



UNIVERSITY OF CAPE TOWN
IYUNIVESITHI YASEKAPA • UNIVERSITEIT VAN KAAPSTAD

DEPARTMENT OF OCEANOGRAPHY

AN INVESTIGATION OF WIND- AND WAVE-DRIVEN DYNAMICS IN ANTARCTIC SEA ICE FROM MULTIPLE TYPES OF BUOY ARRAYS

A research thesis in the fulfilment of the requirements for a Doctoral degree in the field of
Physical oceanography

Prepared for: University of Cape Town

Author:
Miss Ashleigh Womack

Supervisor:
Prof. Marcello Vichi

February 2024

The copyright of this thesis vests in the author. No quotation from it or information derived from it is to be published without full acknowledgement of the source. The thesis is to be used for private study or non-commercial research purposes only.

Published by the University of Cape Town (UCT) in terms of the non-exclusive license granted to UCT by the author.

Dedication

This thesis is dedicated to my late mother

Jodianne Womack

who instilled in me so much love and inspired me to pursue my dreams,
and to my amazing parents, *Julie* and *Scott Womack*,
who loved me like their own and for which I will be forever grateful.

Plagiarism Declaration

- I know that plagiarism is wrong. Plagiarism is to use another's work and to pretend that it is one's own.
- Each significant contribution to and quotation in this document from the work or works of other people has been attributed and has been cited and referenced.
- This thesis is my own work.
- I have not allowed and will not allow anyone to copy my work with the intention of passing it as his or her own work.

Name	Student	Date	Signature
Ashleigh Womack	WMCASH001	February 2024	

Statement of inclusion of publications

I confirm that I have been granted permission by the University of Cape Town’s Doctoral Degrees Board to include the following publications in my Ph.D. thesis, and where co-authorships are involved, my co-authors have agreed that I may include the following publications:

Womack, A., Vichi, M., Alberello, A., and Toffoli, A.: Atmospheric drivers of a winter-to-spring Lagrangian sea ice drift in the Eastern Antarctic marginal ice zone. *Journal of Glaciology*, 1-15. <https://doi.org/10.1017/jog.2022.14>, 2022.

Womack, A., Alberello, A., de Vos, M., Toffoli, A., Verrinder, R., and Vichi, M.: A contrast in sea ice drift and deformation between winter and spring of 2019 in the Antarctic marginal ice zone, *The Cryosphere*, 18, 205–229, <https://doi.org/10.5194/tc-18-205-2024>, 2024.

Name	Student	Date	Signature
Ashleigh Womack	WMCASH001	February 2024	

Abstract

The advance and retreat of Antarctic sea ice is the largest annually recurring event on Earth, and a significant portion of its natural variability appears to be associated with synoptic variability. Therefore, attaining a comprehensive understanding of both the dynamic and thermodynamic processes governing sea ice is vital. Moreover, this is particularly crucial for the Antarctic marginal ice zone (MIZ), where sea-ice behaviour is highly complex and where the interactions between the atmosphere and ocean are more variable. However, the paucity of observations in this region has meant that these drivers are poorly understood and modelled – specifically, the data on sea-ice drift, deformation, and concentration. Therefore, this thesis aims to enhance our understanding of the dynamic nature and response of the MIZ sea ice to extra-tropical cyclones, at both the synoptic and sub-daily scales, as well as to waves propagating into the Antarctic ice cover. This was done using multiple types of buoy arrays, which were operational in different regions around Antarctica and during different seasons and years.

The thesis demonstrated that atmospheric (wind) forcing has a dominant physical control on ice drift. The transfer of momentum from winds at the lower (synoptic) frequencies has been determined to be the primary effect, with the initiation of inertial oscillations of sea ice as the secondary effect. Results indicated that there is a plausible correlation between the presence of cyclones and the onset of the inertial oscillations. However, it was also suggested that the penetration of storm-generated waves may trigger the inertial response of sea ice, or allow the underlying geostrophic currents to keep and sustain the weaker oscillations during the period of storm quiescence.

An investigation into wave-ice interactions revealed that storm-generated waves have enough energy to propagate > 200 km into the ice cover. In the process, they are attenuated at a rate governed by the nature of sea ice, while the energy transferred in doing so leaves behind a wake of broken ice floes. This concomitant relationship is known to promote sea-ice growth during the winter months, when air temperatures begin to decrease. Additionally, it was demonstrated that the same waves-in-ice activity contributes to the break-up and ultimate retreat of the sea-ice cover, when air temperatures increase once again, and large floes are unable to reform after cyclonic events. This in turn impacts the ice cover's susceptibility for drift and deformation by winds and ocean currents.

The evolution and spatial pattern of sea-ice drift and deformation are greatly affected by the balance between external forcing and local ice conditions. The results indicated that for sea ice close to the ice edge, the Antarctic Circumpolar Current (ACC) was able to provide a steady source of significant energy to the ice floes. This significantly modulated the relationship between ice motion and wind forcing, and caused the enhancement of the wind-driven drift. The resulting free-drift and deformation were primarily driven by large-scale atmospheric forcing. Remarkably, there was no discernable shift to the inertial frequency, in response to passing cyclones.

Conversely, ice floes that drifted further from the ice edge and within a less advective ice-ocean system, showcased the expected energy cascade in their velocity power spectral density (PSD). However, counterintuitively, their corresponding deformation PSD demonstrated a strong de-coupling to the large-scale atmospheric forcing. Further analyses, encompassing existing deformation datasets from different regions around Antarctica revealed that, for similar spatial scales, the magnitude of sea-ice deformation varies between seasons, regions and the proximity to the sea-ice edge or the coastline. The observed variability in sea-ice drift and deformation indicates that the Antarctic ice cover is seasonally and regionally dissimilar. However, this study further demonstrates that the local ice conditions (i.e. sea-ice rheology), were also notably different during each buoy deployment. While these findings may lead to conclusions describing the Antarctic sea-ice cover as a non-coherent structure and unpredictable; it is necessary to acknowledge the persisting limitations in the availability of data, especially in contrast to the Arctic. Consequently, despite being the first to present a comprehensive review of sea-ice dynamics across various temporal and spatial scales, this thesis contends that the scarcity of buoy arrays in the Antarctic hinders in making a definitive assertion about sea-ice rheology, and hence, the ice cover's response to external forcing.

Overall, this thesis confirms that the present concentration-based definition, where the sea-ice concentrations range between 15-80 %, is inadequate to describe the sea-ice cover in the Antarctic MIZ. There is an expectation that when sea ice attains 100 % coverage from space it becomes consolidated and subject to strong internal stresses. However, the thesis shows that the MIZ is not uniform and is rather comprised of a disarranged mixture of several ice types, with open-water fractions, that are continuously evolving. Additionally, free-drift conditions were found > 200 km from the ice edge during a winter expansion. This was a result of the complex interactions between sea ice, extra-tropical cyclones, and storm-generated waves.

Acknowledgements

First, I would like to express my deepest appreciation for my amazing supervisor, Prof. Marcello Vichi, who made me believe in myself and provided me with feedback regularly, allowing me to grow as a researcher beyond my own limits. I am truly grateful for your support and your invaluable advice played a key role in the materialisation of this thesis. You have played a momentous role in my life. To my unofficial supervisor, Dr. Alberto Alberello from the University of East Anglia, and my fellow students in the sea-ice team and in Oceanography, thank you for always being ready to lend a help hand. I am extremely grateful for all you have done for me.

Thank you to the Australian team from the University of Melbourne. Dr. Alessandro Toffoli, you have been an incredible help with all of my publications, and in particular thank you and your team with providing guidance and data for my Chapter 4. To the Sound and Vibrations team from Stellenbosch University, I cannot thank you all enough for providing us with the ship motion response data used in Chapter 4.

This thesis would not have been possible without the financial support I received from the European Union's Horizon 2020 research and innovation programme under grant agreement no. 101003826 via project CRiceS (Climate Relevant interactions and feedbacks: the key role of sea ice and Snow in the polar and global climate system). Thank you for your support.

Thank you to the Department of Science and Innovation (DSI) for funding the Antarctic expeditions aboard the S.A. Agulhas II, and for the Department of Environment, Forestry and Fisheries (DEFF) for facilitating them. I would also like to thank the South African National Antarctic Programme (SANAP) and the Captains and crew aboard the S.A. Agulhas II for making the Antarctic expeditions possible.

To my parents who have been my biggest fans: Julie and Scott, thank you for your undivided support during this long journey. Words cannot describe how incredibly fortunate I am for you in my life. To the rest of my family, I am incredibly thankful for your love and constant encouragement.

To Jonathan Rogerson, you were always there for any problem I've had during the past few years, big or small, you were always there to help. So, from the bottom of my heart, thank you.

Last, but certainly not least, thank you to my incredible community of friends. Having you all in my life has been an invaluable source of support, laughter, and love. A special acknowledgement goes to Sina Wallschuss, who has been my rock for the past nine years. Your support has been a huge driving force behind my work, and I am immensely grateful for you.

Preface

This thesis presents the research carried out during my Ph.D. candidature at the University of Cape Town. It includes five main chapters, and is arranged by type of external (wind, wave and oceanic) forcing, scale and complexity. Figure i, below, outlines the how all of these components are arranged and interconnected.

The first chapter provides the required background and introduction to the overall study area, emphasising its validity to several components of the global ocean and climate system. In doing so, it presents the argument for all analyses and investigations which proceed it. The last chapter presents all key findings of this work, and contemplates recommendations for future work. The conducted research is presented in the other three chapters.

Chapter 2 provides an investigation of the influence of atmospheric forcing, specifically ice-landing synoptic cyclones, on sea-ice drift in the Antarctic MIZ. Additionally, it scrutinises the validity of the concentration-based definition for the Antarctic MIZ. This chapter is based on work published as:

Womack, A., Vichi, M., Alberello, A., and Toffoli, A.: Atmospheric drivers of a winter-to-spring Lagrangian sea ice drift in the Eastern Antarctic marginal ice zone. *Journal of Glaciology*, 1-15. <https://doi.org/10.1017/jog.2022.14>, 2022.

Chapter 3 extends the work done in Chapter 2 to provide a better understanding of sea-ice dynamics at the synoptic and sub-daily scales over two 2019 seasons (winter and spring), and in different regions around Antarctica. This chapter is based on work published as:

Womack, A., Alberello, A., de Vos, M., Toffoli, A., Verrinder, R., and Vichi, M.: A contrast in sea ice drift and deformation between winter and spring of 2019 in the Antarctic marginal ice zone, *The Cryosphere*, 18, 205–229, <https://doi.org/10.5194/tc-18-205-2024>, 2024.

Chapter 4 investigates the relationship between wave-ice interactions of various ice types during the 2019 sea-ice formation and retreat phases. This analysis encompasses the dissipative nature of waves through the ice cover and of its concurrent impacts on sea-ice type, which in turn effects sea-ice dynamics. This work is the result of a collaboration with scientists from the University of Melbourne and the University of East Anglia.

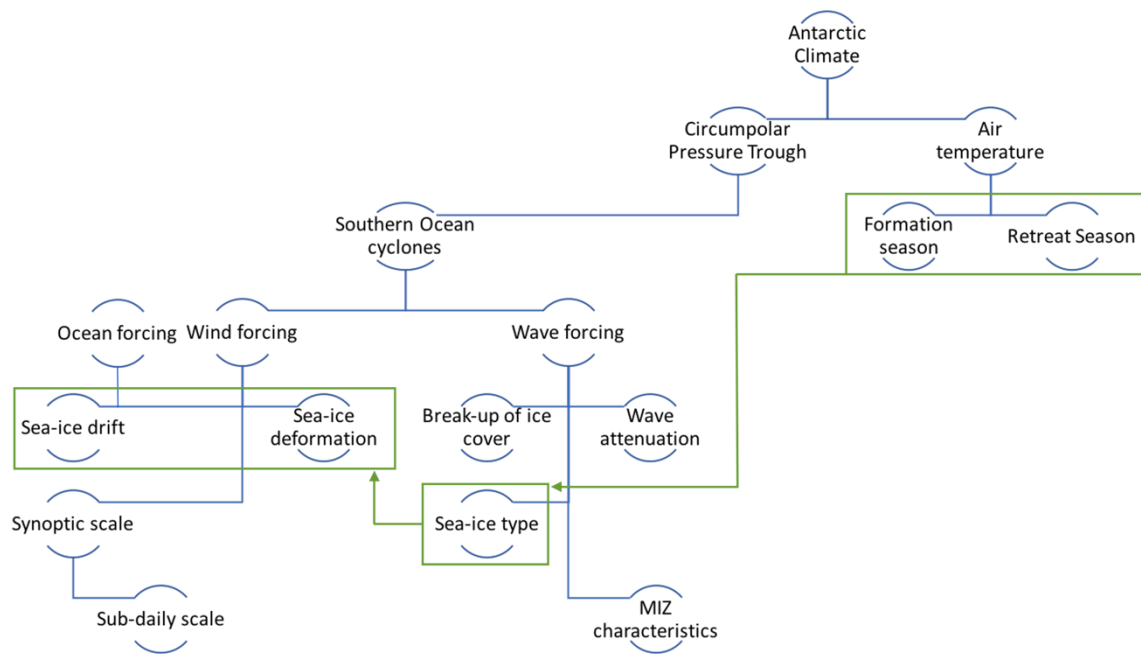


Figure i. The interconnected dynamics and thermodynamics of Antarctic sea ice investigated in the thesis.

Table of Contents

Plagiarism Declaration	iv
Statement of inclusion of publications	v
Abstract	vi
Acknowledgements	viii
Preface	ix
Table of Contents	xi
List of Figures	xiv
List of Tables	xxi
List of Abbreviations	xxii
Chapter 1	1
1. Introduction	1
1.1. Background to study	1
1.1.1. The Antarctic region	1
1.1.2. The lifecycle of Antarctic sea ice	2
1.1.3. Southern Ocean cyclones	5
1.1.4. Atmospheric and oceanic forcing	7
1.1.4.1. Wind forcing	7
1.1.4.2. Wave forcing	8
1.1.4.3. Ocean current forcing	9
1.1.5. Sea-ice dynamics	11
1.1.5.1. Momentum balance equation	11
1.1.5.2. Motion of sea ice	12
1.1.5.3. Free-drift motion	14
1.2. Problem identification	15
1.3. Aims and Objectives	16
1.4. Overview of data and methods	17

Chapter 2	19
2. Atmospheric drivers of a winter-to-spring Lagrangian sea-ice drift in the Eastern Antarctic marginal ice zone	19
Abstract	19
2.1. Introduction	20
2.2. Methods	22
2.2.1. Field measurements and post processing	22
2.2.2. Environmental conditions	24
2.2.2.1. Synoptic cyclones identification	25
2.2.2.2. Estimated significant wave height	28
2.2.2.3. Correlation between buoy drift and wind forcing	29
2.3. Results: drift measurements and analysis	31
2.3.1. The impact of cyclones on sea-ice drift	31
2.3.2. Distance from ice edge	32
2.3.3. Metocean drivers	34
2.3.4. Spectral analysis	40
2.4. Discussion and conclusions	42
Chapter 3	46
3. A contrast in sea-ice drift and deformation between winter and spring of 2019 in the Antarctic marginal ice zone	46
Abstract	46
3.1. Introduction	47
3.2. In-situ observations and environmental data	50
3.3. Methods	53
3.3.1. Extra-tropical cyclone identification	53
3.3.2. Buoy kinematic parameters	53
3.3.3. Wind factor and ocean current's drift	54
3.3.3.1. Comparison of residual currents and the Copernicus GlobCurrent database	55
3.3.4. Lagrangian measures for dispersion and deformation assessment	57
3.4. Results	58
3.4.1. Spatial and temporal changes in atmospheric conditions	58
3.4.2. General drift and meandering	60
3.4.3. Deformation rate estimates	64
3.4.4. Metocean drivers of sea-ice drift	65
3.4.5. Winter-spring differences in the frequency domain	67
3.5. Discussion	70
3.5.1. Detecting ice type from drift features	70
3.5.2. Seasonal and regional comparison of the deformation proxy	72
3.6. Conclusions	75

Chapter 4	77
4. Effects of wave propagation into the Antarctic marginal ice zone during winter and spring of 2019	77
Abstract	77
4.1. Introduction	77
4.2. Field data and methods	80
4.2.1. Records of ship motion and sea states	80
4.2.2. Sea-state estimation from ship motion	82
4.2.3. Response amplitude operator	86
4.2.4. RAO calibration with satellite data	88
4.2.5. Environmental data	88
4.2.6. Waves-in-ice buoys	89
4.3. Results and discussion	93
4.3.1. Calibration of the reconstructed significant wave height	93
4.3.2. Reconstructed waves parameters in the MIZ	94
4.3.3. Comparison between SWIFT buoys and reconstructed parameters	100
4.4. Summary and conclusions	103
Chapter 5	105
5. Key findings and outlook	105
Chapter 6	110
Appendix A	110
Appendix B	115
Appendix C	128
References	131

List of Figures

Chapter 1

- 1.1. Photographs showcasing **(a)** grease ice, **(b)** small pancake ice floes and interstitial frazil ice, **(c)** larger and rafted pancake floes, and **(d)** consolidated first-year ice. These photographs were taken by the South African Weather Service during the SCALE 2019 Winter Cruise aboard the S.A. Agulhas II (de Vos et al., 2019). 4
- 1.2. Stages of the processes driving the asymmetric annual cycle of sea-ice extent in the Southern Ocean. The blue line denotes the monthly mean latitude of the freezing 2-m air temperature (-1.8°C). The black line denotes the monthly mean latitude of the sea-ice edge. The orange line denotes the monthly mean latitude of the CPT. The maximum SAO signifies the period when the SAO causes the CPT to be the most intense and located closer to Antarctica. Figure 1 of Eayrs et al. (2020). 6
- 1.3. Typical force balance for sea-ice drift in a two-dimensional system. This figure has been adapted from Weiss (2013) and Hunkins (1975), but for the southern hemisphere, where Coriolis force acts to the left of sea-ice velocity. Theta (θ) denotes the turning angle between ice drift and wind stress..... 13
- 1.4. Schematic of the locations where each buoy array was deployed and in which year. 17

Chapter 2

- 2.1. The ice conditions during the deployment of the buoy on a pancake floe using the ship crane on 5 July 2017. The diameter of the basket is 1.5 m. Photograph taken by Sebastian Skatulla. 21
- 2.2. The full trajectory of the buoy (time gradient trajectory from light blue to dark blue) from 5 July to 1 December 2017. The shadings show the AMSR2 SIC with values between 0 %-100 % on 1 December. The green circle denotes the end of the advance season on 30 September. The red circle denotes the beginning of the melt season on 1 November. The green line indicates the ice edge at 0 % ice concentration for 30 September, the maximum northward extent of sea ice. The three red boxes indicate the three loops within the buoy's trajectory. 23
- 2.3. Three case studies of large loops in the drift trajectory of the buoy for **(a-b)** 9-20 August; **(c-d)** 26 September-12 October; **(e-f)** 21-31 October. The gradient blue lines in the left column are colour correlated to the time gradient colour in Fig. 2.2, and denote the buoy's trajectory, moving from left to right. The shadings show the AMSR2 sea-ice concentration between 0 %-100 % on the last day for each corresponding meander. The dark-blue points on the right column panels denote the

	position of the buoy at 12:00 pm every day, for approximately ten days; the light-blue points show when each cyclone was closest to the buoy using the labels from Table 2.1.	27
2.4.	(a) The time series of the attenuation distance and its error propagation shaded in blue for the WIIOS from 4-19 July 2017 and (b) its corresponding distribution (blue) and the distribution when the 25th percentile is removed (orange). Both corresponding medians are denoted.	28
2.5.	Latitudinal location of the buoy (green line) and the ice edge, defined as the > 0 % sea-ice concentration contour, along the same longitude, for each day of the buoy's trajectory. The dark blue and light blue lines denote the latitude of the AMSR2 ice edge and 80 % contour, respectively. The red and pink lines are from the SSMIS sea-ice product. The red star symbols indicate the time when the storm cores were closest to the buoy's location, clustered in the three sets listed in Table 2.1. The green box denotes the five storms previously examined by Vichi et al. (2019) and Alberello et al. (2020). The right axis is scaled to show the relative distance of each line from the initial point of deployment.	33
2.6.	The meridional drift velocity of the estimated sea-ice edge (orange) and the buoy (blue). This does not include the last few days when the ice edge started melting (from ≈ 15 November). The grey horizontal line indicates 0 m s^{-1} . The other graphical elements are as described in Fig. 2.5.....	34
2.7.	Velocity components of the buoy (on the left axis) and 10-m wind from the ERA5 reanalyses (on the right): (a) zonal component and (b) meridional component. The other graphical elements are as described in Fig. 2.5.	35
2.8.	Time series of the meander coefficient (shown only from 5 August just before the first large loop until 1 December which was the last day of transmission). The dark-blue line denotes the progressive meander coefficient and the light-blue line denotes the daily meander coefficient. The horizontal grey line at $M = 1$ denotes straight-line drift. The red star symbols indicate the time when the storm cores were closest to the buoy location. The three red rectangular boxes denote the three sets of cyclones indicated in Table 2.1.	36
2.9.	The absolute (single-particle) dispersion statistics for the buoy, from 5 August just before the first large loop until 1 December which was the last day of transmission, depicting the zonal (purple line), meridional (magenta line) and total (cyan line) dispersion. The three red rectangular boxes denote the three sets of cyclones indicated in Table 2.1.	37
2.10.	Scatterplots (left) of ice drift speed as a function of wind speed and (right) of ice drift direction as a function of wind direction. The black line is the 1:1 line. The green line is the turning angle at -19.83°	38
2.11.	Time series of (a) the modelled H_S at the ice edge from ERA5 (orange line) and (b) the estimated H_S at the buoy's position for the four-month drift period. (blue line). The measured H_S of the WIIOS for the first 16 days of the buoy's drift is denoted by the green line. The blue and orange shaded areas denote the H_S of the buoy calculated using the attenuation coefficient ranges from Kohout et al. (2020) for ice concentrations ≤ 80 % and > 80 % respectively. All the other graphical elements are as in Fig. 2.5.	39
2.12.	Power Spectral Density corresponding to the (a) zonal component of ice drift and the (b) meridional component of ice drift of the buoy for four months. The blue line indicates the buoy,	

	and the orange line indicates the ERA5 winds. The vertical red line indicates the peak associated with inertial oscillations (13.47 hours).	41
2.13.	The Wavelet Power Spectrum (left panel) and the Wavelet Spectrum (right panel) of the filtered velocity spectrum of the buoy for its four-month drift, where (a) denotes the zonal component and (b) the meridional component. The red line indicates the cone of influence (left panel), the black contours (left panel) and black dashed lines (right panel) indicate the 95 % significance level. The yellow-dashed line (left panel) denotes the inertial frequency (13.47 hours). All the other graphical elements are as in Fig. 2.5.	42
Chapter 3		
3.1.	The sea ice conditions and deployment of a buoy during (a) the 2019 Winter cruise, on a pancake ice floe, using the ship's crane, and (b) the 2019 Spring cruise on consolidated ice floes. The inset in (b) depicts the frames specifically designed around the standard ISVPs for the 2019 Spring cruise. The diameter of the basket is 1.5 m. The inset is credited to Mardene de Villiers.	50
3.2.	(left-hand side) The mean Copernicus GlobCurrent speed for the deployment period of the Winter buoys (27 July 2019 to 5 October 2019). The trajectories of the three Winter buoys are shown. The black contour denotes the AMSR2 0 % sea ice concentration on 30 September 2019 – the date of approximate austral sea ice maximum. The cyan rectangular box indicates the region used to compute the distribution of the zonal and meridional currents. (right-hand side) The distribution of the spatial mean of the GlobCurrent zonal and meridional currents. The vertical lines denote the magnitudes of the winter residual currents.	56
3.3.	Time series of the ERA5 2 m air temperature (blue) and mean sea level pressure (green) at the location of ISVP 1 in winter (a) , IVSP 4 (b) and ISVP 6 (c) in spring. The yellow stars denote when the winter cyclones were closest to the buoys. The red stars denote when the spring cyclones were closest to ISVP 4. The orange stars denote when the same spring cyclones were closest to the main spring cluster deployed $> 5^{\circ}$ (> 400 km) east of ISVP 4.	59
3.4.	(a) Trajectories of the 2019 Winter buoys. The black (grey) contour denotes the AMSR2 (SSMIS) 0 % sea ice concentration on 30 September 2019 – the date of approximate austral sea ice maximum. The colour circles denote the start position of each buoy. The corresponding colour crosses of ISVP 1 and ISVP 3 denote their positions on 30 September 2019. The termination point of ISVP 2 (25 August 2019) is denoted by the red triangle. (b) Same as (a) but for the 2019 Spring buoys. The colour crosses and concentration contours are for 2 December 2019 – the day when the Trident buoy stopped transmitting data. (c) The operational periods of all buoys in this study, with corresponding colours to the time gradient maps. The deployment and final longitudes for each buoy are included above all operational periods.	61
3.5.	(a-b) Time series of the daily meander coefficient for the duration of the buoys' deployment, where the vertical markers are indicative of the daily time interval and the horizontal grey line at 1 denotes straight-line drift. (c-d) Time series of the absolute dispersion for the period when all buoys were transmitting together (≈ 30 days for both seasons). (e-f) Time series of the TSEs for the duration of the buoys' deployment. The Winter buoys are shown on the left and the Spring buoys on the right. The star symbols are the same as in Fig. 3.3.	62

3.6.	(a) Time series of the standard deviation σ_D of the deformation rate $\dot{D}$ for the Winter buoys (computed from the dispersion of buoy pairs for ≈ 30 days for both seasons). (b) Same as (a) but for the Spring buoys. L_0 denotes the initial length scale of each group of buoys. The stars are the same as in Fig. 3.3.	64
3.7.	Velocity components of the buoys (on the left axis) and 10 m wind from the ERA5 reanalyses (on the right axis): the zonal and meridional component of ISVP 1 in winter (a-b), of ISVP 3 in winter (c-d), and of ISVP 4 in spring (e-f). The star symbols are the same as in Fig. 3.3. Note that ERA5 wind magnitude is similar for the ISVP 1 and ISVP 3 locations.	65
3.8.	(a) Power spectral density corresponding to the zonal and meridional components of ISVP 1, in winter, and ERA5 wind velocity. (b) Same as in (a) but for ISVP 3, in winter. (c) Same as in (a) but for ISVP 4, in spring, where the black vertical line indicates the peak associated with inertial oscillations at 14.22 hours.	67
3.9.	(a) The wavelet power spectrum with time (left panel) and wavelet spectrum (right panel) of the filtered velocity magnitude spectrum of ISVP 1, during winter. (b) Same as (a) but for ISVP 3, during winter. (c) Same as (a) but for ISVP 4, during spring. The red line indicates the cone of influence (left panel), the black contours (left panel) and black dashed lines (right panel) indicate the 95 % significance level. The star symbols are the same as in Fig. A1. Due to the difference in the intensity of the power between the seasons, the ranges are different between the Winter buoys and the Spring buoys.	69
3.10.	(a) Power spectral density of the winter deformation proxy σ_D . (b) Same as in (a) but for spring, where the red vertical line indicates the peak associated with inertial oscillations at 13.57 hours). L_0 denotes the horizontal spatial scale for each cluster of buoys. The line colours correspond to the σ_D time series in Fig. 3.6a and b.	73
3.11.	The standard deviation σ_D of the deformation rate $\dot{D}$ versus the time interval, in days from start of analysis, for buoys from the 2017 Winter Cruise, 2017 PIPERS winter campaign, and the 2019 Winter and Spring Cruises. Each cluster of buoys has been denoted below, with their corresponding initial horizontal spatial scales (L_0).	74

Chapter 4

4.1.	Schematic of the S.A. Agulhas II's transect to the Antarctic MIZ during (a) 2019 winter, and (b) 2019 spring. The inset in (a) shows a closer view of the ship track within the MIZ. The blue line denotes the ship's transect. The black line denotes the AMSR2 sea-ice edge (0 % ice concentration) for the day that the ship entered the MIZ (26 July (a) and 22 October (b)), and the grey line denotes the day the ship left the MIZ (28 July (a) and 8 November (b)). The black and grey circles denote the ships position on the corresponding days. The deployment position of the two SWIFT buoys (20 and 21; light green and teal, respectively) are shown. The deployment position of three Winter buoys (a) and four Spring buoys (b), analysed in Chapter 3 are also shown (colours correspond to Fig. 3.4).	81
------	--	----

4.2.	(a) Open-ocean example of the measured vertical acceleration at Monkey Island, for 20 minutes, during the Winter Cruise, and (c) the corresponding vertical displacement after integration. (b) Same as in (a) but for the Spring Cruise. (d) Same as in (c) but for the Spring Cruise.	83
4.3.	Power spectral density of the measured response of the 20-minute chunk, at Monkey Island, from Fig. 4.2, during (a) winter and (b) spring.	84
4.4.	Schematic of the S.A. Agulhas II – by Soal et al. (2015). The hull of the ship lies below the green horizontal line. The design draught of the ship (7.65 m) is found below the waterline (blue line). The freeboard is found above the waterline.	87
4.5.	The response amplitude operator (RAO) for the heave, with variable drafts. The optimal draught for both seasons is 15.5 m and is denoted by the thick black line.	87
4.6.	Assessment of ECMWF atmospheric analysis product (ERA5) with in-situ observations from the ship, during in-ice activities. Data presented here are the time series of (a-b) the air temperature (in °C) and mean sea level pressure (in hPa), and (c-d) the 10-m wind speed (in m s ⁻¹). The ship's data (from the on-board scientific data system, SDS) are interpolated to an hourly time interval for comparison with the hourly ERA5 reanalysis data. Winter is shown on the left-hand-side, and spring is shown on the right-hand-side.	89
4.7.	Trajectories of SWIFT 20 (light green) and SWIFT 21 (teal) in (a) winter and (b) spring. The inset in (a) shows closer view of the SWIFT buoys' winter trajectories. The corresponding coloured circles along the buoys' trajectories denote their daily positions. The black (grey) contour denotes the AMSR2 0 % sea-ice concentration on the day that the S.A. Agulhas II entered (exited) the MIZ. These dates are the same as in Fig. 4.1.	90
4.8.	(a-b) Time series of the daily latitudinal location of SWIFT 20 (light green) and SWIFT 21 (teal) and the daily ice edge, defined as the > 0 % SIC contour along the same longitude, for each day of the buoys' drift. The black line (dashed) corresponds to SWIFT 20 (SWIFT 21). (c-d) Time series of the relative distance between the buoys and the ice edge, for each day. The horizontal grey line denotes 0 km. (e-f) Time series of the AMRS2 SIC co-located to the buoys' position. (g-h) Time series of the measured significant wave height. (i-j) Time series of the measured mean wave period. The green shaded areas (right-hand-side) denote when SWIFT 20 left the ice cover ship (from on-board cameras). The Winter Cruise is shown on the left, and the Spring Cruise on the right. The red stars denote the spring cyclones previously analysed in Chapter 3.	91
4.9.	Low resolution photographs of sea-ice conditions from SWIFT 20's on-board GoPro camera (≈ 1 m above the surface) during spring – from deployment until it exited the sea-ice cover.	92
4.10.	Low resolution photographs of sea-ice conditions from SWIFT 21's on-board GoPro camera (≈ 1 m above the surface) during spring – from deployment until it was retrieved.	93
4.11.	(a-b) Root Mean Square Error (RMSE) showing the differences between the significant wave height estimated from the ship's motion response and satellite observations (red points) and ERA5 reanalysis (orange points), as a function of the draught. (c-d) Correlation Coefficient (r) relative to (a-b). (e-f) The time series of the mean reconstructed significant wave height across all computations of the RAO (blue points), at each time step, with the standard deviation range (light-blue). The corresponding satellite and ERA5 data at the ship's location is also shown. The blue	

shaded regions denote the period when the ship was in ice. The Winter Cruise is shown on the left and the Spring Cruise is shown on the right.	94
4.12. (a) Time series of the ERA5 2 m air temperature (blue) and mean sea level pressure (green) at the location of the S.A. Agulhas II, during winter. (b) Same as in (a) but for spring. (c-d) Time series of the corresponding ERA5 10-m wind speeds in each season. The blue shaded areas denote when the ship was in ice (Ryan-Keogh and Vichi, 2022). The pink stars denote when the cyclones, previously analysed in Chapter 3, were closest to the ship.	95
4.13. Daily photographs of the local sea-ice conditions during the Winter Cruise taken from the S.A. Agulhas II. The dates and time stamps for each photograph are shown.	96
4.14. (a-b) Time series of the latitudinal location of the S.A. Agulhas (blue line) and the daily ice edge, defined as the > 0 % SIC contour (black line), along the same longitude, for each day of the ship's transect. (c-d) Time series of the relative distance between the ship and the ice edge, for each day. The grey dashed line denotes 0 km. (e-f) Time series of the AMRS2 SIC co-located to the ship's position. The grey crosses denote the observed SIC, estimated to the nearest 10 % (Hepworth et al., 2020). (g-h) Time series of the reconstructed significant wave height measured along the ship's transect. (i-j) Time series of the reconstructed mean wave period, with the inclusion of the peak wave period (red points). The light green and teal circles denote the corresponding values measured by SWIFT 20 and SWIFT 21, respectively, during their deployment and retrieval. The blue shaded areas denote when the ship was in the ice cover (Ryan-Keogh and Vichi, 2022). The red vertical line denotes when the ship began travelling east. The orange vertical line denotes when ship travelled back west. The Winter Cruise is shown on the left, and the Spring Cruise on the right. The pink stars denote when the cyclones, previously analysed in Chapter 3, were closest to the ship.	97
4.15. Daily photographs of the local sea-ice conditions during the Spring Cruise taken from the S.A. Agulhas II. The dates and time stamps for each photograph are denoted. Note, there are missing days.	99
4.16. Time series of the significant wave height during (a) the 2019 Winter Cruise and (b) during the Spring Cruise, as estimated from the ship's motion response (blue points), and measured from SWIFT 20 (light-green line) and SWIFT 21 (teal line). (c) Same as in (b) but showing a closer view of reconstructed wave height, and of SWIFT 21's wave height. (d-e) Time series of the mean wave period during the Winter Cruise and Spring Cruise, respectively. The blue shaded areas denote when the ship was in the ice cover (Ryan-Keogh and Vichi, 2022). The green shaded areas (right-hand-side) denote when SWIFT 20 left the ice cover ship (from on-board cameras). The red vertical line denotes when the ship began travelling east. The orange vertical line denotes when ship travelled back west. The pink (red) stars denote when the cyclones, previously analysed in Chapter 3, were closest to the ship (SWIFT buoys).	101
4.17. Evolution of the environmental conditions during the period when the S.A. Agulhas II travelled through the sea-ice cover. Each panel illustrates the ERA5 mean sea level pressure (black lines, interval 5 hPa), ERA5 significant wave height (coloured shading, interval 1 m), and the AMRS2 0 % SIC – the ice edge (red line). The ship track (blue line), and SWIFT 20 (light-green line) and	

SWIFT 21 (teal line) are shown from their starting position to their current position, denoted by the circles. All data are at 12:00 GMT. 102

List of Tables

Chapter 2

2.1.	Features of the cyclones which occurred during the loops as simulated by ERA5.	31
------	---	----

Chapter 3

3.1.	Operational details of the Winter and Spring 2019 buoys. Herein, all dates and times are Greenwich Mean Time (GMT).	51
3.2.	Parameters describing wind and ocean forcing on sea ice drift.	66

List of Abbreviations

Abbreviation	Definition
ACC	Antarctic Circumpolar Current
ACoC	Antarctic Coastal Current
AMSR2	Advance Microwave Scanning Radiometer 2
ARTIST	Arctic Radiation and Turbulence Interaction STudy
ASI	ARTIST Sea Ice
ASPeCt	Antarctic Sea Ice Processes and Climate
coi	cone of influence
CPT	Circumpolar Pressure Trough
CRiceS	Climate Relevant interactions and feedbacks: the key role of sea ice and Snow in the polar and global climate system
ECMWF	European Centre for Medium-Range Weather Forecasts
FFT	Fast Fourier Transform
GMT	Greenwich Mean Time
GPS	Global Positioning System
IMU	inertial measurement units
IOBL	ice-ocean boundary layer
ISVP	Iridium Surface Velocity Profiler
NSIDC	National Snow and Ice Data Center
Metocean	meteorological and oceanographic
MIZ	marginal ice zone
mslp	mean sea level pressure
PIPERS	Polynyas, Ice Production, and seasonal Evolution in the Ross Sea
PSD	power spectral density

RAO	Response Amplitude Operator
S.A.	South Africa
SAO	Semi-annual Oscillation
SAWS	South African Weather Service
SCALE	Southern oCean seAsonAL Experiment
SIC	Sea-ice concentration
SIE	sea-ice extent
SSMIS	Special Sensor Microwave Imager/Sounder
Sv	Sverdrup
SWIFT	Surface Wave Instrument Float and Tracking
TSE	Trajectory Stretching Exponent
UCT	University of Cape Town
WBA	Wave-Buoy Analogy
WIIOS	Waves In Ice Observation Systems

Chapter 1

1. Introduction

1.1. Background to study

1.1.1. The Antarctic region

The Earth's global climate system is an intricate and interconnected system comprised of the atmosphere, lithosphere, hydrosphere (including the ocean), biosphere, and cryosphere. Interactions between the atmosphere and the ocean are inherently the most vital parts of the climatic system (Arndt, 2016). This is a result of their extensive exchanges of energy and matter, which govern Earth's environmental conditions. The tropics and subtropics are both regions of net radiative input, while on the other hand, there is a net radiative loss over the high latitudes (i.e. the outgoing longwave radiation surpasses the incoming shortwave radiation). The atmosphere's circulation, and to a lesser degree the ocean, restores these imbalances by transporting energy (predominantly in the form of sensible heat, latent heat and potential energy) from the lower latitudes to the poles (Shaw et al., 2016; Simmonds, 2015). As such, the high latitudes play a crucial role in regulating the global climate system (Simmonds, 2015; Kohout et al., 2020), and polar regions are thus required for a comprehensive understanding of global dynamics. Furthermore, these global dynamics are intricately linked to the smaller-scale processes of sea-ice formation as well as its seasonal evolution.

