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Roles of the ocean mesoscale in the lateral supply of mass, heat, carbon and nutrients to the Northern Hemisphere subtropical gyres

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Abstract

Lateral transport at the boundaries of the subtropical gyres plays a crucial role in providing the nutrients that fuel gyre primary productivity, the heat that helps restratify the surface mixed layer, and the dissolved inorganic carbon (DIC) that influences air-sea carbon exchange. Mesoscale eddies may be an important component of these lateral transports; however, previous studies have not quantified the tracer transport due to eddies. Here we assess the physical mechanisms that control the lateral and vertical transport of mass, heat, nutrients and carbon across the North Pacific and North Atlantic subtropical gyre boundaries using the eddy-rich ocean component of a climate model (GFDL’s CM2.6) coupled to a simple biogeochemical model (mini-BLING). Our results suggest that lateral transport across the gyre boundaries supplies a substantial amount of mass and tracers to the ventilated layer of both Northern Hemisphere subtropical gyres, with the Kuroshio and Gulf Stream acting as main exchange gateways. Mass, heat, and DIC supply is principally driven by the time-mean circulation, whereas nutrient transport differs markedly from the other tracers, as nutrients are mainly supplied to both subtropical gyres by down-gradient eddy mixing across gyre boundaries. The budget analysis further reveals that the lateral nutrient transport, combining the roles of both mean and eddy components, is responsible for more than three quarters of the total nutrient supply into the subtropical gyres, surpassing a recent estimate based on a coarse resolution model and thus further highlighting the importance of lateral nutrient transport.

1 Introduction

The dynamics of the primarily wind-driven upper kilometre of the ocean’s subtropical gyres play a critical role in regulating the climate, ocean carbon dioxide (CO$_2$) uptake, and the subtropical ecosystem [Huang and Qiu, 1994; McClain et al., 2004]. More than a third of the Earth’s total meridional heat transport at the latitudes of the subtropical gyres is carried by the ocean, and the convergence of this oceanic heat transport at the subtropical gyre latitudes implies vigorous heat loss to the atmosphere [Trenberth and Caron, 2001]. Subtropical gyres in the Northern Hemisphere comprise the largest carbon sink for the contemporary atmospheric CO$_2$ on an annual basis, due to the low partial pressures of CO$_2$ at their surface [Takahashi et al., 2009]. Moreover, subtropical gyres are home to the ocean’s largest biome owing to their vast surface area, covering roughly 40% of the global ocean [McClain et al., 2004; Letscher et al., 2016]. Thus, understanding what processes supply heat, carbon,
and nutrients to the subtropical gyres is critical to characterize ocean climate, biogeochemistry, and ecology.

In contrast to the neighbouring subpolar gyres and the tropics, where large-scale Ekman suction upwells cold, nutrient-rich water, the subtropical gyres are regions of large-scale downwelling that deepens the thermocline and nutricline [Williams and Follows, 1998; Wilson and Coles, 2005; Omand and Mahadevan, 2015]. The contrasting dynamics of the subtropical gyres and their surroundings leads to strong property gradients at the gyre boundaries. Thus, it is natural to hypothesize that cross-boundary exchange could have profound implications for gyre tracer budgets [Bower et al., 1985; Williams and Follows, 1998; Ayers and Lozier, 2010; Palter et al., 2013]. This cross-boundary exchange, however, is extremely challenging to measure directly, and most global ocean models cannot resolve the mesoscale eddying motions that are ubiquitous at these boundaries, being forced to parameterize their effects instead. Here, we explore the role of cross-boundary exchange on heat, carbon and nutrient budgets, making use of a climate model that resolves a rich spectrum of ocean mesoscale eddies in the subtropics [Delworth et al., 2012; Griffies et al., 2015] and is coupled to a simplified biogeochemical model [Galbraith et al., 2015].

For instance, since cross-boundary heat exchange can influence the subtropical mixed layer temperature, such exchange may impact the reemergence of climate signals, whereby wintertime thermal anomalies sequestered in the deep mixed layer are re-entrained into the mixed layer in the following winter [Alexander and Deser, 1995]. Such reemergence of thermal anomalies is thought to possibly cause persistence of climate signals, such as the North Atlantic Oscillation, on interannual and longer time scales [Kwon et al., 2010; Cassou et al., 2007]. In addition to the potential role of heat supply to the subpolar regions on climate variability, cross-boundary heat exchange also influences the solubility of CO$_2$ in the surface waters. Moreover, anomalies of dissolved inorganic carbon (DIC) crossing the subtropical gyre boundaries can further influence the subtropical uptake of atmospheric CO$_2$ within the ocean [Ayers and Lozier, 2012].

The dynamical supply of nutrients into the subtropical euphotic zone was once thought to be dominated by vertical processes such as mixing and advection from intermediate depths. This paradigm, however, gave rise to a long standing puzzle, as new production outpaced the known vertical supply of nutrients to the subtropical gyre euphotic zone [e.g., McGillicuddy Jr et al., 1998; Oschlies and Garçon, 1998]. An increasing number of recent studies have
replaced this paradigm with one that includes lateral processes in supplying nutrients to the subtropical gyre from the neighbouring nutrient-rich subpolar gyres and tropics [Williams and Follows, 1998, 2003; Oschlies, 2002; Ayers and Lozier, 2010; Palter et al., 2013]. Recently, Letscher et al. [2016] used an observationally-constrained, coarse-resolution model to propose that roughly half of the dissolved inorganic phosphate (PO$_4$) required to close the nutrient budgets in the subtropical gyres is supplied by lateral transport. Understanding lateral transport of PO$_4$ into the subtropical gyres is further motivated by recent studies which revealed that subtropical phytoplankton have elevated carbon to phosphorus ratios, implying a greater organic carbon transport to the deep ocean for every mole of PO$_4$ that enters the gyre than if it were consumed elsewhere [Teng et al., 2014; Galbraith and Martiny, 2015].

This high carbon ratio in subtropical export productivity also has implications for the oceanic storage and air-sea partitioning of CO$_2$. Despite advances in understanding the role of lateral circulation in subtropical nutrient supply, important questions remain unanswered: Where is the lateral exchange strongest? And what is the role of mesoscale eddies?

