SLA frequency spectra have been computed by Stammer (1997) in different areas of the world ocean with T/P data. All spectra show a peak at the annual frequency (mainly related to the dilatation/contraction of surface waters in response to heat flux forcing and to a less extent to wind forcing). Slopes of the spectra generally range between –1 and – 2, the steeper slopes being found in high mesoscale variability areas. Slopes are slightly lower than slopes derived from subsurface currentmeter data. Stammer (1997) suggests that this may be due to near surface signals not seen in subsurface currentmeter data.

4.4.2 WAVENUMBER SPECTRA

4.4.2.1 OBSERVATIONS

While frequency spectra were already well known from in-situ (currentmeter) measurements (although not with the same global coverage but with a much better time sampling), the wavenumber spectra are a unique contribution of altimetry. Following the pioneering work of Fu (1983) with SEASAT, a detailed analysis was conducted by Le Traon et al. (1990) in the North Atlantic with GEOSAT data (Figure 23). They found spectral slopes in the mesoscale band of -4 in the western part of the basin and between -2 and -3 in the other areas between 50 - 200 km and 200 - 600 km. After the break in the slope, which showed a decrease according to latitude, the spectra remain red in the eastern part of the basin. An increased energy at smaller scales (below 100 km) is also observed there. Forbes et al. (1993) did a systematic study of the wavenumber spectra from GEOSAT data in the South Atlantic and found similar shapes of the spectra. A different view of the same spectra was, however, proposed by Stammer and Böning (1992). They suggested that GEOSAT spectra were not significantly different from a k^{-5} law and that noise was probably responsible for the weaker slopes in the eastern basin. This led to some controversy (Le Traon, 1993; Stammer and Böning, 1993). Results obtained with the more precise T/P data (Stammer, 1997) are actually in excellent agreement with GEOSAT results and thus confirmed that noise was not responsible for the observed shape of the GEOSAT spectra (and that GEOSAT was particularly well suited for mesoscale studies). Surprisingly, slopes as weak as k^{-1} can also be observed in very high latitudes from T/P and/or ERS-1/2 data. More recently, Paillet (1999) used an ensemble of 102 high resolution XBT tracks in the North-East Atlantic and showed that the dynamic height wavenumber spectrum was following a k^{-3} law consistent with GEOSAT and T/P results.

4.4.2.2 INTERPRETATION OF WAVENUMBER SPECTRA

The interpretation of wavenumber spectra is closely related to theories of geostrophic turbulence. Geostrophic turbulence is a complex and evolving topic and the reader is referred to Kraichnan (1967), Charney (1971), Rhines (1979), Colin de Verdière (1982), or McWilliams (1989) for a much more detailed review. Geostrophic turbulence is close to 2D turbulence and is characterized by a red cascade of energy towards larger scales while there is an inverse cascade (i.e. towards smaller scales) for the enstrophy. In pure 2D homogeneous turbulence, the cascade would continue up to the scale of the basin which is obviously not occurring. By taking into account the presence of a background planetary vorticity gradient (β-plane turbulence), however, a competition between Rossby waves (dominant at large scales) and turbulence (dominant at small scales) occurs. The cascade of energy is severely reduced at large scales by the dispersion of Rossby waves which prevents the merging of eddies. Rhines defines the β-arrest scale $L_\beta = (\beta/2U)^{-1/2}$ which is the scale for which the β effect becomes more important than non-linearities. At scales larger than L_β, the turbulence is strongly reduced and a linear regime consisting of Rossby waves is observed. While the turbulent regime is characterized by an isotropy, the more linear regime becomes anisotropic (owing to the dispersion relation of Rossby waves) and at the end state of the cascade, flow is mainly zonal ($u>v$). Stratified quasigeostrophic turbulence theories yield a similar cascade in the horizontal (Hua and Haidvogel, 1986; McWilliams, 1989). In addition, there is also a red cascade in the vertical scale. In a two-layer system, eddies in the two layers start to interact when the horizontal scale is close to the internal Rossby radius and to lock in the vertical leading to a barotropic signal. Then the energy cascade proceeds as in the 2D case. More generally, the turbulent cascade in a rotating stratified fluid leads to a 3D isotropy once the vertical coordinate has been re-scaled by N/f and for wavenumbers higher than the wavenumber at which the energy is input (McWilliams, 1989). Those are highly idealized views of the interaction of eddies in the ocean. There are many obstacles to the development of the energy cascade either in the horizontal or vertical scales (coastal boundaries, emergence of localized vortices, bottom topography) (McWilliams, 1989).

