Surface geostrophic currents across the Antarctic circumpolar current in Drake Passage from 1992 to 2004

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Abstract

The Southern Ocean plays an important role in the global overturning circulation as a significant proportion of deep water is converted into intermediate and deeper water masses in this region. Recently, a secular trend has been reported in wind stress around the Southern Ocean and it is thought theoretically that the strength of the ACC is closely related to wind stress, so one consequence should be a corresponding increase in ACC transport and hence changes in the rate of the global overturning. There are no long-term data sets of ACC transport and so we must examine other data that may also respond to changing wind stress. Here we calculate surface currents in Drake Passage every seven days over 11.25 years from 1992 to 2004. We combine surface velocity anomalies calculated from satellite altimeter sea surface heights with measured surface currents. Since 1992, the UK has regularly occupied WOCE hydrographic section SR1b across the ACC in Drake Passage. From seven hydrographic sections surface currents are estimated by referencing relative geostrophic velocities from CTD sections with current measurements made by shipboard and lowered acoustic Doppler current profilers. Combining the seven estimates of surface currents with the altimeter data reduces bias in the estimates of average currents over time through Drake Passage and we show that surface current anomalies estimated by satellite and in situ observations are in good agreement. The strongest surface currents are found in the Subantarctic and Polar Fronts with average speeds of 50 cm/s and 35 cm/s, respectively and are inversely correlated, so that maximum westward flow in one corresponds to minimum westward flow in the other. The average cross-sectional weighted surface velocity from 1992 to 2004 is 16.7 ± 0.2 cm/s. A spectral analysis of the average surface current has only weakly increasing energy at higher frequencies and there is no dominant mode of variability. The standard deviation of the seven day currents is 0.68 cm/s and a running 12 month average has only a slightly smaller standard deviation of 0.52 ± 0.16 cm/s. The Southern Annular Mode (SAM) measures the circumpolar average of wind stress and like the surface currents its spectrum has slightly increased energy at frequencies greater than 1 cpy. A cospectral analysis of these, averaging cospectra of five slightly overlapping 36 month segments improve statistical reliability, suggests that there is coherence between them at 1 cpy with the currents leading changes in the Southern annular mode. We conclude that the SAM and average Drake Passage surface currents are weakly correlated with no dominant co-varying modes, and hence predicting Southern Ocean transport variability from the SAM is not likely to give significant results and that secular trends in surface currents are likely to be masked by weekly and interannual variability.

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1. Introduction

Recent analysis of atmospheric observations in the Southern Hemisphere (Marshall, 2003; Thompson and Solomon, 2002) suggest that since the 1970s there has been an increasing trend toward anomalously strong westerly winds over the Southern Ocean and a number of papers report a Southern Ocean response to this change in forcing (Hall and Visbeck, 2002; Meredith et al., 2004). Theoretical considerations suggest that the strength of the Antarctic circumpolar current (ACC) is related to wind stress or square root of the wind stress (Johnson and Bryden, 1989; Bryden and Cunningham, 2003) so we might anticipate an increase in ACC transport as a response. The most comprehensive observations of ACC transport using current meters and CTD sections were made from January 1979 to February 1980 and gave a total net transport of 133.8 ± 11.2 Sv Whitworth and Peterson (1985) and adjusted by Cunningham et al. (2003) to account for deep flow. However, there are no long term measurements of the total transport of the ACC with which to compare to the wind stress data. A number of estimates of the total transport by geostrophic calculation from CTD sections assuming a deep reference level have been made for the ACC in Drake Passage. For seven CTD sections between 1975 and 1980, the transport was 102.7 ± 12.6 Sv and for seven sections between 1990 and 2000 was 112.2 ± 5.6 Sv (Cunningham et al., 2003). The variability in these few estimates is sufficient that the means cannot be distinguished. Interestingly, Lettmann and Olbers (2005) show that with a simple model of the ACC forced with the observations of rising wind stress these two groups of transports do indicate a statistically significant increase in ACC transport of 0.3 Sv/year from the mid 1970s until 2000. Lacking direct observations of the ACC transport we need to examine other data records such as long term bottom pressure, sea-ice or satellite altimeter records. In this paper, we combine satellite observations of steric sea surface height (SSH) anomalies and measurements of surface currents to obtain an 11.25 year timeseries of ACC surface currents in Drake Passage. Using this we describe the meridional structure of the ACC surface currents and search for a relationship between them and the Southern annular mode.

Lacking accurate knowledge of the earth’s geoid observations of sea surface height (SSH) anomalies from satellite altimeters can be added to a snapshot of the surface currents, giving a timeseries of surface currents e.g. Challenor et al. (1996) across the Antarctic circumpolar current in Drake Passage, Cromwell et al. (1996) across the Azores Current, Laing and Challenor (1999) across the East Auckland Current system, Imawaki et al. (2001) monitoring Kuroshio transport and Uchida and Imawaki (2003) also in the Kuroshio.

