Remote and Local Forcing in the Brazil±Malvinas Region

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(Manuscript received 23 August 1999, in final form 14 June 2000)

ABSTRACT

Origins of the seasonal variability observed in current meter data from the Malvinas (Falkland) Current are sought in the wind field on both a regional and circumpolar scale. A singular value decomposition of the covariance of the fields makes it possible to distinguish between a local and a remote source of variability. The local mode is the result of changes in the wind stress curl around 40°S causing an annual modulation transverse to the mean path of the current, and thus contributes little to the variability in transport. It seems likely that the wind stress curl drives the annual excursions of the Brazil±Malvinas Front and forces the retroreflection of the Malvinas Current. The remote mode features increased transport trailing negative wind stress curl anomalies at Drake Passage with a lag of 20–30 days. Repeating this analysis with the winds on a circumpolar domain suggests that this might be the regional manifestation of a more global feature, associating globally negative wind stress curl anomalies north of 60°S with increased transport. This analysis also shows that the Malvinas Current transport is anticorrelated with the zonal wind stress, hence with the Antarctic Circumpolar Current. Although initially counterintuitive, this is confirmed through coherence analyses with bottom pressure measurements at the south of Drake Passage. These results are further assessed using a 5-yr series of transport in the Malvinas Current derived from TOPEX/Poseidon altimeter data, which is compared with the circumpolar averages of both zonal wind stress and wind stress curl. Coherences are largest with the wind stress curl at latitudes north of Drake Passage (up to the subtropics), and also indicate that it is most influential in the Pacific sector. It is suggested that the Antarctic Circumpolar Current and the Malvinas Current respond differently to the wind forcing, and that these two modes cohabit at Drake Passage. While the barotropic variations of the former have been shown to respond to the zonally averaged wind stress, the fluctuations of the latter at timescales of 100–200 days are more Sverdrupian, that is, sensitive to the wind stress curl (lag ~20 days), explaining why they fluctuate with opposite phase. Superimposed on this, intraseasonal variability at shorter periods (~70 days) results from baroclinic shelf waves trapped along the edge of the Patagonian plateau. These propagate from Drake Passage, and may originate from equatorial Kelvin waves in the Pacific.

1. Introduction

As part of the World Ocean Circulation Experiment (WOCE) Confluence project, current meter measurements were gathered between December 1993 and June 1995 in the Malvinas Current (hereafter MC), near its merger with the Brazil Current (Fig. 1). These showed the mean flow to be equivalent barotropic in form (see, e.g., Killworth 1992, for a definition), whereas the variability is dominated by a surface intensified barotropic-like empirical mode, the structure of which is suggestive of mode coupling caused by the steep topography (Vivier and Provost 1999a, hereafter VP99a). Transport variability was found to be ~12 Sv (Sv = 10⁶ m³ s⁻¹) root mean squared, a significant part of which is due to mesoscale activity, reducing to about half this for timescales beyond 2 months. Transport estimates were extended to 3 yr using TOPEX/Poseidon (T/P) data (Vivier and Provost 1999b, hereafter VP99b), and showed substantial energy near 70 and 180 days and, by contrast, a weak annual harmonic. In this contribution, we investigate the possible sources of variability with periods around one year and shorter. Processes considered include the strength of the Antarctic Circumpolar Current (ACC), wind forcing on a regional scale [down to Drake Passage (DP)], wind forcing on a circumpolar scale, and fluctuations of the Brazil Current. As the current meter


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array was located close to the Brazil–Malvinas Confluence, the latter of these may indirectly affect the velocity records through its action on the subtropical front, which is known to migrate meridionally with a substantial seasonal component (Olson et al. 1988; Garzoli and Garraffo 1989).

Several previous studies discuss the origins of the variability of the MC and/or the location of the subtropical front. From a barotropic model, Matano et al. (1993) suggested that the annual migrations of the subtropical front occur in response to the wind-driven fluctuations in the transport of the Brazil Current as well as possibly the MC, although the contribution of the latter is not clear (VP99b). Analyzing inverted echo sounder data in the frontal area, Garzoli and Giulivi (1994) concluded that, besides the seasonal variations, the marked interannual variability of the location of the front was related to anomalous wind patterns south of the confluence. Moreover, they found no apparent correlation between circumpolar-averaged wind stress at the latitude range of DP (parameterizing wind-forced pulses in the ACC) and the observed anomalous northward penetration of the MC. In contrast, Smith et al. (1994), from an isopycnal model of the South Atlantic, found high correlations at the semiannual period between DP transport, MC transport (with a lag of 90 days), and the seasonal movements of the Brazil–Malvinas Confluence, while the Brazil Current transport exhibited significant energy only at the annual period. They further suggested that this remote forcing of the confluence region at the semiannual period was mediated from DP by a shelf wave (Kelvin-type wave form) propagating cyclonically around the basin.

In this study, we specifically focus on the winds as a potential forcing agent of the MC. For seasonal timescales, it has long been known that the large-scale response of the ocean poleward of 30° latitude is primarily barotropic (Gill and Niiler 1973), although baroclinic processes may be locally important. Specifically for the ACC, support for this view stems both from theoretical (e.g., Clarke 1982) and observational bases (Peterson 1988; Meredith et al. 1996). The adjustment occurs via
fast-moving barotropic Rossby waves for high-frequency forcing, and at longer scales (≥1 month) a Sverdrup balance is obtained in closed basins (Willebrand et al. 1980; Cummins 1991). At high latitudes, topography significantly affects the background potential vorticity gradient, and the adjustment to a topographic Sverdrup balance is mediated along \( f/H \) contours, where \( f \) is the Coriolis parameter and \( H \) the ocean depth. Hughes et al. (1999) considered the effect of topography on the variability in circumpolar transport. They showed that the wind-forced barotropic transport variability will be much more influenced by bottom topography than the the mean (mainly baroclinic) ACC flow and consequently that the transport fluctuations will occur along different paths to the mean ACC transport. A topographic Sverdrup balance constrains the flow across \( f/H \) contours, but does not dictate anything about the flow along these contours, which is usually prescribed by mass continuity. However, in situations where \( f/H \) contours close on themselves (as for a periodic channel or around an island), the flow primarily responds to the wind stress integrated along such a contour, in a balance involving acceleration and friction. Such a situation was shown to occur for the southern portion of the ACC, where the Antarctic continent is nearly surrounded by closed \( f/H \) contours (Hughes et al. 1999; Gille et al. 2000).

