Estimates of Eddy Heat Flux Crossing the Antarctic Circumpolar Current from Observations in Drake Passage

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ABSTRACT

The 4-yr measurements by current- and pressure-recording inverted echo sounders in Drake Passage produced statistically stable eddy heat flux estimates. Horizontal currents in the Antarctic Circumpolar Current (ACC) turn with depth when a depth-independent geostrophic current crosses the upper baroclinic zone. The dynamically important divergent component of eddy heat flux is calculated. Whereas full eddy heat fluxes differ greatly in magnitude and direction at neighboring locations within the local dynamics array (LDA), the divergent eddy heat fluxes are poleward almost everywhere. Case studies illustrate baroclinic instability events that cause meanders to grow rapidly. In the southern passage, where eddy variability is weak, heat fluxes are weak and not statistically significant. Vertical profiles of heat flux are surface intensified with \(50\%\) above 1000 m and uniformly distributed with depth below. Summing poleward transient eddy heat transport across the LDA of \(\approx 0.010 \pm 0.005\) PW with the stationary meander contribution of \(\approx 0.004 \pm 0.001\) PW yields \(\approx 0.013 \pm 0.005\) PW. A comparison metric, \(\approx 0.4\) PW, represents the total oceanic heat loss to the atmosphere south of 60\(^\circ\)S. Summed along the circumpolar ACC path, if the LDA heat flux occurred at six “hot spots” spanning similar or longer path segments, this could account for \(20\%–70\%\) of the metric, that is, up to \(\approx 0.28\) PW. The balance of ocean poleward heat transport along the remaining ACC path should come from weak eddy heat fluxes plus mean cross-front temperature transports. Alternatively, the metric \(\approx 0.4\) PW, having large uncertainty, may be high.

1. Introduction

The Southern Ocean heat balance affects Antarctic climate and glacial melting directly and global climate in general through its effect upon down- and upwelling across the Antarctic Circumpolar Current (ACC). These processes in turn facilitate the biological productivity around Antarctica and govern the sequestration and release of CO\(_2\). A proper understanding of how heat crosses the ACC is crucial to correctly model the ocean’s influence upon climate and has become critically needed owing to uncertainties about how the ACC system responds to changes in atmospheric forcing. Southern Ocean heat losses from ocean to atmosphere and northward heat losses by wind-driven Ekman transport must, in a slowly changing mean state, be balanced by ocean processes. Transient eddy processes and horizontal and vertical overturning circulation contribute to poleward heat transport. The relative role of these heat transfer mechanisms remains uncertain.

Because the ACC encircles the globe with several fronts that signify partial barriers to cross-frontal exchange, meanders and eddies must play an important role in producing meridional fluxes in the Southern Ocean. An early study by de Szoecke and Levine (1981) suggested that along a mid-ACC path defined by the 2°C isotherm, transient eddies were almost entirely responsible for cross-frontal heat fluxes. Exchange across ACC fronts is thought to be particularly concentrated in just a handful of locations with energetic eddies and steep stationary meanders, facilitated by bottom topography (Thompson and Naveira Garabato 2014). Estimates of eddy heat fluxes from observations have

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been made in a small subset of these exchange “hot spots,” notably Drake Passage (Bryden 1979; Nowlin et al. 1985; Walkden et al. 2008; Lenn et al. 2011; Ferrari et al. 2014), south of Tasmania (Phillips and Rintoul 2000), and south of New Zealand (Bryden and Heath 1985). Combining these estimates to elucidate the global role of transient eddies is challenging, not only because of uncertainties in the along-ACC heterogeneity of the transient eddy heat flux magnitude, but also due to the different techniques used to calculate the fluxes themselves; two examples are compensation for mooring drawdown in energetic regimes (e.g., Nowlin et al. 1985) and the separation of rotational and divergent components of ocean eddy heat fluxes (e.g., Marshall and Shutts 1981; Jayne and Marotzke 2002). More recently, analysis of an eddy permitting ocean model by Volkov et al. (2010) showed that the influence of transient eddies has a strong latitudinal dependence, weakening substantially south of ~60°S. Furthermore, Volkov et al. (2010) showed that stationary meanders are an important conduit for heat transport. In a stationary meander, both local and circumglobal, a zonal-mean meridional heat flux ensues if portions flowing northward and others flowing southward have different temperatures (Sun and Watts 2002).

Interaction of the ACC with topography that leads to turning of the mean along-stream ACC with depth provides another mechanism for poleward heat transport. The exact nature of this mean cross-ACC flow is not well understood. It has been hypothesized that this occurs at a few isolated locations by Sekma et al. (2013) based upon their observations in the narrow channel at Fawn Trough. Yet other studies indicate it may occur more ubiquitously. Chereshkin et al. (2012) showed that recirculations are common in the polar frontal zone within Drake Passage with components crossing the upper baroclinic zone of the ACC both northward and southward. Analysis of ocean circulation models shows that mean flow contributions could result from the accumulation of weak, mean, cross-ACC flow (Peña Molino et al. 2014) or from the accumulation of large, localized, positive and negative contributions (Ferrari et al. 2014).

The purpose of this work is to quantify poleward heat fluxes across the ACC by transient and stationary eddies using observations in Drake Passage. Our study, called cDrake, deployed an extensive array of current- and pressure-recording inverted echo sounders (CPIES) to measure the current and temperature structure through the full water column for 4 yr in Drake Passage. It included a transect spanning the channel plus a local dynamics array sited in a region of elevated eddy kinetic energy (Lenn et al. 2007; Firing et al. 2011) in the polar frontal zone between the Subantarctic Front (SAF) and the Polar Front (PF). Sections 2 and 3 present the details of the observations and methodology. In particular, section 2 presents techniques, developed in Gulf Stream and Kuroshio studies (Cronin and Watts 1996; Bishop et al. 2013), that remove a large, rotational, non-divergent contribution from the full eddy heat flux in order to identify the dynamically important divergent eddy heat flux.

Section 4 presents our findings: mapped mean eddy heat flux in central Drake Passage, case studies of cyclogenesis responsible for the observed eddy heat flux pattern, and estimates of eddy heat flux on a transect that spans Drake Passage. Section 5 returns to the issues raised in this introduction with a more thorough treatment of the relative contribution and role of the various ocean heat transfer mechanisms. The cDrake results and existing estimates of eddy and mean flow heat fluxes are discussed in the context of the global heat budget. Section 6 provides a summary of our results.