In Drake Passage, Challenor et al. (1996) calculate a timeseries of geostrophic surface currents by adding satellite SSH anomaly fields relative to $t_0$ and surface currents $u_0$ at $t_0$, estimating the surface currents by adding relative geostrophic currents from CTD measurements and currents measured by a shipboard acoustic Doppler current profiler (SADP). The principal problems are to determine $u_0$ and to verify the timeseries of surface currents with independent observations. We apply the method of Challenor et al. (1996) on WOCE choke point section SR1b across Drake Passage (Fig. 1), extending their work by using independent observations to verify the method and to estimate errors in an 11.25 year timeseries of surface currents. From CTD stations and direct current measurements (Table 1) we estimate the geostrophic surface currents from seven hydrographic sections between 1992 and 2001. The difference between pairs of these in situ velocities should match the altimeter derived velocity anomalies. Independent measurements of surface currents in 7 years gives us 21 comparisons of velocity difference pairs and from these we quantify the error in referencing satellite velocity anomalies to measured surface currents.

Errors from combining satellite and in situ data are large. However, we analyse the temporal and spatial variability in the 11.25 year timeseries of Antarctic circumpolar current surface currents and compare an index of transport-weighted velocities to the Southern annular mode.
2. Method

The following method combines hydrography and altimetry observations that are simultaneous in space and time (Challenor et al., 1996). Steric SSH relative to a level of no motion is $H$ and SSH measured by satellite altimeter relative to the reference ellipsoid is $h$. Since the level of no motion is parallel to the geoid the height of the geoid relative to the ellipsoid is,

$$G + k = H_t - h_t$$

Table 1

Summary of occupations of the WOCE section SR1b from Burdwood Bank at the northern edge of Drake Passage to Elephant Island at the southern edge (Fig. 1)

<table>
<thead>
<tr>
<th>Year</th>
<th>$N$ stations</th>
<th>Measurements</th>
<th>Satellite reference pass date</th>
</tr>
</thead>
<tbody>
<tr>
<td>12–15th November 1992</td>
<td>217</td>
<td>SeaSoar CTD to 400 m profile separation 4 km, SADP</td>
<td>11th November 1992</td>
</tr>
<tr>
<td>21st–26th November 1993</td>
<td>32</td>
<td>CTD, SADP</td>
<td>24th November 1993</td>
</tr>
<tr>
<td>16th–21st November 1994</td>
<td>30</td>
<td>CTD, SADP</td>
<td>23rd November 1994</td>
</tr>
<tr>
<td>15–20th November 1996</td>
<td>32</td>
<td>CTD, SADP LADP</td>
<td>20th November 1996</td>
</tr>
<tr>
<td>29th December 1996–7th January 1997</td>
<td>49</td>
<td>CTD, SADP LADP</td>
<td>7th January 1998</td>
</tr>
<tr>
<td>23rd–28th November 2000</td>
<td>32</td>
<td>CTD, SADP LADP</td>
<td>29th November 2000</td>
</tr>
<tr>
<td>20–26th November 2001</td>
<td>32</td>
<td>CTD, SADP LADP</td>
<td>28th November 2001</td>
</tr>
</tbody>
</table>

The section is 740 km long and the typical CTD spacing is 35 km.

Fig. 1. SR1b section (bold line) across Drake Passage. Bathymetry from Sandwell and Smith (1997). Mean positions of the Subantarctic front and Polar front from the location of the average maximum velocity, and minimum velocity for the Antarctic Polar Frontal Zone (dots) for cross-passage average surface current estimates from 1993 to 2004. The diameter of the dot equals one standard deviation of the mean position.
where \( k \) is a constant and \( t_0 \) is the time of observations. Differentiating along track and using geostrophy to estimate surface velocity anomalies from SSH anomalies (the Geoid is time invariant) gives,

\[
\frac{u_t}{u_{t_0}} = \frac{g}{f} \left( \frac{\partial h_t}{\partial y} - \frac{\partial h_{t_0}}{\partial y} \right)
\]

(2)

where \( f \) is the Coriolis parameter, \( g \) is acceleration due to gravity and \( u \) is the across track surface velocity. Hence, surface geostrophic currents are known for every satellite pass referenced to surface geostrophic currents at \( t_0 \).

For current measurements at two times \( t_{01} \) and \( t_{02} \) we have two estimates of the current at every time \( t \) and from (2) the difference between the measured currents relative to \( t_{01} \) and \( t_{02} \) is,

\[
\frac{u_{t_{01}}}{u_{t_{02}}} = \frac{g}{f} \left( \frac{\partial h_{t_{02}}}{\partial y} - \frac{\partial h_{t_{01}}}{\partial y} \right)
\]

(3)

Therefore, between \( t_{01} \) and \( t_{02} \) the surface geostrophic velocity difference equals the difference of the satellite velocity anomalies.

The extent to which (3) balances is a quantification of this method. We compare 21 pairs of differences of measured surface currents and satellite velocity anomalies to show that a time series of surface currents over many years can be obtained by reference to direct current measurements at one time.

