

Imperial College of Science, Technology and Medicine  
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**The Connected Ocean: comparing inter-ocean transport  
at the surface and at depth**

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# Abstract

The oceans form a global network of inter-connected basins, with currents transporting a range of tracers such as heat and pollution between them. Although the ocean dynamics play a key role in determining transport between basins, the basins are defined by arbitrary geographical borders rather than physical connections. This thesis proposes a method of detecting dynamical boundaries and applies it to the ocean surface and interior. To achieve this, probabilistic models are used to summarise the transport described by vast datasets of water-following trajectories. Such models, known as Markov Chain models, have previously been used to map the distribution of plastic pollution yet are known to induce an artificial dispersion. First, a sensitivity analysis of the effect of model parameters on artificial dispersion is performed, to determine the optimal model set-up. Next, a global dataset of observed trajectories is used to detect dynamic transport barriers in two key areas of inter-ocean surface exchange: the Agulhas Current system and the North Atlantic inter-gyre transport barrier. Connectivity maps are introduced as maps of tracer destinations which highlight dynamical segregation between regions. For example, these are used to identify the source region for Agulhas Leakage and one-way equator-ward transport across the Gulf Stream. Using a new method to extract geostrophic motion from a trajectory dataset, the geostrophic contribution to inter-basin transport can be identified. Finally, connectivity maps are produced for the ocean interior using virtual particles released along isopycnals in the eddy-permitting ORCA025 ocean circulation model. The change in connectivity, in the two study regions, between two target isopycnals is the first step in creating a 3D border of the ocean basins. New basin definitions are then explained by transport processes relevant to the region. Future work should establish the link between dynamical boundaries and the distribution of heat and pollution.



## Declaration

I hereby declare that the work presented in this PhD thesis is my own, and that all other scientific literature drawn upon is cited in the text, and referenced in the reference list.

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It seems to me that a PhD is never really finished. There are many avenues this project could continue down, as there are many applications of the field of physical oceanography. Even after nearly 4 years, it feels like a beginning. Maybe that is how it should be, whether one takes the academic route or not. A PhD requires emotional investment in the work and I know that this will not fade away.

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It was my Father who taught me the wonders of maths and physics. He gained a Masters in Astrophysics, which was unusual for a working-class lad back then, and then became a lecturer of engineering in a polytechnic. He taught thermodynamics and differential equation to adult students from poor and migrant backgrounds similar to his. I suppose that's where I learned that learning science is not just a passion, but also a means to directly improve people's lives. He would have hoped for a few more equations in this thesis.

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*Great are the works of the Lord, studied by all who delight in them.* - Psalm 111:2



# Contents

|                                                        |             |
|--------------------------------------------------------|-------------|
| <b>Abstract</b>                                        | <b>i</b>    |
| <b>Declaration</b>                                     | <b>iii</b>  |
| <b>Acknowledgements</b>                                | <b>v</b>    |
| <b>List of Tables</b>                                  | <b>xiii</b> |
| <b>List of Figures</b>                                 | <b>xv</b>   |
| <b>1 Introduction</b>                                  | <b>1</b>    |
| 1.1 Motivation . . . . .                               | 1           |
| 1.2 Planetary Fluid Dynamics . . . . .                 | 3           |
| 1.3 Global Circulation . . . . .                       | 8           |
| 1.4 Connectivity Analysis . . . . .                    | 11          |
| 1.5 Objectives and Outline . . . . .                   | 14          |
| <b>2 Ocean Connectivity Datasets &amp; Methodology</b> | <b>16</b>   |
| 2.1 Introduction . . . . .                             | 16          |
| 2.2 Observational Global Drifter Datasets . . . . .    | 17          |

|          |                                                          |           |
|----------|----------------------------------------------------------|-----------|
| 2.3      | Virtual Drifter Datasets . . . . .                       | 23        |
| 2.3.1    | Surface Transport from Globcurrent . . . . .             | 24        |
| 2.3.2    | Isopycnal Surfaces from NEMO . . . . .                   | 25        |
| 2.4      | Transition Matrices and Markov Chain Models . . . . .    | 27        |
| 2.5      | Summary . . . . .                                        | 33        |
| <b>3</b> | <b>Markov Chain Model Parameter Sensitivity Analysis</b> | <b>34</b> |
| 3.1      | Introduction . . . . .                                   | 34        |
| 3.2      | Idealised Stommel Gyre . . . . .                         | 35        |
| 3.3      | GDP Drifters and Plastic Accumulation Regions . . . . .  | 39        |
| 3.4      | Summary . . . . .                                        | 45        |
| <b>4</b> | <b>Surface Connectivity</b>                              | <b>46</b> |
| 4.1      | Introduction . . . . .                                   | 46        |
| 4.2      | ADT-following drifters . . . . .                         | 47        |
| 4.3      | Agulhas Current System . . . . .                         | 55        |
| 4.4      | North Atlantic Transport Barrier . . . . .               | 65        |
| 4.5      | Summary . . . . .                                        | 74        |
| <b>5</b> | <b>Isopycnal Connectivity</b>                            | <b>76</b> |
| 5.1      | Introduction . . . . .                                   | 76        |
| 5.2      | Isopycnal Surfaces: $\sigma_0=27.1$ & $27.6$ . . . . .   | 77        |
| 5.3      | Interior Velocity Fields . . . . .                       | 82        |
| 5.4      | Agulhas Current System . . . . .                         | 86        |

|          |                                               |            |
|----------|-----------------------------------------------|------------|
| 5.4.1    | Model surface comparison with GDP . . . . .   | 86         |
| 5.4.2    | Along-isopycnal connectivity . . . . .        | 91         |
| 5.5      | North Atlantic Transport Barrier . . . . .    | 98         |
| 5.5.1    | Model surface comparison with GDP . . . . .   | 98         |
| 5.5.2    | Along-isopycnal transport . . . . .           | 100        |
| 5.6      | Summary . . . . .                             | 105        |
| <b>6</b> | <b>Conclusions</b>                            | <b>107</b> |
| 6.1      | Markov Chain Modelling of the Ocean . . . . . | 107        |
| 6.2      | Surface Connectivity . . . . .                | 109        |
| 6.3      | Isopycnal Connectivity . . . . .              | 111        |
| 6.4      | Future Work . . . . .                         | 112        |
|          | <b>Bibliography</b>                           | <b>116</b> |



# List of Tables

4.1 Percentage of global GDP drifter sub-trajectories which obey the geostrophic  
condition defined by Equation 4.1. . . . . 55



# List of Figures

|     |                                                                                                                                                                                                                                                                                                                                                                   |    |
|-----|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|----|
| 1.1 | Trajectories of Global Drifter Program (GDP) satellite-tracked drifters around the Agulhas Current region between the Indian and South Atlantic oceans. Each drifter trajectory is individually and randomly coloured. . . . .                                                                                                                                    | 4  |
| 1.2 | Time-averaged Absolute Dynamic Topography (ADT) from daily $\frac{1}{4}^\circ$ fields over the period 1993-2017. ADT fields are provided by the Copernicus Marine Environment Monitoring Service (CMEMS). . . . .                                                                                                                                                 | 6  |
| 1.3 | Schematic of the upper ocean response to constant wind forcing, highlighting the change in velocity with depth. A typical depth for the Ekman layer is given. Extracted from the U.S National Oceanic and Atmospheric Administration (NOAA) Education site <a href="http://www.oceanservice.noaa.gov/education">www.oceanservice.noaa.gov/education</a> . . . . . | 7  |
| 1.4 | Schematic of global overturning circulation. Arrows which show simplified transport directions are coloured by density of water being transported (red - light, blue - dense). Transports between basins are shown in Sverdrups ( $10^6 m s^{-1}$ ). Extracted from (Lumpkin and Speer, 2007). . . . .                                                            | 10 |
| 1.5 | Potential density ( $\sigma_0$ ) chart for the upper 1000 m of the Atlantic Ocean. The section corresponds to the 3 A16 research cruises which made a roughly north-south path through the Atlantic between $30^\circ W$ and $20^\circ W$ . Extracted from WOCE Atlas <a href="http://www.woceatlas.ucsd.edu">www.woceatlas.ucsd.edu</a> . . . . .                | 11 |
| 1.6 | Surface basin boundaries derived from modeled trajectories. Map represents an amalgamation of eigenvectors of a transition matrix. Each shaded region corresponds to a dynamically-segregated region on a 60-day timescale. Extracted from Froyland et al. (2014). . . . .                                                                                        | 13 |

|     |                                                                                                                                                                                                                                                                                                                                                                                                                                                        |    |
|-----|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|----|
| 2.1 | Global Drifter Program coordinate data maps. (a) Coverage of coordinate readings per $1^\circ$ cell measured from 1979 to June 2017. White areas contain no coordinate readings (McAdam and van Sebille, 2018). (b) Coordinates of the drifter deployment positions from the same period (blue dots). . . . .                                                                                                                                          | 19 |
| 2.2 | Global Drifter Program coverage increase since 2012, shown as the difference in values between 2012 and March 2017 (McAdam and van Sebille, 2018) . . . . .                                                                                                                                                                                                                                                                                            | 19 |
| 2.3 | Global Drifter Program drogued and undrogued drifter coverage. As for Figure 2.1a, but for (a) drogued and (b) undrogued sub-sets. The coverage represents 45% and 55% of the total dataset respectively. . . . .                                                                                                                                                                                                                                      | 21 |
| 2.4 | GlobCurrent-derived virtual particle coverage maps. Particles are released at the initial time-step only and run for two years, over the 2014-2015 period. Particles were integrated in the daily $\frac{1}{4}^\circ$ GlobCurrent surface velocity fields with the OceanParcels toolkit (a) $\sim 20,000$ particles corresponding to a $1.5^\circ$ release (b) $\sim 200,000$ particles corresponding to a $\frac{1}{2}^\circ$ release. . . . .        | 26 |
| 2.5 | Markov Chain model of tracer release near the Panama Isthmus. This model uses transition matrices constructed from the entire GDP dataset, with a cell-size $dx = 1^\circ$ and time-step $dt = 5$ days. Tracer is released in a $2^\circ \times 2^\circ$ area centred around ( $7^\circ\text{N}$ , $79^\circ\text{W}$ ) and run for 360 days (McAdam and van Sebille, 2018). . . .                                                                     | 30 |
| 2.6 | Markov Chain model output and time-step dependence. (a) The tracer release from Figure 2.5 is repeated with transition matrices $T_{dt=20days}$ , $T_{dt=60days}$ and $T_{dt=180days}$ . Output from day 360 is shown (McAdam and van Sebille, 2018). . .                                                                                                                                                                                              | 32 |
| 3.1 | Stommel's Western Intensification Gyre. Trajectories of particles in a anti-cyclonic northern hemisphere gyre are advected forward in time using Equations 3.1 3.2. Individual trajectories are represented by different colours. Particles are initialised at 100 km spacing in both latitudinal and longitudinal directions and integrated for 5 years. 10% of the trajectories are shown for visual clarity (McAdam and van Sebille, 2018). . . . . | 36 |

|     |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                 |    |
|-----|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|----|
| 3.2 | Markov Chain model applied to the Stommel gyre. The Markov Chain model was initialised with tracer in one cell centred at $(0.5 \times 10^3 \text{ km}, 5 \times 10^3 \text{ km})$ . (a-d) Model output comparison after 7 years for time-steps $dt=5, 20, 60$ & 180 days, with cell size fixed at $dx=100 \text{ km}$ . (e-f) Model output comparison for cell sizes $dx=50, 100$ & $200 \text{ km}$ , with time-step fixed at $dt = 60$ days. The black streamline corresponds to the tracer release cell to indicate the analytical solution. The perpendicular and parallel components of dispersion from Equation 3.3 are shown in red and black text respectively (McAdam and van Sebille, 2018). . . . . | 38 |
| 3.3 | Surface accumulation in gyres. (a) Tracer distribution after 50-years after a global, homogeneous release, using $T_{dt=60days}$ . (b) Areas in black which contain tracer concentrations greater than twice the initial value are defined as “patches”. . . . .                                                                                                                                                                                                                                                                                                                                                                                                                                                | 41 |
| 3.4 | Sensitivity analysis of GDP-derived transition matrix parameters on gyre accumulation over 50 years. The evolution of patch sizes in each ocean basin is shown for the full range of $dt$ values and a fixed $dx = 1^\circ$ . . . . .                                                                                                                                                                                                                                                                                                                                                                                                                                                                           | 42 |
| 3.5 | Sensitivity analysis of transition matrix parameters on gyre accumulation over 50 years. The evolution of patches for a range of $dx$ values and a fixed $dt = 60$ days. . . . .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                | 43 |
| 3.6 | Sensitivity analysis of GlobCurrent-derived transition matrix parameters on gyre accumulation over 50 years. The evolution of patch sizes in each ocean basin for a range of $dt$ values and a fixed $dx = 1^\circ$ . As for Figure 3.4, but using the enhanced GlobCurrent-derived dataset using $\sim 200,000$ particles. . . . .                                                                                                                                                                                                                                                                                                                                                                             | 44 |
| 4.1 | Time-averaged Mean Sea Level Anomaly error from daily $\frac{1}{4}^\circ$ fields over the period 1993-2017. MSLA error fields are provided by the Copernicus Marine Environment Monitoring Service (CMEMS). . . . .                                                                                                                                                                                                                                                                                                                                                                                                                                                                                             | 49 |
| 4.2 | Absolute Dynamic Topography experienced by drifters in the Agulhas System. Examples of individual drifter pathways: an unstable westward-shifted Retroflection (top), a stable central Retroflection (middle) and an Agulhas Ring (bottom). ADT values are interpolated to 6-hourly readings, to correspond to the drifter coordinate readings. . . . .                                                                                                                                                                                                                                                                                                                                                         | 50 |

|     |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                         |    |
|-----|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|----|
| 4.3 | ADT connectivity matrices depicting the change in ADT at beginning and end of all GDP drifter sub-trajectories. Drifter sub-trajectories from 1993 onwards are used to correspond with the available ADT fields. The total number of coordinate data points available for each sub-trajectory length are shown in Table 4.1. . . .                                                                                                                                                                                      | 51 |
| 4.4 | Global distribution of geostrophic drifters. A map of proportion of sub-trajectories, beginning in each $1^\circ$ grid cell, whose ADT values stay within 0.2 m over a $dt$ -day time scale. White areas contain no ADT readings which correspond to the initial or final sub-trajectories readings. . . . .                                                                                                                                                                                                            | 53 |
| 4.5 | Alignment of nongeostrophic and geostrophic velocity fields. (left) Stokes drift and (right) 15 m Ekman velocity vectors are plotted over AVISO absolute dynamic topography. The Stokes drift $\frac{1}{2}^\circ$ field is extracted from European Centre for MediumRange Weather Forecasting (ECMWF) (Dee et al., 2011). The Ekman transport $\frac{1}{4}^\circ$ field is extracted from the GlobCurrent data set. Arrows are plotted every $1.5^\circ$ for clarity. All fields are annual averages from 2014. . . . . | 54 |
| 4.6 | As for Figure 4.3, but for drogued and undrogued subsets of GDP data. . . . .                                                                                                                                                                                                                                                                                                                                                                                                                                           | 55 |
| 4.7 | Distribution of tracer in the GDP-derived Markov Chain model of the Agulhas Current system. Flow exits the region via the Agulhas Retroflection Current (east of $40^\circ\text{E}$ ) or Agulhas Leakage (west of the Good Hope line, other black line). Tracer is released from the red square at $[31^\circ\text{E}, 33^\circ\text{S}]$ and transported by the full-field (left column) and the geostrophic (right) versions of $T_{dt=60days}$ . . . .                                                               | 57 |
| 4.8 | Proportion of tracer transport in the Agulhas Current system which passes through Agulhas leakage (bold) and Return Current (dashed). The boundaries are shown in Figure 4.7. For a range of time-steps, (a) full-field transport and (b) geostrophic Markov Chain model transport. Van Sebille et al. (2009) found 25% of GDP drifters leave via the Agulhas leakage. . . . .                                                                                                                                          | 58 |
| 4.9 | Markov Chain models derived from two GlobCurrent-based particle datasets which differ in the number of drifters deployed. Top: a GDP-sized dataset of 22,560 particles deployed on a $1.5^\circ$ grid. Bottom: A larger dataset of 201,040 particles deployed on a $\frac{1}{2}^\circ$ grid. Both datasets are advected on the $\frac{1}{4}^\circ$ geostrophic and Ekman surface velocity fields from 2012 to 2013 inclusive. . . . .                                                                                   | 58 |

|      |                                                                                                                                                                                                                                                                                                                                                                                                                                                                       |    |
|------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|----|
| 4.10 | Agulhas Leakage connectivity map comparison for full-field and geostrophic flow. Tracer which crosses the Good Hope line (black) is defined as Leakage. A $T_{dt=60days}$ transition matrix was used for both. . . . .                                                                                                                                                                                                                                                | 62 |
| 4.11 | Agulhas Leakage 5-year connectivity maps for a range of GDP-based Markov Chain models. Tracer which crosses the Good Hope line (black) is defined as Leakage. . . . .                                                                                                                                                                                                                                                                                                 | 63 |
| 4.12 | Number of $dt$ -length GDP drifter sub-trajectories entering $1^\circ$ cells. . . . .                                                                                                                                                                                                                                                                                                                                                                                 | 64 |
| 4.13 | Cross-frontal leakage between sub-tropical and sub-polar gyres in the North Atlantic. Left: Full-field GDP-based Markov Chain model. Middle: Geostrophic GDP-based model. Right: The full-field transport minus geostrophic transport. For each cell, the quantity which remains south of the front, or crosses the front if beginning in the sub-polar gyre, is counted. The front is defined by Rypina et al. (2011) from a drifter-derived velocity field. . . . . | 66 |
| 4.14 | Cross-frontal leakage between sub-tropical and sub-polar gyres in the North Atlantic for full flow field and geostrophic flow. For each cell, the quantity which remains (or enters, if beginning north of the front) in the sub-tropical gyre is counted. The front (transport barrier) is the $45N$ zonal line. . . . .                                                                                                                                             | 70 |
| 4.15 | Eigenvector analysis of the GDP-based $T_{dt=60days}$ . The values within the eigenvectors are relative to each other. See text for brief description on eigenvector analysis and Froyland et al. (2014) for a detailed explanation. . . . .                                                                                                                                                                                                                          | 71 |
| 4.16 | Cross-frontal leakage, after 5 years, between sub-tropical and sub-polar gyres in the North Atlantic for full-field and geostrophic drifters flows, using transition matrices $T_{dt=5days}$ , $T_{dt=20days}$ & $T_{dt=180days}$ . . . . .                                                                                                                                                                                                                           | 73 |
| 5.1  | Temperature and salinity charts for the $\sigma_0=27.1$ and $27.6$ isopycnals. Potential temperature and absolute salinity (g/kg) are provided by ORCA025 output files. Here is shown the annual average for the 2001 output. The model output is plotted in its original tripolar grid form Section 2.3.2. for details . . . . .                                                                                                                                     | 79 |

- 5.2 Mixed layer and isopycnal depths in ORCA025. Left column: 5-day average time-step fields when the greatest amount of isopycnal is shallower than the mixed layer depth in the northern hemisphere (5th April 2001). Right column: as before but for minimum amount (13th August 2001). (a) & (b) Mixed layer depths, (c) & (d)  $\sigma_0=27.1$  isopycnal depth, (e)& (f) area where 27.1 isopycnal (I) is shallower than the mixed layer depth (M), (g) & (h)  $\sigma_0=27.6$  isopycnal depth, (i)& (j) area where 27.6 isopycnal is shallower than the mixed layer. The black contour in the isopycnal depths is the "outcrop" line where the isopycnal intersects the mixed layer depth. The overlap area highlights where the mixed layer is deeper than the isopycnal depth (M<I). The model output is plotted in its original tripolar grid form (Section 2.3.2). . . . . 81
- 5.3 Velocity fields in the Agulhas Current System in ORCA025. (a) Annual average speed at the surface. Black lines indicate the 25°S and 32°S parallels and the 25°E and 40°E meridians. (b) Cross-section of the zonal component of velocity through the 25°E meridian, where positive values indicate eastward flow. (c) Cross-section of the meridional component of velocity through the 25°S parallel, where positive values indicate northward flow. (d) As in (b) but for the 40°E meridian (e) As in (c) but for the 32°S parallel. Green lines indicates the depth of the  $\sigma_0=27.1$  and 27.6 isopycnals. All fields are for the ORCA025 2001 annual average. . . . . 83
- 5.4 Velocity fields in the North Atlantic in ORCA025. (a) Annual average speed at the surface. Black lines indicate the 42°N parallel and the 55°W meridian. (b) Cross-section of the zonal component of velocity through the 55°W meridian, where positive values indicate eastward flow. (c) Cross-section of the meridian component of velocity through the 42°N parallel, where positive values indicate northward flow. Green lines indicates the depth of the  $\sigma=27.1$  and 27.6 isopycnals. All fields are for the ORCA025 2001 annual average. . . . . 84

- 5.5 Schematic explaining the difference between horizontal and along-slope velocity components. (a) A horizontal surface with unit vectors  $\mathbf{i}$ ,  $\mathbf{j}$  and  $\mathbf{k}$  in the meridional, zonal and vertical directions respectively. (b) A sloping surface showing the along-isopycnal components,  $\mathbf{i} \perp n$  and  $\mathbf{j} \perp n$ , and the normal to the surface,  $\mathbf{n}$ . . . . . 85
- 5.6 Along-isopycnal velocity fields and isopycnal sloping. Top: Magnitude of horizontal velocity components interpolated to the corresponding isopycnal depth. Bottom: Deviation of the dot product of  $\mathbf{n}$  and  $\mathbf{k}$  from unity, where higher values indicate steeper sloping. Time-step is the first from ORCA025 2001. . . . . 87
- 5.7 Agulhas Leakage connectivity map for the GDP drifter and ORCA025 surface trajectories. Tracer which crosses the Good Hope line (black) is defined as Leakage. A  $T_{dt=60days}$  matrix was used for both and the models were run for 5 years. 89
- 5.8 Agulhas Leakage connectivity map for a range of transition matrices built from ORCA025 surface trajectories. Tracer which crosses the Good Hope line (black) is defined as Leakage. . . . . 90
- 5.9 Agulhas Leakage connectivity map for ORCA025  $\sigma_0 = 27.1$  and  $\sigma_0 = 27.6$  isopycnal trajectories. The depths of the isopycnal regions in these region are, respectively, roughly 1000 m and 1800 m. Tracer which crosses the Good Hope line (black) is defined as Leakage. A  $T_{dt=60days}$  matrix was used for both and the models were run for 25 years. . . . . 94
- 5.10 Time-series of Agulhas Leakage fluxes for the surface, 27.1 and 27.1 isopycnals. Tracer is released from  $[31^\circ\text{E}, 33^\circ\text{S}]$  in  $T_{dt=60days}$  Markov Chain models for each surface. The boundaries for the Agulhas Leakage and Return Current are the Good Hope line and  $40^\circ\text{E}$  respectively (in Figure 4.8). . . . . 95
- 5.11 Agulhas Leakage connectivity map for a range of transition matrices built from ORCA025  $\sigma_0=27.1$  isopycnal trajectories. Tracer which crosses the Good Hope line (black) is defined as Leakage. . . . . 96

|      |                                                                                                                                                                                                                                                                                                                                                                                                         |     |
|------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|-----|
| 5.12 | Agulhas Leakage connectivity map for a range of transition matrices built from ORCA025 $\sigma_0=27.6$ isopycnal trajectories. Tracer which crosses the Good Hope line (black) is defined as Leakage. . . . .                                                                                                                                                                                           | 97  |
| 5.13 | Cross-frontal connectivity map for the GDP drifter and ORCA025 surface trajectories. The map highlights tracer which finishes south of the (Rypina et al., 2016) front (black). A $T_{dt=60days}$ matrix was used for both and the models were run for 25 years. . . . .                                                                                                                                | 99  |
| 5.14 | ORCA025 surface structure compared to GDP-derived front between the North Atlantic sub-tropical and sub-polar gyres. Top: Surface speed. Bottom: Surface ADT. The front derived by Rypina et al. (2011) is super-imposed on both maps. Both maps use the annual average of 2001 for their respective fields. . . . .                                                                                    | 100 |
| 5.15 | Cross-frontal connectivity map for ORCA025 $\sigma_0 = 27.1$ and $\sigma_0 = 27.6$ isopycnal trajectories. The map highlights tracer which finishes south of the (Rypina et al., 2016) front (black). A $T_{dt=60days}$ matrix was used for both and the models were run for 25 years. The blank cells indicate where the isopycnal intersects the mixed layer . . . . .                                | 103 |
| 5.16 | Fate of tracer in the North Atlantic region for ORCA025 $\sigma_0 = 27.1$ and $\sigma = 27.6$ isopycnal trajectories after 25 years. The map highlights tracer which finishes south and north of the (Rypina et al., 2016) front (black). The bottom plots show the tracer consumed by the mixed layer. A $T_{dt=60days}$ matrix was used for both and all maps are for 25 years after release. . . . . | 104 |

# Chapter 1

## Introduction

### 1.1 Motivation

Knowledge of ocean current transport between distant regions is crucial for predicting the distribution and fate of a wide range of tracers, both natural and anthropogenic. Marine wildlife migration often makes use of the most direct and rapid currents (White et al., 2010). Dispersion of oil spills from offshore wells must be tracked to assess ecological and economic damage (Liu et al., 2011). The most abundant marine pollutant, plastic litter, is found throughout the ocean, not just because of its widespread use but because currents have a global reach (Kershaw and Rochman, 2015).

The destination of marine wildlife and pollutants are both, to varying degrees, determined by the ocean currents but it is difficult to detect coherent patterns of movement. For example, the surface accumulation of plastic in the North Pacific is often referred to as a garbage “patch” yet there is no abrupt boundary and the definition of such a feature is based on arbitrary concentration thresholds (Maximenko et al., 2012; Van Sebille et al., 2012). Meanwhile, the largest oceanographic structures, the ocean basins, are bounded by geographical borders defined by the International Hydrographic Organisation, which have little dynamical significance (IHO, 1953).

Besides surface phenomena, transport of important tracers also occurs in the ocean interior. Ocean currents in the northern hemisphere transport roughly a quarter of the heat from solar radiation received at the tropics towards higher latitudes above depths of 1000m (Trenberth and Caron, 2001). Below this, intermediate and deep waters also store and transport heat absorbed from the atmosphere, affecting the rate of anthropogenic climate change (Collins et al., 2013). Carbon dioxide uptake from the atmosphere is also transported through the deep interior for thousands of years (Khatiwala et al., 2012).

The distribution of temperature, salinity and tracer concentration gives some indication of dynamically varying regions within the ocean interior. Typically, the interior is split into water masses which are marked by regionally-distinct temperature and salinity properties. There is, however, mixing between these water masses which blurs their geographical boundaries (Talley, 2013). Instead, it is now common to depict global circulation in a thermohaline coordinate system, as a complement to the geographical coordinate system. This has provided an alternative method of identifying different water masses and estimating fluxes between them (Döös et al., 2012; Zika et al., 2012). Due to the decoupling of water masses from geographical coordinates, water masses in thermohaline coordinates may be spatially discontinuous. Regardless of the coordinate system used, there is nonetheless a lack of a dynamical definition of features within the interior of the ocean.

In recent years, the search for physically-meaningful definitions of surface boundaries has advanced simultaneously with the assessment of marine plastic pollution. Novel definitions are based on the dynamics of floating objects and correspond to regions in the ocean where accumulation is most likely. As a result, they have greatly contributed to explaining the distribution and fate of floating debris (Maximenko et al., 2012; Froyland et al., 2014; Van Sebille et al., 2015). It is therefore reasonable to suggest that dynamically-defined boundaries for the ocean interior would correspond to the distribution of interior tracers such as heat. Understanding the connections between basins and the processes that drive the boundaries is the main aim of this thesis and is an important task which may provide a dynamical context to the distribution of tracers throughout the global ocean.

## 1.2 Planetary Fluid Dynamics

This section focuses on the mathematical tools used to create simplified models of ocean flow. In reality, pathways of water-following tracers are convoluted even on regional scales. The nature of these pathways is uncovered by the trajectories of satellite-tracked drifters. In Figure 1.1, these trajectories show the difference, for example, between the laminar Agulhas Current, running along the eastern coast of South Africa, and the turbulent region to the west in the South Atlantic basin. The former is the fastest western boundary current, reaching surface speeds in its core of up to  $1.8 \text{ m s}^{-1}$  (Beal and Bryden, 1999). In the latter, mesoscale ( $O(10\text{--}100 \text{ km})$ ) eddies give rise to large curvatures and looped trajectories. In the western region of the domain, smaller, sub-mesoscale features ( $O(1 \text{ km})$ ) are present. Dispersion on the sub-mesoscale grows exponentially with time (LaCasce, 2008). As a result, even two objects released a few meters apart can be separated by over 100 km after less than 2 months (van Sebille et al., 2015).

Despite the chaotic nature of circulation, the models discussed here explain many key features by making assumptions about which forces are dominant. The horizontal forces acting on the ocean include horizontal pressure gradients, Coriolis effect and, particularly near boundaries, frictional stresses. Meanwhile, the forces acting in the vertical are gravity and vertical pressure gradients. Acceleration arises from an imbalance of forces. By assuming certain forces are negligible on certain temporal- or spatial-scales, key balances can be identified and expressed in mathematical terms.

A key force balance on the basin scale ( $O(1000 \text{ km})$ ) is the geostrophic balance. It is assumed that frictional boundary forces act over a relatively narrow area near the coast. Throughout the remainder of the ocean interior, the remaining horizontal forces are the pressure gradient force and the Coriolis force; wind stresses will receive a special mention later. The geostrophic balance in the horizontal direction is therefore defined as:

$$\begin{aligned} f u_{gs} &= -\frac{1}{\rho} \frac{\partial P}{\partial y}, \\ f v_{gs} &= \frac{1}{\rho} \frac{\partial P}{\partial x}, \end{aligned} \tag{1.1}$$

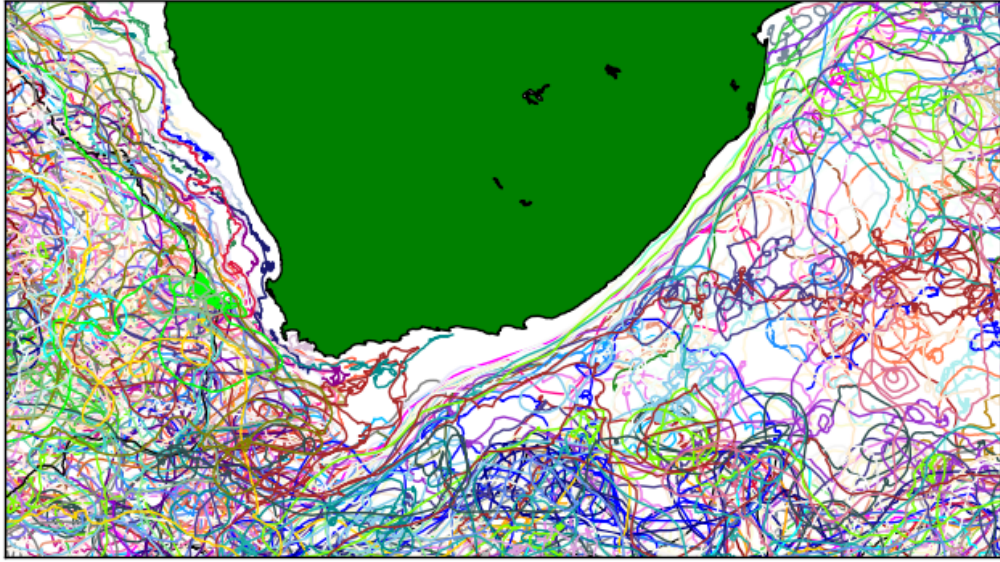


Figure 1.1: Trajectories of Global Drifter Program (GDP) satellite-tracked drifters around the Agulhas Current region between the Indian and South Atlantic oceans. Each drifter trajectory is individually and randomly coloured.

where  $f$  is the Coriolis frequency,  $u_{gs}$  and  $v_{gs}$  are zonal and meridional surface geostrophic velocity components respectively,  $\rho$  is the density and  $P$  the pressure. The Coriolis frequency quantifies the effect of the Earth's rotation. It is proportional to the Earth's angular velocity and the sine of latitude. It therefore increases towards the poles. Thus, closer to the Equator, motion caused by pressure differences for example would be deflected more strongly than nearer the poles.

Equation 1.1 states that geostrophic flow moves parallel to pressure contours. The velocity magnitude also depends on the Coriolis force, which varies with latitude. Pressure differences arise from the weight of the water column relative to the geoid. The height of the water column relative to the geoid, known as the dynamical height, is represented by  $h$ . Pressure at the surface can therefore be stated by the hydrostatic balance:

$$P = \rho_0 g h. \quad (1.2)$$

Substituting Equation 1.2 into Equation 1.1 returns the surface geostrophic velocities with respect to the dynamical topography of the ocean:

$$\begin{aligned} u_{gs} &= -\frac{g}{f} \frac{\partial h}{\partial y}, \\ v_{gs} &= \frac{g}{f} \frac{\partial h}{\partial x}. \end{aligned} \tag{1.3}$$

Geostrophic velocities are exactly parallel to contours of absolute dynamic topography (ADT, also known as sea surface height) and are proportional to the ADT gradient. The Coriolis frequency,  $f$ , varies with latitude and determines the magnitude of the velocities. Figure 1.2 shows the average ADT map for the global ocean. Sharp gradients in ADT correspond to rapid currents such as the Gulf Stream, Agulhas Current and the Antarctic Circumpolar Current. While a steady-state flow is not an accurate representation of the dynamic surface ocean, applying Equation 1.3 to satellite altimetry-derived ADT fields captures boundary current extensions, meanders and mesoscale eddies (Maximenko et al., 2009). Moreover, current velocities calculated from satellite altimetry using Equation 1.3 are accurate to less than  $5 \text{ cm s}^{-1}$ , indicating that the ocean is largely in geostrophic balance (Strub et al., 1997).

The Rossby number is a dimensionless number used to identify which spatial and temporal scales the geostrophic balance can be applied to. Specifically, it measures the relative importance of the Coriolis and inertial accelerations:

$$Ro = \frac{U}{Lf}, \tag{1.4}$$

where  $U$  is the speed and  $L$  the horizontal scale of the flow. A small Rossby number indicates that the effects of planetary rotation are large and that the geostrophic approximation is suitable to describe the flow. For example, the mesoscale features captured by drifters in Figure 1.1, the Agulhas Rings, have a rotational speed of approximately  $0.1 \text{ m s}^{-1}$  and are 50 km wide. The resulting Rossby number is  $O(0.01)$ . Sub-mesoscale features are by definition smaller and the corresponding Rossby number would be  $O(1)$  (Thomas et al., 2008).

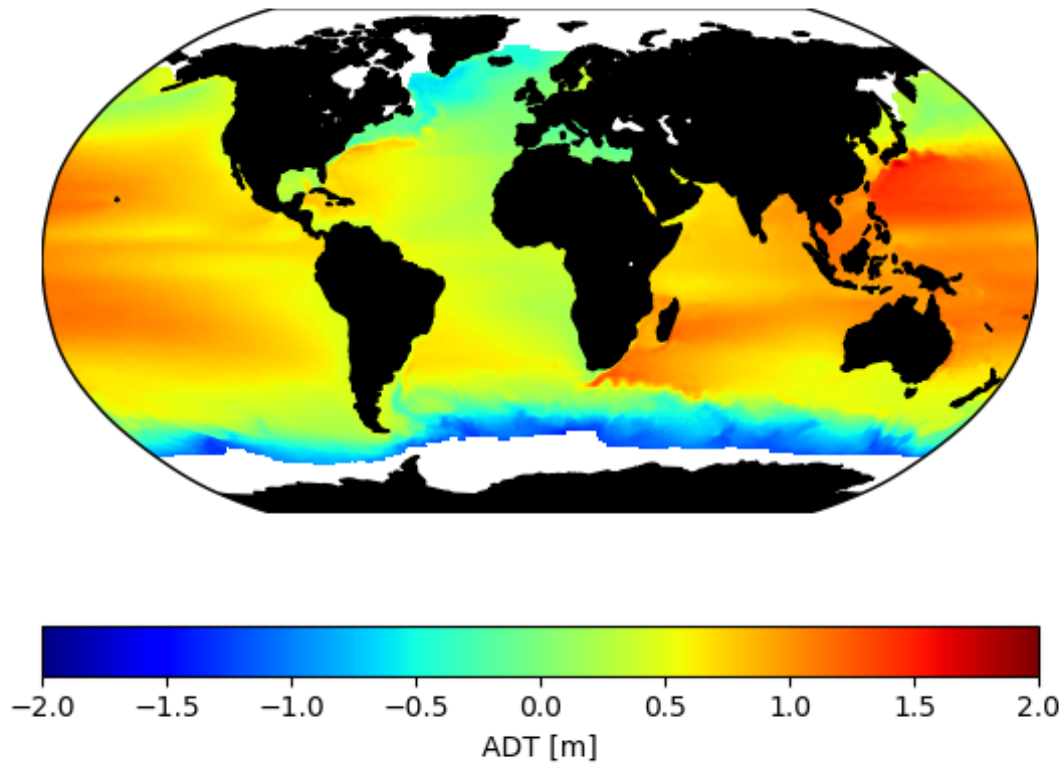


Figure 1.2: Time-averaged Absolute Dynamic Topography (ADT) from daily  $\frac{1}{4}^\circ$  fields over the period 1993-2017. ADT fields are provided by the Copernicus Marine Environment Monitoring Service (CMEMS).

Another key aspect of surface circulation is the response to winds. At the ocean surface, there is a vertical transfer of horizontal momentum provided by the wind. Ekman was the first to derive equations which explained why the resulting surface motion appeared to take place at an angle to the wind direction (Ekman, 1905). In his proof, Ekman assumed the two forces acting were the Coriolis force and the wind shear. Figure 1.3 shows the vertical velocity structure in the near-surface layer affected by wind, which manifests as a decaying spiral motion. The surface velocity is at  $45^\circ$  to the wind while at depth the current direction bends away from the wind and its magnitude decays. The depth of the layer affected by Ekman transport is arbitrary, because the velocity decays exponentially. (Ekman, 1905) defined the depth as where the current velocity is  $180^\circ$  to the original direction. Ekman's theories explain regional patterns of upwelling and downwelling motions in gyres centre, as well as along boundary currents such as the California and Benguela Currents, where a net transport perpendicular to a coastline must be balanced by vertical transport to or from depth. Spiral structures are rarely observed

in the open ocean, however, because near-surface velocity readings are dominated by turbulent mixing and surface waves. In instances where unidirectional winds persistently flow over a body of water, a spiral structure is observed in monthly averages (Chereskin, 1995).

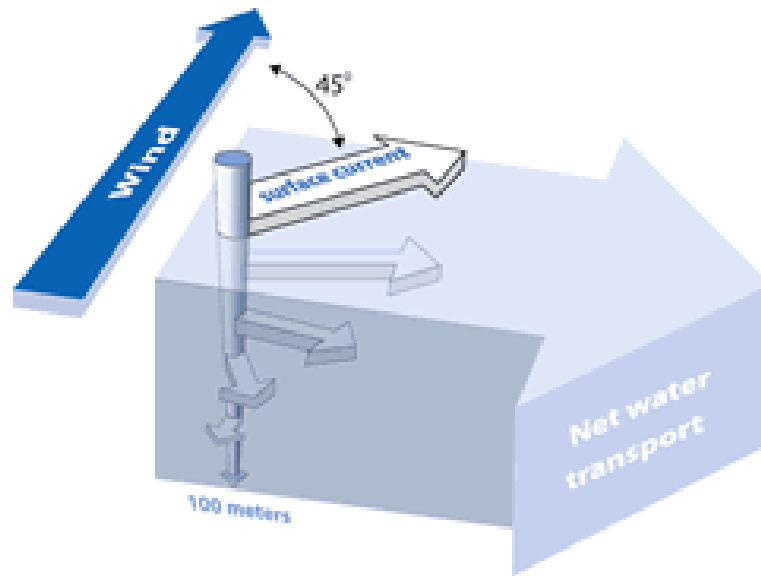


Figure 1.3: Schematic of the upper ocean response to constant wind forcing, highlighting the change in velocity with depth. A typical depth for the Ekman layer is given. Extracted from the U.S National Oceanic and Atmospheric Administration (NOAA) Education site [www.oceanservice.noaa.gov/education](http://www.oceanservice.noaa.gov/education).