2. Observations and data

The cDrake experiment deployed current- and pressure-recording inverted echo sounders in Drake Passage for 4 yr (December 2007 through November 2011), arranged in two configurations (Fig. 1). One, the C line transect that spanned the 800-km passage, included 20 CPIES spaced by 40–60 km. In addition, a local dynamics array (LDA), centered upon the region of highest eddy kinetic energy (EKE) between the PF and SAF, included 24 CPIES sites in a two-dimensional grid with 40-km spacing. During the final year, a closely spaced third array of five CPIES was moored at the base of the Shackleton Fracture Zone (SFZ). A total of 43 CPIES were deployed in the three arrays.

CPIES measure hourly, near-bottom, horizontal currents ($u_{ref}$, $v_{ref}$) (50 m above the seafloor in order to be outside the benthic boundary layer), bottom pressure $p_{bot}$, and surface-to-bottom roundtrip acoustic travel time $\tau$. Most sites returned hourly data for the 4 yr (93%–97% data return for different variables). Low-pass filtering used a fourth-order Butterworth filter, passed forward and backward, and cutoff periods of 3 or 7 days as noted. These data are documented in a comprehensive technical report by Tracey et al. (2013).

Full-depth CTD casts were collected for calibration purposes at the CPIES sites during five annual cruises that visited the array sites. The CPIES data processing produced vertical profiles of the time-varying current and temperature fields, as treated in the following section 3.
The Ssalto/Duacs daily sea level anomaly products were produced by the Copernicus Marine and Environment Monitoring Service, and the mean dynamic topography MDT_CNES_CLS13 was produced by the CLS Space Oceanography Division. Both altimeter products were distributed by AVISO with support from CNES (http://aviso.altimetry.fr).

3. Methods

a. Measurements of temperature and current fields

CPIES determine temperature and horizontal velocity profiles \([T(p), u(p), v(p)]\) following methods presented in Donohue et al. (2010), which were expanded and applied to Drake Passage by Firing et al. (2014). In strong and eddying current systems, \(\tau\) has been applied to determine density profiles using a gravest empirical mode lookup table based on local hydrography (Meinen and Watts 2000; Watts et al. 2001). Applying geostrophy, laterally separated pairs of density profiles produce vertical profiles of baroclinic velocity relative to a near-bottom reference level, chosen here to be 3500 dbar. In a two-dimensional array, the velocities determined are two-dimensional baroclinic current profiles relative to the bottom, designated \(u_{\text{bcb}}, v_{\text{bcb}}\). The velocities and temperatures are mapped at half-daily intervals by optimal interpolation as described by Firing et al. (2014) with careful attention paid to error estimates.

Deep pressure anomalies are leveled to a consistent mean 3500-dbar geopotential, assuming that long time averages of near-bottom currents and bottom pressures are geostrophic. Multivariate mapping that combines the deep pressure and current measurements (Firing et al. 2014) provides the reference velocities \(u_{\text{ref}}, v_{\text{ref}}\) used to render the baroclinic velocity profiles absolute (Fig. 2):

\[
\mathbf{u} = (u, v) = (u_{\text{bcb}}, v_{\text{bcb}}) + (u_{\text{ref}}, v_{\text{ref}}) = u_{\text{bcb}} + u_{\text{ref}}. \tag{1}
\]

It is important to note that while the \(u_{\text{bcb}}\) component flows unidirectionally parallel to the front, the vector sum \(\mathbf{u}\) total current turns with depth due to the \(u_{\text{ref}}\) component.
contribution. The turning illustrated in Fig. 2 can occur instantaneously as well as in the time mean because deep eddies and mean topographically steered currents and recirculations can cross the upper baroclinic ACC structure.

Velocity and temperature time series at depths distributed through the water column from the CPIES LDA agree well with contemporaneous moored current meter measurements on French mooring M4 (Ferrari et al. 2012, 2014), whose location was near site E02 (Fig. 1). Firing et al. (2014) showed this comparison, which we repeat in Fig. 3 with correlations for 3- and 7-day low-pass filtering noted. The current meters measuring $u, v, T$ moved vertically in the water column as the mooring drew down in response to variable drag by the currents. So at each time sample, the CPIES $u, v, T$ measurements were computed to coincide with the pressure level of each current meter in order to conduct a comparison without introducing uncertainties due to mooring motion compensation.

The two time series for each variable at each depth compare point current meter measurements and optimally mapped geostrophic currents. These variables would have small intrinsic differences even if both measurements were perfect. Temperatures agree within 0.3°, 0.1°, and 0.1°C at approximately 520, 930, and 2540 dbar, respectively, and Firing et al. (2014) report the baroclinic velocities at the corresponding levels agree within 0.11, 0.08, and 0.05 m s$^{-1}$. Firing et al. (2014) accounted carefully for the observed differences as the sum of measurement and mapping error plus rapid small-scale ageostrophic processes present in the water column. The high correlation-squared values improve further with 7-day low-pass filtering ($r^2$ ranging 0.74 to 0.90), which removes some of the variability from small-scale processes. Some short-period variability in the moored point measurements can be noted that is not present in the CPIES records. For example, in the temperature records, fluctuations of small vertical scale, such as from internal waves or from filaments advected past individual current meters, would have insignificant effect upon $T$ and geopotential. These results are consistent with the high correlations also found south of Australia during the Subantarctic Front Dynamics Experiment (SAFDE; Watts et al. 2001).

b. Estimates of eddy heat flux fields

Eddy heat flux computed from the total velocity time series is denoted $(u'T', v'T')$, where $(\cdot)'$ indicates difference from the temporal mean $(\bar{T})$, for example, $T' = T - \bar{T}$. Time series of full eddy heat flux agree well between the CPIES and the moored current meter measurements at this same suite of depths through the water column (Fig. 4). This further comparison demonstrates that the covariances between eddy current and temperature are dominated by low-frequency geostrophic variability of large vertical scale rather

![Figure 2](image-url)
than by higher vertical modes or by ageostrophic fluctuations. Just as for Fig. 3, both of these time series follow the varying depth of the moored measurements in order to avoid introducing uncertainties from mooring motion compensation.