3. Sea surface heights

In this study, we use the Ssalto/Duacs altimeter product Delayed Time Maps of Sea Level Anomaly\(^1\) (DT-MSLA). DT-MSLA data are gridded on a 1/3 degree Mercator grid, merging TOPEX/Poseidon or Jason-1 altimetry data with ERS-1/2 or ENVISAT data. Corrections for instrumental, wet and dry tropospheric, ionospheric, sea state bias, inverse barometer effect, tide influence (ocean, earth and pole) and orbit error are applied.

Five hundred and ninety seven day global DT-MSLA grids are available from October 1992 to January 2004. These data were interpolated onto ERS-1/2 track no. 465 in Drake Passage along which we have in situ data (Fig. 1) (Gaussian distance weighting and no temporal interpolation). The dates of satellite passes closest to the cruises are given in Table 1.

The total sea surface height error is determined by the uncertainties of applied corrections and by errors due to merging different data sets. Satellite corrections with the largest uncertainties vary on much larger scales than the currents we are trying to measure and can be neglected (Challenor et al., 1996). We estimate the total SSH error to be 5 cm RMS in the Drake Passage.

4. Geostrophic surface currents

In this section, we estimate surface currents by combining geostrophic currents calculated from pairs of CTD stations with upper and deep ocean current profiles measured by SADP and at each CTD station with a lowered Acoustic doppler current profiler (LADP). Geostrophic profiles are compared to the SADP in the top 300 m and to the LADP within the bottom 250 m and the mean of these gives the reference currents for the geostrophic profiles. Each data type is described, first CTD, SADP and then LADP and measurement errors for each are estimated and the removal of tides and ageostrophic components for the ADP measurements are discussed. In the following section, the final a priori error estimate will be compared to the error inferred by the comparison with the satellite velocity anomalies.

We estimate the surface geostrophic current between station pairs by,

\[
u_{t_0} = u_k + (u_{Son} + u_{Soff} + u_L)
\]

(4)

\(^1\) The altimeter products were produced by Ssalto/Duacs as part of the Environment and Climate EU Enact Project (EVK2-CT2001-00117) and distributed by Aviso, with support from CNES.
where \( u_g \) is the CTD derived surface geostrophic current calculated relative to zero at the deepest common level between CTD station pairs. \( u_{\text{Son}} \) is the difference between the CTD geostrophic and the station pair average of SADP currents at the two CTD station. \( u_{\text{Soff}} \) is the difference between the CTD geostrophic and SADP currents averaged between CTD stations. \( u_i \) is the station pair average of near bottom currents measured by LADP. The range of the three reference current estimates defines the error in our estimate of \( u_i \) and the over-bar denotes the average of these and is referred to as the reference current.

The three estimates of reference currents are likely to be in error because of inaccuracies in measuring instrument heading, either internal to the instrument or relative to independent heading measurements, though the absolute error from heading errors is not known. Station pair averages of ADP velocities (\( u_{\text{Son}} \) and \( u_L \)) will be a poor estimate of the actual average velocity between station pairs \( u_{\text{Soff}} \) if there is strong temporal variability in the currents, or if currents between station pairs vary non-linearly. Heading error from misalignment of the SADP and independent heading measurements maps the forward motion of the ship into across track velocities, so \( u_{\text{Son}} \) are less influenced than \( u_{\text{Soff}} \) as the ship moves only slowly on station. LADP currents \( u_i \) are free from external heading errors but have an unknown heading error from the instrument itself. The sources of error are described in Alderson and Cunningham (1999) and Cunningham et al. (2003), and without a clear preference for any of the three estimates we choose to weight them equally and to average them to provide the reference current.

In 1992, CTD profiles between the surface and 400 m were obtained from a CTD aboard a towed undulator. Averaging the profiles over 12 km reduced the internal wave contamination of the CTD profiles (Challenor and Tokmakian, 1998), and geostrophic velocities were calculated relative to zero at 196 m.

From 1993 full depth CTD stations were obtained with a NBIS Mk3b, Mk3c or SeaBird 911 CTD. Typically, the station spacing is about 50 km in deep water and closer over the continental slopes. In 1997, the station spacing was half of other years. Data quality meets the requirements for WOCE repeat section hydrography (WHPO, 1991).

For a temperature or salinity bias of 0.001 °C and 0.001 and station spacing of 50 km the geostrophic current error is 0.01 cm/s or 0.2 cm/s, respectively. Random errors of this size in temperature and salinity result in negligible velocity errors and we estimate the net error in station pair geostrophic velocities to be less than 0.5 cm/s. For the 1992 data the velocity error is larger due to the 12 km profile separation and potentially larger salinity error associated with CTD measurements made from a towed undulator.

Currents in the top 300 m were collected continuously using a 150 kHz SADP. Ensembles were logged every two minutes in 8 m vertical bins. Heading corrections based on ASHTECH 3D-GPS were continuously available (King and Cooper, 1993) and the data are corrected for heading misalignment angle and speed (Alderson and Cunningham, 1999; Pollard and Read, 1989).