Sea ice and its movement have been shown to strongly influence many of the physical processes that govern climate within the polar seas (Simmonds, 2015; Schlosser et al., 2018). It grows, drifts and melts under the influence of solar, atmospheric, and oceanic forcing (Leppäranta, 2019), and its formation creates a natural radiative barrier between the surface ocean and the cold atmosphere through a high albedo (Allison et al., 1993; Massom et al., 2018). This leads to a profound influence on the exchanges of heat, moisture, gases and momentum, with additional impacts on cloud cover, mean sea level pressure distribution and precipitation (Heil et al., 2009; Parkinson and Cavalieri, 2012; Vaughan et al., 2013; Stroeve et al., 2016). The degree to which sea ice affects these processes depends on the distribution of the ice cover, and its thickness, drift, type and snow coverage (Thorndike et al., 1975; Vichi et al., 2019). These sea-ice characteristics also regulate the freshwater budget in the ocean's surface layer as a result of brine rejection during the formation and freshwater input during ice melt (Massom and Stammerjohn, 2010; Swart et al., 2020). At a first glance, sea ice acts as a protective layer that shields the surface ocean from the atmosphere's effects; it grows and melts in accordance with the heat balance.

However, sea ice is also a complex material in connection with its spatial structure and material composition (Bartels-Rausch et al., 2012). It is a dynamic, porous matrix that holds biota and inorganic particulates within its network of brine pores and channels (Perovich, 2017; Ewert and Deming, 2013). Knowledge of sea-ice dynamics is thus essential for a comprehensive understanding of the ice cover and its climatic interactions (Heil et al., 2001). Since climatic processes take place over several temporal and spatial scales, it is crucial to describe all components as fully and accurately as possible.

The majority of early studies of sea ice and its dynamics have focussed on the Arctic region (e.g. Nansen 1902; Ekman, 1905; Wadhams, 1979; Colony and Thorndike, 1984; Weeks and Ackley, 1986; Thorndike, 1986; Wadhams, 1988). However, since the 1970s with the start of the satellite passive-microwave record (Zwally, 2002; Parkinson and Cavalieri, 2012; Turner et al., 2015), increasing attention has been given to the Antarctic region. These satellite records have shown that the central or principal characteristic of Antarctic sea-ice variability is its annual cycle of the expansion and retreat of the sea-ice extent (Handcock and Raphael, 2020). While both regions exhibit a distinctive annual cycle in response to the seasonal changes in air temperature (Zwally et al., 1983; Maxwell and Barrie, 1989; Overland et al., 2011; Kohout et al., 2014; Eayrs et al., 2019), the patterns of sea-ice formation and melt are notably different between the two hemispheres. In the Arctic, the growth and melt seasons are approximately symmetric, while the Antarctic's seasonal cycle is asymmetric (Eayrs et al., 2020). It undergoes a slow advance during the austral winter months (March to September), and a rapid retreat during the austral summer months (November to February; Eayrs et al., 2019a). This retreat in the Antarctic is almost twice as fast as the autumn advance (Gordon, 1981). The notable contrast between the two hemispheres has indicated that there are vast differences between the Arctic's and Antarctic's sea-ice processes (Eayrs et al., 2019a).

A fundamental difference between these two high-latitude regions is their morphology (Maxwell and Barrie, 1989). Unlike the Arctic, which is a frozen ocean surrounded by continents that confines most of the sea-ice cover, Antarctica is a continent with elevations exceeding 4 km (Simmonds, 2015) that is surrounded by the Southern Ocean. This means that the expansion of its ice extent is not confined by land (Maksym, 2019), but rather it is exposed to some of the largest winds and waves on Earth. The formation of sea ice surrounding Antarctica is one of the largest seasonal changes on Earth (Doble et al., 2003), and dominates the global signal. The ice cover borders Antarctica during the austral summer months at approximately 75-78° S and extends to approximately 55° S during the austral winter months, where it is restricted primarily by the Antarctic Circumpolar Current (ACC; Eayrs et al., 2019a). The ACC is the main current within the Southern Ocean, flowing from west to east around the Antarctic continent (Sect. 1.1.4.3). It is highly dynamic and acts as the northern limit for seasonal ice formation (Eayrs et al., 2019a).

1.1.2. The lifecycle of Antarctic sea ice

During austral winter, when air temperatures begin to decrease, the formation of Antarctic sea ice begins. Sea-ice growth involves many formation stages. However, the growth of Antarctic sea ice is dominated by the formation of pancake ice and interstitial frazil ice at the outer region of the ice cover (Doble and Wadhams, 2006; Petrich and Eicken, 2017; Wadhams et al., 2018). This outer region is known as the marginal ice zone (MIZ). While

having many definitions (see Section 2.1), it is the highly complex and dynamic region characterised by unconsolidated or broken floes, which can extend for hundreds of kilometres. This is due to high winds and the propagation of large ocean waves deep into the ice cover (Squire, 2007; Kohout et al., 2014; Vichi et al., 2019; Kohout et al., 2020; Brouwer et al., 2022). The MIZ exists throughout the year, yet its location and extent is greatly modulated by the asymmetric seasonal cycle (Thomson, 2022). Consequently, a better understanding of the dynamics and thermodynamics, at the different stages of ice formation and melt, is required. Shen et al. (2001) proposed a conceptual model for pancake-ice formation, outlining three distinct stages in the process:

- (1) *The grease-ice stage*: The formation of sea ice starts with the super-cooling of the polar ocean, to approximately -1.8°C (Lemieux et al., 2017), where the first ice crystals known as frazil ice are produced. These then undergo rapid dendritic growth, and in turbulent waters, collisions with each other cause the frazil crystals to adhere to each other (Ashton, 1986). The crystals eventually accumulate to form a thin layer as they float to the surface, where high-frequency waves gather the frazil ice together (Shen et al., 2001). The viscosity of the frazil-ice “slurry” and the inelastic collisions between the ice crystals consume the higher-frequency wave energy. This causes a smoother ocean surface to form, and is therefore termed “grease ice” (Fig. 1.1a; Shen et al., 2001). These two types of sea ice form the sea-ice development stage and are often referred to as new ice, which does not yet make up solid sea ice.
- (2) *The pancake formation stage*: As the ice crystals continue to develop below the ocean’s surface, the accumulation becomes increasingly more buoyant and exposes them to the cold air above (Shen et al., 2001). Under the influence of oceanic waves, this grease-ice layer can become consolidated to form ice floes called pancake ice, with diameters of ~ 2 m in the Antarctic (Alberello et al., 2019). These are small, roughly circular and mobile ice floes that form during wavy conditions at the outer regions of the ice cover (Fig. 1.1b; Alberello et al., 2020). The size of the pancake ice floes is dependent on the high-frequency wave energy travelling into the ice cover (Shen et al., 2001). However, rafting and collisions are thought to eventually consume the high-frequency wave energy, and the process becomes driven progressively by the longer wavelengths forming the larger pancake floes (Fig. 1.1c; Shen et al., 2001). These may then grow to > 4 m in diameter (Alberello et al., 2019).
- (3) *The composite pancake stage*: The longer wavelengths gather and raft the primary pancakes together, which results in the thickening of the floes vertically. Lateral growth, due to the freezing of adjacent floes, takes place concurrently. This forms composite floes (Shen et al., 2001). As the floes increase in diameter, the ice cover functions as a low-pass filter with a decreasing low-frequency cut off. These floes will continue to amalgamate until the wave field is attenuated completely (Shen and Ackley, 1991; Shen et al., 2001), eventually consolidating into a coherent ice sheet (Fig. 1.1d; Weeks and Ackley, 1986; Doble and Wadhams, 2006).

While, the ice cover may grow to an area larger than the size of the Antarctic continent during its winter maximum (Parkinson and Cavalieri, 2012), it forms as a thin cover over the ocean’s surface, and is on average < 1 m thick (Eayrs et al., 2019a). Most of the ice cover surrounding the continent is young ice (10-30 cm thick) and first-year ice (30 cm to 2 m thick), which is made up of consolidated pancake ice and does not survive the summer-time ice melt. For this reason, multi-year sea ice is quite rare, being only found along a few coastal areas around Antarctica, primarily in the Weddell and Ross Seas (Melsheimer, 2023).

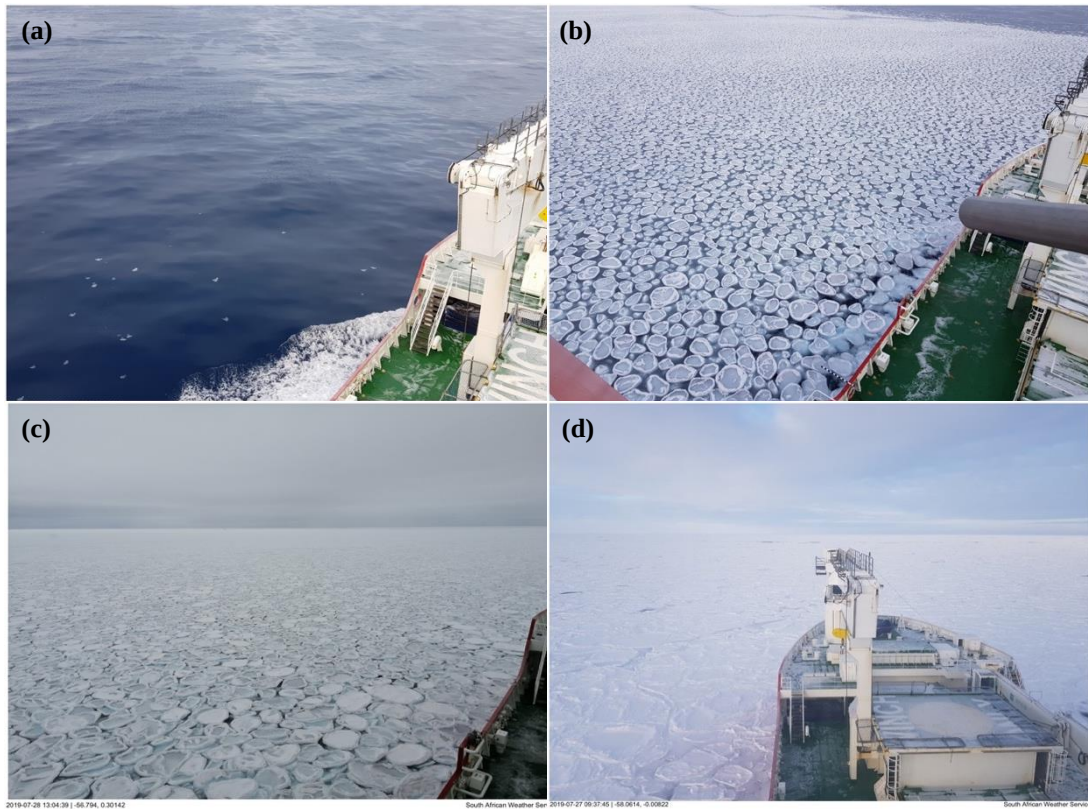


Figure 1.1. Photographs showcasing (a) grease ice, (b) small pancake ice floes and interstitial frazil ice, (c) larger and rafted pancake floes, and (d) consolidated first-year ice. These photographs were taken by the South African Weather Service during the Southern Ocean seasonal Experiment (SCALE) 2019 Winter Cruise aboard the South-African (SA) Agulhas II (de Vos et al., 2019).

Rafting and ridging are two principal processes in the deformation of sea ice, which occur when ice floes converge or collide with one another (Hopkins et al., 1999). These processes are a result of the differential drift of the individual floes (Sect. 1.1.5.2; Hibler, 1974; Geiger et al., 1998; Girard et al., 2009). Rafted ice is the simple (partially) overriding of one ice floe over the other, and is one of the mechanisms for the dynamic thickening of the floes. Ridges, on the other hand, are a result of the flexural failure of ice sheets and the consequent piling of ice formed by the failure at the top and bottom of the two adjacent ice sheets (Hopkins et al., 1999). Deformation of sea ice, through rafting and ridging, establishes the large-scale strength of the ice cover and is responsible for the thickest ice observed (Martin, 2007; Petrich and Eicken, 2017). It produces a mechanism for thickening (Hibler, 1979), even when thermodynamic growth is hindered or halted (Petrich and Eicken, 2017). Thus, within a warming climate, these two processes are particularly important, as they may compensate for some of the thinning of the ice cover as a result of the reduction of ice growth in the winter months and the increase in ice melt during the summer months (Petrich and Eicken, 2017).

Fracturing of sea ice also plays an essential part in the growth, break-up and ultimate retreat of sea ice during its seasonal cycle (Weiss and Dansereau, 2017). When the ice cover is consolidated, compression and/or variations in air temperature can cause the flexure and cracking of the sea ice. Due to the mechanical stress induced by wave activity, ocean currents and winds (Reiser et al., 2019), these cracks may eventually result in large sea ice leads that can extend for several hundreds of kilometres (Bartels-Rausch et al., 2012). Waves are however considered

to be the primary reason for the mechanical break-up of sea ice, and also for the magnitude of these break-ups (Zhang et al., 2015). These break-ups are caused when the bending-induced stresses are larger than the flexural strength of sea ice and/or if the repeated bending causes fatigue failure (Zhang et al., 2015). Subsequently, the floe-size distribution is governed by the propagation of ocean waves in ice (Wadhams et al., 1988; Dumont, et al., 2011; Zhang et al., 2015; Alberello et al., 2019), where floe sizes increase with distance from the ice edge as wave energy concurrently is dissipated and scattered (Shen and Ackley, 1991).

The final stage of the evolution of sea ice is its aging and melting; however, all these stages can occur concurrently (Bartels-Rausch et al., 2012). Contrasting to the Arctic, the more dynamic ice cover of the Antarctic rarely has melt ponds during the summer-melt season (Andreas and Ackley, 1982; Maksym, 2019). Instead, it is more vulnerable to lateral and bottom ice melt when air temperatures begin to rise (Itkin et al., 2017). Sea-ice melt in the Antarctic therefore starts when surface air temperatures reach above 0°C (Andreas and Ackley, 1982). The existing ridges and leads become the weak points at which the ice cover begins to disintegrate (Itkin et al., 2017). Passing cyclones and wave activity can also influence the ice cover during the summertime, however majority of it is preconditioned by the processes that occurred during the winter months (Screen et al., 2011).

1.1.3. Southern Ocean cyclones

The Southern Ocean is home to some of the most extreme and intense planetary storms (Vichi et al., 2019). These storms are predominantly mid- and high-latitude synoptic cyclones. They develop in the Atlantic and Pacific sectors of the Southern Ocean (Hoskins and Hodges, 2005; Uotila et al., 2011; Grieger et al., 2018), and generally follow a south-eastwards path, where they undergo net cyclogenesis in the mid-latitudes and eventually net cyclolysis closer to the Antarctic continent (Hoskins and Hodges, 2005).

These synoptic cyclones play an important role in the weather and its extremes over the Southern Ocean (Murphy and Simmonds, 1993), and contribute to the general atmospheric circulation in the southern polar latitudes by maintaining the Circumpolar Pressure Trough (CPT; Maksym, 2019; Eayrs et al., 2020) – an almost annular low-pressure region (between 60° S and 70° S; Fig. 1.2), with its exact position varying with both longitude and the time of year (Irving et al., 2010). Considering the zonal average, the CPT is typically found furthest south during spring and autumn, and migrates slightly northwards during the summer and winter months (Watkins and Simmonds, 1998; Simmonds and Jones, 1998; Simmonds et al., 2017). This twice-yearly contraction and expansion of the CPT around Antarctica is referred to as the semi-annual oscillation (SAO), and arises from the differences in heat storage between Antarctica and the surrounding oceans (van den Broeke, 1998). Papritz et al. (2014) reported that between approximately 60 % and 90 % of all strong precipitation events in this low-pressure region are caused by synoptic cyclones. Additionally, modelling results reveal that these cyclonic events modify the ocean, sea ice and surface heat fluxes (Uotila et al., 2011), as the low-pressure systems transport heat and moisture poleward from the mid-latitudes to Antarctica (Maksym, 2019), with the latter being particularly important for the mass balance of the Antarctic ice sheet (Grieger et al., 2018).

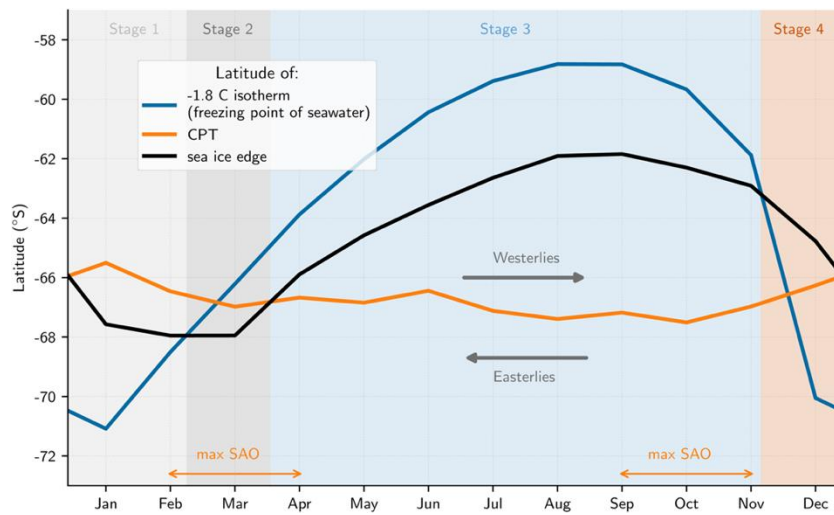


Figure 1.2. Stages of the processes driving the asymmetric annual cycle of sea-ice extent in the Southern Ocean. The blue line denotes the monthly mean latitude of the freezing 2-m air temperature (-1.8°C). The black line denotes the monthly mean latitude of the sea-ice edge. The orange line denotes the monthly mean latitude of the CPT. The maximum SAO signifies the period when the SAO causes the CPT to be the most intense and located closer to Antarctica. This is Figure 1 taken from Eayrs et al. (2020).

Numerous studies have indicated that the Southern Ocean storminess plays a critical role in modulating the Antarctic sea-ice extent (e.g. Eayrs et al., 2019a; Maksym, 2019; Matear et al., 2015; Pezza et al., 2012; Schemm, 2018; Schlosser et al., 2018; Turner et al., 2017), as synoptic events and cyclones contribute to its variability and trends (Holland and Kwok, 2012; Maksym, 2019; Schemm, 2018, Turner et al., 2017; Alberello et al., 2020), particularly in the Atlantic and Indian Ocean regions of the Southern Ocean (Matear et al., 2015). In the northern hemisphere, stationary waves are responsible for a large portion of the transfer of energy to the poles. However, in the southern hemisphere these waves contribute far less to the poleward transfer of energy (Grieger et al., 2018), and therefore it is left to the low-pressure systems to transport the required energy (Peixoto and Oort, 1992). These extra-tropical cyclones act to modulate sea ice, both dynamically and thermodynamically, on short timeframes, with an almost instantaneous response of the ice (Schlosser et al., 2016). Ice-landing cyclones are able to cause rapid variations of sea ice as they (i) warm the overlying air temperatures (Messori et al., 2017), (ii) generate waves close to the ice edge (Kohout et al., 2014), (iii) push the ice edge southwards on the eastern flank (Vichi et al., 2019), while also (iv) accessing a pool of cold air from the Antarctic continent on the western flank. Vichi et al. (2019) demonstrated the result of these processes, which are clearly visible in the daily changes of remotely sensed sea-ice concentration at the ice edge. These relative changes are an indicator of sea-ice drift and deformation induced by intense winds, waves and the advection of warm oceanic air from the mid-latitudes. This ultimately rearranges the MIZ in a clockwise direction (Vichi et al. 2019).

It has been predicted that there will be a continuation in the poleward shift of storm tracks (Turner et al., 2009; Massom and Stammerjohn, 2010; Kohout et al., 2014) as well as the intensification of cyclonic activity in our warming climate (Wang et al., 2014). This may result in increased winds and wave activity over the Antarctic region, especially in the summer and autumn months (Massom and Stammerjohn, 2010). This will greatly affect sea-ice dynamics, particularly in the MIZ due to its dynamical nature and high frequency of storms that across it. However, even with the continuous study of extra-tropical cyclones (Schultz et al., 2019), our current understanding of how they affect sea-ice dynamics and thermodynamics remains incomplete (Vichi et al., 2019).

1.1.4. Atmospheric and oceanic forcing

The dynamics and spatial patterns of sea ice are governed by both the atmosphere and the ocean (Thorndike and Colony, 1982; Wadhams, 2000; Alberello et al., 2020), and are closely associated with extra-tropical and polar cyclones (Heil et al., 2009). The dynamic processes include the drift and deformation of sea ice by winds, waves and ocean currents (see Sect. 1.1.5; Thorndike and Colony, 1982). However, wind forcing is considered to be the main driver of sea-ice drift (Vihma et al., 1996; Uotila, 2001; Doble and Wadhams, 2006; Uotila et al., 2011), and is counterbalanced by the internal stresses of ice floes and the ice-ocean drag (Eq. 1.1; Uotila, 2001; Leppäranta, 2011). Coriolis force also comes into effect when the sea ice is drifting. Ultimately, when investigating sea ice and its characteristics it is therefore essential to first understand its development in context with other components of the polar environment.

1.1.4.1. Wind forcing

Strong, continuous and directionally consistent near-surface winds are one of the most marked features of the Antarctic climate system (Turner et al., 2009). Over the Southern Ocean two primary wind regimes exist. These are the mid-latitude westerly winds, characterised by high wind speeds towards the east, and the polar easterly winds blowing towards the west (Fig. 1.2), along the Antarctic coastline. In the westerly wind belt, the pressure distribution at mean sea level pressure is primarily zonal, and is consistent with the path of cyclones from the mid-latitudes to the Antarctic coastline (Jones and Simmonds, 1993).

During the winter and spring months, the transition zone (the CPT) between the westerlies and the polar easterlies is found over the sea-ice cover. The frequent passage of cyclones over the sea ice imposes continuous changes in local wind conditions. The MIZ, which is dominated by the westerly winds, is affected by a greater number of passing cyclones than the inner ice pack. However, low-pressure systems that do reach the inner ice pack are generally more intense and energetic than those over the outer ice edge region (Heil, 1999). When cyclonic systems pass over the ice-covered ocean, surface winds, particularly the meridional winds play an important role in reshaping the ice edge (Kwok et al., 2017; Vichi et al., 2019). They disturb the ice cover by changing the ice velocity and inducing deformation of ice floes, through the transfer of momentum to the drifting ice and further to the underlying ocean (Leppäranta, 2011). Divergence is associated with winds blowing northwards (southerly winds) from the continent, resulting in offshore transport of sea ice and an increase in the sea-ice extent (Eayrs et al., 2020). This also produces the formation of leads and coastal polynyas near the Antarctic coast. Conversely, southwards blowing (northerly) winds lead to the convergence of sea ice, which increases the ice concentration and thickness through rafting and ridging processes in the inner ice pack (Schlosser et al. 2018). However, strong westerly winds may also have some impacts on the meridional movement of ice through Ekman transport (Watkins and Simmonds, 1998; Schlosser et al., 2018; Eayrs et al., 2020).

Sea-ice drift can vary significantly, and ice floes can accelerate rapidly due to sporadic changes in wind speed and direction (Ackley et al., 2015). Away from the Antarctic coastline, the sea ice is typically close to a state of free-drift (see Sect. 1.1.5.3), moving at a small angle to the winds, which follows contours of atmospheric pressure (Holland and Kwok, 2012). This creates a tendency of sea ice in the outer region to drift eastwards, while the ice along the coastline drifts westwards (Kimura, 2004). The freely drifting ice floes have a strong correlation to the

overlying winds, and can be largely explained by a linear relationship (Farooq et al., 2023). This quantifies the effects of winds on ice motion through both the direct wind forcing and any portion of ocean stress and sea-surface tilt attributed to the local winds (Leppäranta, 2011; Holland and Kwok, 2012).

Surface winds are greatly dependent on cyclonic activity within the CPT (Schlosser et al., 2016). On synoptic time scales, winds that are associated to extra-tropical cyclones can be extremely erratic and intense, especially in the coastal regions of the East Antarctic where cyclones frequently enhance and suppress strong katabatic winds (Turner et al., 2009; Wang et al., 2014). Katabatic winds are by definition gravity-driven air flow (Heil, 1999) and are generated by the downward slope of the Antarctic orography (Turner et al., 2009; Grieger et al., 2018). These winds are increasingly deflected to the left (in the southern hemisphere) of its propagation path by the Coriolis force, which results in a north-westward flow along the coast. As the polar cyclones pass over the ice cover, their northerly flow on the eastern flank suppresses the katabatic winds. After the cyclones have travelled eastwards, their strong southerly flow on the western flank then enhances the katabatic flow. These strong winds can enforce large atmospheric and consequently induce oceanic forcing on sea-ice, with a significant potential to cause rapid changes of extreme sea-ice conditions (Wang et al., 2014) and drive ice drift and deformation.

1.1.4.2. Wave forcing

Ocean surface gravity waves are mechanical water waves that propagate along the sea-surface interface, with wave heights ranging from only a few centimetres to over 10 metres (Babanin et al., 2023). When moderate-to-strong winds blow over the ocean's surface, the majority of energy and momentum are transported to the surface waves before they are passed to the ocean through wave breaking and wave-turbulence interactions (Babanin et al., 2011). With the largest and almost uninterrupted ocean area, waves in the Southern Ocean are the biggest on Earth (Babanin et al., 2019). Intense and frequent storm activity generate these large ocean waves in the Southern Ocean (Massom and Stammerjohn, 2010; Kohout et al., 2014), which have enough energy to propagate hundreds of kilometres into the sea-ice cover (Vichi et al., 2019; Kohout et al., 2014, 2020; Brouwer et al., 2022). Despite the continued study of the interactions between ocean waves and sea ice, our understanding of the impacts that waves impose on ice break-up and the maintenance of the Antarctic MIZ remains incomplete (Montiel and Squire, 2017; Squire, 2020; Thomson, 2022; Horvat, 2022; Shen, 2022).

Ocean waves play a pivotal role in the formation of Antarctic sea ice (Shen et al., 2018). However, they can also leave behind a wake of broken ice floes, which are then more vulnerable and easily deformed by winds and ocean currents (Squire et al., 1995; Langhorne et al., 1998; Kohout et al., 2020). This effectively reduces the natural barrier between the atmosphere and ocean, and enhances heat exchange (Kohout et al., 2014). Vichi et al. (2019) suggested that these propagating waves delay the consolidation and maintain a pancake and frazil ice field, which in turn can accelerate the melt season (Wadhams et al., 1979; Massom and Stammerjohn, 2010; Toffoli et al., 2015). Concurrently, when ocean waves travel into the ice cover – whether composed of discrete ice floes that characterise the MIZ or a consolidated ice sheet – they decrease in amplitude at a rate governed by the nature of the sea ice (Herman, 2017; Bennetts et al., 2015; Squire and Montiel, 2016; Squire, 2020). This reduction or attenuation occurs due to scattering and dissipative processes (Herman, 2017). Scattering redistributes the kinetic energy, but does not remove it, while dissipation eliminates the energy (Lemieux et al., 2017; Dolatshah et al., 2018). Dissipative processes include turbulence; inelastic floe collisions; wave breaking; energy loss, due to ice

deformation under extreme conditions; and ridging, rafting and floe fracturing (Toffoli et al., 2015; Montiel et al., 2018; Squire, 2020). Recent studies have begun coupling sea ice and wave models (e.g. Montiel and Squire, 2017; Boutin et al., 2018; Roach et al., 2018). However, present operational wave models are not yet capable of describing propagating waves through all types of sea ice (Wang and Shen, 2010).

Near the ice edge, high-energy waves create complex sea states resulting in differential ice motion and variable collision patterns (Squire, 2007; Shen et al., 2018; Herman, 2018), which can cause the destruction of ice floes (Squire, 2020). However, because dissipation removes the higher-frequency waves before the lower frequency ones (Meylan et al., 2014; Rogers et al., 2021), this intense zone of energy loss is generally limited to 10-20 km for the Arctic (Squire, 2020). In the Southern Ocean, where the seas are more violent and the sea ice is thinner and more heterogeneous, this zone can be far wider (Brouwer et al., 2022). Attenuation and dissipation rates in these outer ice edge regions are expected to deviate greatly from those in the interior, since they depend on several factors such as ice type, thickness and concentration, as well as wave properties, including wave energy (Ardhuin et al., 2018). To describe the rate of total wave energy attenuation accurately, wave direction, ice conditions, ice extent, wave duration and wave speed must be considered (Kohout et al., 2020). However, due to logistical challenges faced in the harsh Antarctic environment, and the low survival rate of in-situ instruments, these quantities cannot always be comprehensively measured. This regularly leads to the application of assumptions and approximations from satellite imagery and laboratory experiments.

Earlier in-situ measurements of waves in the MIZ have shown that the ice cover attenuates wave energy exponentially over distance (e.g. Robin, 1963). The attenuation rate has become a research focus (Thomson, 2022), as it informs predictions of the width of the ice-covered region that is effected by waves, and hence the MIZ extent (Kohout et al., 2014; Bennetts et al., 2017). Significant advances in measuring the wave attenuation in the Antarctic MIZ have been made, including dedicated scientific expeditions that use arrays of wave buoys deployed on ice floes (e.g. Kohout et al., 2014; 2020). The recent increase in measurements (including satellite remote sensing) has generated a new understanding of wave attenuation in the MIZ (Brouwer et al., 2022), however the dominant sources of attenuation are still a topic of debate. In-situ measurements showcase a large range of attenuation rates (Montiel et al., 2022), even with comparable frequencies. This has been attributed to the dependence on the ice cover properties, as well as the momentum transfer from winds (Alberello et al., 2022). Since larger waves are now occurring more regularly (Dobrynin et al., 2012), as a result of the increased frequency and intensity of polar cyclones, the propagation of ocean waves into the ice cover is potentially growing. It is also known that waves can grow throughout the ice cover itself with the presence of more open water and thinner ice, which become increased fetches over which waves can form and evolve (Herman, 2017; Squire, 2020; Boutin et al., 2021). Therefore, a better understanding of the large-scale MIZ dynamics is an important step to improve our understanding of the inevitable response of Antarctic sea ice to climate change and its vast impacts (Brouwer et al., 2022).

1.1.4.3. Ocean current forcing

The geographic configuration of the Southern Ocean is unique as it possesses a circumpolar channel that is open at all longitudes. This, combined with the strong westerly winds and surface buoyancy leads to the existence of the ACC. The ACC is the largest current system on Earth and continuously transports approximately 130 Sv (1

$Sv = 10^6 \text{ m}^3 \text{ s}^{-1}$) of water eastwards around Antarctica (Wadhams, 2000; Meredith and Brandon, 2017), through the southern parts of the Atlantic, Indian and Pacific Oceans. It consists of relatively fast-moving jets separated by more quiescent zones of water. These jets coincide with oceanic fronts, namely from north to south, the Sub-Antarctic Front, the Polar Front and the Southern ACC Front (Orsi et al., 1995). The southern limit of the ACC is marked by the Southern Boundary. Further south, along the Antarctic coastline, the Antarctic Coastal Current (ACoC) flows westward, transporting approximately 8 Sv of water, as a result of the combined polar easterly winds and the katabatic winds (Sect. 1.1.4.1; Wadhams, 2000). In between these two currents lies a series of sub-polar gyres, most notably the Weddell Sea and the Ross Sea gyres. These gyres become fully covered by sea ice during the winter months and extend up to the southern region of the ACC, where they are strongly steered by the flow of the Southern Boundary (Meredith and Brandon, 2017).

Since sea ice is embedded in the upper ocean, it also interacts with these wind-driven ocean currents through the turbulent boundary layer – the ice-ocean boundary layer (IOBL) – that is influenced directly by rotation through Ekman Spiral (McPhee 2008; MCPhee, 2017). The IOBL arises from the shear (relative motion) between the wind-driven ice drift and the upper ocean (Cole et al., 2014; MCPhee, 2017). It shares many of the same characteristics as the open-ocean boundary layers, apart from the fact that the upper boundary conditions are enforced by a solid-liquid interface rather than a liquid-gas interface. The sub-inertial ocean currents immediately under the ice are described by the combination of a logarithmic boundary layer and an Ekman layer (McPhee, 2008; Cole et al., 2017). The logarithmic boundary layer is typically only a few metres thick. On the other hand, the Ekman layer has currents that decay and rotate with depth, due to Ekman Spiral (McPhee, 2008). This concept was prompted by observations made by F. Nansen during the Fram’s expedition along the Transpolar drift (1883 $\pm$ 1886; MCPhee, 2008; Leppäranta, 2011). Nansen (1902) determined that sea ice was drifting at a 20-40° angle to the right (northern hemisphere) of the predominant wind forcing. This led to the development of the mathematical description of the rotating column of water that forms when water moves at an angle to the left (right) of the wind direction in the southern (northern) hemisphere due to the Coriolis effect (Ekman, 1905).

The complexity of the ocean’s flow due to factors such as Earth’s rotation and the thermohaline circulation (forced by the uneven surface heating, evaporation and precipitation) result in a piling up of water towards the centre of gyres (Wadhams, 2000). This sea-surface tilt leads to the presence of geostrophic currents, which are the steady currents that are formed by the balance between horizontal pressure gradient and the Coriolis force (McPhee, 2008). In ice-covered oceans, Wang et al. (2021) indicated that the impact of surface geostrophic currents on sea ice are predominantly through the varying sea-ice drift and deformation. They also revealed that changes in the freshwater content, and hence, the halosteric height may additionally cause changes in the surface elevation and thus the geostrophic currents, which have strong impacts on sea ice (Wang et al., 2021).

The combined horizontal oceanic forcing on sea-ice motion generally takes place over large scales and is much slower than the direct wind-driven influence. Typical time scales for ocean-induced transport varies from weeks to years (Heil, 1999). Thus, oceanic processes generally contribute to the long-term polar climate state and its variability. However, daily and sub-daily forcing, due to the diurnal or semi-diurnal cycles (McPhee, 1978; Gieger et al., 1998; Heil et al., 2009; Alberello et al., 2020), are also known to affect the surface ocean layers and hence sea-ice dynamics. Much remains unknown about the influence of the IOBL on longer-term scales as in-situ current measurements below the sea-ice cover are sparse, with none currently available in the Antarctic MIZ (Alberello

et al., 2020). Therefore, several recent studies, such as Alberello et al. (2020) and Biddle and Swart (2020), model free-drifting sea ice where the ocean current velocity is assumed to be zero. This leads to significant uncertainties on the complete role that ocean currents play on Antarctic sea-ice dynamics.

1.1.5. Sea-ice dynamics

As reported above, sea ice is affected by the fact that it is not a passive stationary sheet of material covering the ocean's surface, but rather it is driven by winds and ocean currents (Sect. 1.1.4; Wadhams, 2000; Heorton et al., 2019). Fracturing, leads, rafting, pressure ridges and polynyas are all products of the dynamic response of sea ice to these forcing mechanisms. Mathematical theory can describe and predict the motion of sea ice within the IOBL, and together with thermodynamic theory, this allows for the development of ice-ocean models that can define both the velocity of sea ice and its thickness changes. These ice-ocean models deal with the manner in which momentum is transferred through the sea-ice system from external forces (Hibler, 1986). The key features of this transfer may be characterised by equations that represent the following parts (Hibler, 1986; Wadhams, 2000):

- (1) *the momentum balance* – the nature and magnitude of the forces acting on every component of sea ice;
- (2) *ice rheology (a material law)* – the manner in which the ice cover behaves as a material when acted upon by varying stresses;
- (3) *ice strength (a material property)* – what nature and magnitude of stress is required to produce fracturing of the ice cover; and
- (4) *the mass balance* – the reliance of the ice thickness characteristics on the growth, retreat and ice drift (Coon et al., 1974).

Subsequently, these smaller models can be embedded in General Circulation Models (GCMs) of the atmosphere and ocean and predict climate variability (Wadhams, 2000), in which sea ice plays a key role (Simmonds, 2015; Schlosser et al., 2018).

1.1.5.1. Momentum balance equation

The momentum balance describes the forces that determine sea-ice drift and deformation within the IOBL. The momentum balance, for a single ice floe, in a two-dimensional Cartesian coordinate system can be modelled using the Arctic Ice Dynamics Joint Experiment (AIDJEX) model for sea-ice drift with the following equation in the Eulerian form (Coon et al., 1974; Hibler, 1979; Wadhams, 2000; Feltham, 2008; Leppäranta, 2011):

$$m_i \left(\frac{\partial \mathbf{u}_i}{\partial t} + \mathbf{u}_i \cdot \nabla \mathbf{u}_i \right) = -m_i f_c \mathbf{k} \times \mathbf{u}_i + \boldsymbol{\tau}_a + \boldsymbol{\tau}_w - m_i g \Delta H + \nabla \cdot \boldsymbol{\sigma}, \quad (1.1)$$

where each term has units of force per unit area of the sea-ice cover, and t is time. The rate of change of momentum (left-hand-side) is equal to the balance of forces (right-hand-side), which are as follows (from left to right): Coriolis force, wind stress, water stress, sea-surface tilt, and the ice interaction force given by the divergence of the ice stress tensor (Coon et al., 1974; Hibler, 1986; Wadhams, 2000; Feltham, 2008; Leppäranta, 2011). The

Cauchy stress tensor, σ_{ij} , describes the sea-ice stress state (Feltham, 2008), which is dependent on the relevant ice type. In the momentum equation, m_i and $\mathbf{u}_i$ are the mass per unit area and the horizontal velocity vector of ice; $\mathbf{k}$ is the unit vector normal to the ice surface pointed towards the atmosphere; f_c is the Coriolis parameter (Eq. 1.8); g is the gravitational acceleration; and H is the dynamic sea-surface elevation. The mass of the ice is determined by:

$$m_i = \rho_i \int_0^\infty G(h)h \, dh, \quad (1.2)$$

where ρ_i is the density of sea ice, and $G(h)$ is the thickness distribution function expressing the fraction of a given area covered by ice with thickness range of h to $h + dh$ (Coon et al., 1974; Thorndike et al., 1975; Feltham, 2008). This function approximates the thermal and mechanical effects on thickness, and characterises the state of the ice over time (Coon et al., 1974).