Short-period variability in the moored eddy heat fluxes does not contribute substantially to the time-mean eddy heat flux \( u_0 T_0 \). Small-scale temperature features that quickly pass the point measurements are uncorrelated with the currents, which are predominantly geostrophic. For current meter measurements, as well as for CPIES, the fraction of \( u' T' \) covariance in periods shorter than 7 days is only 4%–9%. The \( u' \) versus \( T' \) coherences (not shown) are typically greater than 0.9 for periods longer than 10 days and fall sharply below 0.4 to 0.2 for periods shorter than 7 days. Overall, eddy heat fluxes are mapped and measured well by the CPIES, accounting for 80%–90% of the eddy heat flux variance at all depths.

c. The divergent component of eddy heat flux

Marshall and Shutts (1981) demonstrated that the full eddy heat flux \( \overline{u'T} \) contains a rotational component that, although it may be large, recirculates around contours of temperature variance \( (T')^2 \). As they noted, it is advantageous to remove that rotational contribution because the dynamically important part of eddy heat flux is the divergent component. Cronin and Watts (1996) applied the Marshall and Shutts (1981) method to data from an array of current meters moored beneath the Gulf Stream. Their application removed a large rotational contribution of the eddy heat flux. Additionally, they discussed the dynamical importance of the divergent residual component. Bishop et al. (2013) studied divergent eddy heat fluxes in the Kuroshio Extension using data from an array of CPIES. They showed a natural outcome of expressing the CPIES geostrophic velocities as above in Eq. (1) is that the divergent eddy heat flux arises entirely from joint interaction of the upper baroclinic front and the deep reference current, as we illustrate next.

Referring again to Fig. 2, the velocity component \( u_{bcb} \) is geostrophic and flows along the time-varying front, parallel to geopotential contours \( \phi \). For the cDrake gravest empirical mode lookup table, temperature and geopotential are both functions of \( \tau \) on pressure surfaces, so isotherms are parallel to geopotential contours. Firing et al. (2014, their Fig. 13) show these cDrake gravest empirical mode relationships for \( T, S, \delta, \) and...
versus \( t \). Chidichimo et al. (2014) also illustrate \( \phi \) versus \( \tau \) and \( T \) versus \( \tau \) for several pressure levels. Consequently, the \( \mathbf{u}_\text{ref} T' \) contribution to eddy heat flux is nondivergent:

\[
\nabla \cdot (\mathbf{u}_\text{ref} T') = \mathbf{u}_\text{ref} \cdot \nabla T' = 0.
\]

In contrast, the reference velocity \( \mathbf{u}_\text{ref} \) can advect temperature across the front. Thus, \( \mathbf{u}_\text{ref} \) is perpendicular to \( \nabla T \), and \( \mathbf{u}_\text{ref} T' \) accounts entirely for the divergent contribution of eddy heat flux.

One advantage of this decomposition is that it can be applied to small arrays and isolated instruments. When large fluxes exist at the periphery of a two-dimensional array, ambiguities arise in identifying the divergent component using methods such as those of Eden et al. (2007) or Smith (2008). Determining eddy heat fluxes with \( \mathbf{u}_\text{ref} T' \) captures 100% of the divergence of the full eddy heat flux and substantially reduces the nondivergent contribution without introducing ambiguities by the separation method. This method is not restricted to CPIES; it can also be applied to moorings if they are equipped with a near-bottom current meter.

Because \( \nabla T = \nabla = 0 \) and mass flux is zero, these are heat fluxes, not just temperature fluxes. For comparison with historical data, we present eddy heat fluxes either with dimensional units of \( ^\circ \text{C m s}^{-1} \) or multiplied by \( \rho_o C_p = 4.14 \times 10^3 \) for units of \( \text{kW m}^{-2} \). Here, density \( \rho_o = 1035 \text{ kg m}^{-3} \) and specific heat \( C_p = 4000 \text{ J kg}^{-1} \text{ C}^{-1} \) are representative for the location. The vertical integral from 3500 m near the bottom to the sea surface for quantities like eddy heat flux will be signified by \( \int_z \).

4. Results

a. Mapped fields of mean eddy heat flux

A striking illustration of the different contributions to eddy heat flux can be seen in the 4-yr mean eddy heat flux fields from cDrake (Fig. 5). These fields at 400 m are representative of the strong eddy heat fluxes throughout the upper water column. Although the 4-yr mean fields are temporally stable estimates, the full \( \nabla T' \) field exhibits great spatial variability in both magnitude and direction (Fig. 5a). Individual point measurements within this full eddy heat flux field could invite very different interpretations. Neighboring sites separated only \( \frac{1}{4} \mathbf{e} \) may exhibit qualitatively different full eddy heat fluxes with opposing senses.
The strong baroclinic contribution $\overline{u_{bcb}} T$ (Fig. 5b) can be seen to flow along the contours of the mean $(T)^2$ field. Its rotational nature is in accord with the Marshall and Shutts (1981) prediction and, in this case, happens to produce regions of strong equatorward (albeit rotational) eddy heat flux. Because the full eddy heat fluxes include this rotational component, the orientation of the vectors is spatially variable and confusing. As noted above, it is advantageous to remove this rotational contribution.

The 4-yr mean $\overline{u_{ref}} T'$ field is predominantly down-gradient, across the front, and poleward (Fig. 5c). This dynamically important divergent component of eddy heat flux is strongest in the first 100 km downstream of the cross-channel ridge of seamounts in SFZ, where baroclinic instability processes (discussed below) actively drive growth of meanders and mesoscale eddy variability.

The time-mean eddy heat flux fields are statistically stable because they arise from many short events. Examples of time series of full eddy heat flux with numerous short-term pulses were shown in Fig. 4. Additional time series of meridional $\nu_{ref} T'$, shown in the following subsection, also exhibit many episodic pulses. The relevance for the mean fields is that the integral time scale is short (4 to 6 days). In consequence, statistically stable $\overline{u_{ref}} T'$ field estimates were obtained at most locations in 2- and 3-yr subsets of our 4-yr mean fields (Figs. 6b,c). Even for independent 2-yr subsets the strong poleward $\overline{u_{ref}} T'$ fields are very similar in the lee of the SFZ (Figs. 6d,e). Typical poleward values at 400-m depth are $-0.02$ to $-0.04^\circ$C m s$^{-1}$, equivalent to $-80$ to $-160$ kW m$^{-2}$.

b. Processes that produce the eddy heat flux field: Case studies

Because the field of $\overline{u_{ref}} T'$ vectors are primarily poleward (Fig. 6), we focus here on the meridional component $\nu_{ref} T'$. Time series of $\nu_{ref} T'$ build up their mean values from many short negative events, which can arise from warm water advected southward or cold water advected northward (Fig. 7, note red/blue coloration).

Representative case studies have been chosen in Figs. 7 and 8 to illustrate the responsible processes. The point of the following discussion is that every substantial pulse contributing to the mean poleward eddy heat flux arises where crests or troughs are accompanied, respectively, by deep reference highs or lows tilted ahead of them downstream. These circumstances produce just the right sense of cross-frontal flow $\mathbf{u}_{ref}$, which advects warm water southward or advects cold water northward in the upper baroclinic zone entering a meander crest or, respectively, a trough. One strong, deep cyclogenesis
event in the LDA has been described by Chereskin et al. (2009). Analogous processes of vertical coupling were exhibited in the Gulf Stream (Savidge and Bane 1999), Gulf of Mexico (Donohue et al. 2016), and Kuroshio Extension (Tracey et al. 2012).