Over time GPS navigation has improved and so the random error in each two minute ensemble has been proportionally reduced. For the 1993 data set this random error is estimated to be 5 cm/s and no worse in subsequent years. For a typical averaging period of 90 min the random error of the mean current profile is estimated to be \( 5/\sqrt{45} = 0.75 \) cm/s. For the 1992 data set the 12 km average profiles of SADP have an error of 5 cm/s (Challenor et al., 1996).

The SADP reference currents \( u_{\text{Son}} \) and \( u_{\text{Soff}} \) are calculated as the average difference between the geostrophic and SADP profiles. SADP data shallower than 102 m and deeper than the 75% good quality flag are excluded. The typical mixed layer depth is about 100 m and we assume that shallower than this the direct velocity measurements are contaminated by inertial motions; 75% good is found between 200 and 250 m for our data. The vertical profile of geostrophic currents and SADP currents match closely. The average RMS difference between them for both on and off station SADP profiles is only 0.8 cm/s. Therefore, the estimated reference currents have little error associated with matching the geostrophic and SADP currents between 102 m and the 75% good level.

A 150 kHz broadband self contained ADP was mounted on the CTD frame in 4 years, making top to bottom measurements of currents during the CTD cast (Table 1). Approaching within 250 m of the seabed the LADP also measures the horizontal velocity of the instrument relative to the ground and absolute currents are obtained from a vector addition of the relative water velocities and instrument velocities. Typically near bottom currents are obtained between 50 and 250 m off bottom, in water depths greater than 3000 m. For 16 m vertical bins the standard deviation of each water track and bottom track velocity measurement is about
The velocity error has a contribution from bottom track and water track errors, and typically we obtain 50 independent estimates of near bottom velocity, and we estimate the error of the average near bottom current to be $\sqrt{2/50} = 0.2$ cm/s. At depth in the ACC the vertical velocity shear is small and we assume that station pair averages of $u_L$ can be assigned to the deepest common level.

SADP and LADP data were detided using the barotropic Egbert TPXO.6.2 model (Egbert and Bennett, 1994; Egbert and Erofeeva, 2002). The model predicts strong tidal currents (up to 15 cm/s) at beginning and at the end of SR1b track and very weak tidal currents (~1 cm/s) along most of the Drake Passage transect.

The net surface and reference currents are plotted for all years in Fig. 2. A typical year 1997 (Fig. 2e1) has a clear distribution of jets: the Sub Antarctic Front (SAF) centred around 56°S has maximum currents of about 50 cm/s; the Polar Front (PF) around 57.5°S is a narrower feature with peak currents also about 50 cm/s. South of the PF two large eddies are evident with surface currents of ±15 cm/s. In contrast, 2001 (Fig. 2g1) shows that the SAF and PF merged with peak currents of 64 cm/s. The satellite data will describe the evolution of the surface velocity variability between these realisations.

Comparing the reference currents (Table 2) we see that they are noisier than instrumental errors suggest. The standard deviation over all sections for $u_{Son} - u_{Soff}$ is 8.5 cm/s, from a combination of the assumption of stationery currents and linear horizontal shear between stations and mapping of ship motion into the underway data $u_{Soff}$. The standard deviation of the difference between on station velocities $u_{Son} - u_L$ is 9.5 cm/s. The scatter between the reference currents is large, and we think the overall error is probably set by heading biases,
which could only be reduced by better heading estimates between the ship and the SADP and by calibration of the LADP compass. If the errors in each of the three reference velocity estimates are independent we estimate that the error of each estimated surface current value is approximately \( \frac{\sqrt{2}}{\sqrt{2}} \approx 6.4 \text{ cm/s} \).

### 5. Comparison of in situ and satellite velocity anomalies

Twenty one pairs of differences of satellite derived velocity differences and in situ velocity differences, the left hand and right hand side of (3), are plotted in Fig. 3. The agreement between the in situ and satellite velo-

![Fig. 3. Velocity difference pairs versus latitude of in situ velocity (bold line) left hand side of (3), and satellite velocity differences (dashed line) right hand side of (3). Plots are ordered by decreasing cross correlation.](image)

Table 2

<table>
<thead>
<tr>
<th>Year</th>
<th>( \langle u_{\text{son}}-u_{\text{off}} \rangle )</th>
<th>( \mu ) (cm/s)</th>
<th>( \sigma ) (cm/s)</th>
<th>( n )</th>
<th>( n_1 )</th>
<th>( \langle u_{\text{son}}-u_{\text{L}} \rangle )</th>
<th>( \mu ) (cm/s)</th>
<th>( \sigma ) (cm/s)</th>
<th>( n )</th>
<th>( n_1 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>1993</td>
<td>1.1</td>
<td>9.3</td>
<td>29</td>
<td>28</td>
<td></td>
<td>0.3</td>
<td>10.4</td>
<td>28</td>
<td>25</td>
<td></td>
</tr>
<tr>
<td>1994</td>
<td>−1.9</td>
<td>9.8</td>
<td>29</td>
<td>26</td>
<td></td>
<td>0.8</td>
<td>8.6</td>
<td>49</td>
<td>45</td>
<td></td>
</tr>
<tr>
<td>1996</td>
<td>3.0</td>
<td>9.7</td>
<td>28</td>
<td>24</td>
<td></td>
<td>−2.4</td>
<td>8.1</td>
<td>29</td>
<td>26</td>
<td></td>
</tr>
<tr>
<td>1997</td>
<td>0.4</td>
<td>4.9</td>
<td>49</td>
<td>45</td>
<td></td>
<td>−0.7</td>
<td>10.9</td>
<td>29</td>
<td>27</td>
<td></td>
</tr>
</tbody>
</table>