1.1.5.2. Motion of sea ice

The resulting motion of sea ice occurs in a three-dimensional world, but it is treated rather as a two-dimensional phenomenon on the ocean's surface (Fig. 1.3). This is allowed for two reasons: (i) sea-ice floats on the surface of the ocean, and does not possess a vertical velocity component, and (ii) ice floes move as a rigid body and the horizontal velocity is independent of the vertical coordinate (Leppäranta, 2011). Strictly, with respect to the Earth, the vertical velocity is not zero, however the slowly changing sea level does not lead to any significant inertial effects on ice motion (Leppäranta, 2011). The velocity of sea ice is therefore derived by a two-dimensional Cartesian system:

$$\mathbf{u}_i = \mathbf{u}_i(x, y, t), \quad (1.3)$$

with u and v , respectively, as the x and y components. The dynamics of a single ice floe can be examined using rigid body mechanics (Leppäranta, 2011). This motion has three degrees of freedom comprised of the translational motion $\mathbf{u}_i$ and rotation rate $\dot{\omega}$ about the vertical axis. In granular ice motion, each floe contains individual translational and rotational velocities. Providing that there are enough open areas around the ice floes, they will drift individually. As these open areas reduce, the floes combine into larger and larger groups, and they begin to interact through collisions or overriding (Leppäranta, 2011).

When the size of the drifting ice particles become an order of magnitude greater than the floe size, the continuum approximations can be used for the drift kinematics (Leppäranta, 2011). The continuum method averages the motion of individual ice floes to the continuum length scale. However, the resulting velocity field needs to have continuous spatial derivatives to at least the second order to apply this approach to ice drift (Leppäranta, 2011).

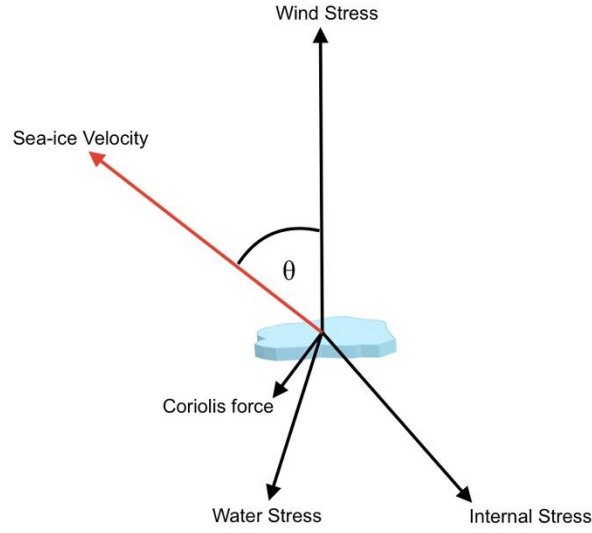


Figure 1.3. Typical force balance for sea-ice drift in a two-dimensional system. This figure has been adapted from Weiss (2013) and Hunkins (1975), but for the southern hemisphere, where Coriolis force acts to the left of sea-ice velocity. Theta (θ) denotes the turning angle between ice drift and wind stress.

The motion of a drift continuum can be split up into rigid translations, rigid rotation and strain. For horizontal movement, translation has two horizontal components, and rotation has one component around the vertical axis (Leppäranta, 2011). Strain denotes the physical deformation of the continuum particles or ice floes. The three invariants of the deformation rate – divergence, shear, and vorticity – can be derived assuming that sea ice is represented as a homogenous continuum with the following equations (Lindsay, 2002; Lei et al., 2020):

$$\dot{\epsilon}_I = \text{div} = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}, \quad (1.4)$$

$$\dot{\epsilon}_{II} = \text{shr} = \sqrt{\left(\frac{\partial u}{\partial x} - \frac{\partial v}{\partial y}\right)^2 + \left(\frac{\partial u}{\partial y} + \frac{\partial v}{\partial x}\right)^2}, \quad (1.5)$$

$$\dot{\epsilon}_{III} = \text{vort} = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}, \quad (1.6)$$

and the total deformation rate is:

$$|\dot{\epsilon}| = D = \sqrt{\text{div}^2 + \text{shr}^2}, \quad (1.7)$$

where $\frac{\partial}{\partial x}$ and $\frac{\partial}{\partial y}$ are the partial derivatives in the x and y directions, respectively.

The relationship between the internal stresses of sea ice, σ (Eq. 1.1), and the large-scale deformation of the ice cover, $|\dot{\epsilon}|$ (Eq. 1.7), is known as sea-ice rheology (Flato, 2003; Feltham, 2008). Rheology is a topic within physics that can be described as the study of deformation and flow of a complex non-Newtonian liquid or gas. A rheology,

also identified as a material law or constitutive equation, defines the relationship between two physical quantities with material-dependent parameters. In sea-ice dynamical models, a rheology describes the relationship between stress and strain-rate, which represents the internal ice forces and corresponding deformation of sea ice (Feltham, 2008; Bouchat et al., 2022). During pack-ice conditions (thicker and stronger ice), internal ice stresses are important, but for low ice concentrations ($< 80\%$; McPhee, 1980; Leppäranta and Omstedt, 1990) and during free-drift conditions, they can be considered negligible ($\nabla \cdot \boldsymbol{\sigma} = 0$; Alberello et al., 2020).

1.1.5.3. Free-drift motion

A free-drift solution to the momentum balance (Eq. 1.1) assumes that our unit reference area of the ice cover moves as if was an isolated system on the ocean surface (i.e. a single floe), where there are no collisions with adjacent floes and no effects of internal ice stresses (Wadhams, 2000; Leppäranta, 2011). Sea-ice drift is then a result of the balance between the Coriolis force and stresses from wind and oceanic forcing (Heorton et al., 2019).

The term *free-drift* was coined by McPhee (1980). However, the Coriolis effect has been analysed for more than a century; since the first introduction by F. Nansen and V.W. Ekman (Sect. 1.1.4.3). This research by Ekman (1905) additionally foretold the presence of inertial oscillations, which are the circular motions of a particle or object within a rotating framework.

Under the action of wind stress, $\boldsymbol{\tau}_a$, an ice floe (simultaneous with ocean currents) will drift in the direction of $\boldsymbol{\tau}_a$ but quickly obtains its own velocity (Wadhams, 2000). Coriolis force will then begin to act perpendicularly to the floe's velocity and divert its motion to the left in the southern hemisphere (Fig. 1.3; Gimbert et al., 2012a). As long as the ice floe holds any component of velocity in the direction of the wind, it will continue to accelerate (Eq. 1.1). Subsequently, the turning effect of the Coriolis force will grow in strength until the velocity of the ice floe starts to oppose the direction of $\boldsymbol{\tau}_a$, and the deviations subside. This process forms inertial oscillations or loops within the drift paths of ice floes, and are characterised by a frequency of:

$$f_c = 2\Omega \sin\phi, \quad (1.8)$$

where ϕ is the latitude and $\Omega = 7.292 \times 10^{-5} \text{ s}^{-1}$ is the angular velocity of Earth. The inertial period is then given by:

$$T = \frac{2\pi}{f_c}, \quad (1.9)$$

and is equal to 12 hours at the poles and infinity at the Equator, where the Coriolis force is zero. These inertial oscillations occur on spatial scales smaller than the Rossby radius of deformation, thus leading them to be a local phenomenon (Heil, 1999).

1.2. Problem identification

While we understand that the advance and retreat of Antarctic sea ice are driven by a range of processes (Eayrs et al., 2019a), the paucity of observations has meant that these drivers are poorly understood – specifically, the data on sea-ice drift, deformation and concentration in the MIZ. This is because the harsh atmospheric and oceanic environment in the Antarctic, especially during the winter months, restricts our accessibility to the ice cover unlike the Arctic. Additionally, the number of in-situ instruments has been limited due to high deployment and maintenance costs. The majority of the available metocean (meteorological and oceanographic) measurements therefore have been collected during the summer months, when sea-ice conditions are calmer. These mild sea states reduce the intensity of many key mechanics dominating the interactions between winds, waves and sea ice. As a result, our knowledge of metocean properties is limited, hindering our ability to accurately model these interactions (Derkani et al., 2021), and understand their impact on sea-ice dynamics. The inability of contemporary models to reproduce the advance and/or retreat of sea ice in the Southern Ocean (Kwok et al., 2017; Hobbs et al., 2016) introduces strong biases in ocean-atmosphere heat fluxes and salt input into the ocean (Doble, 2009), underscoring the critical need for improved observational data and modelling techniques.

Most of our scientific knowledge on the sea-ice drift and deformation surrounding Antarctica additionally has been derived from pack-ice conditions (e.g. Vihma et al., 1996; Uotila et al., 2000; Lindsay, 2002; Doble and Wadhams, 2006; Ackley et al., 2015), and less attention has been given to the dynamics of the MIZ (e.g. Doble et al., 2006; Doble et al., 2013; Alberello et al., 2020), even though their spatial morphology is not equivalent (Vichi, 2022). Earlier studies have attempted to quantify the growth process and elevated-growth rate of pancake ice relative to that of the consolidated pack ice (Doble et al., 2003), as well as the dynamics of pancake ice against the contrasted pack ice (Doble and Wadhams, 2006). However, the lack of in-situ data on sea-ice dynamics and concentration in the Antarctic MIZ has often made satellite remote sensing the only source of information (Vichi et al., 2019). These datasets are unfortunately better suited for understanding large-scale and long-term ice drift patterns (Kwok et al., 2017), rather than the characteristics of daily and sub-daily ice motion, which are more subjected to the passage of cyclones. Therefore, the understanding of the physical processes between the atmosphere, ocean and sea ice in the MIZ still requires a significant amount of attention.

The recent and pronounced reversal of changes occurring in Antarctic sea ice – the gradual increase in the sea-ice extent of approximately 1.5 % per decade, with record highs seen between 2012-2014, and followed by decreases in 2016, 2017 and 2018 (Meehl et al., 2019; Maksym, 2019), as well as in 2022 and 2023 (Purich and Doddridge, 2023) – took many experts by surprise (Stopa et al., 2018). This immense and unexpected loss in sea-ice cover matched the 30-year observed decrease in Arctic sea ice (Eayrs et al., 2021). At a regional scale, the trends exhibit slight differences or even opposing signals (Parkinson and Cavalieri, 2012; Turner et al., 2016). Possible driving mechanisms causing this juxtaposition of trends include changes in surface temperature, wind direction (Holland and Kwok, 2012; Turner et al., 2016), ice-ocean feedbacks (Goose and Zunz, 2014) and ice drift and deformation (Pope et al., 2016). However, this recent shift reflects the lack of understanding of the processes governing Antarctic sea ice (Turner et al., 2013). This is ultimately due to our limited understanding of sea-ice dynamics (Stopa et al., 2018). Therefore, analysing the effects of synoptic cyclones on the dynamical processes of sea ice

at both the synoptic and sub-daily scales are necessary for improving our capabilities to monitor its formation and evolution, and for ensuring a correct parameterisation of momentum transfers as the Antarctic ice extent retreats.

1.3. Aims and Objectives

Although sea ice is such a vital component of the global ocean and climate system, only recently has a line of research been created between different universities and institutes within South Africa and also internationally; despite our unique access to the Southern Ocean and Antarctica. Therefore, by building upon seminal works (e.g. Vihma et al., 1996; Doble et al. 2006; Heil et al., 2009; Leppäranta, 2011; Kohout et al., 2014, 2020; Vichi et al., 2019; Eayrs et al., 2019; Alberello et al., 2020), this project aims to increase our knowledge of the dynamic response of pancake ice to extra-tropical cyclones, and to waves propagating into the ice cover. This is essential to accurately model the evolution of the Antarctic MIZ so that we can improve our climate predictions. Historically, this topic has been not well understood as a result of the seasonal and spatial scarcity of data in the Antarctic MIZ, specifically for the processes driving ice formation in autumn and winter.

In contrast to previous studies, which have focussed on the Weddell Sea and the East Antarctic, this work specifically focusses on sea-ice dynamics in the Atlantic and Indian sectors of the Southern Ocean – a region that is less studied due to its harsh environment and remote location.

In this study, the principal research question is outlined as follows:

How does atmospheric and oceanic (waves and currents) forcing, and local ice conditions affect sea-ice dynamics in the Antarctic MIZ across several temporal and spatial scales?

To elaborate on this principal research question and support its investigation, the key objective of this project is to elucidate the underlying processes that contribute to the temporal and spatial variability of sea-ice drift and deformation in the Antarctic. The interactions between atmospheric and oceanic forcing and sea-ice dynamics in the Antarctic MIZ will be analysed to:

- (1) investigate how polar-cyclone conditions (strong winds and high-wave energy) affect free-drift conditions and reshape the edge of the MIZ at synoptic scales (Chapter 2 and 3),
- (2) investigate how wind-forcing events affect sea-ice deformation at different spatial and temporal scales (Chapter 3),
- (3) investigate how wind-forcing events generate inertial oscillations in sea ice, which affect sea-ice drift and deformation at the sub-daily timescales (Chapter 2 and 3),
- (4) investigate how oceanic-forcing influences sea-ice drift and deformation at the synoptic and sub-daily temporal scales as well as regionally (Chapter 3), and
- (5) investigate the relationship and correlation between wind-ice and wave-ice interactions of pancake ice and interstitial frazil ice during the sea-ice formation and retreat phases, incorporating the dissipative underlying physical processes of sea-ice dynamics and wave attenuation (Chapter 4).

This project intends to demonstrate the highly dynamic and complex nature of the Antarctic MIZ, where the evolution and spatial patterns of ice dynamics are affected by the balance between external forcing mechanisms and local sea-ice conditions.

1.4. Overview of data and methods

Several types of data sources will be used in this study and combined together to answer and obtain the aims and objectives outlined in the section above. Data measurements will be assembled from several clusters of drifting ice buoys deployed in the Atlantic and Indian Ocean sectors of the Southern Ocean (Fig. 1.4); high resolution sea-ice concentration satellite estimates; atmospheric reanalysis data; modelled ocean-current data; and in-situ, satellite and reanalysis wave data. The environmental data will then be co-located to the temporal and spatial positions of the buoy measurements and combined to create comprehensive analyses of the effects of wind and wave forcing on free-drift dynamics in the Antarctic MIZ. All methods required for analyses are fully detailed within each chapter. All data are outlined below:

- (1) *In-situ ice observations*: Remotely tracked ice buoys, which are deployed on and move with ice floes as Lagrangian drifters, are a convenient method to observe ice movement and deformation. They are able to follow individual ice floes, showcasing the erratic motion of sea ice at high temporal and spatial scales, unlike satellite estimates which rather display the general and long-term movement of ice. These buoys (see Table C1) will include the following collected in-situ measurements: the 2017 Winter Cruise buoys (analysed in Chapter 2 and 3), which were deployed by Machutchon et al. (2019) and Eayrs et al. (2019b) in the Indian-Ocean sector of the Southern Ocean; the 2019 Winter and Spring Cruise buoys (analysed in Chapter 3), which were deployed by de Vos et al. (2023) and Womack et al. (2023) in the north-eastern Weddell Sea as part of the Southern oCean seAsonal Experiment (SCALE; analysed in Chapter 3); and the 2017 Polynyas, Ice Production, and seasonal Evolution in the Ross Sea expedition (PIPERS) winter campaign (analysed in Chapter 3), which were deployed by Kohout et al. (2020). The deployed ice buoys have generated temporally and spatially high-resolution datasets of in-situ measurements of metocean conditions in the Antarctic MIZ during the growth and retreat seasons.

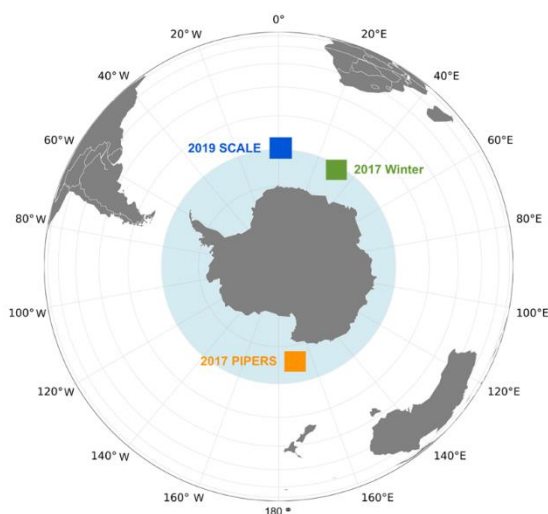


Figure 1.4. Schematic of the locations where each buoy array was deployed and in which year.

- *Analysis of ice observations:* The raw data acquired from the deployed buoys have been derived to fully analyse the spatial and temporal distribution of sea-ice drift and deformation in response to ice-landing cyclones. Additionally, the buoy data contain the sub-daily and true daily information on Lagrangian ice drift and deformation. In turn, these higher-frequency data are vital to differentiate atmospheric and oceanic forcing, as well as to define the inertial signature in sea-ice dynamics. Therefore, these datasets are important for the Antarctic MIZ as they not only supply high resolution measurements, which are vital for understanding the continuously changing MIZ conditions, but also providing long-term observations, which have been historically sparse.
- (2) *Sea-ice concentration data:* This study uses daily sea-ice concentration (SIC) data from remote sensing, acquired from the passive microwave AMSR2 sensor at a 3.125 km horizontal spatial resolution (Spreen et al., 2008), and complemented by the lower resolution Special Sensor Microwave Imager/Sounder (SSMIS) product (Cavalieri et al., 1996), obtained from the Copernicus Marine Service Information (CMEMS, 2014), which produces SIC with a spatial resolution of 25 km.
 - (3) *Atmospheric data:* Synoptic-scale atmospheric data have been obtained from the ERA5 reanalysis product at an hourly frequency and with a 0.25×0.25 degrees horizontal spatial resolution (Copernicus Climate Change Service (C3S), 2017).
 - (4) *Ocean-currents data:* Chapter 2 utilised data of current velocity from Copernicus GlobCurrent database (Copernicus Marine Service Information (CMEMS), 2014).
 - (5) *Wave field data:* Waves-field data have been obtained from a number of different platforms. In Chapter 2, the significant wave height was obtained from a WIIOS buoy, which was deployed as part of the 2017 Winter Cruise, and supplemented by ERA5 reanalysis data at the ice edge. In Chapter 4, wave parameters, namely the significant wave height and mean wave periods, were obtained from ship motion data through the wave-buoy analogy approach. The motion data were measured from an on-board IMU, during the 2019 Winter and Spring Cruises to the MIZ. Chapter 4 additionally utilised ERA5 wave field data, satellite altimetry data that were extracted from the Australian Ocean Data Network (Ribal and Young, 2019), and two Surface Wave Instrument Float with Tracking (SWIFT) buoys, which deployed during the 2019 SCALE cruises by Thomson et al. (2023).

The code used to process the data and produce the figures for Chapters 2 and 3 is available at <https://github.com/mvichi/antarctic-buoys/>.

Chapter 2 ¹

2. Atmospheric drivers of a winter-to-spring Lagrangian sea-ice drift in the Eastern Antarctic marginal ice zone

Abstract

Sea-ice drift in the Antarctic marginal ice zone (MIZ) is discussed using data from a four-month-long drift of a buoy deployed on a pancake ice floe during the winter sea-ice expansion. We demonstrate increased meandering and drift speeds, and changes in the dynamical regimes of the absolute dispersion during cyclone activity, together with high correlations between drift velocities and wind from atmospheric reanalyses. This indicates a dominant physical control of wind forcing on ice drift and the persistence of free-drift conditions. These conditions occurred despite the buoy remaining largely in > 80 % ice concentrations and at distances > 200 km from the estimated ice edge. The drift is additionally characterised by a strong inertial signature at 13.47 hours, which appears initiated by passing cyclones. A wavelet analysis of the buoy's velocity confirms that the momentum transfer from winds at the multi-day frequencies is due to atmospheric forcing, while the initiation of inertial oscillations of sea ice has been identified as the secondary effect. Propagating storm-generated waves may initiate inertial oscillations by increasing the mobility of floes and enhance the drag of the inertial current. This analysis indicates that the Antarctic MIZ in the Indian Ocean sector remains much wider and mobile, during austral winter-to-spring, than defined by sea-ice concentration.

¹ This chapter is based on the work published as: Womack, A., Vichi, M., Alberello, A., and Toffoli, A.: Atmospheric drivers of a winter-to-spring Lagrangian sea ice drift in the Eastern Antarctic marginal ice zone. *Journal of Glaciology*, 1-15. <https://doi.org/10.1017/jog.2022.14>, 2022.

2.1. Introduction

The advance and retreat of Antarctic sea-ice is the largest annually recurring event on Earth (Massom and Stammerjohn, 2010), and a significant portion of its natural variability appears to be associated with synoptic variability (Uotila et al., 2011; Vichi et al., 2019). However, our current understanding of these interactions is still incomplete, especially the effects of passing cyclones on the marginal ice zone (MIZ; Vichi et al., 2019). The MIZ is a highly dynamic and complex region of the (partially) ice-covered ocean, where the interactions between the ocean and atmosphere are stronger and more variable, and where weather conditions such as polar cyclones produce the most rapid and often extreme changes (Sect. 1.1.3; Andreas et al., 2010; Zhang et al., 2015; Vichi et al., 2019, Alberello et al., 2019). The evolution of the Antarctic MIZ, and also the emerging Arctic MIZ (Wadhams et al., 2018), is characterised and governed by the dynamics and thermodynamics of frazil and pancake ice (Sect. 1.1.2; Doble et al., 2003; Doble and Wadhams, 2006) – small, circular and mobile ice floes that form during turbulent ocean conditions at the outer regions of the ice cover. This outer region is conventionally between 5 km to 100 km wide (Feltham, 2005; Heorton et al., 2014), and is usually defined as the area of the ocean covered by 15 %-80 % sea-ice concentration (Strong et al., 2017; Rolph et al., 2020). The use of the concentration-based definition is a conventional operational choice derived from Arctic consideration, which may be less suitable in the Antarctic (Vichi, 2021, and references therein). Its application to the Antarctic MIZ is subject to the limited validation of remote sensing products by in-situ observations. Very little field data of metocean (meteorological and oceanographic) conditions are available in the Southern Ocean and even less in the MIZ (Derkani et al., 2021). Therefore, although the operational definition is applicable in the Arctic, where there is a good agreement between the ice type and ice cover, it is less reliable in the Southern Ocean, where the ice type is less related to the concentration value (Vichi, 2021). This study is therefore not based on a fixed definition of the MIZ, as constrained by concentration thresholds, but it rather considers the Antarctic MIZ extent based on the region of young and/or fractured sea ice that is continuously affected by atmosphere-ocean interactions, in the form of heat and momentum exchanges, wind and ocean current drag, and wave-ice interactions (Kohout et al., 2014; Zhang et al., 2015; Vichi, 2021).

Sea ice moves as a consequence of external forces – atmospheric and oceanic forcing – and internal ice stresses but, the principal factor being the wind forcing (Vihma et al., 1996). Storm-induced waves are generated close to the ice edge, and can propagate hundreds of kilometres into the ice cover, leaving behind a wake of less consolidated freezing floes (Squire, 2007; Kohout et al., 2014). This has been suggested to delay the consolidation of the ice cover and maintain the mixed frazil-pancake field (Vichi et al., 2019), causing the ice cover to become more susceptible to winds and underlying currents (Dumont et al., 2011; Kohout et al., 2014). Winds transfer momentum to the ice cover, and under free-drift conditions i.e. the absence of internal stresses, the sea ice changes from moving to the left of isobars (southern hemisphere) to moving almost parallel to the isobars, with a linear relationship to the surface wind velocities (Wassermann et al., 2006). This ratio between the ice drift speed and wind speed is known as the wind factor and is a key determinant for the dynamics of sea ice (Nakayama et al., 2012), with the angle between the drift vectors being the turning angle (Doble and Wadhams, 2006). The general rule-of-thumb states that the wind factor is 2 % for pack ice (Leppäranta, 2011), however higher values have been reported for pancake ice conditions, both in the Arctic (Lund et al., 2018) and the Antarctic (Doble and Wadhams, 2006).

Sea ice, simultaneously with the ocean, responds to the atmospheric forcing at both the synoptic (multi-day) and inertial (sub-daily) frequencies (periods; Heil et al., 2009). Wind forcing events such as the passing of polar cyclones may resonantly force inertial oscillations of sea ice on the left (southern hemisphere) of their track (Lammert et al., 2009). Therefore, drift trajectories often include circular or elliptical loops – the inertial oscillations – superimposed on an approximately steady translation (McPhee, 1988). This inertial response of sea ice has been reported for both the Arctic (Gimbert et al., 2012; Lammert et al., 2009; Leppäranta et al., 2011) and the Antarctic (Doble and Wadhams, 2006; Heil et al., 2009; Alberello et al., 2020). In free-drift conditions or when sea ice is broken, the ice floe is expected to oscillate as an oceanic fluid parcel. However, the friction between the bottom of the ice and the ocean surface or internal stresses caused by mechanical interactions within the sea-ice cover may cause the inertial oscillations to dampen within a few days (Gimbert et al., 2012).

In this paper, we showcase one of the longest ice buoy trajectories that never left the Antarctic MIZ, as it drifted for four months from the South Atlantic sector to the Indian Ocean sector of the Southern Ocean, under the influence of several synoptic cyclones. We correlate the in-situ drift measurements with atmospheric and oceanic reanalysis data, and quantify the impact of wind forcing on ice drift, through the direct transfer of momentum and initiation of inertial oscillations of the sea ice. This analysis highlights the effects that polar cyclones have on ice drift and the wider Antarctic MIZ.



Figure 2.1. The ice conditions during the deployment of the buoy on a pancake floe using the ship crane on 5 July 2017. The diameter of the basket is 1.5 m. Photograph taken by Sebastian Skatulla.

2.2. Methods

2.2.1. Field measurements and post processing

On 5 July 2017, a Trident Sensors Helix Beacon (Table C1) was deployed on a pancake ice floe ≈ 3 m in diameter (Fig. 2.1) at 62.8° S and 30.2° E from the S.A. Agulhas II as one of five buoys deployed within the MIZ (Sect. 1.4; Machutchon et al., 2019; Eayrs et al., 2019b), ≈ 100 km from the northward ice edge. The buoy was a non-floating device, which transmitted its GPS position data at a sampling frequency of four hours through the Iridium system until its signal was lost on 1 December 2017 at 61.5° S and 55.0° E, due to ice melting. While it stood approximately 1-m in height, the buoy's sensor had little sail effect for a large ice floe. The other four buoys only transmitted information until around 20 July 2017, and have been previously analysed by Vichi et al. (2019) and Alberello et al. (2020), over this shorter timeframe.

The buoy's longitude and latitude positions were used to derive its zonal and meridional velocity components using the linear approximation:

$$u = \frac{\Delta x}{\Delta t} \quad (2.1)$$

$$v = \frac{\Delta y}{\Delta t} \quad (2.2)$$

where Δx and Δy are the zonal and meridional geodesic distances travelled between consecutive points along the buoy's trajectory, and Δt is the time interval (four hours). The error estimation of the ice-drift velocity can be attributed to two sources, the first being the tracking error between two consecutive GPS points. However, when displacements (Δx and Δy) are retrieved from drifting buoys, the tracking error (σ_{tr}^2) is 0, since buoys remain fixed relative to the ice floe of deployment (Dierking et al., 2020). Secondly, there may be errors in the GPS location of the buoy – the geolocation error (Lindsay and Stern, 2003). The calculation of the displacements is prone to the errors of the coordinate readings, making its uncertainty $\sigma_d^2 = 2\sigma_{coord}^2$ (Dierking et al., 2020). However, as seen in Fig. 2.2, the buoy drifted with minimal latitudinal change ($< 2^\circ$) and therefore we may assume identical geolocation errors, which would cancel out when calculating the drift velocity. The major uncertainty of the buoy's velocity would be rather due to the timing of the GPS locations, however Hutchings and Hibler (2008) found that velocity errors are less than 10 % for sampling intervals greater than one hour.

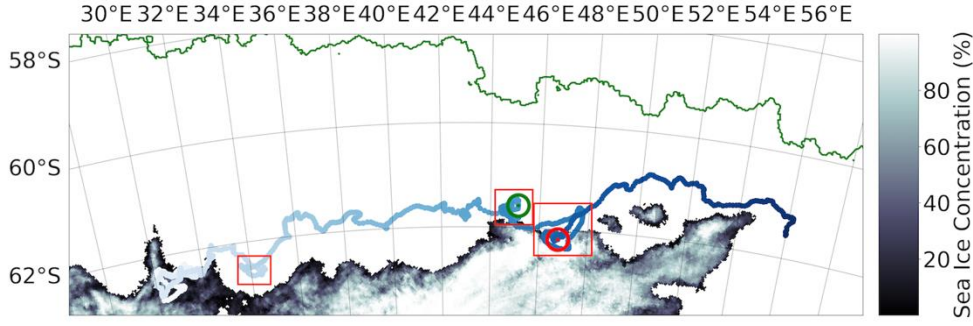


Figure 2.2. The full trajectory of the buoy (time gradient trajectory from light blue to dark blue) from 5 July to 1 December 2017. The shadings show the AMSR2 SIC with values between 0 %-100 % on 1 December. The green circle denotes the end of the advance season on 30 September. The red circle denotes the beginning of the melt season on 1 November. The green line indicates the ice edge at 0 % ice concentration for 30 September, the maximum northward extent of sea ice. The three red boxes indicate the three loops within the buoy’s trajectory.

The meander coefficient acts as a primary quantitative measure for the kinematics of sea-ice drift (Vihma et al., 1996). As ice drifts under the influence of erratic winds, ice floes change direction and deflect from their primary drift direction. This results in their drift trajectories exceeding the geodesic distance between two points. The ratio between these two distances is called the meander coefficient and is calculated as:

$$M = \frac{I}{\Delta D} \quad (2.3)$$

where ΔD is the geodesic displacement between the start and end points of the experiment and I is the total cumulative trajectory length, determined by the four-hourly positions of the buoy, for the four-month period. We computed this as a time series for each four-hourly position of the buoy, showing the progressive meander coefficient as the time window increased. We then additionally computed two discrete meander coefficients over constant intervals of one day and five days for the full trajectory of the buoy in order to analyse how the passing cyclones affected its drift and its seasonal variation. A greater meander coefficient signifies a more erratic trajectory, whereas a value of $M = 1$ indicates that the buoy travelled in a straight line. It must be noted that although the meander coefficient is a ratio between two distances, it is also a function of time, as it is dependent on the sampling intervals of the buoy’s positions and its deployment duration (Heil et al., 2009), as will be presented in Section. 2.3.3.

Additionally, the framework developed by Lukovich et al. (2017) is used to assess the directional changes of the buoy’s drift path in response to the atmospheric forcing, estimated by ERA5 reanalyses. The framework is based on Lagrangian statistical analyses using particle dispersion theory, which shows us whether ice drift is in a sub-diffusive, diffusive, ballistic, or super-diffusive dynamical regime using single-particle (absolute) dispersion statistics (Lukovich and other, 2017). A number of prior studies exist using Lagrangian dispersion and ice-buoy trajectories to quantify sea-ice drift and deformation in the Arctic (e.g. Rampal, 2008, 2009), however it has not been considered to our knowledge in the Antarctic.

Single-particle (absolute) dispersion provides a signature of circulation and organised structure in the flow field, and also shows the linear time dependence in fluctuating velocity variance characteristic of turbulent diffusion theory (Taylor, 1922; Rampal et al., 2009). It is defined as (Taylor, 1922):

$$A^2 = \langle |x_i(t) - x_i(0)|^2 \rangle \quad (2.4)$$

where x_i is the zonal and meridional position of the i -th particle in the ensemble as a function of the passed time, t , and where the angular brackets denote the ensemble mean. However, as this analysis is based on only one buoy, the ensemble mean could not be computed as multiple buoys would be needed. Therefore, the ergodic assumption was applied, which allowed us to calculate the absolute dispersion of the buoy in time instead of space.

Drift dynamics are characterised by the scaling exponent β (the slope between the single-particle dispersion of sea-ice drift and the elapsed time) according to the relation:

$$A^2 \sim t^\beta \quad (2.5)$$

where $\beta > 1$ relates to a super-diffusive dynamic regime, $\beta = 1$ to a diffusive regime and $\beta < 1$ to a sub-diffusive or “trapping” regime (Lukovich et al., 2017). A super-diffusive regime captures long-range correlations and organised structure in the sea-ice drift field that is unrestricted by diffusive energy losses for example through ice mechanics (Lukovich et al., 2017). A diffusive regime describes the behaviour of buoys/ice floes that follow independent random walks, while a sub-diffusive regime denotes the trapping that would occur with dominating contributions from ice-ice interactions (Lukovich et al., 2017). In this study, we compare the scaling exponent of the total absolute dispersion during the periods of high meandering to the overall scaling exponent to identify transitions between the dynamical regimes and the effect that atmospheric forcing has on ice drift.

2.2.2. Environmental conditions

Environmental conditions were obtained from satellite data and reanalysis products for the full duration of the buoy’s drift. The daily sea-ice concentration (SIC) as well as extent (SIE) were obtained from remote sensing, acquired from the passive microwave Advanced Microwave Scanning Radiometer 2 (AMSR2) sensor at a 3.125 km horizontal spatial resolution and a daily frequency (Spren et al., 2008). This was complemented by the lower resolution Special Sensor Microwave Imager/Sounder (SSMIS) product, which produces SIC with a spatial resolution of 25 km. SIC isolines are used in this analysis as subjective reference values to be considered in relation to the buoy location and drift conditions. Given that there are known differences between the products of the order of 10 % during winter and spring (Beitsch et al., 2014) and the known limitations of using a fixed threshold (Kern et al., 2020), we consider SIC reference values from > 0 % to 80 %. We will refer to the 0 % isoline as the ice edge, although this region is acknowledged to be heterogeneous and fragmented. Results are only marginally affected by considering the standard 15 % value, with a maximum deviation of approximately 50 km, which is still in the range of differences between satellite products. We also note that using the 15 % SIC contour would lead to an early melt that was not recorded by the Trident buoy, which is a non-floating device.

The most southerly $> 0\%$ ice concentrations were determined for each longitude within the defined sample region and calculated for each day within the four-month drift. The distance between the buoy and the sea-ice edge was thereafter calculated as the meridional geodesic distance between the two latitudes, along the same longitude, for each day of the buoy's drift. Following this, the meridional drift velocity of the ice edge was additionally computed as the change in the daily ice edge latitudinal positions for the same longitude as the buoy. This was calculated using the following equation:

$$V_w = \frac{(lat_{n+1} - lat_n)\Delta y}{\Delta t} \quad (2.6)$$

where V_w is the average daily meridional change of the ice edge in m s^{-1} , n is the time index of the discretized series of the ice edge latitudes, Δy is the length of one degree latitude in m and Δt is the time interval in seconds.

ERA5 reanalysis was used to obtain synoptic scale meteorological and oceanic conditions – the surface wind velocities (at 10 metres), the mean sea level pressure, the air temperature (at 2 metres), and the significant wave height (up to 15 % ice concentration) – given at an hourly frequency and with a 0.25×0.25 degrees horizontal spatial resolution. The validation of the reanalysed atmospheric variables in the Southern Ocean against the ones measured along the cruise track for the duration of the expedition is available in Vichi et al. (2019).

2.2.2.1. Synoptic cyclones identification

The South Atlantic sector of the Southern Ocean, and particularly the eastern Weddell Sea, is a region of net cyclolysis for cyclones developing near South America and other regions of open ocean genesis (Hoskins and Hodges, 2005; Yuan et al., 2009; Vichi et al., 2019). The majority of severe cyclones occur in winter in the Atlantic at around 0° , 30° , and 90° E, with good agreement among several storm tracking methods (Grieger et al., 2018). These cyclones, with diameters ranging between 500 km - 2000 km (Hoskins and Hodges, 2005; Uotila et al., 2011), are associated with strong winds, and are vehicles of moisture and heat to the high latitudes (Messori et al., 2017; Woods and Caballero, 2016), which in turn greatly affect the thermodynamics and dynamics of sea ice in this region.

The buoy exhibited many loops and meanders (Fig. 2.3) as it drifted under the influence of winds during cyclone activity. As polar cyclones have a typical frequency of one occurrence every five to seven days (Hoskins and Hodges, 2005; Vichi et al., 2019; references cited in the manuscript), and because of the relatively short period and confined trajectory of the buoy, they can be tracked without the need of automatic tracking algorithms. Therefore, to investigate the impacts that the cyclones have on ice drift, we applied a visual inspection method of the ERA5 mslp and 2-m temperature fields at four-hourly intervals, for the four-month drift of the buoy. Wei et al. (2016) reported that the average intensity for cyclones in the Southern Ocean, between 1979 to 2013, was 968.4 hPa in spring, 972.4 hPa in summer, 968.7 hPa in autumn and 967.4 hPa in winter. Therefore, for this study we only considered cyclones with core pressures < 970 hPa, as the buoy did not drift during austral summer. Ten cyclones were therefore identified by low-pressure troughs < 1000 km from the buoy, and by an increase in air temperature to near melting point close to the ice edge, on the eastern flank of the cyclone (Vichi et al., 2019).

Once the cyclones were identified, the dates when the cyclones were closest to the buoy were calculated using a nearest neighbour method and then related to the loops and meanders in the buoy's trajectory. A time series of the mslp in the vicinity of the buoy, as well as maps of the mslp and 2-m temperature fields for when these cyclones were closest to the buoy, overlaid with trajectories of the cyclones from 12 hours before to 12 hours after each cyclone was closest to the buoy, can be found in Appendix A, Figs. A1 and A3, respectively. This analysis however, does not include the analysis of the first 20 days of data which has been analysed by Vichi et al. (2019) and Alberello et al. (2020).