One reason the answers to these questions have remained elusive is that they require observations or models that resolve swift boundary currents and the mesoscale eddies associated with them. These oceanic jets, which mark the western edges of the subtropical gyres, are known to be the regions which serve as both “blender” and “barrier” for cross-frontal tracer exchanges [Bower et al., 1985]. On the one hand, the jets act as “blender” for tracers, in part due to strong down-front westerly winds that drive Ekman transport across the currents [e.g., Williams and Follows, 1998; Ayers and Lozier, 2010; Palter et al., 2011, 2013]. Furthermore, these highly baroclinic western boundary currents are regions of enhanced mesoscale eddy activity due to the heightened available potential energy [Williams and Follows, 2003; Griffies et al., 2015]. These mesoscale motions, collectively referred to as eddies, are ubiquitous features in the ocean and persist over time scales of weeks to months and horizontal scales from tens to hundreds of kilometres [Bishop et al., 2013]. Eddies, including rings shed from frontal systems, can induce significant cross-frontal exchange [Samelson, 1992; Lee and Williams, 2000; Qiu et al., 2007]. Observational studies have estimated that such eddy structures transport a substantial amount of heat and salt [Zhang et al., 2014; Dong et al., 2014; Sasaki and Minobe, 2015]. On the other hand, the swift currents associated with these oceanic jets are thought to act as “barriers” for tracer exchange by suppressing eddy-driven mixing at the shallower depths where the speed of the jet is much faster than the propagation speed of its meanders [Bower et al., 1985; Bower, 1991; Samelson, 1992;
The role of eddies in the Northern Hemisphere subtropical gyres, however, has not been fully resolved:

While the rings shed from the Kuroshio are thought to play a significant role in the potential vorticity budget in the North Pacific subtropical gyre region [Qiu and Chen, 2006; Qiu et al., 2007], rings shed from the Gulf Stream may play only a minor role in the subtropical tracer budgets [Bower et al., 1985]. Given these contrasting properties of western boundary currents and the conflicting reports on the impact of mesoscale rings in tracer budgets over the two Northern Hemisphere subtropical gyre basins, it is difficult to anticipate the role of mesoscale motion in the lateral transport of mass and tracers into subtropical gyres without a model that resolves these motions.

The purpose of this work is to quantify the transport of heat, carbon, and nutrients into the North Pacific and North Atlantic subtropical gyres. Our ultimate goals are to assess the physical mechanisms that control the transport of these tracers into the interior of the subtropical gyres all along their boundaries; evaluate their importance in gyre heat, DIC, and nutrient budgets; and understand the spatial variability of these transports, as well as any contrast that may emerge between the North Atlantic and the North Pacific subtropical gyres.

In order to achieve these goals, we use a preindustrial control simulation from an eddy-rich ocean-atmosphere climate model coupled to a simple marine biogeochemical model (CM2.6-miniBLING). We quantify the annual transport across the boundaries of the subtropical gyres and further decompose it into mean and eddy components. CM2.6-miniBLING provides an adequate tool for our purposes, with its horizontal resolution of 0.1˚ allowing us to sufficiently resolve the mesoscale eddies of interest.

The rest of the paper is organized as follows. In Section 2, we describe the model used (Section 2.1), our definition of the lateral (Section 2.2.1) and vertical extent of the subtropical gyres (Section 2.2.2), and the decomposition of the tracer transport across the gyre boundaries into mean and eddy components (Section 2.3). In Section 3, we discuss the hot spots of the lateral mass transport (Section 3.1), relative roles of advection and down-gradient mixing in the lateral tracer supply to the subtropical gyres (Section 3.2), and the contribution of lateral transport to the tracer budgets in the subtropical gyres (Section 3.3). A discussion and conclusion are provided in Section 4.
2 Methods

2.1 CM2.6-miniBLING

Our primary tool is an eddy-rich coupled climate model, Geophysical Fluid Dynamics Laboratory (GFDL) Climate Model version 2.6 (CM2.6) [Delworth et al., 2012; Griffies et al., 2015], coupled to a simplified version of the Biogeochemistry with Light Iron Nutrients and Gas (miniBLING) model [Galbraith et al., 2015]. The atmospheric and land components of CM2.6 each have a horizontal resolution of 0.5°, whereas the oceanic and sea ice components have 0.1° resolution. The ocean component of the climate model is based on the Modular Ocean Model, version 5 (MOM5) [Griffies, 2012], configured using the Boussinesq approximation with 50 vertical levels and the z* vertical coordinate. The vertical cell thickness increases from roughly 10 m in the upper ocean to 210 m in the deep ocean. There is neither lateral tracer diffusion nor a mesoscale eddy parameterization in this model. However, submesoscale mixed layer eddy transport is parameterized according to Fox-Kemper et al. [2011]. Vertical mixing processes are parameterized in the model using the K-profile parameterization scheme [Large et al., 1994], the internal gravity wave breaking scheme [Simmons et al., 2004], as well as the coastal tide mixing scheme [Lee et al., 2006].

MiniBLING is a simplified version of the prognostic biogeochemical model, BLING [Galbraith et al., 2010], developed in order to reduce computational cost for use in eddy-resolving models, while still simulating essential aspects of bulk ecosystem dynamics [Galbraith et al., 2015]. Its three prognostic tracers are dissolved inorganic carbon (DIC), dissolved oxygen (O_2), and a macronutrient that is initialized from a nutrient climatology as the average of phosphate (PO_4) and nitrate (NO_3), weighted according to their Redfield ratios (PO_4/2 + NO_3/32). Therefore, the macronutrient concentrations have values similar to oceanic PO_4 concentrations. Also, because there is no representation of nitrogen fixation or denitrification, this nutrient tracer is more indicative of PO_4 cycling, and hereafter we refer to this macronutrient as “PO_4”. By acting as the limiting macronutrient in the subtropics, however, the nutrient tracer may behave more like NO_3. The reduction of the number of prognostic tracers was achieved by removing dissolved organic matter, the prognostic iron tracer, and the prognostic alkalinity tracer from the original BLING biogeochemical model. To deal with the absence of a dissolved organic phosphorus pool, the model has a fast-recycling term to return organic phosphorus to phosphate. Alkalinity is diagnosed from a climatological alkalinity-salinity relationship, and the iron cycle has been replaced by prescribing a monthly
iron climatology generated by a 1° version of the climate model ESM2M using BLING [Galbraith et al., 2015]. Despite some biases, including in the seasonal cycle of biomass and export production in comparison to the more comprehensive model TOPAZ, miniBLING was shown to successfully simulate large-scale biogeochemical cycling at the global scale in EMS2M [Galbraith et al., 2015].

The CM2.6-miniBLING preindustrial control simulation is run with atmospheric CO$_2$ fixed at the preindustrial concentration of 286 ppm. The simulation is run for 200 years starting from rest, with the miniBLING component being included from year 48. Temperature, salinity, PO$_4$ and O$_2$ fields are initialized with data from the World Ocean Atlas 2009 (WOA) [Locarnini et al., 2010; Garcia et al., 2010a,b], and DIC fields from the Global Ocean Data Analysis Project (GLODAP), which were adjusted to preindustrial values [Key et al., 2004]. In the following analysis, we utilize the last 10 years of the simulation, corresponding to years 191 to 200. The analysis was limited to ten years due to data storage constraints; however, we assume that the averaging period is long enough to eliminate bias due to most interannual variability, though the averaging period is not long enough to eliminate biases due to decadal fluctuations.

