Use of the 3D Radon Transform to Examine the Properties of Oceanic Rossby Waves

PETER G. CHALLENGOR, PAOLO CIPOLLINI, AND DAVID CROMWELL
James Rennell Division for Ocean Circulation and Climate, Southampton Oceanography Centre, Southampton, United Kingdom

(Manuscript received 16 November 1999, in final form 22 January 2001)

ABSTRACT

One of the most successful applications of satellite-borne radar altimeter data over the oceans in recent years has been the extraction of information about long-wavelength baroclinic Rossby (or planetary) waves, which play a significant role in ocean circulation and climate dynamics. These waves cross ocean basins from east to west at speeds of a few centimeters per second at midlatitudes. The cross-basin propagation time may therefore be several months or even years, and an accurate estimation of the speed of the waves is important. Methods are reviewed for obtaining information on Rossby wave velocity from altimetry data, particularly the two-dimensional Radon transform. Unfortunately, the use of longitude-time plots, although it allows the estimation of the zonal phase speeds, does not give any information on the velocity vector when the propagation of the waves is not purely zonal (east–west). As shown here, the two-dimensional Radon transform can be generalized to three dimensions, enabling not only the true propagation velocity component to be determined but also the direction of the waves and thus any deviation from the pure-westward case. As examples of the application of this extended technique, maps of direction, speed, and energy of Rossby waves in the North Atlantic Ocean are shown.

1. Introduction

Because of its repeat sampling and global quasi-synoptic capability, the satellite-borne radar altimeter is an excellent instrument for looking at large-scale oceanographic phenomena with surface signatures that may be difficult to detect in conventional oceanographic measurements, such as current meter moorings and hydrographic sections. One such dynamical phenomenon is long-wavelength baroclinic Rossby waves. These waves, which are a special type of planetary wave (whose existence is due to earth rotation), play a fundamental role in atmospheric and ocean dynamics. Their name derives from the work of Rossby (1940), who built on the theory of planetary waves first formulated by Hough (1897).

Oceanic Rossby waves are important for a number of reasons: they maintain the strong western boundary currents such as the Kuroshio and the Gulf Stream, they are the main oceanic response to changes in atmospheric forcing, and they are the only means (other than boundary waves) by which information is transferred from the eastern ocean boundaries to the west (e.g., Gill 1982). The speed of Rossby wave propagation is important in determining the oceanic response time to disturbances from the equilibrium state. Rossby waves can even be responsible for disturbing the position of the western boundary currents. By using a general ocean circulation model and data from the space-borne Geodetic Satellite (Geosat) altimeter, Jacobs et al. (1994) present evidence for the existence of an extratropical Rossby wave in the North Pacific, generated by the El Niño event of 1982/83. This wave has also been observed by Jacobson and Spiesberger (1998) in expendable bathythermometric data. Jacobs et al. (1994) suggest that after a decade’s delay this wave induced a shifting of the Kuroshio Current in the northwest Pacific, which may have affected the climate of North America, and which may have been responsible for the dramatic meteorological events in North America in 1993, such as the Mississippi flooding (McPhaden 1994). In short, as well as being one of the ways the ocean itself responds to climate events, Rossby waves may also be a generator of climate change and weather variability.

The first detection of Rossby waves in altimetry was achieved with Geosat data (e.g., Le Traon and Minster 1993; Tokmakian and Challenor 1993). However, tidal aliasing arising from the Geosat sampling pattern meant that any error in the M2 tidal component aliased to a signal with properties similar to the first baroclinic-mode Rossby wave (Parke et al. 1998). The European Remote Sensing (ERS) satellites and TOPEX/Poseidon (T/P) are largely free of this problem and the Rossby wave signal can be clearly seen. Chelton and Schlax (1996) showed that the speed of the Rossby waves as

Corresponding author address: David Cromwell, James Rennell Division for Ocean Circulation and Climate, Southampton Oceanography Centre, European Way, Southampton SO14 3ZH, United Kingdom.
E-mail: ddc@soc.soton.ac.uk

© 2001 American Meteorological Society
detected by T/P is faster than that predicted from linear theory using data atlases of ocean climatology. This observation has prompted a number of theoretical papers that have attempted to explain this discrepancy (Killworth et al. 1997; Qiu et al. 1997; Dewar 1998; White et al. 1998; Killworth and Blundell 1999).