The vertical phase offset is a characteristic signature of baroclinic instability (Holton 1979). The three case studies (Fig. 7) described next illustrate the tendency for the coupled upper and deep features to strengthen during events of poleward, downgradient heat flux. Events tend to grow either until neighboring strong eddies disturb the vertical phase offset, halting the release of potential energy, or until the amplitude has grown to the point that the upper and deep pressure centers shift into vertical alignment and rings pinch off.

During March 2009, the first case study, the satellite sea surface height (SSH) image (Fig. 7) shows a large meander crest extending southward over site A03 at the western edge of the LDA and a large meander trough protruding into the central region near site C06. (In the Southern Hemisphere, a meander crest has a southward or poleward displacement, and a trough has northward or equatorward displacement.) The red coloration of this interval in the A03 time series indicates advection of a warm anomaly southward, contributing to poleward $v'_{ref} T'$. The maps reveal that A03 is situated upstream of a deep high pressure anomaly (pink region), where the deep anticyclonic flow (counterclockwise $u'_{ref}$) crosses the upper meander crest segment with a southward component. The red $u'_{ref} T'$ vectors point poleward across the front in the same direction as $u'_{ref}$. During the

![Image of diagrams showing eddy heat fluxes](image-url)
same time period but slightly downstream, the blue coloration of the C06 time series indicates advection of a cold anomaly northward, also causing poleward $\mathbf{u}'_0 T'$. The maps show C06 is situated upstream of a deep low (purple region) where the deep cyclonic currents (clockwise $\mathbf{u}_C$) flow to the northwest crossing the upper meander trough segment. The blue $\mathbf{u}'_0 T'$ vectors also point poleward across the front but in the opposite
direction to $u_{ref}$. Careful examination of the mapped fields reveals that, on each day, the centers of the upper crest and trough are offset from those of the deep high and low, with the deep features slightly downstream of the upper, which is the signature of baroclinic instability. The sequence of maps from 1 March to 13 March shows that the upper and deep features develop jointly over time. For example, as the meander crest grows in amplitude, the deep high intensifies and the poleward heat fluxes increase.

The SSH image for the second case study shows a period during July 2010 when the orientation of the meander crest and trough is reversed; the trough is in the western portion of the LDA, and the crest is in the central region. The $A03$ $u_{ref}T'$ time series is color-coded blue, indicating northward advection of cold temperature anomalies, while the $C06$ record is red, indicating southward advection of warm anomalies. In this case, $A03$ is located just upstream of a deep low (purple region centered below the upper crest-to-trough meander segment) where the cyclonic deep flow is to the northwest. $C06$, on the other hand, is located just upstream of a deep high (pink region centered below the upper trough-to-crest meander segment) where the deep anticyclonic flow is to the southeast. At both locations, the $u_{ref}T'$ vectors point poleward across the upper front, with the blue arrow indicating that the deep current anomaly is in the opposite direction. Again, the mapped fields show the deep high and low pressure anomalies are tilted downstream ahead of the upper crest and trough. That the features jointly steepen and intensify is clearly evident between 16 and 19 July. Subsequently, however, the features translate northeastward beyond the LDA and are no longer in the same configuration.

Fig. 8. As in Fig. 7, but five events are shown as single-day snapshots rather than time sequences.
over sites A03 and C06, so the vectors rotate and the magnitude of eddy heat flux decreases.

The SSH image for the third case study shows a large-amplitude trough occupying the western half of the LDA about one week before a cold-core ring detaches in July 2011. The $v_{ref} T'$ time series at A03 and C06 both indicate rapidly evolving northward cold advection. For the most part, the sites were located west of a deep low pressure anomaly where $u_{ref}$ has a northward component. The $u_{ref} T'$ vectors point across the front, taking heat southward. In the sequence of maps, the upper meander trough steepens and extends outside the LDA. On 18 July, the contours at the southern edge of the LDA pinch together. In the subsequent days (not shown), strong interactions occurred between the meander front and surrounding eddies, resulting in the cold-core ring that pinches off.

Five additional events of poleward eddy heat flux in record A03 are shown as single snapshots in Fig. 8:

(i) 9 December 2007: The map shows the canonical case of upper wave crest and trough with their centers offset from those of the deep high and low pressure anomalies, which lead downstream. A03 is located west of the deep high (pink) on this date, where poleward $u_{ref}$ crosses the upper crest producing a peak of strong southward warm advection. This crest continued steepening and about 10 days later pinched off a warm-core ring (not shown).

(ii) 22 November 2008: A peak of northward cold advection results from a deep low (purple) located to the east of A03. Northward $u_{ref}$ crosses the upper trough at the southern edge of the LDA. This trough continued steepening and then propagated to the northeast outside the instrumented region.

(iii) 4 July 2009: The strongest peak in the A03 record is attributable to southward warm advection. The intense deep high (dark pink) is tilted downstream of the upper wave crest bringing southeastward deep flow to the site. During the subsequent month, the crest intensified and a warm-core ring separated, which later recoalesced with the front (not shown here).

(iv) 19 October 2010: A strong peak of poleward eddy heat flux is attributable to southward $u_{ref}$ around a deep high crossing the warmer waters in an upper wave crest. The meander crest and trough evident in the SSH image steepened and remained nearly in place for the next month. Thus, this stationary phasing produced a relatively long-lasting poleward peak in the $v_{ref} T'$ time series at site A03.

(v) 16 May 2011: A poleward $v_{ref} T'$ peak is attributable to northward cold advection at site A03 by $u_{ref}$ of relatively colder waters in an upper wave trough. The SSH image shows that a cold-core ring had recently separated, although the mapped fields indicate that the separation process had not completed. Subsequently, the trough near A03 grew in amplitude but remained at the western edge of the LDA, resulting in another long-lasting poleward peak.

In the LDA there are many additional poleward heat flux events in each CPIES record, and close inspection (not shown here) reveals that every substantial contribution to the poleward mean $v_{ref} T'$ comes from events whose upper deep phasing is consistent with baroclinic instability.

c. Transect of eddy heat flux across Drake Passage

The profiles of temperature $T(p)$ estimated from measured $r$ were combined with the directly measured, deep reference velocities $u_{ref}$ at each CPIES along the transect spanning Drake Passage. Hence, each site yields estimates of the dynamically important divergent component of eddy heat flux as time series profiles from sea surface to seafloor. We focus on the meridional eddy heat flux component because the mean field vectors are primarily poleward (Figs. 5c, 6). Representative time series of meridional $v_{ref} T'$ at 400-m depth are shown in Fig. 9. Using the same plotting scales for each panel emphasizes how different the strengths are in the northern and southern portions, where mesoscale eddies are stronger and, respectively, weaker. The strongest ACC jets, the SAF and PF, lie primarily in the northern portion.