Through dimensional arguments, Kraichnan (1967) predicted for 2D turbulence a $k^{-5/3}$ energy spectrum at scales larger than the energy input scale while the enstrophy cascade to smaller scales leads to a k^{-3} energy spectrum. In extending these ideas to 3D quasigeostrophic turbulence, Charney (1971) also obtained a k^{-3} energy spectrum which has been confirmed in numerical simulations (e.g. Herring, 1980; Hua and Haidvogel, 1987). A k^{-3} energy spectrum would yield a k^{-5} one-dimensional (along-track) sea surface height spectrum assuming isotropy (Fu, 1983b; Le Traon et al., 1990). As noted by Le Traon et al. (1990) and Stammer (1997), altimeter wavenumber spectra in high eddy energy regions compare qualitatively with quasigeostrophic turbulence models although the observed slopes are only in k^{-4} rather than the expected k^{-5} law. In these regions, eddies are mainly generated through baroclinic instability and the dynamics are highly non-linear. In low eddy energy regions, where the altimeter slopes are significantly lower (k^{-3}, k^{-2} or even k^{-1}), the turbulence may not be as active and the ocean may be in a more linear regime. The ubiquitous observation of Rossby waves in the subtropical oceans (see Chapter 2) favors this interpretation of a more linear regime outside major ocean currents.

The shape of the spectra can also be different if there is significant wind forcing. Treguier and Hua (1987) in simulations of quasigeostrophic turbulence forced by wind found spectra more in agreement with altimeter observations in low eddy energy regions. In particular, they found that the spectra remain red for wavelengths larger than at the break of the slope. Altimetric spectral slopes may also be weaker than expected because of non-geostrophic effects related to the mesoscale variability of the mixed layer (Klein and Hua, 1988; Le Traon et al., 1990). Glazman et al. (1996) and Glazman and Cheng (1998) thus suggest that internal waves may affect the high wavenumber part of the altimeter spectra. Note that internal tide signals were clearly detected in altimeter data in the tropics (e.g. Ray and Mitchum, 1996) and can also be clearly seen as peaks in the T/P wavenumber and frequency/wavenumber spectra. Filtering of those signals may increase the spectral slopes in low eddy energy regions for intermediate wavelengths.

4.4.3 FREQUENCY/WAVENUMBER SPECTRA

Frequency/wavenumber spectra allow us to associate space and time scales of mesoscale variability as well as the propagation characteristics. Le Traon (1991) calculated frequency-wavenumber spectra in the Atlantic from GEOSAT. The dominant wavelengths of around 200 to 600 km (depending on latitude) are associated with long periods (> 150 days) in the eastern part of the basin, while near the Gulf Stream significant energy is also found at shorter periods. In the Gulf Stream area, propagation velocities can be either westward or eastward for short periods (< 80 days). At longer periods and in the eastern North Atlantic, they are mainly westward. Seasonal signals associated with westward propagation are also observed in these frequency-wavenumber spectra (Le Traon, 1991). These results confirm the interpretation of wavenumber spectra by Le Traon et al. (1990) who suspected a change in dynamic regime after the break in the wavenumber spectrum slope. Indeed, pseudo-dispersion relations deduced from GEOSAT data point to two distinct dynamic regimes, as in numerical models: a turbulent regime for smaller scales (<300 to 500 km), where there is proportionality between space and time scales, and an apparently more linear regime after the spectral peak in wavenumber where an inverse dispersion relation is found in the eastern part of the basin. This latter feature is in agreement with quasigeostrophic models forced by fluctuating winds (Treguier and Hua, 1987). Several investigators have used these analyses to characterize the mesoscale variability in regional areas (e.g. Le Traon and Minster, 1993; Provost and Le Traon, 1993; Morrow and Birol, 1999).