After presenting the data (section 2), we use canonical correlation techniques to establish links between wind forcing and variability of the MC (section 3). More specifically, we analyze the cross-covariance between velocity and regional wind stress curl (WSC) fields at different lags using a singular value decomposition (SVD). Analyses are repeated both with wind stress and WSC on a circumpolar domain in section 4. As will be shown, the direction of the velocity field in the MC for the coupled modes permits a relatively clear distinction between the various sources of variability involved. However, a major drawback of these analyses is the brevity of the current meter records. Thus, to further assess our results, we also use a 5-yr series of MC transport estimates obtained from T/P altimeter data along ground track 26 (Fig. 1). Coherences with various circumpolar measures of the wind field are examined (section 5), as well as with bottom pressure recorder (BPR) data at DP (Fig. 1), whereby we search for evidence of a remote forcing of the MC. Finally, section 6 synthesizes the main results of this study.

2. Data

a. Current meter data

Current meter data used here are from four moorings (M1, M5, M6, and M7) featuring a total of 17 instruments on a section across the MC at about 40°S (Fig. 1; Table 1). The data have been low-pass filtered with a cutoff period of 50 h (whereby tidal and inertial variability has been removed) and subsampled at a daily rate. The mean velocity is largest at westernmost moorings (M1 and M5; reaching 35 cm s\(^{-1}\) at M5) where the flow is steered by topography. At the eastern edge of the array (M6 and M7) it is smaller, with a significant cross-isobath component (Fig. 2).

An overview of the variability of the velocity field is efficiently provided by a principal component analysis (PCA). The PCA is here performed on unnormalized velocity anomalies considering simultaneously each current meter shallower than 2000 m across the section, irrespective of the specific mooring. Although least squares techniques exist for PCA for data containing gaps (e.g., von Storch and Zwiers 1999), these may yield less robust results when the gaps are not scattered throughout the time series, as is the case here with record length varying from 20 days to over 500 days, depend-

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**Table 1. Location of the instruments; M1, M5, M6, and M7 are current meter moorings from LODYC, and ND2 and SD2 are BPRs from POL. The end date for current meter moorings is the end of the longest record at each mooring.**

<table>
<thead>
<tr>
<th>Lat (°S)</th>
<th>Long (°W)</th>
<th>Depth (m)</th>
<th>Start</th>
<th>End</th>
</tr>
</thead>
<tbody>
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<td>450</td>
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<tr>
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<td>1507</td>
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<tr>
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<td>55.35</td>
<td>2200</td>
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<td>10 Jun 1995</td>
</tr>
<tr>
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<td>55.27</td>
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</tr>
<tr>
<td>SD2 60.85</td>
<td>54.71</td>
<td>1020</td>
<td>13 Nov 1992</td>
<td>continuing</td>
</tr>
</tbody>
</table>

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**Fig. 2. Mean velocity in the Malvinas Current and mean wind stress curl. Velocity data have been averaged over the maximum duration of each record (254, 386, 512, and 512 days, respectively). Only the value for the shallowest instrument at each mooring (approximately 300 m) is displayed. The two stars mark the lateral extent of the current meter array (moorings M1 and M7). WSC data have been averaged for the period 1992–96 (units are 10\(^{-7}\) N m\(^{-2}\)). Areas of negative curl are shaded. The 1000-m isobath, marking the path of the MC, is indicated (dark gray line).**

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Fig. 3. Leading EOFs of the velocity field in the MC, computed with all instruments shallower than 2000 m having a record length greater than 386 days (○). Unused instruments are also indicated (+). Arrows represent the horizontal velocity, folded in the vertical plane (north is to the top of the figure). The bold line in the lower left corner represents the average direction of the isobaths (i.e., the direction of the mean flow). Vertical scale in meters; horizontal scale in kilometers.

Fig. 4. Expansion coefficients (principal components) of the leading two EOFs of the velocity in the MC, shown in Fig. 3.

are coherent throughout the section above 2000 m and exhibit a mode of variability transverse to the mean flow (39.7% of the variance) and a mode oriented along the mean flow (33%). These are both surface intensified (Fig. 3), and are consistent with an analysis performed for each mooring separately (VP99a). Both modes are also intensified at the easternmost mooring (M7) where the variability of the flow is largest; this reflects both broadening and shrinking of the current core, and the influence of the variations in the location of the neighboring subantarctic front. Expansion coefficients (or principal components) associated with these two modes are shown in Fig. 4. Mode 1 has an energy peak at about 30–40 days (spectra are not shown but are comparable to those shown in VP99a), but its modulation is mostly annual (47% of the variance) and marginally semiannual (9%). VP99a suggested that this cross-isobath mode was related to annual migrations of the Subantarctic Front, rather than being a natural cycle of the MC. Support for this view is presented in section 3a. Note the sharp anomaly in September 1994 (Fig. 4a), due to an intrusion of a meander of the Brazil Current (VP99a). Mode 2 has the bulk of its energy near 65 days, plus energy at lower frequencies that neither an annual nor a semiannual cycle can properly resolve (13.2% and 1.4%, respectively). However, most of its energy is nearer to the semiannual period: a harmonic function with a period of 140 days accounts for 25% of the variance. Most of the following analyses are performed with time series low-pass filtered at 20 days. This barely changes the structure of the leading EOFs, except for the associated fraction of variance (51.2% and 34.5%, respectively).
b. Wind products

Wind stress fields used in the following analyses are daily-averaged operational forecasts from the European Centre for Medium-Range Weather Forecasts (ECMWF) spectral model. The curl field is computed on a $1^\circ \times 1^\circ$ grid, determining areal averages through Stokes theorem, with grid points closer than $1.5^\circ$ from land being discarded. Analyses are first performed on a regional scale with a domain ranging from $75^\circ$ to $40^\circ$W and from $35^\circ$ to $65^\circ$S, thereby covering the southwest Atlantic and including DP (Fig. 2). The wind fields have been low-pass filtered with a 20-day cutoff period and subsampled at a time rate of 2 days, unless otherwise specified. The 4-yr-averaged (Oct 1992–Oct 1996) WSC field is shown in Fig. 2. A second series of analyses is performed with the wind field on a circumpolar domain covering $30^\circ$–$70^\circ$S.

c. BPRs

As part of WOCE, the ACC transport variability has been monitored at DP using BPRs, and there are several previous studies devoted to the analysis of this dataset (Meredith et al. 1996, 1997; Woodworth et al. 1996; Hughes et al. 1999; Gille et al. 2000). Here, we examine data from two sites only (ND2 and SD2) at either edge of DP (Fig. 1; Table 1), spanning the period November 1992–December 1997. This 5-yr record is obtained by concatenating four shorter duration records, with a typical gap of a few days between each. To obtain continuous series, we first centered each subset about its mean and then used the iterative 30-day low-pass gaussian filtering procedure for irregularly sampled series described by Park and Gambéroni (1995). The concatenated series are therefore cleared from the effect of monthly and fortnightly lunar tides. The gap at site ND2 between the 1994–96 record and the 1996–97 record is $\sim$20 days: interpolated values around this date (late Nov 1996) are therefore regarded as potentially suspect.

