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SATELLITE OCEANOGRAPHY FOR OCEAN FORECASTING

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1. OUTLINE

This lecture aims at providing a general introduction to satellite oceanography in the context of ocean forecasting. Satellite oceanography is an essential component in the development of operational oceanography. Major advances in sensor development and scientific analysis have been achieved in the last 20 years. As a result, several techniques are now mature (e.g. altimetry, infra-red imagery) and provide quantitative and unique measurements of the ocean system.

We begin with a general overview of space oceanography, summarizing why it is so useful for ocean forecasting and briefly describing satellite oceanography techniques, before looking at the status of present and future missions. We will then turn to satellite altimetry, probably the most important and mature technique currently in use for ocean forecasting. We will also detail measurement principles and content, explain the basic data processing, including the methodology for merging data sets, and provide an overview of results recently obtained with TOPEX/POSEIDON and ERS-1/2 altimeter data. Lastly, we will focus on real-time aspects crucial for ocean forecasting. Perspectives will be given in the conclusion.

2. OVERVIEW OF SPACE OCEANOGRAPHY

2.1 WHY DO WE NEED SATELLITES FOR OCEAN FORECASTING?

An ocean hindcasting/forecasting system must be based on the assimilation of observation data into a numerical model. It also must have precise forcing data. The ocean is, indeed, a turbulent system. "Realistic" models of the ocean are impossible to construct owing both to uncertainty of the governing physics and of an initial state (not to mention predictability issues). Continuous observations are required to drive the model towards a realistic state. Ocean forecasting therefore calls for an ocean observing system which should also include forcing data. Both in-situ and satellite data are needed:

- The usefulness of in-situ data is limited by poor space/time coverage, and access to remote regions (e.g. southern oceans) is often difficult. Conventional techniques (ship measurements, ship deployment) also may not be suitable for an operational system. Reliable, autonomous techniques have to be used. Such techniques are already being used or starting to emerge (e.g. XBT, TOGA/TAO, PIRATA, profiling floats such as P-ALACE or PROVOR) and promising new techniques (e.g. gliders, acoustic tomography) are under development. Global operational monitoring of the ocean with in-situ data is still, however, to be developed. The *Argo* project (Roemmich et al., 1999) will be a major step towards this objective.

- Satellite data, on the other hand, can provide long-term, repeated (synoptic) and global measurements of key parameters (e.g. sea level, SST, winds). These are surface parameters but they may be representative of the deep ocean (e.g. sea level from altimeters). An additional advantage is that measurements can be acquired and processed in near-real time.

There is an obvious complementarity between satellite and in-situ data. Satellite and in-situ measurements have different contents (e.g. vertical structure from in-situ data, surface information from satellites) and very different space/time sampling. In-situ data are also required for calibration and long-term validation of satellite data.

2.2 SPACE OCEANOGRAPHY TECHNIQUES

2.2.1 OVERVIEW

Many space oceanography techniques have been developed in the last two decades. I will base my discussion on a broad definition of space oceanography encompassing direct observation of the ocean, ice and ocean forcing terms (wind, heat flux, precipitation, evaporation). Technologies range from passive radiometers measuring the natural electromagnetic signal which radiates from the sea to active radiometers (radars) which transmit a signal and measure the modified echo signal. Passive radiometers allow us to look at different oceanic signals depending on the wavelength. Wavelength bands available are actually limited by absorption of the electromagnetic signal by molecular aerosols and water vapor.

A brief overview of space techniques and of the retrieved oceanographic parameters is given here. For a more detailed description the reader is referred, for example, to Robinson (1985), Weller and Taylor (1993) and Ikeda and Dobson (1995).

In the visible band, between 0.4 μm and 0.8 μm, we find measurements of ocean color (see below) and surface solar radiation flux [one component of the net heat flux at the ocean surface which can be well retrieved from space data (Planton, 1994)]. In the infrared band (0.8 μm to 1000 μm), we find, of course, the Sea Surface Temperature (SST) and the infrared flux, although, due to cloud cover, the latter is difficult to determine precisely from satellite measurements. In the visible and infrared bands, the main limitation is cloud cover. SST and ocean color measurements from space can thus only be obtained for clear skies. Sea ice surface characteristics can also be distinguished in the visible and infrared bands (AVHRR).

In the microwave band (0.1 cm to 100 cm), cloud cover is no longer a problem. Main parameters which can be measured in the microwave band are the integrated atmospheric water vapour (not directly relevant here), sea ice and surface wind speed. The absence of cloud cover is particularly important for sea ice studies (measurement of ice concentration and ice extent). Surface precipitation is a non-radiative parameter which can be inferred from the satellite radiative measurements (microwave radiation is related to atmospheric liquid water content, not directly to rain rate, but a good correlation between the two parameters exists—besides, precipitation is poorly known and satellites provide very useful information). Note that visible and infrared techniques can also provide useful information on precipitation. Latent heat flux or evaporation rate can also be inferred from SST, wind speed and integrated water vapour (although data on surface level humidity would be needed). Finally, one should note that SST can also be retrieved from microwave radiation but with a lower resolution and accuracy compared to infra-red radiation. There is also an interesting capability for measuring Sea Surface Salinity from microwave measurements.

Main active microwave radiometers or radars for satellite oceanography are: altimeters scatterometers and Synthetic Aperture Radars. Satellite altimetry provides measurements of the sea surface topography, significant wave height and wind speed modulus. It also provides measurements of ice sheet topography (e.g. mass balance of Antarctica and Greenland ice sheet). We shall describe altimeter measurement principles in the next section. A scatterometer measures the electromagnetic signal backscatter coefficient in three directions. The backscatter coefficient depends on wind speed and direction. By analyzing backscatter in different directions, we can estimate the wind speed direction. There is a 180° ambiguity because differences in backscatter 180° apart are very small. This ambiguity is generally removed through consistency check from previous measurements and/or using atmospheric ocean models. Scatterometers thus provide wind speed and wind direction measurements. They are also useful for measuring sea ice, although not as useful as microwave radiometers. Synthetic Aperture Radar (SAR) is used mainly to measure the wave spectrum. These measurements may also be used to map the winds and ocean currents at small scales but the signal is complex and difficult to invert (Johannessen, 1995). SAR images also provide useful measurements of sea ice parameters (Johannessen et al., 1995).