Geostrophic balance and Ekman transport are combined to derive the theory of wind-driven surface circulation (Stommel, 1948). A main feature of surface circulation is the ocean gyres which are driven largely by regional wind stress patterns. Stommel (1948) first proved that bottom friction and the variation of Coriolis force with latitude causes a western intensification of flow within the gyres, forming the boundary currents. In the parts of the sub-tropics where winds are mostly zonal, the pressure contours indicate geostrophic flow is also zonal. Ekman transport is directed perpendicular to the current flow, towards the gyre centre. Water therefore converges in the inner gyre regions and downwells. Here, floating objects remain buoyant and instead accumulate. Velocity fields for floating objects are therefore divergent. Meanwhile, surface Ekman flow is directed pole-ward on either side of the Equator, pushing drifting objects away towards the sub-tropics. As a result, the combined motions of Ekman and geostrophic velocity fields can even be used to describe the distribution of floating debris.

Drifting objects also experience accelerations resulting from being only partially submerged in water. First, direct exposure to the wind causes “wind-slip”, which increases the down-wind component of velocity (Poulain et al., 2009). This means that the long-term, basin-scale accumulation of floating plastic debris is sped up compared to debris in suspension in the mixed layer (Beron-Vera et al., 2016). Another phenomenon is Stokes drift, which is a net drift in the direction of wave propagation caused by particles spending more time near the top of the wave where the motion is faster (Van Den Bremer and Breivik, 2017). Again, far from being a minor contribution to basin-scale transport, Stokes drift in storm-rich regions can open up new pathways for transport across ocean basins (Fraser et al., 2018).

### 1.3 Global Circulation

Wind-driven circulation is the surface contribution to a larger, three-dimensional global circulation throughout the ocean interior (Figure 1.4). Winds and waves directly influence the upper ocean in the mixed layer, which is typically no deeper than 200 m. The mixed layer gets its name from the turbulent mixing which occurs within it to homogenise properties such as temperature and salinity. Mixed layer depth depends on a range of factors including incoming radiation, freshwater input and wind speed. The seasonal and geographic variations will be highlighted later, in Chapter 5. Below this, the temperature decays from its mixed layer value to a near-uniform value, in the vertical, in the interior ocean. This transition layer is known as the thermocline. It is here where mixed layer waters are subducted into the interior ocean below (Williams et al., 1995).

Flow below the mixed layer is shielded from frictional forces and is instead approximately in geostrophic balance, with the exceptions being areas near the continental shelves and mid-ocean ridges (Stommel and Arons, 1960). By taking the vertical derivative of the geostrophic balance in Equation 1.3 and substituting pressure for density using the hydrostatic balance (Equation

1.2), the following balance for the interior ocean is found:

$$\begin{aligned} -f \frac{\partial v}{\partial z} &= g \frac{\partial \rho}{\partial x}, \\ f \frac{\partial u}{\partial z} &= g \frac{\partial \rho}{\partial y}. \end{aligned} \tag{1.5}$$

Known as the thermal wind equations, they relate horizontal density gradients to vertical velocity gradients. Density in the atmosphere is a function of temperature, hence the name “thermal” wind. In the ocean, however, density is a function of salinity also. Horizontal gradients in these scalar properties therefore determine the vertical structure of velocity. The velocity is measured relative to a reference depth, typically the depth of no motion. As an example, where density increases eastward, this leads to a positive  $v$  - a northward velocity component. Steeper isopycnals indicate more rapid geostrophic velocities. This is the case for the Gulf Stream and western boundary currents in general, where isopycnals steepen towards the velocity maxima.

Density surfaces therefore play a key role in determining the distribution of nutrients and tracers in the ocean. First, lateral velocities are typically several orders of magnitude larger than vertical velocities. Second, most of the interior is also largely shielded from diabatic processes such as radiative heating, which act to alter the density of water. As a result, the motion of a water parcel should experience negligible change in buoyancy below the mixed layer (Montgomery, 1938). Ideally, therefore, water masses spread along surfaces of constant density. Specifically, a potential density is used instead of in-situ density. Potential values correct for the change in compressibility caused by the pressure of the water column above, relative to a reference pressure. For example, Figure 1.5 shows that surfaces of constant potential density - isopycnals - relative to the surface pressure ( $\sigma_0$ ). They are approximately horizontal in the sub-tropical and equatorial regions but slope towards higher latitudes, reflecting horizontal gradients in temperature and salinity.

Changes in buoyancy drive the exchange of heat and salinity between the mixed layer and deeper waters. These changes are caused by cooling, freshwater-input, sea-ice formation and wind-driven mixing (Wunsch, 2002). Once in the interior, water masses largely retain their

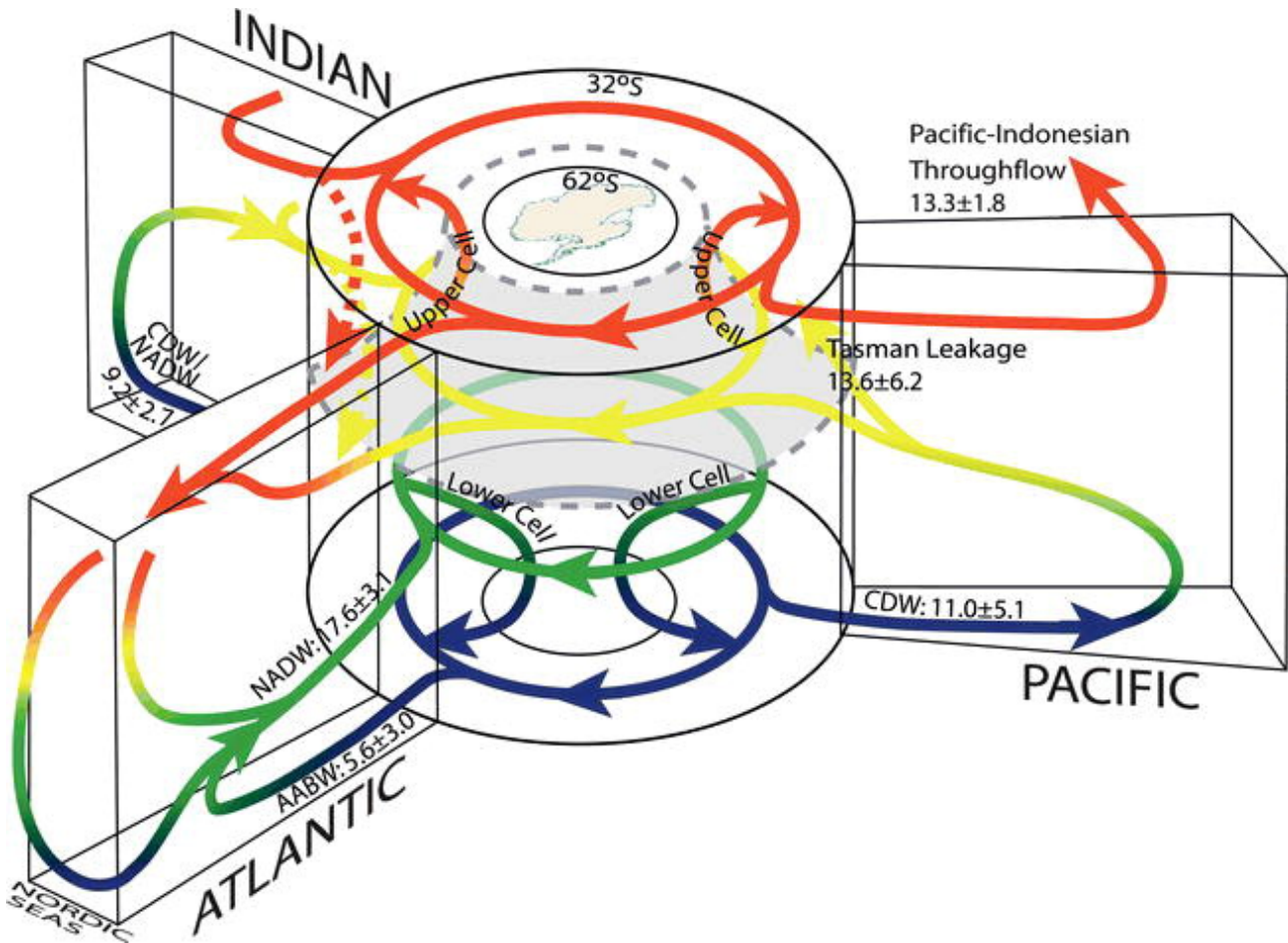


Figure 1.4: Schematic of global overturning circulation. Arrows which show simplified transport directions are coloured by density of water being transported (red - light, blue - dense). Transports between basins are shown in Sverdrups ( $10^6 \text{ m s}^{-1}$ ). Extracted from (Lumpkin and Speer, 2007).

temperature-salinity properties. The general picture of where water rises and sinks globally was consolidated by Gordon (1986). Dense water is formed in cold, salty regions such as the North Atlantic, while warmer water rises, or is mechanically lifted by wind and mixing, in the Southern Ocean. The sinking and spreading of dense water is known as the thermohaline circulation or global meridional overturning circulation. Since the 1980s, the Southern Ocean has also been found to be a crucial region for cooling, the formation of water masses and strong vertical mixing (Talley, 2013). The role of Agulhas Leakage in returning upper ocean water into the Atlantic, known as the “warm” route of global circulation, matches the “cold route” of the Drake Passage (Drijfhout et al., 1996; Beal et al., 2011). The current understanding of this circulation is summarised in Figure 1.4.

For the thermocline to be maintained, the downward mixing of cold water must be matched by a equivalent rate of upwelling (Stommel and Arons, 1960). The requirement for widespread upwelling to close the thermohaline circulation is still contentious to this day yet mechanisms are being tested and quantified. Mixing within the thermocline is enhanced in the Southern Ocean and regions of Ekman upwelling (Liang et al., 2017). Diapycnal mixing is also enhanced in regions with rough and steep bathymetry (Mackay et al., 2018). Frictional stresses at the edges of the ocean basins also enhance vertical mixing in the interior and contribute to upwelling (de Lavergne et al., 2017). The mechanisms for diapycnal motion complements the simplified picture of global circulation presented in Figure 1.4.

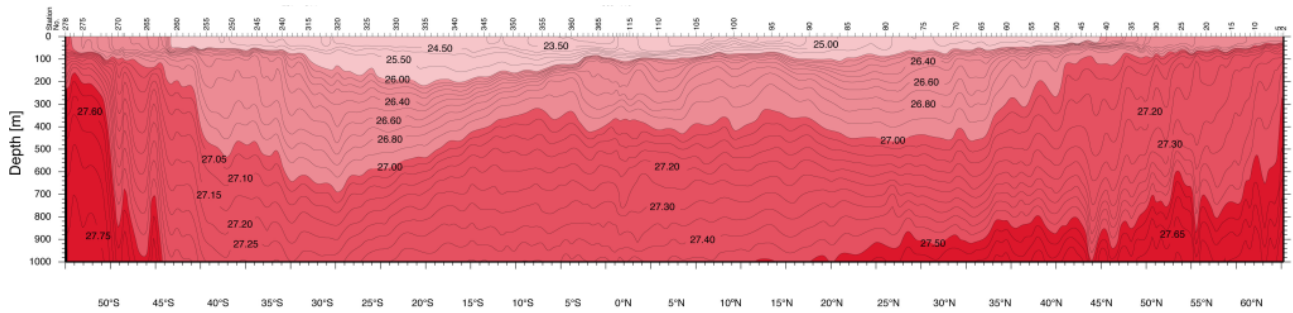


Figure 1.5: Potential density ( $\sigma_0$ ) chart for the upper 1000 m of the Atlantic Ocean. The section corresponds to the 3 A16 research cruises which made a roughly north-south path through the Atlantic between 30°W and 20°W. Extracted from WOCE Atlas [www.woceatlas.ucsd.edu](http://www.woceatlas.ucsd.edu)

## 1.4 Connectivity Analysis

There are two different theoretical and observational ways to study transport. First, Eulerian analysis determines the flow field at specific locations and times. It has a fixed and pre-defined reference frame. Examples include fixed moorings of current meters and satellite-derived fields. Meanwhile, Lagrangian analysis is performed in the reference frame of a moving parcel of fluid. The major example, used throughout this thesis, is the satellite-tracked drifter.

Due to practical and economic limitations, Lagrangian studies were initially limited to regional scales and drifters were sparsely released (e.g. Poulain and Niiler, 1989). Meanwhile, work on inter-ocean exchanges was done by comparing the temperature and salinity properties of water masses, to infer which waters crossed over into other basins (Gordon, 1986). Since then,

two advancements in Lagrangian analysis have vastly improved knowledge of the directions and quantities of transport between oceans: a global dataset of satellite-tracked drifters and Lagrangian analysis of ocean models.

Datasets of observed trajectories now exist for the surface - the Global Drifter Program (GDP) - and the interior - the Argo program (Lumpkin and Pazos, 2007; Ollitrault and Rannou, 2013). These capture the timescales and pathways of connections between parts of the ocean. They may also detect sub-mesoscale dynamics which are not resolved by current circulation models. Meanwhile, Lagrangian analysis of such models is performed by integrating the position of a virtual particle in a modelled velocity field (van Sebille et al., 2018). A key advantage over observational trajectories is that they are used to study pathways of three-dimensional overturning circulation.

New techniques which detect connections from increasingly larger datasets of convoluted pathways have gained traction in recent years. These techniques have been inspired by graph theory and treat the ocean as a network of edges and nodes. The “nodes” are typically rectangular cells (e.g.  $1^\circ$  by  $1^\circ$  areas), while the “edges” are trajectories, either observed or modelled. From this starting point, various studies have approached the concept of connectivity in different ways. For example, (Jönsson and Watson, 2016) solved the shortest-path problem for the ocean, using Dijkstras algorithm to find the path of least “weight”, in this case the time, through the ocean network.

Meanwhile, Maximenko et al. (2012) and Van Sebille et al. (2012) tabulated the entire GDP dataset into a transition matrix and used it to predict motion of tracer released in particular cells. Rows in a transition matrix correspond to probability distribution functions for transport from each cell in the ocean network, for a given time period. By initiating cells with a value corresponding to a tracer concentration, the transport of tracer can be projected into the future. This technique is a form of Markov Chain modelling and is used to track widespread plastic pollution (Van Sebille, 2014).

Froyland et al. (2014) used transition matrices to construct dynamical boundaries of surface circulation. Transition matrices represent transformations applied to the tracer in an ocean

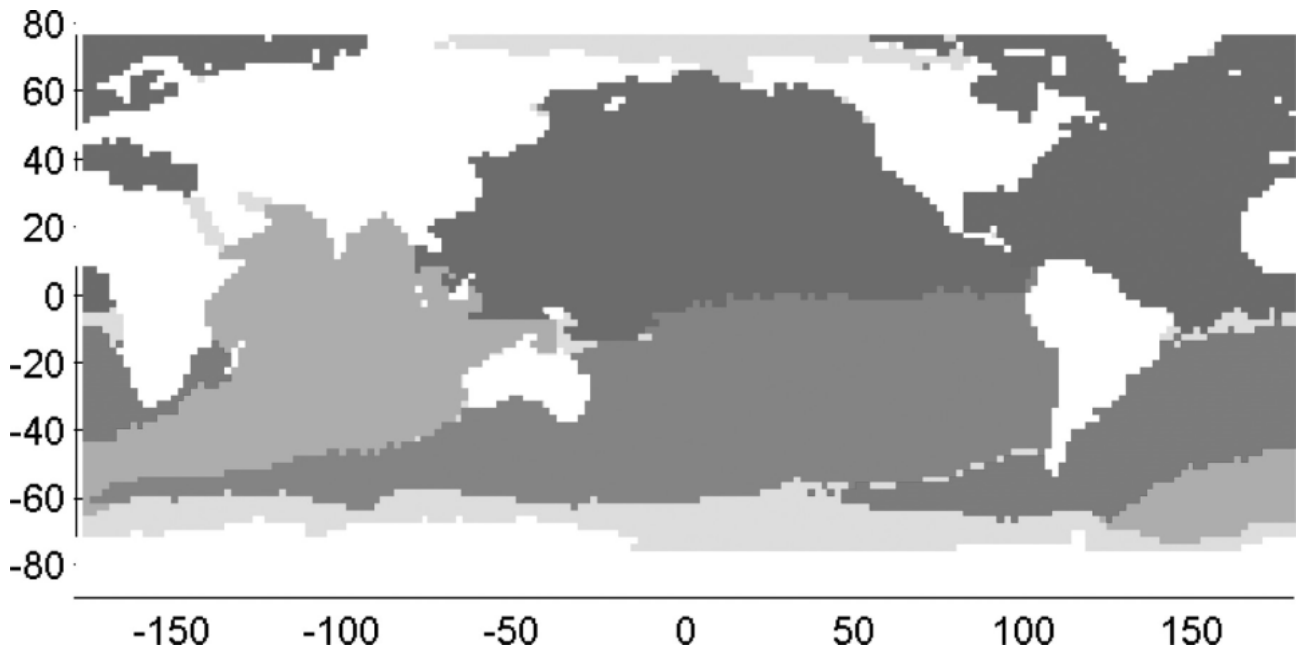


Figure 1.6: Surface basin boundaries derived from modeled trajectories. Map represents an amalgamation of eigenvectors of a transition matrix. Each shaded region corresponds to a dynamically-segregated region on a 60-day timescale. Extracted from Froyland et al. (2014).

network. The eigenvectors of a matrix represent the dominant spatial pattern of the corresponding transformation. Applied to a transition matrix of the ocean domain, the eigenvectors correspond to the dominant convergence and divergent patterns of ocean currents. Froyland et al. (2014) interpreted this as the destination regions for transport in the ocean. Figure 1.6 shows an amalgamation of eigenvectors: the ocean is split into sections and between these sections, transport is minimal. They are, as much as is possible in the chaotic ocean, dynamically segregated areas. Maximenko et al. (2012) achieved a similar result by linking each cell in the ocean to a gyre centre which a particle from that cell reaches first.

The resulting map provides a new definition for dynamical borders in the ocean (Figure 1.6). These borders give a dynamical context to accumulation of tracer in the ocean. For example, the North and South Pacific are split along the Equator due to the strong divergent Ekman transport and so there is minimal connectivity expected between the two basins. However, the equivalent boundary in the North Atlantic is shifted southward due to the presence of the northward Brazil Current. Meanwhile, the southern hemisphere ocean basins each extend westward in the Southern Ocean due to the rapid and direct pathways taken by drifters in the Antarctic Circumpolar Current (ACC).

## 1.5 Objectives and Outline

The main objective of this thesis is to quantify how inter-ocean connectivity changes with depth. This will be achieved using techniques from graph theory which are designed to detect common connections within large datasets of transport pathways. The resulting maps of connectivity will indicate and explain the distribution of dynamically different sections of the ocean. While network-based methods have been applied to the surface circulation, this thesis compares a similar method with more direct dynamical relevance. Following on from this, inter-ocean connectivity within the ocean interior will be studied. Given that surface connectivity helps explain the distribution of floating tracers, understanding interior connectivity will be important in understanding the distribution of interior tracers.

A key advancement on previous work is that the contribution of key dynamical processes will also be considered to identify mechanisms for inter-ocean connectivity. Whereas previous work has focused on global transport, the analysis here will take place on two key regions of inter-ocean exchange, the Agulhas Current System and the North Atlantic basin. The Agulhas Current is the most powerful western boundary current in terms of volume transported and maximum speeds (Beal and Bryden, 1999). It is the sole conduit from the Indian Ocean to the South Atlantic and is therefore a key branch of the global thermohaline circulation (Beal et al., 2011). Specifically, the transport between these regions occurs via Agulhas Rings, also known as Agulhas Leakage, which forms when the Current destabilises at its Retroflexion point (Duncombe Rae, 1991).

Meanwhile, the Gulf Stream and the North Atlantic Current act as the northern boundary of the North Atlantic sub-tropical gyre. Surface transport from the sub-tropical to sub-polar gyres is small (Brambilla and Talley, 2006). While the Gulf Stream plays a key role in transporting tropical heat, pole-ward heat transport in the ocean is roughly a quarter of atmospheric transport between 40°N and 60°N (Trenberth and Caron, 2001). The barrier is therefore manifest in heat transport estimates as well as drifter observations. Diagnosing connections to physical drivers is made simpler on the regional scale rather than on the global scale, where the role of individual currents and processes can be more clearly identified.

Chapter 2 describes the range of observational and model-based Lagrangian datasets used to assess ocean transport. Satellite-tracked surface drifters, satellite-derived fields of geostrophic and Ekman velocities, and circulation model fields will be used throughout the thesis to assess the capabilities of each in detecting connections. This chapter explores the benefits and weaknesses of each dataset. It also provides a more detailed explanation of transition matrix construction and how Markov Chain models summarise extensive Lagrangian datasets.

Transition matrices are a common feature in previous connectivity studies. Given their major role also within this thesis, their capability is rigorously assessed for the first time in Chapter 3. The inherent “leakiness” of this representation is quantified in an idealised oceanic flow, and a mechanism for this behaviour is proposed (McAdam and van Sebille, 2018). The capability of Markov Chain models in the real ocean is then re-assessed by repeating experiments used to detect plastic accumulation spots (e.g. Van Sebille et al., 2012).

Chapter 4 provides a new definition for surface boundaries between the Indian and Atlantic Oceans, and between the sub-tropical and sub-polar gyres in the North Atlantic (McAdam and van Sebille, 2018). These maps highlight new pathways between the regions. The physical mechanisms which drive such transport are explored by comparison to geostrophic transition matrices, which are introduced for the first time here.

Chapter 5 applies this analysis to isopycnal layers to capture the first-order component of interior transport. The chapter focuses on two isopycnals which sample different water masses and depths in the Agulhas and North Atlantic regions, to provide the first analysis of how connectivity change with depth. Given that this Chapter makes use of a  $\frac{1}{4}^\circ$  ocean model, the capability of this model is also assessed.

Finally, Chapter 6 summarises the findings of this thesis in three sections: the methodology, surface connectivity and the change in connectivity with depth. Future work and applications are also discussed.

# Chapter 2

## Ocean Connectivity Datasets & Methodology

### 2.1 Introduction

Some of the earliest oceanographic measurements recorded the drift of floating objects visible from ships (Stewart, 2008). Nowadays, international collaboration and technological innovations have produced global arrays of instruments which autonomously and continuously track the pathways of ocean transport. The first of such datasets discussed in this section is the Global Drifter Program (GDP), operated by the U.S National Oceanographic and Atmospheric Administration (NOAA). As the name suggests, it is an observational record of drifter motion throughout the global ocean. The equivalent of the GDP for the sub-surface is the Argo float program (Riser et al., 2016).

A complementary technique to analysing observed trajectories is the simulation of motion of virtual drifters. To do this, velocity fields are required as input for an interpolation and integration scheme which determines the position of such particles in time. This technique is known as Lagrangian analysis. In this thesis, Lagrangian analysis is performed for the surface ocean using satellite-derived velocity fields. These fields typically represent the motion of one physical driver of transport such as the geostrophic or Ekman component. Analysis of individual

components can diagnose their role in the full-field transport. A major benefit of the Lagrangian analysis, over observational datasets, is that it permits a larger number of trajectories to be determined. For studying the sub-surface, ocean general circulation models (OGCMs) will provide velocity fields for a full 3D ocean simulation. Virtual trajectories will therefore only capture processes represented in the OGCM used. Outside of the target depths of the Argo program however, there is no alternative but to apply Lagrangian analysis to model-derived fields. Trajectories provide an intuitive and visual way to identify pathways between different regions in the ocean and to quantify their relative contribution to regional and global circulation. However, to approximate the full range of connections, a large number of trajectories need to be considered. A transition matrix is a means of storing the information of a large trajectory dataset. It can be used to identify the most-likely routes taken by transportable media, such as drifters. One application is to construct a Markov Chain model, which makes predictions about the future transport. The Markov Chain model will be used throughout this thesis.

In this chapter, the advantages and restrictions of observed and virtual datasets are discussed. Section 2.2 begins with the observed datasets such as the GDP and Argo program. Then, in Section 2.3, Lagrangian analysis is described and relevant input velocities are introduced: geostrophic velocities from satellite-altimetry, the GlobCurrent project and 3D fields from the OGCM NEMO. Lastly, Section 2.4 describes transition matrix construction and how they are used to create Markov Chain models.

## 2.2 Observational Global Drifter Datasets

Perhaps the most intuitive tool for understanding ocean connectivity is a dataset of trajectories of water-following material. For this purpose, NOAA maintains a global network of satellite-tracked floating buoys known as “drifters”. The GDP has collected drifter coordinates on the surface since 1979. Due to motion relative to the satellites, drifter coordinates are inferred by the Doppler Shift of satellites relative to the drifters. An additional shift due to the Earth’s rotation is accounted for. Coordinates are recorded several times per day and are accurate

to within 150 m to 1 km, depending on the position of satellites hosting Argos or the Global Positioning System (Lumpkin and Pazos, 2007). They are then interpolated to provide a 6-hourly data set which is openly available at [www.aoml.noaa.gov/phod/dac](http://www.aoml.noaa.gov/phod/dac). Despite the release of an augmented version of the dataset which contains hourly readings (Elipot et al., 2016), the original 6-hourly data is used here because it provides sufficient temporal resolution for the Markov Chain model analysis described in Section 2.4.

The quality of a trajectory dataset depends heavily on the coverage of coordinate readings throughout the domain. Coverage is a function of drifter lifetime, the convergent nature of two-dimensional surface flow and drifter deployment. Lumpkin and Pazos (2007) stated the average lifetime of a drifter was approximately 450 days but the average length of drifter data available from March 2017 was 382 days. This is nonetheless within the typical time period for drifters to accumulate in gyre centres, meaning coverage will be somewhat biased towards these regions (Beron-Vera et al., 2016). Figure 2.1a shows the map of GDP coordinate data coverage. There is an order of magnitude difference in coverage between the gyre centres and the more divergent sections of the ocean such as the upwelling zones and the Equatorial zone. This is in spite of sparse deployment in some of the well-covered regions, for example the North Pacific gyre (Figure 2.1b). It is clear that the divergent dynamics of ocean currents create a coverage bias.

Moreover, uneven deployment also creates a coverage bias (Figure 2.1b). Deployment preferentially occurs along research cruise tracks and in western boundary currents but is generally sparse throughout the southern hemisphere. Due to the convergence and deployment biases, key areas of inter-ocean exchange such as the Indonesian Throughflow, the Arctic Ocean, much of the Antarctic Circumpolar Current and the Equator are poorly covered by drifter trajectories. Therefore, the transport in these regions will be less well represented by the Markov Chain model, in comparison to the gyre centres.

Coverage has remained uneven in recent years despite the continual deployment of drifters (Figure 2.2). Since the first global surface Markov Chain model study in 2012, the number of available historical drifters has doubled to roughly 20,000 (Maximenko et al., 2012). Since

the creation of the plasticadrift.org online Markov Chain model tool in 2014, the number of coordinate readings has increased from 24 million to 34 million (Van Sebille, 2014). Despite the growth of the drifter program, areas where coverage has increased mostly include previously well-covered regions. This highlights the dominance of ocean current dynamics over deployment bias.

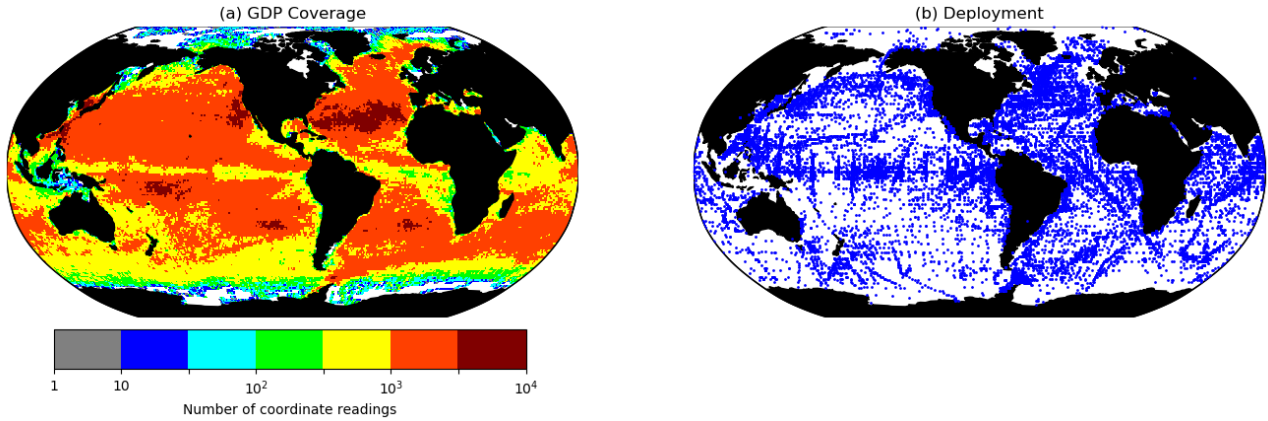


Figure 2.1: Global Drifter Program coordinate data maps. (a) Coverage of coordinate readings per  $1^\circ$  cell measured from 1979 to June 2017. White areas contain no coordinate readings (McAdam and van Sebille, 2018). (b) Coordinates of the drifter deployment positions from the same period (blue dots).

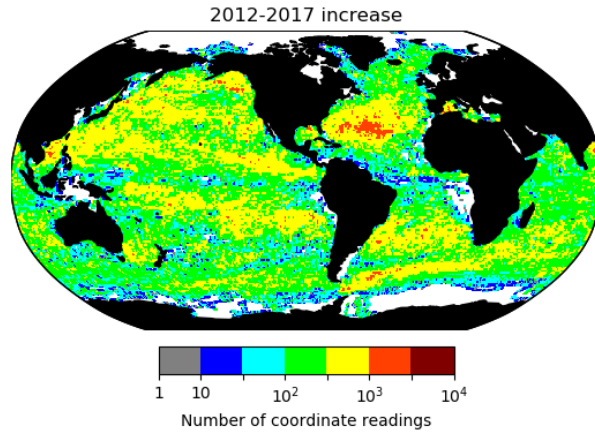


Figure 2.2: Global Drifter Program coverage increase since 2012, shown as the difference in values between 2012 and March 2017 (McAdam and van Sebille, 2018)

Two regions with relatively good coverage and distinct roles in the global ocean circulation are highlighted, for use in later case studies: the Agulhas Current System and the North Atlantic sub-tropical and sub-polar gyres. The Agulhas Current System, at the southernmost tip of Africa, is flanked by regions of high drifter coverage. The Agulhas Current is a western

boundary current which begins between the continent and Madagascar. At the southernmost tip of the continent, the current reaches its retroflexion point where the Agulhas Current splits into an eastward Agulhas Return Current and the north-westward Agulhas leakage. The drifter coverage within the current is relatively low for the area. This is to be expected in a rapid current system if drifters are not consistently deployed upstream. Despite the intense divergence away from the current, recent drifter coverage has increased. This implies drifters are consistently entering the Agulhas system.

Within the North Atlantic, the Gulf Stream acts as a surface transport front between the subtropical and subpolar gyres (Bower et al., 1985; Brambilla and Talley, 2006; Rypina et al., 2011). The surface flow on both sides of the Gulf Stream, as well as downstream, is relatively well sampled. In the sub-polar regions, coverage remains high with the exception of Baffin Bay, north of the Labrador Current. Nonetheless, the Labrador Current region remains moderately covered by hundreds of readings per  $1^\circ$  cell (Figure 2.2).

Besides studying two regions which are distinct in terms of connectivity, two sub-sets of drifters within the GDP dataset are also considered. These two different types of drifters, drogued and undrogued, have different current-following abilities due to the depth of flow they sample. All GDP drifters are deployed with a 15m-deep “sea-anchor” called a drogue, attached by a thin tether, to induce water-following behaviour. Drifters record the date when the tether breaks, typically due to general wear-and-tear or storms (Lumpkin and Pazos, 2007). The average time before the tether breaks is 187 days, while the average time between breakage and the end of a drifter’s lifetime (the “lifetime” of an undrogued drifter) is 216 days. After the date of tether breakage, they are labeled undrogued. Undrogued drifters provide roughly 55% of the total number of drifters coordinate readings available historically. Niiler et al. (1995) showed that they are more susceptible to wind-slip than drogued drifters and have a reduced water-following capability. Therefore, the transport fates of each sub-set will differ.

The coverage of each subset reflects the differing physical processes represented in each. The clearest difference is that the gyre regions contain more undrogued than drogued drifter readings (Figure 2.3). The deployment of undrogued drifters can be considered as the date of drogue

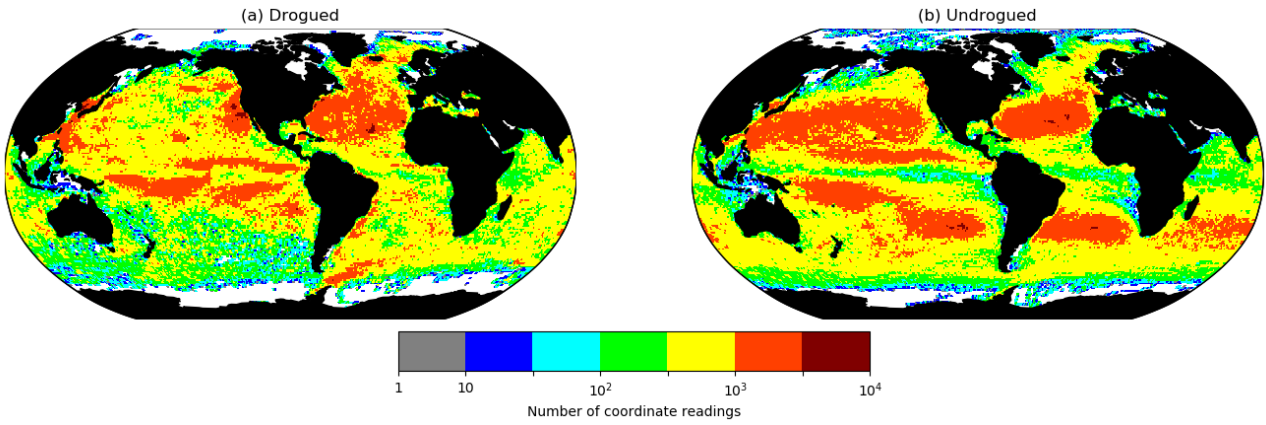


Figure 2.3: Global Drifter Program drogued and undrogued drifter coverage. As for Figure 2.1a, but for (a) drogued and (b) undrogued sub-sets. The coverage represents 45% and 55% of the total dataset respectively.

loss. By the time drifters lose their drogue, they are likely to have already accumulated in the gyre regions. Moreover, wind-slip causes faster accumulation than Ekman dynamics (Beron-Vera et al., 2016). Therefore, while both subsets of drifters are relatively oversampled in the gyres, this bias is larger for undrogued drifters.

It is common to combine both sub-sets in transport studies despite their dynamical differences (Maximenko et al., 2012; Van Sebille, 2014). The main reason for this is to ensure sufficient coverage which reduces the oversampling of the gyres relative to the more divergent regions. In this thesis, drogued and undrogued drifter sub-sets will be combined to construct Markov Chain models which represent a full-field 15m-depth averaged flow with maximum coverage. In some cases, the sub-sets will also be treated separately.

To further understand how connectivity changes with depth, a trajectory dataset which samples intermediate and abyssal waters is also required. One possibility is the ANDRO atlas, which is a dataset of Argo float displacements for depths up to 2750 dbar (Ollitrault and Rannou, 2013). The global Argo program currently consists of roughly 3,000 floats which descend to pre-defined isobars to measure temperature and salinity (Riser et al., 2016). Argo floats descend by means of an internal density-changing mechanism and they remain at their pre-defined isobar for close to 10 days.

The ANDRO dataset is not used in this thesis for the following reasons. Firstly, the time spent below the surface is rarely more than 10 days, which is considerably less than the length of a GDP surface drifter trajectory (Ollitrault and Colin de Verdière, 2014). Previously, Sevellec et al. (2017) created a 10-day transition matrix of the 1000 dbar layer using Argo floats. However, as will be shown in Chapter 2.3, the length of the trajectories plays a key role in determining the quality of the transition matrix and is an effect that has so far been overlooked in the literature. Secondly, the displacements in the 1250 to 2750 dbar range make up only 20% of the total. As a result, there are significant coverage gaps below the 1000 dbar level, particularly in the southern hemisphere. For studies of the global ocean, this restricts the use of the ANDRO dataset to the 1000 dbar layer or to specific regions at depth. While such a study would be interesting, the aim of this thesis is to compare more than one interior layer to the surface.

Before the Argo program, the SOFAR and RAFOS floats were used for tracking transport at depth. The Sound Fixing and Range (SOFAR) floats contained a spring-back piston which altered the compressibility, allowing the float to follow an isopycnal with impressive accuracy. For tracking purposes, the SOFAR floats emitted radio signals which were detected by a mooring station (Rossby and Webb, 1970). This mechanism was later reversed in a bid to improve the cost-effectiveness of the program, meaning that RAFOS floats would detect radio signals emitted from mooring stations (Rossby et al., 1986). In the early 1990s, the ALACE float was introduced as a replacement to the RAFOS floats (Davis et al., 1992). ALACE floats were tracked solely by the Argo satellites and would therefore become the pre-cursor to the Argo float program. Given that RAFOS/SOFAR floats were discontinued and that their use was restricted to smaller regional studies (e.g. Bower et al., 1985), they will not be studied in this thesis.

## 2.3 Virtual Drifter Datasets

Datasets of ocean trajectories can also be produced by numerical modelling of virtual drifters, providing an alternative to the global array of drifters described in Section 2.1. There are several advantages of using virtual drifters, referred to here as particles. Firstly, the practical and economic challenge of creating and maintaining a global array of drifters is avoided. Particle trajectories can also be simulated for a much longer timescale than the existing observational datasets. Moreover, particles can be deployed homogeneously and more abundantly than real drifters. There are drawbacks too. The dynamics represented by the dataset depend on the quality of the virtual velocity fields advecting the particles, which, due to resolution and computational restrictions, do not include the full range of observed physical processes.

Numerical modelling of virtual drifters requires a set of velocity-component fields. Velocity fields are extracted from many sources, either observational or numerically modeled, and may describe 2D or 3D flows. 2D flows include satellite altimeter-derived geostrophic velocities (e.g. Poulain et al. (2012)), mesoscale surface flow from the GlobCurrent repository (e.g. Rio et al. (2014)) and even surface velocities averaged from the GDP dataset (Koszalka et al., 2011). 3D velocity fields are typically extracted from OGCMs based on geophysical fluid dynamics. Many of the seminal applications of virtual particle transport have focused on global transport of water masses. Boning and Cox (1988) were among the first to quantify diffusivity with a Lagrangian analysis. Drijfhout et al. (1996) traced the pathways of thermohaline circulation in an OGCM. Particular successes of this approach have been to highlight the importance of various routes in global circulation and the Southern Ocean in inter-ocean exchange (Blanke et al., 2001; Speich et al., 2001). With recent advancements in computational resources, higher resolution OGCM velocity grids now include topographic and mesoscale effects (e.g. Tamsitt et al., 2017).

In principle, the computation of Lagrangian trajectories of particles is determined by integrating the velocity field:

$$\mathbf{X}(t + dt) = \mathbf{X}(t) + \int_t^{t+dt} \mathbf{v}(\mathbf{x}(\tau), \tau) d\tau, \quad (2.1)$$

where  $\mathbf{X}$  is the Eulerian position of the particle in the domain,  $\mathbf{v}$  is the velocity vector at the particles position and  $d\tau$  is the integral timestep. Discretised numerical integration for particle tracking is typically achieved with a 4th-order Runge-Kutta (RK4) method (van Sebille et al., 2018). The velocity field must be spatially interpolated to determine the value at the particles' locations. Moreover, the velocity field must be temporally interpolated if the value of the time-step ( $dt$ ) does not match the time difference between input velocity fields.

Although Equation 2.1 may be solved with relative ease, it is good practice to use one of the several particle integration codes which have been made open-access by the research groups who produced them. The codes typically employ computationally-efficient solvers and have been validated against theoretical examples. An up-to-date review of Lagrangian analysis and community-developed codes was performed by van Sebille et al. (2018). Throughout this thesis, the OceanParcels toolkit<sup>1</sup> has been employed (Lange and van Sebille, 2017). OceanParcels makes use of the JIT (Just in Time) compilation method for running the integration code in a more efficient manner than standard python modules. OceanParcels also facilitates the creation of custom kernels which can be used for a wide range of applications, from sampling of thermodynamic variables to adding diffusion-style terms to the advection in Equation 2.1.