3. Variability of the coupled fields

The two main EOFs of the velocity field in the MC are nearly orthogonal relative to the topography (Fig. 3), and feature different spectral contents. This suggests that the fraction of variability represented by each of these modes is of different origin: the mode along the mean flow could be of remote origin and therefore correlated with wind patterns at a remote location with some temporal lag. In this section, possible connections between wind forcing and velocity measurements in the MC are investigated using SVD of the cross-covariance matrix of the two fields (Bretherton et al. 1992; Wallace et al. 1992). The essence of this method is to extract the linear combinations in each field that have maximum covariance. For comparison, analyses were repeated using canonical correlation analysis in a variant form proposed by Barnett and Preisendorfer (1987). Similar coupled modes were obtained and the conclusions remained identical; thus only SVD analyses are discussed hereafter.

As the fields have different physical dimensions, the SVD is performed on normalized time series (unit variance). Singular values associated with each mode indicate that the squared covariance (SC) between the fields and singular vectors are the corresponding spatial patterns, from which we derive time series of expansion coefficients by projection of the original fields on each mode. The correlation coefficient between expansion series for each mode is indicative of the strength of the coupling, for which we provide the 95% significance level based on the actual number of degrees of freedom, estimated from the decorrelation timescale of the series, as per Kelly et al. (1996).

The correlation coefficient between expansion series for a given mode is high and almost always significant (see below and, e.g., Fig. 5b). This reflects that the more spatial degrees of freedom are associated to each field, the easier it is to construct linear combinations of the fields that are correlated in time (Wallace et al. 1992). Therefore, to test the robustness of the SVD results and to assess genuine coupling between the fields, we rely instead on Monte Carlo simulations. In this approach, we compare the SC associated with the leading SVD modes to those obtained from a large ensemble of random temporal permutations of the fields (Wallace et al. 1992; Peng and Fyfe 1996; Venegas et al. 1996). In practice we perform 100 random permutations of time indices for the WSC series, and compute the SVD of the cross-covariance matrices between these scrambled
wind series and original MC velocity series. The SC of the leading covarying modes of these 100 drawings is compared to that of unscrambled series. If, for a given covarying mode, the unscrambled SVD yields an SC that exceeds that of the corresponding mode for 99 drawings, then we conclude that the mode is significant at the 99% confidence level. However, as noted by Wallace et al. (1992), a simple random permutation of the time index of each series yields a smaller SC in the SVD analysis, since the temporal autocorrelation of the series is broken. Scrambling will therefore make the SC of the actual series appear more significant than it really is. Therefore the decorrelation timescale of the actual series needs to be considered, and a random permutation by blocks is performed instead: we divide the series into subsets of equal duration, the length of which is the decorrelation timescale.

We investigate both coupling for synchronous wind and current fields, and lagged coupling with wind leading current, whereby we search for evidence of remote wind forcing of the flow. Because we suspect that the variability on the considered timescale is mostly barotropic and involves a quite rapid adjustment of the ocean to changes in the forcing, we consider only lags ranging from 0 days (synchronous fields) up to 50 days (with WSC fields leading the MC velocity series), although a response with longer lags caused by baroclinic processes is not excluded. We focus on the leading two modes at each lag, for which the SC, the SC fraction, and the correlation coefficient of the expansion series are evaluated [Fig. 5]. Most importantly, the significance of SVD modes is also tested from Monte Carlo simulations at the 99% level, and this is the criterion with which we decide whether to seek a physical interpretation for each mode. Henceforth, only two modes are focused on (circled in Fig. 5a), namely the leading SVD mode (SVD1) at zero lag and also at a lag of 30 days. We also note that the second mode (SVD2) at a lag of 40 days is statistically significant; it will not, however, be discussed specifically as it resembles many aspects of the SVD1 mode at 30 days.

a. Synchronous fields: A local mode

The leading coupled mode with zero lag accounts for 73.1% of the total SC, with an associated correlation coefficient of 0.62 [0.53], where the value in brackets is the 95% significance level. Associated spatial patterns for SVD1 are shown in Figs. 6a and 6b, with time series of expansion coefficients shown in Fig. 6c. The pattern map corresponding to the velocity field in the MC (hereafter denoted SVD1_{MC}) (Fig. 6a) closely matches EOF1_{MC} discussed in section 2a (Fig. 3a); furthermore SVD1_{MC} represents 49.3% of the variance of its own field, which is approximately the fraction of variance associated with EOF1_{MC}. As for EOF1_{MC}, SVD1_{MC} is mostly transverse to the 1000-m isobath, and therefore contributes little to the mean equatorward flow except for the westernmost mooring considered here (mooring M5, located in the core of the current) for which there is a more southward contribution, acting to diminish the total equatorward flow. The corresponding pattern for the WSC field (SVD1_{WSC}) is dominated by a strong positive anomaly centered on 50°W and 43°S, that is, over the Brazil Current overshoot (or Brazil–Malvinas Confluence Extension) (Fig. 6b). The structure of the WSC pattern is coherent southward to 54°S except for two zones of slightly negative anomalies, one being centered on the current meter array and the other at about 40°W at the same latitude. The expansion coefficient series for each field (Fig. 6c) are significantly correlated (0.62) and dominated by an annual modulation, whereas, by contrast, there is no suggestion of coherence at shorter periods. The annual harmonic accounts for 60% of the variance of SVD1_{MC} expansion series and for 40% of the variance of SVD1_{WSC} series, in phase with the former.

This covarying mode associates an annually modulated WSC anomaly at 40°S to a broadening of the current to the east and a slight diminishing in its core. A simple interpretation can be given to this mode by considering the seasonal migrations of the Brazil–Malvinas Front. Using a barotropic model of the South Atlantic forced by climatological winds, Matano et al. (1993) showed that the Brazil Current is strongest and reaches its southernmost location during austral summer months (Dec–Feb), and retreats to its northernmost location during austral winter months (Jun–Aug). Furthermore, in their model, annual fluctuations of the Brazil Current are phase-locked with annual WSC modulations at 40°S, with a positive WSC anomaly associated with an increased transport. In fact, they show that the shape of f/H contours is such that the barotropic variations of the Brazil Current are mostly affected by the wind field south of 35°S and to the west of the Mid-Atlantic Ridge, which corresponds to the pole in the WSC field found here. Thus, an increased wind-driven transport during austral summer favors the poleward penetration of the Brazil Current. The Subantarctic Front is in turn rejected more to the south, causing the retroreflection of the MC to occur earlier, closer to the current meter array, yielding southeastward motions and a slight reduction of the MC in its core. Such an annual modulation affects the path of the MC, but not necessarily its transport. Indeed, only a marginal annual cycle appears in transport time series from T/P data compared to the variability near the semiannual period (see section 1). In dimensional units, WSC fluctuations at 40°S associated with this SVD mode are \( \sim 1.2 \times 10^{-7} \text{ N m}^{-3} \), and transverse velocity anomalies are \( \sim 6 \text{ cm s}^{-1} \).