The number of possible comparison pairs is \( n \) and the number used to form the mean is \( n_1 \). Column mean \( \mu \) in last row.
ity anomalies is striking; peaks and troughs align and have similar amplitudes. North of the PF at 58°S the amplitude of velocity anomalies is around 50 cm/s (we discuss the cross passage current and variance in Section 6). At least in the SAF and PF, currents vary by a factor of 10 more than our estimated error of 6.4 cm/s in the reference currents. Therefore, our timeseries of surface currents will not be dominated by errors in the reference currents.

The average difference between in situ velocity difference pairs and satellite velocity difference pairs (Fig. 3) cannot be distinguished from zero using the student-t test with a significance level of 5%. Cross correlations range from 0.93 to 0.34 (with one outlier of 0.12), and the average correlation is 0.6, so generally about 60% of the velocity variance is explained between the in situ and satellite velocity anomalies.

For the 21 pairs of curves the average difference is 0.9 cm/s with a standard deviation of 15 cm/s. The a priori estimates of errors the in situ currents is ±6.4 cm/s and the satellite SSH error is estimated to be ±5 cm which results in a velocity error of \( \left( \frac{g}{f} \right) \times \left( \frac{\sqrt{2} \times 5}{\Delta y} \right) = 12.8 \text{ cm/s} \) (\( \Delta y = 45 \text{ km} \), \( g = 10 \text{ m/s}^2 \) and \( f = 1.2 \times 10^{-4} \text{s} \)). Combining these as root-sum-squares the total error estimate is 14.3 cm/s close to the average difference between in situ difference pairs and satellite velocity difference pairs of 15 cm/s confirming our error estimates for each velocity component and closing the error budget estimates.

Referencing the satellite velocity anomalies to surface currents at one time gives a timeseries in which satellite velocity anomalies and reference velocities can be combined with a net error of around 15 cm/s. Thus the in situ and satellite currents are quite compatible and combining them will produce surface current estimates that have a prescribed error and can be used to analyse the surface current structure of the ACC in Drake Passage. We now examine the meridional current structure of the Antarctic circumpolar current.

6. Velocity variability in the subantarctic and polar fronts

Using the 1997 in situ reference currents as the reference velocities in (2), we have constructed the time mean of the cross passage surface currents using the satellite altimeter data from October 1992 to January 2004.

The two most prominent features (Fig. 4a) are the eastward jets of the SAF and PF and between them the Antarctic Polar Frontal Zone (APFZ). South of the PF velocities are smaller and less variable. The SAF has an average speed and position of 50 cm/s at a 55.7°S, respectively, while the PF has a lower average speed of
35 cm/s and is located at 57.5°S. Between these two fast eastward jets the APFZ has an average speed of 0 ± 20 cm/s, and occasionally exceeds 50 cm/s westward. The mean positions of these features are marked in Fig. 1. This shows the SAF flows through a deep basin north of the West Scotia Ridge, that the APFZ is located close to the ridge crest and that the PF is located on the southern flank of the ridge and that the mean positions do not change much and that in this part of Drake Passage the two main eastward jets are separated by topography.

The variance of the surface currents (Fig. 4b) is smallest at the northern and southern ends of the SR1b section where the surface velocities also decay to near zero. Across both the SAF and APFZ the variance is \( \sim 380 \text{ cm}^2 \text{ s}^{-2} \) and in the PF is \( \sim 250 \text{ cm}^2 \text{ s}^{-2} \). Southward of the PF the variance is less than 100 cm\(^2\) s\(^{-2}\) and reduces steadily southward. So, although hydrographic observations show that eddies are common south of the PF, the variability of the two main ACC jets dominates variability of flow through Drake Passage.

Despite separation by the West Scotia Ridge SAF and PF maximum velocities are inversely correlated. Correlating maximum SAF velocities between the latitudes 55.5 ± 0.5°S, minimum APFZ velocities between 56.5° ± 0.5°S and maximum PF velocities between 57.5 ± 0.5°S (Table 3), we find significant negative correlations between the SAF and PF and between the SAF and APFZ. When the SAF jet has maximum eastward velocities the PF has minimum eastward velocities and the APFZ flow is maximum westward. The SAF and PF are inversely correlated and separated by the APFZ which is more strongly correlated with the SAF than the PF. We conclude from this that there must be an underlying large scale connection between the two main fronts and that the circulation in the APFZ is more controlled by the SAF than the PF.