4.5 COMPARISON WITH EDDY RESOLVING MODELS

The comparison of statistical descriptions such as spectra and EKE with results from eddy resolving models is useful for model validation and for a better understanding of eddy dynamics. It should also help us to better understand the observed frequency/wavenumber of mesoscale variability and its relation with the forcing. Altimeter data have been compared with model data in numerous studies (e.g. Stammer and Böning, 1992; Tréguier, 1992; Hughes, 1995; Fu and Smith, 1996; Stammer et al., 1996, The Dynamo group, 1997; McClean et al., 1997). Comparisons have been between the model and altimetry rms sea level variability and EKE, but a few studies include higher order statistics (space and time scales, wavenumber spectra, Reynolds stresses and eddy/mean flow interaction). These studies show that eddy resolving models still largely underestimate the mesoscale variability. Recent numerical simulations in the North Atlantic with a 0.1° resolution by Smith et al. (1999) show remarkable improvements over previous lower resolution experiments for both the time mean circulation and mesoscale variability. With a 0.1° resolution, the rms sea level variability compares favorably well with T/P observations over most of the domain, including the Gulf Stream region, North Atlantic Current to the east of the Grand Banks, and the Azores

Current. Such a good degree of qualitative agreement should allow a much more detailed comparison with altimetry, including higher order statistics.

4.6 EDDY DYNAMICS

4.6.1 EDDY-MEAN FLOW INTERACTIONS

Eddy-mean flow interactions include the generation of eddies through barotropic and baroclinic instabilities of the mean flow, as well as the convergence of eddy momentum fluxes (Reynolds stresses) to accelerate the mean current. Before the advent of altimetry, these eddy-mean flow interactions had mostly been investigated using analytical or numerical models. Some in-situ studies based on current meter moorings at particular locations had revealed the complexities of the local eddy-mean field. However, there are very few locations with in-situ records longer than 2 years, necessary to derive stable eddy statistics (see Wunsch, 1997, for a summary). Even though we cannot derive an accurate mean current from altimetry, the eddy-mean interactions can still be estimated using a mean derived from hydrography, or from numerical simulations.

In terms of an energy budget, the eddy-mean flow interactions can be separated into 4 components:

1. The transfer from the mean kinetic energy to mean potential energy,
2. The transfer of mean potential energy to eddy potential energy,
3. The transfer of eddy potential energy to EKE,
4. The transfer of EKE to mean kinetic energy (MKE).

The first three components require details of the vertical density structure to define the mean and eddy potential energy fields, which are clearly not accessible from altimetry. Only the final component includes the purely kinematic terms. This barotropic energy conversion of EKE to MKE can be calculated from the horizontal momentum equations (see Wilkin and Morrow, 1994, for details):

$$\overline{u'u'}\frac{\partial \bar{u}}{\partial x}+\overline{u'v'}\frac{\partial \bar{v}}{\partial x}+\overline{u'v'}\frac{\partial \bar{u}}{\partial y}+\overline{v'v'}\frac{\partial \bar{v}}{\partial y} \qquad (10)$$

where the overbar denotes a time mean, and the prime denotes the departure from this mean. When this term is positive, EKE is converted to MKE at this location via the Reynolds stresses doing work on the mean shear so as to accelerate the mean flow. There is also a contribution from the vertical Reynolds stresses $(\overline{u'w'}, \overline{v'w'})$ which are neglected since the vertical velocities are assumed weak near the sea surface.

A similar calculation can be made to derive the eddy momentum flux divergence from the horizontal momentum equations. Again, neglecting the contribution from vertical velocity perturbations, the eddy momentum flux divergence is given by:

$$-\frac{\bar{u}}{|\bar{u}|}\left(\frac{\partial}{\partial x}\overline{u'u'}+\frac{\partial}{\partial y}\overline{u'v'}\right)-\frac{\bar{v}}{|\bar{u}|}\left(\frac{\partial}{\partial x}\overline{u'v'}+\frac{\partial}{\partial y}\overline{v'v'}\right) \tag{11}$$

Here $(\bar{u},\bar{v})/|\bar{u}|$ denotes the unit vector parallel to the co-ordinates of the mean flow $(\bar{u},\bar{v})$, so the first term represents a contribution to the zonal momentum balance, the second contributes to the meridional momentum balance.