The entire northern portion of the ACC transect, particularly the interfrontal zone between the SAF and PF, had large-amplitude variations in $v_{ref} T'$, as exemplified by site C06. There is a stark contrast between the $v_{ref} T'$ amplitudes at C09 and C11 in Fig. 9, even though the PF frequently meandered over each site at different times. Foppert et al. (2016) reported that about 60% of the time the PF took a northern path hovering around site C09, and about 40% of the time, the PF took a southern path hovering around site C11. They found that the PF was seldom located near the midpoint where the transect crossed the ridge of SFZ, even though Chidichimo et al. (2014) had determined the mean position of the PF was between C11 and C09. Cunningham et al. (2003) also found a bimodal PF position at the SR1b line downstream of the cDrake line.

The time-mean eddy heat flux profiles $v_{ref} T'$ can be combined in a transect of Drake Passage along the
CPIES line (Fig. 10a). In the right-hand panel, individual vertical profiles of $v_{ref} T'$ are plotted for representative sites. Strong fluxes at northern and interfrontal sites extend throughout the water column, as illustrated for sites C06, C08, and C09 (Fig. 10b). The northern $v_{ref} T'$ records are laterally correlated between sites, with the horizontal correlation scale $\sim 50$ km. In that northern ACC region, approximately half of the eddy heat flux is...
surface intensified in the upper 850–1000 m, and the other half below 1000 m is almost independent of depth. The vertical integral of mean eddy heat flux \( \int_0^H \bar{u}' \bar{T}' \) (orange curve) in these northern portions of the transect was strongly poleward ranging from −70 to −140 MW m\(^{-1}\). The strongest values occur in the interfrontal region between the mean PF and SAF.

South of the fracture zone (58.5°S, at which C10 was located), the \( \bar{u}' \bar{T}' \) profiles are nearly uniformly distributed with depth. With magnitudes less than 5 MW m\(^{-1}\), they are not, however, statistically different from zero. In the gap between C11 and C13, only a 1-yr-long record was available for site C12, so it was excluded from Fig. 10.

At least three interlinked aspects contributed to the weakness of \( \int_0^H \bar{u}' \bar{T}' \) throughout the southern half of this ACC transect: south of the PF and below 250 m, weak horizontal temperature gradients lead to small temperature signals. The rms deep currents are relatively weak between the PF and the Southern ACC Front (SACCF; Chereskin et al. 2012). The upper mesoscale eddy variability is weak upstream of SFZ, suggesting the absence of baroclinic instability processes with their associated downgradient heat fluxes. Consequently those sites produced insignificant mean values, and the \( \bar{u}' \bar{T}' \) time series at southern ACC sites were not laterally correlated. We will treat \( \int_0^H \bar{u}' \bar{T}' \) as zero everywhere south of SFZ on this transect.

\[ d. \text{Transient and stationary eddy heat transport in LDA} \]

In a zonal sector, the zonal and temporal average eddy flux can be expressed as the sum of two zonal averages: the zonal mean contribution of the transient eddy deviations from the time mean, plus the (separate and independent) contribution of stationary eddy deviations from the zonal mean field (Peixoto and Oort 1992). For atmospheric studies the zonal mean is usually circum-global. However, one can alternatively average over shorter zonal intervals, such as through the wavelength of a stationary meander, as we will do here for our observations in the LDA. The zonal average is signified by square brackets \([\cdot]\), with departure by an asterisk \((\cdot')\). The LDA spanned a stationary mean wavelength, and we calculate the zonal mean of stationary meridional eddy heat transport that arises from the southward or northward flow of slightly warmer or cooler water. Writing the vertical integral explicitly, we follow the methods of Peixoto and Oort (1992, 62–63) and compute the transient and stationary eddy heat transports:

\[
\rho C_p \left[ \int_z (\bar{u}' \bar{T}') \right] \quad \text{and} \quad \rho C_p \left[ \int_z (\bar{u}' \bar{T}') \right].
\]

Here, the left-hand term is the zonal average of transient eddy heat transport, and the right-hand term is the stationary mean meander heat transport. (As with the time average of \( \bar{u}' \bar{T}' \) being zero, the zonal average \( \int_z \bar{u}' \bar{T}' = 0 \), so these are both legitimate mass-balanced heat transports because the net volume flow is zero.) The transient eddy heat transports (left term) sum to −0.005 to −0.014 PW at various latitudes in the LDA. The stationary eddy heat transports (right term) sum to −0.003 to −0.006 PW at different latitudes across the LDA. Their maxima do not coincide, and together they sum to −0.008 to −0.018 PW.

5. Discussion

The literature lacks consensus regarding the relative importance of eddy fluxes and mean flow contributions to Southern Ocean heat transport. Apparently conflicting statements arise, in part, from different methodological approaches and different circumpolar paths of integration as well as from advancement in the interpretation of the temperature fluxes \( \bar{u} \bar{T} \). To reconcile the different viewpoints calls for recognition that eddy and mean temperature transport processes act locally and the conveyance of heat poleward in the ACC arises as a “hand off” between one process and locality to another further along stream as heat completes its crossing to the Southern Ocean.

In the following subsections we (i) consider metrics for gauging Southern Ocean heat transports, (ii) compare our eddy heat flux findings with earlier published results, (iii) examine the relative importance of transient and stationary eddy heat transports, and (iv) distinguish among physical processes driving mean heat transports.

\[ a. \text{Metrics to gauge heat transports} \]

To gauge the relative sizes of ocean eddy and mean heat transport, we offer as one metric the ocean-to-atmosphere heat loss in the Southern Ocean. One can infer from Large and Nurser (2001) a range of estimated total air–sea flux that is roughly 25 W m\(^{-2}\) on average in the Southern Ocean. Note that a global balance adjustment of 16 W m\(^{-2}\) has been applied in this estimate, which indicates \( O(70)\% \) uncertainty. Therefore, a metric with one significant figure is representative and useful. If we multiply 25 W m\(^{-2}\) by the approximate ocean surface area south of 60°S, \( 16 \times 10^{12} \text{m}^2 \), this implied net loss (−0.4 PW) should be offset by ocean net poleward heat transport. For comparable values we cite plots of the total ocean heat transport poleward at 60°S of −0.4 PW in a comprehensive overview of global ocean heat transport by Macdonald and Baringer (2013). A thorough Southern Ocean analysis conducted by Volkov et al. (2010) of output from a data synthesis model, Estimating the Circulation and Climate of the
Ocean (ECCO2), produced values of $-0.3$ PW at 60°S to $-0.4$ PW at 55°S. Gordon (1975) estimated $-0.4$ PW south of the PF. The Ekman and meridional overturning circulation (MOC) zonal average contribution to heat loss at 60°S, as estimated in Volkov et al. (2010), is only about $-0.03$ PW. Hence, our one-digit metric for Southern Ocean heat loss can remain $-0.4$ PW, to be balanced by poleward eddy and mean heat transport.