In the next section, we examine the timeseries of cross passage average velocities.

### 7. Velocity timeseries

We have seven independent sets of reference currents from which we can calculate seven timeseries of the cross passage average cross-sectional weighted surface velocities (Table 4). If there were no errors the average of each timeseries should be identical. There are however small differences, due to biases in the reference velocities in each of the 7 years. We have no way of knowing which set of reference velocities is preferred and rather than referencing each part of the timeseries to the nearest in time reference velocities creating a timeseries with unknown and varying bias, we choose to create a single timeseries by averaging all seven together. The range of the time series averages is 1.3 cm/s (Table 4) and if biases are independent then the error in the time average current can be estimated as a standard error. The time mean cross passage average of the cross-sectional weighted surface velocities from 1992 to 2004 is 16.7 ± 0.68 cm/s and the standard error of the average is ±0.2 cm/s.

Assuming velocity anomalies are depth independent, then the cross-sectional weighted velocity can be scaled to represent transport anomalies (for this section, 1 cm/s gives a transport of 25 Sv). The standard deviation of the seven day transport anomalies (Fig. 5a) is 17 Sv (0.68 cm/s), and the interannual variability (Fig. 5b) is half of this and has a standard deviation of only 8 Sv (0.32 cm/s). For a running average the standard deviation over any 12-month period is \( \sim 13 \pm 4 \text{ Sv} \) (0.52 ± 0.16 cm/s). The standard deviation of the high frequency variability is 13.7 Sv (0.55 cm/s). In 2004, the 12 month averaged transport anomaly is 13.5 Sv (0.54 cm/s) higher than at the beginning. A linear fit has an increase in transport of 0.46 Sv/year (0.018 cm/s/year) with \( R = 0.224 \). However, over some periods the changes can be larger, for example between 1995 and 2004 for the 12 month filtered timeseries transport increases by 40 Sv (1.6 cm/s) equivalent to 4.4 Sv/year. The average baroclinic transport relative to the deepest common level over a similar period is 137.1 ± 6.9 Sv

| Correlation R between the SAF maximum velocity, APFZ minimum velocity and PF maximum velocity |
|----------------------------------|---------|---------|---------|-------|
| SAF&PF                          | −0.35   | −0.42   | −0.28   | 0.0000|
| SAF&APFZ                        | −0.51   | −0.57   | −0.45   | 0.0000|
| PF&APFZ                         | −0.22   | −0.29   | −0.14   | 0.0000|

The 95% confidence intervals are C1 and C2. If \( P < 0.05 \) then the correlation is significant.
so a change of 40 Sv would be approximately 10% of the mean baroclinic transport. The range of baroclinic transports is 123 Sv in 1996 to a maximum of 146 Sv in 2003 (Table 5), a range of only 23 Sv. These sparse measurements limit detection of secular trends and comparison to the continuous surface current time-series. Cunningham et al. (2003) report that there has been no detectable change in transport from the 1970’s to 1990’s based on all available hydrographic data in Drake Passage, thought Lettmann and Olbers (2005), using a simple model suggest that transport has increased over this period at a rate of 0.3 Sv/year. Whitworth and Peterson (1985) present estimates of the total transport through Drake Passage measured by a current meter array, hydrographic sections and bottom pressure recorders from January 1977 to February 1978. During this period they observed a range in transport of 54 Sv with a standard deviation of 9.9 Sv. Subsequently modelled transport fluctuations using the measured cross passage pressure difference from 1977 to 1982 (no data during 1980) had a range of 56 Sv and standard deviation of 8.3 Sv. Between 1981 and 1982 the transport range was 63 Sv with a standard deviation of 12.6 Sv. Over this period large transport changes of

Table 4
Average cross-sectional weighted cross-passage surface current $\bar{u}$ through Drake Passage from 1992 to 2004 referenced to each of the seven in situ surface current estimates

<table>
<thead>
<tr>
<th>Reference year</th>
<th>$\bar{u}$ (cm/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1992</td>
<td>16.9</td>
</tr>
<tr>
<td>1993</td>
<td>16.7</td>
</tr>
<tr>
<td>1994</td>
<td>17.3</td>
</tr>
<tr>
<td>1996</td>
<td>16.7</td>
</tr>
<tr>
<td>1997</td>
<td>16.5</td>
</tr>
<tr>
<td>2000</td>
<td>16.0</td>
</tr>
<tr>
<td>2001</td>
<td>17.0</td>
</tr>
</tbody>
</table>

| $\mu$          | 16.7             |
| $\sigma$       | 0.4              |
| $SE$           | 0.2              |
| Range          | 1.3              |

Mean $\mu$, standard deviation $\sigma$, standard error $SE$ and range.

Fig. 5. Average cross-sectional weighted cross passage currents and transport anomalies of the seven timeseries, one for each reference velocity estimate. Left hand side scale velocity (cm/s) and the transport anomaly on the right (Sv, 1 cm/s = 25 Sv). (a) Average of the seven timeseries and (b) 12 month filtered mean.