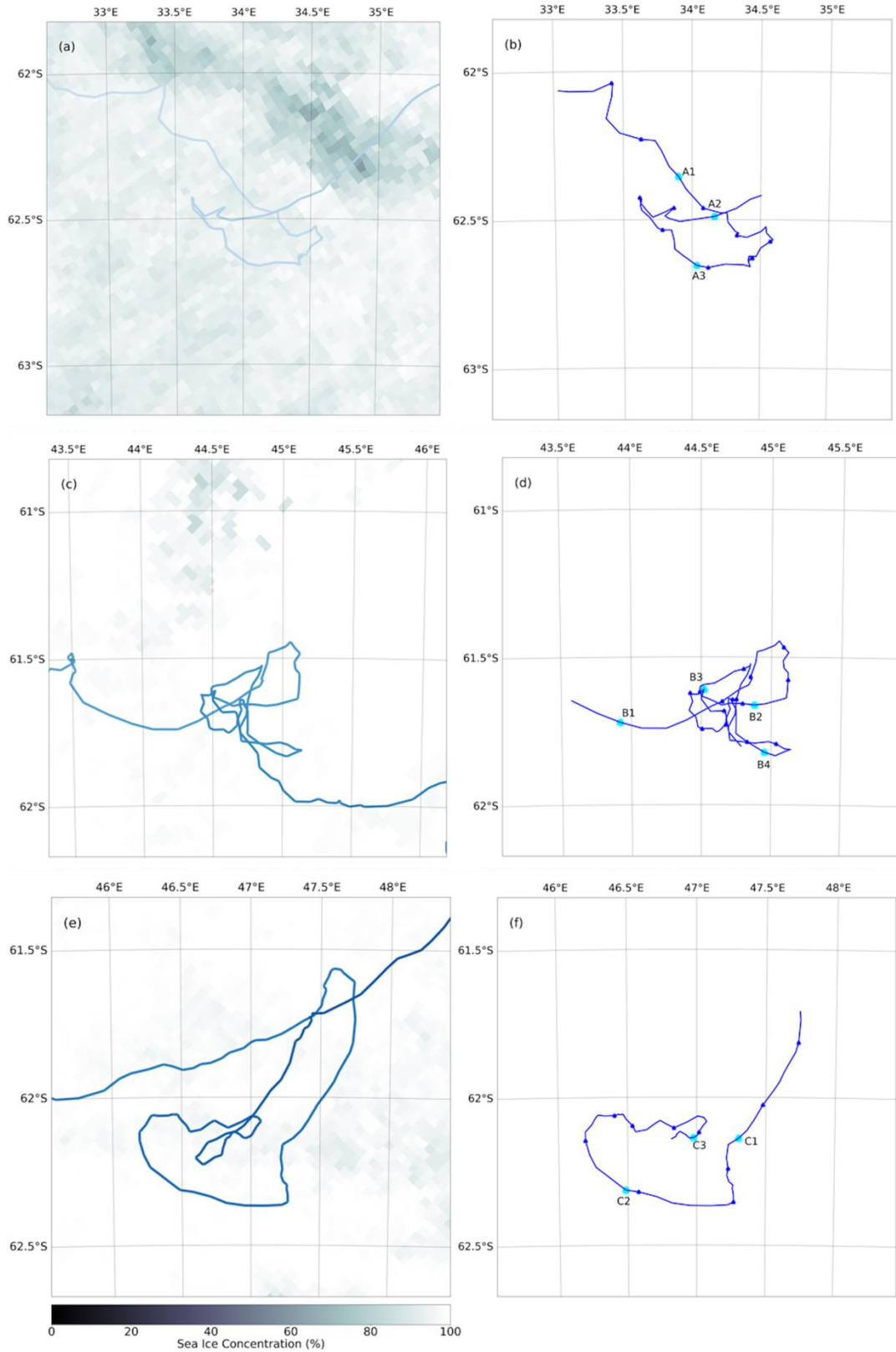


Figure 2.3. Three case studies of large loops in the drift trajectory of the buoy for (a-b) 9-20 August; (c-d) 26 September-12 October; (e-f) 21-31 October. The gradient blue lines in the left column are colour correlated to the time gradient colour in Fig. 2.2, and denote the buoy's trajectory, moving from left to right. The shadings show the AMSR2 sea-ice concentration between 0 %-100 % on the last day for each corresponding meander. The dark-blue points on the right column panels denote the position of the buoy at 12:00 pm every day, for approximately ten days; the light-blue points show when each cyclone was closest to the buoy using the labels from Table 2.1.

2.2.2.2. Estimated significant wave height

The buoy did not measure ocean waves, but waves-in-ice are available for the first 16 days of the deployment through other instruments. As it will be demonstrated in the results section through increased drift kinematics and high mobility, the buoy never left the MIZ and free-drift conditions. Therefore, we assume that there were no major changes in ice conditions from the observed conditions during deployment, and hence waves-in-ice conditions measured during this initial period can be considered representative of the MIZ. In order to estimate the significant wave height (H_s) at the buoy's location, the coefficient of wave attenuation was first estimated, using the modelled ERA5 wave fields (swell and waves) at the sea-ice edge and a 16-day time series of H_s measured by the Waves In Ice Observation System (WIIOs, Eayrs et al., 2019b). This WIIOs, analysed by Alberello et al. (2020) (buoy B2), was deployed ≈ 12 hours prior to the Trident Sensors Helix Beacon's deployment in the MIZ and followed a very similar drift path, which can be seen in the Appendix A, Fig. A3. This allowed for the assumption that the waves-in-ice activity was the same for both buoys during the first 16 days of deployment.

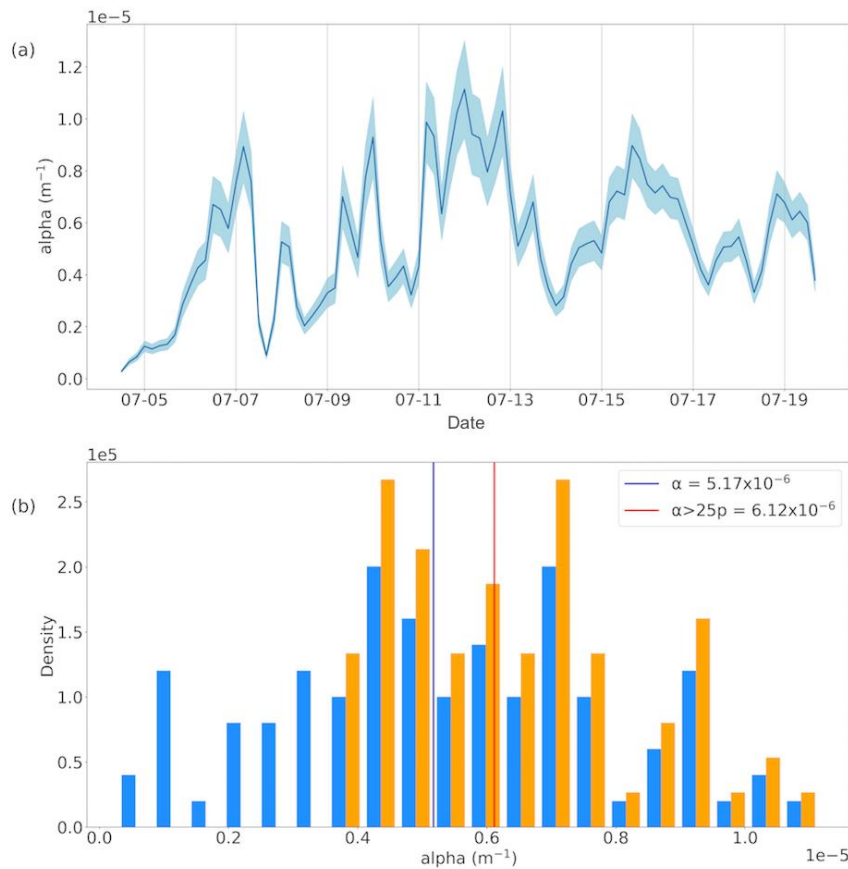


Figure 2.4. (a) The time series of the attenuation distance and its error propagation shaded in blue for the WIIOs from 4-19 July 2017 and (b) its corresponding distribution (blue) and the distribution when the 25th percentile is removed (orange). Both corresponding medians are denoted.

In this simple approximation, the modelled ERA5 wave energy in the ocean at the sea-ice edge is assumed to decay exponentially to the location of the WIIOS (Kohout et al., 2020), and to travel meridionally south along the same longitude. A time series of the wave attenuation was calculated using the following equation:

$$\alpha = \frac{1}{DY} \times \log \left(\frac{H_{Sw}}{H_{Si}} \right) \quad (2.7)$$

where DY is the geodesic latitudinal distance (from the ice edge to the buoy's position) in metres, H_{Sw} is the ERA5 modelled H_S at the 0 % SIC sea-ice edge for the same longitude as the WIIOS, and H_{Si} is the H_S measured by the WIIOS. Wave attenuation should be a function of the ice features, however we did not make this assumption since SIC changes quite quickly from 0-100 % in the MIZ, and waves have been observed to propagate in 100 % SIC (Alberello et al., 2020). As explained in Section 2.2.2, our results are only marginally affected by the SIC threshold used in this study and therefore there is little sensitivity to this choice of the > 0 % sea-ice edge in the incident wave field. The resulting coefficient and its distribution are shown in Fig. 2.4. The median of this time series was then determined to be $\alpha = 5.17 \times 10^{-6} \text{ m}^{-1}$, with an error (shaded region in Fig. 2.4a) calculated through the error propagation formula:

$$std_{\alpha} = \alpha \times \frac{std_{DY}}{DY} \quad (2.8)$$

where std_{DY} is the standard deviation of the latitudinal distance in meters. However, as the median was biased towards the initial period of high waves affected by the extremes (Vichi et al., 2019) we removed the lowest quartile of the time series and once again calculated the median, which was then determined to be $\alpha = 6.12 \times 10^{-6} \text{ m}^{-1}$ (also shown in Fig. 2.4b). For comparison, Kohout et al. (2020) reported attenuation coefficient ranges of $\alpha = 1.6\text{-}5.0 \times 10^{-6} \text{ m}^{-1}$ in low sea-ice concentration (≤ 80 %) and $\alpha = 26.5\text{-}32.7 \times 10^{-6} \text{ m}^{-1}$ in high sea-ice concentrations (> 80 %), in the Ross Sea. The higher estimated attenuation coefficient was then used to compute the H_S time series (H_{Si}) in sea ice at the buoy's location for its full four-month drift. This was calculated using the following equation:

$$H_{Si}(t) = H_{Sw}(t)e^{-\alpha DY(t)} \quad (2.9)$$

where H_{Sw} is the re-analysed ERA5 H_S in the ocean at the sea-ice edge for the same longitude as the buoy.

We acknowledge the limitations of calculating the wave attenuation and the significant wave height within the sea-ice cover using this approach, as one needs to consider the temporal and spatial variability of direction, ice conditions, ice extent, wave event duration and wave speed (Kohout et al., 2020). This estimate is therefore an approximated method to consider the role of waves throughout the buoy's drift, and the possible relationship between waves, ice and passing cyclones.

2.2.2.3. Correlation between buoy drift and wind forcing

The wind factor F and turning angle θ were estimated using the least squares regression method from Kimura and Wakatsuchi (2000). This calculation used the time series of wind and sea-ice drift, neglecting the temporal variation of the ocean current beneath the ice cover, due to the lack of ocean current observations (Nakayama et al., 2012). In this case, the wind factor and turning angle subsequently absorb the influence of the ocean current.

The variables U_w , V_w , and u_i , v_i denote the zonal and meridional components of wind and ice, respectively. The index of the discretised series of wind and ice drift is denoted by k , with n being the upper bound of the summation. These equations can be written as:

$$\theta = \arctan \left[\frac{\sum_{k=1}^n \{U_w(k)v_i(k)\} - \sum_{k=1}^n \{V_w(k)u_i(k)\}}{\sum_{k=1}^n \{U_w(k)u_i(k)\} + \sum_{k=1}^n \{V_w(k)v_i(k)\}} \right] \quad (2.10)$$

$$F = \frac{c_1 + c_2 - c_3 + c_4}{\sum_{k=1}^n \{U_w^2(k)\} + \sum_{k=1}^n \{V_w^2(k)\}} \quad (2.11)$$

where

$$c_1 = \cos\theta \sum_{k=1}^n \{U_w(k)u_i(k)\} \quad (2.12)$$

$$c_2 = \sin\theta \sum_{k=1}^n \{V_w(k)u_i(k)\} \quad (2.13)$$

$$c_3 = \sin\theta \sum_{k=1}^n \{U_w(k)v_i(k)\} \quad (2.14)$$

$$c_4 = \cos\theta \sum_{k=1}^n \{V_w(k)v_i(k)\} \quad (2.15)$$

From here, the vector correlation coefficient R_v^2 can be calculated using the following equation adopted from Kimura and Wakatsuchi (2000):

$$R_v^2 = \left(\frac{c_1 + c_2 - c_3 + c_4}{\sqrt{\sum_{k=1}^n U_w^2(k) + \sum_{k=1}^n V_w^2(k)} \sqrt{\sum_{k=1}^n u_i^2(k) + \sum_{k=1}^n v_i^2(k)}} \right)^2 \quad (2.16)$$

In addition, the Pearson correlation coefficient R_p^2 was also calculated to determine the linear relationship between the modulus of ice drift and wind:

$$R_p^2 = \left(\frac{\text{cov}(w,i)}{\sigma_w \sigma_i} \right)^2 \quad (2.17)$$

where cov is the covariance, σ_w is the standard deviation of the wind speed and σ_i is the standard deviation of the ice drift.

2.3. Results: drift measurements and analysis

2.3.1. The impact of cyclones on sea-ice drift

Over the four-month drift, the buoy travelled a net transition displacement of ≈ 1336 km within the Indian Ocean sector of the Southern Ocean, with minimal latitudinal change ($< 2^\circ$) (Fig. 2.2). During this eastward drift, the buoy's trajectory was characterised by sharp turns and loops, in response to ten cyclones (Table 2.1) which passed over/near the buoy. These cyclones, although not part of the top 1 % of winter cyclones (Wei and Qin, 2016), had minimum pressures within the range of 940-960 hPa and maximum wind speeds between 10-21 m s^{-1} , which greatly impacted the buoy's drift (Fig. 2.3).

The left-hand-side of Fig. 2.3 highlights these three large loops superimposed on the AMSR2 sea-ice concentration for the last day of each corresponding loop, when the buoy was dynamic and mobile at high ice concentrations ($> 80\%$). The right-hand-side shows the buoy's trajectory with its daily position at 12 pm, and its position when each cyclone was closest to it.

Table 2.1: Features of the cyclones which occurred during the loops as simulated by ERA5

Cyclone	Dates	Maximum Wind Speed at the buoy's location	Core Air Pressure
<u>Loop A</u>			
Cyclone A1	9-12 August	15 m s^{-1}	940 hPa
Cyclone A2	15-17 August	15 m s^{-1}	954 hPa
Cyclone A3	18-20 August	21 m s^{-1}	960 hPa
<u>Loop B</u>			
Cyclone B1	26-27 September	20 m s^{-1}	952 hPa
Cyclone B2	30 September-2 October	13 m s^{-1}	956 hPa
Cyclone B3	4-7 October	20 m s^{-1}	944 hPa
Cyclone B4	11-12 October	14 m s^{-1}	960 hPa
<u>Loop C</u>			
Cyclone C1	21-23 October	16 m s^{-1}	964 hPa
Cyclone C2	24-27 October	17 m s^{-1}	952 hPa
Cyclone C3	30-31 October	10 m s^{-1}	968 hPa

The first large cyclonic loop, occurring between 9-20 August, was a result of the three polar cyclones A1-A3 (Table 2.1). The buoy initially travelled south-east during the passage of cyclone A1. Cyclone A2 then caused the buoy to drift north-west before switching north-easterly under the influence of cyclone A3. The second loop along the buoy's trajectory occurred between 26 September-12 October at approximately 10° eastward of the previous loop. This tighter loop was a result of the four cyclones B1-B4, which travelled eastwards impacting the sea-ice cover. Cyclone B1 initially caused the buoy to travel in a south-eastwards direction before switching to a north-eastwards direction. The second cyclone (B2) resulted in several changes in the buoy's drift direction; a southerly, westerly and then a north-easterly drift. During the last two cyclones (B3 and B4), the buoy travelled in two small anti-clockwise loops before finally drifting in a south-easterly direction. The final large loop, occurring between 21-31 October, was much more elongated. Cyclone C1 caused the buoy to drift in a south-westerly direction, before shifting to a southerly direction. The second cyclone (C2) caused the buoy to drift north-westerly, before switching to a north-easterly direction under the influence of cyclone C3.

2.3.2. Distance from ice edge

Figure 2.5 shows the latitudinal extent of the buoy's trajectory and the sea-ice edge from the AMSR2 and the SSMIS sensors, as well as the locations of the 80 % isolines which mark the putative extent of the MIZ. The SSMIS ice edge was generally situated further north than the AMSR2 ice edge, except between mid- to late-September when the AMSR2 ice edge peaked at $\approx 58^\circ$ S, while the SSMIS ice edge shifted south to $\approx 59^\circ$ S. Additionally, the AMSR2 ice edge and 80 % isoline is more variable than the one of the SSMIS sensor. The biases between the AMSR2 and SSMIS sensors can be due to the differences in algorithm sensitivity to ice edge conditions and due to the different spatial resolution of the sensors (Meier and Steward, 2019; Kern et al., 2019).

Despite the difference in the sensors, the buoy was always south of the 80 % ice concentrations, only drifting into lower SICs in mid-November after the melt season began. However, according to the AMSR2 sensor, the buoy left the ice cover on 29 November – 2 days before the buoy stopped transmitting. This discrepancy is likely due to an underestimation of the low ice concentrations by the satellite product. This can be attributed to the limited capability of the AMSR2 sensor in detecting thin ice near the ice edge, particularly during the melt season (Liu et al., 2020). This caused the sea ice to be interpreted by the high-resolution ASI algorithm as ocean. There are inherent uncertainties with ice concentrations lower than 15 % as brash ice, small ice floes or flooding by wave action may cause the ice to be interpreted as open ocean (Meier et al., 2017; Liu et al., 2020).

As the transmission period of the buoy started at the beginning of July, the full asymmetrical seasonal cycle of the Antarctic sea-ice cover is less visible. The ice edge was already quite north in July, therefore only half of the sea-ice advance season is shown. However, the rapid retreat from the beginning of November until the end of the transmission period is more prominent. The buoy's location slowly shifted northward until mid-October. Afterwards, while the ice edge was rapidly retreating for both sensors, the buoy was pushed further northwards (the maximum northward location from the point of deployment was attained in mid-November, see Fig. 2.5 right axes).

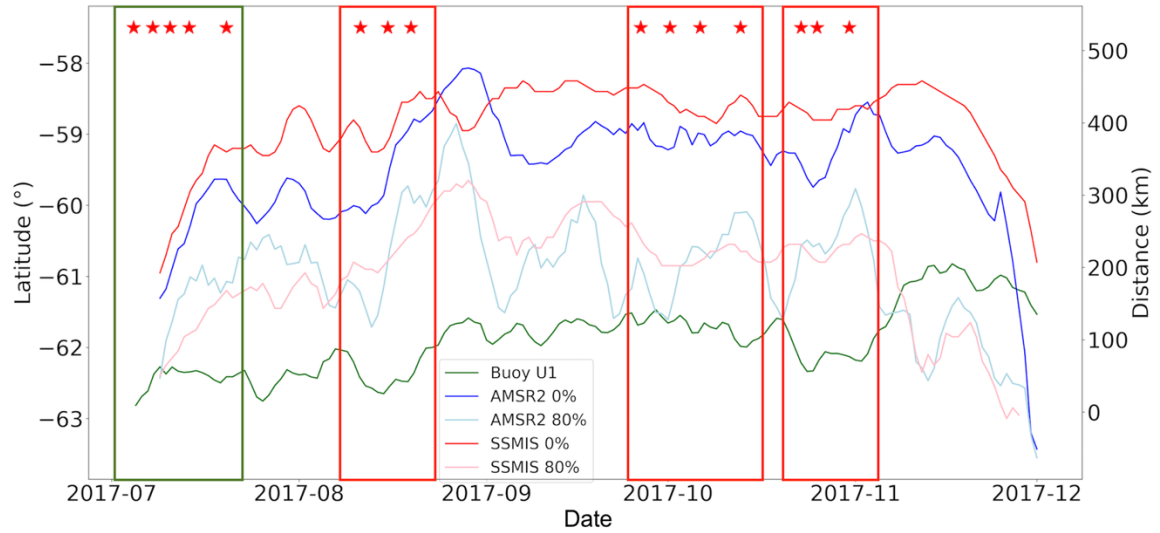


Figure 2.5. Latitudinal location of the buoy (green line) and the ice edge, defined as the $> 0\%$ sea-ice concentration contour, along the same longitude, for each day of the buoy's trajectory. The dark blue and light blue lines denote the latitude of the AMSR2 ice edge and 80 % contour, respectively. The red and pink lines are from the SSMIS sea-ice product. The red star symbols indicate the time when the storm cores were closest to the buoy's location, clustered in the three sets listed in Table 2.1. The green box denotes the five storms previously examined by Vichi et al. (2019) and Alberello et al. (2020). The right axis is scaled to show the relative distance of each line from the initial point of deployment.

Figure 2.5 additionally shows the relative distance from the point of deployment (left axis, assuming a geodesic distance along the meridian). This confirms that the buoy was generally over 200 km south of the ice edge, while additionally reaching distances of ≈ 350 km between September and October, independently of the ice-edge location uncertainty. There was an initial 200 km advance of the ice in less than a month, which was likely due to thermodynamic growth (Doble and Wadhams, 2006), causing the freezing front to move northwards, however the extent of this northwards movement was not followed by the buoy.

2.3.3. Metocean drivers

Figure 2.6 compares the meridional velocity components of the sea-ice edge estimated from the AMSR2 SIC and the buoy meridional drift. The ice edge is displaced meridionally during the passage of cyclones (Vichi et al., 2019), which is partly visible in the buoy displacement.

Although the buoy spent most of its drift over 200 km from the ice edge (Fig. 2.5), where the mobility of ice is assumed to decrease due to consolidation, Fig. 2.6 indicates that the buoy drifted similarly to the ice edge, although with a dampened signal. Fig. 2.7 further shows that the zonal and meridional velocities of the buoy motion varied according to wind forcing. During the periods of frequent cyclone activity, indicated by the four rectangular boxes, the sea-ice velocity notably increased with the wind velocity, while additionally following the direction of the wind's velocity components. The overall drift velocity of the buoy does indicate a high correlation to wind forcing, suggesting minimal if not absent internal stresses of the ice, indicative of ice dynamics being in consistent free drift conditions throughout the trajectory. However, there are a few increases in both the wind and ice drift speeds that do not correspond to the passage of a cyclonic event. Therefore, it is important to quantify the drift response objectively through the analysis of the wind factor and turning angle, which will be done later in this section.

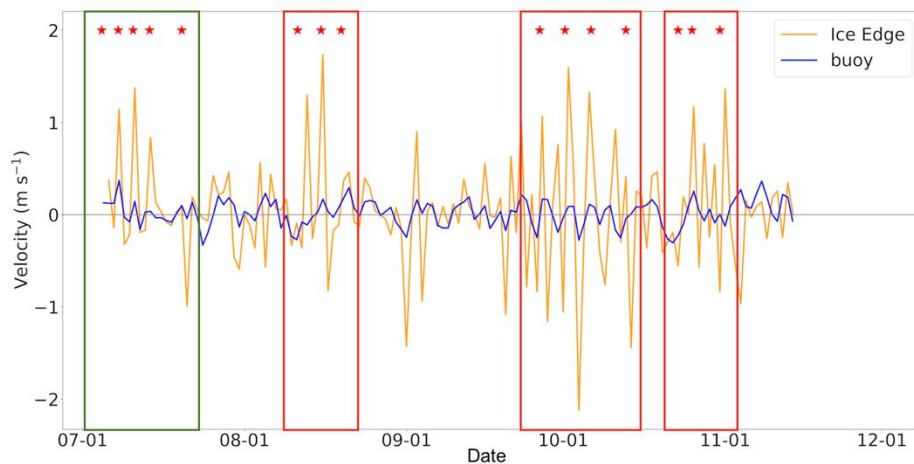


Figure 2.6. The meridional drift velocity of the estimated sea-ice edge (orange) and the buoy (blue). This does not include the last few days when the ice edge started melting (from ≈ 15 November). The grey horizontal line indicates 0 m s^{-1} . The other graphical elements are as described in Fig. 2.5.

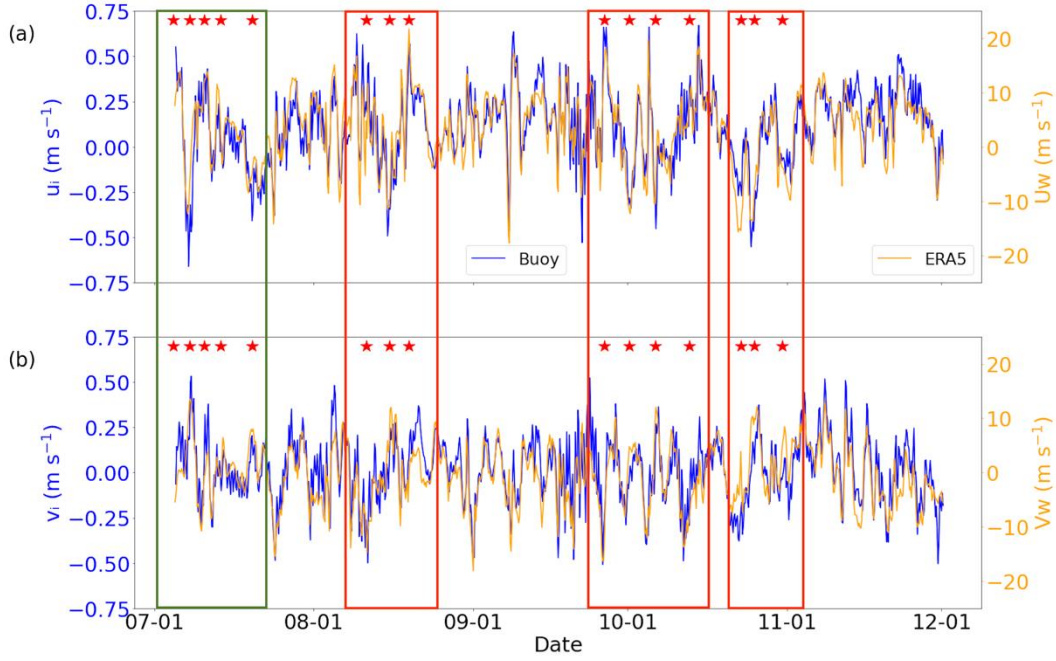


Figure 2.7. Velocity components of the buoy (on the left axis) and 10-m wind from the ERA5 reanalyses (on the right): **(a)** zonal component and **(b)** meridional component. The other graphical elements are as described in Fig. 2.5.

The response of the buoy to the passage of the cyclonic systems is further examined by analysing its meander coefficient (Fig. 2.8) and absolute dispersion (Fig. 2.9) as described in Section 2.2.1. The buoy travelled a total trajectory distance of ≈ 3322 km with a net transition distance of ≈ 1336 km resulting in a final meander coefficient of 2.49 at the end of the buoy's four-month drift (Fig. 2.8). This conformity of the eastwards drift of the buoy is also highlighted in Fig. 2.9 where the total absolute dispersion is governed by the zonal dispersion. This can be a result of the steering influence of the Antarctic Circumpolar Current (ACC). Similarly, Heil et al. (2009), using a 37-day buoy array with a half-hourly frequency, reported low meander coefficients (1.3-2.7) near the coast within East Antarctica, and attributed this relatively straight-line ice drift to be governed by the Antarctic Coastal Current (ACoC). In comparison, Heil et al. (2008) reported that hourly meander coefficients, derived for the western Weddell Sea over a 26-day interval, were considerably larger and varied from 4.4 over the continental shelf break to 9.7 on the shallow continental shelf. We recognize that the computation of the meander coefficient is affected by the period over which it is computed. We have therefore compared the cumulative meandering coefficient with the daily and 5-daily periods. The latter is consistent with the periodicity of storms in this region (Hoskins and Hodges, 2005). We are therefore mostly interested in the relative changes rather than in the absolute values.

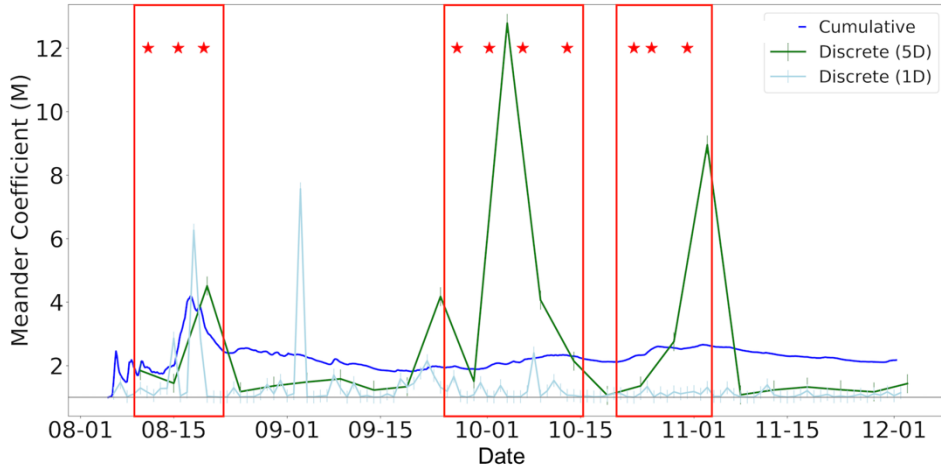


Figure 2.8. Time series of the meander coefficient (shown only from 5 August just before the first large loop until 1 December which was the last day of transmission). The dark-blue line denotes the progressive meander coefficient and the light-blue line denotes the daily meander coefficient. The horizontal grey line at $M = 1$ denotes straight-line drift. The red star symbols indicate the time when the storm cores were closest to the buoy location. The three red rectangular boxes denote the three sets of cyclones indicated in Table 2.1.

The daily meander coefficient increased with the passage of each cyclone, particularly before the fourth (B1) and just after the sixth (B3) cyclones. In comparison, the meander coefficient computed over a 5-day time window rather captured the combined effect of the three groups of cyclones, indicated by the rectangular boxes. These increases in the meander coefficients would be due to the strong winds associated with the cyclones (see Table 2.1). The rises in the daily meander coefficient contributed to the cumulative meander coefficient, which increased to ≥ 3 during the periods of the frequent cyclone activity and caused the three large loops (Fig. 2.3). However, similar to the velocity of the buoy (Fig. 2.7), there were days when the daily meander coefficient notably increased during the periods of reduced cyclonic activity. This can be seen by smaller turns and deflections in the buoy’s trajectory (Fig. 2.2), and is consistent with the indication of free drift conditions and a high correlation to wind forcing as explained above. Towards the end of the trajectory as it started nearing the melt season, the daily and 5-day meander coefficients were close to 1, when ACC would have likely become a more dominant steering influence.

These meandering events greatly affected the meridional dispersion (Fig. 2.9) causing it to notably decrease, when the buoy would have been pushed southwards by winds on the eastern flank of the cyclones. However, the zonal dispersion was not as affected, although there were small decreases during the periods of frequent cyclone activity but it shows an overall gradual increase as the buoy drifted eastwards. Recalling that the scaling exponent β describes sea-ice dynamical regimes, $\beta = 2.77$ for the total absolute dispersion of the full drift period, however the total dispersion is only shown from 5 August, before the first large loop, to 1 December. This indicates that, while drifting within $> 80\%$ SIC, the buoy was characterised predominantly by a super-diffusive regime, denoting organised structure in the flow field. However, during the three groups of cyclones sea-ice dispersion was rather characterised by a sub-diffusive regime $\beta < 1$, which captures interruptions in organised flow due to ice-ice interactions (Lukovich et al., 2017, 2021). These interruptions or “trapping” events would have resulted in the three large loops as the ice cover was compressed by the meridional winds and the displacement of the buoy was

limited. The increase in the meander coefficient during cyclone activity, along with the transitions between these dynamical regimes indicate an erratic nature of ice drift, suggesting that the drift patterns (in Fig. 2.3) are significantly related to atmospheric forcing.

To further investigate the physical control of wind forcing on the buoy's drift, the observed drift speed and direction was compared with ERA5 wind vectors, using the least squares regression method detailed in Sect. 2.2.2.3. The wind factor for the buoy showcased a range of 1 %-6 % with an overall mean wind factor of 2.73 %, and although it is higher than the rule-of-thumb value of 2 %, it is lower than the range of 3 %-3.5 % reported by Doble and Wadhams (2006) for pancake ice in the Weddell Sea. This would be due to the steering influence of the ACC on the buoy. The turning angle was small with a mean value of -19.83° and generally large variations between -50° to $+50^\circ$, which are consistent with Uotila et al. (2000) who reported values from -20° to $+60^\circ$ in the Weddell Sea, using buoy positions that were interpolated at three-hourly intervals. These quantitative measures indicate a high wind response of the buoy throughout its four-month drift.

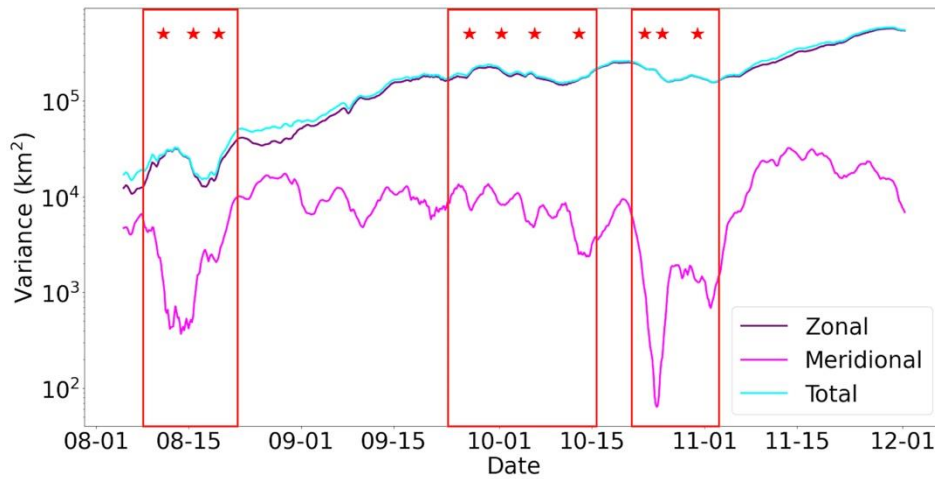


Figure 2.9. The absolute (single-particle) dispersion statistics for the buoy, from 5 August just before the first large loop until 1 December which was the last day of transmission, depicting the zonal (purple line), meridional (magenta line) and total (cyan line) dispersion. The three red rectangular boxes denote the three sets of cyclones indicated in Table 2.1.

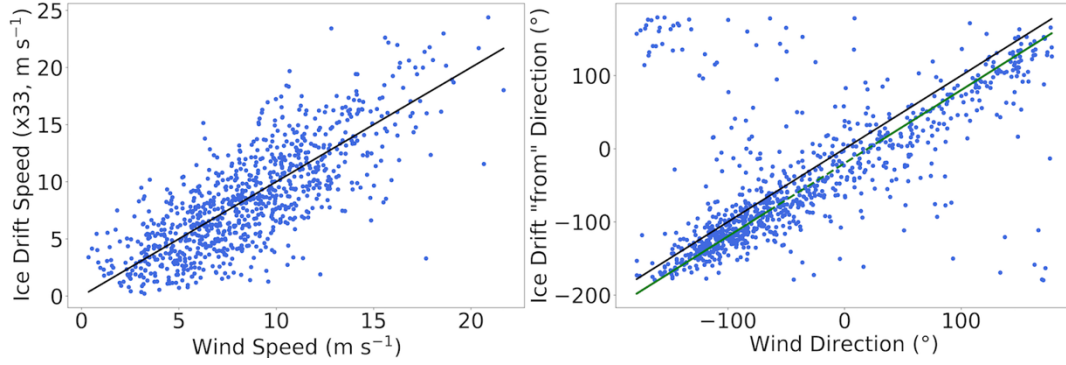


Figure 2.10. Scatterplots (left) of ice drift speed as a function of wind speed and (right) of ice drift direction as a function of wind direction. The black line is the 1:1 line. The green line is the turning angle at -19.83° .

The relationship between the buoy and the wind vectors is shown in Fig. 2.10 (left), based on the indicators shown in Equation 2.10 and Equation 2.11. The ice drift speeds have been multiplied by 33 so that the 1:1 line represents an ice drift of 3 % of the wind speed. The measured drift speeds are dispersed relatively evenly around the 1:1 line, with a Pearson correlation (Eq. 2.17) of $R_p^2 = 0.60$. In comparison, Alberello et al. (2020) and Doble and Wadhams (2006) reported similar but slightly lower values of $R^2 = 0.56$ and $R^2 = 0.5$, respectively, for pancake ice conditions in the Antarctic. They, however, did not report the relationship between ice drift and wind direction, shown in Fig. 2.10 (right). Here the majority of the points are closer to the 1:1 line, with the bulk of them sitting just below it, at the lower portion of the -19.83° turning angle line. However, during periods of quiescence when the wind stresses are smaller, larger turning angles may exist, which could have resulted in a decreased Pearson correlation to $R_p^2 = 0.40$.

The vector correlation, expressed by the correlation coefficient in Eq. (2.16), is high with a value of a $R_v^2 = 0.76$. This is higher than the Pearson R_p^2 for both the speed and direction, and although they cannot be fully compared, it does indicate that the relationship between ice drift and wind vector is significant. This suggests that the wind not only impacted the buoy's speed, through the transfer of momentum, but it additionally impacted the path of the buoy. Both methods confirm the good correlation between ice drift and wind forcing, suggesting that the winds are the primary contributor towards the observed drift of the buoy, even in regions where the eastwards-flowing ACC is expected to have a greater control.

