Mean Circulation of the Coral Sea

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Key points:

- Argo float drift references Coral Sea climatological geostrophic currents
- Interior and boundary current transports largely explained by the Island Rule
- Three distinct pathways carry the SEC through the Coral Sea to the equator

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Abstract

The mean absolute geostrophic circulation of the Coral Sea is constructed from climatological hydrographic data referenced to a 1000m-velocity field derived from Argo float drift. Two branches of the South Equatorial Current (SEC) enter the Coral Sea between New Caledonia and the Solomon Islands: the broad, upper-thermocline North Vanuatu Jet (NVJ), and the narrow North Caledonian Jet (NCJ) extending to at least 1500m. Most of this incoming flow leaves to the Solomon Sea. Four distinct pathways through the Coral Sea are traced by their water properties: (1) The NCJ crosses the Sea to the coast of Australia and turns north at densities sigma 25-27.4 as the main source of the Gulf of Papua (GPC) western boundary current, eventually feeding the New Guinea Coastal Undercurrent; (2) Part of the shallow NVJ turns into the Solomon Sea in mid-basin, carrying high-salinity water above sigma 25.5; (3) Another part of the NVJ continues to Australia, then turns north to join the GPC, extending it to the surface; (4) A shallow finger of NVJ water, traced by low oxygen above sigma 25, turns south along the coast, beginning the East Australian Current (EAC) at 15°S. Total transport from the Coral to the Tasman Sea is small and shallow; instead, most of the EAC is fed from south of New Caledonia, consistent with the Island Rule. However, large transport fractions occur in narrow jets close to coastlines and reefs and are not well sampled, precluding a quantitative estimate of meridional redistribution of the incoming SEC.
1. Introduction

Because it sits at a bifurcation where subtropical South Pacific waters split into equatorward and poleward branches, the Coral Sea has been the subject of organized surveys and analyses for more than 40 years. These studies have shown that the South Equatorial Current carries about 30 Sv west from the subtropical gyre into the Coral Sea between New Caledonia and the Solomon Islands (Fig. 1). This flow bifurcates at or before the coast of Australia, turning north into the Solomon Sea and to the equator, and south into the East Australian Current, though the distribution has been controversial. The present work continues that effort, following its clearest predecessor, Qu and Lindstrom [2002]. Although the earlier findings are broadly confirmed here, the advent of the Argo float array [Roemmich et al., 2009] provides a new piece of information that resolves a critical shortcoming of previous works and makes the study worth revisiting: mid-depth trajectories that allow geostrophic velocities to be referenced. (In addition, Argo has approximately doubled the number of hydrographic profiles in the region). Since it was apparent from very early on that the Coral Sea supports currents of great vertical extent [Wyrtki, 1962; Huang and Qiu, 1998; Sokolov and Rintoul, 2000], the choice of appropriate reference level has been a principal focus of much previous work. The ability to specify absolute currents at the Argo parking depth also emphasizes the narrow, jet-like nature of much of the flow entering and
crossing the Coral Sea, currents that previously could only be broadly delineated in compilations of in situ observations.

Recently, the Coral Sea has been the object of renewed interest, partly in connection with tropical Pacific decadal variability theories. The meridional circulation transporting water masses from the southern subtropics to the equator largely transits through the Coral and Solomon Seas, thus an accurate description of the pathways and transformations of water masses in the Coral Sea is of great importance to tropical climate on the basin scale. A credible diagnosis of the Coral Sea circulation is also necessary to understand flows reaching the Tasman Sea from the subtropics [Hill et al., 2011]. In this context, the international research program SPICE (Southwest Pacific Ocean Circulation and Climate Experiment; Ganachaud et al., [2007]), has begun a series of focused observational efforts to describe specific current systems that are part of the Coral Sea circulation [Gourdeau et al., 2008; Ganachaud et al., 2008; Gasparin et al., 2011; Cravatte et al., 2011; Chroukroun et al., 2010; Hristova and Kessler, 2012; Gasparin et al., 2012; Davis et al., 2012].

*Analyses of Coral Sea cruise data: The evolution of ideas*

Early work in the Coral Sea identified issues that continue to be incompletely understood: the unusually-thick currents and their subsurface maxima, and the question of the proper reference
level to use for geostrophic calculations. Overall, the largest question continues to be the
destination of water carried by the South Equatorial Current when it arrives at the coast of
Australia. In this section we review some key papers interpreting Coral Sea circulation using
in situ hydrographic observations, focusing on the evolution of ideas.

Although there had been scattered observations of Coral Sea hydrography during the 1950s
[Rochford 1960], the earliest organized surveys were conducted around 1960 by ORSTOM
[Donguy et al., 1970] and CSIRO [Wyrski, 1962]. Wyrski [1962] attempted the first synthesis of
these observations, calculating geostrophic flow relative to 1750m and noting several features
that remain valid: the large westward inflow to the Coral Sea between Vanuatu and the Solomon
Islands, the great depth extent of many of the currents, and the general tendency to subsurface
maxima of the westward currents. However, researchers at the time did not appreciate the
possibility of narrow jets carrying substantial transport, and Wyrski’s surveys apparently did not
sample close enough to New Caledonia to observe the North Caledonian Jet, a narrow current at
the northern tip of New Caledonia that carries about one-third the total Coral Sea inflow
[Gourdeau et al. 2008]. Wyrski [1962] also stated that virtually all the westward inflow turned
south into the East Australian Current, with almost no transport into the Solomon Sea (indeed, he
made no attempt to calculate this transport); this has since been shown to be incorrect.
Donguy et al. [1970] included additional surveys carried out by ORSTOM and also examined salinity on the isanostere 340 cl/ton (close to sigma 24.5, at a depth of about 150m in the Coral Sea, and near the subtropical salinity maximum layer). This field provided Donguy et al. an important clue that there was, in fact, a bifurcation of the inflow with a substantial transport northward into the Solomon Sea that could feed the equatorial current system. Like Wyrtki, however, Donguy et al. overestimated the importance of direct (frictional downwind) wind driving and assumed that the upper layer flow was entirely southeastward during austral summer when the trade winds are weak or reversed over the Solomon and northern Coral Seas. A full appreciation of the key role of wind stress curl forcing and Sverdrup (vorticity) dynamics did not inform interpretation of the Coral Sea circulation until several decades later.

Scully-Power [1973] diagnosed Nansen bottle data from a series of cruises by the Royal Australian Navy that attempted to systematically sample the Coral Sea during winter seasons of 1966-71. Confirming Donguy et al. (and presaging conclusions here), Scully-Power found that most of the water entering the Coral Sea from the east between New Caledonia and the Solomon Islands (30 Sv geostrophic flow relative to 1500m in his calculation) left the region to the north, with relatively little flow southward across 20°S. Also like the present work, he found only weak circulation in the region between New Caledonia and the coast of Australia, with the East
Australian Current mainly fed by westward currents from south of New Caledonia. Scully-Power described the northward western boundary current off the coast of north Queensland (the North Queensland Current), and suggested that the current turned offshore in the Gulf of Papua, but was unable to account for the very complex, small-scale flow in the Gulf itself.

Thompson and Veronis [1980] attempted an inverse calculation [Wunsch 1978] based on 24 bottle casts on a cruise describing a closed box from Australia along 30°S to 170°E, then diagonally across the Coral Sea to Papua New Guinea. However, they found that they could not usefully describe the Coral-Tasman Sea connection, nor the EAC, for two reasons that continue to be impediments today: the narrow boundary currents carrying a large fraction of the transport that are not easily sampled by shipboard hydrography or Argo floats, and the extreme eddy variability, especially in the EAC.

Andrews and Clegg [1989] invoked a baroclinic normal mode fit to partial CTD profiles on a series of Coral Sea sections, using an inverse procedure to conserve total (top-to-bottom) volume flux and minimize abyssal velocities. The advantage of this approach is that it avoided the need to choose a reference level for geostrophic flow, but it also had the disadvantages of assuming (a) that observations from a single cruise represented the steady circulation; (b) that abyssal velocities were in fact small; and (c) that they had adequately sampled the narrow jets that we
now know carry much of the transport. However, the method entirely missed the North Caledonian Jet (in fact, their scheme showed *southeastward* flow exiting the Coral Sea between Vanuatu and New Caledonia), and implied a total of 10 Sv of the entering 24 Sv ending up in the Tasman Sea.

Two recent observational studies studied data from single cruises taking high-quality CTD profiles to at least 2000m and came to opposite conclusions about the destination of the water flowing into the Coral Sea. *Sokolov and Rintoul* [2000] analyzed profiles from the WOCE P11S cruise roughly along 154°E (that is, near the western boundary of the Coral Sea) in mid-1993. They found an astonishingly-large transport of 54 Sv westward (relative to 2000db) in the Coral Sea branch of the South Equatorial Current. Since this inflow can only exit eastward through the closed section itself, this implied that about 29 Sv flowed south in the East Australian Current into the Tasman Sea (with the remaining 25 Sv going into the Solomon Sea). On the other hand, *Gasparin et al.* [2012] studied a 2007 cruise that closed a triangular region touching New Caledonia, Papua New Guinea and the Solomon Islands. An inversion based on this cruise’s data showed a transport of 30 Sv entering the Coral Sea above 1300db, of which 29 Sv turned north into the Solomon Sea [*Gasparin et al.*, 2012]. Trying to resolve this discrepancy, *Kessler and Cravatte* [2013] used satellite altimetry to estimate that transients (mesoscale eddies and weak El
Niño conditions) contributed about 1/3 the WOCE P11S westward transport. Considering the
vigorous eddies revealed by altimetry, they concluded that a single cruise may not be a useful
guide to the background circulation.

These analyses of scattered hydrographic sections had established that substantial Coral Sea
water turned north into the Solomon Sea. Tsuchiya [1981] put this observation into basinwide
context by showing that water making up the eastern equatorial Pacific thermostad (13°C water
of relatively high salinity spanning the equator below the main thermocline) must come from the
Tasman Sea. He traced a schematic pathway from the outcrop of the isanostere 160cl/ton (about
sigma-theta 26.4) northeast of New Zealand, circulating around the subtropical gyre, through the
Coral and Solomon Seas to the southern "Tsuchiya Jet" (Subsurface Countercurrent), whence it
flowed east along about 5°S. This suggested that the Coral Sea route was not just a regional
phenomenon but a significant part of the makeup of the tropical Pacific.

Tsuchiya's [1981] paper was the catalyst for the WEPOCS experiments in the mid-1980s
[Lindstrom et al., 1987; Butt and Lindstrom, 1994]. WEPOCS made detailed water property and
velocity observations in the Solomon and Bismarck Seas and in the three straits that form its
northern exits (Vitiaz and Solomon Straits, and St. George’s Channel; Fig. 1). They found
[Tsuchiya et al., 1989; Fine et al., 1994] that the pathway to the equator followed the subsurface
western boundary current in the Coral Sea observed by *Church and Boland* [1983] and *Church* [1987], now known as the Great Barrier Reef Undercurrent (GBRUC). Comparing water properties in the Solomon Sea and in the Equatorial Undercurrent, *Tsuchiya et al.* [1989] found that southern hemisphere water at the core of the Equatorial Undercurrent (near sigma 25) was fed both through Vitiaz St and from the open Pacific north of New Ireland. Water comprising the thermostad below the EUC (near sigma 26.4) was clearly of Coral Sea origin.

*Qu and Lindstrom* [2002; hereafter QL02] is the predecessor to the present work. They gridded available hydrographic data along isopycnals to produce mean fields of temperature, salinity and oxygen on a 0.5° by 0.5° grid. To maximize the data coverage, QL02 computed dynamic height and geostrophic currents relative to 1200m, and presented a description of the circulation in approximately the same region as discussed here. They described the sheared current system across the region, with subsurface maxima of the westward South Equatorial Current and overlying eastward flow south of about 20°S. They emphasized that the northward western boundary current feeding the Solomon Sea begins subsurface near 22°S and flows continuously into the NGCU. All these conclusions presage the present work and are largely confirmed here.