This analysis was repeated (not shown) with only the current meters presenting the longest time series, that is, 512 days: the conclusion is unaltered, with the same annual cycle and the same pattern in the WSC at the same location.
Fig. 6. SVD1 (first mode) for synchronous WSC fields and MC velocity fields. This mode explains 73.1% of the total squared covariance. (a) Spatial pattern for the velocity field in the MC (SVD1MC). The percentage indicated to the left of this subplot is the percentage of variance of its own field accounted for by this coupled mode. (b) Spatial pattern for the wind-stress curl field (SVD1WSC). Areas of negative polarity are shaded, and the two stars sketch the extension of the current meter array. For clarity, the spatial pattern associated to the velocity field in the MC is repeated here, for the uppermost current meters at each mooring only (bold arrows). The percentage indicated on the left of the figure represents the fraction of variance of its own field accounted for by SVD1WSC. (c) Expansion coefficient series for the velocity field in the MC (solid) and WSC (dashed). The correlation coefficient between these two series is 0.62.
b. Fields lagged by 30 days: A remote mode

When the WSC field leads the MC velocity field by 30 days, a significant coupling between the fields is featured in SVD1. This mode accounts for 67.2% of the total SC, and the coupling correlation coefficient is 0.73 [0.44]. Associated spatial patterns for the velocity field (SVD1_{MC}) are oriented to the east-northeast (Fig. 7a), contributing to strengthen the mean equatorward flow shown in Fig. 2. The associated WSC pattern (SVD1_{WSC}) is dominated by a strong negative WSC anomaly in the region of DP, with an intensification of the field centered on 56°S, 55°W [Fig. 7b]. The northern edge of this intensification straddles the 1000-m isobath, indicative of the path of the MC. Note also a positive anomaly to the northwest of the Falkland Islands on the Patagonian plateau; this is a very shallow region, suggesting that despite this pole being as intense as the former, its impact on the MC variability might not be as important. The associated expansion coefficients (Fig. 7c) have substantial annual variability (approximately 30% of the variance for both SVD1_{MC} and SVD1_{WSC}), but even more energy is concentrated close to the semiannual period (a harmonic with period 150 days accounts for 40% of the variance of both series). Overall, a consistent picture emerges from this mode if we assume that the low-frequency variability in the MC is primarily barotropic. The topographic Sverdrup balance for the vertically integrated flow \( u \) is given by

\[
\frac{\rho u}{H} \cdot \nabla \left( \frac{f}{H} \right) = \text{curl} \left( \frac{\tau}{H} \right),
\]

where \( \tau \) is the surface wind stress. A negative WSC anomaly thus induces a flow in the direction opposite to the gradient of \( f/H \), that is, toward shallower depths in the Southern Hemisphere (over a region small enough to neglect the variations of \( f \) compared to those of \( H \)). The \( f/H \) contours are depicted in Fig. 1; the large-scale flow entering DP encounters a \( \nabla f/H \) field oriented to the west. Essentially, DP is a sill, and a negative WSC upon the saddle point causes a convergence of fluid towards lower (more negative) values of \( f/H \). Such a flow is then free to follow \( f/H \) contours that can be traced along the Patagonian shelf break up to the current meter array location, or alternatively, following \( f/H \) contours along Antarctica. Hughes et al. (1999) discussed the important role that the WSC was likely to play over specific key regions along the path of the ACC, enabling the wind stress–driven barotropic flow (“free mode”) to divert from \( f/H \) contours that almost circumnavigate the globe towards other contours that do not. Coherence between the MC and the circum-polarly integrated wind stress is discussed in the next section.

A question raised by this SVD mode concerns the physical meaning of the 30-day lag. The \( f/H \) contours that intercept the current meter array at 40°S range from \(-7 \times 10^{-8} \) (for M5) to \(-3.5 \times 10^{-8} \) m\(^{-1}\) s\(^{-1}\) (for M7), which, according to Fig. 1, have different lengths if we trace them down to DP since those straddled by M6 or M7 have to circle around the edge of the Falkland Plateau before reaching DP. Given the length of these contours (~3000 km), a crude estimate for the phase velocity would be 1.2 m s\(^{-1}\), which is slow compared to the typical speed of long barotropic Rossby waves \( O(10 \text{ m s}^{-1}) \), but perhaps not unrealistic for shorter waves (the constriction of the DP region probably inhibits the development of the largest waves). However, such velocities are more suggestive of a mode of propagation involving stratification. The propagation speeds discussed here must only be taken as order of magnitude estimates, as the SVD analysis might not provide a very accurate measure of the lag [which could range between 20 and 35 days (see below), with corresponding speeds ranging from 1.0 to 1.7 m s\(^{-1}\)].

In the previous section we suggested that the MC transport fluctuations had little impact on the annual migrations of the Brazil-Malvinas Confluence since no clear annual fluctuation could be seen in the equatorward transport. The SVD1 mode considered here, however, exhibits a significant annual contribution (Fig. 7c). Nevertheless, its phase is opposite to what would be expected if the MC were to contribute to the annual drift of the front. Indeed, SVD1_{MC} suggests that the equatorward flow in the MC is in phase with the poleward flow in the Brazil Current. In fact, it is worth underscoring that the SVD yields a decomposition based on the covariance of the two fields and does not necessarily discriminate between the dominant frequencies in each. Hence, as the annual signal is important both in the WSC field and in the component of the velocity field transverse to the mean flow, it is not surprising to have an annual modulation also present in this 30-day lagged mode.

However, this raises the issue, as discussed by Chelton (1982), that the correlation between two geophysical quantities with substantial seasonal variability may merely reflect coincidence rather than causality. Narrowband signals have few degrees of freedom (only two for a pure harmonic function) and require long series for reliable statistics to be established. Here the series are short (1 year), and a simple way to ensure that this SVD mode is not coincidental is to perform the SVD analysis again after removal of the seasonal variability (parameterized here as the best fit to the sum of an annual and a semiannual harmonic). Doing this, we see not only that the SVD1 mode lagged by 30 days is still significant (albeit with a slightly smaller SC fraction and correlation coefficient, for which the significance level is also smaller) but that it also gives a clearer picture of the effect of the WSC at DP on the transport of the MC. The shape of the patterns for both fields as well as the associated expansion series (Fig. 8) are consistent with those found before, with nevertheless subtle and interesting differences. First, changes in SVD1_{MC} feature velocities aligned more closely in the direction of isobaths, with increased magnitude for mooring M5.
located in the core of the current. SVD1_{WSC} has been slightly modified in three ways: 1) the northeastern portion of the window (east of 50°W and north of 54°S) presents no significant variability, with a succession of small-scale patterns of opposite polarity and small amplitude; 2) the area of positive polarity on the Patagonian Shelf has a smaller magnitude, although it has extended up to the latitude of the current meter array; 3) the main pole of variability in the DP region has been shifted to the west and is now centered in the middle of DP, while its magnitude is unchanged. All these observations tend to reinforce the interpretation given earlier to this mode.