Intense polar cyclones have been linked to wave propagation through the ice-covered ocean (Vichi et al., 2019). Since the drifting regime of the buoy was highly correlated to atmospheric variability and did not show major seasonal changes during the trajectory, we have used the methods described in Section 2.2.2.2. to estimate H_5 at the buoy location assuming a constant ice type over the period. Fig. 2.11 illustrates the timing of the storm-generated waves in the open ocean from the ERA5 reanalyses and the estimates of in-ice waves at the buoy location using the mean attenuation coefficient derived from ice conditions at the time of deployment. The reconstructed attenuation coefficient during the measured period (green box) leads to wave heights that are comparable with the observed peak intensity, but overestimates the background value, which is indicative of changes in the ice conditions between the storms. It is therefore likely that the estimated H_5 at the buoy location is also overestimated in terms of the background. For reference, we also show the attenuation rates from Kohout

et al. (2020) for $\leq 80\%$ and $> 80\%$ SIC (see Sect. 2.2.2.2). The buoy drifted mostly in $> 80\%$ SIC, while the estimated H_s (blue line) appears to be more similar to the H_s range for $\leq 80\%$ SIC, rather than the $> 80\%$ SIC range where the H_s dissipates completely also during the wave observation period in the green box. We thus infer that during storm periods, when the buoy drift was not dissimilar from the one measured during the wave observation period, it is possible that some wave energy may have penetrated at distances ≈ 350 km from the ice edge.

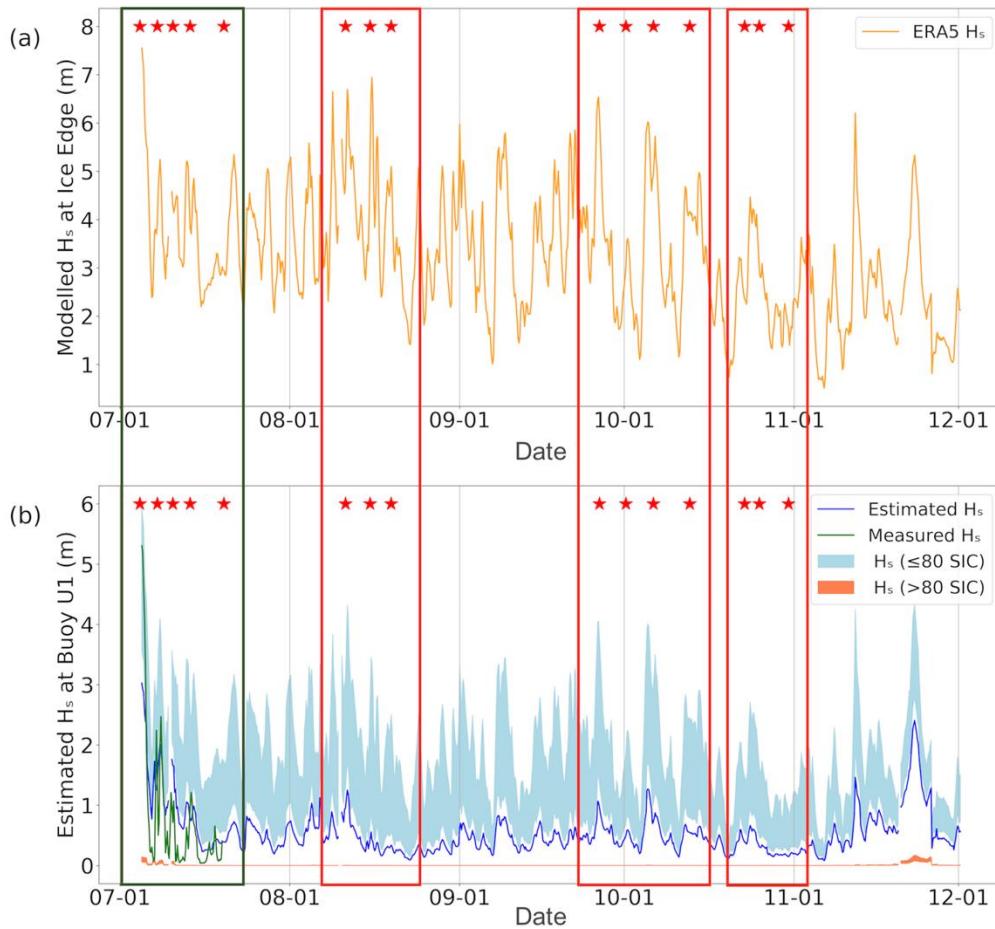


Figure 2.11. Time series of (a) the modelled H_s at the ice edge from ERA5 (orange line) and (b) the estimated H_s at the buoy's position for the four-month drift period. (blue line). The measured H_s of the WIOS for the first 16 days of the buoy's drift is denoted by the green line. The blue and orange shaded areas denote the H_s of the buoy calculated using the attenuation coefficient ranges from Kohout et al. (2020) for ice concentrations $\leq 80\%$ and $> 80\%$ respectively. All the other graphical elements are as in Fig. 2.5.

2.3.4. Spectral analysis

A power spectral analysis was computed for the velocity components of the buoy to further examine how sea ice responds to wind forcing at both the multi-day and sub-daily (inertial) frequencies. Fig. 2.12 shows the power spectral density (PSD) for the four-month drift of the buoy and the ERA5 wind velocity components. The wind velocity shows a classical, continuous energy cascade from the lower (multi-day) frequencies to the higher (sub-daily) frequencies. The ice drift velocity is also subject to a similar energy cascade. However, it indicates a clear energy peak at a frequency just below 2 cycles day⁻¹, at 13.47 hours, as determined by Alberello et al. (2020) for the first two weeks of deployment. The period of these oscillations falls within the theoretical inertial range of 13.47-13.72 hours between 61° S-62° S. As the buoy was deployed and operated in the deep Southern Ocean away from the continental shelf, we did not include the role of tidal forcing that is known to be negligible in off-shelf buoy deployments (Heil et al., 2009; Lei et al., 2021). Tides are effective drivers of sea-ice dynamics in coastal polynyas and close to ice shelves, where they may affect the melting rates (e.g. Hausmann et al., 2020). In addition, the buoy's trajectory was close to an amphidromic point (Kamphuis, 2020: Fig. 7.5), where the tidal fluctuation is notably small (between 20 and 60 cm at most).

A wavelet analysis was additionally performed to yield information about the time series and the frequencies together. To properly isolate the mechanisms that excite inertial oscillations of sea ice in relation to the storm passage, a high-pass Butterworth filter was applied to the buoy's velocity time series in order to filter out the lower frequencies longer than the day. Figure 2.13 shows the filtered wavelet power spectrum (left panel) and the filtered wavelet spectrum (right panel) of both the zonal and meridional velocity components of the buoy. Both components have statistically significant power throughout the time series (being inside the cone of influence), at both the lower and near-inertial, sub-daily frequencies. However, the majority of this power still lies within the lower frequencies despite the use of the high-pass filter. There are three large power intensifications within the lower frequencies (≈ 128 , ≈ 320 and ≈ 512 hours, corresponding to 5, 13 and 21 days respectively) for the zonal component, with a weaker and lagged response for the meridional component. These three power maxima occurred between ≈ 5 -20 July (the period analysed by Vichi et al. (2019) and Alberello et al. (2020)), ≈ 6 -26 August and ≈ 30 September-10 November. The latter two correspond with the dates of the three large loops in Fig. 2.3. This is due to the direct transfer of momentum from the wind forcing, which can be seen in Fig. 2.12 where the sea ice, although with less power, follows the energy cascade of the wind within the lower frequencies.

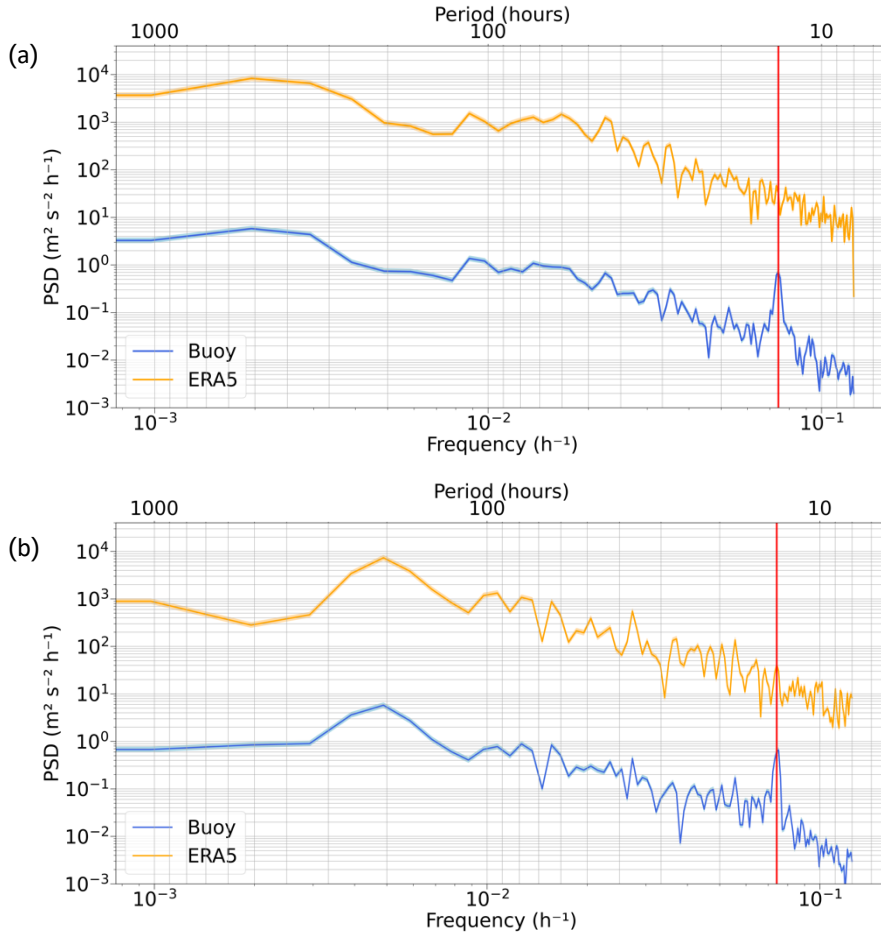


Figure 2.12. Power Spectral Density corresponding to the **(a)** zonal component of ice drift and the **(b)** meridional component of ice drift of the buoy for four months. The blue line indicates the buoy, and the orange line indicates the ERA5 winds. The vertical red line indicates the peak associated with inertial oscillations (13.47 hours).

In addition to the synoptic response of the sea ice, the momentum transfer from the winds at the lower frequencies initiated an inertial response, which appears to occupy a well-defined frequency band (see Fig. 2.13). This is because synchronous with the ocean, sea ice provides a medium that effectively allows the transfer of energy from the lower frequencies (wind forcing) to the semi-diurnal frequencies (Heil et al., 2009). Therefore, a continuous energy cascade from the lower frequencies of wind velocity is needed to produce both the synoptic and inertial frequency responses in ice drift. A closer visual inspection of Fig. 2.13 suggests that the dates of the strongest inertial oscillations coincided with the dates of the lower-frequency power maxima. More notably, these inertial power maxima additionally coincided with the dates of the three large loops (in Fig. 2.3); when (1) the ice drift of the buoy changed to be dominated by high-frequency oscillations (Lammert et al., 2009), and when (2) the rotation velocity was likely greater than the advection velocity (Gimbert et al., 2012), during “trapping” events captured by the sub-diffusive dynamical regimes (Fig. 9), resulting in the increased meander coefficient (Fig. 2.8).

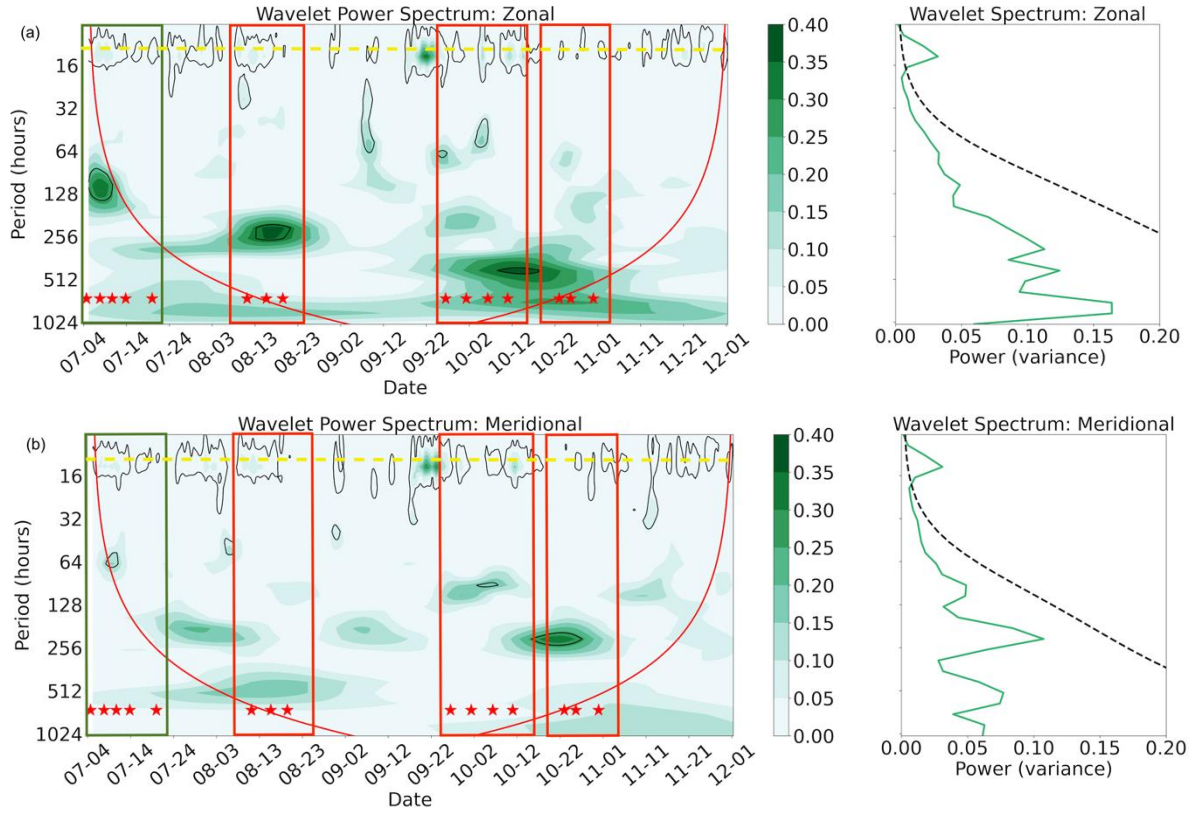


Figure 2.13. The Wavelet Power Spectrum (left panel) and the Wavelet Spectrum (right panel) of the filtered velocity spectrum of the buoy for its four-month drift, where **(a)** denotes the zonal component and **(b)** the meridional component. The red line indicates the cone of influence (left panel), the black contours (left panel) and black dashed lines (right panel) indicate the 95 % significance level. The yellow-dashed line (left panel) denotes the inertial frequency (13.47 hours). All the other graphical elements are as in Fig. 2.5.

Inertial oscillations, although occurring more regularly than the synoptic events at the lower frequencies, are shorter lived and dissipate within a few days (see Fig. 2.13). This dissipation is expected to be due to kinetic energy dissipation within the Ekman layer, friction occurring between ice floes and the surface ocean (Gimbert et al., 2012), as well as internal stresses of the ice floes. By means of the outcome of the meandering coefficient (Fig. 2.8) and the total absolute dispersion (Fig. 2.9), it is speculated that these forces dampen the rotation, which resulted in a more straight-line trajectory of the buoy, before new oscillations were excited.

2.4. Discussion and conclusions

The four-month trajectory from the South Atlantic sector to the Indian Ocean sector of the Southern Ocean, was continuously impacted by polar cyclones. Ten of these cyclones were found visually, using methods described in Section 2.2.1, to determine the coupling between the sea ice and atmospheric forcing occurring when polar cyclones pass over the ice edge. The average daily meridional change of the ice edge estimated from the AMSR2 satellite showed large fluctuations (Fig. 2.6) as it was forced by the passing cyclones as reported previously by Vichi et al. (2019). This indicates that surface winds, particularly the meridional winds associated with polar cyclones, played an important role in reshaping the ice edge (Kwok et al., 2017; Eayrs et al., 2019a; Vichi et al., 2019).

While the buoy spent most of its drift over 200 km from the ice edge (Fig. 2.5), where the mobility of ice is assumed to decrease due to consolidation, the buoy's trajectory showed large meandering and a dynamic response to wind forcing, as it displayed sharp turns and loops while drifting under the influence of the winds (Fig. 2.3). The final meander coefficient was 2.49, indicating a more straight-line trajectory for the buoy's overall drift. However, during periods when the ten cyclones passed near/over the buoy, all methods describing the meander coefficient increased (Fig. 2.8). Additionally, while the overall scaling exponent ($\beta = 2.77$) of the total absolute dispersion showed a super-diffusive regime; during cyclone activity there was a transition to a sub-diffusive ($\beta < 1$) regime, characterised by trapping, resulting in the three large loops (Fig. 2.3). Together, the meander coefficient and the total absolute dispersion indicate a more erratic nature of ice drift during these periods, suggesting that the drift of the buoy was significantly related to wind forcing during the passage of the cyclones. The ice drift speed also notably increased during the passage of the cyclones (see Fig. 2.7). This strong relationship was confirmed by the high mean wind factor of 2.73 %, the small mean turning angle of -19.83° and good linear correlations of 0.60 and 0.40 for the speed and direction, respectively, and 0.76 for the vector correlation. Together, these all indicate that wind forcing had a dominant physical control on the ice floe trajectory throughout the four months, from winter to spring. Additionally, along with the high mobility of the buoy, this high wind response suggests that free-drift conditions likely persisted throughout the buoy's drift.

Although the buoy (and the underlying ice movement) revealed a high wind response, it drifted mostly at distances > 200 km from the ice edge (0 % SIC), while also reaching distances of ≈ 350 km, between August and September. Additionally, according to both the SSMIS and AMSR2 sensors, the buoy remained in > 80 % ice concentrations, only drifting into lower concentrations in mid-November after the melt season began (Eayrs et al., 2019a). During this retreat of the SIE, the buoy was pushed further northwards to its maximum northward location from its point of deployment. This can be attributed to the increased mobility of sea ice during the melt season, where decreases in ice strength and internal stresses lead to more ice deformation, fracturing and increased drift speeds (Tandon et al., 2018). At the end of the transmission period there was however a discrepancy between the buoy and the AMSR2 sensor, as the satellite product indicated the buoy was outside the ice cover two days before the buoy was lost (presumably due to ice melt and hence sinking of the buoy). This has been attributed to the limited capability of the AMSR2 sensor in detecting thin ice near the ice edge during the melt season (Liu et al., 2020). In the overall four months, the SSMIS and AMSR2 sensors often differed from each other (Fig. 2.5). This can be due to the algorithms used by the products and/or their different spatial resolutions. Further analyses will therefore need to be done to better assess the ice extent to determine their accuracy, particularly during the melting period when their error grows. Beitsch et al. (2014) determined that the overall best agreement with ship-based observations and satellite-derived ice concentrations was found using the Bootstrap algorithm, which is in agreement with our finding that SSMIS is a better estimator of the ice edge location in spring.

We speculate that the high mobility of the buoy may have been maintained by heterogeneous ice conditions like the ones observed at deployment. The buoy did not measure waves along its trajectory, and by estimating a constant attenuation coefficient obtained from the first 16 days of deployment we can speculate that wave energy from the edge may have penetrated in regions of apparent full ice coverage from space. This is a coarse approximation, since it assumes that wave direction, ice conditions, ice extent, wave event duration and wave speed are represented by the conditions observed during the period of direct observations in early July. However,

this simple parameterization of wave attenuation suggests that with the high wind response observed, sea-ice conditions during peaks of wave energy in the open ocean may have mechanical properties similar to the pancake-frazil conditions observed during the buoy's deployment, even when the satellite products indicated > 80 % ice concentrations.

A spectral analysis was done to further elucidate the influence of wind forcing on ice drift. Our results demonstrate that the PSD of both the wind and the ice drift velocity components exhibited a continuous energy cascade (Fig. 2.12), with majority of their power found within the lower frequencies and a clear energy peak in the inertial range. As the intense polar cyclones passed over/near the buoy, their strong winds excited inertial oscillations of the ice, since ice is embedded within the surface of the ocean (Heil et al., 2009). The use of a filtered wavelet power spectrum highlighted the response of the buoy to atmospheric forcing at both the lower (synoptic) and inertial frequencies (Fig. 2.13). It showed three large power intensifications at the synoptic scales of 5, 13 and 21 days for the zonal component, with a lagged response in the meridional component. These power maxima coincided with the three groups of cyclones detected in the atmospheric reanalyses (Table 2.1). This suggests that there was a direct transfer of momentum from the winds to the sea ice at these scales, although further analysis involving dedicated numerical models (i.e. built to simulate this phenomenon) will be required in the future.

The analysis revealed that in addition to the synoptic response of the buoy, there is a plausible correlation between the onset of inertial oscillations and the presence of polar cyclones. The most intense inertial responses occurred during the periods of frequent cyclone activity. This relationship cannot be fully demonstrated because sometimes the power at the inertial frequency was found outside of the dates of higher cyclone activity. Additionally, these power intensifications were often excited before the increase in power at the lower frequencies. The propagation of waves into the ice cover can affect the local drift of ice floes and how they respond to wind and ocean forcing. Therefore, we propose that the penetration of storm-generated waves may be able to help initiate the inertial oscillations observed in the sea-ice drift, allowing the power at the inertial frequency to increase before the power at the lower frequencies. Additionally, as waves helped to maintain the pancake-frazil ice conditions, we speculate that this allows the geostrophic current to keep the weaker inertial oscillations during periods of storm quiescence. Alberello et al. (2020) attributed the strong inertial signature at ≈ 13 hours to the geostrophic current and to full free-drift conditions. It is however unclear how these periods of synoptic events (5-20 days) may drive the initiation of inertial oscillations, as for instance by breaking the consolidated ice into smaller floes (e.g. Gimbert et al., 2012), or by maintaining a more fluid sea-ice type throughout the extended Antarctic MIZ. The strong influence at the lower (synoptic) frequencies has however been identified as the primary effect of atmospheric forcing and the initiation of inertial oscillations of sea ice has been identified as the secondary effect. The secondary effect greatly influences ice drift at shorter timescales and causes deviations of the ice drift from a straight-line path.

This analysis confirms that the buoy did not leave the Antarctic MIZ throughout its four-month drift. It also suggests that instead of a local transition from smaller pancake ice floes into a consolidated frozen pack ice (Doble and Wadhams, 2006), there is instead a northward push of a frontal band of unconsolidated sea ice. The exact extent of this frontal band and the features of the ice type left behind cannot be inferred from the available satellite data and would require further observational studies.

Based on these observations, the MIZ in this Antarctic sector is much wider than previously thought (see Sect. 2.1), even throughout the peak period of the winter sea-ice expansion. The buoy drifted within regions of 80 %-100 % ice concentrations, but the ice was not consolidated as suggested by the high mobility of the buoy. This implies that the concentration-based MIZ definition (see Sect. 2.1), where the ice concentration lies between 15 %-80 % is inadequate to describe the sea-ice type in the Antarctic MIZ. This is because the highly dynamic nature of the MIZ is maintained despite the high sea-ice concentrations observed from space, due to the complex interactions between sea ice, polar cyclones and storm-generated waves.

Chapter 3 ²

3. A contrast in sea-ice drift and deformation between winter and spring of 2019 in the Antarctic marginal ice zone

Abstract

Two ensembles of buoys, deployed in the marginal ice zone (MIZ) of the north-eastern Weddell Sea region of the Southern Ocean, are analysed to characterise the dynamics driving sea ice drift and deformation during the winter-growth and the spring-retreat seasons of 2019. The results show that although the two buoy arrays were deployed within the same region of ice-covered ocean, their trajectory patterns were vastly different. This indicates a varied response of sea ice in each season to the local winds and currents. Analyses of the winter data showed that the Antarctic Circumpolar Current modulated the drift near the sea ice edge. This led to a highly energetic and mobile ice cover, characterised by free-drift conditions. The resulting drift and deformation were primarily driven by large-scale atmospheric forcing, with negligible contributions due to the wind-forced inertial response. For this highly advective coupled ice-ocean system, ice drift and deformation linearly depended on atmospheric forcing. We also highlight the limits of commercial floating ice velocity profilers in this regime since they may bias the estimates of sea-ice drift and the ice type detection. On the other hand, the spring drift was governed by the inertial response as increased air temperatures caused the ice cover to melt and break up, promoting a counterintuitively less wind-driven ice-ocean system that was more dominated by inertial oscillations. In fact, the deformation spectra indicate a strong de-coupling to large-scale atmospheric forcing. Further analyses, extended to include the deformation datasets from different regions around Antarctica, indicate that, for similar spatial scales, the magnitude of deformation vary between seasons, regions and the proximity to the sea ice edge and the coastline. This implies the need to develop rheology descriptions that are aware of the ice types in the different regions and seasons to better represent sea-ice dynamics in the MIZ.

² This chapter is based on the work published as: Womack, A., Alberello, A., de Vos, M., Toffoli, A., Verrinder, R., and Vichi, M.: A contrast in sea ice drift and deformation between winter and spring of 2019 in the Antarctic marginal ice zone, *The Cryosphere*, 18, 205–229, <https://doi.org/10.5194/tc-18-205-2024>, 2024.

3.1. Introduction

Antarctic sea-ice forms a natural barrier between the atmosphere and the Southern Ocean, modulating the exchange of heat, gases, and momentum, and contributing to the global climate system balances (McPhee et al., 1987; Kohout et al., 2020). The seasonal sea-ice zone undergoes one of the largest annual changes on Earth (Allison, 1997; Massom and Stammerjohn, 2010), with $\approx 15 \times 10^6$ km² of ice that forms and subsequently melts each year (Eayrs et al., 2019). During the winter advance season, the ice cover is characterised by the formation of frazil ice and relatively free-floating pancake ice floes (Doble and Wadhams, 2006; Wadhams et al., 2018; Alberello et al., 2022), which form during wavy conditions (Meylan et al., 2014). The dynamics and thermodynamics of these roughly circular and mobile floes 1-10 m in diameter (Alberello et al., 2019; Alberello et al., 2022) control the evolution of the marginal ice zone (MIZ; Doble et al., 2003), i.e. the outer sea-ice region, where the interactions between the atmosphere and ocean are most intense and variable (Strong et al., 2017; Wadhams, 1986). The MIZ extent is primarily limited by the Antarctic Circumpolar Current (ACC), which flows clockwise around the Antarctic continent (Vihma et al., 1996), and acts as the northern boundary for seasonal ice formation. However, further into the MIZ, where the influence of open-ocean waves is reduced (Doble and Wadhams, 2006), larger pancakes can freeze together (Shen and Ackley, 1991) and eventually consolidate to form a coherent ice cover (Weeks and Ackley, 1986). Despite this anticipated seasonal consolidation, large variability in sea-ice concentration from space is observed at the monthly scale in regions of 100 % ice coverage (Vichi, 2022), which may increase heat loss from the ocean. During the spring retreat season, the surface heat balance changes, causing the consolidated ice to break up (Squire et al., 1995) and form floes and brash ice with a wide range of diameters. This creates positive feedback through the reduction in albedo and the presence of more open water, where waves can more freely propagate and break the ice (Kohout et al., 2014; Passerotti et al., 2022). This subsequently leads to the further melt and retreat of the seasonal ice cover.

On time scales of a day or more, sea ice moves in response to oceanic and atmospheric forcing (Chapter 2; Thorndike and Colony, 1982; Alberello et al., 2020), and is modified by internal ice stresses, which depend on the characteristics of the ice cover such as ice thickness, concentration and strength (Heil et al., 2009; Heil et al., 2011). However, wind forcing has been regarded as the primary forcing mechanism of ice drift (Sect. 2.4; Nansen, 1902; Allison, 1989; Vihma et al., 1996). The momentum transfer from the winds to the sea ice can be described by a linear ratio, the wind factor (Nakayama et al., 2012), with a turning angle between the ice drift and wind direction ranging between 0-30°, negative in the southern hemisphere (Leppäranta, 2011). Generally, the wind factor is 2 % for pack-ice conditions (Leppäranta, 2011), although larger values have been reported for pancake conditions in both the Arctic (e.g. Wilkinson and Wadhams, 2003; Lund et al., 2018) and the Antarctic (e.g. Alberello et al., 2020; Sect. 2.3.4). On shorter time scales, the inertia of sea ice becomes more important, and drift trajectories often include elliptical loops – inertial oscillations – superimposed on an approximately steady translation (McPhee, 1988). This is because sea ice, simultaneous with the ocean, responds to rapid changes in wind stress (Lei et al., 2021), such as the passage of cyclones (Hibler et al., 2006; Lammert et al., 2009; Gimbert et al., 2012a, b), at both the low (synoptic) frequencies to high (sub-daily) frequencies (periods; McPhee, 1988). However, unlike tidal forcing, the inertial response of sea ice has no direct high-frequency equivalent in neither the oceanic nor the atmospheric spectra (Heil and Hibler, 2002). Rather, a cascade of energy from the low

frequencies to high frequencies, within the wind spectra, is required to generate the inertial-frequency power in the ice drift and deformation (Heil et al., 2009). This cascade arises from non-linear ice dynamics, as ice drift transfers its kinetic energy to the underlying ocean (Leppäranta, 2011). From this, high frequencies can be fed in back into the ice drift, and become trapped close to the semi-diurnal frequencies (Heil et al., 2009).

The ice cover is also highly deformable, as a result of the differential drift of individual floes (Hibler, 1974; Geiger et al., 1998; Girard et al., 2009). Sea-ice deformation is a dynamic phenomenon that is constrained in space and time (Oikkonen et al., 2017; Rampal et al., 2019), but little is known about the heterogeneous Antarctic sea ice. During the last few decades, several methods have been developed to quantify sea-ice deformation. The most common approach to determine deformation is computed from the strain rates – divergence, shear and total deformation (Sect. 1.1.5.2; Lindsay, 2002; Leppäranta, 2011) – which are derived from the spatial gradients in the ice velocity field, at the vertices of polygonal buoy arrays (Stern and Lindsay, 2009; Hutchings and Hibler, 2008; Itkin et al., 2017; Aksamit et al., 2023). However, this leads to large uncertainties, since deploying buoys in pre-defined array shapes is often very challenging in the MIZ and thus not all buoy arrays are deployed sufficiently well to determine deformation in this manner (Rampal et al., 2009) where the patterns align unfavourably (de Vos et al., 2022). In lieu of this, alternative approaches that do not depend on buoy geometry and orientation can be useful, even if to merely determine and differentiate dynamic regions (Aksamit et al., 2023).

Lagrangian dispersion statistics are conventionally used to characterize paths and structures in atmospheric and oceanic dynamical phenomena in order to identify topological and dynamical features within a flow field (LaCasce, 2008; Lukovich et al., 2017). The approaches can be sub-divided into single- (or absolute) and multi-particles methods. However, both single- and multi-particle statistics are needed for a full description of ice floe evolution (LaCasce, 2008). Lagrangian dispersion statistics applied to ice-buoy trajectories have been used extensively to quantify ice drift and deformation in the Arctic (e.g. Rampal et al., 2008, 2009; Girard et al., 2009; Lukovich et al., 2011, 2015, 2017, 2021). However, to our knowledge, only absolute dispersion has been considered for the Antarctic, in Chapter 2 (Sect. 2.3.3), and only for a single-buoy trajectory. In general, absolute dispersion provides a signature of circulation and organised structure in the flow field (Lukovich et al., 2021). It also estimates the linear time dependence of the fluctuating velocity variance, characteristic of turbulent diffusion theory (Taylor, 1922). Variations in ice-fluctuating velocity statistics associated to turbulent diffusion are considered to be related to sea-ice deformation and internal ice stresses (Rampal et al., 2009; Lukovich et al., 2017).

Relative (two-particle) dispersion characterises the deformation of sea ice by using the temporal evolution of the separation between two Lagrangian trackers in the ice (Rampal et al., 2009; Lukovich et al., 2017). Martin and Thorndike (1985) showed that, in a statistical sense, there are similarities between the dispersion properties of sea ice and the dispersion of particles in turbulent fluids, although the underlying physics may be different. By using the separation of buoy pairs as a proxy of the combined strain-rate components, sea-ice relative dispersion is a way to analyse deformation (Rampal et al., 2008), making it a valuable and important measure of sea-ice dynamics.

Another alternative approach to studying sea-ice dynamics is through the trajectory-stretching exponents (TSEs; Haller et al., 2021, 2022) – single-buoy stretching diagnostics. For a trajectory, TSEs are quasi-objective Lagrangian metrics of material stretching (Aksamit et al., 2023). Specifically, they estimate the true material stretching in gradually varying flows, and recently have been extended to approximate deformation of the sea-ice cover (e.g. Aksamit et al., 2023). By resolving Lagrangian coherent structures, it allows for the identification of large sea-ice variation, for example during cyclonic events or springtime ice melt. Large values correspond to periods and areas of significant convergence, divergence and shear (Aksamit et al., 2023).

Synoptic events have a significant influence on the evolution of Antarctic sea ice. However, our current understanding of the interactions between cyclones and sea ice remains limited (Vichi et al., 2019). The majority of the existing literature on extra-tropical cyclones does not consider what happens when they cross the MIZ. Additionally, very little field data of metocean (meteorological and oceanographic) conditions are available in the Southern Ocean, and even less in the MIZ (Derkani et al., 2020). This has had negative feedbacks on the remote sensing and prediction model network of sea-ice concentration, which cannot depend on adequate ground truth to be validated with high confidence (Derkani et al., 2021; Lavergne and Down, 2022). Additionally, this has led to sea-ice dynamics in the Antarctic MIZ being poorly understood – specifically ice drift, deformation and type. Chapter 2 previously attempted to overcome these knowledge gaps by showcasing one of the longest ice-tethered buoy trajectories in the Antarctic MIZ, as it drifted for four months spanning winter and spring within the Indian Ocean sector, and under the influence of several synoptic cyclones. It was demonstrated that wind forcing had a dominant physical control on ice drift, with the persistence of free-drift conditions within regions of 80-100 % remotely sensed ice concentration and > 200 km from the ice edge. Moreover, the drift was characterised by a strong inertial signature at ≈ 13.5 hours, which appeared initiated by passing cyclones. This implied and further corroborated that the concentration-based definition (15-80 %) is inadequate to define the MIZ and its composition, since the highly dynamic nature of the MIZ is maintained despite the high sea-ice concentrations (> 80 %) observed from remote sensing products.

In this study, we extend the work of Chapter 2 to two arrays of ice-buoys deployed during austral winter and spring in the north-eastern Weddell Sea region of the Southern Ocean (Sect. 1.4), to provide information on sea-ice dynamics at the synoptic and shorter time scales over two seasons. We correlate the in-situ drift measurements with atmospheric reanalysis data to investigate the effects of extra-tropical cyclones on ice drift during each season, through the momentum transfer from winds and the generation of inertial oscillations of the sea ice. We also explore the differential drift between the buoys in each season to estimate the ice deformation rates and compare them with the few existing datasets from other regions, in search for general patterns of Antarctic sea-ice deformation.

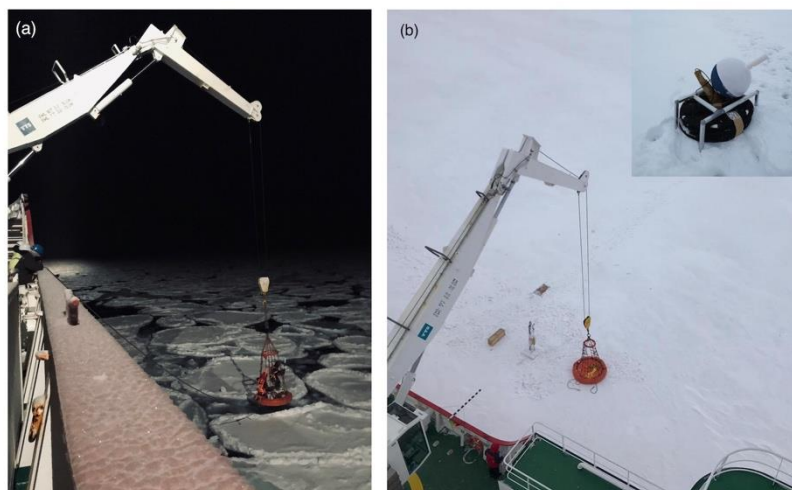


Figure 3.1. The sea-ice conditions and deployment of a buoy during **(a)** the 2019 Winter cruise, on a pancake ice floe, using the ship's crane, and **(b)** the 2019 Spring cruise on consolidated ice floes. The inset in **(b)** depicts the frames specifically designed around the standard Iridium Surface Velocity Profilers (ISVPs) for the 2019 Spring cruise. The diameter of the basket is 1.5 m. The inset is credited to Mardene de Villiers.

3.2. In-situ observations and environmental data

In 2019, winter and spring scientific research expeditions aboard the S.A. Agulhas II were conducted along the Good-Hope line (0° E; Sect. 1.4) to the Antarctic MIZ as part of the *Southern Ocean seAsonAL Experiment* (SCALE; Ryan-Keogh and Vichi, 2022). These expeditions generated a large set of comprehensive data from the physics to the biogeochemistry of the MIZ in the Atlantic sector. In this study, we focus on the analysis of seven GPS-tracked ice buoys deployed during these two expeditions (Table C1), which supplied the sub-daily and true daily information on Lagrangian ice drift and deformation that are needed to validate satellite ice drift products (Heil et al., 2001, 2009).

The Winter cruise departed from Cape Town on 18 July 2019 and reached the MIZ on 26 July, at 56.5° S and 0.1° E. Three polar Iridium Surface Velocity Profilers (MetOcean model; de Vos et al., 2022; de Vos et al., 2023), hereafter simply referred to as ISVP 1-3 were deployed during pancake ice conditions (1-5 m floes, and ≈ 40 cm thick; Fig. 3.1a), along the 0° meridian. Features of the sampling sites have been previously detailed in Skatulla et al. (2022). ISVP 1, was initially deployed in water, in between pancake floes, while ISVP 2 and ISVP 3 were deployed directly onto large pancake ice floes. These ISVPs were expendable devices that recorded position, air and ice temperature, and barometric pressure. Only their GPS positions are used in this analysis due to reliability issues of the meteorological data. Details of the deployments and lifetimes of the buoys are given in Table 3.1.