Altimetry has provided the first global measurement of the surface eddy Reynolds stress terms, ($\overline{u'u'}$, $\overline{v'v'}$, $\overline{u'v'}$) which has been useful for improving our understanding of ocean dynamics, as well as validating estimates from numerical models. Note that energy and momentum budgets also rely on vertically averaged quantities over the water column, so estimates of the full barotropic conversion terms require a relationship for the vertical velocity structure, as well as an estimate of the mean flow.

One of the first studies to investigate the spatial distribution of eddy momentum fluxes from altimetry was made by Tai and White (1990) in the Kuroshio region. They created 0.5 degree latitude/longitude maps of GEOSAT ascending track sea level anomalies for every 17 day repeat cycle, using a spatial decorrelation radius of 200 km. Velocity anomalies are calculated from the mapped SLA assuming geostrophy, and the eddy flux of zonal momentum is estimated from the $\overline{u'v'}$ component. Their maps show positive eddy momentum flux values south of the jet, and negative values to the north, implying a convergence of eddy momentum flux which tends to accelerate the eastward mean jet. The north-south slope in eddy momentum flux $\partial/\partial y\left(\overline{u'v'}\right)$ shows convergence and eastward acceleration between 35°and 37°N (the jet axis), and divergence or westward acceleration farther north from 37.5° to 38.5°N and south from 30° to 33°S. This pattern is consistent with theories of non-linear baroclinic instabilities occurring near the jet axis which favor meander growth, and westward Rossby wave propagation dominating to the north and south of the mean jet. GEOSAT measurements also confirm the convergence of eddy momentum flux at zonal frontal zones in the eastern North Atlantic (Beckman et al., 1994), both at the subpolar front and the Azores Front (see also Le Traon and De Mey, 1994).

The interannual variations in eddy-mean flow interactions have been investigated in the Kuroshio Extension region by Qiu (1995) and Adamec (1998). During the first 2 years of T/P, Qiu (1995) noted significant variations in EKE levels in the Kuroshio Extension and its recirculation gyre (see Figure 6). The EKE in the southern recirculation gyre increased during this period, and energetic analyses showed an energy transfer from the mean field to the eddy field, due to barotropic instabilities. This energy transfer also led to a reduction in the intensity of both the eastward-flowing Kuroshio Extension and the westward recirculation gyre. The relation between the barotropic energy conversion and eddy heat fluxes was also examined by Adamec (1998) using a combination of T/P and AVHRR data. During summer 1994, the convergence of surface Reynolds stresses in the Kuroshio Extension was much weaker than normal, leading to low values of EKE three months later. The time difference is consistent with the characteristic time scales of the eddy energy (about 90-100 days) in this region. The upper ocean heat content was then calculated using a combination of AVHRR sea surface temperature data and XBT data. The lower EKE in summer 1994 was accompanied by a cooler southern recirculation and warming north of the Kuroshio Extension. This decreased the large-scale baroclinicity across the jet and therefore reduced the near surface transport and EKE. Calculations of the baroclinic zonally symmetric circulation indicated that the fronts were also substantially weakened during summer 1994, consistent with the reduced EKE. The primary reason for this change in baroclinicity was due to changes in the convergence of eddy heat fluxes (Figure 24). Adamec (1998) suggested that interannual variations in eddy activity, including eddy heat fluxes, may play a crucial role in modulating the seasonal signal in the Kuroshio Extension.

Another technique which involves less spatial smoothing of the eddy signals is to calculate the surface geostrophic velocity vectors at altimeter crossover points (Morrow et al, 1992, Johnson et al., 1992), as described in Section 2.3.1. Morrow et al (1992, 1994) applied this technique using two years of GEOSAT data in the Southern Ocean, and found anisotropic geostrophic velocity variability in the vicinity of the major currents and strong bathymetric features. The variance ellipse orientation also has important implications for the eddy flux of horizontal momentum: where the ellipse axes are aligned parallel or perpendicular to the mean flow, there is no cross-stream transfer of momentum. Eddy momentum fluxes in the Southern Ocean were calculated using this crossover technique by Morrow et al. (1992;

1994), using streamwise co-ordinates. They found a net convergence of alongstream momentum along the mean axis of the ACC, suggesting that eddies tend to accelerate the mean jet. The GEOSAT data revealed a very complex geographical distribution of Reynolds stress convergence and divergence, suggesting that other instability mechanisms may be important locally. The altimeter data also provided the first quantitative proof that the Southern Ocean eddy momentum flux divergence was too small, and in the wrong direction, to balance the momentum input by the wind.