There are two main differences in the approach taken here: first, we construct absolute velocity fields based on Argo trajectories to reference the geostrophic currents, which generally increases
the estimated speeds and shows the jets to be narrower than could be resolved earlier; second, the
accumulation of Argo temperature and salinity profiles has greatly increased the data coverage in
the Coral Sea. Specific similarities and differences in the conclusions will be noted below.

Consistent with earlier work [Tsuchiya 1981; QL02], we will demonstrate the dynamical and
property unity of the equatorward western boundary current beginning about 25°S and flowing
continuously around the Gulf of Papua and into the Solomon Sea, accumulating mass and water
properties from the zonal jets arriving from the east. It is thus confusing that this fundamental
piece of the basin-wide circulation has been given a series of four local names. From 19°S to
about 15°S along the coast of Australia where the northward flow is subsurface, it has been
called the Great Barrier Reef Undercurrent (GBRUC; [Church and Boland, 1983]). From its
surfacing to Torres St. it is usually called the North Queensland Current (NQC; [QL02; Ridgway
and Dunn, 2003]). The eastward limb along the south coast of Papua New Guinea has been
referred to as the Hiri Current [Burrage, 1993; Schiller et al., 2009; Cravatte et al., 2011], but its
eastern part along the Louisiade Archipelago has also been called the New Guinea Coastal
Undercurrent [Sokolov and Rintoul, 2000]. The connections and boundaries between these
various elements have never been clear. Considering the continuous nature of this boundary
current and its water properties, and its important role in the general circulation of the South
Pacific, it would make sense to describe it with a single name. Consistent with modern oceanographic practice of geographically-descriptive nomenclature, a recently proposed name is the "Gulf of Papua Current" [SPICE community, CLIVAR Exchanges, No. 56, Vol. 16, No.2, May 2011]. At the point where the boundary current turns sharply into the Solomon Sea and joins with substantial inflow from the interior ocean (section 4), it would then make sense for the name to change, and New Guinea Coastal Undercurrent (NGCU) is well-established for this part of the flow [Tsuchiya et al., 1989]. In the following, we will use "Gulf of Papua Current" (GPC) to refer to this boundary current within the Coral Sea, noting the earlier names as appropriate.

2. Data and methods

The foundation of this paper is estimating the absolute velocity at 1000m from drift trajectories of Argo floats, then using that as a level of known motion to reference geostrophic shear from the CARS (CSIRO Atlas of Regional Seas) Atlas [Dunn and Ridgway, 2002; Ridgway et al., 2002; http://www.cmar.csiro.au/cars], giving a field of three-dimensional mean absolute geostrophic velocity in the Coral Sea. Float motion was extracted from the subsurface drift of 208 quality-controlled floats in the region 140°E-180°, 30°S-7°S through May 2011, and objectively mapped. The quality control, treatment of floats with varied parking depths, estimation of surface and ascent/descent times, and objective mapping procedures are discussed
in the Appendix. The resulting mapped 1000m currents are shown in Fig. 2a. In a few places
where the bottom is shallower than 1000m, and thus not sampled by Argo float drift, we chose to
reference geostrophic velocities at the bottom; consequences of this choice for specific regions
are mentioned in the text.

Consistency might suggest using dynamic height and geostrophic shear from the same Argo
floats that provide the trajectory information, which would have the advantage that the times of
the profiles would be the same as those of the velocities. This could be done either profile by
profile or more generally by using a climatology such as the Argo Atlas [Roemmich and Gilson,
2009]. However, Argo shear provides timely gradients only in the alongtrack direction (namely it
only gives the crosstrack geostrophic velocity component), which is less useful for the present
purpose since by definition the velocity direction sampled by the float trajectories is alongtrack.

A further problem occurs in the western boundary currents where the cross-shore density field
resolution provided by existing Argo sampling is inadequate to define the sharp gradients. We
also found that the Argo Atlas, which is intended for studies of large-scale phenomena, had weak
isopycnal slopes along the western boundary where they are known to be steep, with consequent
weak geostrophic shear. The CARS climatology, on the other hand, incorporates many
Australian near-shore profiles (and the 2009 version we used includes Argo profile data through
about 2008) and gives apparently more realistic geostrophic cross-shore shears. As a result of this testing, we concluded that the CARS fields gave a better depiction of the cross-shore property gradients and alongshore geostrophic shear than the Argo Atlas, especially for the western boundary currents. However, the disadvantage of using CARS is that the Argo parking-depth velocities cover primarily the period 2005-2010, while the CARS geostrophic shear spans a longer period, thus mixing temporal and spatial signatures. It is not possible to precisely specify the time distribution of Coral Sea profile data that went into the CARS compilation, but about half of it comes from the Argo floats, and the remainder going back primarily to the 1980s (J.Dunn, personal communication).

On the other hand, the data remains too sparse, and the gridding algorithm used to construct the CARS fields appears too coarse, to adequately describe near-shore shears in a few crucial locations, most notably around the tip of the Louisiade Archipelago at the southeast tip of Papua New Guinea (Fig. 1). Other studies using glider data have shown sharp property gradients and large transports very close (less than 10km) to the reef line as the NGCU turns sharply around the island chain [Gasparin et al. 2012; Davis et al. 2012]. We attempted to embed the Davis et al. glider fields near the Louisiades into the CARS grid, but concluded that this could not be usefully accomplished, primarily because the glider data only extends to 700m. The unfortunate
consequence of this failure is that we do not have confidence in a critical number: the northward transport from the Coral into the Solomon Sea, and this important topic is left for future study.

Although it would have been desirable to extend the study into the Solomon Sea and thus be able to describe the equatorward pathway more fully, we found that Argo sampling inside the Solomon Sea was insufficient to describe the velocity structure at 1000m. A total of 12 Argo floats had entered the Solomon Sea through May 2011: four westward through Solomon St, six northward in the NGCU close to the Louisiade Chain, and two northward in the open water east of the Louisiades. Only one of these has left the Solomon Sea (through Vitiaz St), and three appear to have died on reefs. The rest have spent months or years wandering irregularly in the central basin without any consistent pattern of motion. Although formally the objective mapping produced a solution, it shows abrupt changes of direction with strong velocity convergences at 1000m. Apparently this represents time variability and in our judgment is not realistic or useful; we therefore forwent a description of the velocity north of 9°S in the Solomon Sea.

The resulting three-dimensional flow field of referenced geostrophic currents shows that there is no appropriate “level of no motion” in the Coral Sea. Current magnitudes resulting from this method were at least a few cm s$^{-1}$ down to 2000m in the zonal jets and western boundary currents.
3. Vertically-integrated transports and the Island Rule

3.1 Introduction: Island Rule

The simplest dynamical picture of the flow through the Coral Sea is the linear, wind-driven Island Rule streamfunction [Godfrey, 1989]. The Island Rule is a generalized Sverdrup circulation taking into account the blocking effects of the large islands that are a first-order feature of the southwest Pacific; it also predicts the western boundary currents implied by the interior transports. While Sverdrup dynamics are a simplification, in particular by representing only a vertical integral, assuming that a few years of sampling adequately represents the mean, and that the bottom is flat (which is far from the case in this region [Couvelard et al. 2008]), they have proven remarkably robust in predicting the volume transport and flow structures in many parts of the world ocean. The Island Rule puts the regional circulation into basin and global context, shows how Coral Sea circulation is shaped by basin-scale winds, and also shows how local wind features – modified by the high islands – produce specific regional currents. It provides a linear first-guess hypothesis whose inconsistencies with observations point to where nonlinear or topographic effects might be important, and we bring those out in the sections below. We will therefore describe the predicted Island Rule circulation, using this to introduce the wind forcing that drives the Coral Sea currents, and also as a framework for discussing the currents and transports themselves. In later sections we discuss elements of the observed
circulation that diverge from this simple picture, as well as the baroclinic structure that is not
specified by the vertically-integrated Island Rule.

The Island Rule has been thoroughly discussed in several publications [Godfrey, 1989;
Wajsowicz, 1993; Pedlosky et al., 1997; Firing et al., 1999], and the particular solution shown
here is detailed in the Appendix to Qiu et al. [2009], so here we only mention the specific aspects
necessary for the present story.

Briefly, east of island barriers, the volume transport streamfunction is found from the Sverdrup
relation (i.e., proportional to the westward integral of the wind stress curl from a constant eastern
boundary condition). At an island barrier, where the Sverdrup integration cannot proceed, the
Island Rule determines the value of the streamfunction around its coast; in practice this value is
dominated by the meridional average of the Sverdrup transport east of the island. Thus, while the
initial Sverdrup integration is conducted at each latitude independently, the island value links the
Sverdrup values over the latitude range of the island. The circum-island wind also contributes to
the island streamfunction, though this is usually a small term, even for the large islands of
Australia and New Zealand [Qiu et al., 2009]. Note that it is important to include New Zealand
in the calculation of the Island Rule streamfunction for Australia, since the islands overlap
[Godfrey, 1989]; in the present calculation, New Zealand increases the value of the
streamfunction at Australia by about 8%. Transport between any two islands is given by the
streamfunction difference between them (including South America, here taken as zero).

The streamfunction is constant around the island's coast, so any meridional gradients that might
occur to its east are suppressed to its west, where the Sverdrup westward integration proceeds to
the next island. Thus all the Sverdrup zonal transport east of an island is concentrated into jets
emanating from the island's tips, and we expect relatively quiet regions extending west from the
island.

The inferred western boundary current along an island's coast is the difference between the
Island Rule streamfunction value at the coast and the Sverdrup streamfunction immediately to its
east. In most cases, the meridional averaging that produces the Island Rule value implies that at
some latitude on the island's east coast the coastal and offshore streamfunctions will be equal,
and thus the boundary current there will be zero; i.e., it bifurcates. Thus, in general the net
meridional transport of the western boundary transport at an island is small; typically we find a
simple split of the incoming transport, in some cases partly weighted to the north or south of the
island.
3.2 Uncertainties of the Island Rule calculation

It is worth noting the problems and uncertainties in a practical implementation of the Island Rule for this region. In particular, the sometimes-ambiguous choice of meridional span of an island produces a significant uncertainty in the calculation. South Pacific islands commonly have offshore island chains separated by channels (Vanuatu) or reef systems that greatly extend their effective length (New Caledonia). The resulting — debatable — choices of island length, which may be constrained by the resolution of the gridded wind product used, have a significant effect on the solution [e.g., Couvelard et al., 2008]. The choice principally influences the solution through the meridional averaging mentioned above, where even a 50km span difference can change the estimated streamfunction at the island by up to a few Sv. (For example, note the tightly-packed streamfunction contours east of New Caledonia in Fig. 3c). The choice of island dimensions also affects the estimate of circum-island wind, but this is usually minor. These uncertainties mean, for example, that the bifurcation streamline of the South Equatorial Current at the coast of Australia can appear either north or south of New Caledonia for seemingly slight changes in the assumptions. Thus, we hesitate to ascribe precision closer than a few Sv to Island Rule-estimated values of transport between islands.
The Island Rule provides no solution inside the Gulf of Papua, which is shadowed from westward-propagating long Rossby waves, and where linear Sverdrup dynamics cannot represent the expected western boundary overshoot [Bryan, 1963; Veronis, 1966; Pedlosky, 1987, see his sections 5.6 and 5.7]. For the solution shown in Fig. 3c, we have simply assumed that the land mass of New Guinea fills the Gulf to 10°S, and thereby forced the implied boundary current to flow directly from the coast of Australia along the southern edge of the Louisiade Archipelago.

Similarly, the Island Rule does not encompass the dynamics that cause the EAC to separate from the coast, break into eddies, and recirculate [Godfrey and Golding, 1981]. A second difficulty is that the Island Rule assumes that an "island" is far enough east of the western boundary layer of the next coast to its west that it can be considered independent. That assumption may well not be accurate for 160km-wide Solomon Strait, so the estimate of Solomon Island streamfunction shown in Fig. 3c is open to question.