The Spring cruise departed from Cape Town on 11 October 2019 and was the first time that the Good Hope line was sampled during austral spring. The ship entered the MIZ on 20 October at 55.0° S and 0.0° E and the first standard Iridium Surface Velocity Profiler (Pacific Gyre model; de Vos et al., 2022, de Vos et al., 2023), denoted ISVP 4, was deployed on 24 October along the 0° meridian (for ease of reading, all the drifting buoys are named ISVP and numbered sequentially). During this cruise, specifically designed frames were built around these buoys to allow them to stand firmly on the ice floes without damaging the non-polar battery (inset in Fig. 3.1b), and to ensure that they would operate as Lagrangian-ice trackers. The other three buoys – ISVP 5, ISVP 6 and an ice-

tethered, non-floating Trident Sensors Helix Beacon (Sect. 2.2.1; Womack et al., 2023) – were later deployed more than 5° (> 400 km) east of ISVP 4. All four of these devices were deployed during first-year ice conditions (> 45 cm thick, Fig. 3.1b), and features of these sampling sites have been detailed in Johnson et al. (2023). The standard ISVPs recorded position, air temperature and barometric pressure, while the Trident recorded position and air temperature. Similar to the winter deployments, only the GPS positions are used in the analysis. Details of the buoys’ deployment and lifetime can be found in Table 3.1.

Table 3.1: Operational details of the Winter and Spring 2019 buoys. Herein, all dates and times are Greenwich Mean Time (GMT).

	Deployment date (2019)	Deployment position	Sampling frequency	Analysis end date (2019)	Number of analysed days	Total drift distance (km)
<u>Winter</u>						
ISVP 1	27 July at 01:00	57.05° S, 0.10° W	30 minutes	5 October at 12:00	70 (10*)	2642.85
ISVP 2	28 July at 05:07	57.17° S, 0.00° E	1 hour	25 August at 03:00	28	1036.97
ISVP 3	27 July at 17:09	57.92° S, 0.02° W	1 hour	5 October at 12:00	70	2552.28
<u>Spring</u>						
ISVP 4	24 October at 12:00	59.33° S, 0.06° E	1 hour	7 December at 12:00	44	1101.89
ISVP 5	28 October at 10:00	59.35° S, 6.57° E	1 hour	9 December at 12:00	42	992.44
ISVP 6	29 October at 12:00	59.37° S, 8.16° E	1 hour	7 December at 12:00	39	940.99
Trident	30 October at 12:00	59.47° S, 10.89° E	4 hours	2 December at 00:00	33	650.73

*ISVP 1 left the AMSR2-derived ice edge within 10 days.

The in-situ observations were integrated with environmental data retrieved from satellite and reanalysis products. Larger-scale meteorological conditions in the form of mean sea level pressure (mslp), 10 m wind velocity, and 2 m air temperature were obtained from ERA5 (Copernicus Climate Change Service (C3S), 2017), with an hourly time interval. Prior literature validated this reanalysis product in the Weddell Sea (King et al., 2022) and at $\approx 30^\circ$ E (Vichi et al., 2019). These variables were then bi-linearly interpolated in space to the buoys' locations, hourly for all ISVPs and four-hourly for the Trident, to ascertain the synoptic atmospheric forcing during both seasons. SIC data at 3.125 km spatial resolution were acquired from the passive microwave Advanced Microwave Scanning Radiometer 2 (AMSR2) sensor (Spren et al., 2008), and complemented by the 25-km spatial resolution Special Sensor Microwave Imager/Sounder (SSMIS) product (Cavalieri et al., 1996; NSIDC, 2023).

In addition to their low temporal (and spatial) resolution, the use of satellite products is known to be limited in their application to the broad Antarctic MIZ, where ice type is less related to the concentration value (Alberello et al., 2019; Vichi, 2022). The amount of in-situ data available is not yet sufficient enough to be used for regular validation (Aaboe et al., 2021). This has had drawback effects for prediction models, which are impaired by significant biases in the Southern Ocean (e.g. Yuan, 2004; Li et al., 2013; Zieger et al., 2015). Therefore, in this study we do not use SIC other than to estimate the ice edge, and hence, have rather inferred ice type from the buoys' drift features and wind response (see Sect. 3.5.1).

Herein, the 0 % SIC is used to define the sea-ice edge rather than the conventional 15 % SIC, even though this 0 % SIC region has been recognised as being heterogeneous and fragmented. This is because satellite-product algorithms tend to underestimate the SIC in thin ice as well as close to the ice edge (Pang et al., 2018), where ice melt and growth conditions make up a large component of the sea-ice regime (Agnew and Howell, 2003). Furthermore, in Sect. 2.2.2. it was reported that the results, of the buoy's distance from the ice edge, in the Antarctic were only marginally affected by this choice, with a maximum difference between the 0 % and 15 % SIC of less than 50 km. This is still within the range of differences between satellite products.

By construction, ISVPs continue to drift in the ocean after ice melting and can be further refrozen in between floes. Therefore, there is uncertainty on whether these buoys remained within the MIZ during their drift, especially for ISVP 1 as it was deployed in between ice floes. Using daily SIC data to estimate the ice edge, we determined the dates when the ISVPs left the ice cover. During winter, ISVP 1 first left the AMSR2 ice cover 10 days after its deployment (see Fig. B1 of the Supplement). ISVP 3, initially located the furthest south and thus further from the ice edge, on the other hand remained within the AMSR2 ice cover the longest (until 20 September 2019). ISVP 1 and ISVP 3, eventually both left the SSMIS ice edge on 5 October 2019. We subsequently ended our analysis on 5 October (Table 3.1), but caution was taken when analysing this dataset. ISVP 2 stopped transmitting on the 25 August, presumably due to battery failure. Similarly, the Spring ISVPs also would have been able to function as open-ocean drifters after the ice melted. Thus, we estimated that they left the ice cover between 7-9 December 2019 using the AMSR2 and SSMIS 0 % ice concentrations (see Fig. B2 in Appendix B). The non-floating Trident buoy sank on 2 December 2019 due to ice melting.

3.3. Methods

3.3.1. Extra-tropical cyclone identification

As previously reported in Sect. 2.2.2.1, polar cyclones in the Southern Ocean typically occur every five to seven days (Hoskins and Hodges, 2005; Vichi et al., 2019; and references therein). Additionally, due to the fact that the Winter and Spring buoys' trajectories are confined, the cyclones can also be tracked without the need of an automatic tracking algorithm. Following the procedure done and thresholds used in Sect. 2.2.2.1, a visual inspection method of the ERA5 mslp and 2 m air temperature fields at four-hourly intervals is applied to investigate the time window at which the cyclone cores were closest to the buoys' location. Eight cyclones for winter and seven for spring have been identified by low-pressure troughs with core pressures < 970 hPa and distances < 1000 km from the buoys, and by an increase in air temperature to near melting point on the eastern flank of the cyclones (Vichi et al., 2019). The dates when these cyclones were closest to the buoys have been computed using a nearest-neighbour method, and later associated with the ice drift and dispersion analyses.

3.3.2. Buoy kinematic parameters

The GPS position of all seven buoys was communicated via the Iridium system, with an accuracy of < 5 m. Since some data were irregular, missing or had duplicates, the position data were interpolated to a regular interval of one hour for all six ISVPs and four-hourly for the Trident. The higher temporal resolution of the ISVPs was kept since it provided more accurate data on ice drift. However, the following methods were also repeated for the Spring ISVPs using the four-hourly time interval and the difference was negligible. For each of the buoys, their latitude and longitude positions, like in Sect. 2.2.1 can be used to derive their downwind zonal (u) and meridional (v) components using the standard linear approximation:

$$u = \frac{\Delta x}{\Delta t}, \quad (3.1)$$

$$v = \frac{\Delta y}{\Delta t}, \quad (3.2)$$

where Δx and Δy are the zonal and meridional geodesic distances travelled between points along each buoys' trajectory, at time interval Δt . The speed is taken as the magnitude of the resultant of the velocity components.

For buoys, errors in drift measurements depend on the accuracy of position and time readings. Errors due to the timing of GPS position measurements are generally quite small (Dierking et al., 2020). Hutchings and Hibler (2008) reported that velocity errors are < 10 % for sampling intervals > 1 hour, and therefore it can be neglected (Dierking et al., 2020). The position error estimation can be attributed to the tracking error between two consecutive GPS positions, and the errors in the GPS reference points (Dierking et al., 2020), denoted by Lindsay and Stern (2003) as the geolocation error. The tracking error for buoys is however zero since the buoys remained fixed relative to the ice floe on which they were deployed. The geolocation error for these buoys is also taken to

be negligible as the buoys drifted with little latitudinal change, where we can assume identical geolocation errors, which would cancel out when calculating the drift velocity and speed.

A fast Fourier transform is applied to the buoys' velocity time series to derive the power spectral density as in Heil et al. (2009). As the buoy data were in the form of a discrete time series, a Hamming window in the time domain is used to minimize frequency leakage (Heil et al., 2009; Glover et al., 2011). A Morlet wavelet analysis is additionally computed to examine how the frequency domain changed over time (Liu and Miller, 1996; Torrence and Compo, 1998). To better detect inertial oscillations in the wavelet analysis, a high-pass Butterworth filter is also applied with a cut-off threshold of one day.

Both buoy arrays were deployed and drifted in the deep Southern Ocean away from the Antarctic continent. Therefore, as in Chapter 2, we again do not consider tidal forcing, which is known to be negligible in offshore locations (Heil et al., 2009; Lei et al., 2021; Alberello et al., 2020). Moreover, these buoys also operated near an amphidromic point (Lu et al., 2021; their Fig. 10), which is a geographical location of the ocean where the main tidal fluctuation (M2) is negligible (Martine and Dalrymple, 1994; Kamphius, 2000).

The meander coefficient (M) is computed to assess the effective translation associated with the buoys' drift. This ratio (Vihma et al., 1996; Heil et al., 2009; Heil et al., 2011), as discussed in Sect. 2.2.1, is defined as:

$$M = \frac{1}{\Delta D}. \quad (3.3)$$

This is first analysed as a time series for each time step, showing the cumulative change as the time window increased. It must be noted that M is a function of time over which it is computed, and on the sampling intervals transmitted by the buoys (Heil et al., 2009). For this reason, we also compute a daily discrete meander coefficient to highlight deviations at the synoptic scale.

3.3.3. Wind factor and ocean current's drift

To examine the relationship between ice drift and surface winds, we adopt the linear relation described by Thorndike and Colony (1982). This method relates the drifting buoy velocity (u_i , v_i) and ERA5 10-m wind velocity (U_{10} , V_{10}) as follows:

$$\begin{bmatrix} u_i \\ v_i \end{bmatrix} = F \begin{bmatrix} \cos\theta & \sin\theta \\ -\sin\theta & \cos\theta \end{bmatrix} \begin{bmatrix} U_{10} \\ V_{10} \end{bmatrix} + \begin{bmatrix} \bar{c}_u \\ \bar{c}_v \end{bmatrix}, \quad (3.4)$$

where F is the wind factor, θ is the turning angle and $\bar{c}_u$, $\bar{c}_v$ represent the mean (residual) ocean currents over the analysed period. The counter-clockwise rotation matrix is applied as the angle θ is negative (left deflection) in the southern hemisphere. In Eq. (3.4), the time variations of F , θ , and $\bar{c}_u$, $\bar{c}_v$ are not considered, and thus may vary with the time period over which they are computed, and also the sampling interval of the transmission of the buoys. These constants are calculated using the least squares regression technique described by Kimura and Wakatsuchi (2000) and Kimura (2004) and are fully detailed in Sect. 2.2.2.3 (Eq. 2.10 to 2.15).

As previously done in Sect. 2.2.2.3, Eq. (2.16 and 2.17), we also compute the vector correlation coefficient R_v^2 and the Pearson correlation coefficient R_p^2 (with 95 % confidence interval) to estimate the fraction of the variance of the ice-drift velocity that is explained by the wind velocity, and the linear relationship between the magnitude of ice drift and wind speed.

The mean ocean currents are derived by subtracting the ERA5 wind-driven motion from the in-situ ice motion – that is, the portion that is not linearly related to the variation of wind speed (Kimura, 2004). As such, they carry multiple uncertainties that are not directly quantifiable. To the best of the authors’ knowledge, there are no robust methods to determine under-ice geostrophic currents from remote sensing platforms, and to extract proper statistics. Furthermore, there are currently no in-situ measurements for currents within the Antarctic MIZ region (e.g. Alberello et al., 2020). The only in-situ observations of underlying ocean currents in the Antarctic were reported by Geiger et al. (1998). These observations were measured along the coast in the western Weddell Sea, where the Antarctic Coastal Current is located, and where land-fast and pack-ice conditions dominate. These ice types do not have an equivalent spatial morphology to the MIZ ice conditions (Vichi, 2022), and the coastal current incurs significantly different driving mechanisms.

3.3.3.1. Comparison of residual currents and the Copernicus GlobCurrent database

While there are no in-situ under-ice currents available to validate the residual currents, there is a model product that can be used to compare the estimate of the winter residual currents, and check if they are comparable with the velocities observed in the ACC outside of the sea ice at this longitude. Herein, we use data of current velocity from Copernicus GlobCurrent database in the region of the Winter buoys’ trajectories. This dataset integrates the velocity field of geostrophic surface currents from satellite products recorded from 1993 to 2019 (Rio et al., 2014) and modelled Ekman currents, which contain components from wind stress forcing provided from atmospheric and drifter data (Derkani et al, 2021).

Figure 3.2 (left-hand-side) shows the mean current speed for the period of the Winter buoys’ deployment (27 July 2019 to 5 October 2019). The GlobCurrent database does not have any data for the latitudinal range of the Spring buoys, since they were deployed and drifted further from the sea-ice edge. We reiterate that this product is a model, and as reported in Section 3.2, there is still significant biases for models in the Southern Ocean. Therefore, the GlobCurrent database does have uncertainties (Derkani et al., 2021), including the fact that geostrophic currents cannot be computed from altimetry under the sea-ice. It is more likely that the velocities are biased towards the open ocean components far from the ice edge. The mean zonal and meridional velocities, for the duration and between -55° S to -56° S (the mean of the buoys’ latitudes) and 0-26° E (as indicated by the cyan rectangular box), were extracted to compute the distribution of the velocity components (right-hand-side). The magnitude of the winter residual currents is additionally shown, where it can be seen that they fall within the range of the GlobCurrent distribution, although in the lower tail. This is compatible with the under-ice momentum reduction of ocean currents. The aim was to however demonstrate that the latitude of the 2019 winter deployment is more likely to find more intense ocean currents than at the latitudes of the 2017 winter and 2019 spring deployments. This indicates that currents can be expected to be even higher during the 2019 winter deployment, which supports our hypothesis.

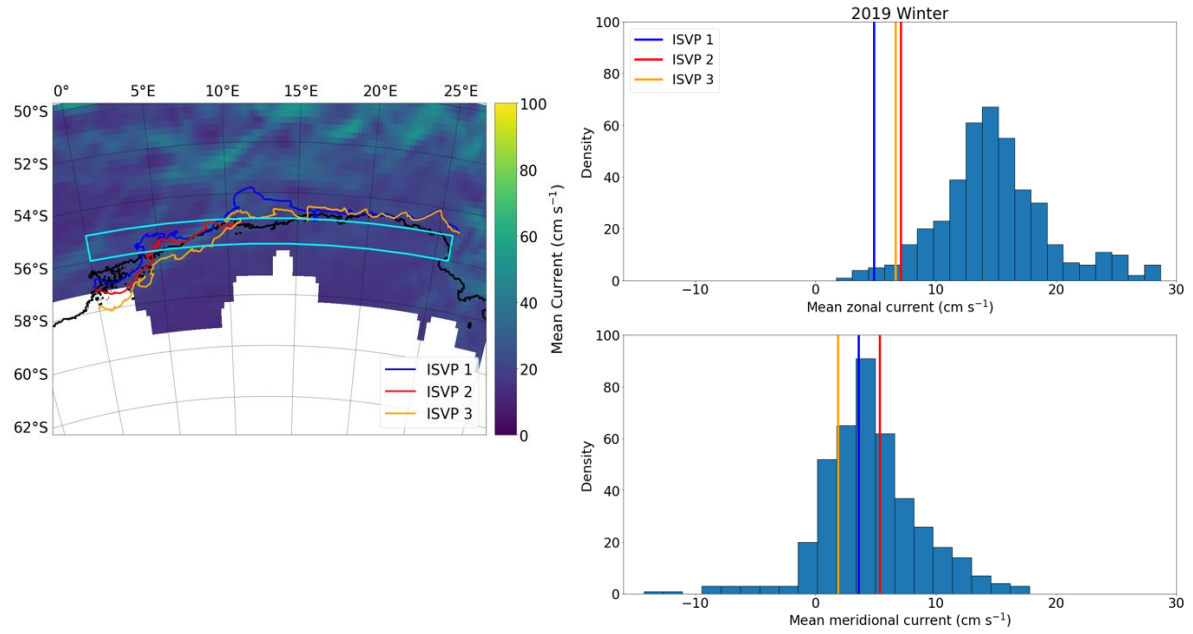


Figure 3.2. (left-hand side) The mean Copernicus GlobCurrent speed for the deployment period of the Winter buoys (27 July 2019 to 5 October 2019). The trajectories of the three Winter buoys are shown. The black contour denotes the AMSR2 0 % sea-ice concentration on 30 September 2019 – the date of approximate austral sea-ice maximum. The cyan rectangular box indicates the region used to compute the distribution of the zonal and meridional currents. (right-hand side) The distribution of the spatial mean of the GlobCurrent zonal and meridional currents. The vertical lines denote the magnitudes of the winter residual currents, which were computed using Eq. 3.4.

3.3.4. Lagrangian measures for dispersion and deformation assessment

The absolute dispersion is used to characterise ice motion, and for an ice buoy in an ensemble of buoys, it is defined as (Taylor, 1922; Lukovich et al., 2017, 2021):

$$AD^2 = \langle |x_i(t) - x_i(0) - \langle x_i(t) - x_i(0) \rangle|^2 \rangle, \quad (3.5)$$

where x_i is the zonal or meridional position of the i -th particle in the ensemble, as a function of the elapsed time t . We also applied the method by Lukovich et al. (2017) to calculate the relative dispersion of any two-buoy arrays:

$$RD^2 = \langle |x_i(t) - x_{i+1}(t) - \langle x_i(t) - x_{i+1}(t) \rangle|^2 \rangle, \quad (3.6)$$

which is defined for adjacent particle pairs x_i and x_{i+1} in the zonal or meridional direction. For both Eq. (3.5 and 3.6), the angular brackets denote the ensemble mean over the number of buoys (pairs) in the array. The totals are computed as the sum of the zonal and meridional components, for the time period when all buoys, in each season, transmitted together.

In this study, Green's theorem could not be applied to compute the strain rates i.e. three-particle dispersion (Sect. 1.1.5.2; Eq. 1.4-1.7). Both the Winter and Spring buoys were deployed in a quasi-linear buoy array geometry, and thus all triangles formed by the buoy positions had small angles ($< 15^\circ$). This would have given unreliable calculations of the strain rates (Itkin et al., 2017), and the reduction in accuracy from a large array to only two buoys is unknown (Alberello et al., 2020). In lieu of this, we further analysed the dispersion of sea ice using the methods proposed by Rampal et al. (2008), which defines a proxy of the full strain rate tensor ($\dot{\epsilon}_{tot}$). They considered two buoys, namely 1 and 2 with absolute positions $\vec{X}_1$ and $\vec{X}_2$ respectively, and with a separation $\vec{Y} = \vec{X}_2 - \vec{X}_1$. If these two buoys, initially separated by $L0 = ||\vec{Y}(0)||$, are observed after a time $t = \tau$, their separation changes to $l(\tau) = ||\vec{Y}(\tau)||$. The change in separation is then defined as:

$$\Delta r = ||\vec{Y}(\tau)|| - ||\vec{Y}(0)|| = l(\tau) - L0. \quad (3.7a)$$

Δr is then computed as a function of τ :

$$\Delta r = l(t + \tau) - l(t). \quad (3.7b)$$

In fluid mechanics, the dispersion process is characterised by the mean square change in separation $\langle \Delta r^2 \rangle$, while from a solid mechanics' perspective, there is a consensus that it is more relevant to express the dispersion in terms of a deformation rate ($\dot{D} = \frac{\Delta r}{\tau L0}$), using the standard deviation (Girard et al., 2009; Rampal et al., 2008, 2009; Weiss, 2013):

$$\sigma_D = \left\langle \left(\frac{\Delta r}{\tau_{L0}} - \left\langle \frac{\Delta r}{\tau_{L0}} \right\rangle \right)^2 \right\rangle^{1/2}, \quad (3.8)$$

where angular brackets again denote the ensemble mean, computed over the number of buoy pairs in the cluster, for the time period when all buoys, in each season, transmitted together. Rampal et al. (2008) demonstrated in the Arctic that σ_D is proportional to $\dot{\epsilon}_{tot}$. We remind that the σ_D and $\langle \Delta r^2 \rangle$ diagnostics are not sensitive to solid rotations. They only quantify the overall deformation due to divergence, convergence and/or shear, but do not allow to distinguish between them.

As an additional proxy for sea-ice deformation, we applied the methods derived by Aksamit et al. (2023), and computed the TSEs for individual buoys in each season as:

$$\text{TSE}_{t_0}^{t_N}(\mathbf{X}_0) = \frac{1}{t_N - t_0} \log \frac{|\dot{\mathbf{X}}(t_N)|}{|\dot{\mathbf{X}}(t_0)|}, \quad (3.9)$$

where $\mathbf{X}_0$ is the initial position, $\dot{\mathbf{X}}$ is the derivative with respect to time (i.e. drift speed), and TSE is the measure of trajectory stretching or contraction (Haller et al., 2021) for discrete data (Aksamit et al., 2023). Similar to that of Lagrangian dispersion statistics, TSEs do not differentiate between contributions from divergence, convergence, or shear to the stretching of the sea ice. Instead, they note the stretching in the direction of vectors, tangent to the trajectory. TSEs are computed forward-in-time from the starting time t_0 and for each time step along the buoy's trajectory. The TSEs in each season were computed over a one-day integration period as to capture the daily-scale Lagrangian stretching, and to relate them to the drift kinematics and meander coefficient.

3.4. Results

In this section, all results from the multiple diagnostics in Sect. 3.3 are presented and explained. The diagnostics are considered complementary a priori, and they are comparatively discussed, analysed, and correlated in Sect. 3.5.

3.4.1. Spatial and temporal changes in atmospheric conditions

The time series of the ERA5 mslp and 2-m air temperature in the vicinity of the Winter and Spring buoys are presented in Fig. 3.3. We only present data from ISVP 1 since all three Winter buoys remained close together (Fig. 3.4a) and their meteorological conditions were very similar. Subsequently, ISVP 6 was also used as a reference for the eastwards cluster of the Spring buoys (Fig. 3.4b), and ISVP 4, that drifted $> 5^\circ$ (> 400 km) west from the rest of the Spring buoys and under slightly different meteorological conditions, has been included as well. The time series for all buoys can be found in Appendix B (Figs. B3 and B4).

During the passage of each cyclone, highlighted by the stars, there was a characteristic drop in the mslp, in both winter and spring. However, there was a significant difference in the time series of the air temperatures between the two seasons. During the winter deployment, the air temperatures initially fluctuated between -15°C and 0°C

as warm air was advected poleward on the eastern flank of cyclones, while the cold polar air was advected equatorward on the western flank (Schlosser et al., 2018; Vichi et al., 2019). However, as the Winter buoys drifted into September and October, this fluctuation became smaller, along with the gradual increase in the background atmospheric temperature. The air temperatures overlying the region of the Spring buoys exhibited a similar fluctuation pattern compared to the end of the wintertime series for the first five days, as shown by ISVP 4 (Fig. 3.4b). After the second cyclone (31 October), the air temperatures fluctuated daily between -5°C and 0°C . Although temperatures increased slightly during the passage of the cyclones, the daily signal appeared to be dominant in contrast to the winter conditions.

All the analysed cyclones (minimum core pressures < 970 hPa) carried substantial energy in their winds, with speeds > 16 m s^{-1} in winter and > 10 m s^{-1} in spring. Cyclones are also prominent transporters of moisture and heat to the polar latitudes (Messori et al., 2017), and generate large waves in the Southern Ocean that can propagate hundreds of kilometres into the ice cover (Kohout et al., 2014). The impacts of these cyclones on the ice cover were however different during each season, as detailed in the succeeding sections.

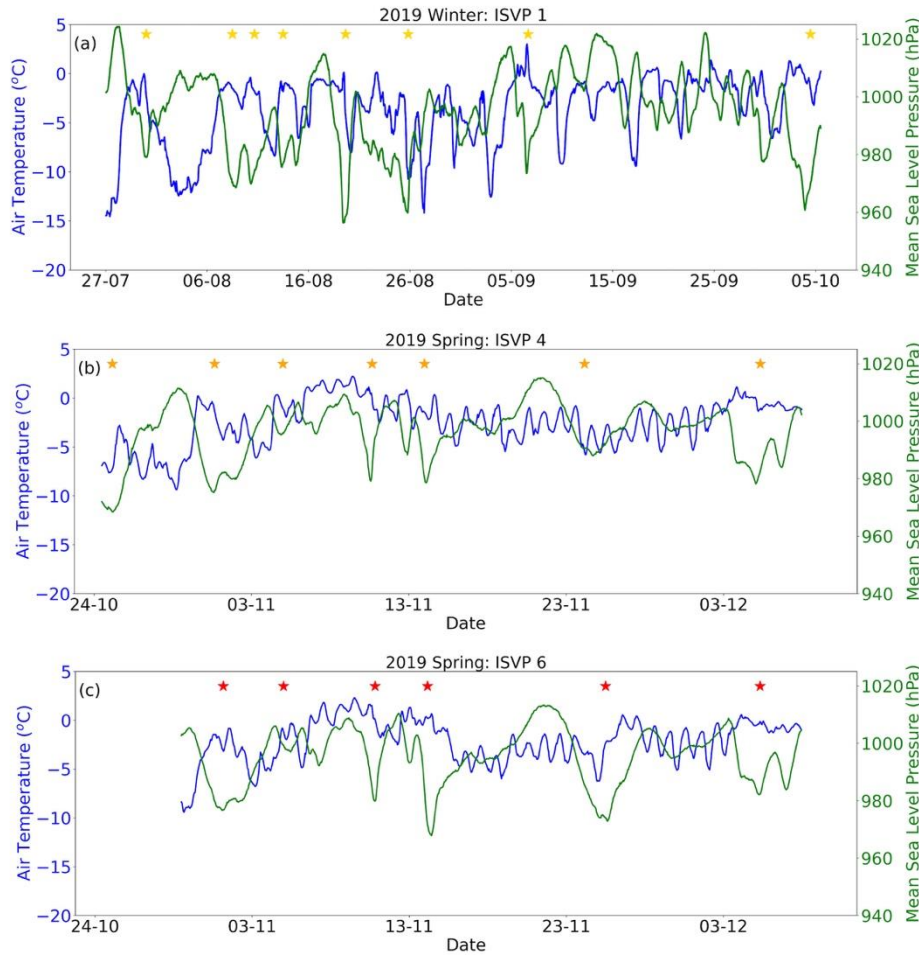


Figure 3.3. Time series of the ERA5 2 m air temperature (blue) and mean sea level pressure (green) at the location of ISVP 1 in winter (a), IVSP 4 (b) and ISVP 6 (c) in spring. The yellow stars denote when the winter cyclones were closest to the buoys. The red stars denote when the spring cyclones were closest to ISVP 4. The orange stars denote when the same spring cyclones were closest to the main spring cluster deployed $> 5^{\circ}$ (> 400 km) east of ISVP 4.

3.4.2. General drift and meandering

All seven buoys were initially deployed in the north-eastern Weddell Gyre region (Fig. 3.4). The Winter buoys experienced a significant eastwards transport, with ISVP 1 and ISVP 3 travelling over 25° (> 1500 km) eastwards, with a smaller latitudinal variation of $\approx 4^\circ$ (≈ 440 km). ISVP 2 only travelled to $\approx 10^\circ$ E (≈ 640 km) since it stopped transmitting on 25 August 2019 (red triangle in Fig. 3.4a). These three trajectories were characterised by large sharp turns and meanders in response to the eight cyclones. The Spring buoys were deployed in the same region as the Winter buoys. However, the prevailing trajectory patterns of the Spring buoys were vastly different. The Spring buoys only drifted 3° - 7° eastwards (≈ 190 - 400 km; Fig. 3.4b), but their meridional drift was significant and contributed to almost half of their total drift, as they travelled between 1° - 2° (≈ 100 - 200 km) almost directly northwards. Their trajectories also exhibited sharp turns and meanders in response to the seven spring cyclones. Additionally, during the prominent northwards drift, the Spring buoys were characterised by small cyclic loops, which are indicative of inertial oscillations (see Sect. 3.4.4).

The overall drift pattern of the buoys can be further described through the buoys' meander coefficients (Eq. 3.3). The final cumulative meander coefficients were < 1.5 for all the Winter buoys, indicating a predominantly straight trajectory. The low meander coefficient indicates that the drift of buoys was influenced by the ACC (see Sect. 2.3.3). Vihma et al. (1996) also reported a meander coefficient value of 1.4 in the region of the ACC. Conversely, the cumulative meander coefficients for the Spring buoys ranged between 2.5 and 3, signifying a more oscillatory trajectory.

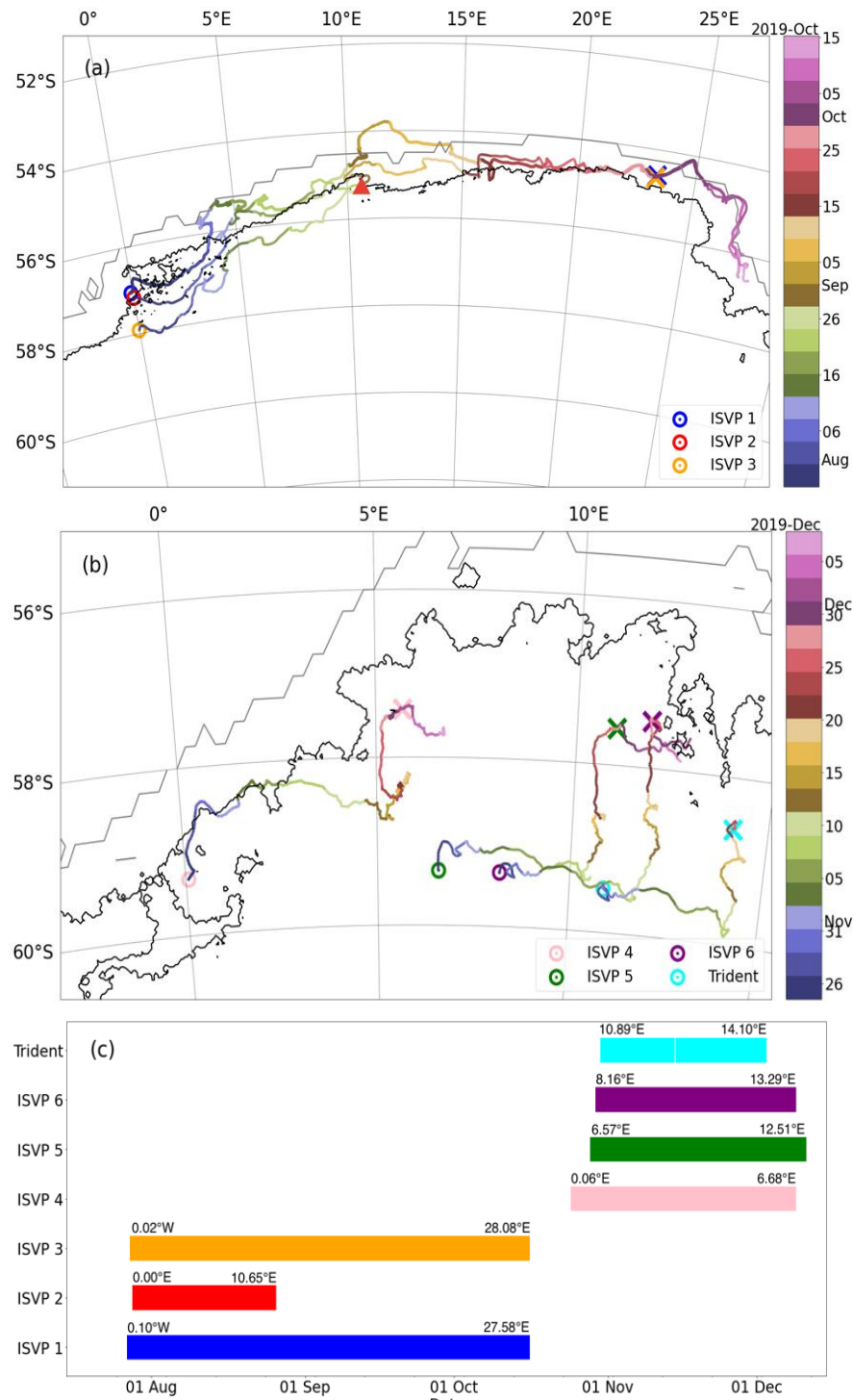


Figure 3.4. (a) Trajectories of the 2019 Winter buoys. The black (grey) contour denotes the AMSR2 (SSMIS) 0 % sea-ice concentration on 30 September 2019 – the date of approximate austral sea-ice maximum. The colour circles denote the start position of each buoy. The corresponding colour crosses of ISVP 1 and ISVP 3 denote their positions on 30 September 2019. The termination point of ISVP 2 (25 August 2019) is denoted by the red triangle. (b) Same as (a) but for the 2019 Spring buoys. The colour crosses and concentration contours are for 2 December 2019 – the day when the Trident buoy stopped transmitting data. (c) The operational periods of all buoys in this study, with corresponding colours to the time gradient maps. The deployment and final longitudes for each buoy are included above all operational periods.

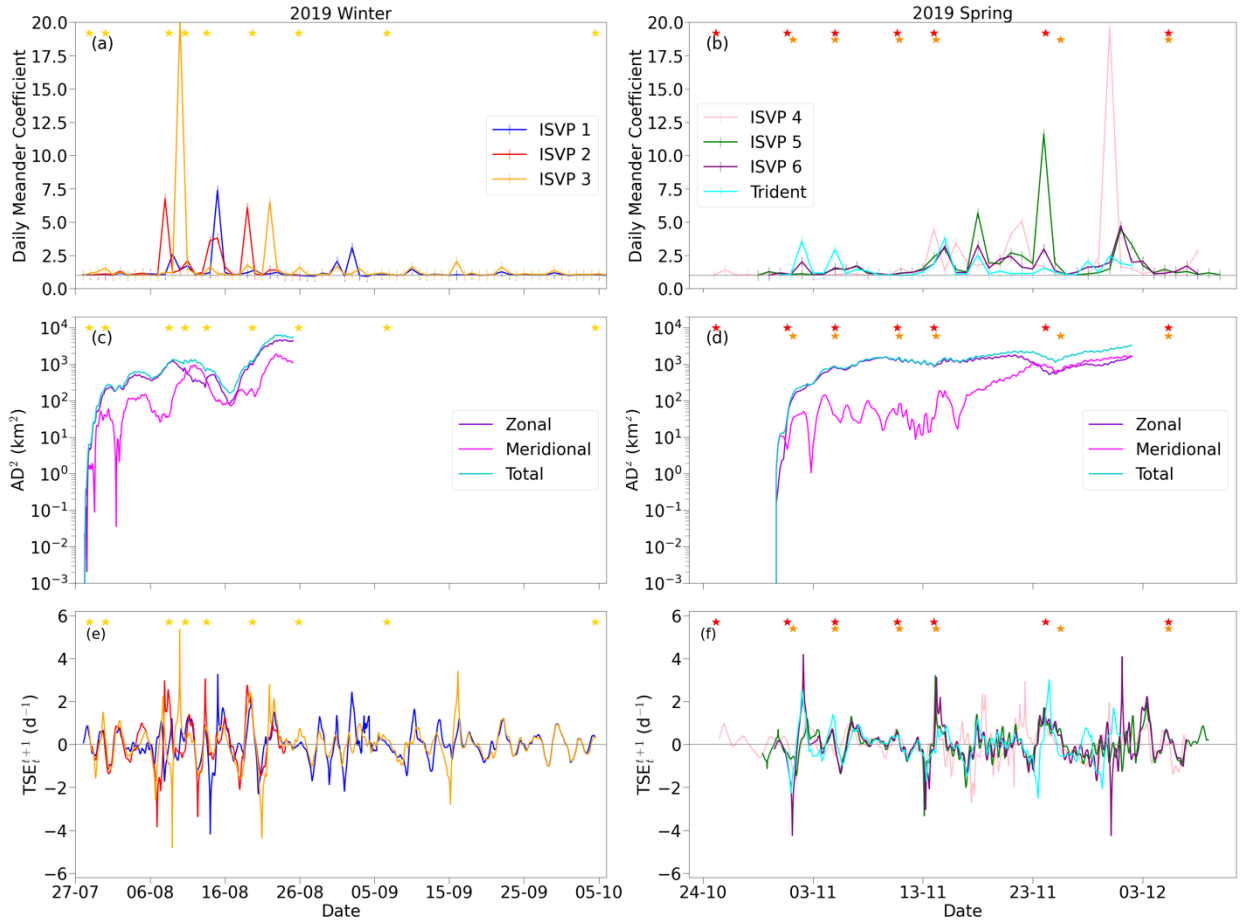


Figure 3.5. (a-b) Time series of the daily meander coefficient for the duration of the buoys' deployment, where the vertical markers are indicative of the daily time interval and the horizontal grey line at 1 denotes straight-line drift. (c-d) Time series of the absolute dispersion for the period when all buoys were transmitting together (≈ 30 days for both seasons). (e-f) Time series of the TSEs for the duration of the buoys' deployment. The Winter buoys are shown on the left and the Spring buoys on the right. The star symbols are the same as in Fig. 3.3.