Finally, one has to consider satellite data collection and/or location systems for in-situ measurements. They are vital for operational oceanography since they allow the near real time data collection and, if needed, the location of drifting instruments for velocity estimations (e.g. surface drifters, ALACE, profiling floats). In particular, it includes the well known ARGOS system.

2.2.2 ALTIMETRY, SST AND OCEAN COLOR

A ranking of satellite oceanography techniques for ocean forecasting objectives would probably give altimetry, SST, scatterometers and probably ocean colour. The reason why altimetry is so useful will be detailed in the next section. We will not discuss further scatterometry as we focus here on ocean parameters. The availability of winds with high space and time resolution is, however, a critical requirement for ocean forecasting (see Milliff et al, 1999 for a review).

It is clear that SST measurements are crucial for understanding and predicting the ocean/atmosphere system. SST is an important factor that influences (and is influenced by) the ocean and atmosphere circulation. The heat fluxes at the ocean/atmosphere interface are also strongly linked to SST. This parameter is operationally measured by the NOAA satellites (AVHRR), and useful data sets are already available. Geostationary satellites (e.g. GOES, METEOSAT) provide a much better temporal sampling but with a degraded spatial resolution and accuracy. SST anomalies are key indicators of changes in the environment (e.g. El Niño events). At smaller scales, SST measurements can provide useful information on the mesoscale ocean flow field and position of the main current system (e.g. Gulf Stream and ocean color measurements) and regions of upwelling. Surface currents may also be derived from the analysis of successive SST images (e.g. Kelly and Strub, 1992).

The usefulness of ocean color measurements was demonstrated by the CZCS instrument on board the Nimbus-7 satellite (1978-1986). There are now several ocean color missions flying, and more are scheduled for the years ahead. Ocean color provides an indirect way of measuring phytoplankton concentration (chlorophyll-a pigment, for the so-called Case 1 waters) but is also sensitive to the influence of sediments, particulate and dissolved organic matter (so-called Case 2 waters found mainly in coastal areas) (see Bricaud, 1995). Phytoplankton is the primary food and energy source for the ocean ecosystem, so it is very important for ecosystem studies. It also plays a significant role in the carbon cycle since phytoplankton dissolves CO_2 through photosynthesis into organic components (biological primary production). Global mean chlorophyll pigment concentration derived from CZCS data shows, for example, regions of high phytoplankton concentration in the upwellings and at high latitudes. Near the coast, the signal cannot be interpreted in terms of phytoplankton only. From the chlorophyll pigment concentration maps, we can derive maps of primary production (production of organic carbon through photosynthesis). Quantitative estimation of phytoplankton and primary production via ocean color measurements is difficult, however, as the retrieved signal is not always dominated by chlorophyll (dissolved organic matter, other pigments). The absorption coefficient of phytoplankton is also variable (according to how it is aggregated). To estimate primary production, we must assume a vertical distribution of chlorophyll and primary production also depends on species, environmental and physiological conditions, etc. There is clearly a need for better sensors and sampling, more wavelengths (SEAWIFS, OCTS, POLDER) but also comprehensive calibration, validation, and new bio-optical algorithms before ocean color measurements can be used quantitatively. These data sets are nevertheless very useful for verifying coupled ecosystem models. Chlorophyll concentration observation is also a useful parameter to be assimilated in these models.

2.3 SPACE OCEANOGRAPHY MISSIONS

Tables below provide an overview of existing and approved satellite missions for altimetry, SST, scatterometry and ocean color (courtesy of A. Ratier) for the next decade [before and during the Global Data Assimilation Experiment (GODAE), Smith and Lefebvre (1997)] (Ratier, 1999). As can be seen, there are a substantial number of missions although only SST could be described as operational. There is not yet a commitment for long term, operational measurements. Note that, as far as altimetry is concerned, gravimetric missions (CHAMP, GRACE and GOCE) were added (see Johanessen et al., 1999). They should allow the estimation of a precise geoid. As we will see later, without precise knowledge of the geoid , altimetry can only provide data on the variable part of the ocean dynamic topography.

Finally, among promising future techniques, one should mention the possibility of measuring sea surface salinity from space. The concept has been demonstrated from plane measurements and there are projects in the Europe (SMOS) and in the USA for flying such a mission in the coming years. Although the precision will be somewhat limited, this will be a very important contribution to the ocean observing system (see Johanessen et al., 1999).

3. SATELLITE ALTIMETRY

3.1 OVERVIEW

There are two main reasons for focusing on satellite altimetry. First, it is probably the most important satellite technique for oceanography, particularly for ocean forecasting. Satellite altimetry has unique capabilities for providing global synoptic view of the ocean circulation. It provides measurements of sea surface topography which is an integral of the ocean interior. These measurements are a strong constraint for the 3D ocean circulation estimation. Second, satellite altimetry is a very mature technique. It started more than 20 years ago with GEOS-3 (1975) and SEASAT (1978) missions. After a seven year gap, these were followed by the US Navy's GEOSAT mission (1985-1989) and the European Agency ERS-1 mission (1991-1996). Two satellites are now operating : ERS-2, the successor of ERS-1 and the US/French TOPEX/POSEIDON (T/P) mission. For the next decade, future missions have already been decided : GEOSAT Follow On was launched in early 1998 and ENVISAT (successor of ERS-1/2) and Jason-1 (successor of T/P) missions are scheduled for launch in early 2001.