This metric for net poleward heat transport can be reexpressed in two ways: the average vertically integrated heat flux per unit length and the average heat flux. This gives us the flexibility to discuss the two- and three-dimensional structure of temperature fluxes in a coherent manner. Using the $-0.4$-PW metric, along the 20,000-km circumpolar path, all ocean heat flux processes combined are balanced by poleward eddy and mean heat transport.

Another useful metric for the mean contribution from ocean eddies was suggested by Johnson and Bryden (1989) who noted the quasigeostrophic relationship between the downward flux of eddy momentum and poleward eddy heat flux. They argued that this interfacial form stress carries the wind stress to the seafloor:

$$
\rho_o f \frac{\overline{vT}}{z} = \overline{fT} \quad \text{or} \quad \rho_o C_p \frac{\overline{vT}}{f} = \frac{C_p \overline{fT}}{f},
$$

where $f$ is the Coriolis parameter, and $\overline{vT}$ is the mean vertical gradient of potential temperature, which we approximate as $0.57 \times 10^{-3} \text{Cm}^{-1}$. By this relation ocean eddies should supply at least $-3.8 \text{kW m}^{-2}$ heat flux to balance the input of wind stress: $\overline{fT} \sim -0.2 \text{N m}^{-2}$. Allowing that uncertainty of $O(50\%)$ exists in this estimate due to circumpolar variability in stratification and wind stress suggests that eddy heat flux should contribute, on average, at least $-2 \text{KW m}^{-2}$. The suggested lower bounds for eddy contributions to the average vertical and circumboreal heat transport integrals would be $-8 \text{MW m}^{-1}$ and $-0.16 \text{PW}$, respectively.

In summary, we have presented two metrics: one for the total ocean heat transport that includes both eddy and mean contributions based upon ocean-to-atmosphere heat loss and one derived from quasigeostrophic relationships that provide a minimum eddy contribution. We expressed each metric equivalently as the total zonal sum, the average flux per meter along the ACC path, and the average flux per unit area of circumboreal section:

- total poleward heat transport, $-0.4$ PW, $-20 \text{MW m}^{-1}$, and $-5 \text{kW m}^{-2}$; and
- minimum eddy contribution, $-0.16$ PW, $-8 \text{MW m}^{-1}$, and $-2 \text{kW m}^{-2}$.

Note that the minimum eddy contribution is $\sim 40\%$ of the total.

### b. Profile of eddy heat flux compared to historical estimates

The observed strong spatial variability in the eddy heat flux fields shown in Figs. 5a and 5b suggests caution in the following comparisons with other locations because we now recognize that values could change magnitude, and even sign, in small spatial distances. Note that this spatial variability arises from the inclusion of the rotational eddy heat flux field (Fig. 5c) as well as from heterogeneity in the divergent eddy heat flux $\overline{vT}$ (Fig. 5b). Vertical profiles of the meridional component of $\overline{vT}$ are repeated with units $\text{KW m}^{-2}$ and plotted together with historical estimates in Fig. 11. Our values in the polar frontal zone between the SAF and PF range from $-80$ to $-140 \text{KW m}^{-2}$ near the surface, weakening to $-10$ to $-25 \text{KW m}^{-2}$ deeper than about 1000 m. Previous studies have reported values from Drake Passage as well as from south of Australia and New Zealand at locations indicated on the lower panels of Fig. 11.

Each of the sparse historical measurements in Fig. 11 has built up observational expertise, developed analysis methodology, and contributed toward understanding eddy fluxes in the ACC:

(i) Drake Passage: Bryden (1979) reported on eddy heat flux estimates of $-6.6 \text{KW m}^{-2}$ averaged across six sites from current meters at 2700-m depth on moorings that spanned Drake Passage near Phoenix Ridge for a year; he found also a significant value, $-12.3 \text{KW m}^{-2}$, at 1520 m on one mooring. Nowlin et al. (1985) analyzed full eddy heat fluxes on current meter moorings on Phoenix Ridge in Drake Passage (sites for both studies are shown in Fig. 1). All their significant values in the 2- to 90-day band (90-day low-passed coordinate system) are grouped and averaged at depths of 800 m ($-38 \text{KW m}^{-2}$), 1400 m ($-3.9 \text{KW m}^{-2}$), and 2600 m ($-4.3 \text{KW m}^{-2}$) in Fig. 11. Walkden et al. (2008) reported current meter measurements in Shag Rocks Passage, a channel through the North Scotia Ridge at the east end of Drake Passage. They found 2- to 90-day band-passed, filtered, cross-front, poleward eddy heat flux of $-12 \pm 5.8 \text{KW m}^{-2}$ at average depth 2800 m. Lenn et al. (2011) analyzed eddy heat fluxes in the upper water column from 50 shipboard ADCP and XBT transects across Drake Passage. They found full eddy heat fluxes generally poleward and surface intensified in all three ACC fronts, with values up
to approximately $-290 \text{ kW m}^{-2}$. Ferrari et al. (2014) reported on nine moorings with three levels of current meters for 2 or 3 yr on a transect spanning Drake Passage, slanting 200 to 400 km downstream of the SFZ (sites shown in Fig. 1). While more than half of their records had insignificant eddy heat flux, all significant values (in 90-day low-passed coordinates) were poleward; a northern slope value was $-71 \text{ kW m}^{-2}$ at 60 m, two midchannel records near 900 m averaged $-12 \text{ kW m}^{-2}$, and four midchannel records near 2500 m averaged $-12.5 \text{ kW m}^{-2}$.

(ii) South of New Zealand: Bryden and Heath (1985) reported eddy heat flux estimates from 2-yr moorings southeast of New Zealand at the confluence of the northern ACC and the subtropical gyre circulation, where eddy variability is highly energetic. Their values ranged from $-35 \text{ kW m}^{-2}$ at 1000 m to $+0.2 \text{ kW m}^{-2}$ at 2000 m, but none of the $vT$ were statistically significant.