4.6.2 EDDY TRANSPORTS, EDDY VISCOSITY AND EDDY DIFFUSIVITY

Eddies are an important mechanism for the transport of heat, salt and momentum in the ocean. However, the eddy transport of different properties has remained a large uncertainty for estimating the global property budgets, due to the difficulty and expense in measuring eddies in-situ. In addition, eddy effects must be parameterized in coarse-resolution ocean models, and the parameterizations are tested on a very limited eddy data base. Altimetry allows us to monitor the global surface eddy characteristics of the flow, and has the potential to improve our understanding of the geographical distribution of eddy transports of different properties.

In turbulence theory, the turbulent Reynolds stresses can be taken as proportional to the local gradient of the mean flow, via a proportionality term given as the eddy viscosity, ν. For example, for the zonal component of the flow, u, the horizontal and vertical eddy viscosity terms are defined from:

$$-\overline{u'u'} = \nu_x \frac{\partial \bar{u}}{\partial x}; \quad -\overline{u'v'} = \nu_y \frac{\partial \bar{u}}{\partial y}; \quad -\overline{u'w'} = \nu_z \frac{\partial \bar{u}}{\partial z} \tag{12}$$

In coarse-resolution ocean models, the eddy viscosity is often assumed constant. Altimetry allows us to investigate the global distribution of the surface eddy viscosity. The formulation given above has been used by Johnson et al. (1992) to estimate eddy viscosity locally in the Pacific sector of the ACC. Using a combination of GEOSAT altimetry for the horizontal Reynolds stresses and climatological data for the zonal mean flow, they estimated an average eddy viscosity of ν_y = 8 x 10^3 m^2/s, consistent with other estimates for oceanic flows. They note however, that point measurements of eddy viscosity undergo much larger fluctuations. Note that this parametrization is obviously oversimplified as eddies or waves can drive mean motions yielding negative viscosity.

The eddy transport associated with any passive tracer can be represented in the form of a Fickian diffusion:

$$v'\lambda' = -\kappa \nabla_h \overline{\lambda} \qquad (13)$$

where $\overline{\lambda}$ is the time-mean passive tracer field (such as salt or temperature) and κ is an isotropic diffusivity tensor. Again, in coarse-resolution ocean models, κ is often taken as spatially uniform in all three spatial dimensions, so estimating its geographical distribution is useful for improving model parameterizations.

A number of different techniques have been used to estimate the eddy diffusivity, κ, for passive tracers directly from altimetry. Assuming conditions of statistically homogeneous, barotropic, β-plane turbulence, Holloway (1986) and Keffer and Holloway (1988) estimated $\kappa \cong C\tau\psi$, where C is a proportionality constant (~0.4) and τ is an O(1) anisotropy tensor induced by Rossby wave propagation. ψ is the r.m.s. value of the streamfunction derived from altimetric sea level anomalies h', where $\psi = gh'/f$, g is the acceleration of gravity, f is the Coriolis parameter. The calculations by Holloway (1986) and Keffer and Holloway (1988) are based on 3 months of SEASAT data, and include various *ad-hoc* scaling factors to take account of the short time series, and the fact that the surface variability overestimates the depth-averaged eddy fluxes. To derive meridional eddy heat and salt fluxes in the Southern Ocean, they apply equation (13) with the meridional gradient in mean temperature and salinity derived from Levitus climatological data. Given the data limitations, they derive surprisingly good estimates of poleward heat flux, sufficient to supply the heat lost to the atmosphere south of the Antarctic Polar Front. The salinity fluxes were plausible at most latitudes, but poleward at high latitudes and thus in the wrong direction to compensate for the observed excess of P-E south of the Polar Front. (The errors in the salinity calculations are not surprising, given that the SEASAT data spanned the austral winter from July to October, when in-situ salinity observations are almost non-existant).