Satellite altimetry is one of the most complex and challenging techniques in terms of accuracy. It requires the range between the satellite and the sea surface to be measured to within a few cm, i.e. assuming a typical satellite orbit altitude of 1000 km a relative accuracy of 10^{-8}. There have been major advances in sensor and processing algorithm performance over the last 20 years. It is important to realize that these advances were made possible through continuous cooperation between engineers and scientists. As a result, accuracy has progressed from several meters to a few cm only. T/P marked a major improvement in accuracy. Its orbit is known with an accuracy of about 2 cm rms and the satellite - ocean surface distance is determined to within 2 cm.

We shall now describe measurement principles, content and errors, and explain data processing techniques. More details can be found in Stewart (1985), Chelton (1988), Rummel and Sanso (1993) or Le Traon (1995). We will then look briefly at results recently obtained with T/P and ERS-1/2 altimeter data before discussing real-time aspects.

3.2 MEASUREMENT PRINCIPLE

The principle of altimetry measurement is simple (although the system is complex). The altimeter measures the range from the satellite to the ocean surface by determining the time taken by the radar pulse to travel from the satellite to the ocean surface. Using a precise orbitography system, we can determine the position of the satellite relative to a reference ellipsoid. Combining these two measurements yields an estimation of the sea level relative to a reference ellipsoid. This estimation comprises the geoid (an equipotential of the earth gravity field to which a motionless ocean would exactly conform) and the ocean dynamic topography. The geoid can vary as much as 100 m for only 1 m for the dynamic topography (the parameter of interest here).

Altimetry also provides measurements of significant waveheight (the radar pulse is first reflected by crests and then by troughs) and wind speed (the power of the returned signal depends on wind speed). These parameters are also used to estimate electromagnetic bias correction (interaction of the electromagnetic wave with the ocean surface).

The satellite repeats exactly the same ground track every cycle. For T/P and Jason-1, this cycle is 10 days, for ERS-1/2 multi-disciplinary phase and for ENVISAT it is 35 days, and for GEOSAT and GFO 17 days. Every cycle, it thus observes the same geoid signal (which does not vary over time) and the dynamic topography (which varies over time).

3.3 CONTENT OF ALTIMETRIC MEASUREMENT

3.3.1 GEOSTROPHY

Assuming the geoid is known (we will discuss this particular problem later on), satellite altimetry provides measurements of ocean surface dynamic topography, which is directly related to ocean currents through the geostrophic approximation. Simple scaling of the Navier Stokes equations shows that for space scales larger than a few tens of km and time scales longer than a few days (except for a thin layer at the surface which is directly driven by winds—the so-called Ekman layer), the circulation is in geostrophic equilibrium (e.g. Pedlosky, 1979). This means that the pressure force is balanced by the Coriolis force. This yields equations 1 and 2. Equation 3 represents hydrostatic balance.

$$\begin{cases} fv = \dfrac{1}{\rho_0}\dfrac{\partial P}{\partial x} & (1) \\ -fu = \dfrac{1}{\rho_0}\dfrac{\partial P}{\partial y} & (2) \end{cases} \qquad f = 2\Omega \sin\theta$$

At the surface, P (the pressure) is equal to $\rho\ g\ \eta$, where η is the dynamic topography (sea level above the geoid, local vertical), ρ is density and g is gravity. The dynamic topography and ocean surface currents are thus simply related :

$$\begin{cases} fv = g\dfrac{\partial \eta}{\partial x} & (1') \\ -fu = g\dfrac{\partial \eta}{\partial y} & (2') \end{cases}$$

$$\left\{ \dfrac{\partial p}{\partial z} = -\rho g \qquad (3) \right.$$

This is, of course, only a diagnostic relationship. It does not tell us what causes the ocean motion (which would require the other terms in the Navier Stokes equations to be taken into account). What it does tell us is that where there is motion there must be a deviation relative to the geoid as expressed in equations 1' and 2'. The physical explanation for this is simple. Pressure forces will tend to cause water to flow to a sea level low. However, because the motion is relatively large-scale, the Coriolis force is large and will deviate the water to the right (left) in the Northern

(Southern) Hemisphere. The water will be deviated up to 90° and an equilibrium between pressure and Coriolis forces will be reached. As a result the flow will be cyclonic (clockwise in the N.H.) for the low and anticyclonic for the high. The same holds for the atmosphere: ocean dynamic topography is related to ocean surface current in oceanography in the same way as atmospheric pressure is related to surface wind in meteorology.

A map of the mean dynamic topography as derived from T/P data and a geoid model (e.g. Stammer et al., 1996) reflects the main features of the ocean circulation. Ocean subtropical gyres which are characterized by high dynamic topography and anticyclonic motions, subpolar gyres and the Antarctic Circumpolar Current which is characterized by a sea level drop of more than one meter from north to south. Using satellite altimetry, we can thus obtain pictures of the ocean circulation on a regular basis (e.g. every 10 days for T/P).

3.3.2 BAROCLINIC AND BAROTROPIC MOTIONS, STERIC HEIGHT

The content of the ocean dynamic topography signal is now detailed to show how it is related to the ocean interior. More details can be found in the Gill and Niiler (1973) paper.

Let us first assume a homogeneous ocean [ρ = constant or $\rho = \rho$ (z)] :

$$fv = \frac{1}{\rho_0}\frac{\partial P}{\partial x} \quad \Rightarrow \quad f\,\frac{\partial v}{\partial z} = \frac{-g}{\rho_0}\frac{\partial \rho}{\partial x} \quad \text{(from 3)}$$

$$\Rightarrow \quad \frac{\partial v}{\partial z} = 0$$

There will be the same velocity over the total water column. These are the so-called barotropic motions which do not depend on the density field.