### 2.3.1 Surface Transport from Globcurrent

The GlobCurrent repository<sup>2</sup> provides global fields for the surface geostrophic currents, surface Ekman currents and 15 m-depth Ekman currents. The combined geostrophic and Ekman flow field at the surface is also available. GlobCurrent datasets were produced by merging satellite-derived current estimates with in-situ measurements such as drifter velocities, as well as producing an empirical model of Ekman transport to which drifter data is fit (Rio et al., 2014). Many advancements in altimeter and drifter data processing have contributed to the GlobCurrent product. The corresponding references are listed on the Bibliography page of the GlobCurrent site. The product fields are provided at a global  $\frac{1}{4}^\circ$  resolution at 3-hourly

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<sup>1</sup>[www.oceanparcels.org](http://www.oceanparcels.org)

<sup>2</sup>[www.globcurrent.ifremer.fr](http://www.globcurrent.ifremer.fr)

time-steps from 2002 to 2014. GlobCurrent products show marked improvement over alternatives such as OSCAR (Ocean Surface Current Analysis Real-time) in detecting near-surface structures such as Ekman spiraling (Rio et al., 2014).

Although the undrogued and drogued GDP sub-sets already provide Lagrangian datasets of the surface ocean at both 0 m and 15 m, deploying virtual particles in GlobCurrent ensures coverage is better and more homogeneous. However, direct comparison between GDP drifters and Globcurrent-derived particles is not viable because the Globcurrent total field contains only the Ekman and geostrophic flow components. The  $\frac{1}{4}^\circ$  spatial resolution does not resolve sub-mesoscale features. The use of GlobCurrent-derived trajectories therefore serves as an estimate to a hypothetical enhancement of the GDP dataset, in which deployment bias is negligible and high data coverage is widespread.

To highlight the benefits of using Lagrangian particles, Figure 2.4 compares the coordinate data coverage of two GlobCurrent-derived datasets. In one, roughly the same number of particles as the GDP drifters is deployed. In the other, an order of magnitude more is deployed. Both sets of particles are advected forward in time using the combined Ekman-geostrophic fields for the 3 year period from 2012-2014, which is longer than the average timespan of the drifters. Both sets are also deployed homogeneously, to remove the deployment bias. As desired, both the gyre regions and divergent regions contain an order of magnitude more readings in the larger dataset than in the smaller. Moreover, coverage in regions such as the Arctic and the Indonesian Throughflow is greater than in the GDP dataset (Figure 2.1). Despite the expected widespread increase in coverage, the circulation bias maintains an order-of-magnitude difference between divergent and convergent regions such as the Equatorial regions and the gyres respectively.

### 2.3.2 Isopycnal Surfaces from NEMO

GlobCurrent products are for the surface only. Moreover, sub-surface trajectory datasets from ARGO and SOFAR/RAFOS floats suffer from poor coverage and restrict analysis to particular depth surfaces. To perform Lagrangian studies in the ocean interior, the alternative is to extract velocity output from OGCMs. The Nucleus for European Modelling of the Ocean

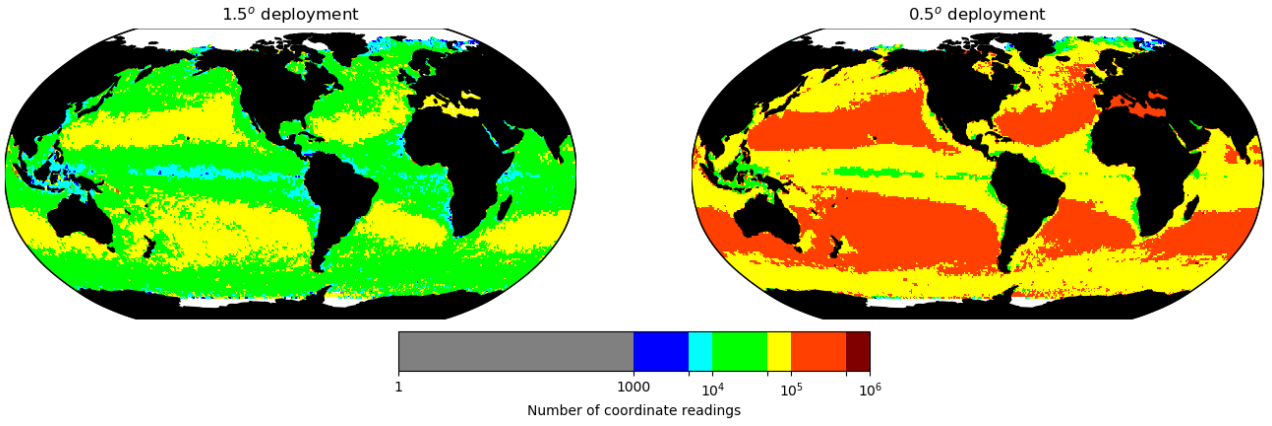


Figure 2.4: GlobCurrent-derived virtual particle coverage maps. Particles are released at the initial time-step only and run for two years, over the 2014-2015 period. Particles were integrated in the daily  $\frac{1}{4}^\circ$  GlobCurrent surface velocity fields with the OceanParcels toolkit (a)  $\sim 20,000$  particles corresponding to a  $1.5^\circ$  release (b)  $\sim 200,000$  particles corresponding to a  $\frac{1}{2}^\circ$  release.

(NEMO) is a modelling framework which consists of various physical engines, such as ocean, sea-ice and biochemistry drivers. It is developed by six institutions across Europe and has contributed to nearly 200 projects which are listed at [www.nemo-ocean.eu/projects](http://www.nemo-ocean.eu/projects). NEMO provides openly-available reference configurations of its various engines. For example, the ORCA model configuration of NEMO is a combination of their ocean physics and the sea-ice drivers (Madec and the NEMO team, 2008).

The ORCA model is run on an Arakawa C-grid with a tripolar structure defined by Madec and Imbard (1996), in which the grid has a regular Mercator structure south of  $20^\circ\text{N}$ . The output used here is from a run performed by the National Oceanography Centre (NOC) System Modeling Group in which two horizontal resolutions, an eddy-permitting  $\frac{1}{4}^\circ$  and an eddy-resolving  $\frac{1}{12}^\circ$  at the Equator, are available. In both cases, due to the tripolar grid, the cells become distorted north of  $20^\circ\text{N}$ . Output is provided at 5-day time-steps and along 64 vertical levels. The  $\frac{1}{4}^\circ$  run covers 1958 to 2001 while the  $\frac{1}{12}^\circ$  run covers 1978 to 2010. Extracting isopycnal velocities will be the focus of Section 5.2.

## 2.4 Transition Matrices and Markov Chain Models

The concept of the transition matrix was introduced in Chapter 1. In the literature, transition matrices are sometimes referred to as transfer operators, and their construction referred to as Ulam’s method (Froyland et al., 2007; Dellnitz et al., 2009). A detailed theoretical treatment of their construction from ocean-based trajectories exists in other oceanographic studies (e.g. Ser-Giacomi et al., 2015). Here, we provide a descriptive overview of the construction of transition matrices and their application to Markov Chain models.

First, the relevant parameters for transition matrix construction are explained in more detail. A transition matrix,  $\mathbf{T}$ , is a summary of information on transport between parts of a domain. The domain of study in this case is the entire Earth, including land and ocean. The domain is gridded into  $n$  cells, each with a size  $dx \times dx$  and labeled from 1 to  $n$ . There is no requirement for cells to be square. The transition matrix has a size  $n \times n$ . Transport between cells is represented by drifter or particle trajectories, which depict the entire lifetime of the drifter or particle. However, these trajectories are then split into sub-trajectories of length  $dt$  so that the transition matrix will represent transport throughout the domain during a time  $dt$ .

The construction of a transition matrix is as follows. The initial,  $i$ , and final,  $j$ , cells of a sub-trajectory of length  $dt$  are determined. The corresponding transition matrix element  $T_{i,j}$  is incremented by one, indicating that so far there is one connection between cells  $i$  and  $j$ . The next sub-trajectory begins at a time  $\tau$  after the beginning of the first. Again, the initial and final cells are determined and the corresponding transition matrix element is incremented. This process is repeated until the end of the drifter trajectory, after which the process is applied to the next drifter in the dataset. Repeating this process until all drifters have been covered will result in a transition matrix which stores the number of connections between each pair of cells. The construction is therefore a spatial discretisation of the transport.

The transition matrix can be made to represent a chosen time-step. Maximenko et al. (2012), for example, used  $dt = 5$  days. Using the method described above, the transition matrix describes a time-averaged flow over the period when GDP data is available (roughly 30 years). Van Sebille

et al. (2012) split the dataset into bimonthly sub-sets and created six transition matrices, each corresponding to the time-average of a bimonthly period. While including seasonality may be a requirement for certain applications, annual-averages are sufficient in this thesis.

Besides  $dt$ , a second temporal parameter,  $\tau$ , was introduced.  $\tau$  is the time difference between sub-trajectories.  $\tau$  is kept smaller than  $dt$ , to allow for more sub-trajectories to be extracted, thereby providing more information on transport. The smaller  $\tau$  is, the larger the number of sub-trajectories available for transition matrix construction. Likewise, choosing a large  $\tau$  reduces the number of sub-trajectories extracted and may reduce the quality of the transition matrix representation. This is equivalent to reducing the coverage of the dataset.  $\tau$ , therefore, also controls the quality of the transition matrix. Throughout this thesis,  $\tau$  will be set as 1 day as a compromise between using more sub-trajectories and matrix construction time.

For the transition matrix to make useful predictions about future transport in the domain, the matrix elements are converted into probabilities. Each element  $j$  in row  $i$  initially depicts the number of drifters flowing out of cell  $i$  and into cell  $j$ . Each element can be normalised by the total number of drifters leaving cell  $i$ . The result is a transition matrix composed of probability distribution functions (PDF) of transport in a time  $dt$  for each cell. Each row  $i$  stores the PDF for the corresponding cell.

A Markov Chain model uses row-normalised transition matrices to make predictions about transport in the domain. Alongside the matrix, which describes the transport, a description of the transportable medium is required. This is provided by an input column-vector  $\mathbf{R}$  of size  $n \times 1$ , which defines the current distribution of a tracer in the domain. Each input vector element value denotes the quantity of tracer in the corresponding cell. The Markov Chain model, which maps  $\mathbf{R}$  into a future state, is a chain of vector-matrix multiplications:

$$\mathbf{R}^{t_0+dt} = \mathbf{T}_{dt} \mathbf{R}^{t_0}, \quad (2.2)$$

where the output column-vector  $\mathbf{R}^{t+dt}$  provides the distribution of tracer after a time  $dt$ . Therefore, the sub-trajectory length  $dt$  is also the model time-step.

The output of the model can be interpreted in two ways. First, output vectors may represent the dispersion of a cloud of tracer after a time  $dt$ . If the input vector consists of just one non-zero cell value, Equation 2.2 also determines all possible destinations for a single particle released from this cell. In the Markov Chain model, a future state depends only on the present. In this sense it is “memoryless”, despite being constructed from drifter trajectories observed in the system’s past. Throughout this thesis, the suitability of describing the ocean in a Markovian way will be assessed.

Another aspect of Markov Chain model theory which is directly relevant to oceanographic applications concerns the choice of timestep. In theory, evolving a system to a given time using different transition matrices should return the same output. For example, four multiplications of  $\mathbf{T}_{dt=5days}$  should return the same output as one evolution with  $\mathbf{T}_{dt=20}$ . In short, the following statement should be true:

$$(\mathbf{T}_{dt})^n = \mathbf{T}_{ndt}. \quad (2.3)$$

To highlight the importance of this concept and to visualise how the Markov Chain model works, a tracer release experiment is performed south of the Panama Isthmus in the central Pacific (Figure 2.5). When observing the drifter trajectories in the area, no coherent flow pattern is visible. Indeed, a Markov Chain model run shows that the tracer cloud extends in all directions and largely remains within a few degrees of the source region. To test Equation 2.3, the experiment is repeated for transition matrices built from  $dt = 5$ ,  $dt = 20$ ,  $dt = 60$  and  $dt = 180$  day sub-trajectories (Figure 2.6). For all scenarios, cell size  $dx$  is kept to  $1^\circ$  and  $\tau$  to 1 day; these parameter choices will be explored in Chapter 3.2. The output for the different transition matrices displays clear variation in the distribution of the tracer. For example, tracer reaches north of  $10^\circ\text{N}$  in the  $\mathbf{T}_{dt=20}$  model but not in the  $\mathbf{T}_{dt=60}$  model. Moreover, an apparent transport across the Panama Isthmus varies between models: 1.24%, 0.109% and 0.004% of tracer is leaked into the Caribbean Sea in the 5-day, 20-day and 60-day models respectively. No tracer is leaked in the 180-day model during the run. The Markovian principle described in Equation 2.3 is therefore not obeyed, at least not when using GDP drifter and the parameter values used here. Chapter 3 begins to explore why this is the case.

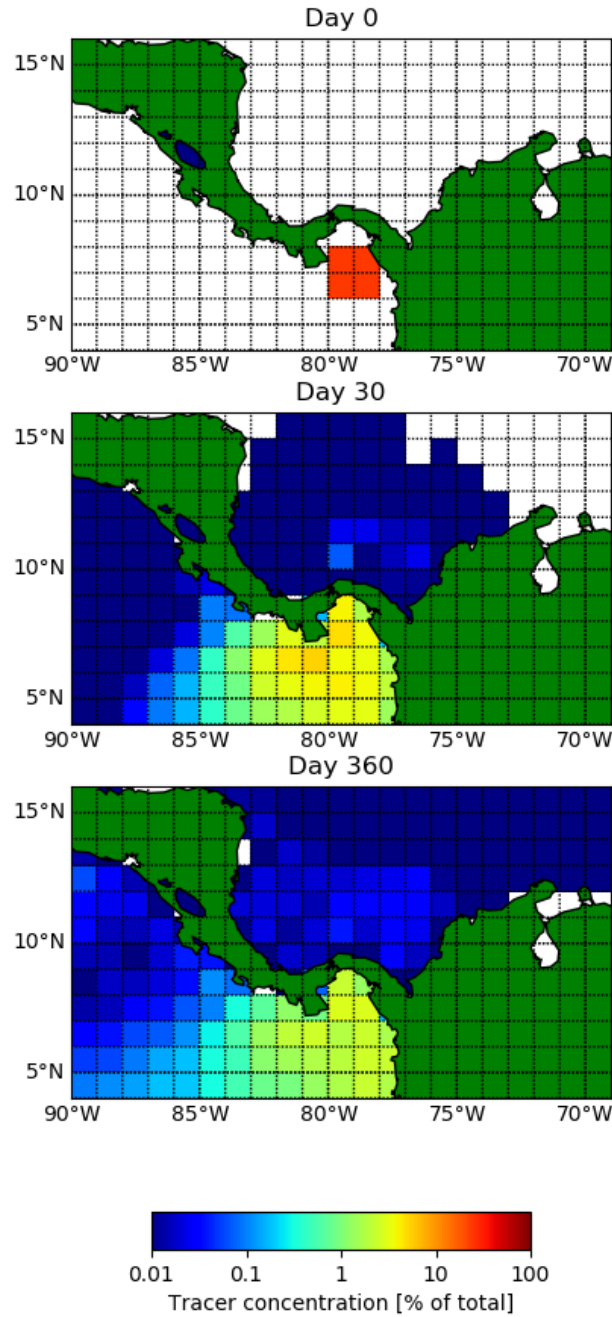


Figure 2.5: Markov Chain model of tracer release near the Panama Isthmus. This model uses transition matrices constructed from the entire GDP dataset, with a cell-size  $dx = 1^\circ$  and time-step  $dt = 5$  days. Tracer is released in a  $2^\circ \times 2^\circ$  area centred around  $(7^\circ\text{N}, 79^\circ\text{W})$  and run for 360 days (McAdam and van Sebille, 2018).

This analysis of the Panama Isthmus highlights another concerning feature of the Markov Chain model (Figures 2.5 & 2.6). Tracer which is released in the central Pacific appears to disperse into the Caribbean Sea, despite there being no physical connection between the two water masses on this timescale. As the tracer transport depends on the drifter connections between cells, this artificial dispersion must be caused by spurious connections. This occurs after one matrix multiplication. An explanation is that a trajectory which briefly passes into the corner of a cell will lead to the Markov Chain model filling the entire area represented by the cell with some tracer. One cell may contain unconnected flows yet they are artificially connected in the transition matrix because of the finite cell size. In this case, cells around the Panama Isthmus contain both Caribbean and Pacific trajectories. The artificial dispersion is therefore a function of cell size and time-step. This is the focus of the next chapter. While the Panama Isthmus problem may be fixed by adjusting cells so that no cell contains two separated oceans, less abrupt ocean barriers such as fronts cannot be treated in the same way. The effect of artificial dispersion in detecting ocean barriers will be studied throughout this thesis.

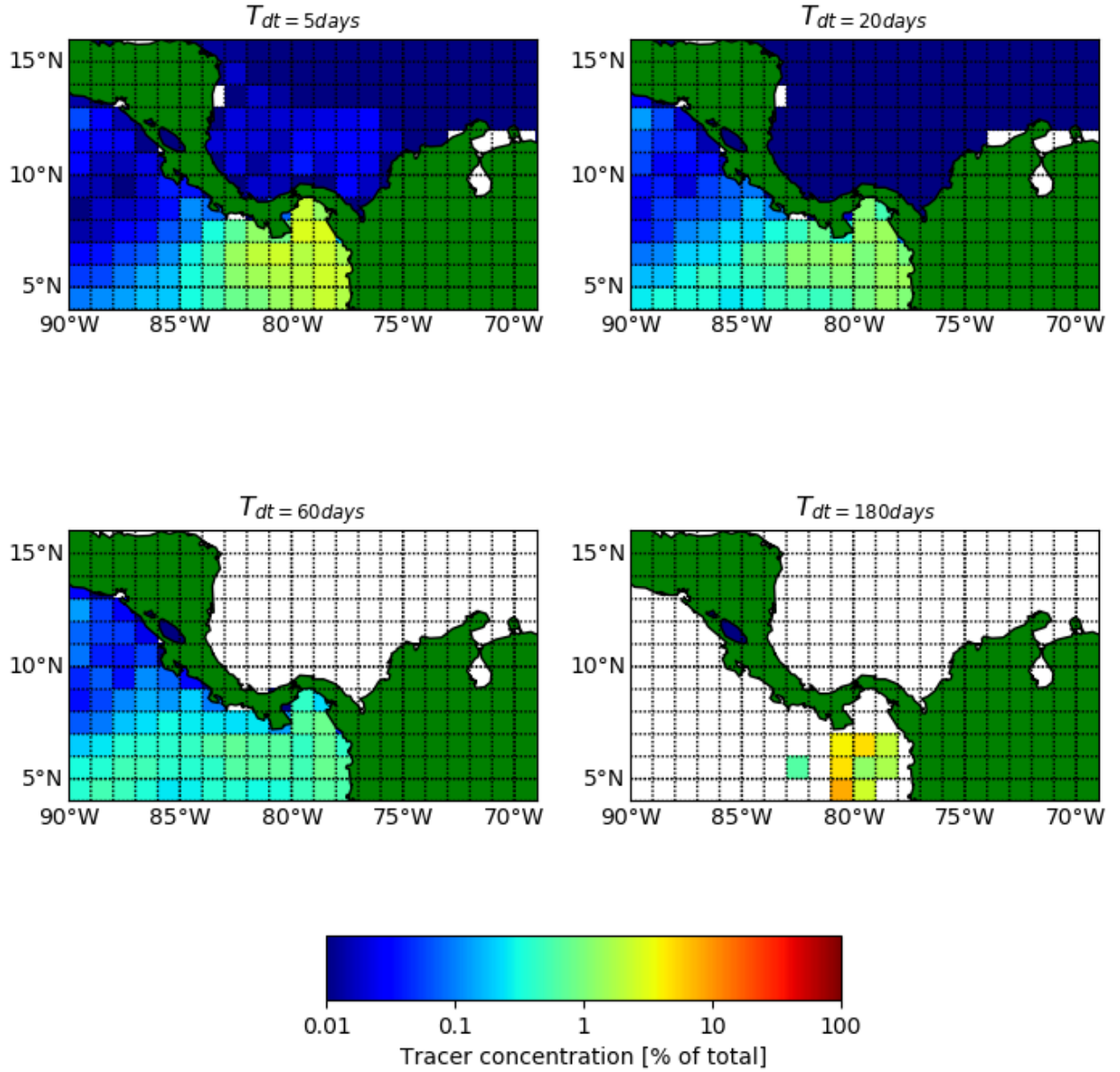


Figure 2.6: Markov Chain model output and time-step dependence. (a) The tracer release from Figure 2.5 is repeated with transition matrices  $T_{dt=20 \text{ days}}$ ,  $T_{dt=60 \text{ days}}$  and  $T_{dt=180 \text{ days}}$ . Output from day 360 is shown (McAdam and van Sebille, 2018).

## 2.5 Summary

Pathways of water-following objects offer an intuitive means of understanding transport dynamics, by explicitly highlighting spatial connections and the timescales of circulation. A range of widely-used observational and numerically-modelled trajectory datasets have been introduced in this section. The Markov Chain model has also been introduced as a means of analysing connections in large trajectory datasets, it will be used throughout this thesis. For the surface ocean, the Global Drifter Program (GDP) stores the trajectories of roughly 30,000 GPS-tracked drifters from the past 40 years and therefore serves as the primary surface dataset. Moreover, it has previously been used for regional and global Markov Chain models. The relatively high and consistent coverage of key regions of inter-ocean exchange, such as the Agulhas current and the North Atlantic, further justifies the use of GDP-based Markov Chain models in these regions.

This thesis also employs a community-led Lagrangian analysis code, OceanParcels, to advect virtual particles along satellite-derived and modelled velocity fields. The GlobCurrent surface flow, a combination of geostrophic and Ekman velocity components, cannot be directly compared to GDP drifters due to missing sub-grid dynamics in the former. However, they are used to create trajectory datasets with greater surface coverage than the GDP. By comparing with virtual particles, the adequacy of the GDP dataset in Markov Chain models can be confirmed. In the absence of a widespread observational dataset for the sub-surface, virtual particle releases will be performed along isopycnal surfaces in the NEMO ORCA velocity fields.

The Markov Chain model will be applied to all the described trajectory datasets in order to assess connectivity. The details of using Markov Chain models to draw conclusions about connectivity will be described in the following chapters. As shown in this chapter, the Markov Chain model is sensitive to the transition matrix construction parameters. There is also an inherent artificial dispersion within the model. Before applying the methodology to the regions of interest, the next chapter will assess these spurious phenomena.

# Chapter 3

## Markov Chain Model Parameter Sensitivity Analysis

### 3.1 Introduction

Markov Chain models highlight connections between regions in the ocean. Like a numerical model based on geophysical fluid dynamic equations, it requires the ocean to be treated as a domain of cells in which the position of some tracer cloud can be propagated in time by a fixed time-step. These values can be adjusted to the needs of the experiment or application. For example, depicting global circulation through the five major gyres may be suited by a  $1^\circ$  cell size and a time-step of several weeks for a decade run. On the other hand, using the model to track local contamination near a pollution source would require a cell size of several meters and a hourly time-step. Unlike numerical models, the Markov Chain model output depends considerably on the parameters used.

The Markov Chain model set-up depends on three factors: the cell size,  $dx$ , the time-step,  $dt$ , and the coverage of trajectory data available. In Section 2.4, it was shown that different time-step values contribute differently to spurious and unphysical connections. Section 2.4 also showed how different time-step models can arrive at different outputs for an equivalent tracer release (Equation 2.3). Previously, Van Sebille (2014) acknowledged the importance of greater

coverage in certain geographical areas, while Rypina et al. (2016) noted the “leakiness” of Markov Chain models changing with time-step. This chapter will quantify model output variation with the parameters and identify a mechanism for this behaviour. It will also determine the impact of trajectory dataset size by comparison with virtual particle dataset.

In Section 3.2, the Markov Chain model will be applied to an idealised flow with an analytical solution. This will ensure that the deviation of the Markov Chain model output from the solution can be easily quantified. In Section 3.3, the ability of the Markov Chain model to detect regional-scale features in a GDP-based model will be tested. The chapter will continually discuss the justification of using Markov Chain models.

## 3.2 Idealised Stommel Gyre

In Chapter 2.3, it was shown that artificial dispersion in Markov Chain models depends on transition matrix parameters. To better understand how artificial dispersion occurs, the output sensitivity to the parameters  $dt$  and  $dx$  is assessed. Instead of performing a sensitivity analysis on Markov Chain models based on the non-linear pathways of observed drifters, a purely advective flow with a pre-defined, analytical solution is used. No stochastic dispersion term representing sub-resolution dynamics is added (e.g. Berloff and McWilliams, 2003). The exact solution is purely advective, and any deviation in the output from the analytical solution will be artificial.

The analytical flow used here is the anti-cyclonic western-intensification subtropical gyre system derived by Stommel (1948). The Stommel gyre is often used to test the quality of numerical solvers in Lagrangian analytical toolkits (Fabbroni, 2009). The stream function is defined as:

$$\Psi = A \sin\left(\frac{\pi y}{b}\right) \left[ e^{(x-\lambda)\pi/b} + e^{-b/\pi} - 1 \right], \quad (3.1)$$

where  $A$  is a constant which incorporates the model wind stress, friction and domain size. The zonal,  $b$ , and meridional,  $\lambda$ , domain dimensions are both 10,000 km.

The Eulerian velocity field is obtained by numerically solving the following:

$$\begin{aligned} u &= -\frac{\partial \Psi}{\partial y}, \\ v &= \frac{\partial \Psi}{\partial x}. \end{aligned} \tag{3.2}$$

Using  $A = 100 \text{ m}^2/\text{s}$  yields velocities which peak at  $\sim 50 \text{ cm/s}$  in the western boundary current and  $\sim 5 \text{ cm/s}$  in the eastern gyre interior. The domain of the velocity field is a 2D horizontal plane with no vertical motion. Particles are deployed at time  $t = 0\text{s}$  and at  $100\text{km}$  resolution, providing 10,000 particles. The integration uses a Runge-Kutta 4 (RK4) time-step of 6 hours, saving particle positions every 24 hours, and runs for 5 years. The resulting trajectories are shown in Figure 3.1. Although this is not enough time for the outermost particles to complete a full loop, the deployment is sufficiently dense and the integration time sufficiently long that all cells are connected to at least several others.

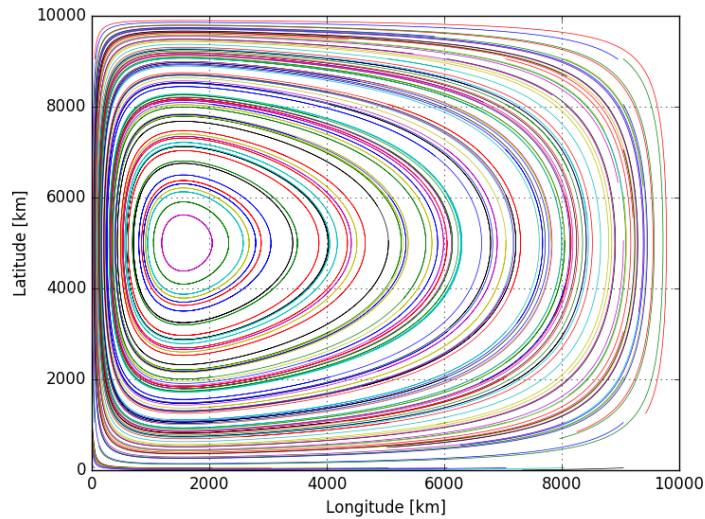


Figure 3.1: Stommel's Western Intensification Gyre. Trajectories of particles in a anti-cyclonic northern hemisphere gyre are advected forward in time using Equations 3.1 3.2. Individual trajectories are represented by different colours. Particles are initialised at  $100 \text{ km}$  spacing in both latitudinal and longitudinal directions and integrated for 5 years. 10% of the trajectories are shown for visual clarity (McAdam and van Sebille, 2018).

The transition matrices are created as described in Chapter 2.4. In this case-study, however, we define spatial resolution in kilometres, using 50, 100 and 200 km. These approximately correspond to  $\frac{1}{2}^\circ$ ,  $1^\circ$  and  $2^\circ$ , which are typical values used in the literature (e.g. Maximenko

et al., 2012). The time steps used are  $dt = 5$  days, 20 days, 60 days and 180 days. For the Markov Chain model experiment, tracer is then released within one cell in the western boundary current and is tracked over a 7-year period. This captures the journey from the boundary current into the slower eastern section of the domain and back again into the western edge.

Figure 3.2 shows the final output for the model run using the range of parameters. The Markov Chain representation of the Stommel flow is poor, regardless of the parameters used in this experiment. The initial cell expands into a cloud of tracer which extends both perpendicular and parallel to the streamline it should be following. The Markov Chain model transport from a cell is an average of the flow in that cell and, therefore, the tracer is not sampling one streamline but all the streamlines which flow through that cell. As a result, it is unreasonable to expect the tracer to follow an individual streamline exactly. Nonetheless, the continual growth of the tracer cloud becomes overly inaccurate. As a wider and wider area is sampled by the tracer cloud, cells which are not connected to the initial cell by any streamline are now covered by the tracer cloud. This process leads to parts of the tracer cloud departing from the streamline path by thousands of kilometres. Moreover, the model output is shown here to deviate further from the desired result for large values of  $dx$  and small values of  $dt$ . The following paragraphs will explain this result.

Across-streamline and along-streamline dispersion, as opposed to meridional and zonal dispersion, are analysed because the streamline is the desired tracer transport. A similar method has been used in the calculation of drifter dispersion relative to potential vorticity contours (LaCasce, 2008; Klocker et al., 2012). The across-streamline and along-streamline dispersion are calculated relative to the tracer cloud's centre of mass using:

$$\begin{aligned} D_{\perp} &= \frac{\sum (c_i x_{\perp})^2}{\sum c_i}, \\ D_{\parallel} &= \frac{\sum (c_i x_{\parallel})^2}{\sum c_i}, \end{aligned} \tag{3.3}$$

where  $c_i$  is the tracer concentration in each cell  $i$  and  $x$  is the distance from the centre of mass.

Specifically,  $x_{\parallel}$  is the along-streamline distance between the cell centre and tracer centre of mass while  $x_{\perp}$  is the distance between the cell centre and the nearest point of the streamline.

Using larger  $dt$  values reduces the dispersion in both directions and thus produces an output tracer cloud which more closely follows the streamlines than a smaller  $dt$  (Figure 3.2a-d). At every Markov Chain multiplication step in Equation 2.2 (Section 2.4), there is some artificial dispersion into adjacent cells. Evolving the system to a given time with a smaller  $dt$  will require more vector-matrix multiplications. Therefore, artificial dispersion occurs more rapidly in models with smaller  $dt$ . This is highlighted by the order of magnitude difference between the  $T_{dt=5}$  and  $T_{dt=180}$  model tracer cloud dispersion components.

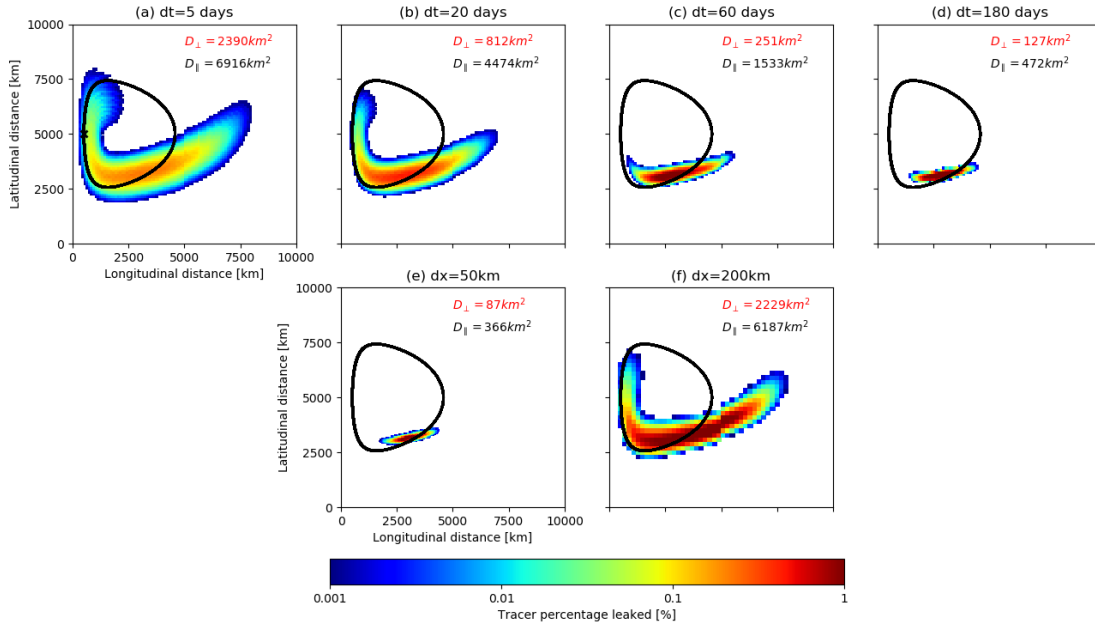


Figure 3.2: Markov Chain model applied to the Stommel gyre. The Markov Chain model was initialised with tracer in one cell centred at  $(0.5 \times 10^3 \text{ km}, 5 \times 10^3 \text{ km})$ . (a-d) Model output comparison after 7 years for time-steps  $dt=5$ , 20, 60 & 180 days, with cell size fixed at  $dx=100$  km. (e-f) Model output comparison for cell sizes  $dx=50$ , 100 & 200 km, with time-step fixed at  $dt=60$  days. The black streamline corresponds to the tracer release cell to indicate the analytical solution. The perpendicular and parallel components of dispersion from Equation 3.3 are shown in red and black text respectively (McAdam and van Sebille, 2018).

Likewise, a similar change in the model output is observed when decreasing the cell size from 200 km to 50 km (Figure 3.2c, e & f). As cell size increases, the amount of artificial dispersion increases. As explained in the previous paragraph, the cloud grows with each timestep by expanding into adjacent cells. However, the size of the cell also affects the dispersion be-

cause using larger resolution further increases the area covered by the tracer. While using shorter time-steps allows more spurious connections between adjacent cells to be included in the transition matrix, using larger cells increases the area covered by tracer because of these connections.

For all parameter choices, the dispersion is larger in the along-streamline than the across-streamline direction by one order of magnitude. Meanwhile, in all experiments the cloud shape is skewed towards the centre of the gyre, where the flow is faster than the outer regions. The part of the tracer cloud which disperses towards the centre will experience a faster advection than the part which disperses away from the centre, and creates an acceleration and a lag on each side respectively. Therefore, the artificial dispersion is also a function of the form of the flow field. Regardless of the quality of the model output, the difference between model output from a range of parameters contradicts the Markovian behaviour condition (Section 2.4, Equation 2.3). Therefore, it is shown here that artificial dispersion is an underlying cause of non-Markovian behaviour in any system which maps trajectories onto grids. In other words, the Markov Chain model is overly dispersive. More importantly for the applications of this thesis, the Markov Chain model output does not provide an accurate representation of idealised flow unless certain parameter conditions are met. To maximise accuracy, the optimal parameter choice is a large  $dt$  and a small  $dx$ .

### 3.3 GDP Drifters and Plastic Accumulation Regions

Section 3.2 showed that the Markov Chain model adds dispersion to a purely-advective flow. Rypina et al. (2016) noted that GDP-based Markov Chain models are also inherently over-dispersive near frontal regions and suggested that the cell size and time-step sizes are a controlling factor. Before such findings were made, Markov Chain models had been used to recreate global surface circulation and to detect the accumulation of plastic waste (Maximenko et al., 2012; Van Sebille et al., 2015). To justify the use of this methodology, in previous drifter studies as well as in this thesis, the capabilities of Markov Chain models with GDP trajectories are

explored. For consistency with later work involving sea surface height observations in Chapter 4, only trajectories from 1993 onwards are considered. This provides 85.7% of the total GDP dataset.

It is not immediately clear if the artificial dispersion manifests itself to the same extent when using a more convoluted trajectory dataset. The GDP dataset differs from the analytical case study in two main ways. First, the analytical scenario contains thousands of times more trajectories and, therefore, more connections between cells. Secondly, physical drivers such as turbulence and sub-mesoscale eddies contribute to dispersion both parallel and perpendicular to the time-averaged pathways of the flow. The presence of natural dispersion therefore makes it difficult to distinguish between artificial and natural dispersion. It is also possible that natural dispersion dominates over artificial dispersion.

To highlight the role of artificial dispersion in GDP-based models, experiments described by Maximenko et al. (2012), used to detect basin-scale accumulation of plastic debris, are repeated for a wider range of transition matrix parameter values. An arbitrary but uniform concentration of tracer is released in every ocean grid cell and evolved forward in time by the transition matrices for 50 years. As for the Stommel gyre case-study, the parameter choices are  $\frac{1}{2}^\circ$ ,  $1^\circ$  and  $2^\circ$  for  $dx$  and 5, 20, 60 and 180 days for  $dt$ . 50 years is sufficient time for gyres to accumulate basin-scale areas of tracer in the North Atlantic, South Atlantic, North Pacific, South Pacific and Indian Ocean basins (Figure 3.3a). The metric for accumulation used here is the total concentration of tracer found in cells which have a concentration larger than twice the initial value (Figure 3.3b), as in the study by Van Sebille et al. (2012). This metric will be herein referred to as a “patch” because, although arbitrary, the resulting areas are similar in size and geographic position to observations of the so-called plastic garbage patches (Law et al., 2014; Van Sebille et al., 2015).

For all parameter choices, the accumulation in each gyre is consistent between different models but only for the first 5-10 years (Figure 3.4). Beyond this time-scale, the evolution of patch mass varies with parameter choices. For example, changing  $dt$  can cause a factor of 2 change in the 50-year patch sizes, while the Indian Ocean Patch does not appear in all configurations.

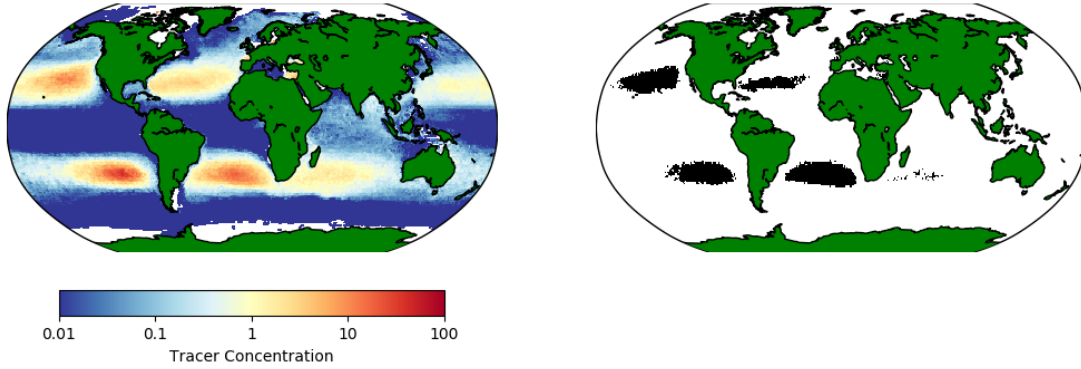


Figure 3.3: Surface accumulation in gyres. (a) Tracer distribution after 50-years after a global, homogeneous release, using  $T_{dt=60days}$ . (b) Areas in black which contain tracer concentrations greater than twice the initial value are defined as “patches”.

The disagreement is smaller between models with different cell sizes than between models with different time-steps (Figures 3.4 & 3.5). In Section 3.1, it was concluded that the shorter time-steps yield a less accurate model output in the case of an analytical flow. However, here, patch mass is not always proportional to time-step. For example, in the South Pacific gyre, the  $T_{dt=180days}$  exhibits less accumulation than  $T_{dt=5days}$  but more than  $T_{dt=20days}$ . Unfortunately, model output cannot be rigorously benchmarked against observations of pollution concentrations because of the difficulty of recreating a realistic input scenario (Jambeck et al., 2015). However, there is a clear difference between Markov Chain model representations using different time-steps, yet much more consistency between models of different cell sizes.

How grid resolution affects model output differs to the way resolution affects OGCMs. Finer cell sizes do not necessarily produce more accurate representations. Instead, regardless of the cell size used, GDP trajectories have sampled the full-flow field including sub-mesoscale turbulence. A primitive-equation model would need sufficiently fine resolution to represent this motion. Here, a  $2^\circ$  grid recreates the same global accumulation pattern as a  $\frac{1}{2}^\circ$  grid.