An alternative technique developed by Stammer (1998) is to calculate the eddy diffusivity, κ, as proportional to the typical horizontal turbulent velocity scale, u', and length scale, l', i.e., $\kappa \propto u'l'$, both derived from altimeter data. This is rewritten in terms of altimetric eddy kinetic energy, K_E, and an averaged eddy integral timescale derived from altimetry on a 5° geographical grid, T_{alt}:

$$\kappa = 2\alpha K_E T_{alt} \tag{14}$$

Here, α is a scaling factor, $\alpha = 0.7 \times (L_{R0}/L)^2$, where L_{R0} and L are the first internal Rossby radius and an eddy mixing scale; α is further scaled by 0.1 to convert altimetric surface EKE values to an averaged EKE over the top 1000 m. The factor of 2 is included to convert the integral timescale, T_{alt}, to the equivalent of the first zero crossing of the auto-correlation function, which is closer to the decorrelation scales observed from current meter data.

Stammer (1998) calculates the eddy diffusivity κ globally from T/P altimetry using equation (14), and compares this with the method of Holloway (1986). Both techniques give a similar geographical distribution, but the Holloway calculation is twice as large as Stammer's. This may be because the streamfunction calculation includes more large-scale variability from seasonal steric changes and planetary wave motion, which are unrelated to eddy transfer processes. Eddy diffusivity values reach 2500 m^2s^{-1} in the energetic western boundary currents, decreasing to 250 m^2s^{-1} in the interior and eastern parts of the basins, and the geographical distribution is quite inhomogeneous. Stammer (1998) also calculates global eddy heat and salt transports using mean meridional gradients of temperature and salinity from Levitus mean climatology. Strong poleward eddy heat and salt transports occur in the energetic western boundary currents: the Gulf Stream, the Kuroshio, and the Agulhas Current (Figure 25). In the equatorial band 5°N to 5°S, eddy heat and salt fluxes are northward in the Pacific and southward in the Indian Ocean, with equatorward eddy transports occurring between 5°-20° latitude.

These altimeter-based eddy transports are consistent with local in-situ estimates but demonstrate the strong spatial inhomogeneity of the eddy diffisivity field. One should note, however, that numerous ad-hoc scaling factors have been included, in particular for the vertical distribution of properties, and as such they provide a first order and qualitative approximation of eddy transports. In addition, these transport estimates are also based on climatological mean meridional gradients, and so should be considered a lower bound on instantaneous eddy transports in the ocean. In the future, better estimates should be possible using eddy resolving models with assimilation of remote sensing and in-situ data.

5. CONCLUSIONS

Satellite altimetry has made a unique contribution to observing and understanding ocean currents and eddies, which stems from its excellent space-time coverage, providing both a quasi-synoptic description and a statistical description of the ocean surface circulation. We now have a much better description of the upper ocean current systems and their associated mesoscale variability. Altimeter data analyses have produced global, quantitative estimates of eddy energy with high spatial resolution, revealing details such as the correlation of EKE with the mean currents and the role of the bathymetry. They have provided, for the first time, a global description of the seasonal/interannual variations in eddy energy. In most regions, significant mesoscale variations appear to be related to changes in the intensity of the mean current instabilities. An additional feature revealed by studying these variations has been the possible role of forcing by fluctuating winds in a few regions of the ocean. The frequency/wavenumber mesoscale circulation spectrum has been characterized, as have been the corresponding time and space scales. The eddy/mean flow interactions have also been mapped for the first time and provide an important ingredient for understanding the western boundary current and ACC dynamics. Our examples also show the potential of synoptic mapping of mesoscale variability, for example for monitoring Agulhas eddies. Such studies are useful to explain the structure of eddies and to better understand eddy dynamics. They provide a good means of testing and validating models and theories.

Despite all this progress, there is still much to learn from altimeter data for mesoscale variability studies. Some suggestions for future work are given below. Recent improvements of eddy resolving model simulations are impressive and models have now a high degree of realism. More detailed comparisons of altimetry (including comparison of higher order statistics such as frequency/wavenumber spectra and Reynolds stresses) with eddy resolving models should now be very instructive. The global frequency/wavenumber characteristics of altimeter sea level variability could also be explained and possibly related to different kinds of forcing. It may also be complemented by a 3D (x,y,t) spectral analysis of combined T/P + ERS-1/2 maps which would allow us in addition to better characterize the anisotropy of the eddy field. A better understanding of altimeter wavenumber spectra will also require more detailed comparisons with models and quasi-geostrophic turbulence simulations and theories. This is important for the interpretation of altimeter data in regions of weak wavenumber spectral slopes. Systematic studies of individual eddies should also be performed (such as studies of Agulhas eddies) and the relation between eddies and Rossby waves should be better analyzed. Finally, the seasonal/interannual variations of the major current systems and the seasonal/interannual variations in eddy energy can be analyzed further since we now have almost ten years of good altimeter data sets (at least for mesoscale studies). All these studies should take advantage of the sampling provided by the combination of several altimeter missions (two and possibly three missions).