The density actually depends on x, y, z and t [$\rho = \rho(x,y,z,t)$]. Let us define ρ' as the density anomaly given by $\rho = \langle\rho(z)\rangle + \rho'$, where $\langle\rho\rangle$ is the water density at a given temperature and salinity (e.g. T = 0°C , S = 35‰). From equation 3, we derive :

$$f\,\frac{\partial v}{\partial z} = \frac{-g}{\rho_0}\frac{\partial \rho}{\partial x} \qquad (4)$$

This is the thermal wind equation. Horizontal density variations are associated with vertical shear (so-called baroclinic motions). Integrating (4) from z_0 to z_1 yields :

$$v(z) = v(z_0) - \frac{g}{f}\int_{z_0}^{z_1} \frac{1}{\rho_0}\frac{\partial \rho'}{\partial x}dz \qquad (5)$$

If density is known, we only need to know the velocity at a given level (reference level) to calculate the velocity over the total water column. From equation 5, we can easily obtain equation 6:

$$(5) \Rightarrow v(z) = v(z_0) - \frac{g}{f}\frac{1}{\rho_0}\frac{\partial}{\partial x}\left(\int_{z_0}^{z_1}\frac{\rho'}{\rho_0}dz\right) \qquad (6)$$

$$\Leftrightarrow v(z) = v(z_0) + \frac{g}{f}\frac{\partial \eta_s}{\partial x} \text{ with } \eta_s(z_0, z_1) = -\int_{z_0}^{z_1}\frac{\rho'}{\rho_0}dz$$

η_s is the steric height. It is generally defined as the integral from the bottom to the surface.

Thus, at the surface:

$$\begin{cases} v(z_0) + \dfrac{g}{f}\dfrac{\partial \eta_s}{\partial x} = \dfrac{g}{f}\dfrac{\partial \eta}{\partial x} \\ with\ v(z_0) = \dfrac{1}{f\rho_0}\dfrac{\partial P_{z0}}{\partial x} \end{cases}$$

The same expression holds for u (with y derivative) which means that:

$$\Longrightarrow \eta = \eta_s + \frac{P_{z0}}{\rho_0 g}$$

The dynamic topography η (measured by altimetry) is the sum of the steric height (baroclinic component) and of a bottom pressure term (sometimes incorrectly referred to as the barotropic component; the exact definition of a barotropic motion is a motion which does not depend on z; a barotropic motion will induce bottom pressure changes, but density changes, if not compensated over the water column, can also induce bottom pressure changes). This shows that the dynamic topography is an integral of density (steric height - baroclinic part), with a contribution from bottom pressure. It is really a measurement representative of the total water column (although it is directly related to surface geostrophic current). This is why satellite altimetry is so useful for constraining the 3D ocean circulation.

3.3.3 NUMERICAL APPLICATIONS

Influence of latitude on sea level slope and ocean current

The Coriolis parameter f is equal to $2\Omega \sin \phi$ (Ω is the angular rotation of the earth, ϕ is latitude). This gives $f \approx 10^{-4}\ s^{-1}$ (at 45°N/S) and $f \approx 10^{-5}\ s^{-1}$ (at 5°N/S). From equation 1' (and 2'), a slope of 1 meter over 100 km will give a current of 1 m/s at 45°N/S. A slope of 10 cm over 100 km only will give the same current speed at 5°N/S.

Influence of temperature gradient on dynamic topography

If we assume a reference level (level of no motion) at 1000 m and no salinity influence, the dynamic topography change measured by altimetry for a track crossing the Gulf Stream (or a ring) with a temperature gradient of 10°C over 100 - 200 km and over a depth of 1000 m will be:

$$\eta = \eta_s = - \int_{1000} \rho' / \rho_0\, dz \;\Rightarrow\; \Delta \eta_s = \Delta \eta = 1000\, \Delta\rho / \rho_0 = \alpha\ 1000\, \Delta T \approx 1.5 \text{ meter}$$

with $\alpha = -1/\rho\ \partial\rho/\partial T$ (thermal expansion coefficient) $\alpha \approx 1 - 2\ 10^{-4}\ s^{-1}$

3.3.4 EQUATORIAL REGIONS

At the equator, the Coriolis parameter f is equal to zero. This means we can no longer assume geostrophy or use dynamic topography to infer the oceanic circulation. It still reflects, of course, the change in density field (e.g. thermocline depth or heat content variations). In particular, the sea level variations (η) in the equatorial regions are closely related to the variations of the depth of the thermocline (H) ($\eta \approx - \Delta\rho / \rho_0\, H$, where $\Delta\rho$ is the density difference between the deep and surface layers). In practice, geostrophy works up to about ± 2 degrees (although the estimation is more sensitive to noise). A relationship between zonal current and dynamic topography has been proposed, however, by Picaut et al. (1990) ("equatorial geostrophy"). This relationship is derived as follows:

$$f\,u = - \partial\eta/\partial y \quad \text{(geostrophy)}$$

$$\text{derivative } /y \Rightarrow \partial f/\partial y\ u + f\, \partial u/dy = -g\, \partial^2\eta/\partial y^2 \;\Rightarrow \text{at the equator } \beta\, u = -g\, \partial^2\eta/\partial y^2$$

This relationship is valid at low frequencies but the estimation is very sensitive to noise (because of the second derivative). It has been shown to provide good results with altimetry (compared to currentmeter data) but altimeter data need to be filtered both in space and time.

3.3.5 STERIC EFFECT RELATED TO HEAT FLUXES

The dominant signal in seasonal large-scale variations in sea level is due to heat fluxes. This has been well illustrated with T/P (e.g. Stammer, 1997). It is important to accurately estimate this effect before analyzing other dynamically more significant effects, like the influence of wind forcing on circulation, for example.

As previously explained, the steric height is given by: $\eta_s = -\int_{bottom}^{surface} \rho' / \rho_0 \, dz$

Density is a function of temperature, salinity and pressure. For small changes of T and S, the variations of density are given by:

$$\frac{\Delta\rho'}{\rho_0} = \frac{\partial\rho}{\partial \mathrm{T}}\frac{\Delta \mathrm{T}'}{\rho_0} + \frac{\partial\rho}{\partial S}\frac{\Delta S'}{\rho_0} = -\alpha\Delta \mathrm{T}' - \beta\Delta S'$$

where α is the thermal expansion coefficient and β the equivalent for salinity. The contribution of temperature to density in the mixed layer is generally much more important than salinity (except at high latitudes).