While it is not reasonable to discount the presence of artificial dispersion in this experiment, its role is unclear. Moreover, the transition matrix parameters are not the only influence on artificial dispersion. As explained in Chapter 2.3, drifter trajectories are split into sub-

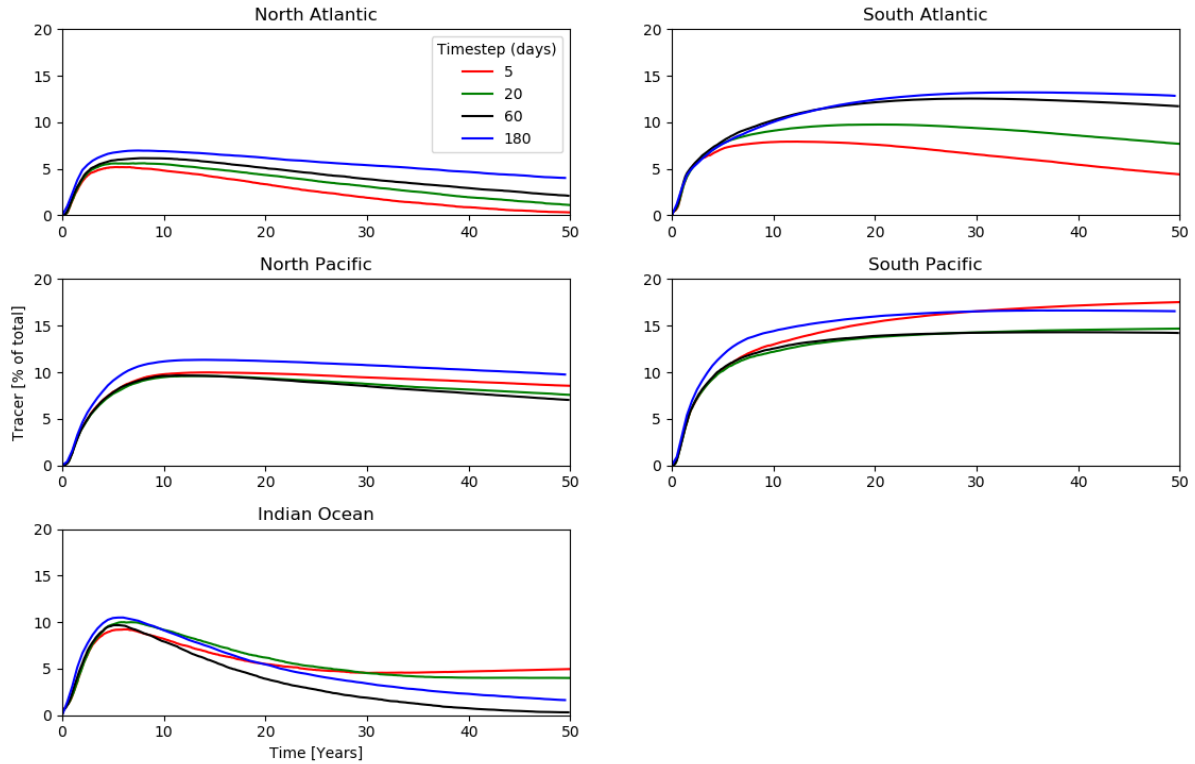


Figure 3.4: Sensitivity analysis of GDP-derived transition matrix parameters on gyre accumulation over 50 years. The evolution of patch sizes in each ocean basin is shown for the full range of  $dt$  values and a fixed  $dx = 1^\circ$ .

trajectories. Using a longer time-step reduces the number of sub-trajectories that can be extracted from a trajectory. The reduction is 40% from a 5-day to a 180-day matrix based on GDP trajectories. In this way, the time-step may reduce the quality of the transition matrix representation.

The effects of artificial dispersion and sub-trajectory availability may counter each other. To remove the influence of the latter, the sensitivity analysis is repeated for a trajectory dataset with an order of magnitude more trajectories than the GDP. The GlobCurrent experiment described in Chapter 2.2.1 is used here to provide a larger dataset and to construct transition matrices. The resulting models produce largely similar patch evolutions for the same range of  $dt$ , and to the GDP-based study in Figure 3.4 (Figure 3.6). The exceptions are the South Pacific and Indian Ocean. Using smaller  $dt$  sees more tracer accumulate in the South Pacific at the expense of the Indian Ocean. In a similar high-resolution global release experiment, Maes et al. (2018) showed that these two gyres are strongly connected to each other. Here, the

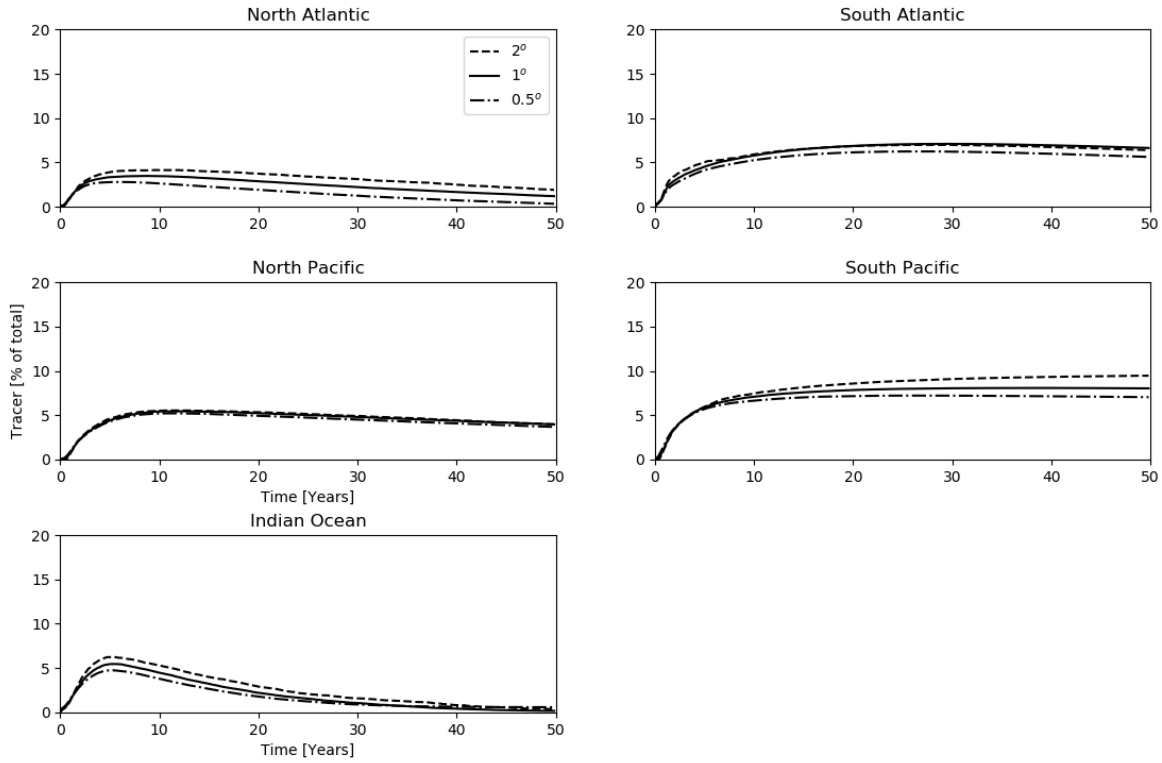


Figure 3.5: Sensitivity analysis of transition matrix parameters on gyre accumulation over 50 years. The evolution of patches for a range of  $dx$  values and a fixed  $dt = 60$  days.

leaky nature of the Indian Ocean is seemingly enhanced in the  $T_{dt=5days}$ . This may indicate that artificial dispersion, more prominent with smaller timesteps, is influencing regions which are considered leaky on large time-scales.

Unlike for the analytical flow in Chapter 3.1, it is not obvious that a particular parameter choice performs best at simulating oceanographic features. One reason may be that reduced number of sub-trajectories, and resulting loss in dataset coverage, counters the reduced artificial dispersion when using larger timesteps. Over the time-scale of 50 years, accumulation by surface currents appears to dominate over the divergent nature of artificial dispersion. For sub-surface flows, to be studied in Chapter 5, this effect may be even more pronounced due to the weak advective regime at depth. Therefore it is not possible to choose an optimal time-step. Instead, throughout the thesis, results from a range of time-steps will be compared.

The ability to represent a transport system with a Markov Chain model therefore depends on a range of factors: the availability of a dataset of particle trajectories, the accuracy of

the representation of the flow it samples, the transition matrix set-up parameters and the interaction of the corresponding artificial dispersion with the flow field. In Section 2.3, the Panama Isthmus represented a geographic barrier between oceans and is a clear but extreme example of artificial dispersion. The role of artificial dispersion across oceanographic transport barriers is beginning to be explored in the literature (Rypina et al., 2016). The next chapter applies the Markov Chain method to the dataset of observed trajectories in order to assess the physical, and non-physical, drivers of surface connectivity. How each of these factors affect Markov Chain models of the non-linear surface motion will be assessed.

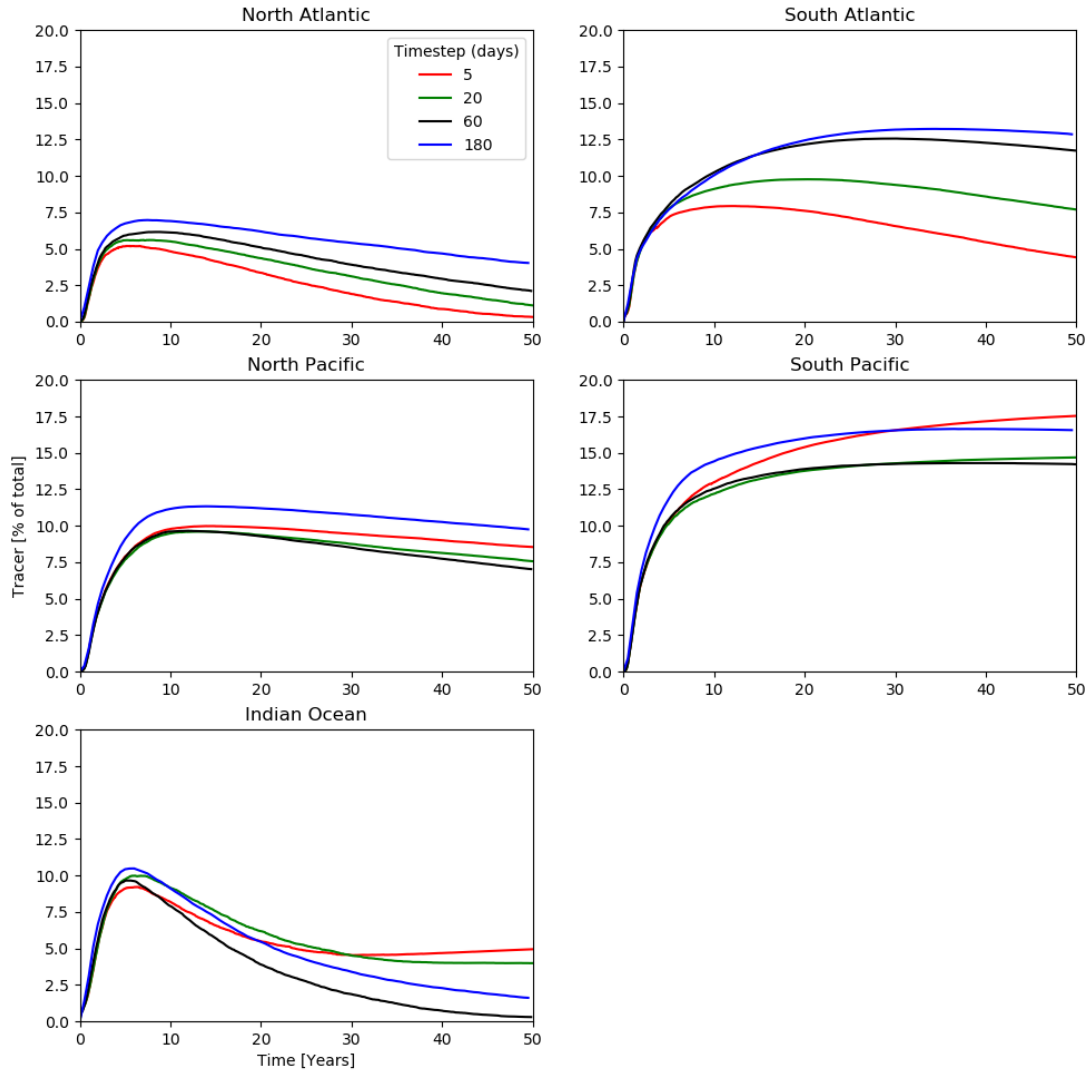


Figure 3.6: Sensitivity analysis of GlobCurrent-derived transition matrix parameters on gyre accumulation over 50 years. The evolution of patch sizes in each ocean basin for a range of  $dt$  values and a fixed  $dx = 1^\circ$ . As for Figure 3.4, but using the enhanced GlobCurrent-derived dataset using  $\sim 200,000$  particles.

## 3.4 Summary

In this section, the sensitivity of Markov Chain model output to the transition matrix parameters has been assessed for idealised trajectories and for the GDP drifter dataset. There appear to be conflicting results between the two case studies. For the idealised Stommel gyre flow, model output differs wildly to the streamlines. The Stommel case study highlights very clearly the contribution of artificial dispersion. The dispersive behaviour is amplified for shorter time-steps and larger cell-sizes. Cells contain a range of streamlines which do not cross or interact but summarising the transport into such cells causes these distinct pathways to be grouped together. As a result, tracer spreads parallel and perpendicular to the pathways used to construct the model. Therefore, the difference in output between models of different parameters arises because of the difference in groupings of distinct trajectories. The violation of the Markovian principle of consistency between solutions (Equation 2.3, Chapter 2.4) is attributed to this spurious behaviour.

The role of artificial dispersion is less clear when using real-ocean GDP trajectories. In an experiment to measure the accumulation of floating debris, traditionally a key application of this method, there is no clear relationship between accumulation and parameters. However, it is clear that parameters can cause regional-scale changes in accumulation. Artificial dispersion is relatively easy to diagnose with idealised trajectories but it is not trivial to detect how important it is for drifters, which already experience physical dispersion.

This chapter has explained the mechanism behind artificial dispersion in Markov Chain models but has not been able to quantify how important it is when using real ocean trajectories. Given that it is not clear which time-step is optimal at reproducing circulation, a range of time-steps will be used for the sake of transparency as well as to further test the sensitivity of results. Nonetheless, the Markov Chain model provides an efficient and intuitive means of regional analysis of trajectories, as demonstrated by the range of experiments which have been shown here. Further case-studies will help to understand whether the “leakiness” is detrimental to the Markov Chain model’s application in oceanography.

# Chapter 4

## Surface Connectivity

### 4.1 Introduction

The focus of this chapter is the surface transport within two distinct but key regions of inter-ocean exchange: the Agulhas Current System and the North Atlantic gyres. Connections and barriers to regional flow, between and within these regions, will be explicitly determined by applying GDP drifter-derived Markov Chain models.

Section 4.2 will focus on identifying the fate of tracer in the Agulhas Current system. In particular, the Markov Chain analysis will provide a fresh look at the extent of the region linked to the South Atlantic gyre and relative strength of the two exit routes from this region, the Leakage and the Return Current. The role of the two tributaries, the Mozambique and Madagascar Currents, along with other transport into the Agulhas Current, will be discussed. Section 4.3 applies connectivity analysis to the North Atlantic ocean, in particular to transport between the sub-tropical and sub-polar gyres. Analysis of GDP drifters will help identify the cross-frontal motion, both pole-ward and equator-ward, throughout the North Atlantic basin.

Both the Agulhas Current and the Gulf Stream are western boundary currents which largely flow along contours of sea-surface height. Unstable meandering and the resulting formation of geostrophic eddies play a key role in transport perpendicular to the currents. This chapter will

also attempt to resolve the role of geostrophy as both a connector and barrier between regions. This will be achieved by creating geostrophic Markov Chain models and comparing output to the full-field models.

This chapter begins with Section 4.2 describing transition matrices which isolate the geostrophic contribution to surface transport. Sections 4.3 and 4.4 focus on the Agulhas Current System and the North Atlantic transport barrier respectively. In both regions, connectivity maps are produced. These highlight barriers and connections based on the vast number of available trajectories. Alongside this, geostrophic Markov Chain models will determine the contribution of geostrophic currents and eddies to inter-basin transport.

## 4.2 ADT-following drifters

This section describes a novel methodology to isolate a key physical driver of surface transport, using observed drifter trajectories. The surface flow field is driven by the geostrophic balance, Ekman transport, Stokes drift and, for floating objects, wind-slip. Here, the geostrophic component will be extracted in order to assess its role in defining transport barriers and key inter-ocean currents on a regional scale. Maximenko et al. (2009) provide an overview of extracting geostrophic velocities from three different methods: altimetry, drifter readings of absolute dynamic topography (ADT) and theoretical models. Poulain et al. (2012) estimated geostrophic velocities by applying a low-pass filter to drifter-derived velocities. However, the impact of the geostrophic component on inter-basin connectivity has not been explicitly studied. This section proposes extracting a sub-set of trajectories which flow within a narrow ADT range. This sub-set will be used to approximate geostrophic flow.

In a purely geostrophic flow, ADT contours and Coriolis force determine the streamlines of flow. In the real ocean, flow is non-stationary and a snapshot of the ADT field is used to determine the velocity field at the particular time. Sokolov and Rintoul (2007) found ACC fronts, defined by high values of altimetry-derived ADT gradient, are aligned with ADT contours. Such analysis has not been performed, to the author's knowledge, on the Agulhas or Gulf Stream

regions. However, Killworth and Hughes (2002) found ADT contours approximate streamlines in equivalent-barotropic flows, using the ACC as an example.

In the real ocean, ADT provides an approximation to regional-scale flow which by definition omits non-geostrophic and sub-scale processes. Daily ADT fields from 1993 onwards are provided by the Copernicus Marine Environment Service (CMEMS) (Merged-Gridded Absolute Dynamic Topography heights in Near Real-Time (MADT-H NRT))<sup>1</sup>. These are the product of two fields: the mean sea level anomaly (MSLA) with respect to the geoid, and the geoid height. The MSLA is derived from altimeter measurements, with an error between 0.1 m and 0.2 m in the western boundary currents and less than 0.05 m throughout the majority of the rest of the ocean surface (Figure 4.1). The geoid is modelled at 100 km resolution to a vertical accuracy of 1cm (Mulet et al., 2012). Although the effective resolution of MSLA is  $2^\circ$  because of the overlap in satellite orbits, the ADT product is provided at  $\frac{1}{4}^\circ$  resolution.

Here, a method is proposed to extract drifter sub-trajectories which have a tendency to flow along contours of ADT. As will be discussed in this section, observed ADT contours may not be exactly equivalent to geostrophic streamlines, the term “geostrophic” will be used to describe the ADT-following trajectories for the sake of brevity. The main advantage of this method is to create Markov Chain models that are based on observed trajectories, offering an alternative to the altimeter-derived virtual trajectories. First, each drifter coordinate reading is mapped to an ADT value using a 2D linear interpolation. No temporal interpolation is performed between the daily ADT maps and 6-hourly drifter readings because the shortest time-period of interest here is 5 days.

Next, a sub-set of drifter sub-trajectories, of length  $dt$ , which remain within a narrow range of ADT values is extracted. To ensure the selected sub-trajectories are predominately driven by geostrophy, it is necessary to choose a suitably narrow ADT range. The threshold chosen here is a 0.2m difference between the beginning and end of a sub-trajectory of length  $dt$ . This threshold,  $dADT = 0.2m$  per time-step, is chosen because it corresponds to the maximum error in the MSLA product. Using a lower value may include erroneous measurements and

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<sup>1</sup>[www.marine.copernicus.eu/faq/ssaltoduacs-integrated-sealevel-anomalies-products-changes-updates](http://www.marine.copernicus.eu/faq/ssaltoduacs-integrated-sealevel-anomalies-products-changes-updates)

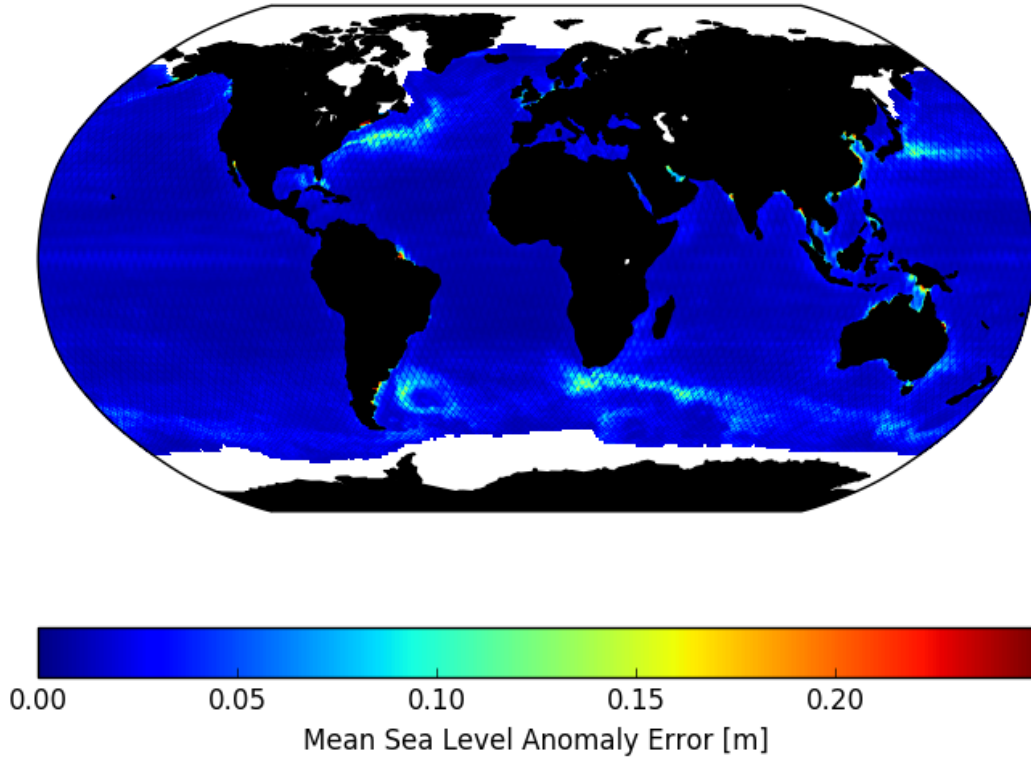


Figure 4.1: Time-averaged Mean Sea Level Anomaly error from daily  $\frac{1}{4}^\circ$  fields over the period 1993-2017. MSLA error fields are provided by the Copernicus Marine Environment Monitoring Service (CMEMS).

potentially cause the mis-identification of geostrophic sub-trajectories as non-geostrophic and vice versa. Specifically, an ADT trend-line is calculated along the length of each sub-trajectory. Using a trend-line smooths the ADT signal and thus reduces the error noise. In summary, a “geostrophic” sub-trajectory is one whose temporal ADT gradient,  $m$ , obeys the following:

$$m < \frac{dADT}{dt}. \quad (4.1)$$

An ADT range threshold is used instead of a specific ADT value because idealised geostrophic streamline values are not conserved in a non-stationary ocean. Instead, observed ADT streamlines are specific to a time-step and will shift in time. As a result, the ADT signal experienced by drifters is often highly variable and contains across-streamline velocity components. This

is also due in part to errors in the ADT fields provided. Due to the relatively small error in the geoid measurement, it is assumed that the MSLA error contributes fully to the total error. The variability is also due to the presence of ageostrophic motion, which is evident in the ADT signal sampled by three typical pathways in the Agulhas Current System (Figure 4.2). Within the core of the current, rapid changes in ADT are observed and the time series varies during different phases of the system's stability. Moreover, the identification of so-called geostrophic trajectories varies with the time-step considered. For example, a 180-day sub-trajectory may be classified as geostrophic but within that time there may be several shorter non-geostrophic sub-trajectories.

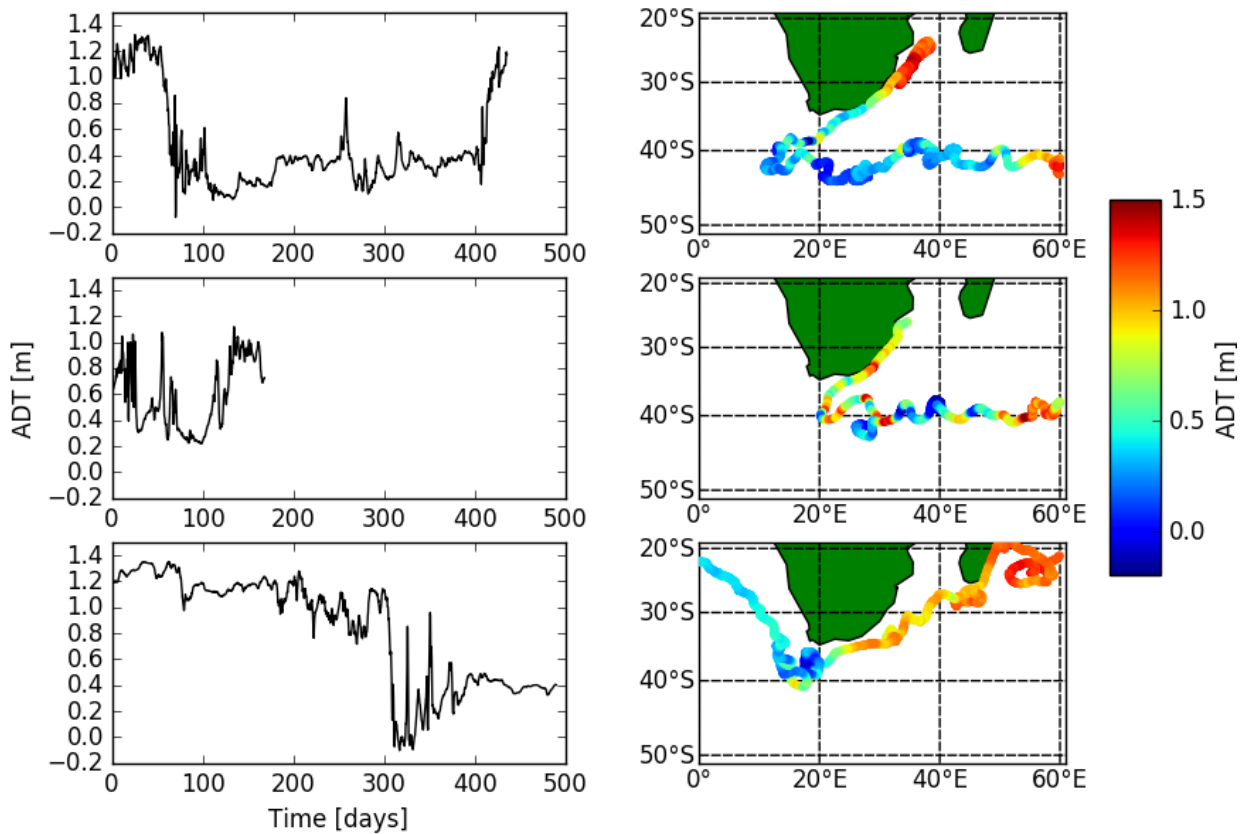


Figure 4.2: Absolute Dynamic Topography experienced by drifters in the Agulhas System. Examples of individual drifter pathways: an unstable westward-shifted Retroreflection (top), a stable central Retroreflection (middle) and an Agulhas Ring (bottom). ADT values are interpolated to 6-hourly readings, to correspond to the drifter coordinate readings.

A key consideration in defining geostrophic trajectories is therefore the time period of interest. Figure 4.3 compares the initial and final ADT values of every sub-trajectory for the range of

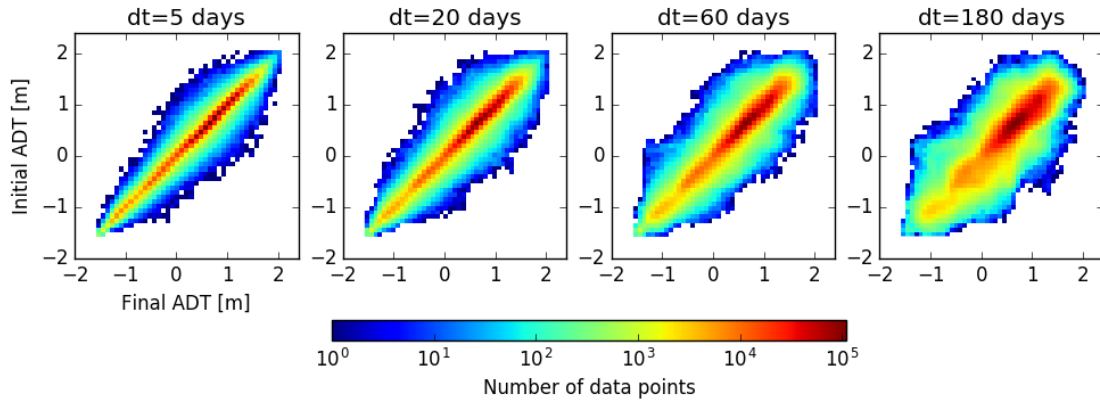


Figure 4.3: ADT connectivity matrices depicting the change in ADT at beginning and end of all GDP drifter sub-trajectories. Drifter sub-trajectories from 1993 onwards are used to correspond with the available ADT fields. The total number of coordinate data points available for each sub-trajectory length are shown in Table 4.1.

$dt$  values used. The before-and-after comparison is shown with connectivity matrices, which are not to be confused with transition matrices. An identity matrix would indicate that all drifters remain on the same ADT contour. Deviation from the diagonal therefore highlights how far drifter transport is from a purely geostrophic state on each time scale. For all time-steps, the majority of sub-trajectories obey Equation 4.1 but there is a decline as time-step increases (Table 4.1). The longer the drift time, the more likely drifters are to deviate over 0.2m from their initial ADT value. Short time-step matrices are less diagonal in the positive ADT regions, showing that drifters are more likely to experience a sharp change in ADT value in these regions. Positive ADT regions are typically those with the sharpest ADT gradient and the most rapid advection, such as western boundary currents. An example of this is in the Agulhas Leakage pathways, where the signal is variable in the high-ADT Agulhas Current but smoother in the low-ADT South Atlantic gyre region.

Generally, the locations where the majority of sub-trajectories obey Equation 4.1 are in the ocean gyre interiors (Figure 4.4). Over longer periods of time, drifters are more likely to experience a great range of ADT if they begin in rapid currents. Meanwhile, those beginning in inner gyre regions are less likely to be entrained into higher-ADT regions. On longer timescales, a wider area contains more sub-trajectories which obey Equation 4.1.

An important caveat to this method is that ageostrophic motions sometimes occur in the same direction as ADT contours. Here, the ADT field is compared to the Stokes drift and the 15 m Ekman transport, which are both key drivers of surface transport (Lumpkin and Pazos (2007), Figure 4.5). Stokes drift is available from WaveWatch3 hindcast runs, while Ekman velocity fields are from the GlobCurrent database<sup>2</sup>. In the Agulhas Current, Stokes drift occurs between  $90^\circ$  and  $180^\circ$  to the flow direction while the Ekman transport is negligible along the current. In the Return Current, Stokes drift is directed in the same direction as the current, eastward, but the average flow field does not meander sharply like the ADT field. The Ekman flow is consistently northeastward south of  $40^\circ\text{S}$ . In the southeastern corner of the South Atlantic gyre, Ekman flow occurs in the same direction as the advection of Agulhas Rings, northeastward. However, material transport in the Agulhas rings predominantly experiences anticyclonic motion and so will occur regardless of Ekman transport.

In the North Atlantic gyre, the Ekman field is mostly perpendicular to ADT contours (Figure 4.5). Along the Gulf Stream, Stokes flow is directed eastward but, much like in the Agulhas Return Current, does not experience sharp meandering.

Although time-averaged fields show encouraging results, they do not indicate how many geostrophic trajectories are caused by non-geostrophic motion. Instead, this section has focused on the differences between geographical patterns on annual-mean fields. In the absence of an equivalent method of extracting Ekman and Stokes sub-trajectories as in Equation 4.1, it is not trivial to compare how many sub-trajectories would be classed as geostrophic, Ekman-dominated and Stokes-dominated. It is possible to compare to virtual particles run solely on Ekman or Stokes flow fields, but such a method is fundamentally different to the GDP-based study performed here.

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<sup>2</sup>Stokes Drift data: [www.datastore.cls.fr/catalogues/wavewatch-3-global-best-estimate](http://www.datastore.cls.fr/catalogues/wavewatch-3-global-best-estimate). Ekman-15 m data: [www.globcurrent.ifremer.fr/products-data/products-overview](http://www.globcurrent.ifremer.fr/products-data/products-overview)

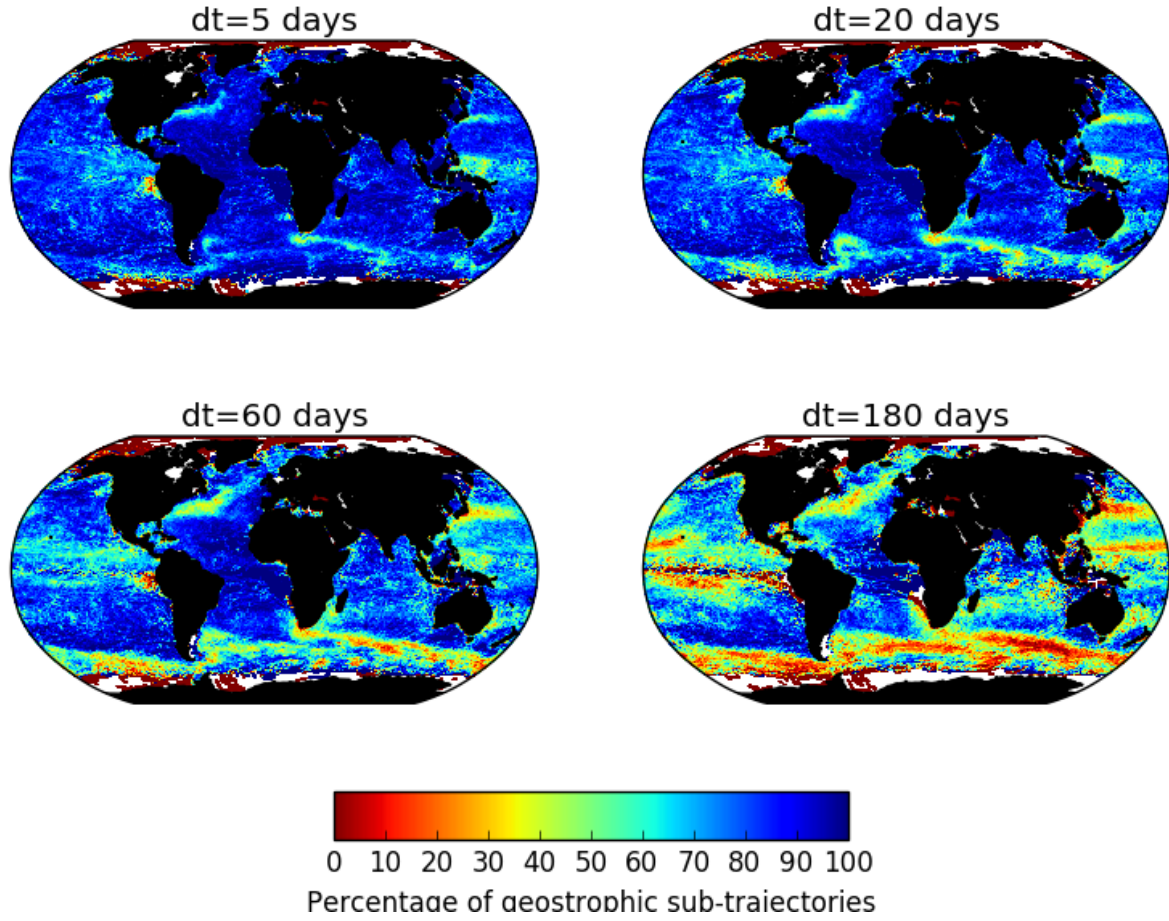


Figure 4.4: Global distribution of geostrophic drifters. A map of proportion of sub-trajectories, beginning in each  $1^\circ$  grid cell, whose ADT values stay within 0.2 m over a  $dt$ -day time scale. White areas contain no ADT readings which correspond to the initial or final sub-trajectories readings.

Another caveat is that the drifter dataset is composed of drifters which experience different forcings, namely the drogued and undrogued types which were introduced in Chapter 2.2. Undrogued drifters are more susceptible to wind-slip and experience the surface Ekman field, whereas drogued drifters are shielded from wind-slip and experience a deeper Ekman field (Poulain et al., 2012). It may be expected that undrogued drifters are not in geostrophic balance, or less so than their drogued counterparts, and that fewer undrogued drifters would obey Equation 4.1. The proportion which does obey Equation 4.1 is less than the proportion for the drogued sub-set, for all time-steps, but the difference is less than 5% in all cases (Table 4.1). The difference between the two datasets is shown in their corresponding connectivity matrices (Figure 4.6). Only on seasonal timescales and longer, a minority of undrogued drifters

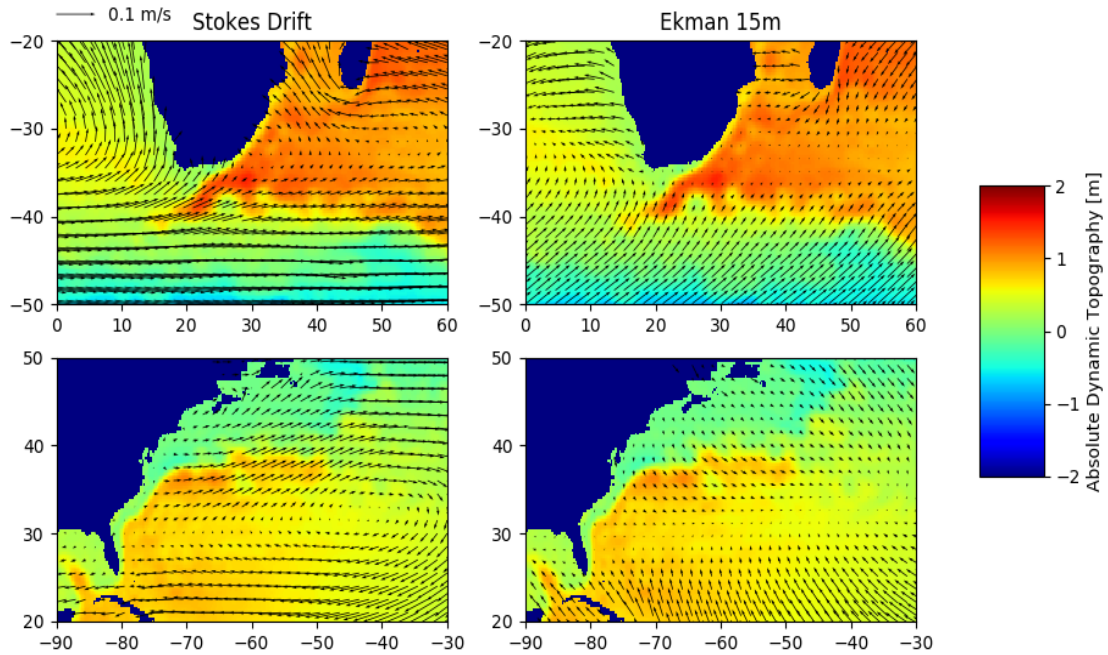


Figure 4.5: Alignment of nongeostrophic and geostrophic velocity fields. (left) Stokes drift and (right) 15 m Ekman velocity vectors are plotted over AVISO absolute dynamic topography. The Stokes drift  $\frac{1}{2}^\circ$  field is extracted from European Centre for MediumRange Weather Forecasting (ECMWF) (Dee et al., 2011). The Ekman transport  $\frac{1}{4}^\circ$  field is extracted from the GlobCurrent data set. Arrows are plotted every  $1.5^\circ$  for clarity. All fields are annual averages from 2014.

in negative ADT regions experience a larger change in ADT values. Therefore, is it not deemed necessary to restrict the analysis to drogued drifters alone.

Some geostrophic trajectories may also be mis-identified if erroneous ADT values dominate the signal on the scale of the time-step, which is more likely for the shorter time-steps. Mis-identification would effectively cause a random sampling of the total drifter dataset, which would approximate full-field flow as opposed to extracting geostrophic flow. As a result of greater mis-identification when using shorter time-steps, geostrophic Markov Chain model results for  $dt = 5$  days may approach the full-field results. In a region where there is expected to be weakened geostrophic transport, geostrophic and full-field models will differ. If there is no clear difference between full-field and geostrophic transport models in such regions, it may be a symptom

| %         | dt = 5 days | dt = 20 days | dt = 60 days | dt = 180 days |
|-----------|-------------|--------------|--------------|---------------|
| All       | 98.2        | 94.7         | 87.9         | 72.4          |
| Drogued   | 98.6        | 95.9         | 89.7         | 76.0          |
| Undrogued | 97.9        | 93.8         | 86.9         | 71.6          |

Table 4.1: Percentage of global GDP drifter sub-trajectories which obey the geostrophic condition defined by Equation 4.1.

of mis-identification. In the following sections, mis-identification of trajectories is taken into account when applying the technique to the two case-study regions.