Nevertheless, given the robustness of the Sverdrup circulation in much of the world ocean, the importance of the islands in determining the South Pacific circulation, and the clarity of the Island Rule's depiction both of the blocking effects of the islands and the specific role of the western wind stress curl, it is well worthwhile to examine this picture of the vertically-averaged wind-driven currents.
3.3 Island Rule results

Mean wind stress over the Coral Sea (Fig. 3a) is dominated by southeasterly tradewinds whose maximum is roughly along 20°S: from south of Fiji, just north of New Caledonia, then trending slightly north to Australia (the zero wind stress curl line in Fig. 3b). Mean zonal wind stress is much weaker in the South Pacific Convergence Zone (SPCZ) along 6°-10°S. The resulting negative (upwelling) wind stress curl that overlies the northern Coral Sea lifts the thermocline towards the north (Fig. 4a) to a peak near 10°S (not shown); the resulting slopes away from this peak intensify westward gyre flow from about 10°-14°S and also produce the eastward South Equatorial Countercurrent north of the peak along 7°-9°S (Fig. 3c, and Chen and Qiu [2004]).

Broadly speaking, this entire pattern moves with the sun: displaced about 5° northward in southern winter and southward in southern summer. The high mountains of Fiji and Vanuatu, and to a lesser extent New Caledonia, create wind speed minima in the lee to the northwest of these islands (Fig. 3a), resulting in wake dipoles of positive and negative curl (Fig. 3b), similar to features that have been noted behind other mid-ocean islands [Chelton et al., 2004]. These dipoles have an enhanced effect in the South Pacific because they occur west of the islands, where the Island Rule resets the Sverdrup integration to a constant value; thus the spatial variation of the streamfunction in the west is determined largely by the local dipoles rather than by the large-scale winds.
Most importantly, when the Sverdrup integration is carried westward from Vanuatu (where the
Island Rule streamfunction value is about zero), the southern negative lobe of the dipole quickly
increases the value of the streamfunction while the northern positive lobe retards it. The result
(Fig. 3c) is a zig-zag of streamfunction contours, resulting in the eastward Coral Sea
Countercurrent (CSCC) flowing towards Vanuatu along 16°S [Qiu et al., 2009]. The same
zigzag phenomenon is seen west of Fiji to a lesser degree.

The Island Rule solution predicts that the westward inflow to the Coral Sea is split twice: first by
Fiji into branches north and south of about 17°S, then the southern branch is split again by New
Caledonia [Webb, 2000]. These jets are known as, first, the North Vanuatu Jet (NVJ), which
flows through the relatively broad opening between Vanuatu and the Solomon Islands; second,
the North Caledonian Jet (NCJ), concentrated at 18°S just at the tip of the New Caledonian reef;
third, the South Caledonian Jet (SCJ) at 24°S. In the Island Rule representation the NVJ carries
about 10 Sv, the NCJ about 12 Sv, and the SCJ about 8 Sv.

With Sverdrup transport approaching Fiji concentrated to the south (Fig. 3c), the meridional
averaging that determines the Island Rule streamfunction value there is smaller (less negative)
than if the transport were uniform, thus more of the incoming transport passes south of Fiji
[Stanton et al., 2001]. This means that transport arriving at Vanuatu is even more strongly weighted to its south (the negative wind stress curl between Fiji and Vanuatu also contributes to this southward weighting; Fig. 3b), and almost all passes south of Vanuatu. As a result, a strong flow approaches New Caledonia near its northern end; the bifurcation along New Caledonia's east coast is shifted to a more negative value and the NCJ/SCJ split is weighted towards the northern jet.

The bifurcation of the western boundary current along the coast of Australia has been the subject of much discussion [Church and Boland, 1983; Church, 1987; Tsuchiya et al., 1989; QL02]. In the present Island Rule solution, the bifurcation streamline is the $-11.3$ Sv contour, which arrives at the coast near $19^\circ$S (Fig. 3c). (Note that this $11.3$ Sv equals the transport of the Indonesian Throughflow in the Island Rule solution). That contour passes just north of New Caledonia where the coastal streamfunction value is $-12.0$ Sv. Thus, a net change only slightly larger than $0.7$ Sv in the estimated Australia-New Caledonia streamfunction difference would switch the bifurcation contour south of the island; since this is well within the uncertainty of the calculation the distinction is meaningless and we simply say that the bifurcation is approximately at New Caledonia. (Another recent Island Rule calculation, using QuikSCAT winds, found a similar flow pattern, with streamfunction value of $-13$ Sv at Australia and $-15$ Sv at New Caledonia;
According to this vertically-averaged representation, westward flow north of New Caledonia largely turns north at the coast of Australia and flows as a western boundary current into the Solomon Sea, while flow south of New Caledonia turns south in the East Australian Current.

As the wind stress curl west of New Caledonia is small (Fig. 3b), only weak transport is predicted in the lee of the island. Interior transport is about 1 Sv northward, and the vertically-averaged western boundary current along the Australian coast between 23°S and 19°S is 2-3 Sv southward (Fig. 3c). The Island Rule-inferred East Australian Current begins at 23°S with the 8 Sv of the SCJ and grows to about 18 Sv by 30°S, while the Gulf of Papua Current begins at 19°S with the 12 Sv of the NCJ and grows to 25 Sv in the Gulf of Papua (Fig. 3c). Little net transport flows from the Coral to the Tasman Sea, according to the Island Rule.

Thus from this linear, vertically-integrated point of view there is a clear dynamical southern boundary of the Coral Sea in the latitude band west of New Caledonia, where transport arriving from the east in the South Equatorial Current has split irrevocably into poleward and equatorward branches. In the sections below, we will show that the baroclinic structure of these flows in fact produces a current system more complex than this, and in particular that shallow southward and deeper northward transport along the western boundary does occur west of New
Caledonia. Nevertheless, the basic picture remains realistic and we will henceforth consider the southern boundary of the Coral Sea to be about 21°S.

### 3.4 Observed interior transports compared to the Island Rule prediction

Observed vertically-integrated currents broadly show the same features as the Island Rule, with some clear modifications by topography (Fig. 5a,b). The same three westward jets are seen splitting around the islands, as are the weak currents west of New Caledonia and the western boundary current bifurcation near 19°S. West of Vanuatu, the eastward Coral Sea Countercurrent is observed along 16°S to 155°E, similar to the Island Rule solution (most clearly seen in the streamfunction of Fig. 3c). The Island Rule jets are constrained to be one gridpoint wide and flow due west from the tip of their generating island, while the observed currents may be narrow at their origin but broaden with distance; some of this is probably due to nonlinear, mixing and instability processes not encompassed in the Island Rule [Qiu et al., 2009].

Some of the differences between the Island Rule and observed currents can be ascribed to ocean bottom topography, which is absent from Island Rule dynamics; Couvelard et al. [2008] discuss and explicitly model the dynamical effects of Coral Sea topography. Most prominently, the roughly 200 by 400km Queensland Plateau centered at 17°S,150°E (Fig. 1) has extensive shallows and reefs that effectively block large-scale flow, and is separated from the Great Barrier
Reef by a 1000m deep, roughly 100km wide channel. The observed NCJ, broader than the single-gridpoint Island Rule jet, encounters the Queensland Plateau and partly splits around it (Fig. 5a; and see [Choukroun et al., 2010]).

Observed and Island Rule westward transports entering the Coral Sea can be compared on sections northward and southward from New Caledonia (Fig. 6a,b). In both cases, the Island Rule jets (North and South Caledonian Jets) have realistic magnitudes (however note that the jet latitudes in the Island Rule solution depend on the ambiguous choice of island endpoints mentioned in section 3.1.2). Near New Caledonia, the observed jet cores have approximately the same 1° latitude width as their Island Rule counterparts. In the north (Fig. 6a), the Island Rule correctly places the eastward CSCC at 15°-16°S, but its magnitude is smaller than observed. The most important transport difference is the too-weak Island Rule representation of the NVJ at 11°-15°S (Fig. 6a). While observed NVJ transport is about 13 Sv, the Island Rule gives only 9 Sv; it is not known why this discrepancy occurs. However, both our gridded observed currents and the Island Rule prediction indicate that the NVJ is a broader current than the narrow island jets, consistent with it being an ordinary Sverdrup current largely unaffected by the islands. (We show in section 4.2 that the NVJ is also confined much shallower than the deeply-extending NCJ and CSJ). Most of the NVJ does not emanate from the tip of an island, and thus it should not be
considered an "Island Rule jet". It is, however, worth noting that some high-resolution OGCMs simulate narrow jets at the northern tip of the Vanuatu Archipelago and through some of its channels (e.g., the 1/12° OCCAM model shown in Fig. 6 of Ganachaud et al. [2008]). In our gridded product, there is a poorly-resolved suggestion that a small part of the NVJ does originate at the north end of Vanuatu at 14°S (Fig. 2a).

An apparently small difference will turn out to be significant: the turning of much of the NVJ northward into the Solomon Sea near 155°E (Fig. 5a). This feature is absent in the Island Rule solution (most clearly seen in Fig. 3c), as Sverdrup currents must have a southward component in this region where the wind stress curl is negative (Fig. 3b), and there is no other linear mechanism to produce meridional flow. The northward turn of much of the NVJ directly into the Solomon Sea before encountering the western boundary has been noted previously in observations [Andrews and Clegg, 1989; Gasparin et al., 2012] and models [Melet et al., 2010], and we will show below that this is a significant feature of the Coral-Solomon Sea circulation and its property transports.

The question of flow in the latitude band 24°S-19°S west of New Caledonia remains vexing. The Island Rule predicts essentially zero flow in this band, either in the interior or along the western boundary (Fig. 3c). Unfortunately, Argo sampling of this region is the weakest in the southwest
Pacific (Fig. 7). There have been only 6 float deployments west of New Caledonia (19°S-24°S); one of these is very recent and two died on the Chesterfield Reefs along 158°E. Few floats drift into the region: not a single float has passed completely through it from south to north, and only one from north to south (float 5901627, which bounced along the bottom against the Great Barrier Reef in depths of 200-600m, so its trajectory information is not useful). Several passed partway along the western boundary (in both directions) but died on the reef. Float paths usually bifurcate around New Caledonia either north or south, then head due west to Australia and continue northward or southward along the boundary; that pattern of float drift is consistent with the Island Rule picture of westward jets at each end of the island, with New Caledonia marking a final split of the SEC. However, the sparse sampling means our confidence in the observed circulation pattern west of New Caledonia is low.

From Argo trajectory data, then, one would conclude that flow west of New Caledonia is weak and disorganized (Fig. 2a). Geostrophic relative currents from CARS indicate the same thing, with two important (and dynamically linked) exceptions: First, shallow eastward flow is found everywhere in the western South Pacific south of about 15°S (Fig. 2b; also see Reid [1961] and Ridgway and Dunn [2003]). Second, strong upward slopes of isopycnals are observed along the coast of Australia south of 15°S, in particular above the roughly 400m-deep Marion Plateau at
19°-23°S (Fig. 1). These imply shallow southward transport along the boundary, and are the source of the southward coastal vectors above topography in Fig. 5a. This will be further discussed below in the following sections on the baroclinic circulation.

4. Observed baroclinic circulation - in the interior

4.1 Water masses entering the Coral Sea

Water properties are used in this work to trace pathways of flow that has entered the Coral Sea; here we briefly introduce the characteristics of water arriving in the SEC between New Caledonia and the Solomons; for more complete discussions, see Donguy and Henin [1977], Tomczak and Hao [1989], Sokolov and Rintoul [2000], and Maes et al. [2007].

Salinity and oxygen entering the Coral Sea from the east have very different structures, but both exhibit a front near 16°S at all depths above about sigma 27 (Fig. 8). At thermocline levels, the front is found in the eastward CSCC between the NVJ and NCJ. On the southern side of the front, higher-oxygen water arriving in the NCJ suggests more recent ventilation than water arriving north of the front in the NVJ. Two distinct high-salinity tongues enter the Coral Sea at different densities. At sigma 24.5 high salinity is on the northern side of the front in the core of the NVJ, while at sigma 26.5 high salinity is on the southern side in the NCJ (Fig. 8). At sigma 27.2, where the NCJ is the only significant flow, the sign of the gradient switches again. The
front and reversing gradient are most clearly seen in salinity sampled by a glider section from the Solomons to New Caledonia, described by Gourdeau et al. [2008] (Fig. 9, right).