In dimensional units, WSC fluctuations at DP for this SVD mode are \( \sim 10^{-3} \) m s\(^{-1} \) and from \( \sim 7 \) cm s\(^{-1} \) (at M5) to \( \sim 4 \) cm s\(^{-1} \) (at M7) for the velocity field.

Finally, we note that after removal of the seasonal
cycle SVD modes at neighboring lags, from 20 to 35 days, are also significant (although associated with a smaller SC). Examination of these modes is sufficient to realize that they duplicate the 30-day lagged mode, suggesting that the measure of the lag provided by the SVD analysis is more suggestive of a timescale rather than being an accurate measure.

4. Winds over a circumpolar domain

Although the analysis of the previous section suggests an effect of the WSC at DP on the fluctuations of the MC, it seems unlikely that this regional process should be the sole factor determining the strength of a current that branches directly from the ACC, which is known

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**Fig. 8.** Same as Fig. 7 (SVD1 for 30 days lagged time series) except that the seasonal variability, parameterized here as the best fit to an harmonic function with annual and semiannual periods, has been removed from each field prior to the SVD analysis.
to respond primarily to the zonally averaged wind stress (e.g., Wearn and Baker 1980; Hughes et al. 1999; Gille et al. 2000). To examine whether the WSC over other parts of the Southern Ocean covaries with the MC transport, the previous analyses are repeated here on a circumpolar domain. In addition, in order to test whether the MC is also a direct extension of the ACC from the viewpoint of its barotropic fluctuations (which are responses to the large-scale zonal winds), coupling with the wind stress is also investigated using the same technique. Significant modes for the lagged-SVD analysis both with wind stress and WSC, with and without their seasonal variations, are shown in Fig. 9.

a. Wind stress curl

Focusing first on the WSC field, Fig. 9a shows a significant SVD1 mode at zero and 10-day lag. These two modes are almost identical (not shown) and are consistent with the “local” mode obtained on the regional domain for synchronous fields. They feature a positive WSC anomaly over the Brazil Current extension at 40°S associated with an annually modulated transverse velocity anomaly for the MC, contributing little to the equatorward transport. Over the rest of the Southern Ocean, the distribution of the WSC is globally positive north of 50°S and negative south of this line. It is not possible to assess whether the WSC over the rest of the Southern Ocean has a real effect on the MC. We simply note that this mode is compatible with its counterpart for the regional analysis and is consistent with the interpretation given above [relying on Matano et al.’s (1993) study] as being mostly a regional effect on the variability of the Brazil Current, and on the annual drift of the Brazil–Malvinas Front.

From Fig. 9a, we also note that SVD1 mode at lag 30 days (where we would expect to see the “remote” mode) is no longer significant but that, instead, the SVD2 mode is significant at all lags except one. Since all these SVD2 modes show only subtle differences in their associated patterns, it is legitimate to focus on the one that is associated with the largest fraction of the SC (35%) and with the largest coupling correlation coefficient of 0.82 [0.42], that is, the mode at a lag of 20 days, shown in Fig. 10. Both the expansion coefficients and the velocity field pattern resemble those for the remote mode in the regional analysis (Figs. 7a,b, the difference is here for the WSC pattern that is not only negative over DP but also over most of the Southern Ocean, north of 60°S (Fig. 10c). After seasonal variations are removed (Fig. 9c), this remote mode remains, showing as the leading mode at a lag of 10–20 days and accounting for 60% of the total SC with a correlation coefficient of 0.8 [0.37].

b. Wind stress

The lagged SC for the wind stress (Fig. 9b) resembles that of the WSC. The SVD1 mode at a lag of 0–10 days (not shown) is a repeat of the local mode discussed for the WSC analysis. Both the MC velocity pattern and expansion coefficients are comparable to those in Fig. 6, whereas the wind stress pattern has no circumpolarly coherent structure but is such that (in the regional domain considered in the previous section) its curl is consistent with the WSC pattern featured in Fig. 6c.

The remotely forced mode appears again in SVD2 at lags from 0 to 20 days, with the largest SC fraction (35%) and correlation coefficient (0.74) at 20 days. The pattern for the MC velocity and expansion coefficients (Fig. 11) are again barely changed from what they were for the SVD analysis with the WSC in both regional and circumpolar domains. A striking aspect of the wind stress pattern (Fig. 11c) is its circumpolarly coherent structure, featuring a clear negative anomaly of the strong westerlies over the path of the ACC, here associated with an increased transport in the MC. We also note that this wind stress pattern is consistent with the WSC pattern of Fig. 10c.

After removal of the seasonal cycle, unlike for the WSC analysis, this mode is no longer significant at the 99% level (Fig. 9d). Although it does appear in SVD1 at a lag of 20 days, with the largest SC fraction, it is only significant at the 92% confidence level (which is the most significant mode of all those in Fig. 9d). The lack of robustness of this mode suggests that the WSC is perhaps a more important parameter for the MC than the wind stress.

c. Conclusion

These analyses of the wind field on a circumpolar scale are consistent with those carried out on the re-
Fig. 10. SVD2 mode at a lag of ~20 days for the WSC on a circumpolar domain and the velocity field in the MC. Conventions are the same as in Fig. 7.

Regional scale. They feature two main modes, the first of which is an annually modulated local mode that we believe reflects the action of the wind-driven annual fluctuations of the Brazil Current on the velocity records through the seasonal displacement of the Subantarctic Front. It provides little contribution to the equatorward transport of the MC. The second mode of interest shows the effect of negative WSC anomalies at DP causing the MC to accelerate. The circumpolar analysis suggests that this mechanism is effective is, as was previously stated, somewhat uncertain. Nevertheless, a lag of 10–20 days is suggested by the circumpolar analysis, whereas it was 20–35 days for the regional analysis.