All of these papers, bar one, have been concerned with the zonal (east–west) speed of Rossby waves. (The exception is Dewar 1998, who produced characteristics of long baroclinic Rossby waves not restricted to constant latitude lines.) In practice, however, Rossby waves do not always propagate solely westward: their speed may have a north–south component as well. It is expected that this will be particularly important when they interact with topography, over midocean ridges for example (Killworth and Blundell 1999). In this paper we review the methods of extracting velocities as seen through zonal sections, in particular the 2D Radon transform (RT), and present a method that handles real 3D datasets as functions of longitude, latitude, and time, in order to cope with nonzonal propagating signals. We conclude by giving an example from the North Atlantic Ocean.

2. Two-dimensional analyses

The basis for the detection of Rossby waves in altimeter measurements of sea surface height (SSH) anomaly is the Hovmöller diagram or longitude–time plot. (The SSH anomalies are normally calculated with respect to a multiyear mean SSH.) To construct such a diagram, altimeter data are interpolated to a grid and zonal sections are “stacked” to give a two-dimensional plot whose x axis is longitude and y axis is time. Westward propagating signals can clearly be seen in such diagrams (not shown) as areas of local maxima and minima moving linearly from right to left across the diagram. The speed of these signals can easily be calculated by hand by simply taking the slope of the diagonal alignments of maxima and minima. While such a procedure is easy to carry out on a few diagrams, a more automated and objective method is required to calculate propagation speeds for entire ocean basins.

A number of two-dimensional data processing techniques can be applied to the longitude–time diagrams in order to study the properties of zonally propagating signals. For example, the two-dimensional Fourier transform (FT) yields the different spectral components of longitude–time plots, which appear as peaks in the frequency–wavenumber domain, and has been used for a number of years for the detection of Rossby waves (e.g., Le Traon and Minster 1993; Tokmakian and Challenor 1993). One advantage of this technique is that it allows for the detection of single components of the propagating signals, which may correspond to different baroclinic modes as suggested by Cipollini et al. (1997). Distinct baroclinic modes of propagation are also observed by Subrahmanyam et al. (2001), who applied the same technique to longitude–time plots of T/P SSH anomalies in the tropical Indian Ocean.

An alternative method is to mathematically define the technique used to measure speed by hand. Such a signal processing method is the RT first described by Radon (1917).
plot and $x, y$ are longitude and time, respectively.] If $L$ is an arbitrary line at an angle $\theta$ with respect to the $x$ axis, RT (Deans 1983) is defined as the projection of $f(x)$ on $L$; that is,

$$p(s, \theta) = \int_x f(x) du,$$

where $u$ is the direction orthogonal to $L$ (along which the integral summation is performed) and $s$ is the coordinate on $L$. Note that for a given $\theta$, the RT is a 1D function of the line coordinate $s$. We can rewrite (1) in terms of coordinates $x$ and $y$, as shown in Fig. 1:

$$p(x', \theta) = \int_y f(x, y)|_{x=x' \cos \theta - y' \sin \theta} dy'. (2)$$

The RT has many varied applications—for example, in medical imaging, in the processing of seismic data to find subsurface rock interfaces, and in the detection of ship wakes in satellite images. In satellite remote sensing of the oceans, Chelton and Schlax (1996) applied the RT to their global analysis of Rossby waves observed in T/P data, as did Polito and Cornillon (1997) in their intensive study of Rossby waves in the North Atlantic. Cipollini et al. (1997, 1999) applied it to T/P data in the North Atlantic and found strong Rossby wave propagation near the Azores Current; while Hill et al. (2000) used the method to detect Rossby waves in global sea surface temperature data from the Along-Track Scanning Radiometer on board ERS-1.

In all these applications, the essential method is to transform two-dimensional images with lines into a domain of possible line parameters (e.g., slope, which equates to speed in a longitude–time plot), whereby each line in the image will yield a peak at the corresponding line parameter in the transform domain. When the focus of the analysis is on the propagation speed, rather than on the wavelength and period characteristics of the single components, the RT is more appropriate than the FT.