The equation of heat (or potential temperature) conservation at the ocean surface reads:

$$\frac{\partial \mathrm{T}}{\partial \mathrm{t}} + \mathrm{u}\frac{\partial \mathrm{T}}{\partial \mathrm{x}} + \mathrm{v}\frac{\partial \mathrm{T}}{\partial \mathrm{y}} + \mathrm{w}\frac{\partial \mathrm{T}}{\partial \mathrm{z}} = \mathrm{K_v}\frac{\partial^2 \mathrm{T}}{\partial \mathrm{z}^2}$$

$$\textbf{with at z} = 0 \quad Q = \rho_0 C_\rho K_v \frac{\partial T}{\partial z}$$ with Q = air/sea heat flux

It is easy to show (e.g. Gill and Niiler, 1973) that at large scales (> 1000 km), advection is negligible. Integrating over H, the mixed layer depth, thus yields:

$$\int_{-H}^{0} \frac{\partial T'}{\partial t} dz = \frac{\partial}{\partial t}\int_{-H}^{0} T' dz = \frac{Q}{\rho_0 C_p}$$

Ignoring the salinity contribution on density, we obtain the following approximation :

$$\Rightarrow \frac{\partial}{\partial t}\eta_s' = \frac{\alpha}{\rho_0 C_p} Q \qquad (11)$$

The net heat flux induces (steric) sea level changes according to the simple equation given above.

This also means that sea level can provide useful information on heat fluxes. Thus, if properly used, satellite altimetry should allow us to correct for ocean heat flux errors (e.g. by assimilating altimetry and SST data). For ocean forecasting, this also shows that ocean dynamic topography is a superposition of different signals with very different vertical and horizontal scales. The complexity of the measurement content must be taken into account in the assimilation procedure.

3.4 ALTIMETER DATA PROCESSING

So far, we have considered that altimetry provides measurements of the surface dynamic topography. We now describe the basic altimeter data processing techniques used to extract the dynamic topography from measurements of sea surface topography. Our analysis includes a discussion of measurement errors.

3.4.1 OCEANIC SIGNAL EXTRACTION FROM ALTIMETRY AND MEASUREMENT ERRORS

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

$$S = N + \eta + \varepsilon$$

where N is the geoid, η the dynamic topography and ε the measurement errors (orbit error, propagation effects in the troposphere and ionosphere, tides, electromagnetic bias, inverse barometer effect, altimeter measurement noise).

Present geoids generally are not accurate enough to estimate the absolute dynamic topography η globally, except at very long wavelengths (> 2000 km) (see also discussion in section 2.3). The variable part of the dynamic topography η' ($\eta' = \eta - \langle\eta\rangle$) (or Sea Level Anomaly, hereafter SLA) is, however, easily extracted since no prior knowledge of the geoid height is needed. The most commonly used method is the so-called repeat track method (collinear analysis), 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 thus obtained by removing the mean profile, which contains the geoid and the quasi-permanent dynamic topography, from each profile.

Altimeter measurements of sea surface topography are affected by a large number of errors. These include errors on the range measurement due to propagation effects in the troposphere and ionosphere; electromagnetic bias; errors due to inaccurate ocean and terrestrial tide models; cross-track geoid errors, and inverse barometer effect. Most are large-scale, and do not limit the use of altimetry for ocean mesoscale studies. For studies of large-scale oceanic variability, most of these corrections must be taken into account, however, since they can significantly contaminate the oceanic signal. Some of these errors can be corrected with dedicated instrumentation (dual-frequency altimeter for ionospheric corrections, microwave radiometer for wet tropospheric corrections). Electromagnetic bias can be deduced by analyzing altimeter data.

The inverse barometer effect is actually a large, real oceanic signal which reflects the response of the sea level to changes in atmospheric pressure. The static response assumes a non-dynamic adjustment of the ocean due to atmospheric pressure (about 1 cm for 1 mb change in atmospheric pressure). Recent numerical and empirical studies suggest that the ocean generally responds as an inverted barometer except over very short time scales and in semi-enclosed seas (e.g. Gaspar and Ponte, 1997). This correction should thus generally be applied when studying the dynamic response of the ocean.

The tidal signal is the most important variable signal in altimetric data. This signal is only partially corrected using global ocean tide models. The residual errors are then aliased at certain periods depending on the repeat-period of the satellite. The satellite repeat-period should be chosen to avoid aliasing near dominant oceanic periods (e.g. annual or semi-annual periods). The M2 tide is thus aliased near 60 days for T/P. This is much less of a problem than the GEOSAT 317-day aliasing period which can cause major problems for the interpretation of altimetric signals. More accurate global tide models based on hydrodynamic models and/or T/P data are now available and should be used. Still, aliasing may occur and this has to be taken into account when interpreting the data.

Last, but not least, comes the orbit error. This error is caused by imperfect knowledge of the spacecraft position in the radial direction. It is actually the largest error on altimetric measurements of sea surface topography. It depends on the quality of the satellite tracking system. For T/P, precise orbit determination is achieved via three distinct tracking systems (DORIS, Laser, GPS) providing almost global coverage of satellite orbits. The radial orbit error obtained is thus accurate to about 2 cm, compared to 10 cm accuracy for the most recent GEOSAT and ERS-1/2 orbits available. Orbit errors are long-wavelength errors (about 40 000 km) that can be reduced by analyzing the altimeter data. This can be done via global or regional minimization of crossover or repeat-track differences (relative to a mean or to a given cycle). These differences do not contain any geoid signal and are dominated by the orbit error. Crossover or repeat-track differences also contain the oceanic signal. The main problem is thus to separate the orbit error from this oceanic signal. We must first assume an a priori analytical form or an a priori spectrum of the orbit error. A more empirical approach, commonly used, is to approximate the orbit error by fitting a first or second degree polynomial over a given arc length. However, this method also removes the large-scale oceanic signal and is only suitable for mesoscale studies. To minimize ocean signal removal, more complex methods using cross-track information are required. Global crossover minimization can thus be used to estimate the orbit error without removing too much of the large-scale oceanic signal. More generally using inverse techniques, the orbit error signal could be obtained through a global adjustment taking into account not only the spatial but also the temporal characteristics of the orbit error and oceanic signal.