Another point of interest from this section is the pattern of the wind stress associated with the mode that forces the MC transport. Besides being consistent with the WSC patterns, it shows a most surprising and quite paradoxical feature: an acceleration of the MC is here associated with a deceleration of the circumpolar westerlies. Remembering that the ACC variability, at least close to the Antarctic continent, is coherent with the circumpolarly averaged wind stress (e.g., Hughes et al. 1999), this suggests that the ACC and MC vary with opposite phase. At this stage, it is clear that further elements are needed to assess whether this is a genuine oceanographic feature rather than a statistical curiosity. In the next sections, distinct (and longer) datasets and different techniques of analysis are used for this purpose. However, we note here that this paradox disappears if we accept that the WSC (rather than the wind stress) is the important parameter: a globally negative WSC anomaly over the Southern Ocean is associated with equatorward velocity anomalies in the MC. This is the predicted behavior of a western boundary current subjected to a time-varying Sverdrup balance.

5. Comparison with BPRs at Drake Passage

To assess the potential for the variability in the MC to be remotely forced, current meter data are compared to bottom pressure measurements at sites ND2 and SD2, at each edge of DP. These instruments were aimed at monitoring the large-scale barotropic transport of the
ACC, but Meredith et al. (1996, 1997) showed that while the BPR at site SD2 was well suited for this purpose, instruments at ND2 recorded significant baroclinic variability at periods shorter than 100 days. These were attributed to meanderings and excursions of the northernmost jet of the ACC, and/or some coastal process. The northernmost jet of the ACC directly branches to the MC shortly after the ND2 site, thus it is possible to verify whether this variability affects the South Atlantic as far north as 40°S. The following analyses are based on coherences (i.e., magnitude squared coherence) for which both spectral and cross-spectral estimates have been determined using the adaptive version of the multitaper method (Thomson 1982; Percival and Walden 1993), using seven data tapers, with a duration × half-bandwidth product of 4. Significance levels for coherences have been estimated based on the hypothesis of a harmonic process drown in a background red noise, for which the random variable

\[(N - 1)\hat{\gamma}(f)/[1 - \hat{\gamma}(f)],\]

where \(N\) is the number of tapers and \(\hat{\gamma}\) is the coherence estimator, is distributed like \(F_{2,2(N-1)}\) (Fisher–Snedecor law) so that the \((1 - \alpha) \times 100\%\) confidence level on the coherence is given by \(1 - \alpha^{1/(N-1)}\).

**a. BPR and velocity data in the MC**

A compact way to investigate possible relationships between current meter and BPR data is to estimate the coherence between bottom pressure and the principal components associated with each of the two leading EOFs of the velocity field (EOF\(_{MC}\)) shown in Fig. 4. As suggested in the previous sections, the only component of the velocity anomaly in the direction of the mean MC is likely to be remotely forced and, in fact, no significant coherence is found between EOF\(_{MC}\) (transverse mode) and bottom pressure anomaly at either the ND2 and SD2 sites, or the pressure difference ND2–SD2 (used in some studies as proxy for the total geostrophic transport variability through DP). This is consistent with the more local origin of this cross-isobath velocity mode. By contrast, EOF\(_{MC,1}\) (along-isobath mode, Fig. 3b) exhibits a significant coherence with bottom pressure at ND2 of 0.75 (with a 99% confidence level of 0.54), for periods ranging between 50 and 75 days.
Fig. 12. Bottom pressure at Drake Passage and principal component associated with the leading EOF of the velocity at site M5. The principal component associated with EOF(M5) (dashed line) is shown together with the bottom pressure [solid lines at site (a) ND2, (b) SD2 (solid), as well as (c) the pressure difference between ND2 and SD2]. Series are 30-day low-pass filtered to remove lunar tides in the BPR records, centered, and normalized to unit standard deviation.

days, with a phase lag of $\pi/6 - \pi/3$, corresponding to a delay of 5 to 12 days (BPR leading), increasing with increasing periods. There is no frequency for which significant coherences can be evidenced with either pressure at SD2 or ND2 − SD2 pressure difference.

Since significant coherence is found with the northern BPR only, we now test whether this is purely a coastal process trapped near shallower depths or if it affects the whole current equally. The analysis is therefore repeated with each current meter mooring separately, for which EOFs are evaluated (see VP99a for a detailed discussion of EOFs computed separately at each mooring), focusing only on the principal component corresponding to the mode in the direction of the mean MC (denoted EOF). At M5, located in the core of the current and lying approximately on the same $f/H$ contour as ND2, EOF accounts for more than half of the variance (see Fig. 8a of VP99a), and the corresponding principal component is shown in Fig. 12, together with the bottom pressure anomaly at ND2 and SD2, as well as ND2 − SD2. The coherence found with ND2 in the range 45–80 days is more robustly featured (up to 0.85 at 70 days) than for EOF$_{2\text{MC}}$ before (Fig. 13) with a phase lag of $\sim \pi/3$ (BPR leading), approximately constant in this interval (the corresponding phase delay ranges from 8 to 15 days). Again, there is no significant coherence either with SD2 or with ND2 − SD2. For EOF$_1$ at M6 (2200-m depth), coherences with ND2 are not as strong, although still significant for the same period range, whereas no significant coherence is found with M7 (3000-m depth). This suggests that this coherent mode at 45–80 days is trapped along the shallowest part of the continental shelf break, from DP up to the confluence area.

b. BPRs and the MC transport from T/P

1) MC transport estimate

The BPR/MC analyses presented so far are hindered by the fact that the velocity series are short, tending to render the results less conclusive. Accordingly, we now focus on the longer series of MC transport obtained from altimeter data. The ability of T/P data from ground track 26 (Fig. 1) to monitor the MC transport was tested against in situ measurements (VP99b), and the transport obtained from altimetry was well correlated (0.8) with direct estimates, suggesting that transport time series could be extended with a reasonable accuracy (a 3-yr series was obtained). Here, to take advantage of the length of the BPR series, we extend this estimate up to 5 years (Fig. 14). Although it is possible that the mean transport from the year-long current meter record differs from the mean transport for the 5-yr period represented by T/P, we nevertheless expect the variability to be relatively well depicted. Power spectral density for the transport features a broad peak at 70 days and a substantial semiannual cycle, stronger than the variability at the annual period (Fig. 15), in agreement with the results from the 3-yr series (VP99b).

2) Coherence with BPRs

Visual correlations are apparent between the 30-day low-pass filtered, 10-day sampled, MC transport series and BPR data at both ND2 and SD2 sites (Fig. 16). Coherences with each of the BPR series are estimated both from the full series as well as for the series truncated in mid-November 1996 in order to avoid the 20-day gap in the BPR data (Fig. 17). Coherences with ND2 data are in general agreement with those obtained from one year of direct velocity measurements, featuring a coherent band at periods ranging from 60 to $\sim$120 days. The broad coherence peak at periods shorter than 80 days obtained before is only robustly featured for the series truncated to 4 years (thin line in Fig. 17a), whereas the full series suggests significant coherences for periods larger than 80 days only, due probably to a poor interpolation in late November 1996. The phase spectrum is also consistent with that shown in section 5a, with a mean phase lag of $\sim \pi/3$ (8–20 days delay) in the frequency range where the coherence is significant. The distance separating both sites along $f/H$ con-
tours is ~2500 km, yielding a propagation speed of 1.4–3.6 m s\(^{-1}\).