(Table 5) so a change of 40 Sv would be approximately 10% of the mean baroclinic transport. The range of baroclinic transports is 123 Sv in 1996 to a maximum of 146 Sv in 2003 (Table 5), a range of only 23 Sv. These sparse measurements limit detection of secular trends and comparison to the continuous surface current timeseries. Cunningham et al. (2003) report that there has been no detectable change in transport from the 1970’s to 1990’s based on all available hydrographic data in Drake Passage, thought Lettmann and Olbers (2005), using a simple model suggest that transport has increased over this period at a rate of 0.3 Sv/year. Whitworth and Peterson (1985) present estimates of the total transport through Drake Passage measured by a current meter array, hydrographic sections and bottom pressure recorders from January 1977 to February 1978. During this period they observed a range in transport of 54 Sv with a standard deviation of 9.9 Sv. Subsequently modelled transport fluctuations using the measured cross passage pressure difference from 1977 to 1982 (no data during 1980) had a range of 56 Sv and standard deviation of 8.3 Sv. Between 1981 and 1982 the transport range was 63 Sv with a standard deviation of 12.6 Sv. Over this period large transport changes of
46 Sv and 56 Sv occur in August 1977 and June 1981, respectively. These estimates of the total transport variability and range are similar to the variability and range in our timeseries over similar periods. Hence it seems plausible that transport variability, derived from the surface currents, reflects the total transport variability.

Like Whitworth and Peterson (1985) we do not find any significant seasonal signal (not shown) in the surface currents.

In the next section, we compare the timeseries of surface currents to the Southern Annular Mode which measure the atmospheric sea level pressure gradient around Antarctica to investigate if the variability in surface currents can be related to atmospheric forcing.

8. Southern annular mode

The southern annular mode (SAM) is an inherent characteristic of the climate system and is a measure of the large scale alterations of atmospheric mass between the mid and high latitudes (Gong and Wang, 1999; Hall and Visbeck, 2002; Thompson and Wallace, 2000a; Thompson and Wallace, 2000b; Visbeck and Hall, 2004; White, 2004). The SAM index is defined as the difference in normalised zonal mean sea level pressure between 45°S and 65°S, and has been shown to explain more than 50% of the atmospheric sea level pressure (aSLP) variance around Antarctica. A high SAM index corresponds to a steeper aSLP gradient across the Southern Ocean which implies a stronger eastward wind stress, which provides a potential coupling to Southern Ocean transport variability (Bryden and Cunningham, 2003; Johnson and Bryden, 1989).

To compare our transport (or current) timeseries to the index of the SAM we create an ACC index (ACI) by subtracting the average and normalising by the standard deviation (Fig. 6a). The SAM and ACI indices have significant monthly variability, which smoothing using a 12 month filter (Fig. 6b) suppresses revealing that both timeseries vary in phase between 1993 and 2000, but for the last 3 years having a different relationship. The maximum crosscorrelation coefficient between the SAM and ACI is 0.4 with the SAM lagging the ACI by 11 months. Both indices have significant autocorrelation, reducing to zero after about 14 months, and we estimate the degrees of freedom in the 11.25-year timeseries as approximately 10. The annual means (Fig. 6c) should therefore be independent and show no significant correlation at any lag.

Spectra of the two indices are shown in Fig. 7. Both have almost white spectra with nearly equal energy at all frequencies but hinting at increased energy at annual and shorter periods. Hurrell and van Loon (1994) also calculate the power spectra the SAM and find that there is no semi-annual oscillation and that only at periods of 2.7, 4.2 and 45.7 months does the power spectral density exceed the 95% confidence limits.

Despite the absence of significant correlations between the SAM and ACI and the white character to their spectra we have calculated their cospectra and to improve the statistical reliability of these estimates we split

<table>
<thead>
<tr>
<th>Year</th>
<th>Month</th>
<th>Day</th>
<th>Transport (Sv)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1993</td>
<td>November</td>
<td>23.5</td>
<td>131.4</td>
</tr>
<tr>
<td>1994</td>
<td>November</td>
<td>18.5</td>
<td>140.4</td>
</tr>
<tr>
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<td>February</td>
<td>17.5</td>
<td>144.0</td>
</tr>
<tr>
<td>1996</td>
<td>February</td>
<td>17.5</td>
<td>131.0</td>
</tr>
<tr>
<td>1996</td>
<td>November</td>
<td>17.5</td>
<td>123.1</td>
</tr>
<tr>
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<td>January</td>
<td>2.5</td>
<td>143.8</td>
</tr>
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<td>November</td>
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<td>140.4</td>
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</tr>
<tr>
<td>2002</td>
<td>December</td>
<td>29.5</td>
<td>135.2</td>
</tr>
<tr>
<td>2003</td>
<td>December</td>
<td>13.5</td>
<td>146.0</td>
</tr>
</tbody>
</table>