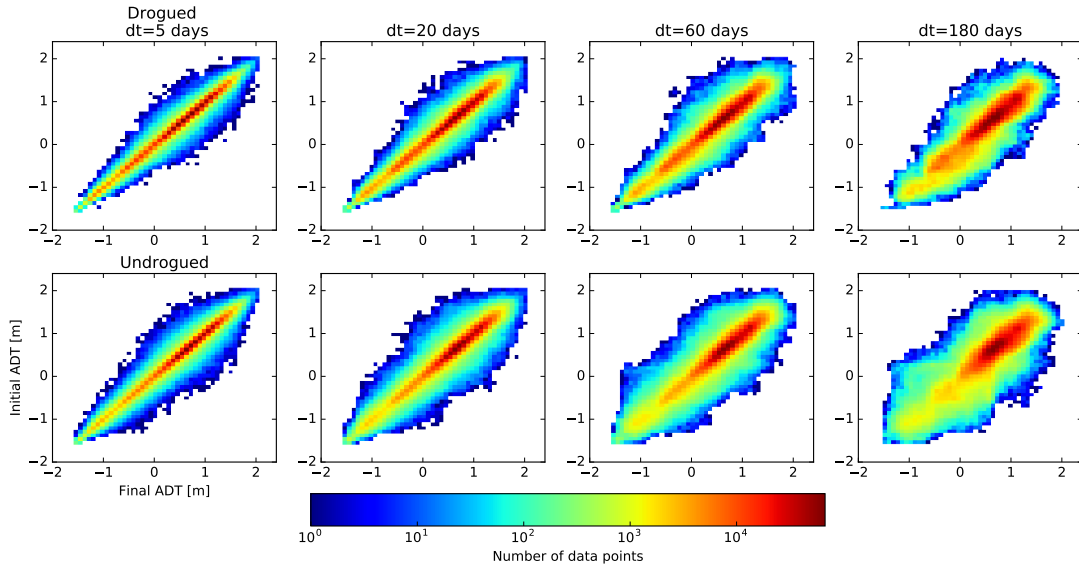


Figure 4.6: As for Figure 4.3, but for drogued and undrogued subsets of GDP data.

### 4.3 Agulhas Current System

Flow from the Agulhas Current splits into two pathways at the retroflection point: (1) north-westward Agulhas Leakage into the South Atlantic through Agulhas rings and other coherent structures and (2) the eastward Agulhas Return Current into the Indian Ocean. 25% of drifters which are advected by the Agulhas Current leave through Agulhas leakage and the remainder flow into the Return Current (Van Sebille et al., 2009). Agulhas rings are transient features, of which 3-6 are formed annually when the Retroflection reaches an unstable state (Boebel et al., 2003a). As they transport water northwestward, away from the Retroflection and into an area

rich in submesoscale activity, the rings decay before reaching the South Atlantic subtropical gyre (van Sebille et al., 2010). Despite the decay of these coherent structures, which initially carry water and tracer into the Atlantic, transport continues northward. R  hs et al. (2013) found Agulhas-to-Gulf Stream transport time scales to be roughly a decade.

To assess the capabilities of the geostrophic Markov Chain model, the full-field and geostrophic models are used to track a tracer release within the Agulhas Current at (31  E, 33  S), upstream of the average position of the Retroflexion (Beal et al., 2011). The tracer is released at the start only and the Markov Chain models run for 5 years. Agulhas Leakage is defined as tracer which flows northwest of the Good Hope line. The Good Hope line defines a fixed mooring array and the path of research cruises designed to monitor exchanges between the Indian and Southern Atlantic Oceans (Ansorge et al. (2005) & Figure 4.7). Specifically, here, the northern section of the line is used. Meanwhile, Return Current flow is defined as tracer which flows east of 40  E.

In both the full-field and geostrophic models, the tracer cloud grows perpendicular to the path of the Agulhas Current, as well as being advected along it (Figure 4.7). Because this dispersion has occurred even after only one time-step, at 60 days for the  $T_{dt=60days}$  matrix, artificial dispersion cannot be the reason behind such transport. Instead, tracer is dispersed into the eddy-rich recirculation region (Lutjeharms and Ansorge, 2001). After 60 days, the tracer has already bifurcated at the Retroflexion, although two distinct clouds have not formed. Although the Return Current is observed to be 60-80 km wide and only meanders up to 3   before 40  E (Boebel et al. (2003b)), the tracer cloud consistently covers the region between the source and Return Current region past 40  E. Again, this indicates the influence of the recirculation region.

Additionally, mesoscale eddies formed by the meandering of the Return Current may return tracer to the source region. For the Agulhas Leakage, although individual rings and filaments are narrow features ( $\sim 50$  km wide), they spread and decay throughout the South Atlantic. Hence, the time-averaged flow field shown here displays a broad cloud of tracer.

The proportion of tracer transported through the Leakage and Return Current in the full-field model are in agreement with observations (Figure 4.8). 18-25% is leaked into the South Atlantic

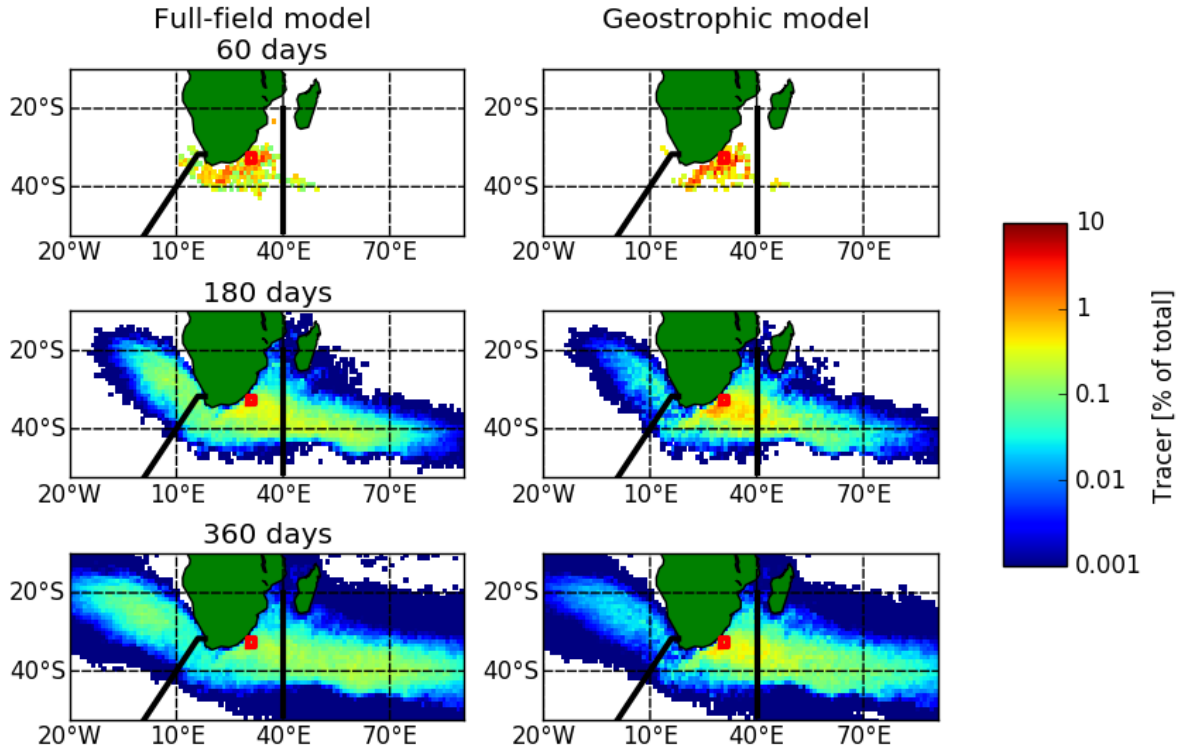


Figure 4.7: Distribution of tracer in the GDP-derived Markov Chain model of the Agulhas Current system. Flow exits the region via the Agulhas Retroflexion Current (east of 40°E) or Agulhas Leakage (west of the Good Hope line, other black line). Tracer is released from the red square at [31°E, 33°S] and transported by the full-field (left column) and the geostrophic (right) versions of  $T_{dt=60days}$ .

compared to 25% of GDP drifters. 55-61% of tracer leaves the region via the Retroflexion. The remainder recirculates with the Agulhas recirculation region, between the Return Current, Madagascar and the Agulhas Current. Within the 5-year timescale, tracer re-enters the source regions from the Return Current and is then redistributed through both the Leakage and Return Current. Tracer in the latter begins decreasing by  $\sim 10\%$  after 2.5 years, followed by a rise in the Leakage.

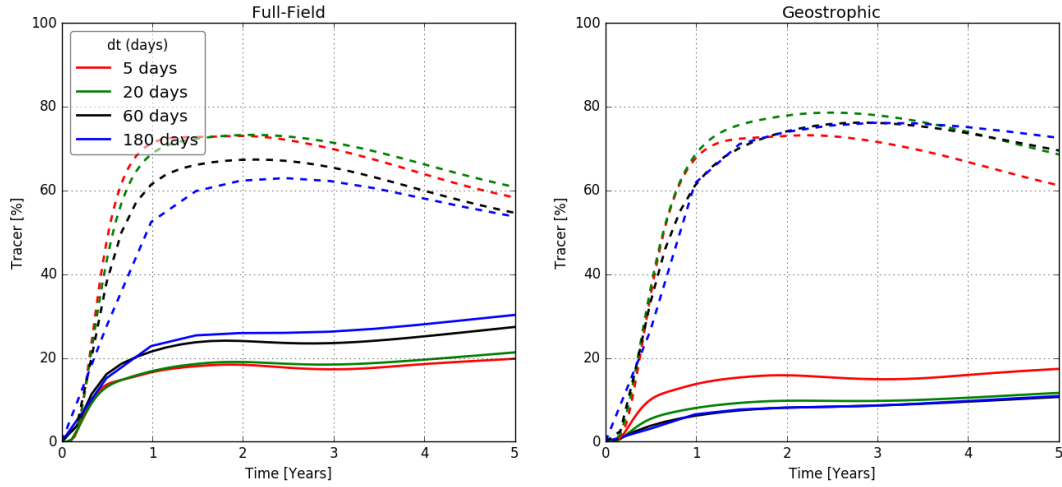


Figure 4.8: Proportion of tracer transport in the Agulhas Current system which passes through Agulhas leakage (bold) and Return Current (dashed). The boundaries are shown in Figure 4.7. For a range of time-steps, (a) full-field transport and (b) geostrophic Markov Chain model transport. Van Sebille et al. (2009) found 25% of GDP drifters leave via the Agulhas leakage.

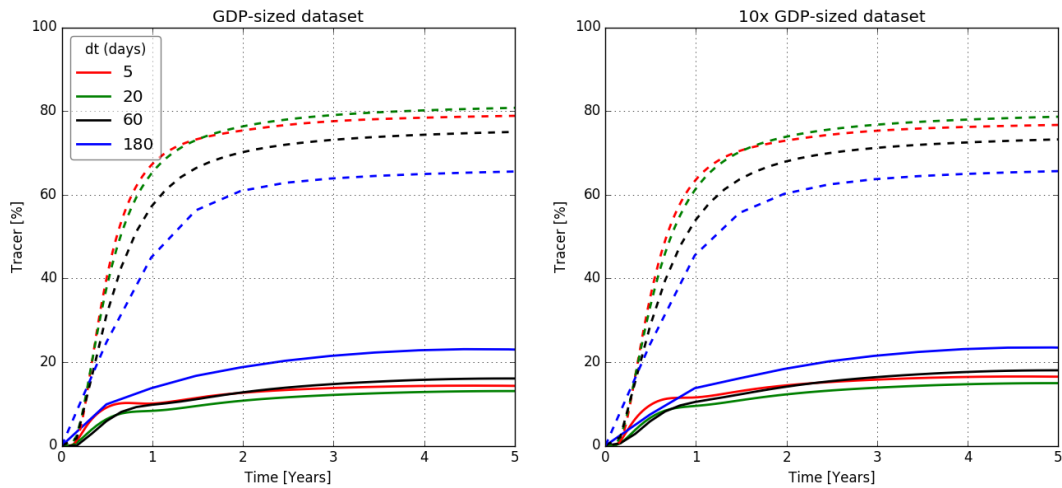


Figure 4.9: Markov Chain models derived from two GlobCurrent-based particle datasets which differ in the number of drifters deployed. Top: a GDP-sized dataset of 22,560 particles deployed on a  $1.5^\circ$  grid. Bottom: A larger dataset of 201,040 particles deployed on a  $\frac{1}{2}^\circ$  grid. Both datasets are advected on the  $\frac{1}{4}^\circ$  geostrophic and Ekman surface velocity fields from 2012 to 2013 inclusive.

The size of the Agulhas Leakage tracer cloud and the corresponding proportion within the geostrophic model are less than the full-field model (Figure 4.7). The tracer distribution shifts from 21.1-29% to 10.3-17.2% in the Agulhas Leakage and from 55.8-61.4% to 61.6-74.1% in the Return Current. This implies geostrophy drives tracer towards the Return Current. This result is seemingly counter-intuitive, given that Agulhas Leakage rings are geostrophic features. However, rings typically undergo a rapid decrease in ADT during their decay (van Sebille et al., 2010). In fact, one-third of observed rings do not escape the Cape Basin (Schouten et al., 2000). Drifters being transported in Agulhas rings will initially maintain their ADT value but, as they travel into the South Atlantic gyre, their ADT value decreases. The number of geostrophic trajectories in the northwestern direction will therefore decrease, and the corresponding transport in the Markov Chain model will be reduced.

In Figure 4.8, the  $T_{dt=5days}$  geostrophic model results are notably different to the longer timestep models. Although the evolution and final proportions of tracer throughout the region are consistent between the  $T_{dt=20days}$ ,  $T_{dt=60days}$  and  $T_{dt=180days}$  model runs, the  $T_{dt=5days}$  run is more similar to the results for the full-field model. This is evidence that using shorter timestep trajectories can lead to a higher probability of mis-identifying geostrophic trajectories (Section 4.2). As mentioned in Section 4.2, mis-identification causes a random sample of sub-trajectories to be picked. This, in turn, is likely to depict the full-field flow. Therefore, the  $T_{dt=5days}$  geostrophic matrix is less able to represent geostrophic motion than the other matrices and behaves similarly to the full-field model.

To confirm that the size of the GDP dataset is sufficiently large to detect inter-ocean transport, the Agulhas Current experiments are repeated for transition matrices derived from the GlobCurrent experiments described in Section 2.2 (Figure 4.9). The combined Ekman and geostrophic velocity field for the surface is used to advect two particle sets. The first has a similar number of drifters to the GDP dataset. The second has an order of magnitude more. The resulting transport through exit regions are near identical to the GDP-based results and to each other. This result suggests that the coverage of the GDP dataset is suitable for constructing Markov Chain models in this region. Figure 4.9 also shows that the sensitivity of transport to the choice of time-step is still evident when using a larger trajectory dataset.

The GlobCurrent-based model also displays a weaker recirculation of Return Current flow (Figure 4.9). It is expected that the  $\frac{1}{4}^\circ$  resolution of the velocity fields does not resolve sub-mesoscale features present in the recirculation region. However, GlobCurrent is eddy-permitting. This suggests that the mesoscale processes detected by GlobCurrent do not contribute to the majority of transport away from the recirculation region. The comparison with GlobCurrent has not highlighted deficiencies in the GDP-based models but, instead, suggested that submesoscale processes are key to inter-ocean transport in the Agulhas Current System.

While the Agulhas Leakage is the pathway to the Atlantic basin, single-cell experiments do not show where the origin for this transport may be. To identify the origin points for inter-basin transport, Markov Chain models are repeated for every cell from  $[5^\circ\text{E}, 45^\circ\text{S}]$  to  $[50^\circ\text{E}, 20^\circ\text{S}]$ . The resulting maps for each transition matrix, labelled as connectivity maps, are shown in Figure 4.10. The maps highlight how much transport from individual cells crosses into the south Atlantic. For example, the area north-west of the Good Hope line is in the South Atlantic and tracer flows towards the Equator, so the cells are fully connected to the Atlantic. In contrast, cells in the Agulhas Current transport roughly a quarter of tracer past the Good Hope Line.

The transport in the geostrophic model suggested a preference for tracer to exit via the Return Current. Comparing the full-field and geostrophic connectivity maps highlights a region-wide difference between the two (Figure 4.10). In the full-field model, at least 20% more tracer is transported into the Agulhas Leakage from the region adjacent to the Agulhas Current. In this region, a notable proportion of tracer transport is still caused by geostrophic eddies, as shown by the geostrophic model. The southern border of this region is the time-averaged path of the Return Current. The region is more strongly connected after two years while the Agulhas Current remains strongly connected from the beginning, indicating again that pathways in the recirculation region are less direct.

The region in which Leakage tracer originates covers the north-eastern part of the domain, adjacent to and including the Agulhas Current. While the Agulhas Current acts as the main conduit for inter-basin transport, the region it connects is of course much wider and is clearly highlighted in the connectivity maps. The features and quantities are detected in all models

in Figure 4.11, regardless of timestep. However, the shorter time-step models appear to define a smoother transition between the Leakage-connected region and the surroundings. Artificial dispersion is expected to be more prominent between cells in these shorter time-step Markov Chain models. In a “leakier” model, boundaries between such features would be less abrupt. So, this is further evidence of the role of artificial dispersion in shorter time-steps.

The inhomogeneous nature of connectivity when using longer time-steps (see  $T_{dt=180days}$ , Figure 4.11) may also suggest an uneven sub-trajectory count within the domain. Cells which are strongly connected to the South Atlantic neighbour those which are less so. Varying tracer transport values may arise where there are fewer sub-trajectories, and thus less transport information. The number of sub-trajectories present in this domain is indeed larger for the shorter timesteps (Figure 4.12). The poorly sampled regions include the Benguela Current and upstream of the Agulhas Current. However, the regions where connectivity maps are patchy, such as the Recirculation region, show small difference in sub-trajectory availability between time-steps. Artificial dispersion, therefore, is causing the smoothing of connectivity between regions, as opposed to GDP transport information.

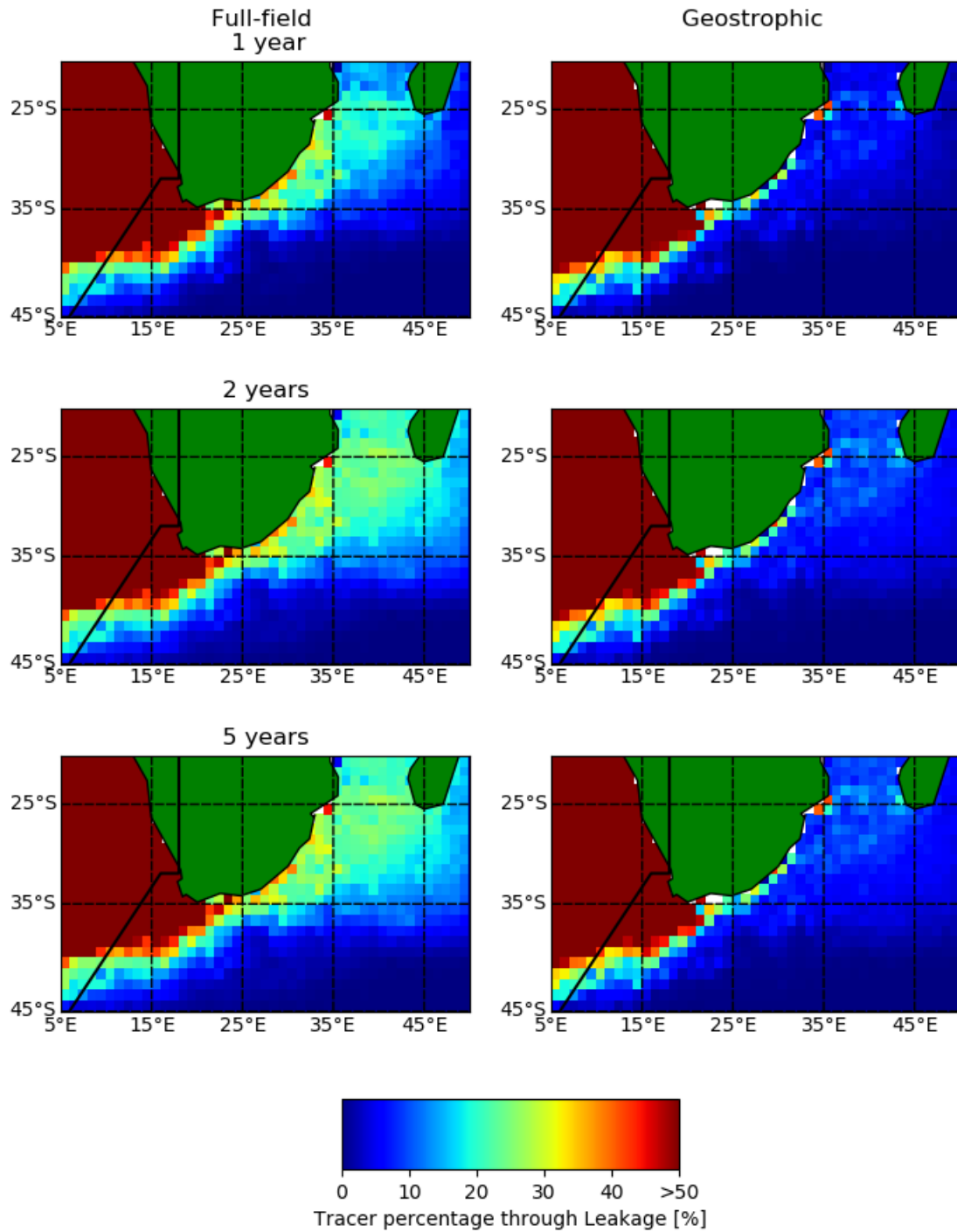


Figure 4.10: Agulhas Leakage connectivity map comparison for full-field and geostrophic flow. Tracer which crosses the Good Hope line (black) is defined as Leakage. A  $T_{dt=60days}$  transition matrix was used for both.

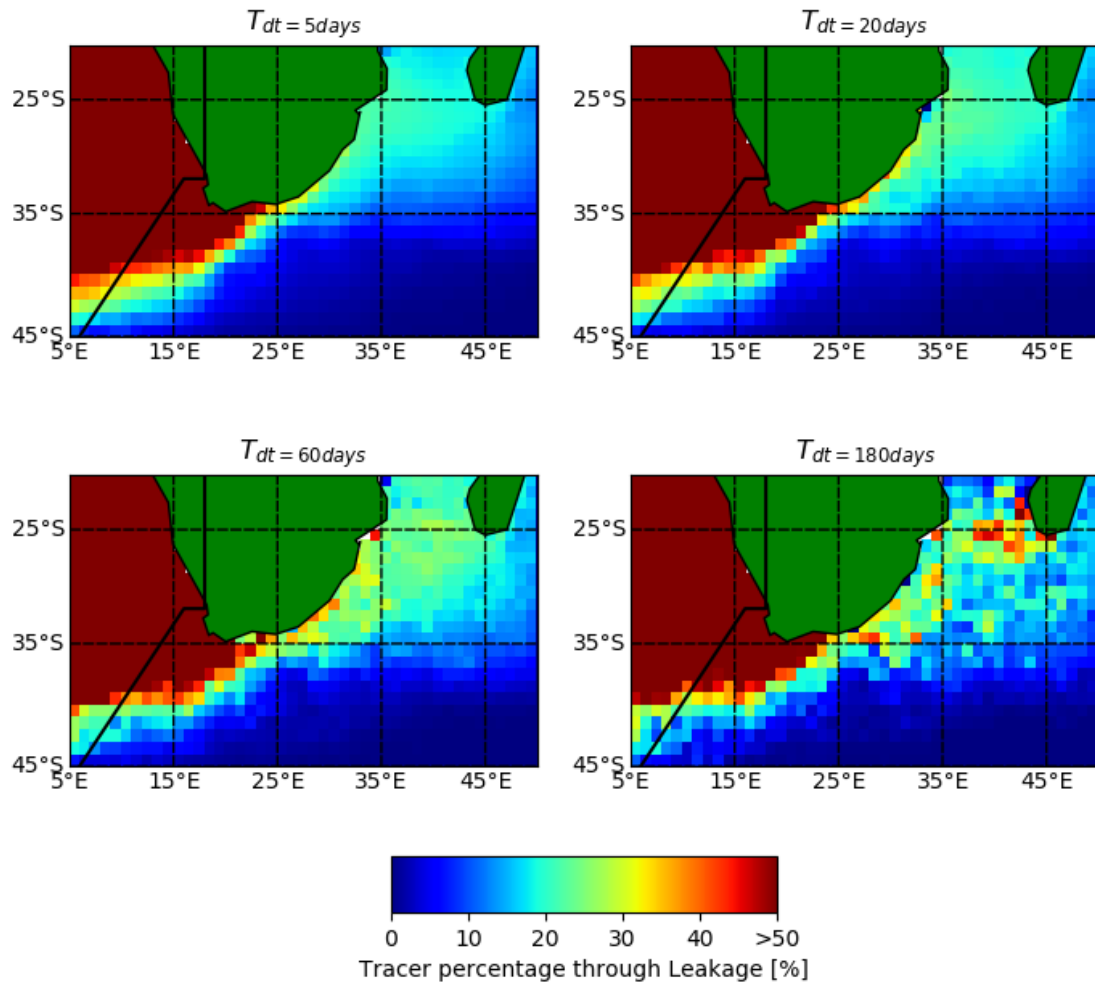


Figure 4.11: Agulhas Leakage 5-year connectivity maps for a range of GDP-based Markov Chain models. Tracer which crosses the Good Hope line (black) is defined as Leakage.

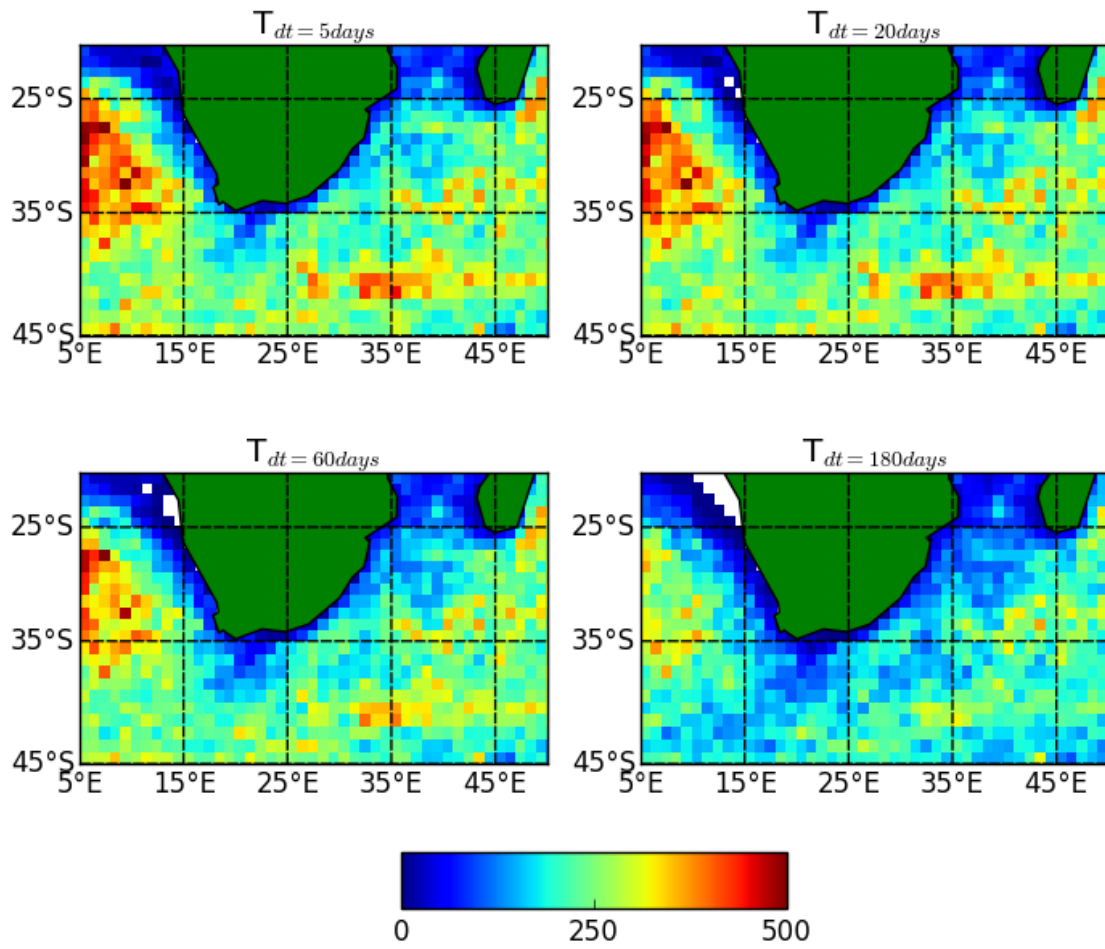


Figure 4.12: Number of  $dt$ -length GDP drifter sub-trajectories entering  $1^\circ$  cells.

## 4.4 North Atlantic Transport Barrier

Defining oceanographic fronts is not trivial. Fronts may be based on abrupt changes in either scalar value fields, such as temperature and salinity, or dynamical properties such as potential vorticity (Bower et al., 1985). Rypina et al. (2011) identified a front in the time-averaged velocity field derived from drifter measurements, which segregates sub-tropical cyclonic and sub-polar anti-cyclonic gyre streamlines. In their study, they first calculate the two-dimensional, time-mean drifter-based velocities on a  $1^\circ$  grid. From this, a discretised streamfunction was determined which clearly separated the sub-polar and sub-tropical gyres into anti-clockwise and clockwise flows respectively. A front between the two gyres was taken to be the zero-streamline which straddled both gyres. It is, however, easier to calculate transport proportions across a particular latitudinal line (Brambilla and Talley, 2006; R  hs et al., 2013). Previous studies focused only on tropical-to-polar transport across the Gulf Stream barrier (Hakkinen and Rhines, 2009). Here, cross-frontal transport analysis is extended to study the fate of tracer which originates in both gyres.

To understand cross-frontal connectivity throughout the North Atlantic, connectivity maps for the region are created similar to Section 4.3. First, tracer is released one cell at a time in a rectangular domain extending from ( $80^\circ\text{W}$ ,  $30^\circ\text{N}$ ) to ( $0^\circ\text{E}$ ,  $60^\circ\text{N}$ ). The model is run for 5 years, which is considerably longer than the time taken for drifters to travel from the Gulf Stream to the sub-polar gyre (Brambilla and Talley, 2006). For each cell release, the quantity of tracer which traverses the Rypina et al. (2011) front is determined. The domain experiment is repeated for the range of time-steps and for geostrophic transition matrices. As suggested by the Agulhas Current case study, the geostrophic  $T_{dt=5days}$  model is expected to approximate the full-field model results, while the longer time-step models should highlight differences between the full-field and geostrophic models (Figure 4.13).

In the full-field model, nearly all the tracer which originates south of the dynamical front remains in the subtropical gyre system after 5 years. The same is true for tracer which originates between the front and the coast of the northern U.S and Canada, west of  $50^\circ\text{W}$ , and which traverses the Gulf Stream into the sub-tropical gyre. Labrador Current tracer also ends up

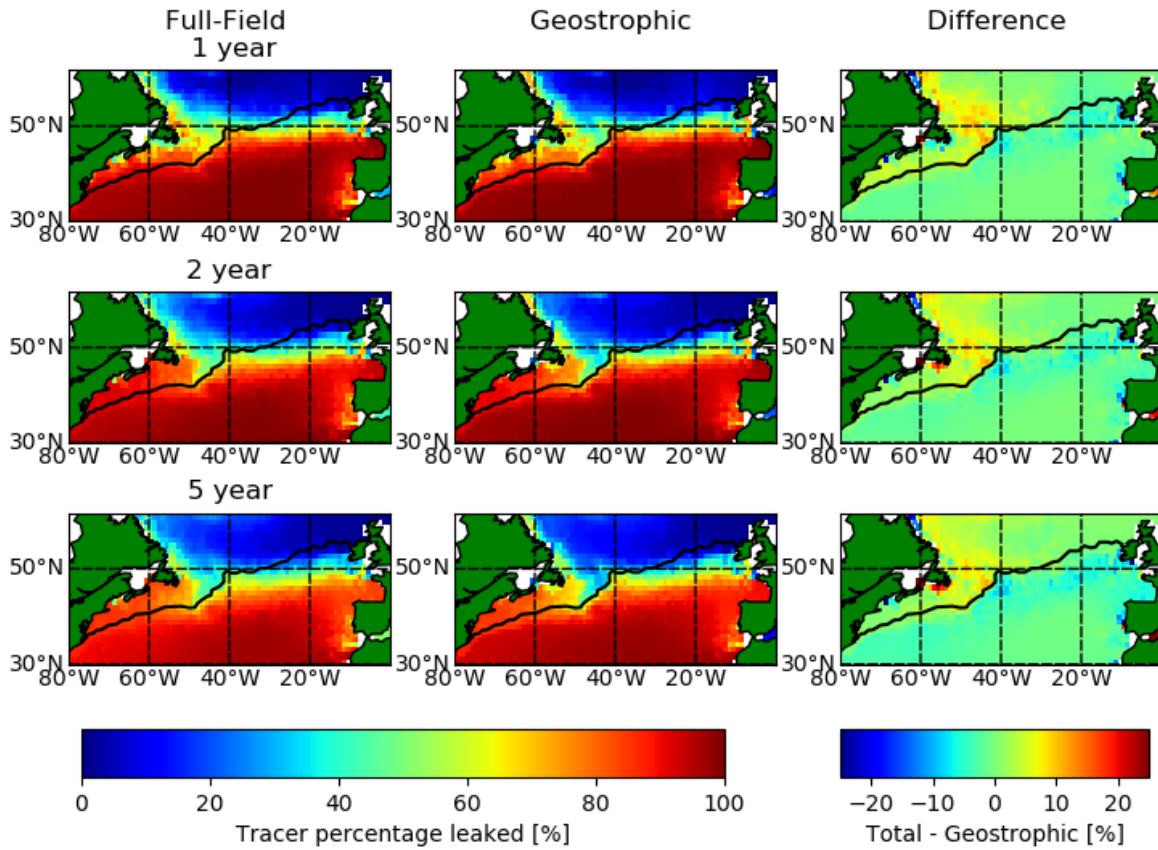


Figure 4.13: Cross-frontal leakage between sub-tropical and sub-polar gyres in the North Atlantic. Left: Full-field GDP-based Markov Chain model. Middle: Geostrophic GDP-based model. Right: The full-field transport minus geostrophic transport. For each cell, the quantity which remains south of the front, or crosses the front if beginning in the sub-polar gyre, is counted. The front is defined by Rypina et al. (2011) from a drifter-derived velocity field.

south of the front. Thus, the sub-polar region is connected to the sub-tropical gyre through the Labrador Current and west of  $50^\circ\text{W}$ . In the central North Atlantic around  $[45^\circ\text{N}, 45^\circ\text{W}]$ , the separation implied by the Markov Chain model analysis and the front are aligned. The separation is not rigorously defined here because the transition from no-northward tracer to no-southward tracer is several cells wide. To the east, the model-derived separation barrier is further south than the stagnation line, perhaps because the physical front is weaker here. There is little change in the size of this separation as tracer is advected beyond a year.

In the geostrophic model, the implied separation barrier largely follows the stagnation line to the same extent as in the full-field model. There are also clear differences between the full-field

and geostrophic field connectivity (Figure 4.13, right column). In the Labrador Current, 25% more tracer is transported equator-ward in the full-field than the geostrophic model. However, north of the Gulf Stream where the majority of tracer is transported pole-ward in the full-field model, the geostrophic model behaves similarly.

The tracer carried by the Labrador Current is brought to between 50°W and 40°W, where the Gulf Stream bifurcates and forms the North Atlantic Current. In the geostrophic model, the quantity of Labrador Current tracer crossing the front is reduced. This indicates that the cross-frontal processes around the North Atlantic Current are non-geostrophic in nature. However, cross-frontal transport for tracer originating on the northern flank of the Gulf Stream is equally strong in both models. This indicates that cross-frontal processes here are geostrophic in nature. Eddies which form from meandering instabilities of the Gulf Stream are the most likely processes causing geostrophic cross-frontal transport.

Crucially, the Markov Chain comparison between full-field and geostrophic models does not explicitly highlight where processes are contributing to cross-frontal transport. However, there is evidence which points towards the North Atlantic Current. At 50°W, the Gulf Stream deviates pole-ward sharply to transition into the North Atlantic Current (see front defined by Rypina et al. (2011) in Figure 4.13). This feature is also marked by time-averaged ADT (Hakkinen and Rhines, 2009), eddy diffusivity (Abernathey and Marshall, 2013) and the velocity field. North Atlantic eddies are considered, in model studies, to be strongly anisotropic, with the maximum intermediate-scale (<150 km) transport occurring perpendicular to the current. However, drifter-based calculations show weakened anisotropy at the bifurcation of the Gulf Stream at 50°W (Rypina et al., 2012). Meanwhile, mesoscale eddy diffusivities are similar either side of the Gulf Stream core and remain locally high along the North Atlantic Current extension (Abernathey and Marshall, 2013).

The area where the Gulf Stream transitions into the North Atlantic current, near 40W, plays a key role in the connectivity map (Figure 4.13). In the geostrophic model, the northern flank of the Gulf Stream is fully connected to the sub-tropical gyre while the Labrador Current is less well-connected. An equal amount of Labrador Current tracer recirculates in the sub-polar

gyre. However, tracer originating closer to the coastline and within the core of the current is more likely to cross the front than on the edge nearer to the Labrador Sea. This tracer may intersect flow from the north of the Gulf Stream, and shares the same fate, explaining the similar quantities of cross-frontal transport from each source region. The intersection occurs upstream of the northward deviation of the Gulf Stream, where it becomes the North Atlantic current. East of this, equator-ward transport is negligible. Therefore, cross-frontal transport is implied to take place at the transition of the Gulf Stream to the North Atlantic Current.

The barrier implied by the Markov Chain model experiment disagrees in places with the connections proposed by Rypina et al. (2011). In particular, Rypina et al. (2011) highlight the region between the Gulf Stream and US and Canadian coastlines as a source of sub-polar water. Here, in the analysis of observed GDP drifters, it is instead strongly linked to the sub-tropical gyre. It is important to note that the coverage of drifters onshore of the Gulf Stream core is over 500 readings per  $1^\circ$  cell, meaning there is at least 1 reading every  $20 \text{ km}^2$ . In the Agulhas Region, the coverage is of the same order of magnitude and is sufficient to capture the relative strength of Leakage and Return Currents as well as the sub-mesoscale dynamics. Rypina et al. (2011) used virtual trajectories run on drifter-derived background flow fields with altimeter-based mesoscale perturbations. Here, the North Atlantic front is extracted from their background velocity field. Thus, it is possible that differences in the two method's outputs arise from the treatment of sub-mesoscale features. Although the method presented here uses fewer trajectories, these trajectories capture sub-mesoscale variability.

Another point in favour of the new results is bathymetry. This region, along with the Labrador Current, are on the raised Grand Banks shelf region (Townsend et al., 2015). There is a steep gradient towards the deeper basin, where the Gulf Stream and the Labrador Sea are found. The sub-polar connectivity therefore also mirrors bathymetry. On the shelf, surface transport is on average towards the equator. The tracer originating between the Gulf Stream and the coastline may therefore also head south-westward, anti-parallel to the Gulf Stream.

An often-cited caveat of drifter studies is the coverage bias of drifters, which has been discussed in Chapter 2. Reverdin et al. (2003) compared drifters released north of the Gulf Stream and

north of the North Atlantic Current, finding each sub-set of drifters have different fates, with the majority in each ending up in the sub-tropical and sub-polar gyres respectively. At the time of their study, GPD drifter coverage was smaller than now and they cite deployment bias as a caveat to their findings. The analysis here uses additional drifter data (see Figure 2.2, Section 2.1) and may add evidence to their findings. Nonetheless, there remain discrepancies with methods based on unbiased virtual particle releases such as Rypina et al. (2011).

Here, it is shown that the fate of tracer originating in the Labrador Current and the northern flank of the Gulf Stream is the sub-tropical gyre. In terms of two-way transport, this would suggest a different, and new, picture of connectivity between the two regions. This conclusion is tested in two ways. First, the results may be somewhat dependent on the definition of the Rypina et al. (2011) boundary used here so the connectivity maps are recreated with the  $45^{\circ}\text{N}$  latitudinal line serving as the boundary between the sub-tropical and sub-polar gyres (Figure 4.14). The distance between the front and  $45^{\circ}\text{N}$  is up to  $15^{\circ}$  latitude yet the picture of regional connectivity is almost identical. The amount of tracer from the Labrador Current to cross the  $45^{\circ}\text{N}$  is, before 2 years, less than the amount to cross the Gulf Stream. This is presumably due to cross-frontal processes occurring further north than  $45^{\circ}\text{N}$ . The longer-term model output proves not to be very sensitive to the choice of boundary.