Contrary to the analysis with short series, significant coherences with SD2 are found at periods between ~120 days and one year (Fig. 17c), spanning a spectral region almost complementary to that for which ND2 is coherent except for a sharp peak at 70 days that is only marginally significant and therefore not discussed. Most surprisingly, the associated phase spectrum (Fig. 17d) reveals that the pressure at the southern edge of DP is
coherent with the MC transport with only a small lag ($\leq \pi/4$), where we would have expected a lag close to $\pi$ if the circumpolar barotropic transport fluctuations had been responsible for changes in the MC transport (recall that the circumpolar transport variations are proportional to the opposite of those of the bottom pressure at the south of DP). This clearly appears in the coherence plot with ND2 − SD2 pressure difference where significant coherences for periods larger than ~120 days feature an opposite phase ($\pi$), whereas for periods shorter than ~100 days a more customary 20-day phase delay is featured. Although again perplexing, we note that the ACC appearing to vary with an opposite phase to that of the MC is consistent with the SVD analysis of the circumpolar wind stress (section 4), where equatorward velocity anomalies in the MC appeared to be associated with a negative anomaly of the circumpolar westerlies over the path of the ACC.

c. MC transport and circumpolar measures of the wind field

A last point we wish to discuss before concluding this study is the relationship between the wind field on a circumpolar scale and our 5-yr T/P transport time series. Several studies have focused on coherence analyses between various circumpolar measures of the wind field and BPR data at DP (e.g., Hughes et al. 1999; Gille et al. 2000) showing that the ACC is predominantly sensitive to the variations of the zonally averaged wind stress at the DP latitude range, responding with only a small lag, suggestive of a contribution both from acceleration and friction [a purely accelerating flow would instead feature a $\pi/2$ phase lag (Gille et al. 2000)]. A similar approach is undertaken here, comparing instead the MC transport; this complements section 3, with the advantage that the series are now five times longer.

Coherences are estimated both with the zonally averaged wind stress (zonal component only) and WSC at each latitude between 70$^\circ$ and 35$^\circ$S. Although showing significant coherence with the wind stress in the DP latitude range at periods of 100–200 days (Fig. 18), the lag is here again close to $\pi$. As for the analysis with the BPR data, this suggests an anticorrelation between the ACC and MC transport in this frequency range. Conversely, coherences with the zonally averaged WSC are significant in the 100–200 day band outside the range of latitudes of DP. They are larger than those obtained with the wind stress and are maximum between 40$^\circ$ and 50$^\circ$S. Although the coherence pattern is globally the same when the WSC is zonally averaged over the three oceans, they are strongest when the average is performed over the Pacific–Indian sector (Fig. 19), whereas there is virtually no coherence with wind stress curl averaged over the Atlantic sector only. The phase spectrum (Fig. 19b) shows that the MC transport lags the WSC variations with a phase $\approx \pi/3$ for winds south of DP, whereas it is close to $-\pi$ for winds north of DP for which coherences are largest.

If the MC is a boundary current whose temporal variability depended on Sverdrup dynamics, then its variability should be opposite to the velocity anomaly in the interior (assumed here to be regions north of DP). Consequently, it should be proportional to the opposite of the WSC for regions north of DP, yielding a lag close to $\pi$ between the two fields. To account for this expected phase lag, we consider the phase relationship shown in Fig. 19b relative to a lag of $\pi$. Accordingly, it appears that the MC transport trails the WSC in the 45$^\circ$–50$^\circ$S band, by only a small lag (0–$\pi/6$, i.e., less than 20 days for the coherent frequency range), which increases with increasing distance from DP latitude. The phase lag is larger ($\sim \pi/3$) with the WSC north of 45$^\circ$S but, since the coherent frequency range is shifted towards shorter periods (~100 days), the temporal lag is approximately

![Fig. 14. Volume transport of the MC at 40°S from T/P data. Units are Sv (Sv = 10$^6$ m$^3$ s$^{-1}$).](image1)

![Fig. 15. Power spectral density of the MC volume transport at 40°S from altimeter data. The 95% confidence level is marked.](image2)
the same. Figure 19 also suggests that positive WSC to the south of DP may also contribute to the MC fluctuations, with a somewhat longer lag ($\pi/6-\pi/3$). Negative WSC anomalies to the north of DP contributing to the MC transport are, however, more robustly featured in the period range 100–200 days, although the difference between positive WSC anomalies to the south of DP and negative WSC anomalies to the north of DP (parameterizing the convergence of fluid at the range of latitude of DP) seems to be an influential process for longer periods (250–300 days), resulting in an increased coherence with the MC transport.

6. Discussion and summary

Several analyses have been undertaken, aiming at understanding the sources of variability in an 18-month-long current meter record from the Malvinas Current. With the array being located near the Brazil–Malvinas Front, annual frontal motions cause the retroreflection point of the MC to migrate, affecting the velocity records. This effect is indirectly evidenced through a SVD of the covariance of velocity and wind stress curl fields over a regional domain, from which an annually modulated coupled mode associates a positive wind stress curl anomaly over the Brazil Current extension (40°S) to a cross-isobath velocity mode in the MC (contributing only little to the equatorward transport). Relying on Matano et al.’s (1993) study, this WSC anomaly is considered as a proxy for the strength of the Brazil Current, whose southward excursions constrain the MC to leave the shelf break earlier, yielding transverse velocity anomalies in the records. As this annual variability is important in the cross-isobath component of the velocity and is not clearly visible in the along-isobath component of the flow (which contributes to its transport), we conclude that the wind-driven Brazil Current is the main agent for the annual variability observed in the frontal motions.

The along-isobath component of the velocity field (which contributes a majority of its transport) instead appears to be remotely forced in a distinct SVD mode. This mode associates a negative WSC over DP to an increase of the equatorward flow in the MC with a lag of 20–35 days. This suggests that the input of negative vorticity from the wind at DP favors fluid motions against the gradient of background potential vorticity, with the fluid being subsequently free to flow along $f/H$ contours up to the confluence area. However, the SVD analysis repeated with the wind field over the whole Southern Ocean suggests that this mode is a local manifestation of a more global feature, associating a negative WSC field north of a dividing line at 60°S (including DP) to an increased transport of the MC.