3.4.2 MAPPING

Most applications need maps (and associated formal error) of the altimetric signal (SLA) on regular space/time grids. This can be done using optimal interpolation methods which use a priori knowledge of the space and time scales of the ocean signal. It is preferable, particularly in low eddy energy regions and when several altimeter data sets are merged, to take into account an along-track long wavelength error (correlated noise, e.g. due to orbit, tidal or inverse barometer residual errors) in the method (Le Traon et al., 1998).

3.4.3 MERGING OF MULTIPLE-SATELLITE ALTIMETER DATA

The merging of multi-satellite altimetric data sets is necessary for a better mapping of sea level and oceanic circulation variations. To merge muti-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 (e.g. same tidal models, same meteorological models, etc) should be used. Inter-calibrated means that relative biases and drifts be corrected for at the mm level and also that the orbit be reduced (to have consistent data sets). A methodology we proposed and successfully tested 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).

Once altimetric data have been homogenized and inter-calibrated, the next step is to extract the Sea Level Anomaly for the different missions and to merge the data using a mapping or assimilation technique. To extract the SLA, it is preferable to use a common reference surface (e.g. a very precise mean sea surface or mean profiles consistent between the different missions) to obtain the SLA relative to the same ocean mean. Then data can be merged using a mapping technique as previously described.

3.5 RESULTS FROM T/P AND ERS-1/2

An overview of results obtained with T/P and ERS-1/2 is now given to illustrate the contribution of altimetry to ocean forecasting. It contains T/P results on mean sea level variations, large scale sea level and oceanic circulation variations, Kelvin/Rossby waves, El Nino and the comparison with global eddy resolving models. T/P and ERS-1/2 results deal with the combination of T/P and ERS-1/2 for regional and mesoscale analyses. The reader is referred to the two T/P (1994, 1995) and the ERS-1/2 (1998) Journal of Geophysical Research special issues for a more complete overview.

To forecast possible global sea level rise, you need to observe the mean sea level trend over a very long period with accuracy better than 1 mm/year. To achieve this accuracy from 1000 km altitude is a real challenge. T/P is the first altimetric mission which enables us to observe the global mean sea level variations with such an accuracy. The signal over the first five years is between 1 and 2 mm/year (10 to 20 cm per century), consistent with results derived from tide gauges which were (and still are) extremely useful for validating altimeter measurements. The main advantage of altimetry is global coverage, which makes measurements much more representative of the global mean sea level. Satellite altimetry also offers a unique ability to observe mean sea level variations at smaller regional scales. We know that global warming will not induce a uniform rise in the mean sea level. T/P estimations show that the sea level rise can be much higher in some places (a few cm/year) although there are some regions where the signal is clearly related to an interannual ocean signal (not a secular drift). This also shows why it is so difficult to monitor the global mean sea level from a limited number of tide gauges. T/P data have also provided very useful estimates of mean sea level variations in regional seas (e.g. Mediterranean, Black Sea, Caspian Sea). The ability to forecast the mean sea level change in these high density regions is of utmost importance.

One of the main achievements of T/P has been its ability to measure large-scale sea level variability. Seasonal variations are mainly due to steric effects related to heat fluxes (see section 3.3). They are 180° phase-shifted between the two hemispheres and have a typical amplitude of 3 to 5 cm. These variations are correlated with sea surface temperature (expansion and contraction of surface waters) but with a phase lag of 1 to 2 months. The amplitude of the signal is smaller in the southern hemisphere, because of its greater expanse of ocean, and because of the role of continental air masses in the northern hemisphere. In the tropical oceans, seasonal variations mainly result from the seasonal variation of the circulation in response to wind forcing. Interannual variations can also be as large as seasonal variations. El Niño events and associated Kelvin waves are also well observed by T/P (Fall 1992 and Winter 1993, Fall 1994 and Winter 1995 and mainly Fall 1997 and Winter 1998). Several studies of tropical Kelvin and Rossby waves using T/P data have shown that they are a potentially important factor in understanding the coupled ocean/atmosphere system related to the El Niño Southern Oscillation (ENSO). Rossby waves and their propagation characteristics have also been studied in the mid-latitudes. One of the main results was to show that the propagation was faster than that of free linear waves (Chelton and Schlax, 1996). This may have major effects on the way the ocean adjusts to atmospheric forcing.

T/P also provides a unique data set for validating global ocean models and understanding their main deficiencies (Stammer et al., 1996; Fu and Smith, 1997). A detailed model/data comparison is a fundamental step before assimilation. Comparison of T/P results with global high-resolution (1/4 or 1/6 degree) simulations shows that the models still underestimate variability in mid- and high-latitude regions. The seasonal signal is qualitatively well reproduced. At high frequencies (higher than 100 $days^{-1}$), the barotropic response of the ocean to wind forcing has been characterized globally, for the first time, by comparing T/P and model results.

We have also learned a lot from altimetry about the mesoscale ocean circulation variations (e.g. Gulf Stream, Kuroshio, Antarctic Circumpolar Current, Canary Basin, Mediterranean sea). T/P alone cannot provide a good mapping of mesoscale signals: at least two satellites are needed. The contribution of merged T/P and ERS-1/2 data in providing a detailed description of mesoscale dynamics has been illustrated in, for example, the Mediterranean sea (Ayoub et al., 1998). The ability to map mesoscale signals was also very well demonstrated by the comparison of T/P and ERS-1 data with in-situ measurements in the Canary basin during the Semaphore experiment (Hernandez et al., 1995). The agreement was better than 3 cm rms, which is about the accuracy we expect from in-situ measurements. Satellite altimetry is actually the only way of precisely monitoring and possibly forecasting mesoscale signals.