The second comparison is with an eigenvector decomposition of the transition matrix used to run the domain experiments. The eigenvectors of a transition matrix represent the regions of attraction in a domain, in this case the whole ocean. An interpretation of the eigenvector is that tracer released within the corresponding region is most likely to remain with this region. For example, Froyland et al. (2014) produced 5 eigenvectors of a drifter-based transition matrix, each representing one of the 5 ocean basins. Here, a larger number were computed to find the eigenvector which corresponds to the North Atlantic sub-tropical gyre, shown in Figure 4.15. The lower values in the eigenvector highlight regions which are less strongly attracted to the higher values, or which have some affinity to another regions represented by lower values. This is evident in the Labrador Current, Labrador Sea and along the North Atlantic front all share low eigenvector values. These correspond to areas where tracer in the full-field Markov Chain model is split between the two gyre systems. Moreover, the distinct values in the gyre centres

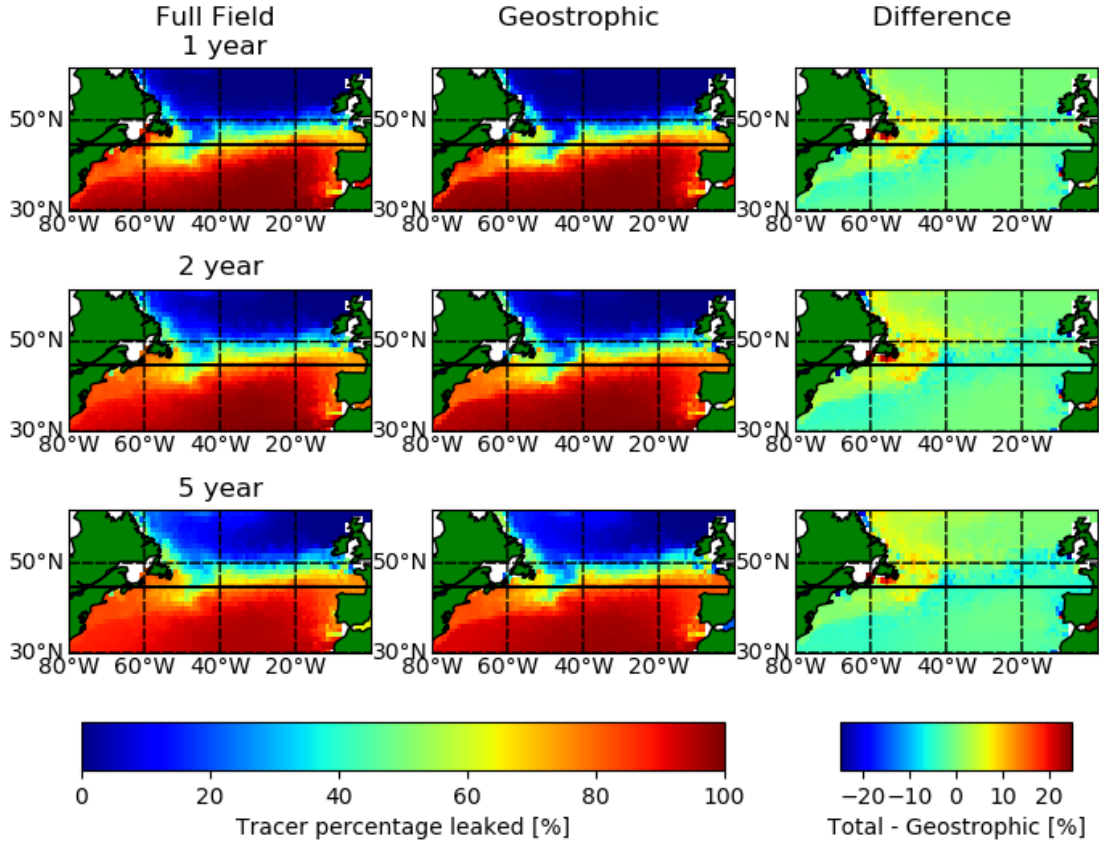


Figure 4.14: Cross-frontal leakage between sub-tropical and sub-polar gyres in the North Atlantic for full flow field and geostrophic flow. For each cell, the quantity which remains (or enters, if beginning north of the front) in the sub-tropical gyre is counted. The front (transport barrier) is the  $45^{\circ}\text{N}$  zonal line.

indicate that these areas belong to almost completely different regions of attraction. Given that both methods are based on the same GDP-derived transition matrices, it is expected that results are similar. The benefit of the Markov Chain approach is that it can depict connectivity for a given time, as in the 1,2 and 5 year connectivity maps shown here.

As discussed in the previous chapters, some cross-frontal transport is expected to be caused by artificial dispersion. Moreover, oceanographic fronts are typically less abrupt than the example of the Panama Isthmus (Section 2.4). The North Atlantic front oscillates around the fixed definition used here, but the seasonal range of latitudes remains below  $1^{\circ}$  from the average (Kelly et al., 1999). This is less than the cell size used in the Markov Chain model. In a time-averaged transport field, this motion of the front leads to increased artificial connectivity

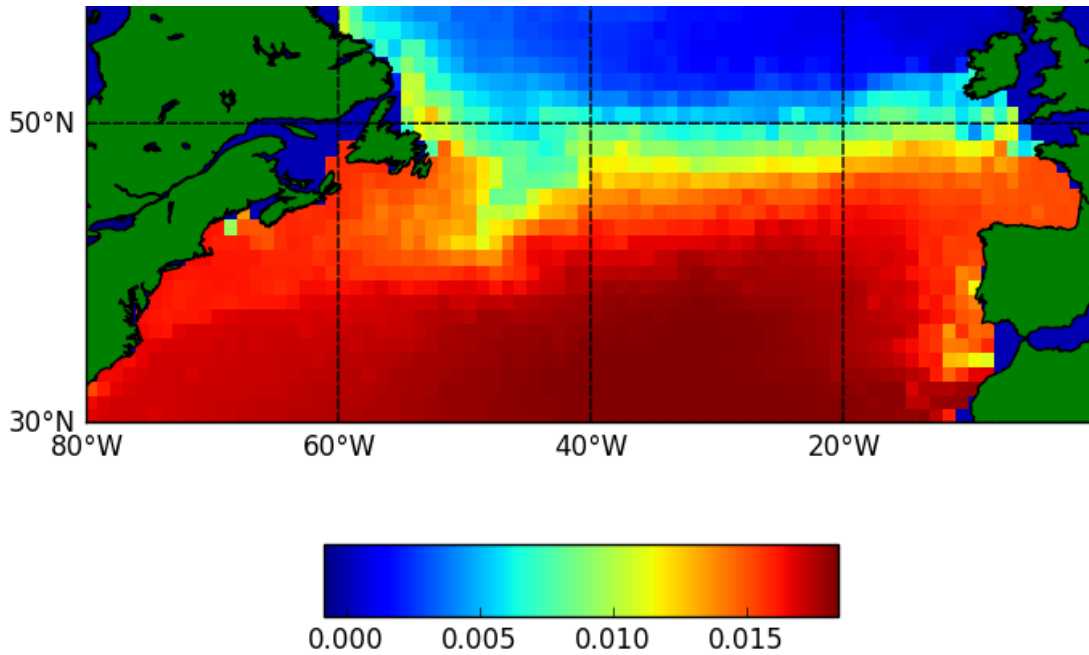


Figure 4.15: Eigenvector analysis of the GDP-based  $T_{dt=60days}$ . The values within the eigenvectors are relative to each other. See text for brief description on eigenvector analysis and Froyland et al. (2014) for a detailed explanation.

between the regions it oscillates between. If artificial dispersion were a key component in the cross-frontal transport, it would be expected that equal amounts of transport occurs in both directions. However, in this scenario there are geographically distinct transport patterns along the front. The widespread asymmetrical connectivity suggests that there is real, net, equatorward transport on the surface of the North Atlantic. This conclusion is also backed up by the magnitude of polar-to-tropical transport and the wide area of tracer origins which are connected to the sub-tropical gyre.

Section 3.3 concluded that the optimal time-step for Markov Chain analysis of real ocean drifters is difficult to assess. The domain release experiment is now repeated for  $T_{dt=5days}$ ,  $T_{dt=20days}$  and  $T_{dt=60days}$  (Figure 4.16, left column). As the time-step increases, the transition between areas connected to either gyre become less smooth. For example, in the  $T_{dt=5days}$  model, the Labrador Current connectivity to the sub-tropical gyre decays toward 40°W. However, this decay is more abrupt in the  $T_{dt=180days}$  model. The geographical distribution of inter-ocean connections are present regardless of timestep. Artificial dispersion, expected to be more prominent in shorter time-step models, may act to “over-connect” the Labrador Sea to the Labrador Current.

Because tracer which reaches the Labrador Current is transported to the sub-tropical gyre, the artificial dispersion indirectly induces more cross-frontal transport. Regardless of the amount of artificial dispersion, net transport at the surface is still equator-ward.

In the large time-step geostrophic models, the same abruptness of transition between sub-tropical and sub-polar connections is visible (Figure 4.16, middle column). However, there is also a larger difference between the full-field and geostrophic models at larger time-steps. As mentioned in the previous sections in this chapter, the  $T_{dt=5days}$  geostrophic matrix may approximate the full-field matrix. The results for both  $T_{dt=5days}$  models are almost identical, which support this hypothesis. Yet the same is true for the  $T_{dt=20days}$  as well. Therefore, the growing difference between full-field and geostrophic models is not explained by the ADT error. Instead, it is worth noting that the coverage of geostrophic drifters decreases with larger time-steps (Figure 4.4, Section 4.2). The  $T_{dt=180days}$  model predicts more equator-ward flow in the full-field model. A reason for this may be that the geostrophic model is built from considerably fewer trajectories in this region, it does not accurately recreate transport. The 60-day time-step is therefore expected to be the most reliable model, due to there being sufficient drifter readings and the reduced risk of mis-identification of geostrophy.

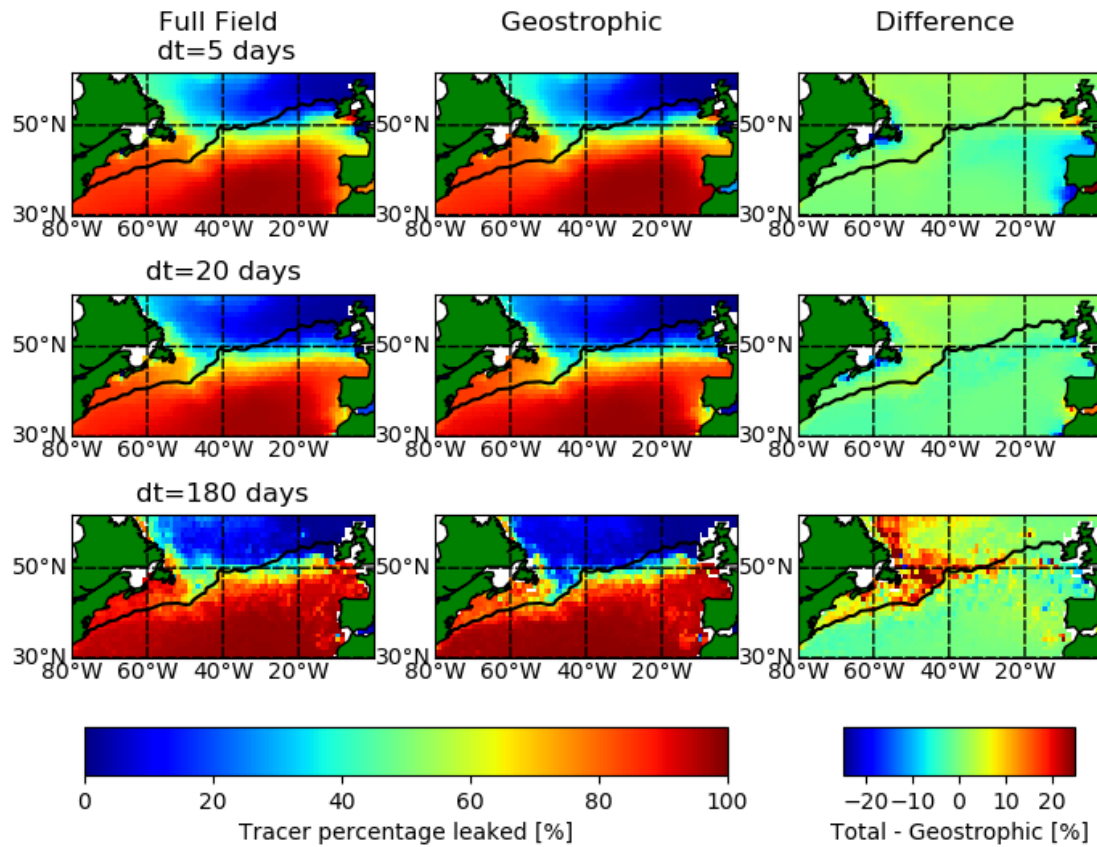


Figure 4.16: Cross-frontal leakage, after 5 years, between sub-tropical and sub-polar gyres in the North Atlantic for full-field and geostrophic drifters flows, using transition matrices  $T_{dt=5days}$ ,  $T_{dt=20days}$  &  $T_{dt=180days}$ .

## 4.5 Summary

In this section, two key areas of inter-ocean surface transport have been re-assessed with a methodology that analyses a vast number of trajectories. The Markov Chain analysis produced connectivity maps of the Agulhas Current system and the North Atlantic region. This chapter also introduced geostrophic Markov Chain models which, by comparison with the full-field transport, detects where geostrophic processes contribute to connectivity. Here, geostrophic motion is approximated by extracting drifters which follow a narrow range of ADT contour values. This approach uses observed trajectories and can therefore be directly compared with the full GDP dataset.

In the Agulhas Region, the extent of the source region of Agulhas Leakage was found to extend from the Agulhas Current towards the western Indian Ocean, with its southern border marked by the meandering Return Current. Within this region, geostrophic eddies are found to contribute to the flow towards the Agulhas Current, as much as non-geostrophic flow. As well as the re-circulation region, the connectivity to the Atlantic is also driven by the Madagascar and Mozambique Currents, which all act to transfer tracer into the Agulhas Current. In the North Atlantic region, there is a “one-way” system of cross-frontal surface transport. The northern flank of the Gulf Stream is almost fully-connected to the sub-tropical gyre, while the entire sub-tropical section remains blocked off from the sub-polar gyre. The Labrador Current transports roughly 50% of tracer passing through it to the sub-tropical gyre, with tracer on its western flank more likely to continue equator-ward. The connectivity map detected two distinct gyres and the border between them which, due to the aforementioned role of the Labrador Current, takes on a different pattern than was previously understood. Therefore, according to observations, the Gulf Stream acts as a surface barrier but only to northward transport. It is known that northward transport occurs at intermediate depths (Bower and Rossby, 1989), thus this result may still be consistent with the conservation of mass in the region.

In both examples of inter-basin transport, it was possible to conclude where geostrophic transport mostly contributed to connectivity. Within the Agulhas Current, wider recirculation region and the Labrador Current, geostrophic models contributed an important proportion of inter-

basin transport. The exact nature of the non-geostrophic transport was not determined here, although detached eddies from the Return Current are the likely mechanism in the Agulhas. However, in the case of the North Atlantic, it has been possible to infer where cross-frontal transport occurs. Source regions of tracer bound for the sub-tropical gyre are found west of the bifurcation of the Gulf Stream, while further east along the core of the North Atlantic Current the gyres are sharply divided.

The conclusions also agree with eigenvector decomposition of transition matrices, as well as for a range of parameters used to construct the transition matrices. Given the discussions in Chapter 3 regarding the "leakiness" of such models, it is encouraging that differences are minimal. However robust the regional analysis might be, there is a discrepancy between models on the exact magnitudes of transports between cells. The accuracy of Markov Chain models and the connections shown here are therefore unclear. Moreover, the geostrophic matrices were shown to be less accurate when using shorter time-steps, due to the mis-identification of so-called geostrophic trajectories at the 5-day period. While the quantity of geostrophic transport varied between parameters, differences between geostrophic and full-field models are still evident throughout the parameter space.

# Chapter 5

## Isopycnal Connectivity

### 5.1 Introduction

The role of currents as connectors or barriers of flow between regions is expected to change with depth for several reasons. First, the strength of western boundary currents typically decays and their paths shift laterally (Imawaki et al., 2013). The core of the Agulhas Current, for example, recedes from the continental slope and slows gradually with depth (Bryden et al., 2005). At a certain depth, the horizontal component of velocity reverses to become a sub-surface counter-current (Beal and Bryden, 1999). The Ekman transport, a key component of surface flow and accumulation, is confined to the Ekman layer, typically no more than 100 m deep. Eddies ejected from unstable boundary currents can depress the isotherm throughout the full depth of the ocean but water masses are only trapped and carried with eddies at certain depths. For the Agulhas Rings, this depth is approximately 1000 m (Duncombe Rae, 1991).

It is also known that the Gulf Stream acts as a front near the surface, although the previous chapter suggested this was true for pole-ward flow only. Model and observational float studies show that cross-frontal, pole-ward transport instead occurs in the interior and mostly along isopycnals (Bower and Rossby, 1989), although diapycnal motion is more likely at depths shallower than 1000 m (Burkholder and Lozier, 2011).

Montgomery (1938) and Iselin (1939) first proposed that motion in the interior was suppressed by stratification and a lack of radiative fluxes in the ocean interior. Given the lack of diabatic processes, they suggested that water parcels flow along surfaces of constant entropy. As a result, parcels do not have their potential density altered. Isentropic surfaces are therefore closely approximated by surfaces of constant potential density - isopycnals. Sub-surface water masses are expected to approximately maintain their densities and thus their distribution can be largely explained by isentropic transport analysis. Understanding along-isopycnal flow is also key to explaining the distribution of certain nutrients, such as oxygen (Richards, 1957).

In this chapter, the Markov Chain model experiments will be applied to isopycnal surfaces in a  $\frac{1}{4}^\circ$  circulation model. The aim is to gain a first-order understanding of the inter-basin connectivity caused by along-isopycnal flow. In Section 5.2, the motivation for choosing particular isopycnals is justified. Sections 5.3 and 5.4 focus on the Agulhas Current and the North Atlantic domains respectively. Connectivity maps for the isopycnal surfaces will be compared to the those obtained at the surface.

## 5.2 Isopycnal Surfaces: $\sigma_0=27.1$ & 27.6

The NEMO ORCA025 ocean circulation model output used is composed of 5-day averaged fields including the three-dimensional velocity components, potential temperature and absolute salinity. The latter two are used to calculate the potential density throughout the model ocean using the TEOS-10 Equations of State (Jackett and McDougall, 1995). Specifically, the python module `gsw.thermodynamics_from_t.sigma0_pt0_exact` is used. The isopycnals of interest in this thesis are the  $\sigma_0=27.1$  and 27.6 isopycnals.

The choice of two different isopycnals is motivated by the different water masses present on both, as well as the sloping of the isopycnals. Observations of water masses indicate some similarities and some differences between the two (Curry and Oppo, 2005). For example, the North Atlantic Deep Water (NADW) overflow from high latitudes extends further south on isopycnals deeper than 1000 m. To check for the presence of certain water masses within

ORCA025, the temperature and salinity properties along the isopycnal surfaces are shown in Figure 5.1. Here, names of observed water masses are used for brevity, although the exact T-S characteristics of the model masses are not confirmed to be equivalent to the real ocean.

On the 27.1 isopycnal, the North Atlantic sub-tropical gyre is warmer and saltier than its southern hemisphere equivalent, which indicates that the reach of the fresher Antarctic Intermediate Water (AAIW) does not extend north of the Equator. There is a northward warming gradient within the South Atlantic which originates from the Agulhas Current System. The Agulhas Leakage therefore transports Indian Ocean water into the Atlantic at the depth of the 27.1 isopycnal. Meanwhile, within the North Atlantic sub-tropical gyre, there is a westward cooling and freshening. This may be caused by southward pathways of the NADW and Labrador Sea Water (LSW) as well as the overflow of Mediterranean Sea Water (MSW) from the east. The presence of a local warm and salty intrusion into the Arctic Sea indicates that MSW is also transported pole-ward. It should be noted that the depth of the 27.1 isopycnal captures the shallower parts of these water masses, which extend deeper.

On the 27.6 isopycnal, there is less distinction in the T-S characteristics between hemispheres, with the North Atlantic appearing cooler than on the shallower isopycnal (Figure 5.1). This is indicative of the increased southward extent of the NADW. The warm, salty tongue extending from the Mediterranean is still present at 2000 m, although not as dominant in the gyre as on the 27.1 isopycnal. On the 27.6 isopycnal, the NADW appears to extend past the Equator and to share similar properties to waters from the Antarctic. Moreover, there is a much weaker signal of warmer waters entering the South Atlantic from the Indian Ocean via the Agulhas Leakage. In ORCA025, therefore, there are significant differences between the extent and types of water masses present on the 27.1 and 27.6 isopycnals.

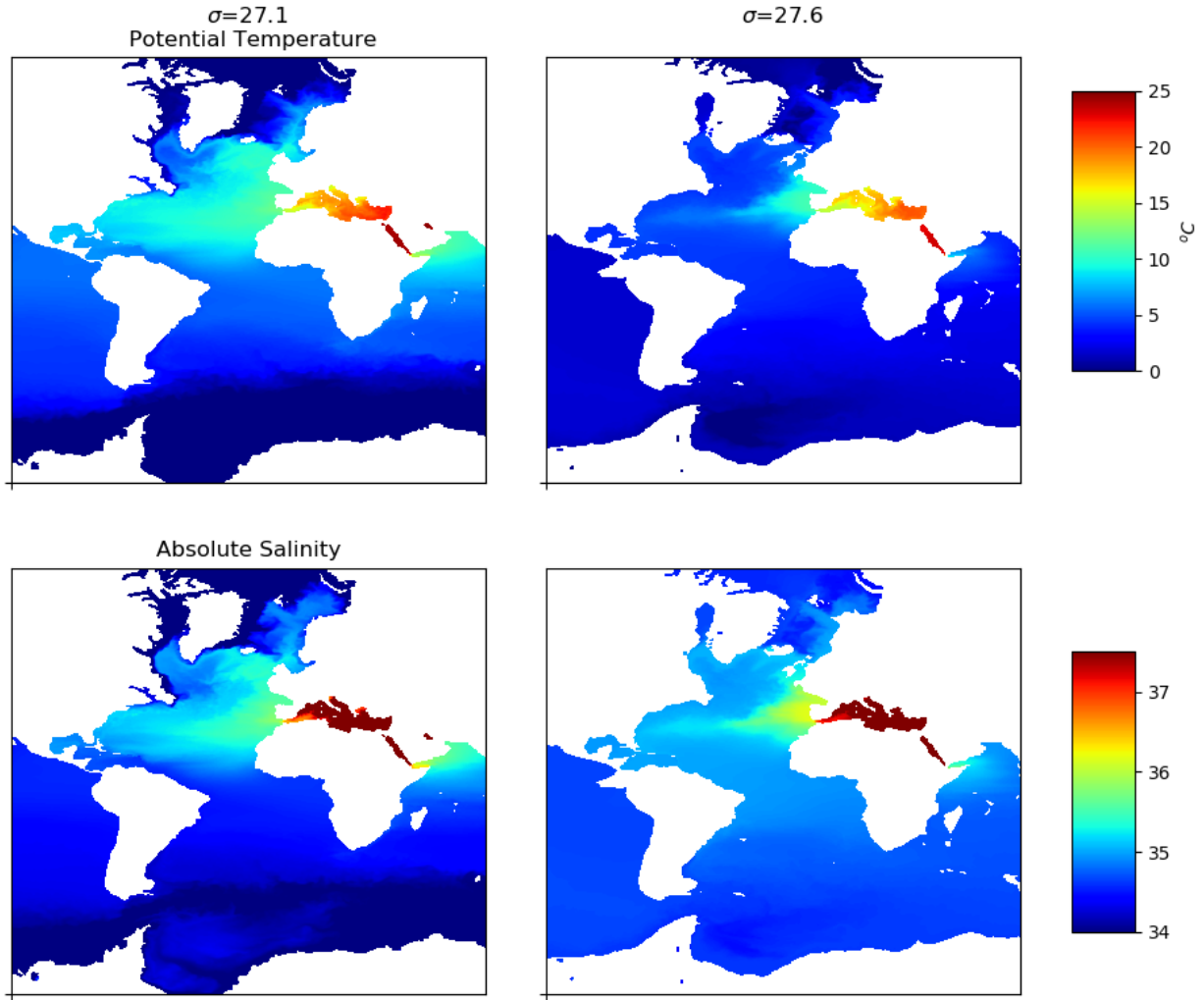


Figure 5.1: Temperature and salinity charts for the  $\sigma_0=27.1$  and 27.6 isopycnals. Potential temperature and absolute salinity (g/kg) are provided by ORCA025 output files. Here is shown the annual average for the 2001 output. The model output is plotted in its original tripolar grid form Section 2.3.2. for details

The water mass characteristics found on isopycnals depend largely on the source regions. Water masses are formed at the base of the mixed layer, introduced in Section 1.3. In practice, NEMO's definition of mixed layer depth is based on a joint temperature-density threshold criterion (de Boyer Montégut et al., 2004). Water is entrained during the spring months when the isopycnal outcrop shifts equator-ward, while during autumn the outcrop propagates pole-ward at a faster rate than the newly-entrained water is advected (Williams et al., 1995). In other words, the mixed layer consumes isopycnal water and re-mixes it. In Lagrangian experiments, the mixed layer is usually treated as a termination point for particles, particularly in studies

which focus on along-isopycnal transport (e.g. MacGilchrist et al., 2017). Given the differences in outcrop position and depth ranges between the 27.1 and 27.6 isopycnals (shown in Figure 5.2), mixed layer consumption will act differently on each.

The interaction between isopycnals and the mixed layer is therefore a key process in defining the water mass characteristics (Figure 5.2). The times shown here correspond to the ranges of outcropping positions in the Northern Hemisphere. Because the mixed layer definition used in NEMO is effectively uniform in density, the outcrop position indicates where the isopycnal intersects the base of the mixed layer and not the surface. In mid-spring, the isopycnal outcrop is at its southernmost point and begins propagating pole-ward. In mid-autumn, it begins receding equator-ward. There is negligible seasonal variation in the isopycnal depths or slope in the interior. However, there are notable differences in the outcrop positions of both surfaces and the mixed layer depths and, as a result, the overlap. The 27.1 isopycnal outcrops in the Southern Ocean, oscillating between Antarctica and the sub-tropical front of the ACC. It also outcrops in the North Atlantic between the Gulf Stream front and the north of the sub-polar gyre in the Barents Sea. Meanwhile, the 27.6 isopycnal outcrops only in the inner Labrador Sea. Because of its reduced exposure to the mixed layer, the source region for deeper isopycnals is restricted. In this case, it leads to more uniform thermohaline properties.

The 27.1 and 27.6 isopycnals at high latitudes are shallow enough to interact with the mixed layer within the sub-polar North Atlantic gyre (Figure 5.2). The mixed layer is deepest when the isopycnal outcrops closer to the equator. The area where water parcels are consumed by the mixed layer is therefore greatest in spring and at a minimum in autumn. This does not necessarily mean consumption rate is greatest at this time. As a result, on the 27.1 isopycnal, the south Atlantic Ocean is cut-off from the neighbouring ocean basins from August to October. The overlap area is considerably smaller for the 27.6 isopycnal because of its greater depth but parts of the Labrador Sea and the sub-polar gyre are deeper than the mixed layer in the spring. The consumption of water parcels, or in this case virtual particles, will therefore affect connectivity differently on either isopycnal.

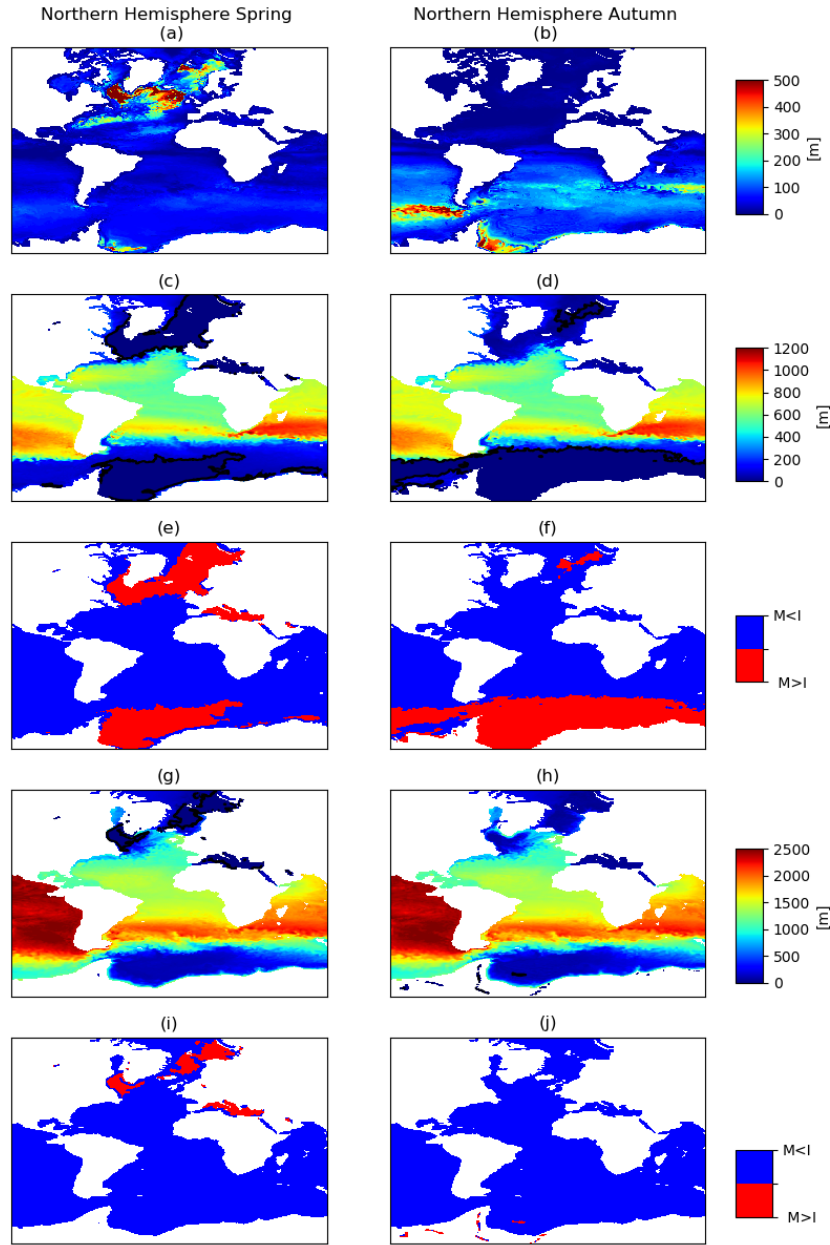


Figure 5.2: Mixed layer and isopycnal depths in ORCA025. Left column: 5-day average time-step fields when the greatest amount of isopycnal is shallower than the mixed layer depth in the northern hemisphere (5th April 2001). Right column: as before but for minimum amount (13th August 2001). (a) & (b) Mixed layer depths, (c) & (d)  $\sigma_0=27.1$  isopycnal depth, (e) & (f) area where 27.1 isopycnal (I) is shallower than the mixed layer depth (M), (g) & (h)  $\sigma_0=27.6$  isopycnal depth, (i) & (j) area where 27.6 isopycnal is shallower than the mixed layer. The black contour in the isopycnal depths is the "outcrop" line where the isopycnal intersects the mixed layer depth. The overlap area highlights where the mixed layer is deeper than the isopycnal depth ( $M < I$ ). The model output is plotted in its original tripolar grid form (Section 2.3.2).

### 5.3 Interior Velocity Fields

Besides differences in the water masses sampled, the 27.1 and 27.6 isopycnals will also sample different currents or the same currents at different depths and strengths. In the real ocean, surface currents decay and shift laterally with depth, and may be replaced by an undercurrent flowing in the reverse direction. Examples include the Agulhas Undercurrent (Beal and Bryden (1997)) and the Deep Western Boundary Current below the Gulf Stream (Richardson, 1977). The structure of ocean currents with respect to the target isopycnals will be described here, to help analyse the connectivity studies in the following sections.

ORCA025 captures key coherent features of the surface flow in the Agulhas Region including the Mozambique and Madagascar Currents, the formation of Agulhas Rings and the meandering of the Return Current (Figure 5.3). At 25°S, the Agulhas Current is a narrow feature that extends to the depth of the 27.1 isopycnal. Meanwhile, the 27.6 isopycnal crosses just below the core of the Agulhas Undercurrent, which is also a narrow current adjacent to the coastline. The Madagascar Current, a tributary to the Agulhas Current, also begins decaying below the 27.1 isopycnal. On the eastern side of the South Atlantic, where the northward Benguela Current is expected, there are instead discontinuous northward velocities along the 25°S. Further south, at 32°S, the Benguela Current northward component is clearer and the Agulhas Current widens. There is no apparent undercurrent but this may be because it does not give a strong signal on the zonal velocity map. The meridional velocity field at 25°E does detect a reversal in the flow direction, albeit below the 27.6 isopycnal.

In the North Atlantic, the Gulf Stream is a coherent feature which turns into the more discontinuous North Atlantic Current at the intersection with the Labrador Current (Figure 5.4). Surface transport is largely eastward, with the exception being between the Gulf Stream and the coast where transport is westward at the depths of both isopycnals. Both the Gulf Stream and the westward flow are maintained, or in the case of the latter, widened, at the depths of both isopycnals. In the gyre interior, the eastward surface flow component decays before the depth of the 27.1 isopycnal and is replaced by a westward velocity. Between the Gulf Stream and the coast, there is locally enhanced zonal transport around the mount at 50°W. Further

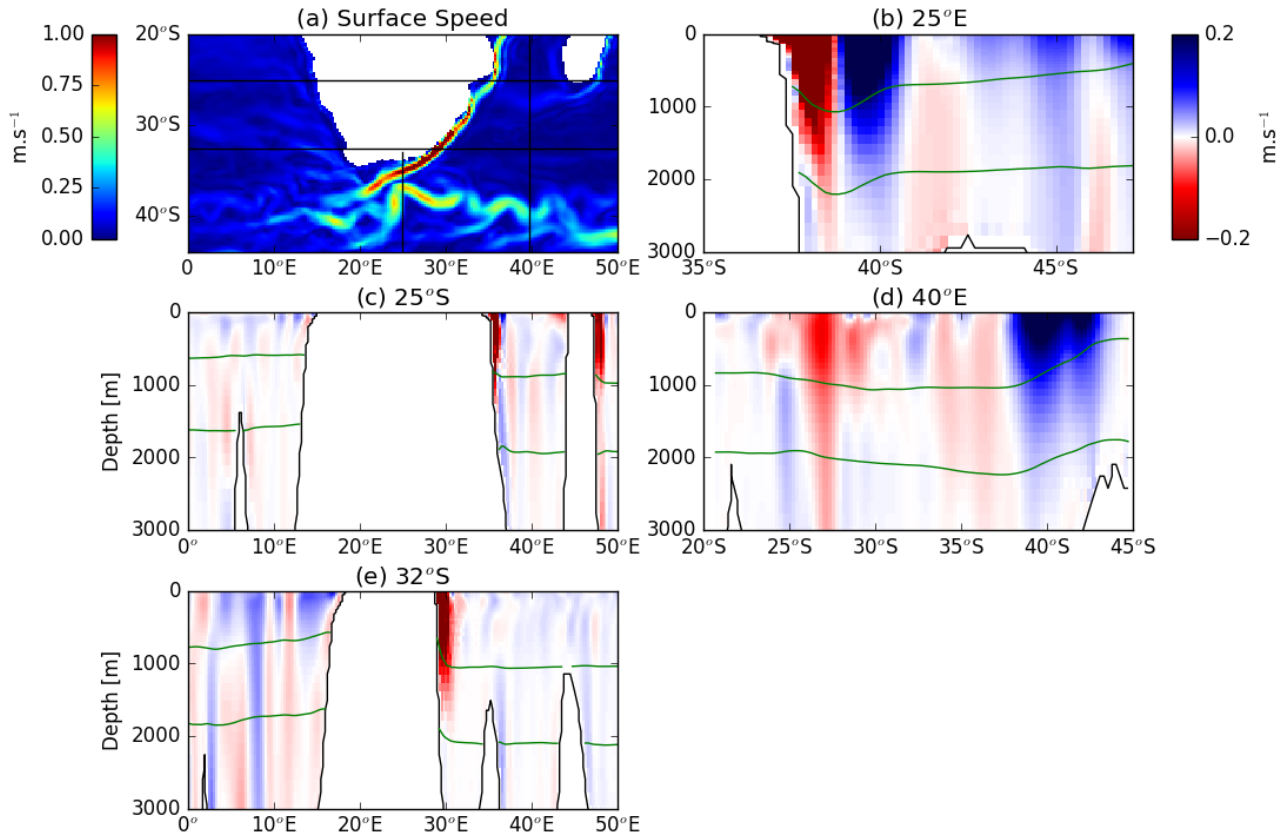


Figure 5.3: Velocity fields in the Agulhas Current System in ORCA025. (a) Annual average speed at the surface. Black lines indicate the 25°S and 32°S parallels and the 25°E and 40°E meridians. (b) Cross-section of the zonal component of velocity through the 25°E meridian, where positive values indicate eastward flow. (c) Cross-section of the meridional component of velocity through the 25°S parallel, where positive values indicate northward flow. (d) As in (b) but for the 40°E meridian (e) As in (c) but for the 32°S parallel. Green lines indicates the depth of the  $\sigma_0=27.1$  and  $27.6$  isopycnals. All fields are for the ORCA025 2001 annual average.

eastward, yearly-averaged north and south flows are found in proximity to each other, indicative of a highly-meandering pathway for the tail of the Gulf Stream.

For the purpose of advecting particles along the isopycnal surfaces, the velocity fields along the isopycnals are assumed here to be equal to the horizontal velocity components. ORCA025 velocity fields are interpolated to the depth of the desired isopycnal,  $z_\sigma$ , to provide  $\mathbf{V}_{z_\sigma} = (u_{z_\sigma}, v_{z_\sigma}, w_{z_\sigma})$ . The vertical velocity is not used here. Therefore, the velocity field input for Lagrangian advection codes is defined as:

$$\begin{aligned} u &= u_{z_\sigma}, \\ v &= v_{z_\sigma}. \end{aligned} \tag{5.1}$$

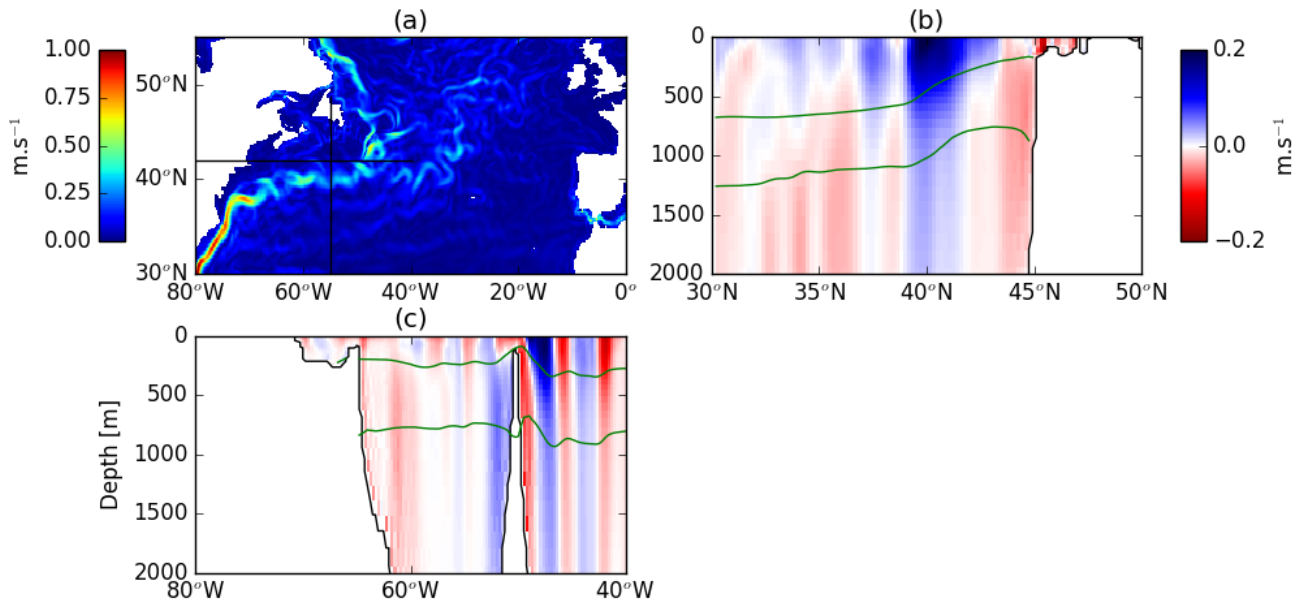


Figure 5.4: Velocity fields in the North Atlantic in ORCA025. (a) Annual average speed at the surface. Black lines indicate the  $42^\circ\text{N}$  parallel and the  $55^\circ\text{W}$  meridian. (b) Cross-section of the zonal component of velocity through the  $55^\circ\text{W}$  meridian, where positive values indicate eastward flow. (c) Cross-section of the meridional component of velocity through the  $42^\circ\text{N}$  parallel, where positive values indicate northward flow. Green lines indicates the depth of the  $\sigma=27.1$  and  $27.6$  isopycnals. All fields are for the ORCA025 2001 annual average.