As with any transform, the selection of the longitude–time subdomain on which it is applied requires a tradeoff between resolution and homogeneity. In the FT case the longitude–time space is transformed into a wave-number–frequency space so any periodic signal with a constant wavelength and period will correspond to a distinct peak in that space. If the selected subdomain is too narrow, adjacent peaks will be indistinguishable from each other. On the other hand, if the interval is too large, so that the propagation characteristics vary significantly, or if the waves are not periodic in the first place, the interpretation of the transform becomes difficult. A consequence of this is that the FT is not suited for the study of single waves, such as the soliton observed by Jacobs et al. (1994) in the Pacific as a result of the 1982/83 El Niño.

As with the FT, the RT needs homogeneity to show...
but no information on the true velocity vector of the waves. Figure 2 depicts a wave crest travelling at a direction $\varphi$ relative to pure westward, at two subsequent instants, $t$ and $t + \Delta t$. In the time $\Delta t$, the crest travels a space vector $\Delta s_{\text{TRUE}}$, so the velocity vector is $v_{\text{TRUE}} = \Delta s_{\text{TRUE}}/\Delta t$. The space traveled by the crest as seen through a zonal section is $\Delta s_x = |\Delta s_{\text{TRUE}}|/\cos \varphi \geq |\Delta s_{\text{TRUE}}|$, and the corresponding zonal speed is $v_x = \Delta s_x/\Delta t = |v_{\text{TRUE}}|/\cos \varphi \geq |v_{\text{TRUE}}|$. It is important that we remove the inherent limitation of longitude–time plots because, as mentioned above, Rossby wave speed (not only the zonal phase speed but also the magnitude of the velocity vector) is an important parameter related to the response time of the oceans to forcing. Moreover, possible deviations of the direction from the pure westward propagation predicted by simple theory have been suggested by recent theoretical work such as Killworth and Blundell (1999), who studied the effects of varying bottom topography on Rossby wave propagation. Observations of the deviation of speed from the pure-westward case are thus needed. To study the directional properties of Rossby waves and estimate the magnitude of their velocity vector we need to look at cuboids of data, not just longitude–time slices. Consequently, we have to use three-dimensional techniques such as the 3D FT and the 3D RT. In the present paper we deal with the latter, which, as we showed in the 2D case above, is more appropriate for evaluating the magnitude and direction of velocities.

The 3D RT is defined in the same way as in Eq. (1), except that $x$ is now a vector in 3D and the integral is a projection on a plane, whose orientation is identified by a pair of angles $(\theta, \varphi)$, rather than on a line (see Fig. 3):

$$p(x', y', \theta, \varphi) = \int f(x, y, z) \, dz'. \quad (3)$$

The formal details are in Deans (1983). The transformation of coordinates is expressed by

$$x' = x \cos \varphi \cos \theta + y \sin \varphi \cos \theta + z \sin \theta,$$

$$y' = -x \sin \varphi + y \cos \varphi,$$

$$z' = -x \cos \varphi \sin \theta - y \sin \varphi \sin \theta + z \cos \theta. \quad (4)$$

The above equations can be easily obtained by considering the transformation of coordinates as the result of two consecutive rotations, the first one of an azimuth $\varphi$ around the $z$ axis and the second one of an elevation $\theta$ around the $y'$ axis, as shown in Fig. 3.

When applying this concept to our dataset of gridded satellite data, rather than generate two-dimensional plots with longitude and time axes, we will now work with a cube of data, retaining the latitude dimension. Our objective is to find linear alignments of crests and troughs within the cube—the signature of possible Rossby waves. This can be done in our 3D space by projecting the data onto a $(x', y')$ plane oriented at $(\theta, \varphi)$ and summing the projected data into 2D bins (see Fig. 3). Doing so we have a discretization of the 3D RT. We then look for the pair of values $\theta_{\text{max}}$ and $\varphi_{\text{max}}$, which maximize the energy of the projected sum. These identify the plane orthogonal to the main direction of the alignments, that is, the main direction $z'$ of propagation of crests and troughs in the cuboid. Elevation $\theta$ can then be converted into speed as for the 2D case, while azimuth $\varphi$ yields the geographical direction of propagation.

In practice, as seen for the 2D transform, a trade-off has to be made between the assumption of homogeneity and spatial resolution. So the 3D analysis is carried out, not on the complete cuboid of data, but on successive “subdomains” defined by a sliding window in latitude–longitude space, for example, $5^\circ \times 11^\circ$. Hence, the Rossby wave propagation characteristics that are determined are the averaged properties in a spatial window of this size. In the present paper, we include the complete time series in each spatial window. In principle, however, one could choose to apply the 3D RT to subsets of the complete time interval spanned by the data and thereby investigate temporal changes in Rossby wave propagation characteristics, although it must be recalled that shortening the time interval also lowers the resolution.