A comparison of the modelled fields and observations is shown in Figure 1. The overall observed global pattern is captured by the model, particularly over the regions of interest indicated by black contours on the panels on the right column. However, some deviations are apparent: In general, the model has a cold bias, while it overestimates the DIC field. The cooler simulated surface temperature compared to the present-day observations is an expected outcome of the model spin-up with the preindustrial level of CO$_2$. The discrepancy in DIC likely results from the cooler temperatures and higher CO$_2$ solubility. A band-like negative temperature anomaly in the North Atlantic suggests a southward shift of the North Atlantic Current in the model relative to the observations. However, as discussed later in Section 2.2.1, our choice of gyre boundaries uses a dynamical definition based on model fields, so that such misalignment of observed and simulated currents should not influence our interpretation of the processes giving rise to cross-boundary transport. Some potential causes of errors for our cross-gyre boundary study include more pronounced gradients of the modelled nutrients and DIC at the northern subtropical gyre boundary in the North Pacific. This deviation from the observations could induce larger down-gradient eddy-induced tracer mixing in our result compared to the observations. The model, however, adequately captures the magnitude of the cross-gyre contrasts and the spatial scales that separate them. For a de-
Figure 1. Surface fields for (left) the average over the last 10 years of the CM2.6-miniBLING preindustrial control simulation, (centre) observations, and (right) the difference between the CM2.6-miniBLING simulation and observations (modelled field - observation), for (top to bottom) DIC ($\mu$mol kg$^{-1}$), PO$_4$ ($\mu$mol kg$^{-1}$), O$_2$ ($\mu$mol kg$^{-1}$), and temperature (°C). The modelled PO$_4$ field is compared with the observed PO$_4$/2 + NO$_3$/32, as described in Section 2.1 in the text and labelled “PO$_4$”. The observational fields of temperature, PO$_4$, NO$_3$ and O$_2$ are from the World Ocean Atlas 2009 [Locarnini et al., 2010; Garcia et al., 2010a,b] while DIC is from GLODAP corrected to preindustrial values [Key et al., 2004]. The observational fields are interpolated onto the CM2.6 grid cells. The black contours shown on the right panel indicate the location of the 10-year mean subtropical gyre boundary in both basins (see Figure 2 for the gyre boundary positions of all months and Section 2.2.1 for a definition of the gyre lateral boundaries).

2.2 Definition of the Subtropical Gyres

The subtropical gyres are pools of nutrient-depleted warm water, circulating anticyclonically, driven primarily by winds [Stommel, 1958]. Although the gyres have been the subject of study for decades, there is no standard definition for the subtropical gyre boundaries. For example, previous studies have investigated subtropical gyre properties inferred from hydrographic or satellite observations with a range of definitions, typically based on surface properties, such as chlorophyll content and wind stress curl [Nicholson et al., 2008;
Lozier et al., 2011; Kwon et al., 2016; Oschlies, 2002; McClain et al., 2004; Irwin and Oliver, 2009; Letscher et al., 2016]. Here, we define the subtropical gyre boundaries combining the fundamental ideas that the western edges of the subtropical gyres are the swift western boundary currents, and that the surface of the subtropical gyre contains a pool of warm, well-oxygenated, low-nutrient water, whose vertical limit is marked by the base of the ventilated pycnocline. Thus, the subtropical gyre boundary definition in this study combines both the property-based and dynamical definitions of the subtropical gyres.

2.2.1 Lateral Extent

In the current study we define the subtropical gyre’s lateral extent using the monthly climatology of the two-dimensional barotropic mass quasi-stream function (Ψ) integrated over the upper 1,000 m. We chose 1,000 m for the base of the quasi-stream function, as our interest lies in the upper ocean. Because the Gulf Stream and Kuroshio extend over a depth of approximately 1,000 m, this definition aligns better with the jets than integrating to the ocean bottom. Note that the true mass stream function is defined only for divergence-free flows, and since the 1,000 m lateral transport is not required to be divergence-free, we call the integral of this transport the mass quasi-stream function. The gyre boundary is insensitive to the depth of integration above 1,000 m: For instance, integrating over the top 500 m instead of 1,000 m does not significantly change the overall gyre’s areal extent (not shown). For each basin, we picked one value of the mean mass quasi-stream function which encloses the largest areal extent for every climatological month. In this manner, Ψ = 20 Sv (1 Sv = 10^9 kg s^{-1}; note that here we use a mass Sverdrup) and Ψ = 25 Sv were chosen for the North Pacific and the North Atlantic, respectively (Figure 2). This approach is similar to that used by Palter et al. [2013], who defined the subtropical gyre using the largest closing sea surface height contour. This definition of subtropical gyres provides a dynamical definition and has the advantage that the boundary is precisely aligned with the monthly-mean meandering jet on the western and northern edges of the subtropical gyre. The qualitative results and interpretation of the mass and tracer transport across these boundaries are not sensitive to which stream function is chosen for the boundary, so long as it is within the swift jet. However, if the chosen boundary did not align with the jet, then any jet meanders across the boundary would appear as local hot-spots of cross-boundary exchange (not shown).

Significant month-to-month variability in the areal extent of the gyres for both basins is evident in Figure 2, ranging from 2.7 × 10^6 to 6.9 × 10^6 km^2 in the North Atlantic and 1.2
Figure 2. Monthly climatological positions of the subtropical gyres in this study, defined by using monthly mean two-dimensional barotropic mass quasi-stream function integrated over the upper 1,000 m. a) $\Psi = 20$ Sv was chosen for the North Pacific and b) $\Psi = 25$ Sv was chosen for the North Atlantic subtropical gyres as the quasi-stream function that encloses the largest areal extent over the whole year.

$10^7$ to $2.1 \times 10^7$ km$^2$ in the North Pacific. The seasonal meridional shift in the southern gyre boundaries corresponds to that of the surface wind stress curl (not shown), while the northern gyre boundaries are closely tied to the Gulf Stream and Kuroshio dynamics.