These high-salinity tongues have distinct origins. The shallow NVJ carries water that was subducted in the eastern subtropics near 20°S, 120°W, at the highest surface salinity in the Pacific [Tomczak and Hao, 1989; Delcroix and Henin, 1991; Kessler, 1999; QL02; Grenier et al., 2013]. The deeper high-salinity tongue, in the NCJ south of the front, originated north of New Zealand near 30°S, 180° [Donguy and Henin, 1977; Tsuchiya, 1981; Roemmich and Cornuelle, 1992; Donguy, 1994; Qu et al., 2008].

4.2 Velocities

The observed circulation reflects the tilted subtropical gyre [Roemmich and Cornuelle, 1990]. The bowl of the gyre is centered near 15°S at 20°C, but is near 20°S at 10°C, and is south of 30°S at 5°C (Fig. 4a); thus the westward gyre flow deepens to the south. Above about 500m (~sigma 26.5), isotherms successively break off towards the surface from the general downward slope of the thermocline; south of Vanuatu (15°S) isotherms above 12°C increasingly slope up (Fig. 4a). That upward slope indicates eastward geostrophic shear (Fig. 2b), implying that the overall westward flow of the gyre is weakened or reversed toward the surface, so westward currents south of about 15°S have subsurface maxima [Reid, 1961].
The North Vanuatu Jet, north of 15°S, is surface-intensified and limited to the upper ocean; while it is the strongest of the three jets at sigma 24.5 (Fig. 10a), it is much weaker at sigma 26.5, and has disappeared in the directly-mapped Argo drift velocities at 1000m (Fig. 2a); note that there is almost no slope of sigma 27.4 below the NVJ near 1000m (Fig. 4a). In other words, the NVJ appears simply as the westward limb of the shallow thermocline gyre, spanning about 400km from 15°S to 11.5°S (Fig. 11g), and is adequately represented by geostrophy relative to 1000m. A slight upward-to-the-south slope of isopycnals above sigma 24.5 at these latitudes, caused as much by salinity as temperature (Fig. 4a,b), gives it a maximum speed of about 16 cm s\(^{-1}\) at 50-75m, fairly broad but centered at 13°S.

By contrast, the North Caledonian Jet (and to a lesser extent the SCJ) is narrower but much thicker vertically (Fig. 11h,i), and both are seen to be prominent features at sigma 27.2 (Fig. 10c), where they are identified by the two tongues of low-salinity Antarctic Intermediate Water (AAIW) they carry past the tips of New Caledonia (Fig. 4b; Fig. 8c). The isopycnal slopes associated with these two currents remain steep at 1000m, well below the subtropical thermocline (Fig. 4). This large depth extent is consistent with direct observation of the NCJ by lowered ADCP profiler, which showed strong westward flow in the NCJ to 1500m [Gourdeau et al., 2008; also see Ganachaud et al., 2008; Gasparin et al., 2011].
The NCJ flows across the Coral Sea to at least 154°E as a distinct jet, much narrower than the NVJ, with a core only about 100km wide, centered at 17.5°S, and carrying about 13 Sv above 1000m (Fig. 6a). This climatological view is confirmed by motion of the individual float tracks (also see Maes et al. [2007]). Twenty-one of the Argo floats studied here passed through the start region of the NCJ at the northern tip of New Caledonia, all with parking depths near 1000m. Of these, eight are either too recent or died quickly. Of the 13 useful floats, four had irregular motions that took them out of the current. The remaining nine floats (70%) drifted from 162°E to 154°E without crossing either 16°S or 20°S; i.e., they can be seen as being carried by the NCJ at 1000m. Their mean zonal drift speed over this distance (counting subsurface distances and times only) was 3.1 cm s\(^{-1}\), or somewhat less than a year to cross the Coral Sea. Clearly the NCJ is a coherent feature of the circulation at 1000m; this is borne out by the mapped Argo trajectories (Fig. 2a). On the other hand, isopycnals in the upper NCJ flatten and reverse slope above about 275m (Fig. 4a,b, and note that the near-surface salinity gradient also contributes to this change). This gives an NCJ maximum speed of about 11 cm s\(^{-1}\) near 275m depth, decreasing to about 7 cm s\(^{-1}\) at the surface, and remaining strong and well-defined with speeds of 5 cm s\(^{-1}\) to 1500m.

About 2-3 Sv of westward flow joining the NCJ above 300m appears to come through the Grand Passage, a gap roughly 35km wide and up to 600m deep through the northern New Caledonia
reef system near 18.75°S. This is seen as the shallow southern lobe of westward flow in Fig. 11h at 19.5°S-18.5°S (comparable synoptic estimates of transport through the Grand Passage are found in Fig. 2 of Gourdeau et al. [2008]; also see Ganachaud et al. [2008] and Gasparin et al. [2011]).

The eastward-flowing Coral Sea Countercurrent, between the NCJ and NVJ, begins near 157°E and increases in magnitude approaching Vanuatu between 15° and 16°S (Fig. 10a), consistent with the Island Rule solution (Fig. 3c). The CSCC is seen only in the upper layer above about 250m, intensified towards the surface where its mean speed is near 14 cm s⁻¹ (note strong eastward near-surface shear along 16°S; Fig. 2b). Being shallow and less than 150km wide, its transport is only about 1.5 Sv (Fig. 6a). Isopycnal slopes associated with the CSCC are upward to the south only above sigma 26; deeper isopycnals at this latitude are flat or slope downward (Fig. 4a). Apparently the upwelling effect of the negative curl west of Vanuatu (Fig. 3b) is felt only in the upper thermocline. A similar band of upper eastward shear is seen west of Fiji (Fig. 2b), apparently due to the similar curl pattern downwind of Fiji (Fig. 3b; and see Qiu et al. [2009]). Although not detected in the present mean fields (nor obvious in the individual Argo trajectories), a band of high mesoscale eddy activity is commonly seen in satellite altimeter data
along the axis of the CSCC at 16°S, and was attributed to barotropic shear between the mean
eastward CSCC and its westward-flowing neighbors by Qiu et al. [2009].

The observed CSCC is a clear boundary between the NVJ and NCJ east of its origin near 157°E
(Fig. 10a). Approaching the coast of Australia, velocities on sigma 24.5 and 26.5 suggest a
partial merging of the two westward currents: the NVJ tends slightly south consistent with the
negative wind stress curl over this region (Fig. 3b,c), while part of the NCJ appears to turn north,
perhaps because of the blocking Queensland Plateau [Choukroun et al., 2010].

Within 50-100km of Makira Island (the southernmost of the Solomons chain), a thin band of
eastward flow is seen north of the NVJ, with about 2 Sv apparently exiting the Solomon Sea
(Fig. 5a). This counterflow is found in the mapped Argo trajectories (Fig. 2a) and also in the
geostrophic shear (note the downward slope of the sigma 27.2 and 27.4 isopycnals at the
northern end of the 162°E section (Fig. 4). Both measured and geostrophic depictions suggest a
subsurface core; a similar current, closely trapped to the coast and centered near 400m, was seen
in the cruise data of Gasparin et al. [2012].

This picture of the currents north of New Caledonia is generally consistent with a detailed
absolute velocity section made by an ocean glider crossing from Guadalacanal to New Caledonia
roughly along 162°E in July-October 2005 [Gourdeau et al., 2008]. In that snapshot, the NVJ
was comparably 300km broad and shallow, with a maximum speed of more than 20 cm s\(^{-1}\) at the surface but virtually absent below about 350m. Its transport was 21 Sv, about 40% larger than the mean from the present Argo/CARS data, but at a time of year when this current is known to be seasonally stronger \cite{Kessler and Gourdeau, 2007}. By contrast, the NCJ depicted by the glider was about 100km wide, centered at 17°S, with a similar 20 cm s\(^{-1}\) maximum speed and was almost unsheared from the surface to 600m; its transport was 12 Sv, the same as found here. The 2005 glider snapshot showed a weak CSCC, mostly indicated by very irregular flow between 14.5° and 16.5°S, consistent with the eddies known to occur there \cite{Qiu et al., 2009}.

The South Caledonian Jet is the third westward element of the South Pacific subtropical gyre, carrying slightly less transport than the NVJ or NCJ. Unlike the NCJ, flow above the SCJ is strongly sheared and reduces almost to zero at the surface (Fig. 11i; also see Fig. 2); this reflects the reversal of isopycnal slopes in and south of the SCJ core (Fig. 4). The mapped SCJ in the present data is less clearly seen than the NCJ. This may be an observational limitation due to the high magnitude of mesoscale eddy variability across this latitude band, as suggested by the sparse in situ observations of the SCJ \cite{Ganachaud et al., 2008}, and borne out by satellite altimetry \cite{Qiu et al., 2008}.
The SCJ passes over two topographic barriers: the Norfolk Ridge extending south from New Caledonia along 167.5°E, with sill depths about 1500m and many shallower obstructions, and then the Lord Howe Rise along 160°-162°E, about 1500m deep but with a shallower seamount chain just to its west (Fig. 1). It funnels through the deepest gap in the Norfolk Ridge at 25°S, but appears to break up at the Lord Howe Rise seamount chain (Fig. 10, and Ridgway and Dunn [2003]). In contrast to the NCJ, Couvelard et al. [2008] noted a much larger change in the SCJ between a flat-bottom model, where it flowed due west like a typical Island Rule jet, to a model with realistic topography where the current path became convoluted and meandering.

Although the linear Island Rule solution predicts the SCJ to be an island jet emanating directly from the southern tip of New Caledonia (Fig. 3c), this character is much less evident in the present observations, where the current appears further offshore than does the NCJ (most clearly seen in the mapped Argo trajectories of Fig. 2a). It is not clear how much of the SCJ is fed from the southward branch of the SEC splitting against the east coast of New Caledonia (namely as an Island Rule jet), versus how much derives from other westward offshore flow funneling through the Norfolk Ridge. A further difficulty in characterizing the SCJ in the west is that it approaches the coast of Australia at a latitude where there is a significant western boundary recirculation.
[Ridgway and Dunn, 2003; Godfrey and Golding, 1981]. Clearly much about the SCJ remains an unsolved puzzle whose fundamental properties and dynamics are poorly understood.

These three currents continue generally westward across the Coral Sea to the coast of Australia. However, the high-salinity tongue of the NVJ branches into the Solomon Sea (Fig. 8a). This is consistent with the velocity on sigma 24.5 (Fig. 10a) which suggests that part of the NVJ turns north, joining the western boundary current flow entering the Solomon Sea. A similar current structure was seen in glider observations across the mouth of the Solomon Sea [Davis et al. 2012]. This glider sampling also showed that the salinity pattern across the mouth of the Solomon Sea is virtually a mirror of that entering the Coral Sea (Fig. 9). We interpret this to indicate substantial transport from the NVJ directly into the central Solomon Sea at isopycnals above sigma 25.5. As noted above (section 3.2.2), there are no apparent linear mechanisms that would lead to such a northward turn in mid-basin. Because of the difficulty in describing the velocity near the tip of the Louisiades from the present data (section 2 above), we are unable to quantify the transport making this northward turn. The picture is further confused because the shallow velocities seem to show a cyclonic eddy in the southeastern Solomon Sea, with southward flow exiting the Sea in the east (Fig. 10a, and Fig. 2b), as was also suggested by the glider analyses of Davis et al. [2012], the SADCP compilation of Cravatte et al. [2011] and the
modeling study of Melet et al. [2010]. Nevertheless, high-salinity water from the NVJ quite clearly enters the Solomon Sea east of the boundary current.