It has been shown in several studies that the circumpolar barotropic transport fluctuations are driven by the wind stress rather than the WSC. Although these fluctuations occur predominantly near the Antarctic continent where $f/H$ contours are almost circumpolarly closed, Hughes et al. (1999) have suggested that the WSC may be important over areas such as DP, where
it can act to shift the flow toward another range of \( f/H \) contours where it would be free to flow (typically the MC), featuring a so-called “almost free mode.” As the influence of the WSC at DP is suggested by the SVD analysis, we repeated it with the circumpolar wind stress field instead, to examine the importance of the Hughes et al. (1999) theory here. Paradoxically, this analysis associates an increase of the MC transport with a negative anomaly of the westerlies over the ACC, suggesting that (contrary to expectations), the ACC and MC are 180° out of phase. This puzzling situation becomes understandable if we consider that, rather than the wind stress, the WSC is the driving parameter for the MC and that a globally negative WSC north of 60°S
Fig. 18. Coherence and phase spectra between MC transport from T/P and circumpolarly averaged zonal wind stress as a function of latitude. (a) Coherences larger than the 99% significance threshold. Solid lines indicate the latitude range of DP. (b) Phase spectrum (in radian) as a function of latitude, for significant coherences. A positive lag indicates wind stress leading on the MC transport. For clarity, the spectrum has been divided into only three frequency ranges: periods of 30–100 days (circle), 100–200 days (squares), and 200–400 days (triangles). The shaded annulus shows the latitude range of DP. To avoid overloading this panel, only the upper half (with the highest coherence) of spectral points of (a) are plotted.
associated with an increase in the equatorward flow in the MC at a small lag (~20 days) qualifies the latter as being a return flow of a time-varying Sverdup balance, a scenario proposed by Stommel (1957) for the mean flow of the ACC.

The above analyses can only be considered to be suggestive rather than conclusive, owing to the brevity of the time series considered. Thus we sought to assess their results from longer datasets and different techniques of analysis. BPR data from each edge of DP were
compared to current meter observations and showed a very different cross-spectral content depending on whether the northern or southern BPR was considered. The northern BPR is coherent with MC velocity data for periods ranging from 50 to 80 days, leading with a lag of no more than 20 days. As coherences are strongest with shallowest current moorings, a shelf-trapped wave is suggested, and its phase speed (1.4–3.6 m s\(^{-1}\)) is more reminiscent of a baroclinic influence. By contrast, no significant coherence is found with the southernmost BPR from the short current meter series. This analysis was, however, repeated with a 5-yr series of transport in the MC derived from T/P altimeter data, following procedures described by VP99b. Coherences with the northern BPR are again found, but the situation has changed for the southern BPR, where significant coherences are featured in a complementary spectral band, ranging from \(-120\) days to one year. However, the phase spectrum suggests a small lag, whereas a lag close to \(\tau\) would have been expected if the ACC and MC transport fluctuation had been directly correlated. Although again initially surprising, this finding is consistent with the analysis performed with the circumpolar wind stress above.

Finally, the 5-yr series of transport in the MC was compared to both the zonally averaged wind stress and WSC at each latitude from 70° to 35°S. While coherences with the wind stress are significant in the band 100–200 days, in the DP latitude range, the phase spectrum again features an anticorrelation. Largest coherences are, however, found with the zonally averaged WSC between 40° and 50°S at periods between 100 and 200 days. These are strongest when zonal averages are taken over just the Pacific Ocean (they are barely significant when the zonal average is performed over the Atlantic). An anticorrelation is also featured, which is not surprising if the MC is considered as the return flow of time-varying Sverdrup balance. Consistent with this hypothesis, the phase spectrum shows that the MC trails WSC variations by only a small lag (\(\approx \pi/6\) or 20 days), suggesting a rapid barotropic adjustment.

Two main broad spectral peaks had been identified in the MC transport variations (VP99b). The first one, near 70 days (and coherent with bottom pressure to the northern edge of DP), is identified as a shelf wave propagating along the continental margin, whose phase speed (1.4–3.6 m s\(^{-1}\)) suggests the influence of stratification. Similar intraseasonal coastal variability (periods of 30–70 days) has been evidenced propagating along most of the western coast of South America at a phase speed of 2–3 m s\(^{-1}\) (Clarke and Ahmed 1999), originating from incoming Kelvin waves in the equatorial Pacific. It now seems probable that these waves go around the edge of South America and continue their journey into the Atlantic Ocean (at least up to 40°S). Although not for the same period range or phase speed, Johnson (1990) showed that coastal waves could be traced from 12°S as far south as DP along the western coast of South America.

The second broad spectral peak in the MC transport, near the semiannual period, appears to reflect a barotropic adjustment to changes in the WSC north of 50°S, mostly in the Pacific sector. This suggests that two distinct regimes occur at DP: while barotropic fluctuations on the southern side are driven by the zonal wind stress, changes on the northern side would instead be driven by the WSC over subtropical regions of the Pacific [as speculated by Hughes et al. (1999)], subsequently forcing the MC. A profound difference affecting data from the two edges of DP is the presence of almost circumpolarly closed \(fH\) contours close to Antarctica, for which the Sverdrup balance does not apply, whereas over most of DP the bottom relief is high enough to block \(fH\) contours, a situation for which Sverdrup dynamics are not irrelevant (e.g., Wang 1994; Krupitsky and Cane 1994) despite the absence of a meridional boundary. We note that, while suggesting that a time-varying Sverdrup balance is the most influential process for the MC transport, the full dynamics of it have not been established here; in particular, there is no reason why this balance would be established along zonal contours, as the analysis of section 5c might suggest. The \(fH\) contours in the Southern Ocean are extremely contorted, making it virtually impossible to verify this hypothesis using the available observational database. However, we note that due to the presence of the mid-Pacific ridge, a range of contours (\(fH = -2.5 \times 10^{-8}\) m\(^{-1}\) s\(^{-1}\)) connects midlatitudes to the latitude of DP, which might be the beginning of an answer (Fig. 20). It seems also likely that stratification may play an im-
important role in “smoothing” the topography, specifically in regions such as Drake Passage, which could explain why the effect of wind forcing is not confined to the Pacific sector and reaches the MC.

Acknowledgments. The authors gratefully thank the science and technology groups of the Proudman Oceanographic Laboratory who provided the BPR data from the ACCLAIM Program. It is also a pleasure to thank C. Boone (IPSL, Paris) for preprocessing the ECMWF model outputs and A. Kartavtseff (LODYC, Paris) for calibrating and assessing current meter data. Useful discussions with E. Kestenare and N. Sennechal (LODYC, Paris) about the multitaper method are also duly acknowledged. We also thank the two anonymous reviewers for their valuable comments. The Confluence program is part of the French contribution to WOCE. It was funded by CNRS/INSU within the PNdEC program (Programme National d’Etude du Climat). Additional funding from the PNTS program (Programme National de Télédétection Spatiale) is also acknowledged. Financial support for FV was provided by the Ministère de l’Enseignement Supérieur et de la Recherche and by the Société des Secours des Amis de la Science.

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