The combined use of altimeter data with other data sets such as in-situ hydrographic data (T and S profiles), current meters and drifters, SST from infra-red imagery and forcing fields should also be developed further. The comparison/combination of altimeter data with TOGA and WOCE surface drifter data should provide, for example, a means of estimating the ageostrophic component of the surface circulation. The vertical structure of mesoscale variability should also be obtained (at least in a statistical sense – see Wunsch, 1997) from an analysis of in-situ data such as the ARGO profiling float array (ARGO Science Team, 1998). More generally, the integration of altimeter data with in-situ data (T and S profiles, drifters, floats) and forcing data is crucial for a better understanding of ocean currents and mesoscale variability dynamics. Assimilation into models is a powerful means for performing such an integration. This aspect is developed in the last chapter of this book and is obviously highly relevant to the above discussion. It is related to the development of operational oceanography which will benefit to scientific studies on ocean currents and eddies by providing (as it is today for the atmosphere) a regular description of the three dimensional ocean dynamically consistent with altimeter data, in-situ data and forcing data. This is one of the challenges of the future Global Ocean Data Assimilation Experiment (GODAE) (Smith and Lefebvre, 1997).

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FIGURE CAPTIONS

Figure 1 : Impact of a SLA white noise of 3 cm rms on the error on the cross-track velocity for different choices of SLA low-pass filtering cut-off wavelengths and for different latitude bands. Velocities are calculated from slopes of filtered SLA over 14 km. Units are cm/s.

Figure 2 : Mean and standard deviation of Sea Level Anomaly (SLA) mapping error for single and multiple altimeter missions. The calculation assumes a space scale of 150 km (zero crossing of the correlation function) and an e-folding time scale of 15 days. Units are percentages of signal variance (from Le Traon and Dibarboure, 1999).

Figure 3 : Schematic showing the major current systems described in this section.

Figure 4 : Sketch of the Gaussian jet model. (a) the Gaussian velocity profile is characterised by the maximum velocity a_1, the central position, a_2, and the jet width, a_3. These parameters are allowed to vary, and are adjusted to fit the altimetric residual heights data, using a least-squares technique. (b) Sea surface height is obtained by integrating the velocity profile. (c) The angle between the groundtrack and a perpendicular to the Gulf Stream axis. (from Kelly and Gille, 1990). The bottom panels show geostrophic velocity profiles from altimetry residuals and synthetic mean (solid) and the Gaussian model (dashed) in the Gulf Stream region for the first six GEOSAT cycles.

Figure 5 : Seasonal variations of sea level difference across the Gulf Stream from TOPEX/POSEIDON data (dashed line) and estimated from ECMWF net surface heat fluxes (solid line). The component due to Gulf Stream seasonal position variations has been removed. (from Kelly et al., 1999).

Figure 6 : Eddy Kinetic Energy distribution in the Kuroshio Extension and its recirculation for the periods: (a) Oct 92 to Apr 93, (b) Apr 93 to Oct 93, (c) Oct 93 to Jun 94. The contour unit is 0.01 m^2s^{-2}, areas with energy levels greater than 0.2 m^2s^{-2} are shaded. (from Qiu, 1995).

Figure 7 : (a) The ASUKA Group's in-situ observations along a TOPEX/POSEIDON satellite groundtrack in the Kuroshio, deployed from October 1993 to November 1995. (b) Time series of the absolute volume transport estimated from TOPEX/POSEIDON data, using (c) the regression relation between transport variations and SLA derived from the in-situ data. (from Imawaki and Uchida, 1997).

Figure 8 : Mean frequency/wavenumber spectrum of SLA in the Brazil/Malvinas Confluence area as derived from GEOSAT data. Units are in cm^2. There is a peak at the semi-annual frequency while little energy is found at the annual frequency. (from Provost and Le Traon, 1993).