5. Western Boundary Currents

5.1 Vertically-integrated boundary transport

Observed and Island Rule-predicted western boundary transports are compared in Fig. 12a. While the Island Rule boundary transports are deduced by conserving mass in the boundary layer; i.e., from the difference between the inshore (Island Rule) and offshore (Sverdrup) streamfunction values (Fig. 3c), observed boundary currents were calculated from the mapped Argo float drift and climatological geostrophic shear exactly as was done for interior-ocean locations. To estimate "western boundary transport" from the observed currents, we defined the boundary current as the integral of meridional current from the coast to the first zero-crossing that was at least 0.5°, but not more than 3°, from the boundary at each depth and latitude. The results were very similar to a simpler definition of 125km from the boundary (Figs. 11a-f show that 125km is a reasonable choice for boundary current width everywhere except in the complex topography at the Queensland and Marion Plateaus). The channel inshore of the Queensland Plateau is about 1200m deep with a sill just below 1100m; Argo floats drift freely and rapidly through it, with speeds of 5-8 cm s⁻¹. However, below 1200m, the westernmost point is 300km
further east, offshore of the plateau. It does not seem meaningful or useful to join these two
regions, and we have therefore blanked them in Fig. 12b. The interesting question of a double
western boundary layer – inshore and offshore of the Queensland Plateau – is not yet adequately
sampled by the Argo float drift. While there may be a secondary western boundary current east
of the Plateau, this topic must await additional data.

Latitudes west of New Caledonia

As expected from the vectors of Fig. 5b, there is little vertically-integrated boundary transport
across the latitude range of New Caledonia (23°S-19°S) in either observed or Island Rule
representations (Fig. 12a), consistent with the idea that the vertical-average bifurcation latitude
of incoming flow at the coast of Australia cannot be specified more precisely than that. However,
as noted above (section 3.2.2), sampling in this region is considerably sparser than in the Coral
Sea itself; also, much of the (southward) western boundary transport occurs over the shallow
Marion Plateau – unsampled by Argo – so our flow estimates there come entirely from CARS
geostrophy with an assumed zero reference level at the bottom (Fig. 5a).

The EAC south of 23°S

South of the bifurcation, both observed and Island Rule-predicted transports show the East
Australian Current growing to about 18 Sv by 28°S with the arrival of the southward-turning
inflow from the SCJ (Fig. 5a, and Ridgway and Dunn [2003]). Further south, the observed boundary transports (estimated from the coast to the zero-crossing) become much larger (Fig. 12a), but this growth includes only the inshore (southward) half of an increasingly-recirculating EAC with substantial northward flow offshore [Ridgway and Godfrey, 1994], and therefore does not represent the entire western boundary system (Figs. 5a, 11a). If the offshore northward recirculation is included, the total observed transport of the EAC system remains about 18 Sv to 34°S.

Much of the EAC turns offshore at the Tasman Front near 34°S (Fig. 5a, and Ridgway and Dunn [2003], Tilburg et al. [2001]), and observed southward transport drops rapidly (Fig. 12). This occurs at 34.5°S (north tip of New Zealand) in the Island Rule solution but over the broader band 35°-32°S in the observed transports. Considering the large contribution of eddies on EAC transport in this latitude range [Godfrey et al., 1980; Ridgway and Dunn, 2003], we do not expect to estimate a useful mean transport from the presently-available sampling.

**The western boundary current in the Coral Sea**

North of the bifurcation in the Coral Sea, Island Rule-predicted transport grows rapidly to 12 Sv as the NCJ is represented arriving at the boundary in a single gridpoint at 18.5°S (Fig. 5b), then gradually increases to about 26 Sv at 11°S (Fig. 12a). Observed transport grows more slowly, as
the NCJ splits around the Queensland Plateau at 16°S-18°S (Fig. 5a) and its transport arrives over a wider latitude range.

Along the Queensland coast north of 13°S, the observed boundary transport decreases as the current turns offshore and breaks into loops and eddies (Fig. 10). The complex currents in the Gulf of Papua have been previously noted [e.g., Scully-Power, 1973; Melet et al. 2010] but their dynamics not diagnosed, although an inertial overshoot [Bryan, 1963; Veronis, 1966; Pedlosky, 1987, see sections 5.6 and 5.7] seems likely north of 12°S, where the coast of Australia is shadowed by the Louisiade Archipelago. We are unable to determine reliably if the loops and eddies are permanent or have preferred locations, but note that many Argo floats spend months or years revolving around the Gulf of Papua in both cyclonic and anti-cyclonic directions with apparently no consistent pattern. As the climatological geostrophic shear is a rather smooth cyclonic eddy filling the entire Gulf (Fig. 2b), but the mapped Argo trajectories show a double loop with smaller scales (Fig. 2a), we do not have great confidence that the details of the currents in the interior Gulf shown here (Fig. 5a) represent a true mean. Since these phenomena are not encompassed by the steady, linear Island Rule dynamics, the discrepancies between the Island Rule and observed transports (Fig. 12a) north of 12°S are to be expected.
Perhaps more surprisingly, the observed western boundary transport never reaches the magnitude predicted by the Island Rule at any latitude north of the bifurcation (Fig. 12a). This apparently reflects the partial loss of NVJ transport directly into the Solomon Sea near 155°E before encountering the western boundary (section 4.1). Thus some of the westward gyre transport never reaches the western boundary as it does in the Sverdrupian prediction; it is not at all apparent why this should occur. Nevertheless, the property tongues extending into the Solomon Sea (high salinity and low oxygen on sigma 24.5, low salinity and low oxygen on sigma 26.5; Fig. 8) confirm its reality, so the explanation for this important piece of the Coral Sea circulation and property transport remains elusive.

5.2 Baroclinic boundary currents

QL02 (see their Fig. 9) showed that the bifurcation of the South Equatorial Current slopes southward with depth along the Australian coast. Over about 10° latitude, from 25°S to 15°S, upper flow is southward (the start of the East Australian Current), overlying deeper northward flow that becomes the Gulf of Papua Current and eventually the New Guinea Coastal Undercurrent (Fig. 12b).

Cross-shore sections of velocity (Fig. 11) show the evolution of the western boundary flows. The NCJ (Fig. 11h) supports subsurface maxima of the western boundary currents all around the
Coral Sea, with relatively high oxygen (and high salinity on sigma 26.5; Fig. 8b) while the shallower NVJ (Fig. 11g) brings near-surface flow with low oxygen (and also high salinity; see section 4.1). An alongshore coastal section of oxygen concentration shows the sheared advection along the coast of Australia as these inputs arrive: low-oxygen NVJ and upper NCJ water carried south along the coast above sigma 26, while high-oxygen water from the NCJ is carried north along sigma 27 (Fig. 14). Compared to the analogous section at 162°E (Fig. 4c), which is similar to other offshore sections and describes the arriving water properties, the boundary pattern differs by the effect of upper southward and mid-depth northward alongshore advection.

**Southward boundary flow: the EAC**

The East Australian Current begins as a shallow (< 200m, above sigma 25.5) southward flow at 15°S in the Queensland Passage (Fig. 11c; Fig. 12b; Fig 10a; and *Church and Boland* [1983; 1987]). It is traced south along the coast across the shallow Marion Plateau (Fig. 11b) by low-oxygen water from the North Vanuatu Jet (Figs. 13b and a); the oxygen sections confirm that the tongue is closely confined to the western boundary and has no possible offshore source of low oxygen south of the NVJ (Fig. 8).

The nascent EAC is relatively unchanged until about 23°S, where it thickens rapidly (Fig. 12b) and its oxygen concentration increases at all depths (Fig. 14). Both these changes would be
consistent with either of two influences: First, the arrival of the high-oxygen South Caledonian
Jet at the western boundary (Figs. 5a and 4c), and second, the beginning of the EAC offshore
recirculation carrying deeply-mixed (and thus highly-oxygenated) waters and reinjecting them
into the EAC beginning near 23°S (Fig. 5a). Further south, the EAC continues to deepen to at
least 1500m (Fig. 12b; Fig. 11a), which apparently represents the growing influence of the
eddying recirculation [Ridgway and Dunn, 2003].

**Northward boundary flow: the GPC**

Northward flow along the western boundary (the Gulf of Papua Current; GPC) begins as a weak
(∼1.5 cm s⁻¹) current below 800m near 25°S, flowing under the EAC and expanding upward in
the tilted bifurcation region (Fig. 12b). The tongue of high oxygen along sigma 27 (Fig. 14)
marks this northward flow. As the North Caledonian Jet arrives at the western boundary near
19°S, most of its transport below 300m (∼sigma 26) turns north, into the deep Queensland
Passage where GPC strength greatly increases (Fig. 12b; Fig. 10b,c; Fig. 11c); this sheared
structure was called the Great Barrier Reef Undercurrent by Church and Boland [1986].

Further north along the Queensland coast, additional flow from the shallow NVJ joins the
boundary current, which grows towards the surface (Fig. 11d). Around the Gulf of Papua and
south of the Louisiades, the eastward boundary current is still about 100km wide and largely
subsurface (Fig. 11e). Further downstream, as the GPC turns the corner into the Solomon Sea (Fig. 11f), much more shallow NVJ water merges with the inflow (e.g., the high-salinity tongue at sigma 24.5 in Fig. 8a; also note the shallow inflow of low-oxygen water arriving east of the Louisiade Chain near 157°E in Fig. 13f); the equatorward transport grows and thickens, still with a tendency to a subsurface maximum, but extending to the surface. This same character is observed at the northern exits of the Solomon Sea: Vitiaz St. [Lindstrom et al., 1987; Tsuchiya et al., 1989; Lindstrom et al., 1990] and Solomon St. [Butt and Lindstrom, 1994; Cravatte et al., 2011].

The bifurcation

Fig. 12b generally confirms the earlier result of QL02, with the bifurcation contour in nearly the same place. The present boundary current velocities are about 50% larger and they maintain this magnitude much deeper than the QL02 result. This difference is presumably because currents here are referenced to the significant Argo drift velocities at 1000m (Fig. 2a) while the QL02 currents are relative to a zero at 1200m. The present results show additionally that the tilted bifurcation is not a smooth slope but is closely related to the arrival of the three westward currents and can be traced by their transports and water properties on density surfaces. Perhaps the bifurcation would be better described not as a tilt but as something closer to two separate
layers: Above about 250m depth (approximately sigma 26) the bifurcation occurs near 15°S where the NVJ splits, with its southward part carrying a distinct low-oxygen tongue south along the coast, inside the Queensland Plateau, into the shallow EAC as described above (Fig. 14). Below 250m, the bifurcation occurs near 23°S (i.e., south of the NCJ) and the northward subsurface boundary current is dominated by high-oxygen, high-salinity water from the NCJ and SCJ. Additionally, the tilt and sheared boundary current appears to extend deeper and further south than previously realized, with northward flow seen below the EAC, at depths of 1000m or more, to at least 25°S (Fig. 12b). We have less confidence in the shears below 1500m so have cut off Fig. 12 there, but those suggest that the tilt continues south with depth, to about 29°S at 2000m.

6. Discussion

6.1 Overview

We have described two principal pathways through the Coral Sea: a broad, shallow (tropical thermocline) route that splits north and south, and a deeper jet-like route that quickly crosses to the coast of Australia and feeds a northward western boundary current that eventually becomes the New Guinea Coastal Undercurrent.
The deeper route enters the Coral Sea in the narrow North Caledonian Jet at the tip of the New Caledonia reef system at 18°S (Fig. 11h). This Island Rule jet is about 100km wide; its maximum is near 275m and strong westward flow extends below 1500m [Gourdeau et al., 2007; Ganachaud et al., 2011]. At all levels the NCJ carries higher oxygen than the NVJ to its north (Fig. 8). Below about sigma 25.5 (about 250m), it transmits high-salinity water created northeast of New Zealand [Donguy and Henin, 1977]. The NCJ flows west across the Coral Sea as a distinct jet, so that most of the Argo floats in it go directly and rapidly to the coast of Australia. Although it seems to split around the shallow Queensland Plateau, both branches turn north at the coast and rejoin as a western boundary current that circles the Gulf of Papua clockwise. This grand loop around the Coral Sea is the most prominent feature in the mapped Argo trajectories at 1000m (Fig. 2a). This subthermocline boundary flow is traced by high oxygen (AAIW at sigma 27.1, centered at about 650m) and high salinity (in the layer sigma 26-27) tightly trapped to the coast, entering the Solomon Sea to begin the New Guinea Coastal Undercurrent.