(Pena-Izquierdo et al., 2015), for example, expands on this method by vertically averaging the horizontal velocities over a  $0.05 \text{ kg.m}^{-3}$  layer. Both method ignore the vertical velocities. The assumption that flow at the isopycnal surface is defined solely by Equation 5.1 is only true if the slope of the isopycnal is horizontal and the vertical velocity component is zero. In reality, isopycnals slope relative to the horizontal plane. The omission of vertical velocity, while orders of magnitude smaller than the horizontal components, will affect trajectories on the timescales of interest ( $\geq O(10 \text{ years})$ ). In particular, isopycnals slope relatively steeply at higher latitudes towards outcropping positions, towards gyre centres and near rapid currents (Figure 5.2).

Due to isopycnal sloping relative to the horizontal, the direction of along-isopycnal transport is not aligned with the meridional and zonal axes. In this study, the aim is to force particles to remain on the isopycnal. Ideally, velocity fields would consist of along-isopycnal velocity components which are aligned with the  $\mathbf{i}$  and  $\mathbf{j}$  components (Figure 5.5). However, the offset of the horizontal velocity components from true along-isopycnal value is quantified with the the normal to the isopycnal surface. This is defined as the unit vector in the direction of maximum

density gradient:

$$\hat{\mathbf{n}} = \frac{\nabla\sigma}{|\nabla\sigma|}. \quad (5.2)$$

The regions where this misalignment is prominent are shown by the offset of the isopycnal normal to the vertical (Figure 5.6). Sloping is greatest typically where surface currents extend deep enough to reach the isopycnal surface, as seen in the western boundary currents and along the northern flank of the Antarctic Circumpolar Current. The exceptions are the Equatorial currents. Where the sloping is greatest, the horizontal velocities will differ more from the true values.

Along-isopycnal velocities would need to be on the plane of the isopycnal and therefore perpendicular to the normal (Figure 5.5). They would also need to be aligned with the unit vectors  $\mathbf{i}$  and  $\mathbf{j}$ , in order to be used by Lagrangian advection codes. In the absence of such a method, the horizontal velocities are used. The benefit of this is that Lagrangian advection codes such as OceanParcels are designed to work with velocity components in the standard Cartesian reference frame defined by unit vectors  $\mathbf{i}$ - $\mathbf{j}$ - $\mathbf{k}$ . Schemes which adapt Cartesian velocity components on a curvilinear plane, such as north of 20°N on ORCA tripolar grids, exist and have been validated for use in OceanParcels (Delandmeter and van Sebille, 2019).

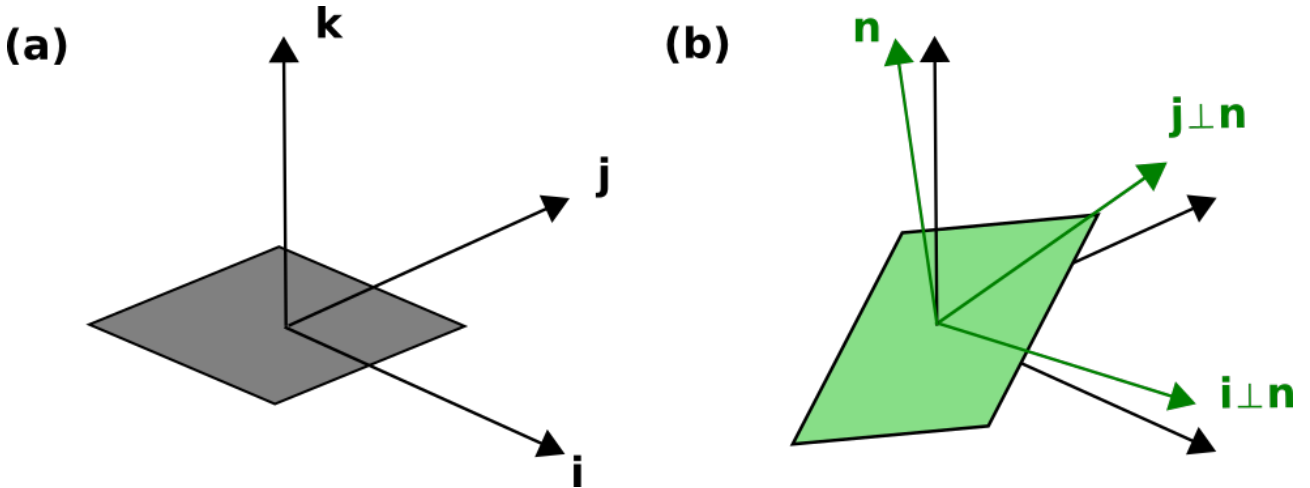


Figure 5.5: Schematic explaining the difference between horizontal and along-slope velocity components. (a) A horizontal surface with unit vectors  $\mathbf{i}$ ,  $\mathbf{j}$  and  $\mathbf{k}$  in the meridional, zonal and vertical directions respectively. (b) A sloping surface showing the along-isopycnal components,  $\mathbf{i} \perp \mathbf{n}$  and  $\mathbf{j} \perp \mathbf{n}$ , and the normal to the surface,  $\mathbf{n}$ .

Diapycnal transport is not studied here. Model studies show that diapycnal transfer rates are greater at shallower depths. On the 27.1 isopycnal, a  $0.1 \text{ kg.m}^{-3}$  change in potential density is expected after 4 years (Burkholder and Lozier, 2011). (Pena-Izquierdo et al., 2015) performed a similar study, in which they interpolate the 2D and 3D velocity fields to the depth of the  $\sigma_0 = 26.8$  isopycnal. They also used horizontal velocity fields for the 2D case and, by comparing the 2D and 3D trajectories, assess the role of diapycnal mixing. The diapycnal flow component is not explicitly considered in this thesis. As a first step in understanding interior connectivity, this chapter focuses on along-isopycnal flow.

The OceanParcels Lagrangian toolkit, introduced in Chapter 2.3, is used to advect particles in the along-isopycnal velocity fields. Particles were deployed every  $\frac{1}{2}^\circ$  and run for 2 years between the years 2000-2001, the latest available years from the ORCA025 database. This timescale is roughly equivalent to the average lifetime of GDP drifters but the number of particles is an order of magnitude larger than the number of surface drifters available. Particles which enter an area where the mixed layer is deeper than the isopycnal are deleted. As a result, the model runs retain over 70% of particles after 2 years. Transitions matrices are constructed from this dataset, including “pre-death” trajectories, in the same way as described in Chapter 2.4.

## 5.4 Agulhas Current System

### 5.4.1 Model surface comparison with GDP

First, it is necessary to assess the capability of ORCA025 to simulate inter-ocean connectivity. In the absence of observed trajectories of isopycnic floats, a comparison is restricted to surface flow. The connectivity map for the Agulhas Leakage is here recreated for a surface ORCA025 particle dataset and compared to the GDP-based connectivity (Figure 5.7). In the GDP-based model, Agulhas Leakage tracer originates from a wide recirculation region adjacent to the Agulhas Current. In the ORCA025 model, tracer destined for the Agulhas Leakage originates mostly along the narrow Agulhas Current. There is less than 10% tracer transport originating

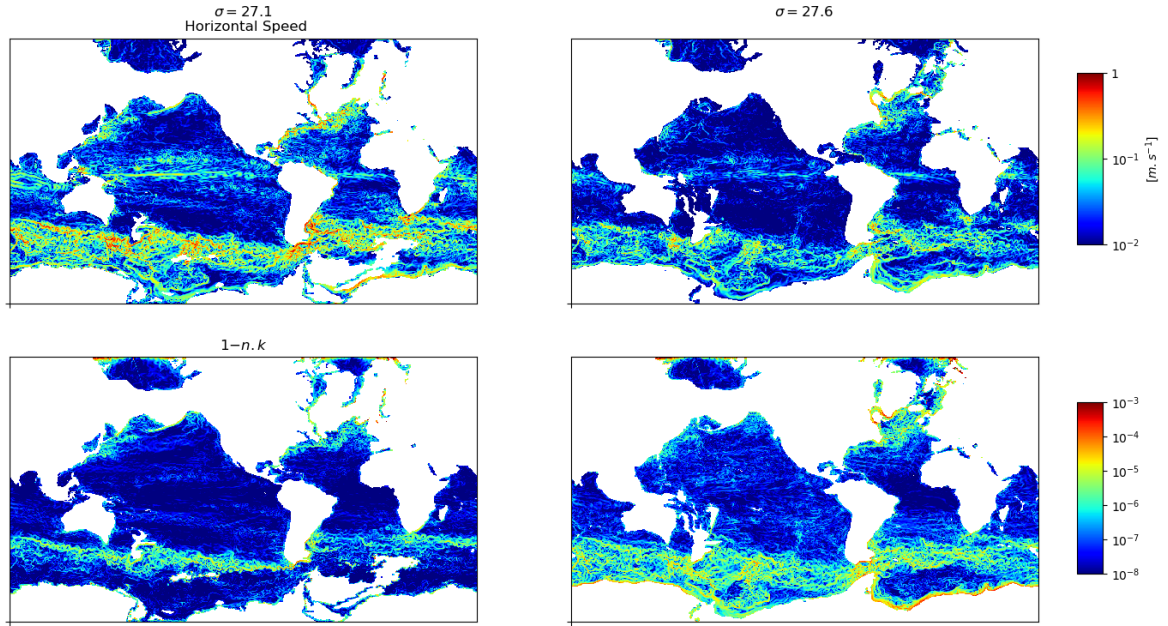


Figure 5.6: Along-isopycnal velocity fields and isopycnal sloping. Top: Magnitude of horizontal velocity components interpolated to the corresponding isopycnal depth. Bottom: Deviation of the dot product of  $\mathbf{n}$  and  $\mathbf{k}$  from unity, where higher values indicate steeper sloping. Time-step is the first from ORCA025 2001.

from cells in the adjacent region, as opposed to 20-35% in the GDP-based experiment. The difference in transport is evident even after 5 years when, as shown in Section 4.3, transport out of the Agulhas system no longer changes in the time-averaged Markov Chain model. This suggests that the processes which bring tracer into the Agulhas Current are not resolved in the ORCA025 model. This comparison shows that the region where these processes are not resolved extends eastward as far as  $45^\circ\text{E}$ . In the real ocean, tracer emitted from the Return Current is re-circulated to the Agulhas Current by submesoscale processes (Pollard and Read, 2017). Therefore, the most likely reason for the reduced area of connectivity in the ORCA025 is that it only resolves larger eddies formed from the meandering of the Return Current.

Another key comparison at the surface is that a smaller proportion of tracer takes the Leakage route than in the observations. This could indicate that the Agulhas Current is more stable in the model, thereby producing fewer Agulhas rings. Another reason may be that even within strongly advective flows, unresolved physics may contribute to transport. For example, Casal et al. (2009) observed onshore-to-offshore eddies moving through the Current affected

the downstream transport. However, it is also important to note that the region which is solely connected to the South Atlantic gyre (denoted by the separation between red and blue areas in Figure 5.7) has its boundary between  $5^{\circ}\text{E}$  and  $20^{\circ}\text{E}$  shifted northward by several degrees latitude in the ORCA025 model. Given that the Agulhas Leakage is the sole connection between the two basins, the northward-shifted boundary between the basins is evidence that Rings are formed more northward in ORCA025.

In theory, a model in which the Agulhas Current loses inertia too rapidly will cause the Current to bend westward towards the South Atlantic, shedding Rings further north and less frequently (de Ruijter, 1982; van Sebille et al., 2009). Here, however, the ORCA025 model appears to transport less tracer via Agulhas Rings while pushing the retroflection point northward (Figure 5.7). It is possible that the virtual particle release does not detect a representative number of Agulhas Rings over the two year lifetime. Either way, the ORCA025 model does not accurately recreate the Agulhas Current and its exit routes. The comparison with the GDP-based model therefore agrees with previous findings that, to capture a representative sample of Agulhas Leakage features, a  $0.1^{\circ}$  model is desirable (Biaosoch et al., 2008). Figure 5.8 repeats the experiment for the range of time-steps ( $dt$ ), to check if artificial dispersion affects the amount of Leakage. Reassuringly, the conclusions drawn here are true regardless of the transition matrix parameters used.

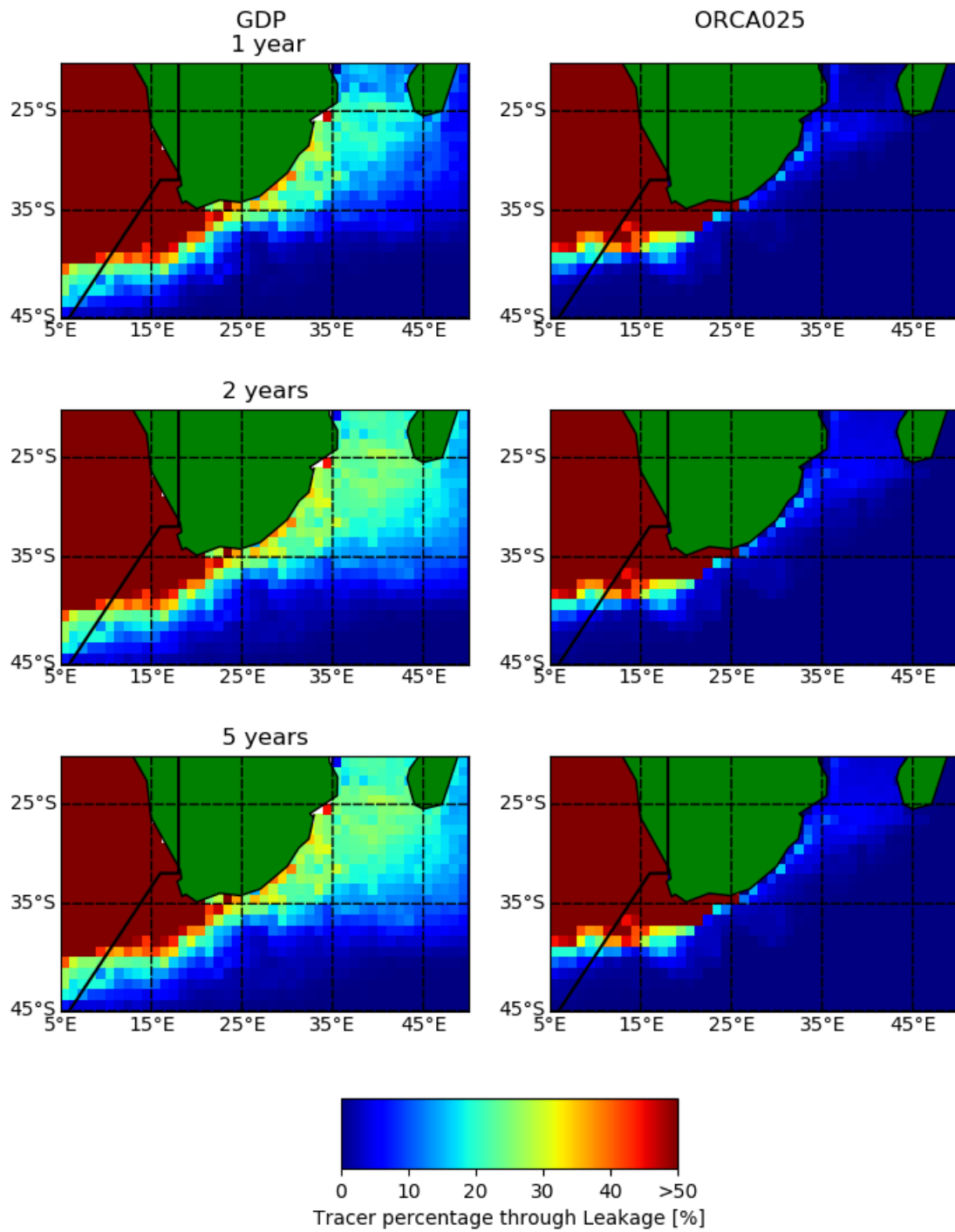


Figure 5.7: Agulhas Leakage connectivity map for the GDP drifter and ORCA025 surface trajectories. Tracer which crosses the Good Hope line (black) is defined as Leakage. A  $T_{dt=60days}$  matrix was used for both and the models were run for 5 years.

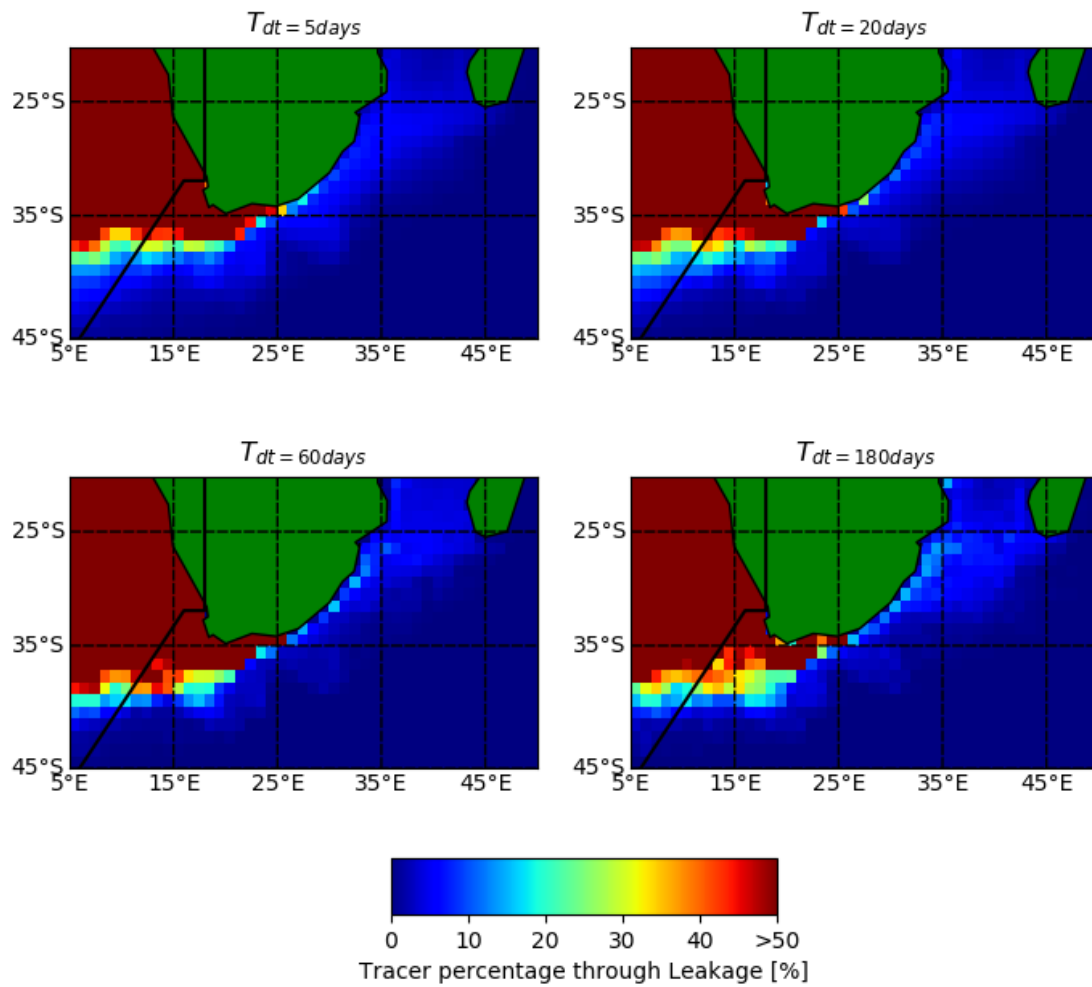


Figure 5.8: Agulhas Leakage connectivity map for a range of transition matrices built from ORCA025 surface trajectories. Tracer which crosses the Good Hope line (black) is defined as Leakage.

### 5.4.2 Along-isopycnal connectivity

ORCA025 does not fully represent inter-basin connectivity but it provides the sole dataset to highlight differences between the surface and target isopycnals in the absence of widespread observations. The most recent Coupled Model Inter-Comparison project (CMIP5) used an ensemble of  $0.2 - 2^\circ$  ocean models (Taylor et al., 2012). The next-generation set of global climate models, from CMIP6, still refer to “high” resolution as finer than  $\frac{1}{4}^\circ$  (Haarsma et al., 2016). ORCA025 therefore still provides a relatively modern and widely-used ocean simulation (Section 2.3.2).

Given the typically lower velocities at depth, the analysis is performed for a 25 year timescale (Figure 5.9). On the  $\sigma_0=27.1$  isopycnal, the connectivity between the Indian Ocean and South Atlantic basins becomes stronger and more widespread with time. Tracer originating in the Agulhas Current escapes the region within a year, with roughly 25% leaving via Leakage. At 5 years, up to 30% of tracer from the recirculation region and western part of the Indian Ocean gyre is transport by Leakage. The influence of the Return Current is visible from the meandering southern boundary of this recirculation region throughout the model run. Eventually, the region of connectivity expands throughout the domain, with the Return Current path being the exception. Another difference is found in the boundary to the South Atlantic. Whereas this boundary on the model surface was found north of  $40^\circ\text{S}$ , the same boundary on the 27.1 isopycnal is found further south, towards  $45^\circ\text{S}$ .

The importance of the magnitude of Agulhas Leakage is highlighted further by considering the evolution of tracer transport in time for each surface (Figure 5.10). On the model surface, Agulhas Leakage makes up less than 10% of the exit flow and does not match the strength of the observed flow (van Sebille et al., 2009). Agulhas Leakage is stronger on the 27.1 isopycnal than the 27.6 isopycnal. After 25 years, shallower Leakage transport is roughly 5 times larger than on the deeper isopycnal (Figure 5.10). The time at which fluxes no longer vary are also different between the three surfaces. For the surface and 27.1 isopycnal, the system reaches a state where fluxes between the Agulhas Current, Return Current and the re-circulation region are steady after 15 years. On the 27.6 isopycnal, after 25 years, the amount of tracer leaving

via the Return Current increases but the Agulhas Leakage transport does not counter this.

The inter-basin connectivity is stronger and more widespread on the 27.1 isopycnal than at the surface (Figures 5.7, 5.9 & 5.10). This may be surprising, given that the lack of sub-grid scale processes is expected to reduce inter-basin transport compared to the observations. However, the T-S distribution at these depths implied stronger connectivity for the 27.1 isopycnal than the deeper 27.6 isopycnal. Again, the experiment is repeated for a range of parameters to ensure artificial dispersion is not the cause of the regional pattern of tracer distribution. The artificial dispersion acts only to smooth out the regions of connectivity without changing their extent or magnitude sufficiently to change the understanding of connectivity (Figures 5.8, 5.11 and 5.12). Therefore, the enhanced dispersion on the 27.1 isopycnal is a domain-wide feature in ORCA025 and not an artefact of the Markovian representation.

On the  $\sigma_0=27.1$  isopycnal, the source region of Agulhas Leakage is more widespread than on the deeper isopycnal (Figure 5.9). This is due in part to the relative advection speeds. However, even after 5 years, connectivity is restricted to the area close to the Retroflexion. This indicates that Leakage Rings in the ORCA025 still transport water at the depth of the 27.6 isopycnal. After 25 years, the connectivity on the deeper isopycnal extends up the Agulhas Current by only several hundreds of kilometers. Throughout the run, less tracer exits via Leakage compared to the shallower isopycnal. Additionally, a region of tracer to the northwest of the Good Hope line is transported towards the Agulhas Region. This feature is only visible on the 27.6 isopycnal.

Several features on the connectivity map can be explained by the presence of undercurrents at the isopycnal depths. Cross-sections through the region highlight the Agulhas and Benguela Undercurrents at the depth of the 27.6 isopycnal (Figure 5.3). There is pole-ward transport on the 27.1 isopycnal along the eastern coast because the Agulhas Undercurrent is not present at this depth. The undercurrent does however play a key role in driving inter-basin transport on the 27.6 isopycnal, shown by the limited connectivity along the coastal region (Figure 5.9). Meanwhile, only 25% of tracer in the neighbouring region northwestward of the Good Hope line remains in the South Atlantic gyre. Instead, the majority is transported toward the Agulhas System and into the Indian Ocean. In the Markov Chain model, time-averaged Agulhas Leakage

successfully enters the South Atlantic but the Benguela Undercurrent acts to counter this Leakage by returning tracer back, north of the Retroflexion. As shown in Figure 5.3, the northward Benguela Current, still present on the 27.1 isopycnal, decays and is replaced by discontinuous southward jets.

Another key physical feature to be explained is the widespread region of strong connectivity on the 27.1 isopycnal (Figure 5.9). On the surface, ORCA025 does not adequately depict this transport towards or into the Agulhas Current (Section 5.3.1). As it is unlikely that dispersive processes such as eddies “grow” to a resolvable scale at the depth of the 27.1 isopycnal and shrink again before the 27.6 isopycnal, the model connectivity must be derived by other, resolvable features. At 25°S, in ORCA025, between the continent and Madagascar, southward jets appear between roughly 200m and 2000m depth, therefore missing the 27.6 isopycnal (Section 5.3). These jets act as tributaries to the Agulhas Current along the 27.1 isopycnal. Crucially, there are westward jets flowing between 33°S and 37°S in the re-circulation region which reach a maximum width at the depth of the 27.1 isopycnal. They decay below this depth. While this does not fully explain the extent of the Leakage source region, it does explain why it is present only on the 27.1 isopycnal.

In the previous chapter, it was shown how artificial dispersion can impact connectivity maps. Namely, more artificial dispersion acts to smooth the region of connectivity. Figures 5.11 and 5.12 indicate that this is true for the isopycnal layers too. Lagrangian statistics are expected to differ between surface and deeper waters because of the reduced value of real lateral diffusivity and generally slower currents (LaCasce, 2008). Artificial dispersion, however, appears to play the same role as at the surface. Fortunately, the conclusions drawn about regional connectivity are unchanged by the choice of parameters used (Figures 5.11 & 5.12).

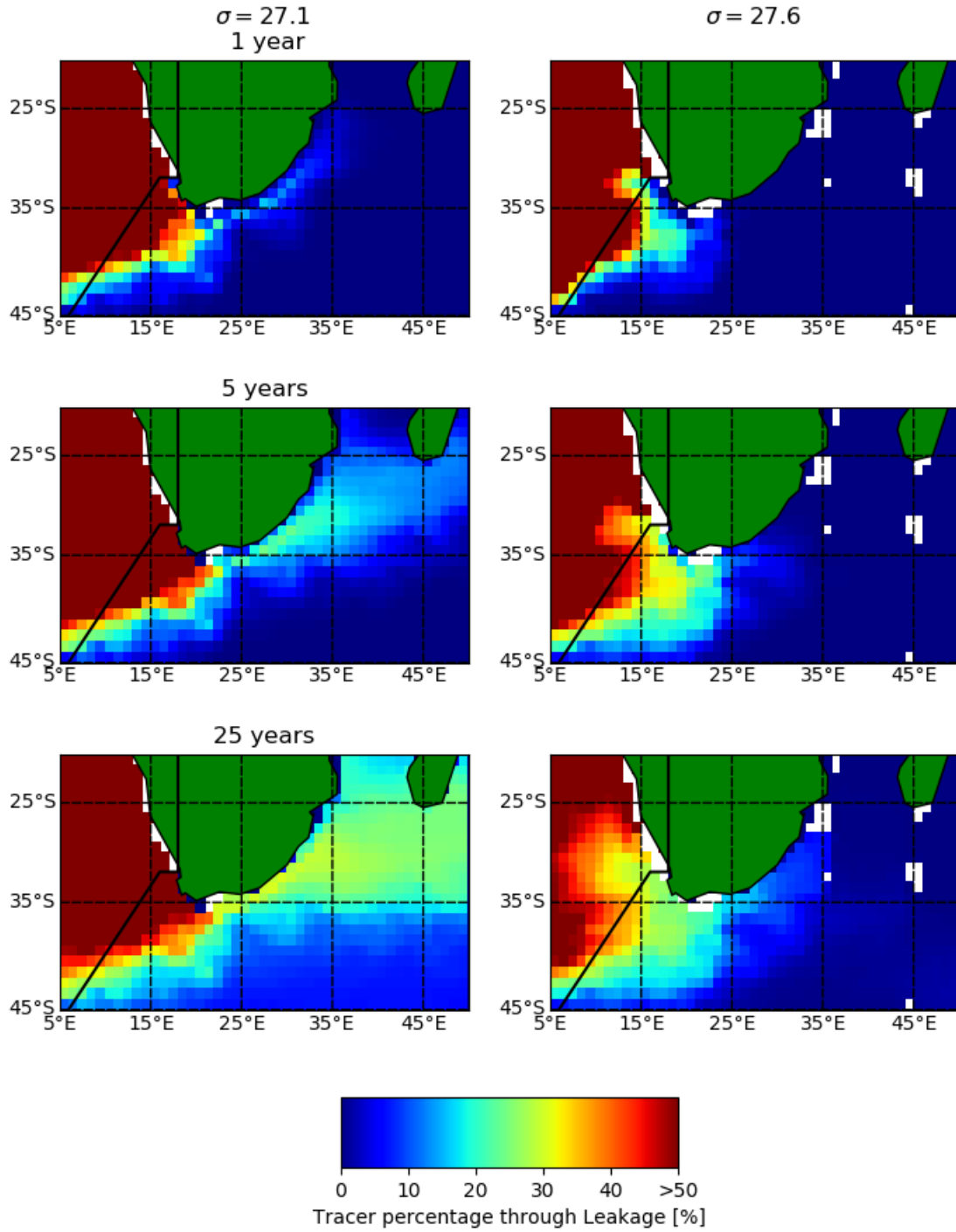


Figure 5.9: Agulhas Leakage connectivity map for ORCA025  $\sigma_0 = 27.1$  and  $\sigma_0 = 27.6$  isopycnal trajectories. The depths of the isopycnal regions in these region are, respectively, roughly 1000 m and 1800 m. Tracer which crosses the Good Hope line (black) is defined as Leakage. A  $T_{dt=60days}$  matrix was used for both and the models were run for 25 years.

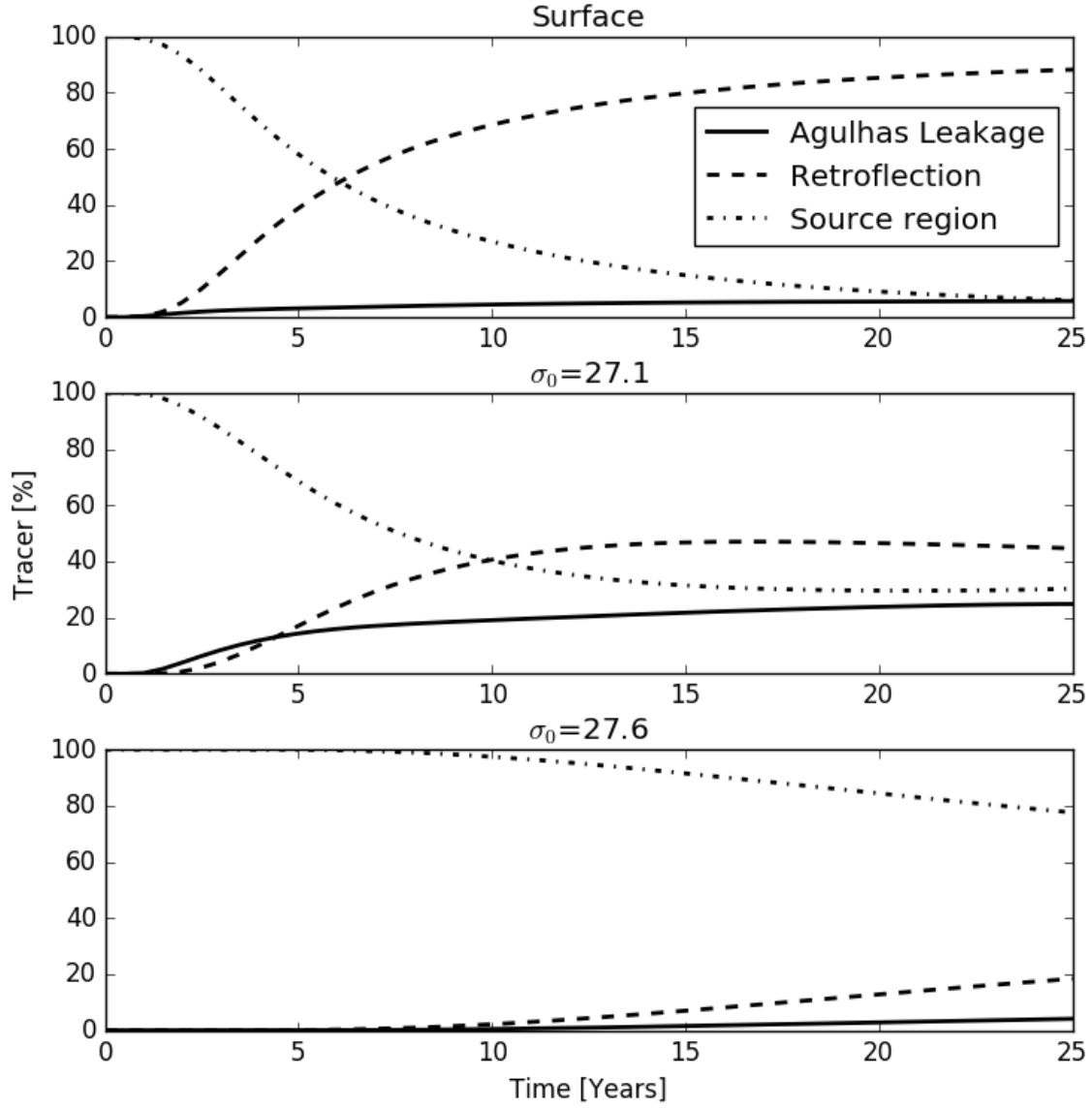


Figure 5.10: Time-series of Agulhas Leakage fluxes for the surface, 27.1 and 27.1 isopycnals. Tracer is released from  $[31^\circ\text{E}, 33^\circ\text{S}]$  in  $T_{dt=60days}$  Markov Chain models for each surface. The boundaries for the Agulhas Leakage and Return Current are the Good Hope line and  $40^\circ\text{E}$  respectively (in Figure 4.8).

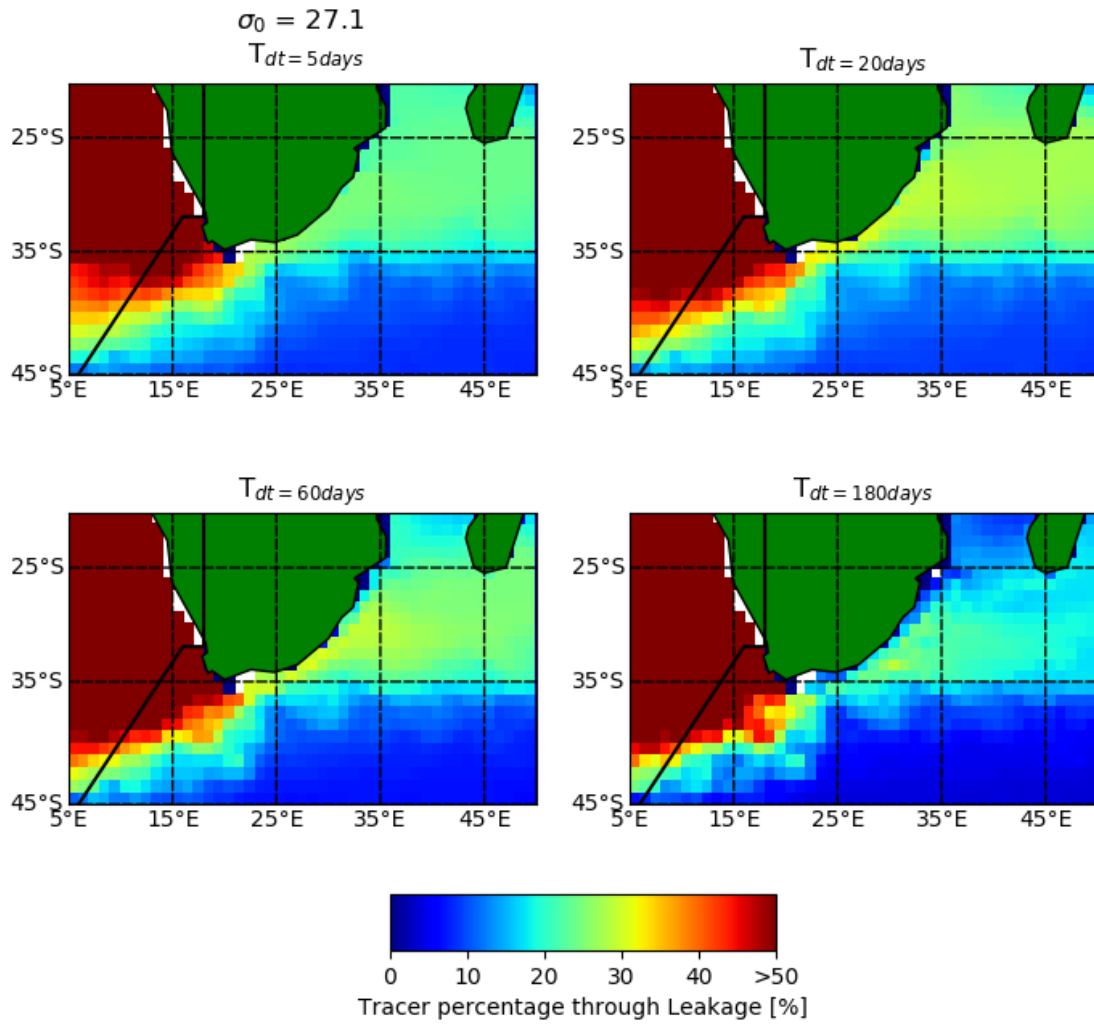


Figure 5.11: Agulhas Leakage connectivity map for a range of transition matrices built from ORCA025  $\sigma_0=27.1$  isopycnal trajectories. Tracer which crosses the Good Hope line (black) is defined as Leakage.

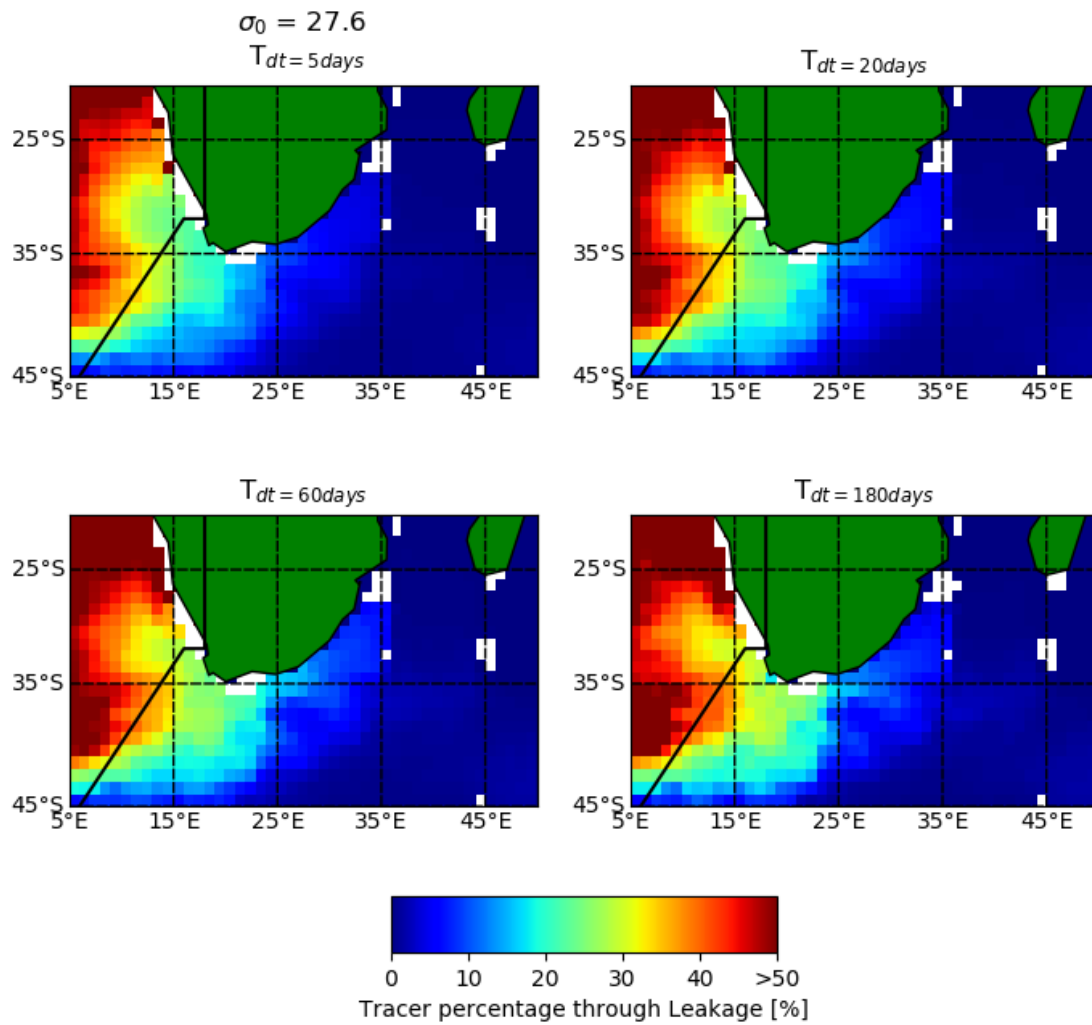


Figure 5.12: Agulhas Leakage connectivity map for a range of transition matrices built from ORCA025  $\sigma_0=27.6$  isopycnal trajectories. Tracer which crosses the Good Hope line (black) is defined as Leakage.