A detailed study of the resolution of the RT will be the object of a subsequent study.

4. Application of the 3D radon transform

a. Processing of TOPEX/Poseidon data

The T/P satellite, launched in August 1992, carries two altimeters (TOPEX and Poseidon; see Fu et al. 1994, for details). However, because they have been carefully intercalibrated we may treat them as one al-
timetric dataset. The satellite lays down a ground track of passes 2.7° apart in longitude, repeated every 9.92 days. We applied a standard set of corrections for orbit errors, atmospheric delays, tides, and sea state effects (see Cipollini et al. 1997). A mean height profile is calculated for the three calendar years (1993–95) and SSH anomalies are computed relative to that. The accuracy of the SSH retrieval with T/P is of the order of 2 cm, as found by Cheney et al. (1994) by comparison with tide gauges. Each cycle of data is then interpolated onto a 1° × 1° grid. The interpolation, which uses a Gaussian-weighted mean of all the data within 200 km of a grid point, reduces the instrument and correction errors while leaving the larger-scale signal relatively unaffected. Chelton and Schlax (1996) suggest that signals with an amplitude of ~1 cm may be observable. Two hundred thirty-four 10-day T/P cycles were used, spanning October 1992–February 1999.

b. 3D RT of T/P data in the North Atlantic

We used the discretized form of the 3D RT described above and applied it to the processed, gridded T/P data volume in the North Atlantic. Tests were undertaken with a variety of spatial windows: 5° lat × 5° long, 5° lat × 9° long, 9° lat × 9° long, and 5° lat × 11° long. This last choice of 5° lat × 11° long was found to be a reasonable compromise between smoothing of signal properties and noise reduction to enable propagating signal detection. Limiting the latitude range to 5° is a reasonable choice (e.g., Polito and Cornillion 1997) to suppress excessive smoothing of the latitudinally dependent Rossby wave speed. Note that a Fourier analysis method would require two, or better, three samples per wavelength to satisfy the Nyquist criterion, and that a window length of 11° in longitude would be insufficient to resolve the Rossby waves in low latitudes where wavelengths can reach 2000 km or more. The RT suffers a similar limitation at low latitudes. A way to overcome this limitation would be the use of different spatial windows (with longer longitude spans closer to the equator, which obviously means reducing the resolution in the location of the waves), but here, for the sake of simplicity, we restrict our analysis to a single window.

The analysis presented here was therefore applied to a 5° lat × 11° long × 234-cycle subdomain, which in turn was stepped across the whole dataset in 2° × 2° intervals. Figure 4 shows a polar plot of the RT energy centered at 34°N, 36°W, revealing strong quasi-westward propagation of energy, with a maximum at a deviation of 10° to the south relative to a pure-westward propagation. The speed corresponding to the maximum energy is ~2.6 cm s⁻¹, consistent with the results of Cipollini et al. (1997), who found exactly the same value from a 2D analysis of T/P data over the 38°–8°W longitude span at 34°N.

The 3D RT over the whole basin was performed using a complete azimuth range of 0°–360° in increments of 5°. The elevation range in the 3D RT calculation was 0°–80° in steps of 1°. This comfortably covers the range of possible Rossby wave speeds over the latitude range presented below (16°–54°N).

After the 3D RT energy was calculated for the North Atlantic basin, we applied various screening of the results in azimuth, speed, and energy to highlight propagating Rossby waves. We selected only those speeds within an interval in velocity space, thus suppressing too-fast and too-slow propagating signals, including the stationary and near-stationary ones related to seasonal heating and cooling. This screening was based on a priori knowledge of propagating baroclinic Rossby waves, namely, the known latitudinal dependence of propagation speed. Chelton and Schlax (1996) demonstrated that observed zonal Rossby wave speeds exceed predicted speeds from the linear theory, often by a factor of 2 or more, particularly at midlatitudes. We then selected only those energy maxima whose speed was within 0.5 and 3 times the expected (linear theory) first-mode baroclinic Rossby wave speed at that latitude. We believe that this interval includes the vast majority of signals that are genuine Rossby waves. To highlight westward or quasi-westward propagating waves, the velocity screening was combined with screening of results in terms of azimuth. Again, a conservative approach was adopted, choosing 230° < φ < 310°, that is, only those signals with a propagation direction within ±40° from pure westward. A further screening was done on the energy of the observed maxima, retaining only those above a threshold. The threshold was selected after Monte Carlo trials with fields of pure Gaussian noise with an rms value of 2 cm, and ensures that the peaks observed are significant above the 95% confidence level. Different types of screening may be required for different datasets, for example, sea surface temperature, which we do not consider in this paper but clearly shows the signature of Rossby waves (Hill et al. 2000).