2.2.2 Vertical Extent

We set the vertical limit of our study region to coincide with an isopycnal level just denser (0.1 kg m$^{-3}$) than the densest isopycnal level that outcrops in March within the subtropical gyre. We refer to the layer above this isopycnal as the ventilated pycnocline, following Luyten et al. [1983]. By choosing an isopycnal that is just denser than the outcropping isopycnal, we ensure that the lateral transport of mass and tracers that we wish to quantify are across the lateral gyre boundary set by the streamline defined in Section 2.2.1, rather than across an isopycnal outcrop within that boundary (a summary schematic at the end of the
paper, Figure 9, shows the isopycnal and gyre boundary). Additionally, under surface buoyancy loss, fluxes can extend into the stratified interior [Large et al., 1994], so choosing an isopycnal slightly denser than the outcropping density ensures that we capture these fluxes. The oxygen cross sections in Figure 3 suggest that the deepest ventilated layers are above \( \sigma_\theta = 25.8 \text{ kg m}^{-3} \) (which corresponds to a seawater potential density of 1.025.8 kg m\(^{-3}\)) in the North Pacific, and \( \sigma_\theta = 26.6 \text{ kg m}^{-3} \) in the North Atlantic. Thus, we set the base of the vertical layer at these isopycnals, which divide the warm, nutrient- and DIC-depleted, and oxygen-rich upper layer from the cool, nutrient- and DIC-rich, and oxygen-poor lower layer (Figure 3). Not coincidentally, this is the deepest layer that the atmosphere can influence directly in the subtropical gyre, and in our analyses we integrate the subtropical gyre properties and the cross-boundary transports from the surface to these isopycnal levels. The annual mean depth of the ventilated pycnocline for each basin is shown in Figure 4a and b, illustrating that the deepest layers are found just south of the Gulf Stream and the Kuroshio. Like our lateral gyre boundaries, the depth of the ventilated pycnocline varies monthly according to a 10-year average seasonal cycle. In both gyres, the maximum area-weighted mean depth is found in late winter (Figure 4c).

This definition of the ventilated pycnocline in both subtropical gyres also encapsulates the subtropical mode water (STMW). STMW is a layer of weakly stratified water and it exists by virtue of wintertime convection at the poleward edge of the subtropical gyres [Worthington, 1959; Hanawa and Talley, 2001]. At the time of formation, nutrients are consumed at the northern flank of the subtropical gyres in both basins, and this nutrient-depleted water mass is advected southward while being gradually eroded by mixing [Palter et al., 2005; Oka et al., 2015]. STMW in each basin is known to have a characteristic temperature, which is roughly 16 - 19.5°C in the North Pacific [Masuzawa, 1969; Oka et al., 2015] and 18°C in the North Atlantic [Worthington, 1959]. In the model, these temperatures are found between the isopycnal levels \( \sigma_\theta = 24.4 - 25.6 \text{ kg m}^{-3} \) in the North Pacific, and between \( \sigma_\theta = 26.2 - 26.4 \text{ kg m}^{-3} \) in the North Atlantic (Figure 3), thereby roughly corresponding to our ventilated pycnocline. Thus, the model simulates similar densities bounding observed hydrographic properties.
Figure 3. Ten year mean modelled properties along a) 137°E in the North Pacific and b) 66°W in the North Atlantic. (Top to bottom) DIC (µmol kg⁻¹), PO₄ (µmol kg⁻¹), O₂ (µmol kg⁻¹), and temperature (°C) plotted in colour, with potential density overlaid with black contours. The isopycnal level chosen for the vertical extent of the subtropical gyre for each basin in this study (σθ = 25.8 kg m⁻³ in the North Pacific, and σθ = 26.6 kg m⁻³ in the North Atlantic) is highlighted with a thick white contour on each panel.

2.3 Decomposition of the transport into mean and eddy components

The horizontal transport of seawater mass and tracer mass through a vertical grid cell face of the lateral gyre boundaries defined in Section 2.2.1 is written

\[ \Phi = \rho_0 d \hat{n} \cdot UC. \]  

(1)

In this equation, \( \rho_0 \) is the constant reference density for the Boussinesq approximation (\( \rho_0 = 1.035 \text{ kg m}^{-3} \)), \( C \) is the tracer concentration (set to unity for seawater mass transport), and

\[ U = u dz \]  

(2)
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Figure 4. Annual-mean depth of the ventilated pycnocline for a) the North Pacific and b) the North Atlantic basins (colour shades). We divide the gyre boundary (coloured thick lines in panels a and b) into three segments: magenta for the jet region (i.e. the Kuroshio and Gulf Stream), green for the northern gyre boundary (north of 25°N for North Pacific and 30°N for North Atlantic), and black for the southern gyre boundary. These lines are drawn using the 10-year mean barotropic quasi-stream function, as described in Section 2.2.1.

c) Monthly variability of the area-weighted mean ventilated pycnocline depth for (cyan) the North Pacific and (purple) the North Atlantic subtropical gyres (thick lines) and one standard deviation of the interannual variability (shaded envelopes).

is the horizontal velocity field weighted by the grid cell thickness. The normal vector $\hat{n}$ is perpendicular to the vertical area of the cell face, $dzdl$. The horizontal distance $dl$ is either $dx$ or $dy$ according to the orientation of the gyre boundary. For the subtropical gyres, we
choose the normal vector $\mathbf{n}$ to point inward, so that a positive transport means mass is added to the gyre interior.

Following Griffies et al. [2015] and Dufour et al. [2015], we decompose the total transport averaged over 10 years into two terms:

$$\Phi_{\text{total}} = \Phi_{\text{mean}} + \Phi_{\text{eddy}}. \quad (3)$$

The transport

$$\Phi_{\text{total}} = \mathbf{\Phi} = \rho_0 dl \mathbf{n} \cdot \mathbf{UC} \quad (4)$$

is the total transport of seawater mass and tracer mass averaged over 10 years crossing the gyre boundaries. The lateral gyre boundaries are set according to the monthly climatological quasi-barotropic mass streamfunction (Section 2.2.1), which in turn determines the geometric factors, $dl \mathbf{n}$. We accumulate the horizontal tracer transport, $\mathbf{UC}$, every model time step, thus ensuring a proper accounting of temporal correlations. The contribution, $\Phi_{\text{mean}}$, measures transport from the time-mean mass transport acting on the time-mean tracer concentration,

$$\Phi_{\text{mean}} = \rho_0 dl \mathbf{n} \cdot \bar{\mathbf{UC}}. \quad (5)$$

We diagnose the time-mean tracer concentration and time-mean horizontal mass transport directly from the model output, so that the time-mean component captures the climatological seasonal cycle of the transport. Finally, the eddy component is the residual,

$$\Phi_{\text{eddy}} = \Phi_{\text{total}} - \Phi_{\text{mean}} = \rho_0 dl \mathbf{n} \cdot (\mathbf{UC} - \bar{\mathbf{UC}}). \quad (6)$$

This eddy term arises from correlations between fluctuations in tracer concentration and mass transport relative to the climatological monthly-mean transport. This term is dominated by mesoscale fluctuations (e.g., jet meanders, anomalous Ekman transport, rings, etc.), though it also contains fluctuations acting on interannual and longer time scales.