## 5.5 North Atlantic Transport Barrier

### 5.5.1 Model surface comparison with GDP

In the North Atlantic, as for the Agulhas Region, the ORCA025-derived Markov Chain model is compared to the GDP-derived model (Figure 5.13). There are several noticeable differences in this region which either highlight the missing sub-grid scale dynamics or a shift in the position of the Gulf Stream. First, in ORCA025, less tracer released in the path of the Gulf Stream remains south of the front. In the first year, there is minimal cross-frontal transport from within the Gulf Stream. This later increases to 25% after 5 years. Given the rapid advection of the current, this timescale indicates that cross-frontal transport occurs in the eastern part of the domain. The GDP-based Gulf Stream tracer instead remains within the sub-tropical gyre. Likewise, the northern flank of the Gulf Stream shows 25-35% less southward transport than the GDP model.

Another key component of the inter-gyre connectivity which is altered in ORCA025 is the front eastward of the the Gulf Stream bifurcation at 40°N. The position of the derived front between the gyres appears to be almost 10° south of the observed front (Figure 5.13). The model-derived separation is found further north of 50°N whereas the GDP model remains level below 50°N from 40°W eastward. ORCA025 also produces a weaker and narrower Labrador Current connection. The entire Labrador Sea is less well connected, indicating that the role of the Labrador Current in transporting adjacent tracer southward is reduced. This is presumably because the front is shifted southward in ORCA025. The key components of connectivity according to observations - the position of the North Atlantic front and the contribution of the Labrador Current - are therefore less well-represented in the ORCA025 model.

The front used in Chapter 4, derived from GDP drifter readings by Rypina et al. (2011), is again used as to define the North Atlantic front. To justify the use of an observed front in a model simulation, it is compared to the velocity structure and ADT from ORCA025 (Figure 5.14). The observed front, deduced from local ADT and velocity maxima, is found shifted slightly southward west of 40°W. ORCA025 appears to contain a stronger southward branch of

the Gulf Stream bifurcation from 40°W, although this signal is less prominent in the velocity map. Towards the eastern part of the domain, the observed front again aligns with both ADT and velocity maxima. The previous Chapter compared the use of this front to the 45°N parallel and achieved very similar geographical patterns for connectivity. It is therefore expected that the slight differences in the ORCA025 will not change the quantity of cross-frontal transport measured.

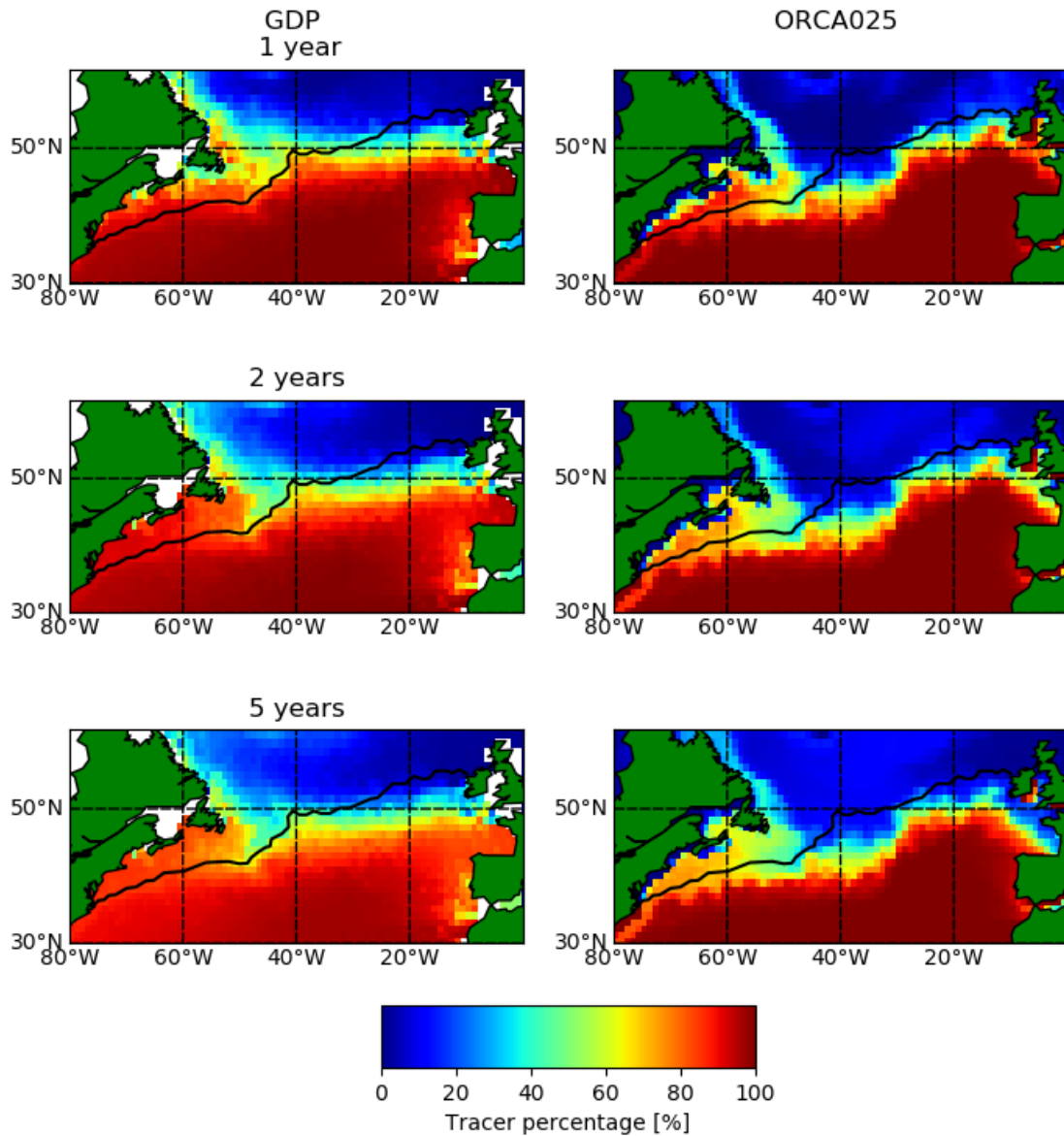


Figure 5.13: Cross-frontal connectivity map for the GDP drifter and ORCA025 surface trajectories. The map highlights tracer which finishes south of the (Rypina et al., 2016) front (black). A  $T_{dt=60days}$  matrix was used for both and the models were run for 25 years.

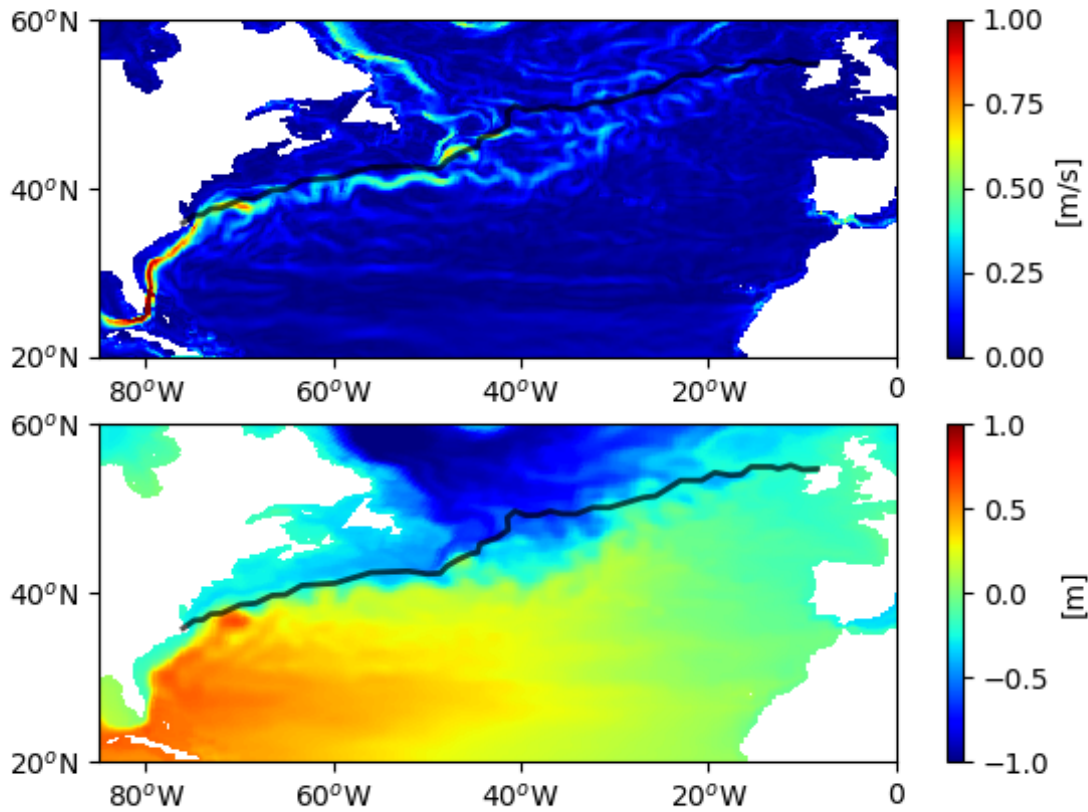


Figure 5.14: ORCA025 surface structure compared to GDP-derived front between the North Atlantic sub-tropical and sub-polar gyres. Top: Surface speed. Bottom: Surface ADT. The front derived by Rypina et al. (2011) is super-imposed on both maps. Both maps use the annual average of 2001 for their respective fields.

### 5.5.2 Along-isopycnal transport

While inter-basin transport within the Agulhas System is different for ORCA025, it was still possible to draw conclusions about the isopycnal transport in the  $\frac{1}{4}^\circ$  degree model. The same is true for the North Atlantic region. The Labrador Current remains a source of sub-tropical gyre tracer on both isopycnals (Figure 5.15). On the  $\sigma_0=27.1$  isopycnal, the Labrador Current tracer transport appears to decrease with time while on the 27.6 isopycnal it is maintained with time. Another notable difference is the apparent extent of the sub-tropical gyre, as defined by the separation between no-pole-ward and no-equator-ward transport. On the 27.1 isopycnal, the sub-tropical region shrinks southward with time. By 25 years, a large area between 40°W and 20°W sees less than 50% of tracer originating there remains in the sub-tropical gyre. Meanwhile,

a similar behaviour is seen on the 27.6 isopycnal but does not extend as far south.

To fully understand the connectivity along isopycnals at high latitudes, it is necessary to consider the quantity of tracer which has been consumed into the mixed layer (Figure 5.16). In the North Atlantic, equator-ward cross-frontal transport is not compensated by pole-ward transport alone. On the 27.1 isopycnal, 30-40% of tracer in the Labrador Current remains pole-ward of the front. Meanwhile, Labrador Sea water is completely consumed by the mixed layer. In the central part of the sub-tropical gyre, the quantity of tracer consumed increases northward. This mirrors the southward reduction in equator-ward tracer. This also aligns with the extent of isopycnal outcropping seen in Figure 5.1. On the 27.6 isopycnal, less tracer is consumed by the mixed layer. In the Labrador Sea roughly 30% is consumed, while the rest mostly remains within the sub-tropical gyre region. In the eastern part of the domain, the majority of tracer escapes mixed layer consumed and moves poleward.

The picture of connectivity changes considerably with depth and this is primarily caused by reduced interaction with the mixed layer (Figure 5.16). From the surface to the 27.1 isopycnals, the Labrador Current sub-tropical connection weakens and the entire Labrador Sea connection on the timescale of 5 years. By 25 years, there is no sub-tropical connection from the Labrador Sea on the 27.1 isopycnal. Sub-tropical and sub-polar gyres are therefore less well-connected along the 27.1 isopycnal. However, the inter-gyre connectivity along the deeper isopycnal is more similar to the surface. The 27.6 isopycnal outcrops further north than the 27.1 isopycnal and, as a result, the mixed layer consumption of tracer is confined to the northern part of the domain.

The interaction of the isopycnal tracer with the mixed layer is based on outcrop area and the details of the outcropping. On the 27.6 isopycnal, the mixed layer outcrops an area enclosed by the Labrador Sea and Current (see patch between 60°W and 40°W, north of 50°N, Figure 5.16). On the 27.1 isopycnal, the region where mixed layer consumption occurs spread across the ocean domain. Subtle changes in the velocity field of the deeper isopycnal may therefore drive tracer toward or away from the mixed layer intersection, changing the connectivity for the region. On the shallower isopycnal, tracer consumption is widespread and any change in

velocity within this region will not change the fate of tracer. These regional considerations will be useful when comparing new methods to extract velocity fields, or undertaking seasonal studies.

The connectivity pattern shown here will differ from the true ocean due to the resolution of the ORCA025 model. The trajectories used here are forced to flow along the isopycnal surfaces. The closest observation-based comparison is with RAFOS floats, which are designed to be isobaric and so do not always approximate potential density surfaces (Swift and Riser, 1994). Expected connections can instead be compared from similar depths to the isopycnals. Labrador Current water at 700m and 1500m is more likely to take gyre-interior pathways instead of following the equator-ward Deep Western Boundary Current (DWBC), with over 70% remaining in the sub-polar gyre (Bower et al., 2009, 2011). This agrees well with the tracer transport derived by ORCA025 models (Figures 5.15 and 5.16). However, their experiments using virtual particles in the FLAME model found re-circulation added to an increase in DWBC pathways on time-scales longer than 10 years. It is possible that ORCA025 more accurately depicts the relative importance of the DWBC. This point may be assessed by performing a connectivity analysis in a range of models.

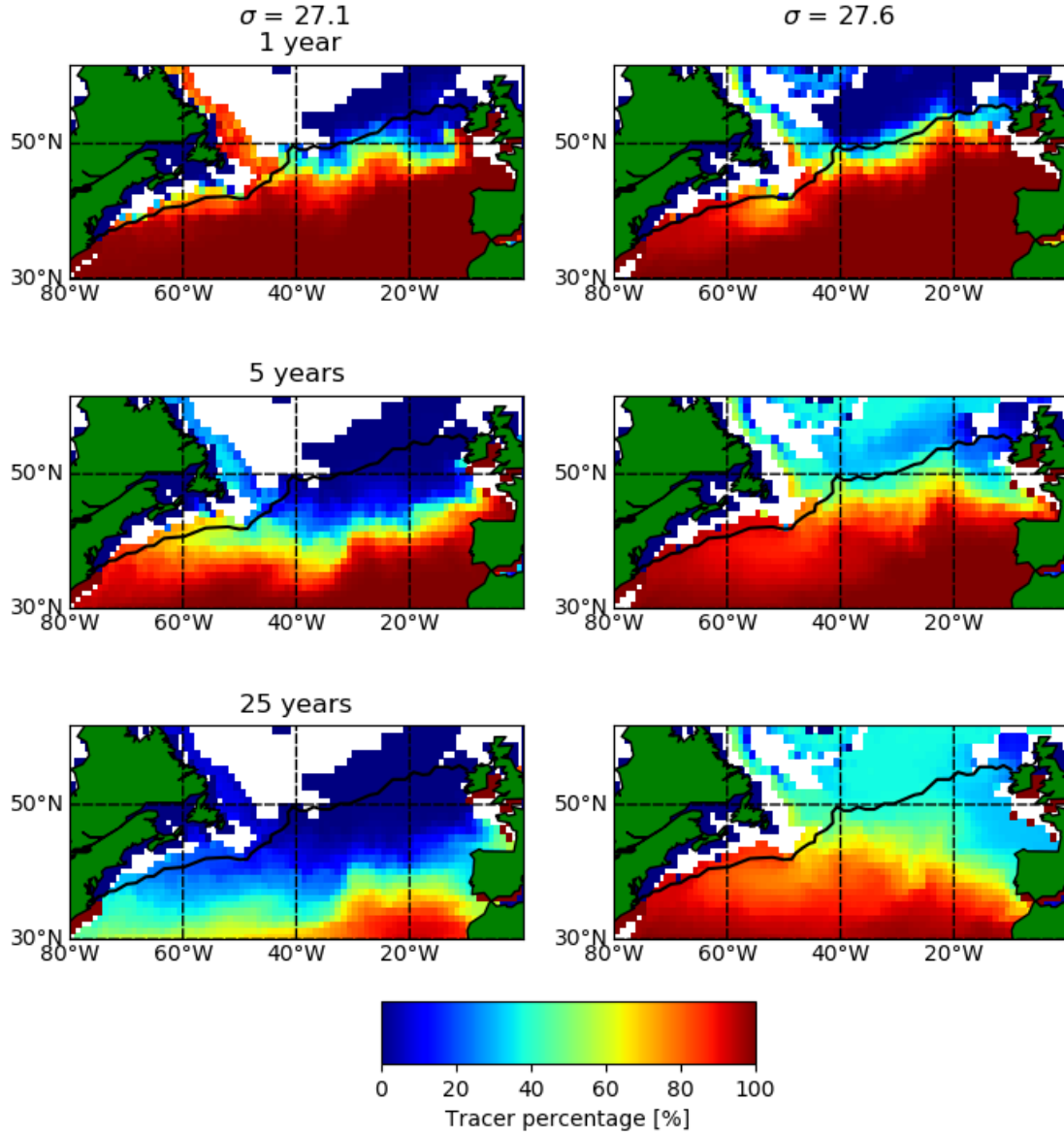


Figure 5.15: Cross-frontal connectivity map for ORCA025  $\sigma_0 = 27.1$  and  $\sigma_0 = 27.6$  isopycnal trajectories. The map highlights tracer which finishes south of the (Rypina et al., 2016) front (black). A  $T_{dt=60days}$  matrix was used for both and the models were run for 25 years. The blank cells indicate where the isopycnal intersects the mixed layer

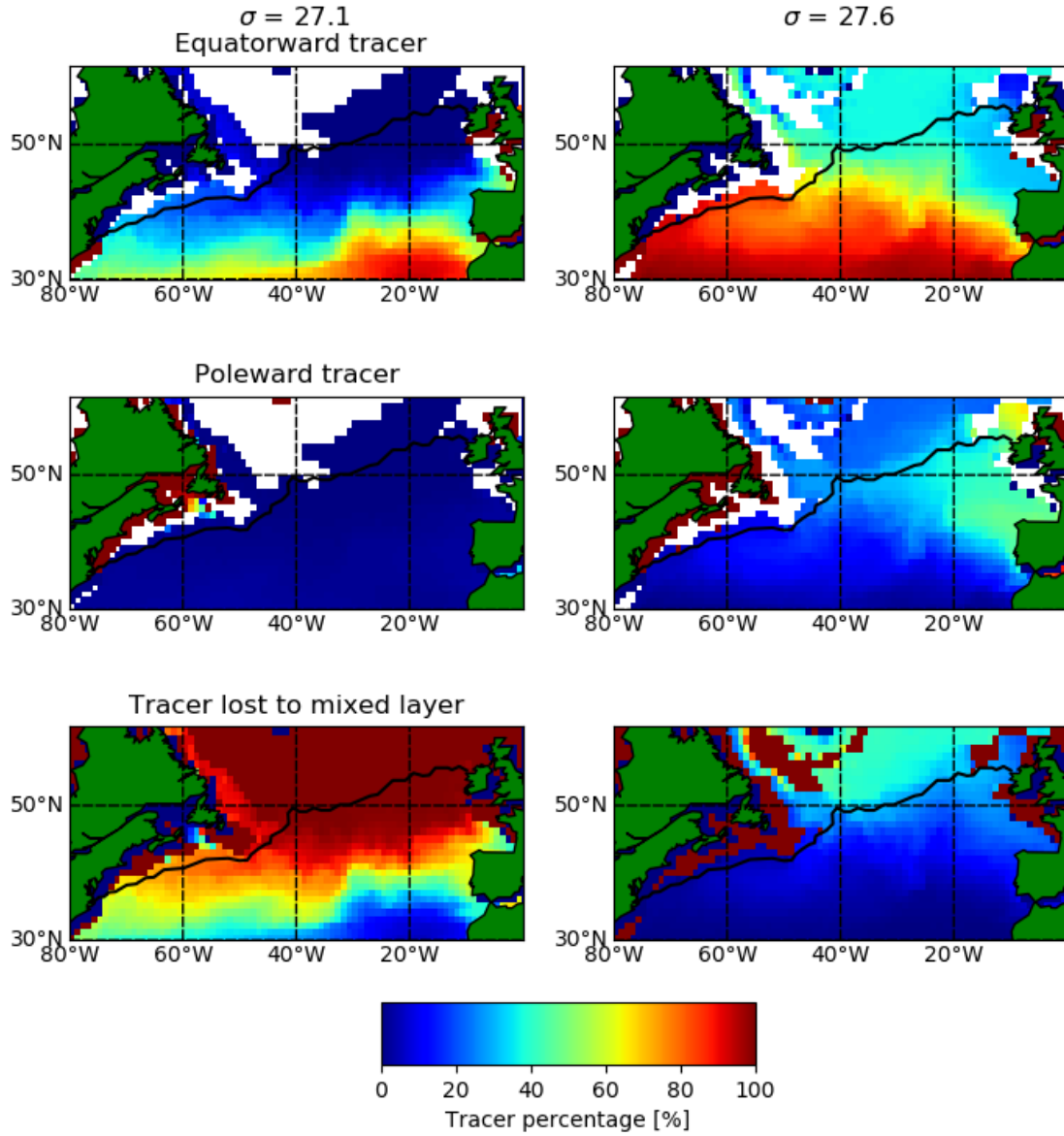


Figure 5.16: Fate of tracer in the North Atlantic region for ORCA025  $\sigma_0 = 27.1$  and  $\sigma = 27.6$  isopycnal trajectories after 25 years. The map highlights tracer which finishes south and north of the (Rypina et al., 2016) front (black). The bottom plots show the tracer consumed by the mixed layer. A  $T_{dt=60days}$  matrix was used for both and all maps are for 25 years after release.

## 5.6 Summary

In this chapter, inter-basin connections along two surfaces of constant potential density in a  $\frac{1}{4}^\circ$  model were identified with the Markov Chain model. The  $\sigma_0=27.1$  and  $27.6$  isopycnals were chosen because of the water masses which are advected along them. The  $27.1$  isopycnal is important for the shallow branch of NADW, MSW and the transfer of Indian Ocean water into the Atlantic. Meanwhile, the  $27.6$  isopycnal predominately captures NADW crossing into the South Atlantic. Both are important for global circulation. The North Atlantic serves as a source region for deep water and also contains a barrier between tropical and polar regions, while the Agulhas Current serves as a connector between two distinct basins and water masses. The connectivity in both regions, as well as the role they play in global circulation, changes with depth.

Horizontal velocity components were used to approximate the along-isopycnal velocities in order to create a first-order view of basin-wide connectivity. A densely-seeded virtual particle release was run in OceanParcels. Conclusions made about inter-basin connectivity are largely restricted to features present in ORCA025. The surface transport field does not capture the same extent of connectivity as drifter-based models, largely due to the model's inability to resolve mesoscale processes. Given that  $\frac{1}{4}^\circ$  ocean models are still used in climate model ensembles, knowledge of the connectivity in such models is useful in itself. In the absence of widespread observations of transport pathways at depth, use of an ocean model provides the sole way to produce the connectivity maps presented here.

Agulhas Leakage is present on the  $27.1$  and  $27.6$  isopycnals. The region from where Leakage-bound tracer originates, however, varies between surfaces. On the surface, this area covers the Agulhas Current and its tributaries, the East Mozambique and Madagascar Currents. On the  $27.1$  isopycnal, connectivity is spread throughout the recirculation region although the unresolvable recirculation processes are not contributing to this. Below this, the connectivity is very weak outside the Agulhas Current.

For along-isopycnal transport in the North Atlantic, the mixed-layer plays a key role in consuming water masses and redistributing them on different density surfaces. Here, rates of obduction and subduction are not taken into account (Qiu and Huang, 1995). However, the connectivity analysis highlights the extent of mixed layer interaction with the isopycnals of interest. Nonetheless, the pattern of connectivity is largely similar to the surface. The Gulf Stream acts to block northward flow while the North Atlantic current and the eastward extension of the front permits Labrador Current tracer to cross into the sub-tropical gyre.

Crucially, the full extent of cross-frontal transport is not depicted by along-isopycnal flow alone because water parcels experience changes in buoyancy. Early RAFOS float releases in the Gulf Stream show meandering causes along-isopycnal cross-frontal transport (Bower and Rossby, 1989). Water parcels, however, will be expected to experience a temperature change, and therefore a density change, particularly towards the North Atlantic Current (McCartney and Talley, 1982). In model runs, most cross-frontal trajectories experience a small (less than  $0.2 \text{ kg.m}^{-3}$  in 4 years) densification (Burkholder and Lozier, 2011). Nonetheless, the analysis presented here does not present the full, three-dimensional picture of connectivity. However, the along-isopycnal pattern of connectivity can be used as a comparison to fully 3D future studies to identify the role of the diapycnal component in inter-basin transport.

# Chapter 6

## Conclusions

The final Chapter highlights the main findings of Chapters 3, 4 and 5. In each section, there is discussion of the caveats of the findings, the contributions of connectivity theory to oceanography and the dynamics of the two regions studied. Finally, there is discussion of the implications for future work.

### 6.1 Markov Chain Modelling of the Ocean

The GDP drifter dataset has become considerably larger since previous studies (e.g. Maximenko et al., 2012), including those which cited a lack of coverage as a caveat to their findings (Reverdin et al., 2003). Throughout this thesis, GDP-based Markov Chain models have been compared to those constructed with larger datasets of virtual particles based on GlobCurrent products. The results suggest that a larger number of particles in altimeter-derived fields do not change the probabilistic representation of circulation. The size of the GDP dataset is seemingly sufficient in two intensely-studied regions, the North Atlantic and the Agulhas Region. However, the Indonesian Throughflow and much of the Southern Ocean remain relatively poorly sampled.

Markov Chain models have been used here to display the quantity, extent and time evolution of one region's connection to another. In this sense, this method is most similar to analysis by

Maximenko et al. (2012), which constructed basin boundaries based on the shortest-time path to gyre centres. The method used here involves repeating tracer releases throughout a domain and measuring the quantity which ends up in a particular destination. The benefit is that results are based directly on the quantity of transport pathways and are dynamical in nature. Because coverage varies between cells, the statistical significance and therefore the accuracy of the method may not be uniform (Koszalka et al., 2011). The regions of interest in this thesis are relatively well covered and Markov Chain models results vary little when using different cell sizes (Section 3.3).

An issue with the Markov Chain methodology is the exaggeration of dispersion (Rypina et al., 2016). Here, a rigorous sensitivity analysis was carried out to determine the mechanism for this artificial dispersion. Grouping trajectories within a cell causes an inherent spatial-averaging of dynamic features which are distinct to each other and do not affect each other over the scale of the cell. As shown in Chapter 3, in the analytical Stommel gyre, linear streamlines, which by definition do not intersect, were nonetheless grouped together. This causes dispersion in both the parallel and perpendicular directions to the transport. Because dispersion occurs at each time-step, shorter time-step models will exhibit greater artificial dispersion than longer time-step models.

While Markov Chain models of analytical flows are strongly affected by artificial dispersion, the real ocean is naturally dispersive. and the effect of artificial dispersion is not immediately evident. Artificial dispersion causes inaccuracy in determining transport proportions between regions. The sizes of accumulation regions of floating debris in the gyres can differ by up to 50% depending on the time-step used and the gyre of interest. However, it has not been possible to state which time-step model is most accurate or why. It may be that artificial dispersion is enhanced in certain areas, depending on the non-linearity of local trajectories and the resulting spurious connections. Steadily increasing the non-linearity of trajectories, from Stommel streamlines to GDP-like trajectories, using a stochastic diffusion term, would improve understanding the interaction between natural and artificial dispersion.

Overall, artificial dispersion in GDP-based Markov Chain models acts to “smooth” the sharpness of boundaries in connectivity maps. With regards to connectivity studies, the presence of artificial dispersion has not affected the conclusions drawn here about ocean dynamics. For example, the section of the Indian Ocean connected to the South Atlantic is observed in the Markov Chain models used here regardless of the time-step used.

## 6.2 Surface Connectivity

In Chapter 4, connectivity maps were used to identify the extent and strength of regional connections and to detect the physical drivers of this transport. To achieve this, Markov Chain models built from all available drifter trajectories were compared to models built from a sub-set of ADT-following drifters. This sub-set was used to approximate the geostrophic component of flow. Rather than conduct experiments based on relatively coarse altimeter-derived velocity fields, this thesis developed a methodology for identifying observed drifter trajectories whose transport is largely due to the geostrophic balance. The key benefit of this is that it allows direct comparison to reliable, full-field GDP-based Markov Chain models.

Chapter 3 discussed several reasons why drifters which remain within a specific ADT range over a given time are not solely influenced by the geostrophic balance: ageostrophic motion along ADT contours, the choice of ADT threshold and the effects of wind on drifters. LaCasce (2000) compared drifter trajectories to contours of  $f/H$  to display where topographic steering occurred. Likewise, comparing drifter pathways to ADT contours highlights where geostrophic motion steers the flow, and how this picture changes with timescale. Over seasonal timescales, drifters leave western boundary currents for parts of the ocean with lower ADT values. In this case, drifters are classified as “non-geostrophic”. A short-coming of this method is therefore that the geostrophic steering is only evident on shorter timescales.

The Agulhas Leakage has long been understood to be a major conduit for Indian and Pacific Ocean waters to enter the Atlantic. Here, it was shown that the recirculation in the area found between Madagascar and the Return Current contributes to inter-basin transport by supplying

the Agulhas Current. Tributaries such as the Madagascar and Mozambique Currents, as well as sub-mesoscale and geostrophic processes, all transfer waters into the Agulhas Current, at both upstream and downstream sections. Although this information is stored in the GDP dataset, the Markov Chain model makes clear the extent of this connectivity.

In the North Atlantic basin, a new definition for the surface boundary between the sub-tropical and sub-polar gyres is proposed, which differs from definitions which follow the core of the Gulf Stream and North Atlantic Currents (Rypina et al., 2011). In previous studies there has been a focus on northward transport across the Gulf Stream originating from the Florida Current (Brambilla and Talley, 2006; Hakkinen and Rhines, 2009). There have also been discrepancies between model studies and observations about the fate of Cape Hatteras water (Reverdin et al., 2003; Rypina et al., 2011). Here, the Gulf Stream waters and the entire sub-tropical gyre are shown to remain within this region south of the Gulf Stream and North Atlantic Current. New findings from this thesis show that the Labrador Current, bringing water from the Labrador Sea, and Cape Hatteras areas are also strongly connected to the sub-tropical gyre.

Cross-frontal transport is inferred to occur before the Gulf Stream morphs into the North Atlantic Current, which is also where the Labrador Current intersects the two. This is because there is no equator-ward cross-frontal transport further east of the bifurcation point. Differences between the full-field and geostrophic models show that Cape Hatteras water is transported by geostrophic processes, while Labrador Sea water is transport by both ageostrophic and geostrophic processes. There is no evidence to suggest whether they occur in proximity to each other. However, eddies formed from the Gulf Stream instabilities occur west of the intersection point. Warm-core eddies transport warmer sub-tropical water poleward. Roughly 20 warm-core eddies are formed each year (Liu et al., 2018) yet a similar number of equator-ward cold-core rings formed. Apparently, drifters are not transported with the warm-core rings. Another option is that equator-ward transport occurs anti-parallel to the Gulf Stream, between the current and the coast, and is entrained upstream.

The conclusions drawn for surface transport correspond strictly to drifter motion, which is divergent and buoyant. Water parcels differ from drifters in that they experience up- and

down-welling in gyre centres and, due to the changes in thermohaline properties, are forced into the ocean interior. The drifter-derived connectivity maps will therefore be relevant to the fate of floating debris and passive wildlife migration. However, the definition for a surface front for ocean circulation depends also on the vertical motions, which is not accounted for here.

## 6.3 Isopycnal Connectivity

In Chapter 5, physical definitions for ocean basins, as described for the surface, have been produced for the transport of intermediate water masses on the  $\sigma_0=27.1$  and 27.6 isopycnals. A  $\frac{1}{4}^\circ$  ocean circulation model was used to provide trajectory datasets which represent adiabatic motion in the ocean interior. The observational alternative, the Argo dataset, is restricted to one depth layer (1000m) and its floats do not strictly follow isopycnals. Moreover, using the ORCA025 allowed the connectivity analysis to assess a state-of-the-art model's ability to capture processes which drive connectivity.

Validation of the model was only possible on the surface by comparison of drifter-derived and ORCA025 Markov Chain models. It highlighted the lack of submesoscale processes in  $\frac{1}{4}^\circ$  OGCMs and their contributions to global circulation. In the Agulhas System, the absence of submesoscale dynamics reduces the extent of connectivity to the South Atlantic. Instead, Indian Ocean waters are more likely to return via the Return Current. Moreover, the inertia of the Agulhas Current is underestimated, a common symptom of inadequate resolution. In the North Atlantic, the Gulf Stream core is shifted southward by several degrees compared to observations and, again, inter-basin connectivity is not fully captured, particularly in the Labrador Sea.

The 27.1 and 27.6 isopycnals were targeted because of their differences in the key regions of interest. Patterns on T-S charts, which allow a basic isentropic analysis (Montgomery (1938)) are also explained by connectivity analysis of ORCA025 trajectories. Agulhas Leakage transport decays below the depth of the 27.1 isopycnal, leading to a reduction of warmer Indian Ocean water masses entering the South Atlantic on the 27.6 isopycnal. Moreover, the Agulhas Region

connectivity is significantly narrower on the 27.6 isopycnal. A surprising finding in the model connectivity was that westward drift in the recirculation is stronger on the 27.1 isopycnal than on the surface.

Meanwhile, on the 27.1 isopycnal, the adiabatic advection of water masses from the high-latitude regions is hindered by the extent of the mixed layer, which acts to re-work water parcels into other isopycnal layers. Nonetheless, the connectivity analysis highlights differences in the effective boundary between the sub-tropical and sub-polar gyres. The transport of MSW into the sub-polar gyre, which is greater on the 27.6 isopycnal, pushes the transition between the gyres further south. This is expected, as NADW extends further south on the 27.6 isopycnal and dominates the T-S properties throughout the Atlantic. The Labrador Current remains the sole conduit for equator-ward flow. Meanwhile, the Cape Hatteras connection to the sub-tropical gyre is not present on the surface nor the interior.

## 6.4 Future Work

The Markov Chain models of surface drifter motion used here represent time-averaged transport from over 30 years of data. Previous studies which isolate the physical drivers of transport highlight the seasonal variations and peaks in transport due to such drivers (Brambilla and Talley, 2006). To maintain ample coverage, this thesis focused on time-averaged flow. Time-averaged structures are crucial for understanding circulation and often provide the basis for schematics. A natural extension of this thesis would be to study seasonal variations of transport boundaries, using seasonal sub-sets of trajectories to detect when certain pathways are enhanced or specific to certain times of the year.

Chapter 5 focused on the adiabatic, along-isopycnal component of transport which is a simplified picture of interior circulation. It was shown that the mixed layer plays a key role in inter-basin connectivity, particularly on the shallow isopycnals. The surface area of the isopycnal which is shallower than the mixed layer varies seasonally. This is where the assumption that transport is largely adiabatic breaks down. Using seasonal sub-sets of trajectories would also be necessary

to accurately depict the rate of mixed layer consumption. Diapycnal transport is also enhanced where water masses interact and near the solid boundaries of the ocean interior (de Lavergne et al., 2017). Moreover, given that enhanced sloping has been shown to determine transport structure in the Agulhas as well as in the Atlantic, along-isopycnal transport alone does not describe the full picture of connectivity. As a result, the next logical step in interior connectivity analysis is to construct a three-dimensional Markov Chain model from trajectories which are not restricted to isopycnals but instead are also affected by diapycnal motion.

There are already methodologies for constructing three-dimensional transition matrices in oceanographic contexts (Khatiwala et al., 2005). This thesis has shown the changes in inter-basin borders with depth along 2D surfaces. A full 3D picture of, for example, the Agulhas Leakage boundary from Section 5.4 would display the region's role in transporting water at all depths. Such a map would display spatial connections at depth, accompanying inter-basin volume studies such as (Blanke et al., 2001). A comparison of isentropic analysis and a 3D analysis which included diapycnal motion would also highlight the role of the latter in regional connectivity. This aim should be approached with caution however, as 3D Markov Chain models may include numerical diapycnal diffusion.

A major consideration for inter-ocean exchange is the conservation of mass. In the North Atlantic, surface cross-frontal transport is found to be blocked from north to south but not in the other direction. Return flow to balance the mass transport across the front will not occur at the same depth level. Mass conservation in Markov Chain models can only be assessed when considering transport across all surfaces, i.e. in 3D. Moreover, transport quantities have been quantified in a probabilistic way with percentages of a tracer cloud given as the measure. This measure provides no information on the mass fluxes between areas, although it may indicate relative strengths or currents. The connectivity between regions may also be studied with Lagrangian-based mass flux calculations Blanke and Raynaud (1997). Such experiments would prove a further validation of the conclusions drawn from Markov Chain models.

In the absence of sufficient trajectory data in the ocean interior, this thesis has relied on a global circulation model. Further comparison with next-generation, higher-resolution models will be

necessary to determine whether current state-of-the-art models are sufficiently depicting inter-ocean dynamics. The interior velocity structure in ORCA025, although useful for analysing the connectivity maps, differs from the real ocean. For example, mooring studies show that the Agulhas Current expands laterally with depth (Bryden et al. (2005)), the Madagascar Current has pole-ward transport at the depths of the 27.1 and 27.6 isopycnals (Nauw et al. (2008)) and the Agulhas and Benguela Undercurrents are present at the depths of both isopycnals (Majumder and Schmid, 2018). These features were altered or missing in the ORCA025 model world. Therefore, near-future connectivity studies based on models cannot be compared rigorously with observations.

In recent years, the use of graph theory in oceanography has advanced alongside research into marine plastic pollution (Van Sebille, 2014; Froyland et al., 2014). Treating the ocean as a network provides a framework for highlighting origins and destinations of chaotic transport. New definitions of ocean basins are also relevant for inter-basin transport of pollution, highlighting where a particular country’s plastic is most likely to end up. Future work in the field of marine plastic pollution will require 3D particle tracking to determine the quantities of pollution which is submerged or sinking. Advancement in Lagrangian analysis will add sinking parameterisation terms to the advection of water-following particles (van Sebille et al., 2018). Previously, the Markov Chain model has been used to identify surface accumulation regions in the gyres. The next step is to identify regions where plastic accumulates at depth and on the sea floor. 3D Markov Chain models and connectivity maps based on plastic-like particle trajectories will highlight the “hot-spots” of plastic in the ocean interior.

The continued application of transition matrices to interior transport may also enhance our understanding of interior transports such as the thermohaline circulation. Anthropogenic climate change has led to an increase in the heat content of the upper and intermediate ocean (Collins et al., 2013). Dynamically-segregated areas in the interior may help detect regions where heat storage is enhanced, and highlight pathways for stored heat. Depicting the ocean circulation in thermohaline coordinates provides a means of detecting water masses which complements the traditional geographical picture of circulation (Groeskamp et al., 2019). Döös et al. (2012) and Zika et al. (2012) used this method to identify regions with distinct thermohaline characteris-

tics, such as the Southern Ocean as the freshest and coolest ocean basin. Connectivity maps of the ocean interior would provide a geographical complement to thermohaline analysis.

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