The shallow route enters the Coral Sea in the North Vanuatu Jet at 11°S-14°S, surface-intensified and occurring mostly above sigma 26 (300m). It is not an Island Rule jet, but is simply an unobstructed part of the South Equatorial Current, carrying high-salinity, low-oxygen water that was created in the eastern subtropics [Kessler, 1999; Qu et al., 2008]. About half of this inflow
breaks off in the central Coral Sea, turning north into the Solomon Sea, where it is the source of the high-salinity water found on thermocline isopycnals. The other half continues to the coast of Australia where it splits north and south. The northward part joins the subsurface flow from the NCJ, extending the boundary current towards the surface (Fig. 11, compare panels c and d). A smaller part turns south, to begin the East Australian Current. This flow is traced by a tongue of low oxygen in the layer sigma 25-26 along the boundary as far south as 30°S (Fig. 14).

We have taken the viewpoint of a climatological mean circulation, while knowing that the Coral Sea, like much of the rest of the ocean, is vigorously turbulent and variable at all timescales. In particular, we note observations of energetic eddies along the front at 16°S [Qiu et al., 2009], and in the Gulf of Papua [Scully-Power, 1973]. And although we see a mean South Caledonian Jet in the referenced geostrophic currents reported here, synoptic observation of this current (or model snapshots), do not show this current except episodically [Ganachaud et al., 2008]. These regions of turbulent flow can be expected to accomplish significant mixing. Nevertheless, the observations reported here and elsewhere show that such mixing is not overwhelming (e.g., Tsuchiya's [1981] demonstration that water properties formed near New Zealand remain coherent enough to be readily identifiable after passing through the North Caledonian Jet, the Gulf of Papua Current, the New Guinea Coastal Undercurrent, and the near-equatorial "Tsuchiya
jets", to the far eastern Pacific). Clearly the role of eddy mixing in the Coral Sea remains to be fully explicated.

6.2 Mass transport balance entering and exiting the Coral Sea

We have refrained from specifying numerical values for total transport leaving the Coral Sea to the north and south, despite the desirability of doing so. One of the main findings here is that much of the transport occurs in narrow jets close to coastlines and reefs, through constricted channels, or over shallow topography, that are often not well sampled. In our judgment, the coast-to-coast transport values depend too sensitively on these endpoints; despite the ability to reference with the Argo trajectory velocities, experiment showed that apparently small changes in methodology produced quantitatively different estimates of transport.

The inability to specify the mean mass transports exiting the Coral Sea is an unfortunate failure of this work; indeed, one of our principal motivations for using the Argo trajectory velocities as a reference was to do exactly that. Credible estimates of the exiting transport distribution would resolve the essential question of the destination of South Equatorial Current transport as it bifurcates at the coast of Australia, which has occupied much previous work beginning with Wyrtki [1960] and continuing to the present [e.g., Gasparin et al., 2012]; see the Introduction. They would also anchor studies of annual, interannual and lower-frequency variability that are
highly relevant for the climate of the western and equatorial Pacific. However, we do not have high confidence in the numerical transport values of the meridional distribution of exiting flow, and therefore forwent such specification in the description above.

Nevertheless, it is worth showing the transport balance resulting from the velocity field deduced here, to demonstrate its uncertainty and for comparison with future observations that hopefully will produce more confidence. Fig. 15 shows the estimated transport in density layers across the sides of a box enclosing the central Coral Sea. Strong incoming flow is found in all layers, totalling 27.4 Sv above sigma 27.5. Transport exiting north into the Solomon Sea above this isopycnal is about 14.7 Sv, and is dominated by flow between sigma 26-27 (roughly 300-600m depth, consistent with other observations of a subsurface maximum in the New Guinea Coastal Undercurrent; Gasparin et al. [2012]; Davis et al. [2012]). Flow exiting to the south totals 7.3 Sv and is weak in all layers. The net inward transport of 5.3 Sv is the imbalance of this estimate; thus about 20% of the incoming flow cannot be accounted for in the exit sections. We believe that at least most of this unmeasured transport in fact leaves to the north, principally in the New Guinea Coastal Undercurrent. It is also possible that some of the 7.3 Sv of the southward flow implied by this analysis in fact leaves to the north as well.
On the other hand, relatively good sampling (Fig. 7) gives more confidence in our estimate of the mean absolute geostrophic transport entering the Coral Sea between New Caledonia and the Solomon Islands, which we find to be 24 Sv above 1000m (Fig. 6a), 26 Sv above 1500m, and 27 Sv above 2000m. These values include about 2 Sv of shallow westward transport through the Grand Passage (section 4.2), and a slightly-larger eastward transport at the northern end of the section next to the Solomon Island chain (Fig. 6a). Note that these calculations of coast-to-coast transport are similar to, but not the same as, transports through the east side of the box shown in Fig. 15 and cited above.

These estimates were tested by comparing results from integrating along different pathways between New Caledonia and the Solomon Islands, examining both the part due to relative geostrophy from CARS and that from our mapped 1000m velocities based on Argo trajectories. As the NCJ is well-sampled by many Argo trajectories close to the tip of New Caledonia where the 1000m flow is less variable, most of the ambiguity causing differences between pathways occurs at the northern end of the section where fewer Argo floats have passed near the southernmost Solomon Island (Makira). Kessler and Cravatte [2013] compiled all historical estimates of transport between New Caledonia and the Solomon Islands and found values varying between 25-30 Sv, relative to 2000m (their Table 1), similar to those found here. They
also found ENSO variability of up to ±10 Sv, which could affect the estimate of mean transport reported here. Overall, the uncertainty of estimated mean transport entering the Coral Sea is at least ±3 Sv, or about 10% of the total.

The situation is not as good for the flows exiting the Coral Sea north and south. A particular problem in estimating the southward transport is that much of the boundary current appears to occur over the roughly 400m-deep Marion Plateau at 19°-22°S, where water properties and isopycnal slopes indicate southward flow but there is no Argo sampling, and we chose to reference geostrophy at the bottom (Fig. 11b). The resulting estimated boundary transport values have alongshore fluctuations of 3-5 Sv in this latitude range (Fig. 12a) that could as easily be due to the choice of bottom referencing as to real flow variations. These wiggles imply an effective minimum uncertainty on how well we can estimate transport between New Caledonia and Australia (namely the flow from the Coral to the Tasman Sea). Meridional flow outside the coastal boundary layer in this latitude band appears weak, from both Argo trajectories and from CARS geostrophy (Fig. 2 and see section 3.2.2 above). However, even if the uncertainty is larger than this, there is little evidence of significant transport that has entered the Coral Sea between New Caledonia and the Solomons turning south into the Tasman Sea, as some have argued [Wyrtki, 1962; Andrews and Clegg, 1989; Sokolov and Rintoul, 2000].
Salinity and oxygen on isopycnals show a strong contrast between Coral Sea water and that south of 18°S, at all levels (Fig. 8). Except for the finger of low-oxygen water from the NVJ that penetrates south along the western boundary near sigma 25 (about 250m depth; Fig. 14; Fig. 13a,b), water properties indicate only weak southward boundary transport exiting the Coral Sea.

We estimate that net transport between New Caledonia and Australia is probably southward but is unlikely to be larger than about 6-8 Sv. This is still substantially more southward transport than the Island Rule estimate of 0.7 Sv. But in any case, the main growth of EAC transport (and its increasing thickness) occur with the arrival of the South Caledonian Jet near 23°S, not from the Coral Sea (Fig. 12b), and most of the NCJ appears to turn north at the coast of Australia.

This inference is consistent with other recent estimates from climatological data and repeat sampling. Ridgway and Dunn [2003] used an earlier version of the CARS climatology and showed a similar intensification of the EAC with the arrival of SCJ inflow (their Fig. 10 and section 6.2). Roemmich et al. [2005] studied high-resolution repeat XBT sections made during 1991-2002 on the PX30 line. This line sampled from just south of New Caledonia to Brisbane, Australia (27°S); that is, diagonally across the SCJ and then across the EAC south of the SCJ’s arrival at the coast of Australia. Roemmich et al. found about 7 Sv total southward geostrophic
transport across this line (relative to 2000m depth) above the 12°C isotherm (about 450m). The PX30 data showed northwestward crosstrack flow near New Caledonia (apparently the SCJ), then a northward recirculation just east of a strong (14 Sv), narrow EAC (their Fig. 1). An analogous section from the present data is similar, though our EAC is weaker: about 10 Sv crosstrack above 12°C. Overall, then, modern data compilations do not suggest a large southward transport from the Coral to the Tasman Sea as had been proposed by some early studies or analyses of individual cruises.

A different problem occurs at the northern boundary of the Coral Sea, where one would like to know the transport entering the Solomon Sea that subsequently feeds the equatorial circulation [Fine et al., 1994; Tsuchiya et al., 1989; Melet et al., 2010; Grenier et al., 2011]. In this case the difficulty is that the Louisiade Archipelago that forms the western boundary is narrow (50-60km across), and the CARS climatology appears to interpolate properties across it, especially at the eastern tip (in fact CARS includes "ocean" values across much of the eastern Louisiades). As mentioned in section 2, recent glider and shipboard observations make clear that the northward boundary current is very strong close to the tip of the Louisiades [Davis et al., 2012; Gasparin et al., 2012], and we consider the result from the present data (about 15 Sv exiting the Coral Sea to the north) an underestimate.
We note that transports estimated from individual ship or glider sections can be badly aliased by eddies and internal wave noise, and that ENSO variability amplitude is at least 25% of the mean transport [Davis et al., 2012; Kessler and Cravatte, 2013]. We conclude that neither individual sections nor presently-available climatologies produce accurate representations of the mean transport exiting the Coral Sea either north or south. Our focus here has therefore been on the characteristics and mechanisms that produce and control the Coral Sea circulation.

6.3 Dynamics of the tilted western boundary current bifurcation

The meridional structure of the Coral Sea western boundary current can be understood as governed by two basinscale influences: the zonal currents bringing and removing mass from the western boundary layer, and the net meridional transport through the South Pacific required by the Indonesian Throughflow (ITF). These two factors determine which way a particular element of zonal inflow or outflow to the boundary region will turn at the coast, and thus determine the shape of the bifurcation. Neither of these factors can be understood considering the Coral Sea region in isolation.

The shallow southward western boundary flow that begins the East Australian Current in the tilted bifurcation region (Fig. 12b) demonstrates the significance of non-local forcing. This current is geostrophically associated with isopycnals sloping up to the coast; for example, the
sigma 24.5 isopycnal is at least 20-30m shallower at the coast than 100km offshore, everywhere from 30°S to 15°S. This upwards-to-the-coast slope cannot be the result of the local winds, which are strongly alongshore to the northwest – and thus Ekman downwelling – all year along this coast north of 28°S (Fig. 3a). The shallow southward boundary current and resulting southward turn of the NVJ through the Queensland Passage therefore must be remotely forced.

Since the deeper northward boundary current also cannot be locally forced, this sheared system, uniquely among the subtropical gyres of the world ocean spanning at least 8° latitude, represents a distinct characteristic of the South Pacific as a whole.

From a steady, linear (Island Rule) point of view, transport between Australia and South America must be the same at all latitudes from Tasmania to the equator, and that is also the transport of the Indonesian Throughflow (ignoring flow through Bering Strait). Vertically-integrated western boundary transport is thus the difference between the interior Sverdrup transport at each latitude and the Island Rule-estimated ITF (11.3 Sv for the winds studied here; see section 3). With the zonally-integrated zero wind stress curl line near 18°S (Fig. 3b shows the western part), one can infer that the boundary current will transport about 11.3 Sv northward at that latitude (Fig. 12a). Since the Sverdrup transport is southward north of that line, and
northward to its south (Fig. 3), the integrated western boundary currents will tend in the opposite sense.