Baroclinic eddies are one category of motion that gives rise to such mesoscale fluctuations, and there are two key processes associated with these eddies. First, baroclinic eddies act to convert available potential energy (input by wind and buoyancy forcing) to kinetic energy [Gent et al., 1995]. Potential energy reduction flattens isopycnals, and thus reduces thermal wind shear. This adiabatic mesoscale eddy transport opposes the wind-driven time-mean flow. As suggested by Gent et al. [1995], this process is commonly parameterized by an eddy-induced advective velocity added to the tracer equation. Therefore, eddies transport
mass and tracer by this isopycnal slumping effect, and this tracer transport can be either up-gradient or down-gradient. In addition to slumping isopycnals, mesoscale eddies stir tracers along isopycnals, thus enhancing the fine scale tracer features (i.e., tracer gradients are intensified). In turn, the enhanced fine scale tracer features allow tracers to be efficiently mixed by molecular diffusion. This aspect of mesoscale eddy transport is commonly parameterized by down-gradient isopycnal tracer diffusion [Solomon, 1971; Redi, 1982; McDougall et al., 2014]. CM2.6 does not use parameterizations of the advective or diffusive transport due to mesoscale eddies; rather, both effects of eddies are at least partially represented through the resolved eddying flow.

3 Results

3.1 Hot spots of lateral mass transport across the subtropical gyre boundaries

Mass and tracers may enter or leave the subtropical gyres at any point along the gyre boundaries. By examining the 10-year averaged transport of mass and tracers across the subtropical gyre boundaries (hereafter cross-boundary transport) as a function of distance around the gyre, we first aim to shed light on hot spots of exchange and their associated physical mechanisms.

Figure 5 shows the 10-year averaged cumulative sum of the total, mean, and eddy mass and tracer cross-boundary transports into the subtropical gyre for each basin. The figure was constructed by taking the cumulative sum of the cross-boundary transport at each 0.1° longitude, starting from the western edge of the gyre boundary at 25°N for the North Pacific and 30°N for the North Atlantic and moving in the clockwise direction. 120°W (North Pacific) and 30°W (North Atlantic) correspond to the easternmost extension of the gyre; the boundary only extends this far east for a couple of months of the year (see Figure 2), so that the curves appear flat around these longitudes for the months when the gyre does not extend that far. The Kuroshio and Gulf Stream regions are labelled “jet”, and the northern and southern gyre boundaries are labelled for transport occurring to the north and south of 25°N for North Pacific and 30°N for North Atlantic (see Figure 4a and b for mean boundary locations). The annual-mean is then weighted by the fraction of the year that the quasi-stream function reaches each longitude. The figure shows that the total mass transport (first row in Figure 5) is almost always inward across the gyre boundaries integrated over the ventilated pycnoclines (defined in Section 2.2.2) for both of the basins. Inward mass transport is
an expected consequence of the large-scale wind stress that causes subtropical convergence and downwelling. This inward transport is dominantly set by the mean transport, consistent with monthly-mean negative wind stress curl and convergent Ekman transport over the entire annual cycle (not shown). The vertical profile of the cross-boundary transport (Figure 6) reveals that this inward transport occurs principally in the upper layer, above \( \sigma_\theta = 23.5 \) in North Pacific and \( \sigma_\theta = 26.2 \) in North Atlantic, further supporting the idea that the Ekman transport is the critical inward transport mechanism.

There is intense cross-boundary mass transport across the Kuroshio and Gulf Stream, accounting for 57\% and 82\% of the total mass transport, respectively. A comparison of the longitudes where cross-boundary transport is strongest in Figure 5 to the jet meander locations in Figure 2 reveals that the mean component of the cross-jet transport is correlated with the curvature of the meandering of the jet system: Anticyclonic curvatures in Kuroshio (e.g. at 133\(^\circ\)E) and Gulf Stream correspond to peaks of inflowing mass, whereas the locations of cyclonic curvatures correspond to local minima of the inflowing mass (e.g. at 136\(^\circ\)E), consistent with observations from Bower and Rossby [1989] and Bower [1991].

The spatial patterns of cross-boundary transport are associated with well-known oceanographic features, even outside the jet regions. In the North Pacific, approximately 5 Sv of mass is transported into the gyre across its southern gyre boundary, which is almost entirely driven by the mean component. This large mass input likely arises from the persistent alignment of the location of the monthly southern gyre boundary in the low latitudes (see Figure 2) with the prevailing easterly winds, which drive mass input by Ekman transport. This pronounced mass input is followed by a dramatic mass export at around between 120\(^\circ\)E and 130\(^\circ\)E, acting to cancel approximately all the mass transported across the Southern Gyre boundary. The location where mass is exported from the subtropical gyre coincides with the location where Kuroshio intrudes into the South China Sea through Luzon Strait [Nan et al., 2015], as well as where the westward North Equatorial Current, which constitutes the southern gyre boundary of the North Pacific subtropical gyre, reaches the Philippine coast and bifurcates into the northward Kuroshio and the southward Mindanao Current [Qiu and Lukas, 1996]. We therefore hypothesize that the large leakage of mass at this location is due to the divergence of the flows associated with the bifurcation of the North Equatorial Current as well as Kuroshio intrusion through Luzon Strait. In the North Atlantic, the mean mass transport across the southern gyre boundary is more uniform, and accounts for 18\% of the total mass input into the gyre basin.
Figure 5. The cumulative sum of the 10-year average transport across the subtropical gyre boundaries, integrated vertically to the base of the ventilated pycnocline (Section 2.2.2) for a) the North Pacific and b) the North Atlantic. The total transport is shown in black, the mean transport in blue, and the eddy transport in orange. The envelopes around the total and eddy components indicate one standard deviation of their interannual variability. The cumulative sum is taken along the boundary in the clockwise direction, starting from the westernmost edge of the jet region and plotted as a function of longitude (see Figure 4a and b for the mean position of the jet and the northern and southern gyre boundaries). The annual mean is constructed by weighting the monthly transport by the fraction of the year that the quasi-stream function reaches each longitude (Figure 2). The right-hand y-axis gives the cumulative sum of the cross-boundary transport divided by the annual-mean gyre area. For the mass transport, division by a reference seawater potential density of $\rho_0 = 1035$ kg m$^{-3}$ converts mass $\text{Sv}$ ($10^9$ kg s$^{-1}$) to a velocity with a unit of m s$^{-1}$. Temperature transport ($\text{Sv} \ ^\circ \text{C}$) is expressed as equivalent heat transport in PW after multiplying by heat capacity. The sign convention is such that an increase in the cumulative sum indicates inward transport into the subtropical gyre.
In both gyres, the most pronounced eddy mass transport is found in the jet regions, as
may be expected from the enhanced mesoscale activity there, with a large interannual vari-
ability. Much of this mesoscale transport, however, does not contribute to the net mass con-
vergence or divergence, presumably due to the large rotational component of mass fluxes.
The rotational component circulates around the eddy potential energy contours, and it does
not contribute to the divergence of advective fluxes [Marshall and Shutts, 1981; Bishop et al.,
2013]. Although any advective flux consists of rotational and divergent components, no
unique method exists for decomposing advective fluxes into these components [Fox-Kemper
et al., 2003]. The effect of the rotational component of the advective flux, however, vanishes
for the sum of these fluxes around the closed contour encircling the gyre, leaving only the
divergent component [Dufour et al., 2015]. After summing around the entire gyre boundary,
the average eddy mass transport is slightly negative (i.e. outward), which is in line with ex-
pectations that the mass export by the eddy component opposes the wind-driven time-mean
flow, as as discussed in Section 2.3. However, the eddy-driven mass transport is so small that
it is statistically indistinguishable from zero, after accounting for the interannual variability.
The small outward eddy mass flux is due to transport near the base of the ventilated pycno-
cline, which competes with inward mass input on lighter layers (Figure 6).