(iii) South of Tasmania: Phillips and Rintoul (2000) reported on four current meter moorings with 3 to 5 levels instrumented from 420 to 3320 m in the SAF south of Australia for 2 yr. They grouped their best estimated values, as calculated in shear coordinates and band-passed filtered between 2 and 90 days, ranging from $-34.8 \text{ kW m}^{-2}$ at 420 m down to $-2.1 \text{ kW m}^{-2}$ at 2240 m, as plotted in Fig. 11.
(iv) Fawn Trough, Kerguelen Plateau: Sekma et al. (2013) found negligible eddy heat fluxes inside the narrowly constrained Fawn Trough channel at Kerguelen Plateau. They focused instead upon mean temperature transport and mean current angling across a bottom slope.

Plotted together, as they are in Fig. 11, on a scale that includes near-surface values, the range among deeper measurements looks small. Upon closer inspection, there is a factor of 5 or greater range among the estimates below 1000 m. The wide range of variation in the above estimates of eddy heat flux arises for three main reasons: the divergent eddy heat flux field differs greatly from the full eddy heat flux in its spatial structure (Fig. 5), the full eddy heat flux field estimated by previous studies can change sign and magnitude in short lateral distances, and fluxes change substantially with depth.

Most of the above values exceed in magnitude the metric value of $-5 \text{ kW m}^{-2}$, indicated by the vertical dashed line, which indicates the required average heat flux if it were all supplied by eddies. Many estimates greatly exceed that metric. Several authors in the above studies have made the case that the eddy heat flux estimates, if representative of the circumpolar ACC, are more than enough to balance Southern Ocean heat loss to the atmosphere. Our divergent eddy heat flux estimates could arguably more than meet that criterion; however, recognizing now the great variation in EKE, however this peak weakens sharply poleward to relative insignificance (<+0.03 PW) at and south of 60°S.

The zonal average MOC around the Southern Ocean comprises northward surface Ekman transport, outflow of Antarctic Bottom Waters below ridge depths where geostrophic transport can exist, and middepth balancing ageostrophic transports of southward Circumpolar Deep Water and northward Intermediate Waters (Speer et al. 2000). Volkov et al. (2010) show the latitudinal dependence of heat transport associated with the MOC reaches a strong northward peak (+0.8 PW) at 45°S; however, this peak weakens sharply poleward to relative insignificance (<+0.03 PW) at and south of 60°S.

An early conceptual hydrographic study regarding the ACC concluded that eddies, not the mean, should supply most of the poleward heat transport (de Szoeke and Levine 1981). They based this upon the observation that a path defined by vertical average temperature equal to 2°C approximately follows a streamline in the middle of the ACC, and if $\mathbf{u} \cdot \nabla T$ were near zero everywhere along that streamline, the mean horizontal heat transport would be insignificant.

In contrast, Sun and Watts (2002) concluded that mean baroclinic flow along ACC transport streamlines could, in principle, account for $-0.14 \text{ PW}$, about 35% of the above metric for Southern Ocean heat loss. They used the Olbers et al. (1992) compilation of circumpolar hydrographic data in streamfunction projections to...
estimate profiles of temperature and baroclinic velocity. They also confirmed the de Szoeko and Levine (1981) prediction of insignificance of mean transport across the 2°C circumglobal path. Then, choosing instead zonal transects such as 56°S, the flow along mass transport streamlines crosses northward and southward in its global stationary meander path. Owing to the northward/southward component of mean along-stream flow, an equal mass transport of cold water passes northward into the Atlantic and relatively warmer water returns southward into the Pacific, producing the above substantial mean poleward heat transport.

The existence of along-stream temperature changes that produce the mean poleward heat transport noted by Sun and Watts (2002) is indicative that cross-stream convergent or divergent heat fluxes are vital to raise or lower the middepth water temperatures. Eddy and mean fluxes act collaboratively to produce that heat transport. Several recent papers (discussed below) emphasize that the turning of the mean along-current stream with depth is another mechanism for transporting heat across stream. Vertical turning can arise as noted in Fig. 2 and would carry different temperatures at different depths across stream. Turning, in the mean, occurs through interaction with topography on both large and small scales.

At Fawn Trough, Sekma et al. (2013) found mean turning with depth associated with bottom currents angling across sloping topography of the channel. Their observed mean transports should be identified, however, as temperature transport, not heat transport, because the mass transport was not zero. In the cDrake LDA, we calculated mean temperature transports \( \tau_p T \), where \( \tau_p \) is perpendicular to the mean front, in the same manner as Sekma et al. (2013). We find mean temperature fluxes of similarly large magnitude (from \(-1350\) to \(+740\) kW m\(^{-2}\) at 400-m depth). The sign changes at different locations according to direction of the strong, mean, deep recirculation as mapped in Chereskin et al. (2012). The cross-front mean flow occurs without requiring a deep channel. There is a partial cancellation of these mean temperature fluxes across the LDA, but in order to interpret these values as a mean heat flux, we would need a circumglobal integration to confirm that there is zero mass transport.

Ferrari et al. (2014) found mean vertical turning in their observations in Drake Passage and used the ORCA numerical model (\(1/6^\circ\)) to analyze mean horizontal heat flux. They highlighted the role of bottom topography to produce the mean turning on small horizontal scales. The ORCA12 model-estimated heat transports were poleward, \(-0.31\) PW between the SAF and PF, decreasing sharply southward to \(-0.08\) PW across the PF and \(-0.04\) PW near the SACCF.

Peña Molino et al. (2014) used the Southern Ocean State Estimate (SOSE) model (\(1/6^\circ\)) and computed the direction of the geostrophic velocity vector as a function of depth relative to the surface flow within the circum-polar envelope of the time-mean ACC. They separated the flow into baroclinic contributions relative to the bottom and a deep barotropic reference component, which typically is not aligned with the baroclinic component. Relative to the surface streamline, the cross-stream components of barotropic and baroclinic flow are by construction equal and opposite at the surface, and their vertically integrated cross-stream transports partially cancel, with baroclinic cross-stream transport being about 30% smaller. Moreover, the mean baroclinic shear itself exhibited turning with depth, particularly in regions at the edges of jets, where the along-stream flow was less than 0.02 m s\(^{-1}\). The effect was systematic in the broad basins where bathymetry rose or fell very gradually; the vertical integral of baroclinic flow turned relative to the surface shear and typically contributed, respectively, equatorward or poleward transport of magnitude 10 m\(^2\) s\(^{-1}\). While the equivalent vertically averaged current in 4000-m depth is only 10/4000 = 0.0025 m s\(^{-1}\), the circumpolar sum along streamlines is not negligible, contributing between 30% and 50% of the amplitude of the baroclinic cross-stream transport. The total sum of geostrophic poleward barotropic and smaller equatorward baroclinic transport, \(-5\) to \(-20\) Sv (1 Sv = \(10^6\) m\(^3\) s\(^{-1}\)) on various streamlines, exactly balances Ekman and ageostrophic residual mean circulations. Cross-stream mean temperature transport arises because different temperatures are advected cross stream at different depths. The sum of geostrophic cross-stream temperature transports is poleward \(-0.2\) PW at the northern SAF and decreases in magnitude smoothly southward across the ACC to near zero south of the PF. As Peña Molino et al. (2014, p. 8024) summarize, “These temperature transports by the time-mean geostrophic flow are small compared to the temperature transports [at lower latitudes] by the Ekman flow (equatorward) and transients (poleward), but comparable to the residual between Ekman and transients, hence an important component of the heat transport.”