The shear and resulting tilt of the western boundary current is due to the tilted South Pacific subtropical gyre and the resulting pattern of zonal inflows and outflows to the boundary layer. The tilted gyre produces upper eastward shear overlying deeper westward currents; while we have seen that flow at thermocline level and below is almost entirely westward (Fig. 10), surface flow is largely eastward south of 15°S (Fig. 2b shows the upper shear; and see Donguy and Henin [1977]; Ridgway and Dunn [2000]). This eastward interior transport — removing mass from the western boundary layer — must be balanced by shallow alongshore convergence, while the deeper westward inflow is balanced by divergence in the boundary layer. These produce the alongshore and vertical shear in the boundary layer, which can be referenced by the known total boundary transport.

An appropriate place to start the integration is the Tasman Front at 34°S, where the boundary current is southward (because the wind stress curl is most negative there) and deep mixing produced by the eddying EAC recirculation (e.g., Cronin [1996]) means that the current is southward at all depths (Fig. 12b). At successive latitudes to the north, the shallow convergence forced by the interior flows means that the upper boundary current becomes more southward
\( \frac{dV}{dy} < 0 \), while the deep divergence means that the lower boundary current becomes more northward \( \frac{dV}{dy} > 0 \). These tendencies must stay in balance with the weakening of the total boundary transport as the interior wind stress curl decreases to the north (Fig. 3b). At some latitude south of the zero wind stress curl line, the interior transport will equal 11.3 Sv (northward), and the vertically-integrated boundary transport will therefore be zero. This implies that the boundary current at that latitude will have upper southward and deeper northward flow, as seen in the observed boundary current profile (Fig. 12b). Further north, in the Coral Sea itself, the interior currents are strongly westward at all depths (Fig. 10) and thus convergent in the boundary layer. The resulting boundary current shear is thus northward, strengthening the deep northward alongshore current, and quickly weakening and reversing the upper southward current to produce the thick GPC that feeds the NGCU.

In this light, the southward near-surface EAC north of about 28\(^\circ\)S is seen as a response to shallow eastward (offshore) flow and consequent alongshore convergence in the south.

Similarly, the deep origin of the GPC near 25\(^\circ\)S is the result of subthermocline westward transport in the central subtropical gyre. The tilted bifurcation is the result of this interplay between the total boundary transport (determined by the ITF) and the pattern of interior zonal currents adding and removing mass from the boundary layer. The presence of New Caledonia
near the zero boundary current region complicates the picture locally, but does not change the fundamentals.

We can now ask what would happen to the tilt of the bifurcation if the ITF were closed. If winds over the Pacific remained the same, the interior wind-driven (Sverdrup) flow would be unaltered; the only change could be in the western boundary current. As in the North Pacific, the time-mean boundary current would exactly balance the interior meridional Sverdrup transport at each latitude (again ignoring Bering Strait). With the ITF closed, the Island Rule western boundary transport would be uniformly more negative along the entire coast of Australia: the red curve in Fig. 12a would be shifted downward 11.3 Sv. The bifurcation streamline in the Coral Sea would be the zero Sverdrup transport contour (thus definitely north of New Caledonia; Fig. 3c), and the NCJ would turn south at the coast of Australia, increasing the transport of the EAC by 11.3 Sv and decreasing the equatorward western boundary transport towards the Solomon Sea by the same amount. If this extra 11.3 Sv were uniformly distributed over the upper 1500m of the boundary current (equivalent to subtracting 5.4 cm s$^{-1}$ from the current shown in Fig. 12b), the resulting western boundary flow would be southward at all depths from about 18°S, while shifting the surface bifurcation less than 1° north. The tilt of the bifurcation would then span only about 3° latitude and be comparable to the small tilt in the North Pacific [Qu and Lukas 2003].
Even if the distribution of this 11.3 Sv was weighted more heavily towards the surface, the resulting bifurcation latitude range would be greatly lessened. *McCreary et al.* [2007] did the experiment of closing the ITF in a hierarchy of models and found that the bifurcation tilt indeed nearly disappeared.

In this work we have thought of the Coral Sea in a Sverdrupian context with its Island Rule generalization. Sverdrup dynamics utilize the great simplification of top-to-bottom vertical integration, but these ideas can be extended to an isopycnal layer framework that provides additional insight into the vertical structure of a subtropical gyre. *Luyten et al.*, [1983] models distribute the wind-driven Sverdrup transport over layers constrained to conserve potential vorticity (PV). They postulate that wind stress curl establishes the PV of upper layer water columns while they are exposed to the atmosphere, with those values conserved after the column is subducted into the gyre circulation.

While the simplest interpretation of a Sverdrupian subtropical gyre driven by downwelling wind stress curl is a bowl of isopycnals defining a single active layer, the LPS view extends this to a second, deeper, layer. Subduction occurs where each isopycnal outcrops at the poleward edges of the gyre, forming a ventilated subsurface layer under part of the upper layer gyre. The subducted water columns carry the surface-forced downwelling signal into the second layer, where they...
spread westward and equatorward along PV-conserving Rossby wave ray paths [LPS; Luyten and Stommel, 1986; McCreary and Lu, 1994]. The result is a deepening of the lower interface, forming a secondary bowl of isopycnals below the initial single active layer bowl. Because the subduction signal originates at the poleward side of the gyre and spreads equatorward only slowly, this second bowl is shifted poleward from that in the upper layer, and the combination makes a tilted gyre, as suggested by the slopes in Fig. 4 where the deepest point of each isotherm or isopycnal is further south than that of the one above.

The geostrophic flow in each bowl-shaped layer of such a gyre is anti-cyclonic, but the poleward shift of the lower layer bowl gives a band across the center of the gyre where the lower flow is westward but the upper is eastward; note that the upper-layer eastward flow in this band is enhanced so the vertically-integrated transport remains Sverdrupian. That is the situation in the western South Pacific from about 30°S to 15°S, where the upper eastward flow is often called the Subtropical Countercurrent [Qiu, 1999; Qiu et al., 2008]. Thus the situation at the outcrops where subduction occurs – especially in spring when the deep winter mixed layer becomes isolated from the atmosphere as the surface restratifies – is crucial in determining the shape and characteristics of the gyre.
Huang and Qiu [1998] performed an LPS calculation for the South Pacific forced by climatological winds. They noted the very deep late winter mixed layer along the sub-polar front (between 40°S and 50°S), where the sharp meridional gradient of outcropping isopycnals allows subduction and ventilation to at least sigma 27, contrasting with the North Pacific where the outcrops occur at warmer temperatures. This has two important consequences from the LPS point of view: First, the South Pacific wind-driven gyre extends very deep. The 1500m depth extent of the NCJ noted here as well as by other observational studies [Gourdeau et al., 2008; Ganachaud et al., 2010; Gasparin et al., 2011] can be seen as due to the influence of the sub-polar front, carried to the subtropics by equatorward flow in subsurface layers. Second, the strong tilt of the South Pacific subtropical gyre is implied by the ventilation of a larger range of isopycnals than in the North Pacific. The strong tilt results in more shear between the upper and lower layers and thus a broad latitude band where eastward surface currents overlie deeper westward flow. In turn, these vertically-sheared zonal currents contribute to the tilted bifurcation of the western boundary current system by their pattern of inflow and outflow to the boundary layer.

6.4 Remaining open questions

This work leaves several important observational and dynamical questions that could not be answered from the present data, and that present fruitful topics for further research.
The “direct” inflow from the northern Coral Sea to the Solomon Sea

The observations presented here show that a substantial portion of the shallow North Vanuatu Jet turns north into the Solomon Sea before reaching the coast of Australia. This is attested to by several pieces of evidence: First, a high-salinity tongue on sigma 24.5 (about 150m depth in this region) extends into the Solomon Sea east of the western boundary current (Fig. 8a); second, glider observations show the correspondence of high salinity marking the NVJ along 163°E with that across the mouth of the Solomon Sea (Fig. 9); third, the measured velocities on sigma 24.5 in the heart of this water mass break off near 156°E from the main NVJ inflow (Fig. 10a).

Inflow from the NVJ is consistent with the upward extension of the New Guinea Coastal Undercurrent maximum: from near 400m south of the Louisade Archipelago to a much shallower value inside the Solomon Sea (compare Figs. 11e and f), which is also seen in Vitiaz St. [Lindstrom et al. 1990]. Similar conclusions have been reached by others studying several different data sets [QL02; Cravatte et al., 2011; Gasparin et al., 2012; Davis et al., 2012; Hristova and Kessler, 2012] and in models [Kessler and Gourdeau, 2007; Couvelard et al., 2008; Melet et al., 2010]; this has been called the "direct" inflow to the Solomon Sea. Gasparin et al. [2012] estimated this element of equatorward transport at 7 Sv during a single cruise in August 2007; given our large uncertainties, that value is consistent with the results here. The
issue is important because theory and repeat glider observations suggest this direct connection between the North Vanuatu Jet and Solomon Sea transport varies more quickly and substantially with changes of the ENSO cycle than the deeper western boundary connection [Davis et al., 2012]. Thus the dynamics controlling this northward turn of the NVJ are key to understanding the link between the South Pacific and the equator, especially at annual and interannual timescales.

It is not clear why this direct transport occurs at all, since the wind stress curl is strongly negative across the mouth of the Solomon Sea and remains so throughout the climatological annual cycle; thus the Sverdrup transport is southward (Fig. 3b,c). But given the quite different data sets that have identified this flow over many years, it seems unlikely to be an artifact of inadequate sampling. Thus one is left to conclude either that a deeper balancing southward flow exists, that topographic effects turn the current northward (though this is not obvious in the nonlinear model experiments of Couvelard et al. [2008] that explored the role of topographic steering), or that another nonlinear process is responsible; perhaps that the northern edge of the North Vanuatu Jet "feels" the fast boundary current wrapping around the Louisiade Archipelago – about 100km outward from the island chain itself – further south and east than the actual coastline.
A coherent dynamical explication of the northward mid-basin turn of the NVJ remains to be accomplished, and is essential to an understanding of the subtropical-equatorial connection.

**Southward flow exiting the Solomon Sea in the east**

The observations presented here, like those of others, have shown a shallow southward flow exiting the eastern Solomon Sea (Fig. 10a); vertical sections (not shown) show this current to be surface-intensified and concentrated above 200m. A similar current pattern was described from the ADCP compilation of Cravatte *et al.* [2011], synoptic cruise data of Gasparin *et al.* [2012], surface drifter observations of Hristova and Kessler [2012], and glider observations of Davis *et al.* [2012], possibly indicating a semi-permanent cyclonic eddy filling the eastern Solomon Sea. Gasparin *et al.* [2012] suggested that the southward flow had entered the Solomon Sea from the north, but it also may be a recirculation of the direct inflow from the NVJ discussed above.

Is this southward flow entering the Coral Sea from the north a consistent feature? What is the origin and destination of its transport? Is it an additional source of mass or property transport to the Coral Sea, or does it simply recirculate within the eastern Solomon Sea?

**A western boundary current on the eastern side of the Queensland Plateau?**

The North Caledonian Jet splits around the Queensland Plateau, about 100km east of the Australian mainland at 15.5°-18°S (section 4.2); most of the flow seemed to turn north, but we
were unable to quantify the split or verify if flow east of the Plateau is boundary-current-like. Is there a double western boundary current east and west of the Plateau? Do they occur in the same depth ranges? Do they vary synchronously or at different timescales? Are there consequences for property transport and the redistribution of arriving properties along the western boundary?

_Eddies and circulation in the Gulf of Papua_

Strong eddies extend to at least 1000m depth in the Gulf of Papua [Scully-Power, 1973, and the Argo float trajectories noted in section 5.1), but the properties carried north by the GPC seem to pass through unaffected (note the similarity of the oxygen sections in Fig. 13c,d,e). Is it true that these eddies do not accomplish significant mixing? Dynamically, do the eddies represent boundary current overshoot (section 3.2) or do they propagate into the Gulf from the east? Do they have a preferred size, direction of rotation, or timescale?