In total, 10.2 Sv and 3.0 Sv of mass is imported across the gyre boundaries for the
North Pacific and North Atlantic subtropical gyres, respectively. After normalization by the
annual-mean gyre area (right axis on Figure 5), these transports are similar in magnitude,
with the North Pacific gyre laterally receiving 19.6 m yr$^{-1}$ and the North Atlantic 17.8 m
yr$^{-1}$. Following mass conservation, these values roughly correspond to mean downwelling,
and they are within the expected range of Ekman downwelling in these subtropical gyre re-
regions [Marshall and Plumb, 2007].

3.2 Relative roles of advection and down-gradient mixing in the supply of DIC,
heat, and PO$_4$ to the subtropical gyres

The spatial pattern of mean- and eddy-driven DIC and heat transport resembles that
of mass transport, as can be seen in Figure 5 and 6. This is not surprising, since the tracers
are advected by the residual mean (or total mass) transport, which is the sum of mean and
eddy advective mass transports. Therefore, in the absence of strong down-gradient, along-
isopycnal tracer mixing, mass and tracer transport should follow similar patterns. The lateral
input of DIC and heat by the mean component into the subtropical gyre together reduces the
capacity to take up atmospheric CO$_2$ in both gyres, given that cross-boundary input of DIC
and heat increase the sea surface partial pressure of CO$_2$ (pCO$_2$) in the gyre [Bates et al.,
1998; Takahashi et al., 2009; Ayers and Lozier, 2012]. Eddies act to remove 27% (31%) of
the DIC supplied by the mean flow in the North Pacific (North Atlantic), and 13% (21%) of
the heat. Much of this outward eddy transport occurs across the jets and the southern gyre
boundaries of both gyres, near the base of the ventilated pycnocline. As was the case for the
mass transport, roughly three times more DIC and heat are transported into the North Pacific
subtropical gyre compared to the North Atlantic. After normalizing by the annual mean gyre
area, these supply terms come to similar values for the Atlantic and Pacific (right axis on
Figure 5).

The eddy-driven transport of PO$_4$ provides the majority of the cross-boundary nutrient
transport in both gyres, a behaviour that is unique among the tracers. The inward PO$_4$
transport by the eddy component explains 59% and 120% of the total transport of PO$_4$, for
the North Pacific and the North Atlantic, respectively (in the North Atlantic the eddy supply
is opposed by the mean, such that it exceeds the total). This eddy-driven flux of PO$_4$ enters
the subtropical gyre principally across the Kuroshio and Gulf Stream, and, to a lesser degree,
over the southern gyre boundary in the North Pacific (Figure 5). Most of the eddy-driven
supply occurs near the base of ventilated pycnocline in both gyres (Figure 6). In the North
Pacific, the mean component also supplies PO$_4$ into the subtropical gyre, while in the North
Atlantic, the transport by the mean flow mainly acts to remove PO$_4$ from the gyre, even while
the mean supply of mass is inward. This difference in the role of the mean flow in the PO$_4$
budget between the gyre basins arises from the fact that the circulation at the base of the ven-
tilated layer in the North Atlantic is flowing out of the gyre, whereas the North Pacific flow
over the entire layer is inward, as can be seen from the vertical profile of the cross-boundary
mass transport in Figure 6. Because the PO$_4$ at the base of the layer is elevated relative to the
shallower depths where concentrations are essentially zero, this small outflow of the mean
circulation at the base of the layer translates to an overall outward PO$_4$ transport due to the
mean circulation.

As was briefly discussed in Section 2.3, the eddy-driven tracer transport that is not
carried by the eddy mass transport can be explained by the down-gradient along-isopycnal
tracer transport by the eddies, which arises due to the correlation between the velocity and
tracer anomalies, $\tilde{h}u'C'$. This down-gradient eddy transport of a tracer $C$, $\chi_C$, across the
gyre boundary on an isopycnal layer can be roughly scaled as follows [Griffies, 2004; Dufour
Figure 6. The 10-year averaged tracer transport across the subtropical gyre boundaries for total (black), mean (blue), and eddy (orange) components for a) the North of Pacific and b) the North Atlantic, binned into isopycnal levels. The deepest density level shown corresponds to the isopycnal at base of the ventilated pycnocline, as defined in Section 2.2.2. The envelopes around the total and eddy components indicate one standard deviation of their interannual variability. The sign convention is such that positive values indicate inward transport into the subtropical gyre. Temperature transport (Sv °C) is expressed as equivalent heat transport in PW.

\[ \chi_C = -\rho_0 \kappa \bar{h} \cdot \nabla \bar{C} \cdot \hat{n} \, dl, \]  

where \( \rho_0 \) is a reference seawater density, \( \kappa > 0 \) is an along-isopycnal eddy diffusivity coefficient, \( h \) is the isopycnal thickness, \( dl \) is the segment of the gyre boundary, and \( \hat{n} \) is the vector normal to \( dl \). The eddy diffusivity coefficient, \( \kappa \), is a function of space and time [e.g. Ferrieri and Nikurashin, 2010], and its variations should be consistent across all tracers. CM2.6 resolves much of the mesoscale variability responsible for this down-gradient mixing in the subtropics, and includes no diffusive closure for such fluxes. Hence, we expect the size of
the tracer gradients to explain the difference between the importance for the eddy-driven PO$_4$ transport relative to that for DIC and heat.

Figure 7 shows the normalized 10-year averaged tracer gradient at every 0.1° longitude along the gyre boundary, with positive values indicating higher concentrations outside the gyre than inside. With this sign convention, down-gradient mixing along a positive gradient would provide a source of tracer to the subtropical gyre. The figure reveals that the tracer gradient is positive for DIC and PO$_4$ (higher values outside the subtropical gyres), whereas it is dominantly negative for temperature (higher values inside the subtropical gyre), as expected from Figure 1. The strongest tracer gradients are found across the Kuroshio and Gulf Stream regions relative to the rest of the gyre boundary for all the tracers. The gradients in DIC concentration across the gyre margins are relatively subdued compared to the other tracers, since DIC variations are small relative to background concentrations (> 1,800 µmol kg$^{-1}$ everywhere in the ocean, as simulated under preindustrial atmospheric CO$_2$). The normalized PO$_4$ gradient along the gyre boundary is approximately 1.5 times larger than the normalized DIC gradient for the North Pacific and nearly 10 times larger in the North Atlantic on average. Therefore, the larger eddy-driven supply of PO$_4$ to the gyres relative to the eddy-driven DIC flux is very likely due to down-gradient mixing across a stark biogeochemical divide.