| μ | 137.1 |
| σ | 6.9 |

Day is the mid-point day of each section, of a nominal 5-day cruise. Mean μ, standard deviation σ.
each timeseries into five slightly overlapping 36 month segments at the expense of revealing any low frequency variability. Each segment is detrended and the final cospectra are averages of estimates from the five segments. There is significant coherence between the two timeseries (Fig. 8a) at frequencies of 1 cpy and less and at frequencies of 2.5 cpy. For annual and lower frequencies the phase changes from $+20^\circ$ and $-20^\circ$, which implies a change in the lead-lag relationship between 1 cpy and 0.5 cpy.
We conclude from this and from the simpler cross-correlations that the SAM and Southern Ocean transport variability are, at very best, only weakly related. There are no dominant modes of variability in either, but at annual frequencies the SAM and ACI do covary. Changes in Southern Ocean transport (or surface currents) in Drake Passage are not clearly linked to the circumpolar large scale atmospheric forcing as measured by the SAM. This contrasts with a modelling study that found a strong correlation between the SAM and Southern Ocean sea ice (and other properties) (Hall and Visbeck, 2002) and an observational analysis (Meredith et al., 2004) showing a relationship between the changing seasonality of the SAM and ocean SLP variations from tide gauge measurements on the western side of the Antarctic Peninsula. We repeated the analysis of Meredith et al. (2004) for our surface current data but did not find evidence that there was a change in seasonality related to the SAM. White (2004) argues that the Antarctic circumpolar wave and teleconnected El Nino-Southern Oscillation dominate the SAM and so will ultimately be responsible for driving observed Southern Ocean variability – rather than the SAM itself.

9. Conclusion

Timeseries of surface geostrophic currents in the Southern Ocean can be monitored by combining satellite altimeter SSH anomalies with current measurement. Here we estimate surface currents by referencing CTD derived geostrophic currents to near surface currents measured by SADP and near bottom currents measured by LADP. The resulting timeseries depends on the initial accuracy of both the surface currents and the satellite derived velocity anomalies and we estimate the error in the surface currents to by ±6.4 cm/s and in the satellite anomalies to be twice as large. Hydrographic and ADP measurements were taken along the WOCE SR1b section across Drake Passage on seven occasions between 1992 and 2000. The difference in surface velocity between any pair of years must be matched by the same difference calculated from the altimeter. The seven occupations give twenty one pairs of velocity differences which we compare to the satellite derived velocity anomaly differences. Cross-correlations between each pair of difference curves ranged from 0.92 to 0.34 with a mean value of 0.6 and a student-t test could not distinguish between them. The mean difference in situ and satellite velocity anomalies for the 21 comparisons is 0 ± 15 cm/s, and the range of 15 cm/s is consistent with the a priori error estimate of 14.3 cm/s. From this favourable comparison we conclude that the in situ velocities and satellite velocity anomalies measure currents on spatial and temporal

Fig. 8. Cospectral analysis of the SAM index and Drake Passage transport anomalies (equivalently surface current anomalies). The timeseries were divided into five slightly overlapping 36 month segments and each segment is demeaned and the final cospectra are averages of estimates from the five segments. (a) Coherence squared versus frequency (cycles per year). Dashed lines are 90% and 75% confidence limits and (b) phase (degrees).
scales that allow the two data sets to be combined to create a timeseries of surface currents through Drake Passage.

The meridional structure of the ACC is dominated by two jets, the SAF whose average speed is 50 cm/s and position is 55.7°S and the PF which has a lower average speed of 35 cm/s and is located to the southward at 57.5°S. The APFZ between these currents has an average speed of 0 ± 20 cm/s but occasionally exceeds 50 cm/s westward. In the eastern end of Drake Passage the West Scotia Ridge divides the passage in two, with the SAF in the northward basin and the PF located on the southern flank of the ridge and the APFZ separating the two jets centred over the ridge crest. Despite the separation of the SAF and PF by the APFZ and the ridge, the jets are anticorrelated, so then the SAF has maximum eastward velocities the PF has minimum eastward velocities, indicating a large scale connection between the jets. Variability of the currents is highest in a zone approximately 2.5° wide including the SAF, APFZ and PF where the variance is between ~380 cm² s⁻² and ~250 cm² s⁻². Immediately south of the PF the variance of the surface currents drop rapidly in about 0.5° to less than 100 cm² s⁻². At the north and south of the SR1b section the surface current variance is low and velocities decay to near zero.

The average cross-sectional weighted average velocity of the ACC through Drake Passage is 16.7 ± 0.2 cm/s with a standard deviation of 0.68 cm/s. Interannual changes in surface current are typically ~1 cm/s. The largest secular trend in this record occurs between 1995 and 2004 when there is an increase in the average current of 1.6 cm/s, however over the whole record length the net change in surface current is only 0.2 cm/s (0.018 cm/s/year), and there is no evident secular increase in surface currents.

A cospectral analysis of an index of the transport variability (equivalently surface currents) and the SAM index shows that the timeseries covary in phase for frequencies of 1 cpy and less and at frequencies of 2.5 cpy. Because the timeseries are only weakly correlated and there are no dominant frequencies of variability predicting Southern Ocean transport variability from the SAM is not likely to give significant results.

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