3.6 REAL-TIME ASPECTS

Real-time aspects are crucial for operational oceanography. Altimeter data (if required during the mission design) can be acquired and processed in near-real time (2 days) (e.g. ERS-1/2, Jason-1, ENVISAT). However, the data are less accurate because orbit computation requires environmental parameters which cannot be obtained in real time. Dedicated processing is thus needed to correct the orbit error, along with continuous comparison of real-time products with precision delayed-time products to assess accuracy. Results recently obtained with T/P and ERS-2 near-real time data are almost as accurate as with delayed-time data thanks to dedicated processing techniques (Boone and Le Traon, 1997; Cheney et al., 1997). T/P data are now used in near-real time for El Niño prediction for which they were shown to be very useful (Cheney et al., 1997).

4. PROSPECTS

Space oceanography provides unique tools for ocean forecasting. Some techniques are very mature (altimetry, SST) and can be readily used in models. However, new space oceanography missions (e.g. for measuring salinity), and better sampling of sea level and winds are required. Operationality for space oceanography missions is also not yet guaranteed (as it is for meteorological satellites), although it is hoped that this may be achieved in the foreseeable future.

In the future, ocean forecasting, through data assimilation, will be an appropriate and powerful means to assess the impact of the satellite (and in-situ) observing systems, to identify gaps and to improve the efficiency/effectiveness of the observing system. A better use of space (and in-situ) data using effective data assimilation techniques and a better assessment of the ocean observing system components are among the main objectives of GODAE, the Global Ocean Data Assimilation Experiment (Smith and Lefebvre, 1997; Le Traon et al., 1999).

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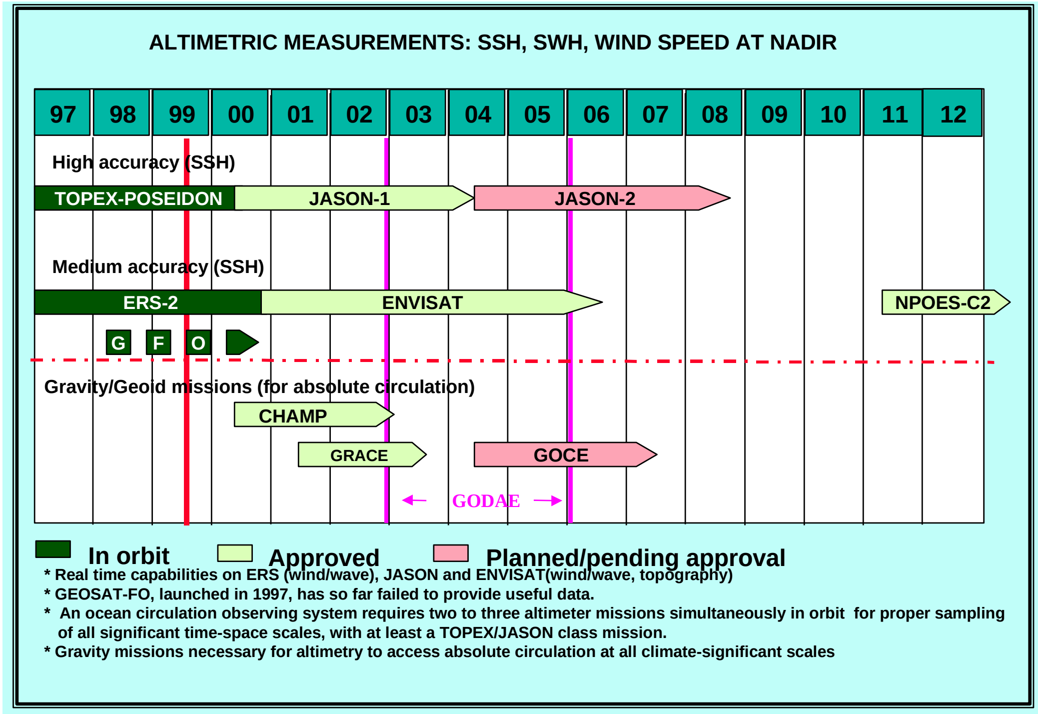

Table 1 : Schedule of altimeter and gravity missions (courtesy of A. Ratier)

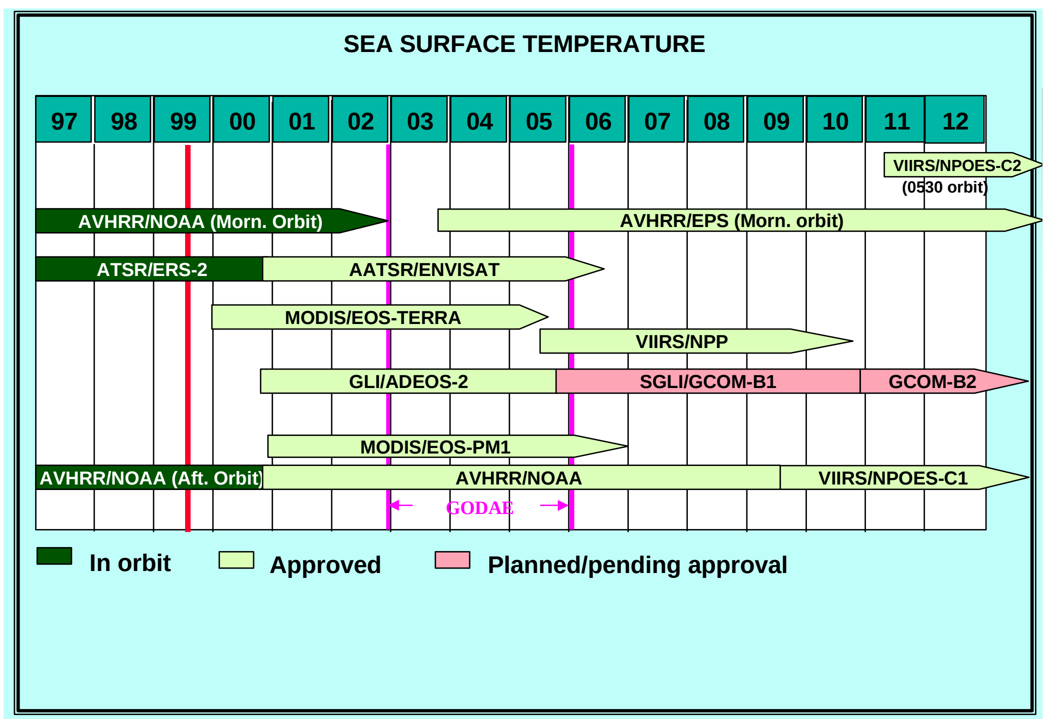

Table 2 : Schedule of SST missions (Courtesy of A. Ratier)