The Gulf is a meeting-place of water from the NCJ and part of the NVJ, with the westward inflow, northward boundary current, and eddy circulations occurring close to each other in a small region (Fig. 10). It is likely that the strong seasonal and ENSO signals of the NVJ [Kessler and Cravatte, 2013] make this system quite variable. What observational tools are required to adequately sample this complex set of interactions? Can we test and trust model representations of it?
**What causes the great depth of South Pacific circulation?**

We and others have noted the deep extension of westward currents in the South Pacific compared to other subtropical gyres (section 4.2), especially the NCJ, and cited the role of deep isopycnal outcropping at the sub-polar front as in the LPS model of Huang and Qiu [1998]. Is this theory an adequate mechanism to distinguish the South Pacific from the other subtropical oceans? Does interannual, decadal or climate change variability at the front imply corresponding changes at depths to sigma 27 in the western gyre? Can this be tested from observations?

**How can the redistribution of the SEC be measured?**

We have described two main sources of Coral Sea inflow that are relatively well-defined, but were not able to quantitatively specify the mass or property fluxes that turned poleward or equatorward (section 6.2). Quantifying this bifurcation, which is the final split of South Pacific subtropical gyre water, is the principal unsolved climate problem of the Coral Sea and was a primary aim of this study. In the end, despite being able to reference much of the geostrophic flow using Argo trajectories, we found that significant fractions of the transport occur close to coastlines and reefs, or over shallow topography; since these regions are sampled only sporadically, we were unable to make a credible estimate of the bifurcating transports. As the Argo time series lengthens, and with the aid of other broadscale tools (e.g., satellite altimetry),
the interior circulation will become adequately resolved and its sampling uncertainty better
defined, but many of the boundary currents and jets will remain poorly observed by these
networks. It is apparent that a quantitative description of the bifurcating mass and property
transports will require specifically-designed ongoing programs to measure the boundary currents
and near-coastal jets, so developing these programs should be a priority. Several of these efforts
are underway in the context of the SPICE program.

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Appendix

Steps to map the Argo trajectories

Three-dimensional velocity fields were constructed according to the following steps:

1) Argo trajectory, metafile and profile files were obtained from the USGODAE Argo data browser (http://www.usgodae.org/cgi-bin/argo_select.pl), to obtain all floats that ever entered the region 140°E-180°, 35°S-7°S, through May 2011. 208 floats were found, with a total of 15574 dives in the Coral Sea region.

2) For each dive of each float, the parking depth and endpoint times and positions were determined from the trajectory, metafile and profile files, when available. Positions on land were discarded. A float might not drift at its assigned parking depth because it could be bouncing over shallow topography or have internal problems or inconsistent file information. We found that the grounded flags and cycle numbers in the trajectory files, which would have been very useful in this step, were inconsistent and could not be relied upon. In some cases, the parking depth could not be determined, in which case the assigned parking depth was assumed. Dives that had a parking depth determined in this manner at or deeper than Smith and Sandwell bathymetry at their location were discarded. Also, any floats with parking depths less than 800m were discarded in this step. After 452 discards, a total of 15122 usable dives were found.
3) Dive time and distance were estimated from the float’s transmissions. Since a float typically does not dive immediately after its last transmission, nor transmit immediately after arriving at the surface, we tried to estimate the unsampled surface time and drift following Park *et al.* (2005). However, fewer than half of the floats had the required time and pressure information in the trajectory files. Therefore we found the average unsampled surface time from the floats with this information (~2 hr on each end), and applied it to all dives of all floats. (In fact, experiment showed that the final speeds were not sensitive to choices from 2 to 12 hr.)

4) Drift during ascent and descent were estimated. Again we attempted to follow Park *et al.* but for most floats the necessary information (time of start and end of ascent and descent) was missing. As for the surface time, we assumed the average over the floats with these values available (12 hr). We used mean geostrophic shear relative to 1000m from CARS over these average times to estimate the drift during ascent and descent. The Ekman transport could have been incorporated into the estimated ascent/descent drifts, but its value was more than an order of magnitude smaller than the other uncertainties and was ignored.

5) All dives resulting from this step were visually checked for indications of errors (high speeds, large surface drifts, dive times longer than 12 days, etc.) All the raw velocities and trajectories were individually plotted and visually inspected, and 266 questionable dives removed (about 1.8%), leaving 14856 dives finally available for mapping.
6) For floats whose parking depth was not at 1000m, CARS mean geostrophic shear was used to “re-reference” each dive to 1000m (but note that individual dives with parking depths shallower than 800m had been discarded in step 2 above).

7) Raw velocities were found from the resulting distance and time values.

8) Lagrangian zonal and meridional velocities at 1000 m were gridded independently to produce a mean climatological field on a 0.25° longitude, 0.25° latitude grid, using an optimal objective analysis method (De Mey and Menard, 1989). This method gives a value of subsurface velocity at each grid point, using nearby data only. Around each grid point, data were mapped taking into account the spatial scales of the relevant dynamics. The decorrelation scales in the zonal and meridional directions are 50 km where data coverage was good, and 100 km where data coverage was weaker. The scales were also chosen to avoid mapping across islands or reefs. The gridding method gives, in addition to the interpolated velocities, a normalized error at each grid point allowing us to estimate the confidence we can place in the currents we observe. Only currents for which the error is less than 40% of the variance at that point were considered.

9) To reduce small-scale noise, the mapped velocities were smoothed in the direction of each vector by a gaussian weighted average with e-folding scale of 100km.

10) The smoothed mapped 1000m velocities were used as a reference for CARS climatological geostrophic shear (i.e., adding the mapped 1000m velocities to CARS shear relative to 1000m),
to give a 3-dimensional velocity field from the surface to 2000m. Where bathymetry occurs above 1000m (and thus there are no float trajectories to provide referencing), we used CARS geostrophy relative to the bottom. This was especially significant over the roughly 300m-deep Marion Plateau from 19° to 21°S on the coast of Australia, where isopycnals slope upward to the coast and the geostrophic shear is thus strongly southward (Fig. 5a). This appears to be a reasonable choice since it connects regions of like boundary transport at these depths to its north and south (Fig. 10 or 12), however there is no direct physical justification for this choice.

We note that steps 2, 3 and 4 above required ad hoc assumptions about the parking depths, the times of leaving and arriving at the surface, and the ascent/descent times that will have introduced errors of unknown magnitude.
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FIGURES:

Fig. 1. Reference map of the Coral Sea, showing features discussed in the text. Brown shading shows shallow bathymetry at depths of 100, 500 and 1000m. Names of countries are in upper case. “M.P.” indicates the Marion Plateau near 20°S, 151°E, and circled "V", "S" and “G” abbreviate Vitiaz and Solomon Straits, and St. George’s Channel, respectively.

Fig. 2. (a) Mapped Argo drift velocity at 1000m. The scale vector is at lower left, and vectors with zonal component larger than 1.5 cm s$^{-1}$ are colored by their zonal component: red for westward, blue for eastward. (b) CARS geostrophic velocity at the surface relative to 500m. The coloring is the same as for the top panel, except it begins for magnitudes greater than 4 cm s$^{-1}$.

Bathymetry as in Fig. 1.

Fig. 3. (a) Mean wind stress vectors from ERS winds during 1991-2000 (scale vector at bottom left); (b) Mean wind stress curl (10$^{-7}$ N m$^{-3}$) based on the winds above; (c) Sverdrup transport streamfunction (Sv) calculated from Godfrey’s Island Rule and the wind stress curl field (middle).

Fig. 4. Vertical meridional sections of temperature (°C, a), salinity (psu, b) and oxygen concentration (mL/L, c) along 162°E, from the CARS climatology (section 2). White overlaid contours (same on all three panels) show the depths of the isopycnals 24.5, 26.5, 27.2 and 27.4.
This section is just west of New Caledonia, whose latitude range is indicated by a solid black bar at the top.

**Fig. 5.** (a) 0-1000m vertically-integrated transport from the referenced geostrophic currents (scale vector at lower left). (b) Island Rule transport vectors comparable to (a), with the same scale (but half as densely plotted). Note that Island Rule-inferred western boundary currents and jets at the tips of islands are necessarily a single gridpoint wide. Bathymetry as in Fig. 1.

**Fig. 6.** Observed (dashed) and Island Rule-estimated (line) transport (Sv) extending north (a, along 162.5°E) and south (b, along 165.5°E) from New Caledonia. The Island Rule jets occupy the 1° latitude-wide region immediately adjacent to the island.

**Fig. 7.** The number of Argo drift velocity samples in each 1° by 1° bin.

**Fig. 8.** Maps of salinity (psu, left panels) and oxygen concentration (ml/L, right panels) on the isopycnals 24.5 (a,d), 26.5 (b,e) and 27.2 (c,f), from the CARS climatology (section 2). Each map has its own color scale and contour interval. Regional-mean depths of each isopycnal are indicated at bottom left. Values on the shallow isopycnal 24.5 are only plotted for depths greater than 100m. Bathymetry as in Fig. 1 is shown if the bottom is shallower than the isopycnal.
**Fig. 9.** Salinity anomaly sections (psu, common scale at bottom) along glider lines from the southern Solomon Islands to New Caledonia (right), and to the tip of the Louisiade Archipelago (left), oriented so the Solomon Island start point is at the center (bottom keymap shows the glider lines in green). Each upper panel shows the anomaly along isopycnals from the mean over both sections. The section on the right is from the single glider mission in mid-2005 [Gourdeau et al., 2008], showing latitude along the section. The section on the left is a composite of many sections over 4 years, binned according to fractional distance from the Louisiades (0) to the Solomons (1) [Davis et al., 2012].

**Fig. 10.** Referenced geostrophic velocity vectors on isopycnal surfaces. (a) on sigma 24.5, with mean depth 126m over this region. (b) on sigma 26.5, with mean depth 398m. (c) on sigma 27.2, with mean depth 822m. Scale vectors are shown in white over Australia; the upper two panels use the same scale, the lower panel has an increased scale length. Bathymetry as in Fig. 1.

**Fig. 11.** Velocity sections across the key regional currents. Each panel is 3.5° latitude or longitude wide, and all have the same color scale (cm s$^{-1}$, bottom). Red colors indicate eastward or northward velocity, blue westward or southward, and the light-blue arrows pointing from the core of each current to green bars on the reference map (middle) show the location of each section. The three panels at lower right (g,h,i) show the incoming branches of the SEC (NVJ: g,
NCJ: h, SCJ: i). The other sections all have the coast on the left. White overlaid contours show sigmatheta, every 0.5 kg m$^{-3}$. The specific current abbreviations are given below the reference map.

Fig. 12. (a) Island Rule (red) and horizontally and 0-1500m vertically integrated observed (black) western boundary transport (Sv) along the coast of Australia (see section 5.1). (b) Meridional component of observed velocity (cm s$^{-1}$) averaged across the western boundary layer (see section 5.1). Red indicates northward, blue southward; nonlinear color scale at right. White contours show isopycnals. The blank (white) region near 20°S has been omitted because the western boundary jumps eastward too far to be considered continuous.

Fig. 13. Oxygen concentration sections (mL/L, scale at lower middle) from the CARS climatology, corresponding to the velocity sections in Fig. 11. White overlaid contours show sigmatheta, every 0.5 kg m$^{-3}$.

Fig. 14. Mean oxygen concentration (mL/L) along the coast of Australia, from the CARS climatology. White overlaid contours show sigmatheta. Dark solid contour indicates 4.0 mL/L; the overlaid dark dashed contour shows the corresponding value from a section along 157°E.

Fig. 15. Approximate (and unbalanced) transport budgets (Sv) in density layers across the sides of a box enclosing the Coral Sea. Black shading is land, and brown shading indicates depths at
100m, 500m and 1000m. The colored transport bars are oriented inward or outward according to the direction of transport, with the same color scheme for each side (labeled for the 162.5° section) and dashed lines indicate the transport scale. The central outlined box shows the net inward transport (imbalance) in each layer. Bathymetry as in Fig. 1.
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