The time series of the daily meander coefficient varied between the two seasons (Fig. 3.5a and b). The highest values in winter occurred as a result of the second to the fifth cyclones (8-19 August), but rapidly decreased afterwards to fluctuate between 1-2. This reduction and the mostly straight-line drift of ISVP 1 and ISVP 3 can be seen in Fig. 3.4a, where from late August, their trajectories became predominantly eastwards with only small turns and deflections. However, ISVP 1 did exhibit increased meandering between 26 August and 2 September, which correlates with dates when ISVP 1 drifted in the open water outside of the AMSR2 estimated sea-ice edge (Fig. B1 in Appendix B). The largest meander coefficient values in spring occurred during two periods when inertial oscillations were present (see Fig. 3.4b). There were, however, two earlier increases in the meander coefficient particularly for the eastward cluster of the Spring buoys, during the second and third cyclones (31 October and 5 November, respectively). Since the meander coefficient is also affected by the sampling time step, the Spring ISVPs were therefore recomputed using the Trident's four-hourly time interval (Fig. B5a in Appendix B). The magnitude and timing of the peaks remained the same, indicating that the relative motion in connection with the passage of the cyclones is realistic.

The evolution of the buoy trajectories is also described by the absolute dispersion (Fig. 3.5c-d) and the relative dispersion (Fig. B6 of Appendix B) – from the time when all buoys in each season began transmitting together. For this reason, they both have been computed for ≈ 30 days using Eq. (3.5 and 3.6). During winter, the total absolute dispersion (i.e. zonal and meridional components combined in quadrature) was predominantly influenced by the zonal displacement (Fig. 3.5c), while its corresponding relative dispersion was largely governed by the meridional separation (Fig. B6a). This was a result of these buoys being deployed meridionally along 0° E (Fig. 3.4a) and drifting coherently eastwards with the winds and the ACC. However, both methods indicated substantial fluctuations of all components in response to the passage of cyclones, indicating a highly deformable ice cover. During spring, the total absolute dispersion and relative dispersion were both predominantly governed by their zonal component (Figs. 3.5d and B6b, respectively). However, when the Spring buoys switched to drift northwards (Fig. 3.4b), after the fourth cyclone (10 November), this led to the eventual dominance of the meridional displacement. Concurrently, the meridional separation exhibited greater changes in response to these cyclones. However, its magnitude was smaller and thus its contribution to the total was minor in comparison to the predominantly constant zonal separation. This indicates that the Spring buoys moved more coherently and as an aggregate despite the initial larger deployment distance. It is also noteworthy that there are minor fluctuations in all of the dispersion components during the spring season, which correspond to the dates of the cyclic loops and meanders in the buoys' trajectories (Figs. 3.4b and 3.5b). Overall, we however find the absolute dispersion is less sensitive to the spatial orientation of the deployments, and thus more useful to characterise the dominant dispersion direction.

During both seasons, increased local TSEs occurred with the passage of the cyclones, often before and/or after the cyclone core was closest to the buoys' locations (Fig. 3.5e and f). This is because cyclones push the ice edge southwards, as warm air is advected poleward on their eastern flank. This compression of the ice cover is then followed by the relaxation and northwards movement of the ice edge when southerly winds, on the western flank, prevail (Vichi et al., 2019). Although being conceptually different, the TSE and the meander coefficient show the same evolution. TSEs allow to further distinguish the compression and the stretching, with the latter always observed after the former, and the stretching events are generally linked to the increase of the meander coefficient. In winter, ISVP 1 and ISVP 3 indicated increased stretching outside of cyclonic activity (after 26 August), which is not visible neither in the meandering coefficient nor in the dispersion. Sea ice is still deformable in this period, likely due to oceanic influences (Aksamit et al., 2023) or changes in wind direction (Itkin et al., 2017) that are not directly related to cyclones. Also during spring, the buoys indicated increased and more erratic TSEs during periods of no cyclonic events (13-24 November and 2 December). These however, like their corresponding meander coefficients, coincided with the two periods of inertial oscillations (attended to in Sect. 3.4.4). We note that, like the meander coefficient, TSEs are influenced by the sampling frequency. Therefore, the Spring ISVPs were recomputed using the Trident's four-hourly time step (Fig. B5b in Appendix B). Their magnitudes varied slightly but their time series continued to indicate the same phenomena in connection with the passage of the cyclones.

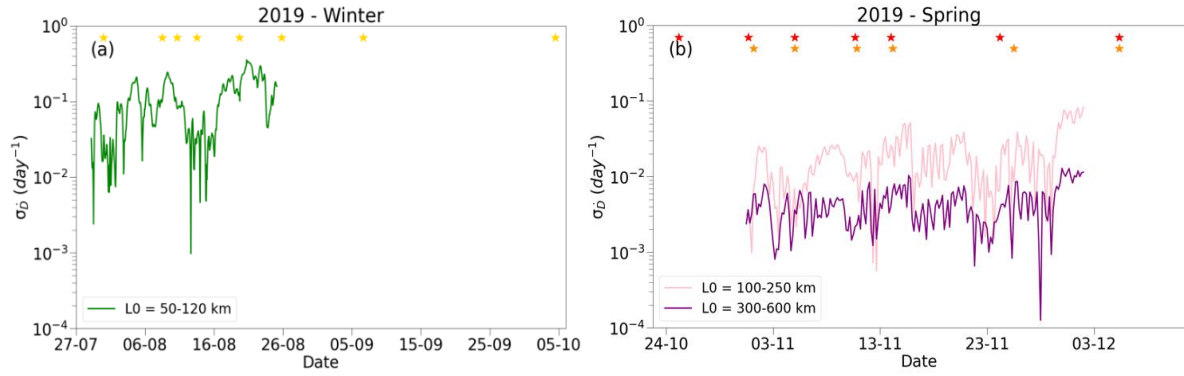


Figure 3.6. (a) Time series of the standard deviation σ_D of the deformation rate $\dot{D}$ for the Winter buoys (computed from the dispersion of buoy pairs for ≈ 30 days for both seasons). (b) Same as (a) but for the Spring buoys. L0 denotes the initial length scale of each group of buoys. The stars are the same as in Fig. 3.3.

3.4.3. Deformation rate estimates

The time series of the deformation rate σ_D for both seasons is shown in Fig. 3.6a and b. As ISVP 4 was deployed $\approx 5^\circ$ west of the other Spring buoys, we separated the analysis into two smaller clusters based off their initial length scales (L0 = 100-250 km and 300-600 km), as shown in Fig. 3.6b. This allowed for a more comprehensive analysis of sea-ice deformation at different spatial scales, and not only between the two seasons.

The σ_D measured by the Winter buoys was not only larger than during spring, but it also exhibited greater fluctuations in relation to the passage of cyclones (Fig. 3.6a). The most notable decrease of the σ_D occurred during the passage of the second to the fourth cyclones (8-13 August) that came in close succession. As this proxy only quantifies the magnitude of the total deformation rate, we cannot discern whether this was predominantly due to divergence, convergence and/or shear. This did however occur in relation to the decrease in both the absolute (Fig. 3.5c) and relative dispersion (Fig. B6a), along with the large peaks in the meander coefficient (Fig. 3.5a), which is indicative of the compression of the ice cover by the cyclones. Overall, these large fluctuations of the σ_D indicate that the winter ice cover was deformable in relation to the changing winds at the scale of 50-120 km, allowing for opening of leads and rafting of floes.

Although the σ_D sampled by both spring spatial scales varied due to the passage of cyclones, it also exhibited regular and relatively uniform fluctuations throughout the analysed period that were not associated with the cyclones (Fig. 3.5b). Therefore, the deformation of the spring ice cover was less correlated to the cyclones and their changing winds, relative to during winter. Additionally, the σ_D from the larger cluster exhibited a more “flattened” trend in comparison to the smaller cluster. Thus, its time series exhibited a smaller σ_D . This indicates that there was a smaller magnitude of deformation at these larger length scales.

3.4.4. Metocean drivers of sea-ice drift

Figure 3.7 shows the zonal and meridional velocity components of ISVP 1 (a-b) and ISVP 3 (c-d), during winter, and ISVP 4 (e-f), during spring, compared to the co-located ERA5 wind-velocity components. The time series of all Winter and Spring buoys can be found in Appendix B (Figs. B7 and B8, respectively). The overall drift velocity of the Winter buoys showed a good correlation to the wind (Table 3.2), suggesting low to absent internal stresses in the ice cover, which is typical of free-drift conditions. ISVP 1 showed large peaks ($> 1 \text{ m s}^{-1}$) on 11 August and between 26 August and 5 September. These periods correlate with the dates when ISVP 1 drifted outside of the AMSR2 estimated sea-ice edge (Fig. B1 in Appendix B).

As a reference for the Spring buoys, ISVP 4 initially drifted similarly to the wind velocity (Fig. 3.7e and f), although with a dampened signal because these buoys were deployed on consolidated sea-ice conditions. After the second spring cyclone (31 October), when air temperatures increased (Fig. 3.3b), ISVP 4's velocity fluctuations began to amplify. However, after the fourth cyclone (10 November) these fluctuations appear to be less correlated with the wind vectors and indicate a semi-diurnal signal.

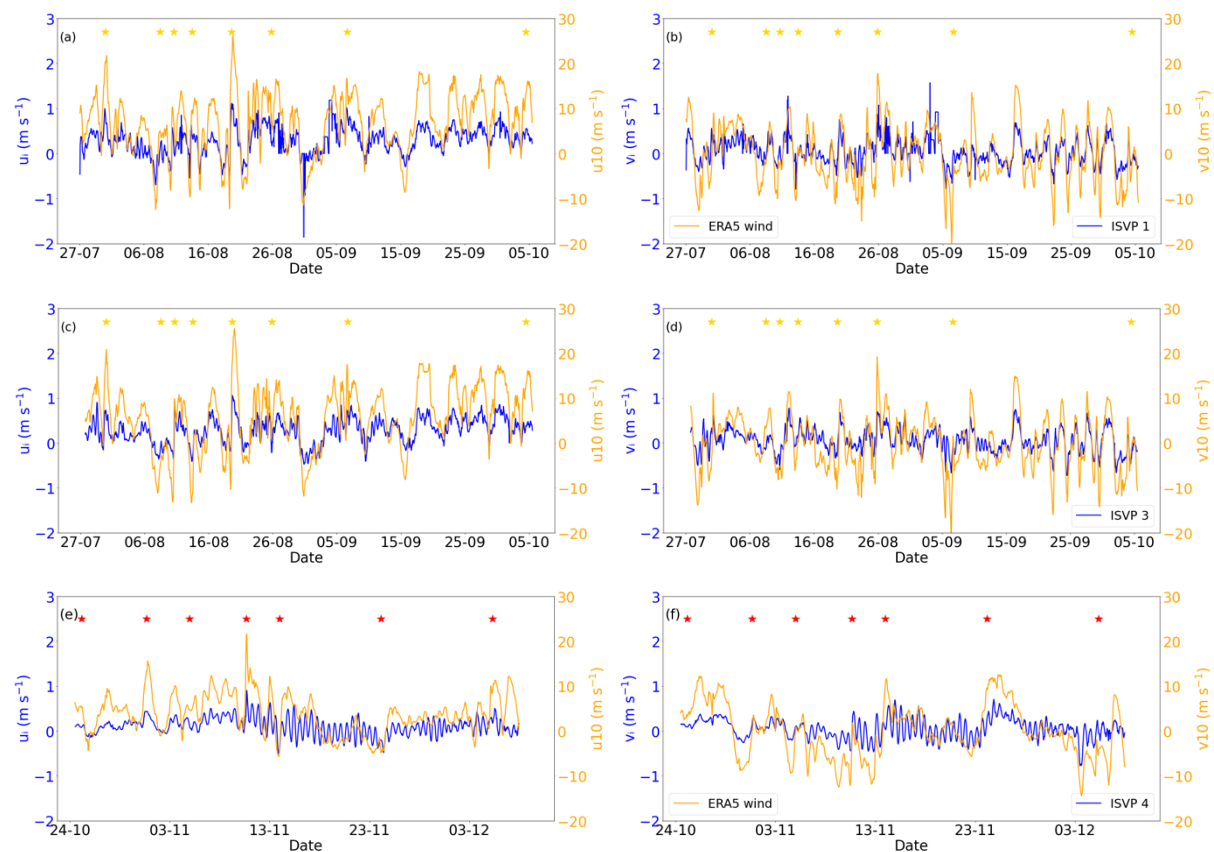


Figure 3.7. Velocity components of the buoys (on the left axis) and 10 m wind from the ERA5 reanalysis (on the right axis): the zonal and meridional component of ISVP 1 in winter (a-b), of ISVP 3 in winter (c-d), and of ISVP 4 in spring (e-f). The star symbols are the same as in Fig. 3.3. Note that ERA5 wind magnitude is similar for the ISVP 1 and ISVP 3 locations.

The observed buoy speed and direction have been also compared with the ERA5 wind vectors to quantify the physical control of atmospheric forcing on sea-ice drift (Sect. 3.3.3). In order to remove the period of oceanic drift, ISVP 1 was only analysed for the first 10 days, when it plausibly drifted within both the SSMIS and AMSR2 ice edges. The main results of this analysis are summarised in Table 3.2. The Winter buoys exhibited high wind factors ranging between 3.16 % and 3.78 %, with small turning angles ranging between -7.89° and -11.19°. The Spring buoys exhibited lower wind factors ranging between 2.42 % and 3.05 % and larger turning angles, between -21.18° to -27.00°. This indicates that the drift of the Spring buoys had a lower response to wind forcing but still in the range of the Antarctic values (Sect. 3.1) and high compared to the Arctic.

The relationship between the buoys' drift and the wind vectors can be quantified by both the vector R_v^2 and Pearson R_p^2 correlations (see Table 3.2). All p-values computed for the Pearson correlation were less than 0.05, thus indicating a statistically significant correlation, and a linear relationship between the buoy and wind velocities.

Table 3.2: Parameters describing wind and ocean forcing on sea-ice drift.

Buoy	Mean	Mean	Vector	Pearson	Mean current		Mean current
	wind factor (%)	turning angle (°)	correlation R_v^2	correlation R_p^2	velocity (cm s ⁻¹)		speed (cm s ⁻¹)
					$\overline{c_u}$	$\overline{c_v}$	
<u>Winter</u>							
ISVP 1 (first 10 days only)	3.16	-7.89	0.53	0.50	4.68	3.57	5.88
ISVP 2	3.77	-7.47	0.64	0.56	7.06	5.33	9.83
ISVP 3	3.62	-11.19	0.74	0.59	6.62	1.84	6.88
<u>Spring</u>							
ISVP 4	3.03	-21.18	0.54	0.27	-0.73	0.73	1.04
ISVP 5	2.89	-23.59	0.56	0.35	-1.33	-1.07	1.71
ISVP 6	3.05	-21.87	0.54	0.37	-3.12	-0.01	3.12
Trident*	2.42	-27.00	0.44	0.37	0.13	-2.11	2.11

*Four-hourly time interval

The mean residual current velocity components ($\overline{c_u}$, $\overline{c_v}$) are estimated as described in Sect. 3.3.3 (see Table 3.2). The mean current speed for the Winter buoys was larger than the Spring buoys, due to the Winter buoys drifting within the region of the ACC. Moreover, as the mean velocity components were positive in both the zonal and meridional directions, the underlying currents most likely also enhanced the wind-driven drift of the ice. On the other hand, the mean current velocity components of the Spring buoys varied greatly, and in most circumstances, they opposed the direction of wind-driven ice drift. This can be indicative of a region characterised by the presence of oceanic eddies. We note that while these residual currents are a rough estimate of the under-ice currents (see Sect. 3.3.3.1), they do suggest that the ACC had a significant impact on the drift of the Winter buoys in comparison to the Spring buoys drifting further from the ice edge.

3.4.5. Winter-spring differences in the frequency domain

Figure 3.8 shows the spectra of the ERA5 wind and ice drift velocities of ISVP 1 (a), ISVP 3 (b), and ISVP 4 (c). The other buoys in each corresponding season indicated similar results to the ones displayed here and can be found in Figs. B9 and B10. The wind velocity spectra for both seasons form a typical, continuous energy cascade from the lower frequencies to the higher frequencies. In contrast with winter data from Womack et al. (2022), the drift of the Winter buoys revealed no apparent inertial oscillations (Fig. 3.8a and b), although a weak and statistically non-significant peak at 15.05 hours is detected. However, this is not large enough to make a comprehensive analysis of a possible shift to the inertial range. Also noteworthy is the “flattening” of ISVP 1’s energy cascade in the high-frequency range (with a period smaller than 9 hours).

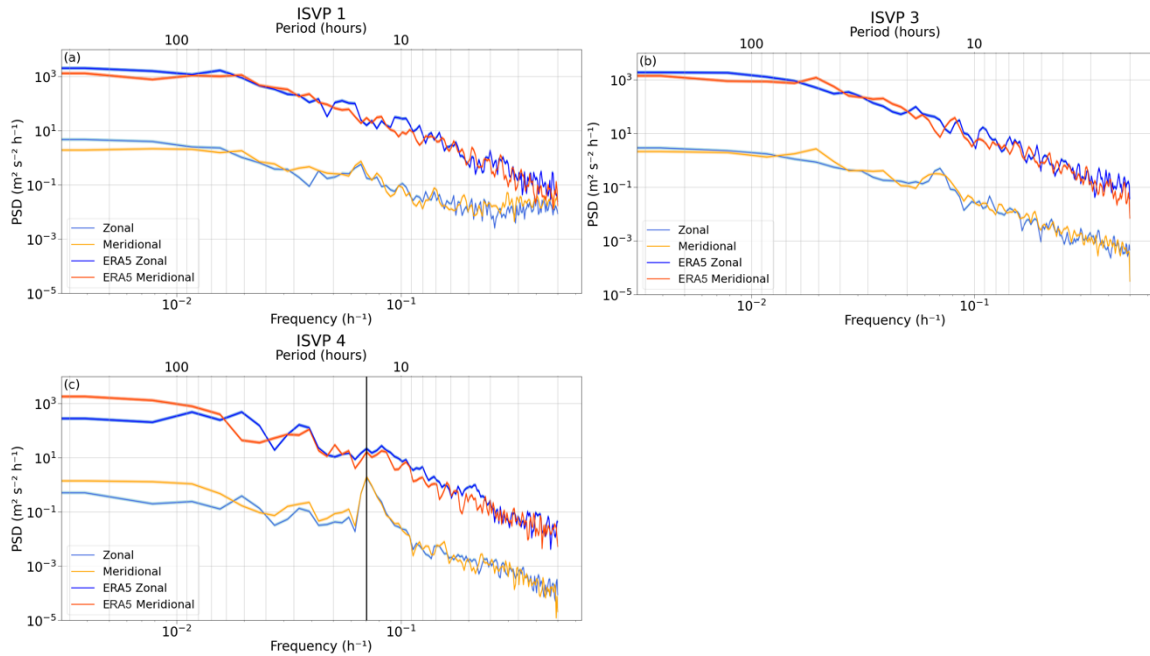


Figure 3.8. (a) Power spectral density corresponding to the zonal and meridional components of ISVP 1, in winter, and ERA5 wind velocity. (b) Same as in (a) but for ISVP 3, in winter. (c) Same as in (a) but for ISVP 4, in spring, where the black vertical line indicates the peak associated with inertial oscillations at 14.22 hours.

ISVP 4 and the rest of the Spring buoys exhibited an energy peak at the inertial frequency (highlighted by the vertical black line in Fig. 3.8c). The period of these oscillations is 14.22 hours for the ISVPs and slightly lower at 13.93 hours for the Trident, as it was deployed further south. In comparison, Heil et al. (2009) reported a period of 13.19 hours for the East Antarctic at $\approx 65^\circ$ S. However, all Spring buoys indicated a similar inertial response within the theoretical inertial range, determined by the Coriolis parameter, at 57° S- 60° S (14.30-13.85 hours). This can be clearly seen by the cyclic loops during their northward drift (Fig. 3.4b).

Figure 3.9 shows both the wavelet power spectrum (left panel) and the wavelet spectrum (right panel; this corresponds to the power spectrum integrated over time) of the filtered velocity magnitude for the reference buoys. The other buoys in each corresponding season indicated similar results and can be found in Figs. B11 and B12 of Appendix B. Despite the use of the Butterworth high-pass filter, majority of the power (found within the cone of influence) in the Winter buoys' velocity spectra remained at the multi-day periods with peaks at 64, 128 and 256 hours (≈ 3 , ≈ 5 and ≈ 10 days respectively). While the Winter buoys did respond to the cyclones differently, these intensifications found at the low frequencies can be associated to passing cyclones. This response was strongest during the fifth and sixth cyclones (19 and 25 August, respectively) when the Winter buoys switched from drifting north-eastwards to predominantly eastwards (Fig. 3.4a), under the action of winds with speeds $> 25 \text{ m s}^{-1}$ (Fig. 3.7a-d). After the seventh cyclone (6 September), ISVP 3 continued to exhibit increased power at the low frequencies. We attribute this to the long period of high pressure between the 7-30 October (Fig. 3.3), when strong winds persisted (Fig. 3.7a-d). ISVP 1 measured less power during this period, possibly because it drifted in between ice floes. Overall, these increased power intensifications at the lower frequencies were due to the direct transfer of momentum from the wind forcing. This can be seen in Fig. 3.8a and b where the drift spectra, although with less power, followed the energy cascade of the wind within the lower frequency range. This is in agreement with previous literature, where it was shown that the cascade of energy arises from non-linear ice dynamics (Heil and Hibler, 2002; Geiger and Drinkwater, 2005), and can eventually lead to inertial oscillations in the ice motion (Heil et al., 2009). However, as indicated by the wavelet spectra (Fig. 3.9a and b; right panel), the Winter buoys (including ISVP 2 that drifted for half the time; Fig. B11) continued to exhibit no statistically significant power within the inertial range, even after the high-pass filter was applied. There were a few "pulses" of energy at the inertial frequency (Fig. 3.9a and b; left panel), but these were very short-lived and much weaker than the synoptic response of the ice cover. Furthermore, ISVP 1 continued to show increased and statistically significant power at the very high frequencies, which were not always associated with passing cyclones.

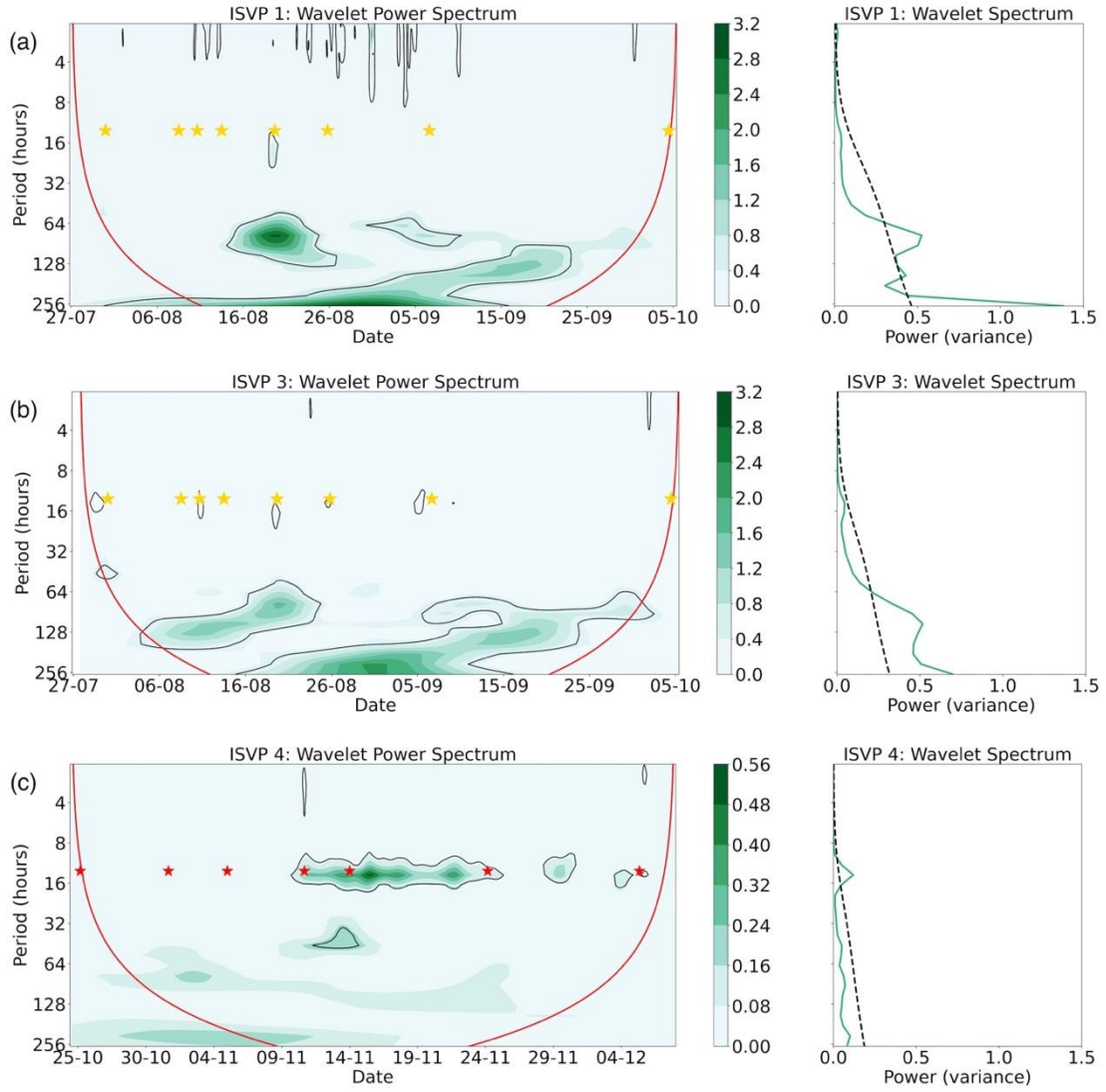


Figure 3.9. (a) The wavelet power spectrum with time (left panel) and wavelet spectrum (right panel) of the filtered velocity magnitude spectrum of ISVP 1, during winter. (b) Same as (a) but for ISVP 3, during winter. (c) Same as (a) but for ISVP 4, during spring. The red line indicates the cone of influence (left panel), the black contours (left panel) and black dashed lines (right panel) indicate the 95 % significance level. The star symbols are the same as in Fig. A1. Due to the difference in the intensity of the power between the seasons, the ranges are different between the Winter buoys and the Spring buoys.

The spring buoy, ISVP 4 (Fig. 3.9c), also indicated low-frequency power intensifications (within the cone of influence) that were associated with the passage of cyclones. However, compared to winter, the power of the Spring buoys at the multi-day periods was far weaker, and also statically non-significant in the wavelet spectrum (Fig. 3.9c; right panel). Rather, the majority of the power was found at the inertial frequency. This is because in addition to the synoptic response of the Spring buoys, the fourth to sixth cyclones (10-24 November) also generated strong inertial oscillations, which occupied a well-defined frequency band (≈ 2 cycles day⁻¹). At a closer inspection, the dates of these strong inertial oscillations corresponded to the northwards drift of the Spring buoys when the ice drift velocity (Fig. 3.7e and f) was governed by high-frequency oscillations, with increased meandering and TSEs (Fig. 3.5b and f, respectively). These inertial oscillations continued until after the passage of the sixth cyclone, when they then dissipated within a few days, due to friction at the ice-ocean interface and/or the internal stresses caused from mechanical interactions within the ice (Colony and Thorndike, 1980; Leppäranta,

2011; Gimbert et al., 2012b; Lei et al., 2021; Marquart et al., 2021). The Spring buoys thus returned to a more straight-line drift path (Fig. 3.4b). However, another shorter-lived inertial response did occur before the passage of the seventh cyclone (5 December).

3.5. Discussion

3.5.1. Detecting ice type from drift features

We will first discuss the four Spring buoys, which were deployed during well-developed, first-year ice consolidated ice conditions (Fig. 3.1b), where the internal ice stresses are expected to be significant (Kawaguchi et al., 2019). Thus, while the sea ice drift followed the winds, the drift velocity exhibited a dampened signal (Fig. 3.7e and f). However, after the second cyclone (31 October), the air temperatures increased and began oscillating with a diurnal frequency near melting point (Fig. 3.3b and c). These sustained higher air temperatures, along with increased waves-in-ice activity during the passage of the second and third cyclones (31 October and 5 November; Sect. 4.3; Thomson et al., 2023), most likely caused the consolidated ice cover to break up and melt. Subsequently, the ice cover became more susceptible to drift and deformation by the winds and ocean currents. This can be seen by the two earlier peaks in the daily meander coefficient time series (Fig. 3.5b), with the concurrent notable increases in the TSEs (Fig. 3.5f) and in the drift velocities (Fig. 3.7e and f). Since the ice cover would have become more susceptible to wind forcing, this led to the growth of the zonal component of the absolute dispersion (Fig. 3.5d) and hence, the initial 3°-5° eastwards displacement of the ice floes (Fig. 3.4b). Consequently, while the mean wind factors and mean turning angles indicated a dynamic response of the ice floes to wind forcing, we found little correlation of the overall drift velocity to the wind velocity, using the two different methods (Table 3.2). We later recomputed these two correlation parameters using ISVP 4 from its deployment date to 9 November (before the inertial oscillations were excited), and they both increased from $R_v^2 = 0.54$ and $R_p^2 = 0.27$ to $R_v^2 = 0.89$ and $R_p^2 = 0.78$, respectively. Therefore, although the power found within the lower frequencies of the wavelet power spectra (Fig. 3.9c; left panel) was non-significant, these measures suggest that during the first few days the spring sea ice was correlated to the wind forcing at the synoptic scale, as observed for the winter ice cover.

During the passage of the fourth to sixth cyclones (10-24 November), the associated transient winds allowed for the development of inertial oscillations within the drift (Fig. 3.7e and f). This is indicative of strong Coriolis forcing, which was likely larger than the advection term in the momentum balance. The inertial response influenced the spring sea ice at the shorter timescales and caused deviations of the ice drift from its initial more straight-line path, as seen by the large peaks in the daily meander coefficient (Fig. 3.5b), the erratically varying TSEs (Fig. 3.5f), and the change to the predominately northwards drift of the ice floes (Fig. 3.4b). This led to the eventual dominance of the total absolute dispersion by the meridional displacement (Fig. 3.5d), and notable changes to the relative dispersion's meridional component (Fig. B6b). Additionally, the inertial response resulted in larger turning angles (Table 3.2), as Coriolis forcing notably deflected the ice floes left from dominant the wind forcing. The use of the filtered wavelet spectra highlights this strong response of the ice cover to the atmospheric forcing at the inertial frequency (Fig. 3.9c). This further indicates that the spring ice cover continued to break up

into smaller floes and brash ice. Our results demonstrated that the spectra of both the drift and wind exhibited an energy cascade from the lower to higher frequencies; however, sea-ice motion exhibited majority of its power at the inertial frequency, rather than at the same frequencies at which the atmospheric forcing was occurring (Fig. 3.8c). We attribute this to the weak geostrophic currents (Table 3.2), which were also observed in Geiger et al. (1998) and Alberello et al. (2020), and the increased mobility of the ice floes (Johnson et al., 2023). Furthermore, while our analysis indicates a plausible correlation between the presence of cyclones and the onset of the inertial oscillations, the power intensifications at the inertial frequency in some cases occurred outside the dates of higher cyclone activity (Fig 3.9c; Fig. B12 in Appendix B). Following the analysis done in Sect. 2.4, we attribute this to the northwards drift of the sea ice while the ice edge was concurrently melting and retreating, where a likely increase in the propagation of waves may also have triggered the inertial oscillations or allowed the geostrophic current to keep the weaker oscillations during the periods of quiescence.

In winter, the mobility of sea ice is generally assumed to decay quickly due to its consolidation (Doble and Wadhams, 2006; Weeks and Ackley, 1986). However, the sampled sea ice continued to exhibit high drift velocities (Fig. 3.7a-d) and a dynamic response to wind forcing, i.e. the trajectories displayed periods of sharp turns and meanders (Fig. 3.4a), and periods of significant stretching and compression (Fig. 3.5e). This erratic nature of ice drift was particularly evident in the period between the second and fifth cyclones (8-19 August), as the daily meander coefficient and TSEs notably increased (Fig. 3.5a and e, respectively). This suggests that the ice drift was more tightly linked to the wind forcing during the passage of these three cyclones. In this case, contrasting to the 2017 buoy analysed in Chapter 2, the Winter buoys were forced closer to the ice-edge region, and the mean turning angles (Table 3.2) were notably smaller than the value of -19.83° reported in Sect. 2.3.3 for pancake ice, and the values presented by the Spring buoys, drifting further from the ice edge. This indicates that the ice floes near the ice edge drifted more closely to the direction of the winds. Moreover, the mean wind factors (Table 3.2) exhibited higher values than 2.73 % in Sect. 2.3.3. They were instead closer to the median wind factor of 3.9 % reported by Wilkinson and Wadhams (2003) for low sea-ice concentrations ($\leq 25\%$). Since the wind factor and turning angle are known to be modified by the underlying ocean current (Nakayama et al., 2012), and in conjunction with the final cumulative meander coefficient of < 1.5 , we suggest that the high velocity of the ACC (Table 3.2) provided a steady source of significant energy. This in turn modulated the relationship between the buoys' drift velocity and the wind vectors. Subsequently, this led to the enhancement of the wind-driven ice drift by the eastwards flowing ACC, which allowed ISVP 1 and ISVP 3 to travel over 25° (> 1500 km) eastwards in 70 days (Fig. 3.4a; Table 3.1). Together, the analysis indicates that the 2019 winter sea ice was under a much stronger steering influence of the ACC than the ice floe analysed in Chapter 2, which drifted > 200 km from the sea-ice edge and further away from the more intense region of the ACC.

The spectral analysis provides an additional argument. While the wind- and ice-drift velocity components exhibited the typical continuous energy cascade from the lower frequencies to the higher frequencies (Fig. 3.8a and b), this energy cascade did not generate any statistically significant peaks within the inertial range (Fig. 3.9a and b). Majority of the power instead continued to be found at the same frequencies at which the large-scale atmospheric forcing occurred. The amount of power found within these lower frequencies of the wavelet power spectra was also considerably larger than in Sect. 2.3.4 (Fig. 2.13). Geiger et al. (1998), in the western Weddell Sea, also found that the Antarctic Coastal Current provided a steady source of moderate low-frequency power to

the ice drift. Therefore, the results further confirm that the strong wind and oceanic forcing together caused the nonlinear velocity terms to remain much larger than the Coriolis term. Consequently, the winter sea ice did not exhibit a clear energy peak within the inertial range. Together with the significant correlation values (Table 3.2), this high mobility of the winter ice cover suggests that the heterogeneous ice conditions like the ones observed during deployment (Fig. 3.1a) were maintained, and hence free-drift conditions are likely to persist throughout winter in this region. However, the evidence of no inertial oscillations is in stark contrast to the data reported Sect. 2.4 for free-drift conditions. We therefore confirm the stronger role of the underlying current in the region of this study.

We now briefly discuss the features of the Winter buoy, ISVP 1, which was deployed in the interstitial ice between ice floes. Its diagnostics indicated a significantly lower correlation to the winds of only $R_v^2 = 0.53$ and $R_p^2 = 0.50$ (Table 3.2), even though these parameters were computed for the first 10 days when it drifted within both the SSMIS and AMSR2 ice edges (Fig. B1 in the Appendix B). It also exhibited an elevated higher frequency portion in its drift spectra (Fig. 3.8a). This phenomenon was also observed by Doble and Wadhams (2006) for an outer ice-edge buoy, during the pre-consolidation phase. When we recomputed the power spectrum for the first 10 days, the elevated higher frequency portion was reduced (Fig. B13 in Appendix B). This is indicative of increased oceanic forcing in the signal. Moreover, the dates of this higher-frequency signal in the wavelet power spectrum (Fig. 3.9a) additionally corresponded to the dates of the extreme peaks ($> 1 \text{ m s}^{-1}$) in ISVP 1's drift velocity (Fig. 3.7a and b), and peaks in its daily meander coefficient and TSE (Fig. 3.5a and e, respectively). Therefore, we can confirm that ISVP 1 left the ice cover during these periods. In this regard, we warn on the limitations of determining ice type conditions from floating drifters, since they would confound the relationship to wind forcing and eventually the deformation rate.

3.5.2. Seasonal and regional comparison of the deformation proxy

The differential drift of sea ice in each season was principally examined through the evolution of the deformation proxy σ_D (Fig. 3.6), in relation to the passage of cyclones. During spring, although the buoy clusters indicated coherence in its total relative dispersion (Fig. B6b), the associated spring σ_D exhibited large variations that coincided with the occurrences of the cyclones (Fig. 3.6b). Whilst this proxy cannot discern whether this is due to divergence, convergence and/or shear, it is likely due to the varying meridional winds which resulted in the changes in the meridional separation, and more erratic changes in the TSEs (Fig. 3.5f). The σ_D was additionally characterised by regular higher-frequency changes. A spectral analysis of the σ_D (Fig. 3.10b) confirmed these changes to be at a period of 13.57 hours, which is situated within the theoretical inertial range. In contrast to the analysis of the Spring buoys' drift velocities themselves, the spectra exhibited no energy cascade from the lower to the higher frequencies. Power associated at the lower frequencies, prevalent for the sea-ice velocity, was effectively dampened out. The lower frequencies were strongest at ≈ 5 days, and associated with the occurrences of cyclones, with secondary peaks at the lower and inertial frequencies. Similarly, Heil et al. (2008) found that, for buoys further from the Antarctic coastline, the power spectrum was dominated by a peak at the diurnal frequencies with secondary peaks occurring at the inertial and lower frequencies. On the other hand, in coastal regions around Antarctica, the power in sea-ice deformation is rather driven by sub-daily processes without any

low-frequency contributions for all kinematic parameters (e.g. Geiger et al., 1998; Heil et al., 2008, 2009, 2011). This difference is attributed to the different bathymetry, where buoys drifting near the coast are more susceptible to tidal forcing (Heil et al., 2008), which can enhance the sub-daily signal. Collectively, our results indicate a counterintuitive strong de-coupling between the large-scale atmospheric forcing and the sea-ice deformation in spring, despite the expectation that fractured sea ice would be prone to following wind forcing. In the western Weddell Sea, Geiger et al. (1998) showed that while moderate low-frequency currents must also have an effect, the sub-daily and daily deformation processes were driven by the wind-induced inertial oscillations of the ice-ocean system. They additionally demonstrated that spatial features in the underlying current, due to topological features, showed in the non-linear interactions. Thus, while bottom topography would not have had any effect on the deformation of spring sea ice in this open-ocean region, Swart et al. (2020) speculated that following the ice melt, the interactions of freshwater input and intense winds of the Southern Ocean can promote and alter sub-mesoscale eddies. This may be an energy source contributing to the shape of the spring deformation spectra at the higher frequencies (Fig. 3.10b) in addition to the high-frequency ocean oscillations, while lower frequencies were more likely coupled to the intermittent winds and the weaker geostrophic currents.