The PO$_4$ gradient is sharper across the Kuroshio and Gulf Stream, relative to the temperature and DIC gradient, because PO$_4$ concentrations are drawn down to near-zero on the subtropical side of the boundary currents nearly to the base of the ventilated layer. Here, light does not seem to limit productivity, as chlorophyll blooms during deep mixing in the darkest months of winter [Palter et al., 2005]. Iron is also relatively abundant, particularly in the North Atlantic, permitting complete PO$_4$ depletion [Mahowald et al., 2005; Sedwick et al., 2005]. The PO$_4$-depleted water masses formed just south of the jet regions are subducted as STMW and recirculate well below the euphotic zone south of their formation region. Therefore, PO$_4$ is restored on those layers via remineralization [Palter et al., 2005]. These layers rejoin the Gulf Stream and Kuroshio, where they are swiftly transported in what has been referred to as a “nutrient stream” [Pelegri and Csanady, 1991; Williams et al., 2006; Palter and Lozier, 2008; Guo et al., 2012]. Thus, the stark nutrient gradients across the Kuroshio and Gulf Stream are sustained by the presence of these nutrient streams and the depletion of the nutrient just south of the boundary currents. Finally, unlike DIC and tem-
perature, PO$_4$ is not subject to air-sea flux exchange, which tends to smooth out lateral tracer gradients.

It is also noteworthy that the cross-boundary PO$_4$ flux is nearly twice as large in the North Atlantic subtropical gyre as the North Pacific after normalizing by gyre area, as can be seen in the right axis on Figure 5. In contrast, mass, heat and DIC cross-boundary transport per unit area is similar for both gyres. This difference between the two gyres likely stems from the fact that in the Gulf Stream, unlike in the Kuroshio, nutrient concentrations are further enriched by waters imported from outside the subtropical gyre, such as from the tropics and Southern Ocean, in the shallow return pathway of the Atlantic Meridional Overturning Circulation [Palter and Lozier, 2008]. As a result, the nutrient transport along the Gulf Stream is approximately 4 - 5 times larger compared to the Kuroshio [Guo et al., 2012], leading to the sharper PO$_4$ gradient across the Gulf Stream than across the Kuroshio. In addition, the gyre-to-gyre difference may be further enhanced by the greater availability of iron in subtropical North Atlantic due to atmospheric deposition of iron-rich Saharan dust [Mahowald et al., 2005; Sedwick et al., 2005]. Iron is an important limiting nutrient, and the difference in the iron availability between the basins likely leads to a more complete PO$_4$ drawdown in the North Atlantic subtropical gyre and a sharper PO$_4$ gradient at its boundaries, compared to its North Pacific counterpart.

3.3 Importance of the cross-boundary tracer transports for the subtropical gyre budgets

Finally, the importance of the cross-boundary transport is evaluated by performing an annual budget analysis over the gyre regions (Figure 8). Both DIC and heat are almost exclusively supplied by the mean component of the lateral transport across the gyre boundary. The eddies oppose the mean transport, removing about a third of the DIC and a quarter of the heat from both the North Atlantic and North Pacific gyres. Two thirds of the DIC supplied laterally downwells, as expected from the large scale downwelling in the subtropical gyre regions, with the air-sea exchange of carbon as well as the biological consumption and remineralization of DIC playing only very minor roles. We remind the reader that this simulation is run with a preindustrial atmospheric CO$_2$ concentration, and does not include the anthropogenic rise in CO$_2$. Heat is lost from both subtropical gyres to the atmosphere at a rate greater than 30 W m$^{-2}$, which is the same order of magnitude as the heat export by the eddy component and the downwelling. As noted earlier, the supply of PO$_4$ is dominantly set
Figure 7. Cross-boundary tracer gradients. Gradients are computed along the subtropical gyre boundary for a) the North Pacific and b) the North Atlantic basin, averaged over the ventilated pycnocline, and plotted as a function of longitude in a clockwise manner starting from the westernmost edge of the jet region (see Figure 4a and b for the mean boundary locations). The tracer gradient for each month is averaged at each longitude and the annual mean is constructed by weighting the monthly-mean gradients by the fraction of the year that the quasi-stream function reaches each longitude. We choose a sign convention such that a positive gradient means the tracer has lower value inside the subtropical gyre compared to outside. With this convention, down-gradient mixing would correspond with a positive tracer input into the gyres. The thick line indicates the annual mean, while the shading indicates one standard deviation of its interannual variability. The tracer gradients are normalized by the average, large-scale, basin-wide gradient (tracer concentration difference between inside the subtropical gyre and outside, divided by the entire length scale from the centre of the subtropics to the exterior) within [10°N - 45°N]×[120°E -120°W] for North Pacific and [10°N - 45°N]×[90°W - 30°W] for North Atlantic over the top 500 m depth.

by the eddy component of cross-boundary lateral transport for both gyres, but particularly so for the North Atlantic. The total lateral transport, combining the roles of both mean and eddy components, makes up 77% of the total nutrient supply in the North Pacific and 86% in the North Atlantic. These values are roughly 1.5 times larger than the estimate by Letscher et al.
Figure 8. Annual-mean DIC (left), PO$_4$ (middle) and temperature (right) budgets for the North Pacific and North Atlantic subtropical gyres. "Mean" (blue) corresponds to the mean lateral advective transport of each tracer, “eddy” (orange) the eddy lateral advective transport, and “downwelling” (white) the transport across the isopycnal at the base of the ventilated pycnocline, computed as the residual between the 3D advective convergence within the gyre bowl and the total cross-boundary lateral transport. “CO$_2$ flux” (green on the left panel) is the air-sea CO$_2$ flux; “biological sink” (grey on the left panel) is the biological consumption and remineralization of DIC; “net biological sink” (grey on the middle panel) corresponds to the net biological consumption of PO$_4$ integrated over the vertical layer; “vertical diffusion” (green on the middle panel) is the flux of PO$_4$ across the bottom of the layer via parameterized vertical mixing; “submesoscale” (yellow on the middle panel) is the PO$_4$ supply due to the parameterization of submesoscale mixed layer instabilities; and “net sfc heating” (green on the right panel) is the net surface heating including the effect of longwave and shortwave penetrating radiation, as well as sensible and latent heat fluxes. Additional terms, including vertical diffusion and submesoscale transport of DIC and temperature are all at least three orders of magnitude smaller than the shown terms, and therefore excluded from the figure. Note that the imbalance between the source and sink particularly evident for PO$_4$ is due to the non-negligible tendency term, amounting to -2.3 mmol m$^{-2}$ yr$^{-1}$ for the North Pacific and -1.5 mmol m$^{-2}$ yr$^{-1}$ for the North Atlantic subtropical PO$_4$ budgets.