The results are presented in Fig. 5. Empty 2° × 2° grid cells correspond to locations where no maxima in the 3D RT passed the screening described above. Visual cross-checking with polar plots of the type shown in Fig. 4 confirmed that there were no distinct westward or quasi-westward propagating signals at such locations. Figure 5a shows the direction of the resulting signals. As can be seen from the figure, most energy at midlatitudes actually propagates within 10° of due west. Figure 5b shows the speed of the waves corresponding to the signals selected in Fig. 5a. The energy of these waves (in arbitrary units) is shown in Fig. 5c. It is worth noting that the results are robust in that they are in agreement with westward-propagating features observed in Hovmöller plots at a number of selected locations (not shown here). Figure 6 shows a comparison between zonal averages of speeds over the North Atlantic basin, including the zonal speed derived from our results (Fig. 5b divided by the cosine of the deviation in Fig. 5a), the zonal phase speed predicted from the linear theory,
and the zonal phase speed predicted from the extended theory by Killworth et al. (1997). The same figure also includes the three propagation speeds from T/P data measured in the extratropical North Atlantic by Chelton and Schlax (1996).

5. Discussion

In the plot of the direction of propagation of the waves (Fig. 5a), we observe a wide midlatitude zonal band (approximately 20°–40°N) where the 3D method retrieves a quasi-westward signal (within ±10° from pure westward). In this band, the topography has no clear effect, possibly because the longitudinal extent (11°) of the 3D RT window is greater than the length scale of potentially important topographic features, for example, the width of the mid-Atlantic ridge.

The propagation speed of the waves (Fig. 5b) increases equatorward in accordance with the theory. We also note at midlatitudes (around 34°–36°N) a distinct increase in speed as the waves move to the western part of the basin, consistent with previous results (Chelton and Schlax 1996; Cipollini et al. 1996). There are no clear effects of topography on speed, again because they are probably averaged out within the 3D RT window. It is known, however, that topography can have important local effects on Rossby-wave velocity (e.g., Killworth and Blundell 1999).

From Fig. 5c we note high energy in the region of the Gulf Stream. While this region is obviously very energetic at a number of different scales, it has to be stressed that the energy plotted here is only that associated with the westward propagating signals passing the screening described in the previous section, so it is most likely to be associated with Rossby waves. We also observe that in the eastern North Atlantic there is a zonal band of enhanced Rossby wave energy centered on 34°N. This zonal waveguide of stronger Rossby wave propagation just south of the Azores Current is consistent with previous T/P observations analyzed using stan-
Fig. 5. Direction of (a) propagation, (b) velocity magnitude $|v_{\text{true}}|$, and (c) energy of the strongest energy maximum in a 3D RT analysis of gridded TOPEX/Poseidon data. The signal at each location has been selected after the screening described in section 4, that is, from those propagating within $\pm 40^\circ$ of pure west (i.e., within the azimuthal range $230^\circ \leq \varphi \leq 310^\circ$) with speeds between 0.5 and 3 times the speed predicted at each latitude by the linear theory, and energy above a confidence threshold. The isobaths are for depths of 2000 m and 3500 m.
Our results for the basin-wide averages of zonal speed at different latitudes, plotted in Fig. 6, show that the waves tend to go significantly faster than predicted by the linear theory at latitudes between 27° and 45°. In the same latitudinal band the results are in broad agreement with the zonal mean for the North Atlantic from the extended theory by Killworth et al. (1997), although the latter shows greater variability with latitude (see Eq. (12.3.16) in Gill 1982). Moreover, the propagation speed decreases, the effects of the mean barotropic flow on the speed and direction of the waves become more evident; in other words, some of the scatter might be due to noticeable advection of very slow-moving disturbances by the mean barotropic flow at the depth of the thermocline.