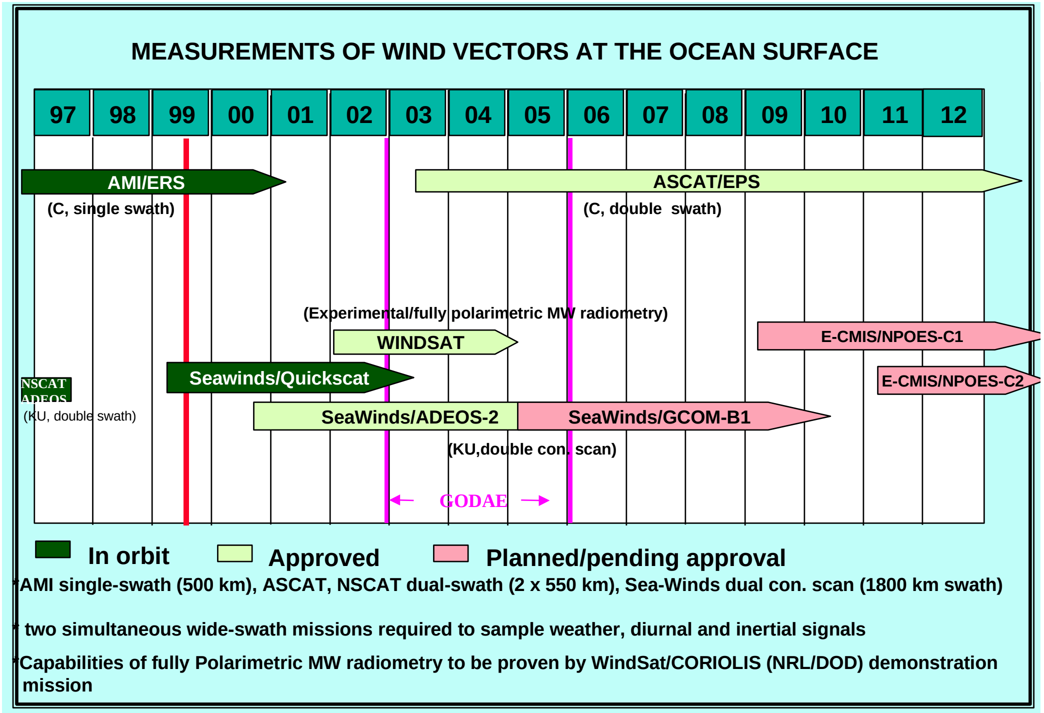

Table 3 : Schedule of scatterometry missions (courtesy of A. Ratier)

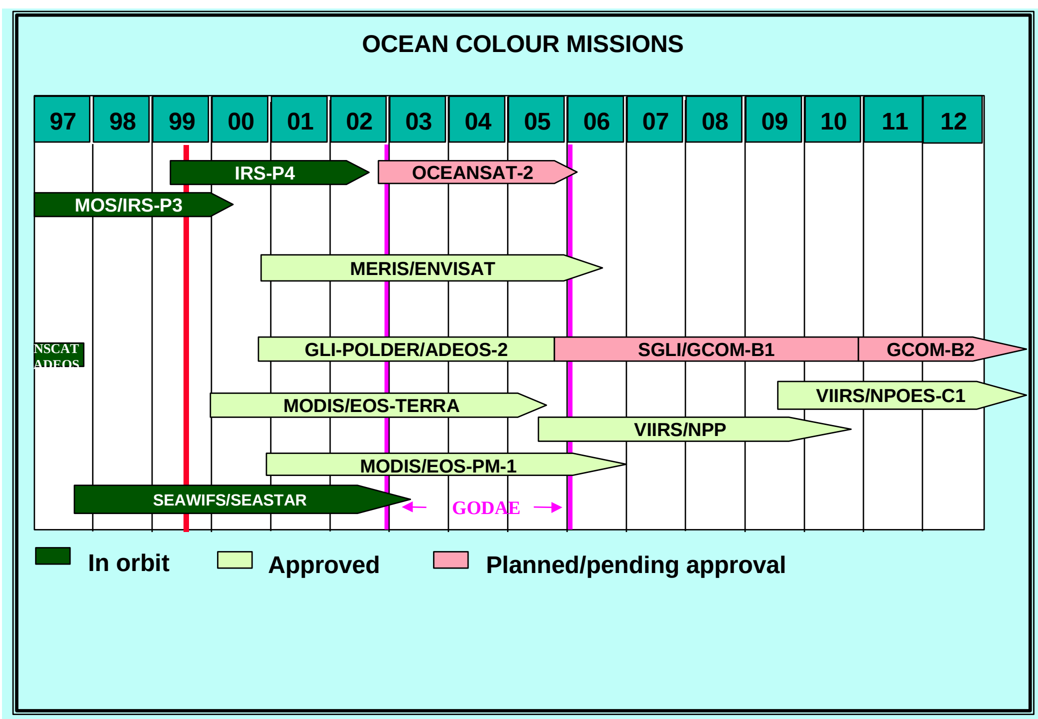

Table 4 : Schedule of ocean color missions (courtesy of A. Ratier)