Figure 9 : Velocity variance ellipses in the East Australian Current region from (a) GEOSAT observations at crossover points and (b) long-term surface drifter data plotted over bathymetry. (from Wilkin and Morrow, 1994).

Figure 10 : (a) Forcing terms for the poleward Leeuwin Current for the 3-year period 1993-1995 : the alongshore pressure gradient from TOPEX/POSEIDON data (solid line) drives onshore geostrophic flow which is balanced by the offshore wind forced flow from the alongshore wind stress from ERS-1 (dashed line). Positive forcing indicates conditions for a poleward coastal current. (b) similar fields from climatology, with no semi-annual signal (from Morrow and Birol, 1999).

Figure 11 : (a) Zonal gradients of residual sea surface heights from the Fine Resolution Antarctic Model (FRAM) compared with (b) an EOF reconstruction of similar fields from TOPEX/POSEIDON, for 5 Feb 1993. The pattern highlights Rossby waves, with wavelengths of 200 - 400 km (from Hughes, 1995).

Figure 12 : Mean sea surface height across the Antarctic Circumpolar Current reconstructed from GEOSAT height variability using a meandering Gaussien jet, after Gille (1994). Contour interval is 0.2 m.

Figure 13 : Agulhas eddy trajectories computed from the T/P data for 1993 to 1996. Note the changes from a broad Agulhas eddy corridor in 1993/1994 to a narrower corridor in 1996. Different colours represent individual eddy tracks. (from Witter and Gordon, 1999).

Figure 14 : The meandering of the Azores front east of the Mid-Atlantic ridge as observed from (a) hydrographic and (b) drifter data from the Semaphore experiment and (c) ERS-1 and (d) TOPEX/POSEIDON data (from Hernandez et al., 1995).

Figure 15 : General circulation in the Mediterranean sea and main ocean circulation features. Permanent features are solid lines and recurrent features are dashed lines. (from Iudicone et al., 1998).

Figure 16: Seasonal/interannual variations of the Mediterranean circulation as derived from the combination of T/P and ERS-1/2. The maps are 3-month averages and correspond to the the summer 1993 and the summer 1996. Units are in cm. (courtesy G. Larnicol).

Figure 17 : Rms Sea Level Anomaly from 5 years of TOPEX/POSEIDON and ERS-1/2 combined maps. Units are in cm. (from Ducet and Le Traon, 1999).

Figure 18 : Time series of monthly KS ($K_S=\sin^2(\phi)\, K_E$, where K_E is the T/P cross-track geostrophic velocity variance and ϕ is the latitude) and τ^2 (where τ is the wind stress) in the North Atlantic (from Stammer and Wunsch, 1999).

Figure 19 : Time series of the EKE in the STCC region (from Qiu , 1999).

Figure 20 : Seasonal change in the vertical shear between the STCC and its underlying NEC (solid line) versus the seasonal change in the EKE (dashed line) (from Qiu, 1999).

Figure 21 : Eddy scales estimated from TOPEX data as the integral scale of the Sea Level Anomaly autocorrelation function (up to the first zero crossing) and averaged zonally between 0° and 360° in longitude (from Stammer, 1997).

Figure 22 : Scatter diagram of the first zero crossing of the SLA autocorrelation function against the corresponding first internal Rossby radius (from Stammer, 1997).

Figure 23: Along-track Sea Level Anomaly wavenumber spectra in the North Atlantic as derived from two years of GEOSAT data (from Le Traon et al., 1990).

Figure 24 : Wavelet transform of the energy conversion for 33° N –35° N due to (a) barotropic instability and (b) downgradient heat flux. The abscissa it time in 10-day TOPEX/POSEIDON cycle numbers: Cycle 11 corresponds to Jan 1993, cycle 120 to Dec 1995, and the change in response occurs around cycle 70 (August 1994). The ordinate ("dilation") is a function of frequency: frequency doubles for each integer value of dilation, and zero dilation corresponds to mesoscale frequencies. (from Adamec, 1998).

Figure 25 : (a) Meridional eddy heat transport, v'T'; contour interval is 5 x 10^6 W m^{-1}. (b) Meridional eddy salt transport, v'S'; contour interval is 5 kg m^{-1}. Negative values are dashed. (from Stammer, 1998).

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