The observed results for the basin-wide averages of zonal speed at different latitudes, plotted in Fig. 6, show that the waves tend to go significantly faster than predicted by the linear theory at latitudes between 27° and 45°. In the same latitudinal band the results are in broad agreement with the zonal mean for the North Atlantic from the extended theory by Killworth et al. (1997), although the latter shows greater variability with latitude (the meridional variability in our analysis is obviously smoothed by the use of a 5° window in latitude, as discussed in section 4b). The speeds measured by Chelton and Schlax (1996) at 29.73° and 32.37° are significantly greater than both our results and those by Killworth et al. (1997); however, the three speeds (Chelton and Schlax’s, Killworth et al.’s, and this study) almost coincide (at ~1.5 times the speed expected from the linear theory) in the most energetic waveguide around 34°N.

At latitudes equatorward of 25°N, the 3D RT method picks up waves slower than the theory (except at 16°N, where it detects a fast signal close to the western boundary). Interestingly, Killworth et al.’s (1997) phase speeds are also lower than the linear theory predictions in the latitudinal band 20°–22°N. However, our estimates are still approximately 15%–20% higher than Killworth et al.’s in the 20°–24°N interval. One possible explanation of this discrepancy could be that at those latitudes the transform is highlighting higher-order, slower baroclinic modes of propagation, but the behavior of the 3D RT at low latitudes and any issues related to its resolution will obviously have to be investigated further.

Before going into too much detail on the particular oceanographic results, it has to be recalled that the main purpose of this paper was to describe the 3D RT methodology used and to show (with the North Atlantic example) that it allows a study of the directional properties of propagating features like baroclinic Rossby waves. We believe that this has now been established. We are now planning to apply the 3D RT analysis to the global ocean and to study quantitatively the correlation between wave properties and topography. We hope that such observational results will continue to stimulate the development of theoretical analyses of Rossby wave propagation.

Acknowledgments. We thank NASA and AVISO for the provision of TOPEX/Poseidon altimeter data; and Jeff Blundell, Katy Hill, Peter Killworth, Graham Quartly, Ian Robinson, and Stefano Raffaglio for their kind help and suggestions. We are grateful to Helen Snaith for support of the altimeter processing software.

REFERENCES
Hill, K., I. S. Robinson, and P. Cipollini, 2000: Propagation char-
characteristics of extratropical planetary waves observed in the
105, 21 927–21 945.
Hough, S. S., 1897: On the application of harmonic analysis to the
dynamical theory of the tides. Part I: On Laplace’s ‘oscillations
of the first species’, and on the dynamics of ocean currents.
Mitchell, W. J. Teague, and A. J. Wallcraft, 1994: Decade-scale
trans-Pacific propagation and warming effects of an El Niño
Niño–Southern Oscillation induced Rossby waves in the north-
24 596.
Killworth, P. D., and J. R. Blundell, 1999: The effect of bottom
topography on the speed of long extratropical planetary waves.
———, D. B. Chelton, and R. de Szoeke, 1997: The speed of observed
and theoretical long extratropical planetary waves. J. Phys.
Le Traon, P.-Y., and J.-F. Minster, 1993: Sea level variability and
semiannual Rossby waves in the south Atlantic subtropical gyre.
327.
Parke, M. E., G. Born, R. Leben, C. McLaughlin, and C. Tierney,
Polito, P. S., and P. Cornillon, 1997: Long baroclinic Rossby waves
detected by TOPEX/POSEIDON. J. Geophys. Res., 102, 3215–
3235.
Qiu, B., W. Miao, and P. Müller, 1997: Propagation and delay of
forced and free baroclinic Rossby waves in off-equatorial oceans.
Radon, J., 1917: Über die Bestimmung von Funktionen durch ihre
Kl., 69, 262–267. [For English translation, see Deans (1983),
204–217.]
Rossby, C. G., 1940: Planetary flow patterns in the atmosphere.
Subrahmanyam, B., I. S. Robinson, J. R. Blundell, and P. G. Challenor,
2001: Indian Ocean Rossby waves observed in TOPEX/PO-
SEIDON altimeter data and in model simulations. Int. J. Remote
Tokmakian, R. T., and P. G. Challenor, 1993: Observations in the
Canary Basin and the Azores Frontal Region using Geosat data.
White, W. B., Y. Chao, C.-K. Tai, 1998: Coupling of biennial oceanic
Rossby waves with the overlying atmosphere in the Pacific basin.