6. Summary

Eddy heat fluxes were estimated from CPIES observations in a well-resolved local dynamics array (LDA) and on a transect spanning Drake Passage. The LDA was centered on the interfrontal zone between the Polar Front and Subantarctic Front, where satellite altimetry and repeat shipboard ADCP measurements (Lenn et al. 2007; Firing et al. 2011) indicate high EKE. The 4 yr of
measurements produced statistically stable eddy heat flux estimates, judged by the integral time scale (4 to 6 days) producing abundant degrees of freedom and confirmed by the close resemblance of fluxes calculated from 2- and 3-yr subsets of measurements.

The horizontal currents in the ACC turn with depth because the deep geostrophic reference current vector \( u_{\text{ref}} \) can be as large as 9 to 14 cm s\(^{-1}\) rms throughout the LDA, of which a substantial component can cross the upper baroclinic current relative to the bottom \( u_{\text{bcb}} \). CPIES directly each measure \( u_{\text{ref}} \), and maps of the \( \phi, T \) and \( u_{\text{bcb}} \) fields are obtained from measured \( \tau \), at half-day intervals, from surface to bottom. The \( u_{\text{bcb}} \) component flows parallel to isotherms and contributes nondivergent (rotational) heat fluxes. The dynamically important divergent component of eddy heat flux is contributed by \( u_{\text{ref}}T \).

The mean field of \( u_{\text{ref}}T \) has large spatial variability; it is strong in the lee of the Shackleton Fracture Zone and directed across-front poleward at almost all of our measurement sites. This mean field is built up from many short pulses of eddy heat flux produced by baroclinic instability. That is to say, these events in which a deep pressure anomaly (high or low) was tilted ahead downstream of, respectively, an upper meander crest or trough. That characteristic phase offset is the signature of baroclinic instability, for which the deep reference current crosses the upper baroclinic front and releases available potential energy via poleward downstream of eddy heat flux.

In the northern half of our ACC transect, the poleward component of divergent eddy heat flux \( u_{\text{ref}}T \) is strong, arising from frequent events of baroclinic instability that appear to be triggered in the polar frontal zone, which lies in the lee of the SFZ. The southern half of the transect crosses upstream (east) of SFZ and in a region of much lower EKE, at which sites \( u_{\text{ref}}T \) is weaker by nearly two orders of magnitude and not significantly different from zero.

Vertical profiles show eddy heat flux is distributed throughout the water column, with 50% surface intensified above 1000 m, and the balance almost uniformly distributed with depth below 1000 m. North of SFZ, the vertical integral of divergent eddy heat flux \( \int_{z} u_{\text{ref}}T \) ranges between \(-70\) and \(-140\) MW m\(^{-1}\).

The vertical and zonal integral of poleward transient eddy heat transport across the 250-km length of the LDA is \(-0.010 \pm 0.005\) PW, equivalent to average flux \(-10 \pm 5\) kW m\(^{-2}\). The stationary meander contribution in the LDA adds \(-0.004 \pm 0.001\) PW (with maxima at different latitudes), summing to \(-0.013 \pm 0.005\) PW. Thus, in this segment of the ACC, whose length is about 1.2% of the circumglobal path of the ACC, its eddy heat transport accounts for 3.3% of a metric, \(-0.4\) PW, representing total oceanic heat loss to the atmosphere south of 60\(^{\circ}\) S.

For comparison, the metric \(-0.4\) PW corresponds to average vertical integral heat flux \(-20\) MW m\(^{-2}\) for all ocean processes along the 20000-km ACC path. The corresponding metric average heat flux over the 4000-m depth is \(-5\) kW m\(^{-2}\).

While one cannot extrapolate circumglobally from local measurements in such a spatially varied ACC, the eddy heat flux values are in good accord with the few other historical measurements, particularly when one accounts for the host of different depths and approaches. There are important dynamical/interpretation differences between full or divergent eddy heat flux, historical developments of methods back through time, or working in shear coordinates, slowly varying (90-day low-pass filtered) coordinates or geographic coordinates, with or without mooring motion compensation, and attempting to estimate vertical average heat fluxes from rather limited depths.

Eddy heat fluxes are likely to be concentrated in 4 to 10 hot spots of eddy-driven exchange around the ACC, such as Drake Passage. Thompson and Sallée (2012) use AVISO sea level anomaly fields to estimate the zonal distribution of numbers of Lagrangian particles that cross the SAF and PF, illustrating peaks downstream of six major topographic features. De Souza et al. (2013) parameterize eddy diffusivity and estimate the mean lateral gradients across the ACC using satellite altimetry, also illustrating six peaks of eddy heat flux. Abernathey and Cessi (2014) found divergent eddy heat fluxes calculated using the Southern Ocean State Estimate (Mazloff et al. 2010) were an order of magnitude larger at four locations downstream of topography than elsewhere around the ACC. In these examples, amplitudes vary greatly among peaks, and the region where peak exchange occurs extend zonally 2 to 5 times longer than the stationary meander observed by our LDA. Using these as a guide, one might “scale up” the LDA eddy transport measurements by a factor of 6 to 20 to obtain a circumglobal estimate for hot spots.

Our well-estimated vertical and zonal integral divergent eddy heat transport through a stationary meander suggests, applying the above reasoning, that six hot spots of eddy-driven exchange around the ACC might account for as little as 20% to about 70% of the metric, that is, approximately \(-0.08\) to \(-0.28\) PW. Recall that the wind stress and quasigeostrophic interfacial form stress balance suggests a lower bound of \(-0.16\) PW for the eddy contribution to heat flux. The balance of ocean poleward heat transport should come from weak eddy and mean cross-front transports distributed along
the remaining 75%–90% of the ACC path. Alternatively, the metric of −0.4 PW (traced to air–sea heat fluxes with 70% adjustment factors) may be too high.

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