[2016], who evaluated the lateral supply of nutrients in a data-constrained, coarse-resolution ocean model, using fixed lateral subtropical gyre boundaries and the fixed local annual maximum mixed layer depth as the vertical limit. It is unclear if the source of this difference arises from differences in the simulated PO$_4$ gradients, differences in the definition of the gyre boundaries, and/or the fact that our simulation resolves a large spectrum of mesoscale fluctuations that were parameterized in the simulations of Letscher et al. [2016]. In any case, both studies agree that the supply of nutrients to the subtropical gyres is largely governed by lateral exchange across its boundaries, despite important differences in the model used to evaluate the supply and distinct definitions of gyre boundaries.
4 Discussion and Conclusions

In this study we evaluated the transport of mass, heat, carbon, and nutrients into the North Pacific and North Atlantic subtropical gyres over the annual-mean timescale, in a preindustrial simulation of an eddy-rich climate model coupled to a simplified biogeochemistry model (CM2.6-miniBLING). We proposed a new gyre boundary definition based on a combination of dynamical and property-based criteria: The lateral gyre boundaries are defined by the largest region of anticyclonic circulation, whereas the vertical extent is set by the isopycnal that underlies a pool of warm, oxygenated, oligotrophic water, which we call the ventilated pycnocline following Luyten et al. [1983]. The transport across the subtropical gyre boundary thus defined was decomposed into a mean component and a mesoscale eddy component following Griffies et al. [2015] and Dufour et al. [2015].

A schematic summary of the transport of mass and each tracer is shown in Figure 9. Lateral cross-boundary transport supplies a substantial amount of mass, heat, DIC, and PO$_4$ to the subtropical gyre regions in both the North Pacific and the North Atlantic basins on annual-mean basis. Mass, heat, and DIC supply into the gyres are mainly set by the mean component of the transport, and largely achieved across the Kuroshio and Gulf Stream. The concurrent transport of heat and DIC together act to reduce the subtropical ocean uptake of atmospheric CO$_2$ below what it would be in the absence of such lateral exchange, by increasing the pCO$_2$ in the subtropical seawater. Mesoscale eddies tend to remove mass, heat and DIC supplied by the mean component from the gyres. This cancellation of the mean circulation by the eddy-driven circulation is in line with the expected role of baroclinic eddies in flattening isopycnals, which are steepened by the time-mean, wind-driven circulation [Gent and McWilliams, 1990], and agree with previous studies evaluating tracer budgets using the same model [Griffies et al., 2015; Dufour et al., 2015].

Transport of PO$_4$ differs markedly from mass and the other tracers, in that the eddies are the primary supply mechanism of PO$_4$ into the subtropical gyres of both basins. We attribute this greater role for the eddies to the sharp PO$_4$ gradients across the gyre boundaries, which allow for strong down-gradient mixing. The effect of the outward eddy-induced mass transport on PO$_4$ transport is, thus, seemingly swamped by the down-gradient PO$_4$ diffusive transport. In turn, the strong PO$_4$ gradient across the Gulf Stream and Kuroshio relative to the other tracers is interpreted as a consequence of biological activity: PO$_4$ serves as a limiting nutrient for primary productivity in the subtropical gyre regions, resulting in
Figure 9. Schematic of mass and tracer lateral, cross-boundary transport on a cross-section of the subtropical gyres. The lateral boundary, defined in Section 2.2.1, is denoted by the vertical straight lines labelled $\Psi(x, y, t)$. The curved line labelled as $\sigma_\theta(x, y, z, t)$ shows the time-varying vertical limit of the subtropical gyre (the grey shaded area), defined in Section 2.2.2. The blue straight arrows indicate the mass and tracer time-mean advective cross-boundary transport. Wavy arrows show the eddy cross-boundary transport for DIC and heat (orange) and PO$_4$ (green).

The near-complete depletion of PO$_4$ to the base of the subtropical euphotic zone or mixed layer, whichever is deeper. On isopycnals beneath the euphotic zone or mixed layer, nutrients are restored via remineralization during recirculation before rejoining the Kuroshio and Gulf Stream. Because the deepest mixed layers are formed just south of the Gulf Stream and Kuroshio, the largest nutrient gradients are found across these jets. In contrast, air-sea CO$_2$ exchange tends to equalize surface DIC concentrations and subdues their gradients. Our result agree with inferences from a previous study by Lee and Williams [2000], who used an idealized model to illustrate that down-gradient eddy diffusive transport should play the leading role over eddy advection for tracers with a short lifetime, such as PO$_4$, which is rapidly consumed by photosynthesis in the subtropics.

Lateral transport was shown to significantly contribute to the budget of all tracers in both subtropical gyres. Lateral PO$_4$ input is the dominant nutrient source to both subtropical gyres, amounting to 77% of the total nutrient supply in the North Pacific and 86% in the North Atlantic, with the remaining supply due to vertical mixing. Thus, our study agrees with previous work that underscored the importance of the lateral nutrient supply into the subtropical gyres (e.g. Letscher et al. [2016]), and shows this qualitative result is not sensi-
tive to the model used to estimate the lateral fluxes or the boundaries drawn around the gyres.

Our work also examined the spatial pattern of the nutrient exchange and the mechanisms by which it enters the gyre, which point toward a critical role for eddy exchange across the Kuroshio and Gulf Stream in supplying a large quantity of nutrients to the northern fringe of the gyres. In contrast to the supply of nutrients, DIC and heat are transported into the subtropical gyre almost exclusively by the mean circulation. Approximately one third of DIC supplied to the subtropical gyres is removed by the eddy lateral transport on an annual-mean basis, while the rest downwells. The air-sea exchange of carbon contributes little to the subtropical carbon budget in both basins in this preindustrial model simulation. On the other hand, the heat sink balancing the cross-boundary supply by the mean flow is approximately equally split between removal by eddies, ocean to atmosphere heat loss, and downwelling.

In summary, our results confirm the importance of lateral transport in the tracer budgets of the Northern Hemisphere subtropical gyres. It is notable that the majority of PO$_4$ that fuels new primary productivity is provided by lateral transport across the Kuroshio and Gulf Stream, with eddies being the principal supply mechanism. Another important result of this work is that cross-boundary transport provides the primary source of heat to the gyre on an annual basis; some of this heat is removed by eddies and downwelling, while roughly a third is lost to the overlying atmosphere. Given the first-order role of cross-boundary fluxes in the upper-ocean heat budget, we expect these transport processes to be critical to restratification of the subtropical mode waters, with the potential to influence the subduction of thermal anomalies that give memory to the climate system.

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