The winter deformation proxy (Fig. 3.6a), along with its corresponding dispersion analyses (Figs. 3.5c and B6a), showed pronounced fluctuations, throughout the analysed period, that were notably related to the occurrences of cyclones. This was due to the high mobility and dynamic response of the winter ice floes to the winds and ACC. Therefore, the winter ice cover was unable to transmit the stress necessary to resist deformation by the passage of the cyclones, and the associated wind directional changes (Fig. 3.7a-d). In agreement with Hutchings et al. (2011), we find that since the internal ice stresses were not significant in winter, the sea-ice velocity (Fig. 3.7a-d) and deformation time series followed the wind forcing on the ice cover, and thus remained linearly related to the large-scale atmosphere. This was further exemplified by the power spectrum of the winter σ_D (Fig. 3.10a). The σ_D spectra similar to the velocity spectra formed a typical, continuous energy cascade from the lower to higher frequencies, with majority of the power found at the same frequencies at which the large-scale atmospheric forcing was occurring (Fig. 3.8a-b). In accordance with Geiger et al. (1998), we suggest this was additionally due to the steady source of low-frequency energy from the ACC, which was in strong coherence with the wind. Therefore, unlike the spring σ_D and previous literature, the winter σ_D like its drift velocity remained strongly coupled with the atmosphere in winter.

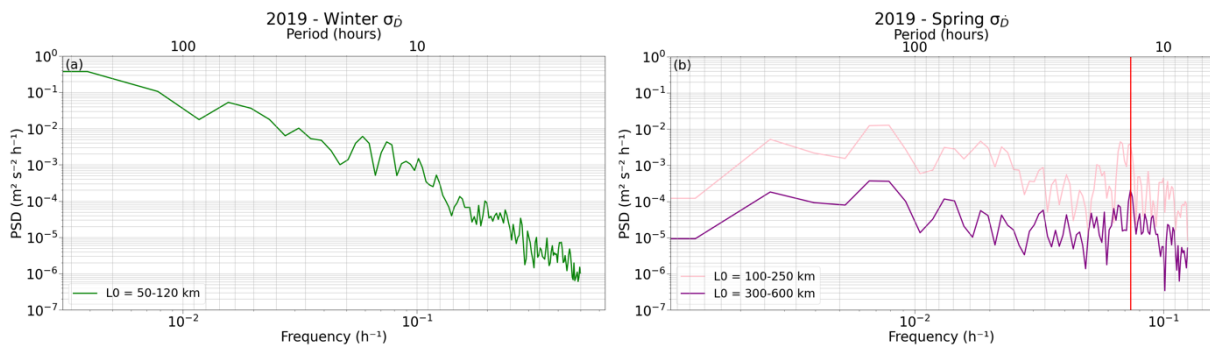


Figure 3.10. (a) Power spectral density of the winter deformation proxy σ_D . (b) Same as in (a) but for spring, where the red vertical line indicates the peak associated with inertial oscillations at 13.57 hours). L0 denotes the horizontal spatial scale for each cluster of buoys. The line colours correspond to the σ_D time series in Fig. 3.6a and b.

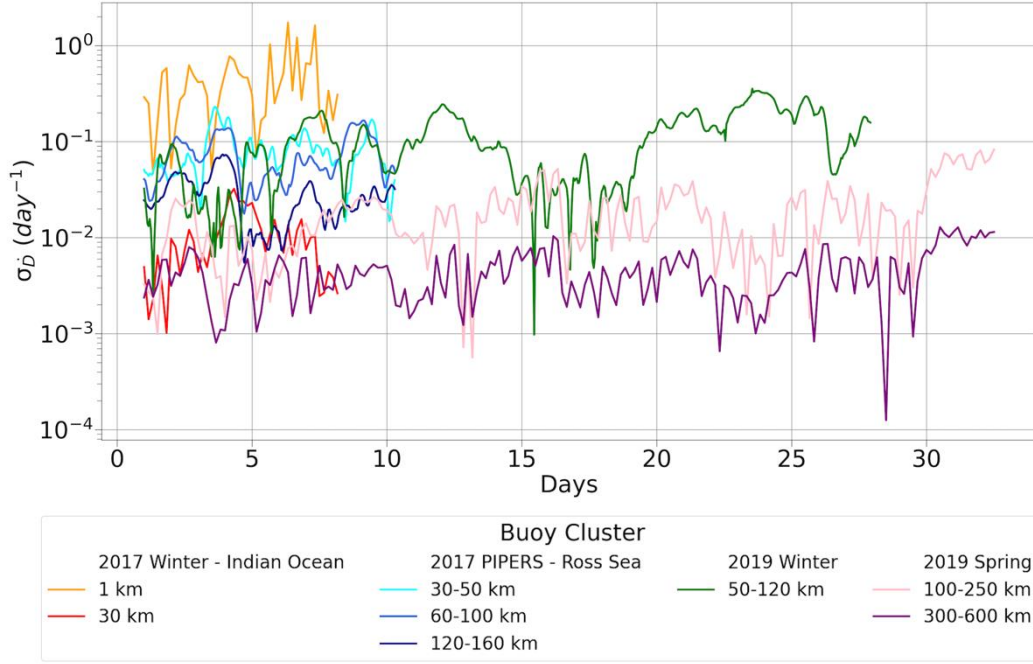


Figure 3.11. The standard deviation $\sigma_{\dot{D}}$ of the deformation rate $\dot{D}$ versus the time interval, in days from start of analysis, for buoys from the 2017 Winter Cruise, 2017 PIPERS winter campaign, and the 2019 Winter and Spring Cruises. Each cluster of buoys has been denoted below, with their corresponding initial horizontal spatial scales (L_0).

The sea-ice deformation is scale dependent as the ice velocity field is known to be spatially discontinuous (Lindsay et al., 2003). The two horizontal spatial scales of the spring cluster allowed for a comparison of the deformation across different spatial scales. Our analysis shows that as the spatial scale decreases, larger strain rates become apparent, which we find in agreement with previous literature (e.g. Rampal et al., 2008; Hutchings et al., 2011; Lindsay et al., 2003). This indicates that the largest strain rates (and gradients) are held by the smaller portion of the sea-ice area. Another signature of this deformation localisation at smaller spatial scales, is the increase of $\sigma_{\dot{D}}$ power as the scale is reduced, which is in agreement with Marsan et al. (2004). However, very few buoys have been deployed in the Antarctic MIZ to allow for a full comparison of the deformation state under different seasonal conditions or across regions. We provide an initial summary by including additional data available to us from two buoy arrays deployed in different regions around Antarctica (see Fig. 1.4). The first was deployed during the 2017 Winter Cruise at approximately 62° S and 30° E (Sect. 2.2.1; Machutcheon et al., 2019; Vichi et al., 2019; Alberello et al., 2020), and are analysed in this study between 5-13 July 2017. The second array included six of the fourteen buoys from the 2017 Polynyas, Ice Production, and seasonal Evolution in the Ross Sea expedition (PIPERS) winter campaign, which were deployed at approximately 67° S and 180° E (Kohout et al., 2020). These six buoys are analysed in this study between 13-23 June 2017.

The results from Fig. 3.11 show an overall reduction of deformation as the area, over which it is computed, is increased. However, the 2017 Winter Cruise buoys, and more noticeably at the 30-km spatial scale (red line), exhibited a lower magnitude than the 2017 PIPERS and the 2019 Winter buoys of similar spatial scales. While all three of these buoy arrays were deployed on pancake-ice conditions (Alberello et al., 2020; Kohout et al., 2020), where we would assume a similar rheology of the ice cover, the $\sigma_{\dot{D}}$ of the 30-km cluster was rather more similar

to that of the 2019 Spring buoys, but with significantly larger spatial scales. As waves help to maintain the pancake-frazil ice conditions (Weiss and Marsan, 2004; Kohout et al., 2014; Vichi et al., 2019), we attribute this to the varying waves-in-ice conditions occurring during the drift of the three winter-buoy arrays. Alberello et al. (2020, their Fig. 3) reported significant wave heights of 0-6.25 m measured by one of the 2017 Winter buoys, and drift speeds which predominantly fluctuated between 0-0.5 m s⁻¹ in mostly 100 % ice concentration. Kohout et al. (2020; their Fig. 4a) reported larger maximum significant wave heights of $\approx$ 2-9 m and drift speeds reaching $\approx$ 1 m s⁻¹, between 13-23 June 2017, in a region of > 80 % ice concentration. Therefore, as the wave heights and drift speeds were generally lower during the 2017 Winter-buoy drift, their freedom to respond to wind and ocean forcing most likely was reduced, possibly due to the close packing of the floes. This would have resulted in a more significant ice rheology and thus, a higher resistance to deformation by the winds and ocean currents. Therefore, while we can confirm that the estimated deformation is a function of the area over which it is calculated, we find that in the Antarctic it is also strongly determined by the rheology of the ice cover, which can be preconditioned from previous seasons (Lei et al., 2020), as well as the atmospheric and oceanic forcing under which it is governed. Consequently, Antarctic sea ice may not always exhibit a classic de-correlation length scale, which is in agreement with Hutchings et al. (2011) for the Arctic. Thus, the magnitude of deformation, for similar spatial scales, may vary between different seasons, regions and proximity to the sea-ice edge and the Antarctic coastline.

3.6. Conclusions

In this study, we analysed a unique dataset of drifting sea-ice buoys in the Atlantic MIZ in winter and spring 2019, which showed how the evolution and spatial pattern of sea-ice drift and deformation were affected by the balance between external atmospheric and oceanic forcing and local ice conditions. We compared the estimates of deformation with the few datasets from other Antarctic regions and seasons. Here, we highlight our results and conclusions:

- During winter, the ACC modulated the relationship between ice drift and wind forcing, near the sea-ice edge. This led to a highly energetic and mobile ice cover that was characterised by free-drift conditions, predominantly in the zonal direction. The resulting drift and deformation were primarily driven by large-scale atmospheric forcing, with a negligible inertial response. The relationship between wind forcing and sea-ice drift and deformation both remained linear in winter, with the kinematic parameters strongly coupled to the atmospheric forcing.
- During spring, sea-ice drift was initially driven by large-scale atmospheric forcing, but with a dampened signal due to the consolidated ice conditions at deployment and stronger internal stresses. As the surface heat balance changed, it caused the ice cover to melt and break up. This led to an increase in the drift kinematics, and the ice drift changed to be dominated by the inertial response. However, the deformation spectra in spring indicated a strong de-coupling to large-scale atmospheric forcing, which is counterintuitive in melting conditions. We interpret it as an increased response to the ocean forcing, with higher frequencies driven by ocean oscillations and possibly influenced by sub-mesoscale flows.

- A comparative analysis of the existing datasets revealed that Antarctic MIZ deformation is strongly determined by the rheology of the sea ice, as well as the atmospheric and oceanic forcing on the ice cover. This indicates that a classic decorrelation length scale may not always exist. Therefore, the magnitude of deformation may vary with ice type between seasons, regions and the proximity to the ice edge and the coastline, for similar spatial scales.

In summary, the present study highlights the need for a better understanding of the impacts of ocean currents and waves on the Antarctic marginal ice zone to help fully understand and quantify the effects of atmospheric forcing, which is one of the main drivers of drift and deformation. The paucity of oceanic observations, especially in the ice-covered Southern Ocean has meant that these drivers are poorly constrained. Therefore, the collection of oceanic in-situ observations is vital. These should include (1) measurements surface currents beneath the ice so that we can discern ice drift and deformation due to wind or ocean currents, and (2) the continuation of waves-in-ice measurements to not only understand how waves maintain a mobile ice cover, but to also understand how sea ice attenuates the wave energy. Additionally, buoys that are ice-tethered and sink when ice melts, acting as an effective Lagrangian tracker of the ice itself, should become a priority, although environmentally undesirable. They avoid the aliasing of drift information and a more reliable detection of the ice floe lifetime, without making assumptions based on the remotely sensed edge detection. This is particularly important as there are known errors in satellite products, especially at the ice edge and during the melt season.

Chapter 4

4. Effects of wave propagation into the Antarctic marginal ice zone during winter and spring of 2019

Abstract

This chapter applies a novel method for measuring waves in the Antarctic MIZ using alternative data analogous to buoy's measurements. Ship motion was measured from an on-board IMU, during the 2019 Winter and Spring Cruises to the MIZ. Using the wave-buoy analogy (WBA), this was converted into an incident wave spectrum through a ship-specific response amplitude operator, which acts as a transfer function. Estimates of the related wave parameters – significant wave height and mean wave period – are obtained along the ship's transect through the MIZ. While this method has been previously applied in the open ocean, the WBA technique allows for the unique opportunity to estimate waves-in-ice. The reconstructed wave parameters have been calibrated with satellite-based and reanalyses estimates in the open ocean and compared with measurements from two concurrent SWIFT buoys, drifting within the ice cover. The results show that wave parameters respond to changing ice conditions, distance from the ice edge, and cyclone passage. Larger waves and shorter periods are observed closer to the ice edge, with wave energy penetrating over 200 km when mobile floes dominate. In compact ice conditions, the significant wave heights were negligible. Thus, the study highlights the well-known complex wave-ice relationship, where waves decay based on the nature of the sea ice, and vice versa. The ship's motion response did however unexpectedly detect higher-frequency waves in consolidated-ice conditions. This was attributed to non-wave related phenomena. Analyses of the longer Spring Cruise highlights the impact of passing cyclones as they generate increased wave energy near to the ice edge, and make the ice cover more mobile and susceptible to drift and deformation. Additionally, it is shown that these storm-generated waves play a crucial role in influencing the retreat of sea ice during the melt season. Overall, this study emphasizes the importance for more high-resolution measurements for understanding air-ice interactions in the Antarctic MIZ.

4.1. Introduction

The significance of the interactions between open-ocean waves and sea ice has provided a strong incentive to observe, understand, simulate and predict current and future MIZ conditions and processes (Brouwer et al., 2022). Over the past decade, pivotal experimental campaigns have supplied much needed measurements of wave

evolution in the MIZ (Thomson, 2022). However, there are still prominent knowledge gaps of wave-ice interactions within the Antarctic MIZ. This is driven (partially) by the scarcity of in-situ wind, wave and sea-ice data (Montiel et al., 2016; Stopa et al., 2016). However, this is changing with the introduction of inexpensive sensors that can be more easily developed (Waseda et al., 2022). Modelling efforts concerning waves-in-ice have also progressed significantly over the years, although the heterogeneous nature of the MIZ and the vast variety of wave attenuation processes continue to make wave prediction in ice difficult (Squire, 2018; Boutin et al., 2021). Predictions near the ice edge are firstly limited by the uncertainty of the position of the ice edge (Schweiger and Zhang, 2015). This is supplemented with additional challenges such as the incomplete understanding of the growth and evolution of a heterogeneous sea-ice surface, as well as the attenuation of wave energy into the ice-covered ocean (Voermans et al., 2019; Rogers et al., 2021; Alberello et al., 2022). Understanding these processes within the MIZ and parametrising their effects in models is imperative not only for reproducing and predicting local conditions in the MIZ but also at the larger scales (e.g. Bateson et al., 2020).

The interactions between waves and sea ice contribute to several aspects of the polar climate (Kousal et al., 2022). Ocean surface waves have been shown to govern the extent of the Antarctic MIZ and its properties (e.g. Kohout et al., 2014). Thus, also the atmosphere-ocean fluxes and the seasonal cycle of the ice extent (Brouwer et al., 2022). During the sea-ice growth season, the mobility of the ice-covered surface by storm-generated waves contribute to sea-ice concentration changes (Vichi et al., 2019), which may lead to an intensified cooling of the ocean surface. This allows the surface to refreeze during periods of quiescence, after ice floes have drifted and have been gathered in different areas under the action of winds and ocean currents (Kousal et al., 2022). In turn, wave generation and attenuation is suppressed. Li et al. (2021) demonstrated that this wave-induced break-up of sea ice can lead to increases in sea-ice concentration and thickness during the growth season. On the other hand, during the melt season, wave-induced ice break-up leads to the warming of the ocean surface and the consequent acceleration of sea-ice melt (Kousal et al., 2020). A reduction in albedo of the ice-ocean surface further accelerates this ice melt, due to the increase in area of the darker ocean below (Curry et al., 1995). As a result, ocean surface waves have larger fetches and grow bigger, and hence, promote the break-up and retreat of sea ice in the MIZ (Kohout et al., 2014; Passerotti et al., 2022). Access to this information on sea-ice dynamic properties is therefore necessary to the understand ice drift and deformation.

Research activity on wave-ice interactions in the Southern Ocean has taken place through numerical wave modelling, theoretical studies, laboratory studies, and field observations, either in-situ or from remote sensing (Rogers et al., 2021; Rossi et al., 2021; Thomson, 2022). However, historically, research and observational field work on wave-ice interactions in the Southern Ocean has gained far less interest than the Arctic Ocean because of its more remote location and harsher environment. The earliest experiment conducted in Antarctic sea ice was that of Robin (1963), which described measurements using a shipborne wave recorder in the Weddell Sea during 1959-1960. The results were described in terms of the distance of penetration of swells in the ice cover, which Robin (1963) found to depend on wave period, ice thickness and floe size. In 1986, a break-up event was observed in the Weddell Sea's pack ice > 560 km from the ice edge, and this motivated the analysis by Liu and Mollo-Christensen (1988). Years later, Doble and Wadhams (2006) found that the propagation of waves had a notable impact on ice growth in the Weddell Sea, during the autumn advance season. Kohout et al. (2014) analysed data from the deployment of waves-in-ice observation systems (WIIOS) in the East Antarctic MIZ, and provided

evidence that wave-induced break-up events are responsible for the correlation between the trends in local significant wave height and trends in the contraction and expansion of the ice edge. In July of 2017, two WIIOS were deployed by Eayrs et al. (2019b) in the Atlantic-Indian Ocean sector of the Southern Ocean. This provided information on storm-generated waves and how they contributed to the maintenance of an unconsolidated ice field (Vichi et al., 2019), and additionally, how the wave field maintained $\approx 50\%$ of its energy after 100 km of propagation into the MIZ (100 % SIC; Alberello et al., 2020). Lastly, Kohout et al. (2020) deployed an extensive array WIIOS buoys in the Antarctic MIZ to quantify the rate of attenuation in sea ice. They further elucidated that waves decayed exponentially with distance from the ice edge – a phenomenon known for many years (e.g. Robin, 1963). Ice-tethered wave buoys, while supplying high-resolution temporal data, can only endure the harsh polar environment for a short period (usually few weeks at most; Nelli et al., 2023). They are additionally limited to a certain threshold of high-frequency waves and thus can underestimate maximum wave displacement (Campbell, 2014). Therefore, considering the several ship expeditions that take place every year, ship-based observations – whether visually or through technological platforms – can be used as a vital source of wave data for the polar latitudes (Derkani et al., 2021; Alberello et al., 2022). This includes, remote sensing methods, using the ship's marine radars, which are widely used to measure the directional wave spectrum in the open ocean with good accuracy (Nelli et al., 2021). However, strong backscatter from the ice cover is a major source of uncertainty in the MIZ (Derkani et al., 2021). Stereo imaging has also been applied successfully to detect two-dimensional spectral wave properties during ship expeditions (Alberello et al., 2022).

Recently, a new and promising technique, referred to as the wave-buoy analogy (WBA), has been reported in literature, where a ship acts as a wave-rider buoy (Nielsen and Jensen, 2011; Nielsen and Dietz, 2020; Nelli et al., 2023). The WBA uses standard ship sensors and consists of converting the ship's six-degree-of-freedom motion response from on-board inertial measurement units (IMU) into an incident directional wave spectrum, through the response amplitude operator (RAO; Nielsen and Jensen, 2011; Newman, 2018; Nelli et al., 2023). The RAO is a transfer function (i.e. a filter that transfers waves into responses, usually in the frequency domain; Nielsen et al., 2021) that defines the behaviour of the ship at sea. The WBA principle also depends on linear wave theory. This theory provides a linearised description of the propagation of surface gravity waves in water with idealised physical properties and motions, and with gravity as the only external force (Holthuijsen, 2007). Linear wave theory allows for the reasonably accurate description of wave characteristics such as wave height, period and speed, using a sinusoidal (harmonic) model of the wave form (Hedges, 1976; Mayhew, 2015). However, the reconstruction of the wave spectrum using the WBA can still be difficult since the ship itself acts as a low-pass filter. This hinders the accuracy in the upper portion of the spectrum, and applications linked to non-linear wave dynamics (e.g. estimates of wave height distribution and probability of the occurrence of large waves) and wave-structure interaction (Nelli et al., 2023; and references within). Another reason for inaccuracies lies in the estimation of the RAO itself, since it requires accurate knowledge of the ship's hull lines (Nielsen et al., 2021), which are further complicated by changes in ship speed and bearing during navigation. Nevertheless, the sea-state estimations derived with estimated RAOs have been compared against co-located measurements from marine radar to validate the WBA technique. This method has been shown to provide reliable measurements of the integrated wave parameters in the open ocean such as significant wave height and mean wave period (e.g. Nielsen and Jensen, 2011; Nielsen and Dietz, 2020; Nielsen et al., 2021; Nelli et al., 2023).

The WBA method allows for estimations of wave parameters that are not affected by meteorological conditions, and it is comprised of a system with a low initial cost due to simple instrumentation that generally needs little maintenance (Nielsen and Jensen, 2011). Moreover, ship-board measurements are extremely practical and offer a platform that is typically not spatially constrained. Thus, while the WBA has been previously applied, calibrated and validated for estimating sea states, its application also may be extended to determine waves-in-ice activity, since concurrent in-situ meteorological and ice conditions can be measured along the ship's transect within the ice cover. Previously, Shestov (2019) applied this technique to estimate wave attenuation through pancake-ice conditions in the Barents Sea. The author reported lower attenuation coefficients compared to prior literature for the same region, which was attributed to the smaller floe diameters. Overall, these measurements can provide a more comprehensive understanding of the air-ice-ocean interactions in the polar regions and will help to increase in-situ wave observations in the Antarctic MIZ.

In this study, information on wave-ice interactions are computed using measurements of the S.A. Agulhas II's vessel motion, obtained during the 2019 Winter and Spring Cruises to the Antarctic MIZ (Sect. 1.4; Ryan-Keogh and Vichi, 2022). The reconstructed sea-state conditions are computed following the WBA methods described by Nelli et al. (2023) and Nielsen et al. (2021). These are then used as proof of concept to estimate the wave parameters in the Antarctic MIZ. Two drifting Surface Wave Instrument Float with Tracking (SWIFT) buoys, that were deployed during these two expeditions (Wahlgren et al., 2023), are used in conjunction with these reconstructed estimates. These data are insufficient to compute and analyse the attenuation rates of propagating waves in the ice cover, but can help to characterise and better understand the relationship between sea-ice type and waves-in-ice, and additionally the seasonal evolution of the Antarctic MIZ. We thus correlated the wave measurements with atmospheric reanalyses data and in-situ sea-ice observations to investigate (i) how the propagation of wave energy may have impacted the MIZ extent and its properties, (ii) how sea-ice type may have affected the propagation of waves in the ice cover, and (iii) how the passage of cyclones over the MIZ influenced these wave-ice interactions, during each of these two 2019 seasons.

4.2. Field data and methods

4.2.1. Records of ship motion and sea states

The term “inertial sensor” is used to represent the combination of a three-axis accelerometer and a three-axis gyroscope. It is often supplemented by a magnetometer, which indicates the direction of the magnetic field (Kok et al., 2017). The gyroscope measures the sensor's angular velocity i.e. the rate of change of the sensor's orientation, and the accelerometer measures the external specific forcing that acts on the sensor. The specific force is comprised of the sensor's linear acceleration and the Earth's gravitational force. Currently, the majority of gyroscopes and accelerometers are built using microelectromechanical system technology (Kok et al., 2017). These instruments are small, light and inexpensive, and their reliability has grown over the years. Therefore, by processing the signals from the gyroscope and accelerometer, IMUs can be employed to derive the dynamic positioning and orientation of the ship during navigation (Figuero et al., 2016).

Observations of the S.A. Agulhas II's ship motion were collected on board during the two 2019 SCALE cruises to the Antarctic MIZ (Ryan-Keogh and Vichi, 2022; Chapter 3). During these two expeditions (see Fig. 4.1), the Sound and Vibrations Research Group from Stellenbosch University installed a full-scale measurement system on the S.A. Agulhas II to measure ship responses during different operational conditions (Ryan-Keogh and Vichi, 2022, p. 280; Taylor et al., 2023a, b). All IMUs were initially aligned with the ship's coordinate system and stored in five-minute-long consecutive batches with a sample rate of 2048 Hz. Data were then processed, using kurtosis and band-pass filtering (Taylor et al., 2023a, b), to attain the vertical translation motion (heave), rotation about the longitude axis (pitch) and rotation about the transverse axis (roll). Only these three components were freely accessible. However, Nielsen et al. (2023) reported that for the WBA any global wave-induced ship response can be considered; for example, one of the six DOF motion components. Therefore, given that gyroscopes are known to incur large uncertainties in their measurements near the poles, and considering the fact that the ship's active stabilisation primarily affects pitch and roll rather than the heave, this study, like the one done by Shestov (2019), only considers heave motion.

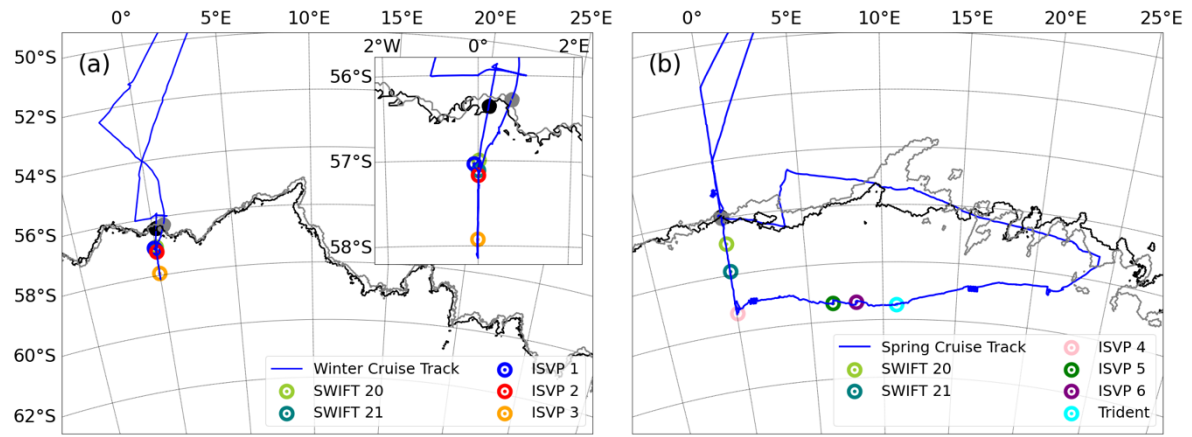


Figure 4.1. Schematic of the S.A. Agulhas II's transect to the Antarctic MIZ during **(a)** 2019 winter, and **(b)** 2019 spring. The inset in **(a)** shows a closer view of the ship track within the MIZ. The blue line denotes the ship's transect. The black line denotes the AMSR2 sea-ice edge (0 % ice concentration) for the day that the ship entered the MIZ (26 July **(a)** and 22 October **(b)**), and the grey line denotes the day the ship left the MIZ (28 July **(a)** and 8 November **(b)**). The black and grey circles denote the ships position on the corresponding days. The deployment position of the two SWIFT buoys (20 and 21; light green and teal, respectively) are shown. The deployment position of three Winter buoys **(a)** and four Spring buoys **(b)**, analysed in Chapter 3 are also shown (colours correspond to Fig. 3.4).

Taylor et al. (2023a, b) validated the data at the locations of interest against an accelerometer not in the six-accelerometer array, and they found the measured and computed signals to agree well (both their Fig. 2.2). They additionally reduced the sampling rate to 10 Hz for operational use (strictly 9.99 Hz). Two different IMU locations were obtained: from Monkey Island, which is located just above the bridge (at a height of 34 m above ocean surface; Alberello et al., 2019); and at the estimated ship's centre of gravity (COG; 8.86 m above baseline of the ship). The COG location is an approximation due to unspecified loading conditions that resulted in an unknown exact center of gravity for the ship, necessitating estimation through various equations. These locations are detailed fully in the user manual supplied by Taylor et al. (2023a, b). However, the wave parameters, reconstructed from the ship's COG, were far smaller than those from Monkey Island. This is presumably due to the ship's increased stability at this location, as it is closer to the waterline. However, its estimation likely would have added additional uncertainties in the RAO calculations. Therefore, herein, we use the local IMU measurements from Monkey Island.

4.2.2. Sea-state estimation from ship motion

With the assumption that the motion response of the ship is linearly dependent on the incident wave field, and that navigation and sea-state conditions are stationary and ergodic within a given sampling interval (Nielsen and Jensen, 2011; Nielsen et al., 2021; Nelli et al., 2023), the WBA builds upon a comparison between the measured response and the directional wave spectrum (Nelli et al., 2023), according to the following governing equation:

$$R(\omega_e) = \int_0^{2\pi} |Z(\omega_e, \vartheta)|^2 E_e(\omega_e, \vartheta) d\vartheta, \quad (4.1)$$

where $R(\omega_e)$ is the measured response spectrum for heave motion, $Z(\omega_e, \vartheta)$ is the ship's RAO related to the heave motion in multiple directions (Sect. 4.2.3), and $E_e(\omega_e, \vartheta)$ is the unknown encounter directional wave spectrum (Faltinsen, 1993). ω_e (in rad s^{-1}) is the angular frequency and ϑ (in radians) is the wave direction. It is important to note that $E_e(\omega_e, \vartheta)$ is the spectral density detected by the moving ship, and therefore it is a distortion of the real spectra due to a Doppler shift (Faltinsen, 1993).

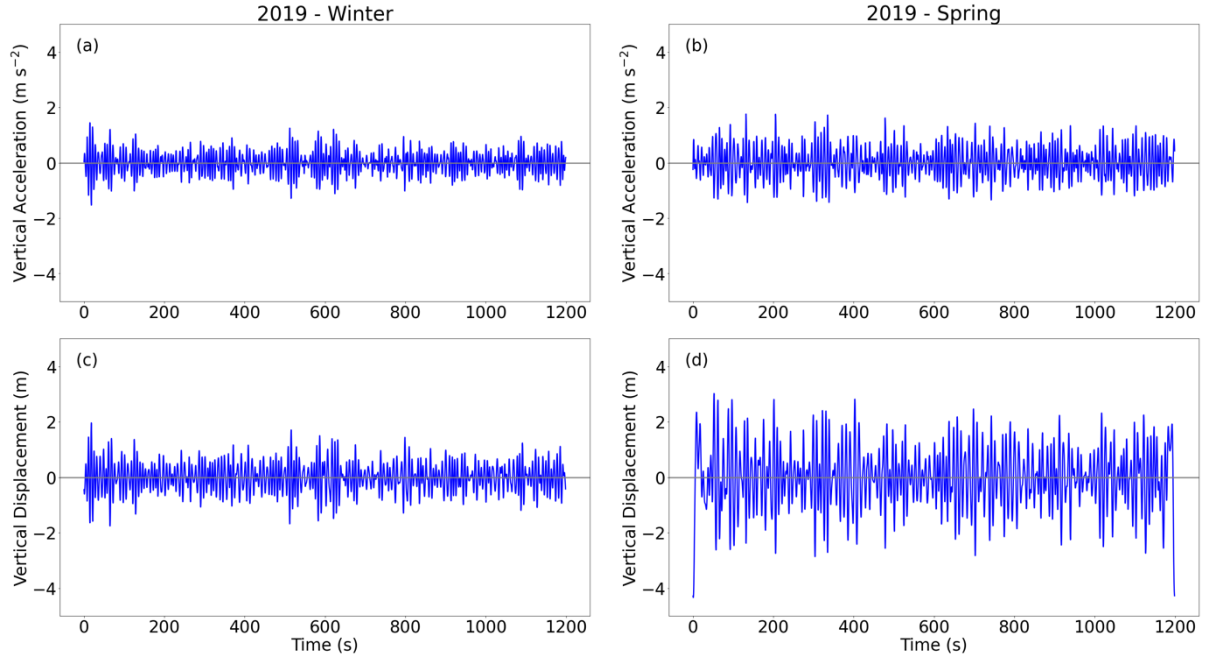


Figure 4.2. (a) Open-ocean example of the measured vertical acceleration at Monkey Island, for 20 minutes, during the Winter Cruise, and (c) the corresponding vertical displacement after integration. (b) Same as in (a) but for the Spring Cruise. (d) Same as in (c) but for the Spring Cruise.

While varying environmental and operational conditions impact and limit the above assumptions, a window smaller than 30 minutes allows for the assumptions to be met for practical purposes (Nielsen et al., 2021). For instance, the World Meteorological Organisation has established a 20-minute window as the standard for the analyses of wave measurements (World Meteorological Organisation, 1998), as it balances the stationarity of wave conditions and enables the collection of an adequate number of waves for a statistically robust analysis (Alberello et al., 2023). Subsequently, the IMU data in each season are sectioned into 20-minute chunks (i.e. $20 \text{ min} \times 60 \text{ s} \times 10 \text{ Hz}$). Following this, the vertical acceleration (m s^{-2}) can be integrated twice to calculate the vertical displacement (in metres; Fig. 4.2). In order to remove any effects of a trend term in the integration process, the chunks are first de-trended and then a Fourier transform is used to transform the heave into the frequency domain (Qihe, 2019). The frequencies are then shifted to centre around zero, filtered to remove undesired frequencies from the acceleration spectra and then integrated twice in the frequency domain. An inverse Fourier transform can be used to get the inverse result i.e. time domain. When the real part is transformed, the necessary displacement in the time-domain is acquired. Lastly, a power spectral density using Welch’s method, with 1024 intervals and 50 % overlap, is applied to the vertical displacement to obtain $R(\omega_e)$ and its corresponding frequencies (Fig. 4.3). Since the ship filters high-frequency modes, components with encounter frequencies $> 2 \text{ rad s}^{-1}$ (i.e. 0.3182 Hz) are removed to reduce noise.

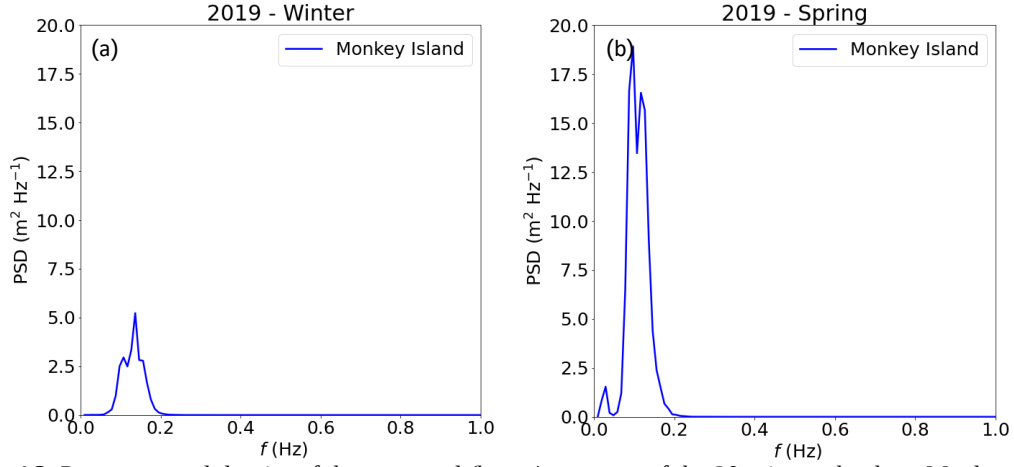


Figure 4.3. Power spectral density of the measured (heave) response of the 20-minute chunk, at Monkey Island, from Fig. 4.2, during (a) winter and (b) spring.

The governing Equation 4.1. can be rearranged to:

$$S(\omega_e) = \frac{R(\omega_e)}{|\zeta(\omega_e)|^2}, \quad (4.2)$$

where the numerator on the right-hand-side denotes the measured response spectrum calculated for the heave motion,

$$\zeta(\omega_e) = \frac{1}{2\pi} \int_0^{2\pi} Z(\omega_e, \vartheta) d\vartheta, \quad (4.3)$$

the denominator is the RAO in the frequency domain (i.e. integrated over all directions), and

$$S(\omega_e) = \int_0^{2\pi} E_e(\omega_e, \vartheta) d\vartheta \quad (4.4)$$

is the encounter wave energy spectrum. A weighted-average approach based on the actual encountered wave direction is desired (Eq. 4.4; Nielsen et al., 2021). In this case, model hindcasts would be the only source for wave direction along the ship's transect, although this has been shown to contribute additional uncertainties (Derkani et al., 2021). In this work, as done in Nelli et al. (2023), Eq. (4.2) is applied without any directional weighting.

The significant wave height (H_S) is computed from the zero-th order moment (m_0) of the wave spectrum (Eq. 4.2) as:

$$H_S^{IMU} = 4\sqrt{m_0}, \quad (4.5)$$

and the mean wave period as:

$$T_m^{IMU} = \frac{m_0}{m_1}, \quad (4.6)$$

where m_1 is the first-order moment of the scalar wave energy spectrum (Thomson et al., 2019). The n -th order spectral moments are given by:

$$m_n = \int_0^\infty \omega_e^n S(\omega_e) d\omega_e. \quad (4.7)$$

The peak period (T_p^{IMU}) is computed as the inverse of the maximum frequency of the scalar wave energy spectrum.

Although the Doppler shift's distortion of the spectral shape does not affect the energy content or the zero-th order moment, the encounter frequencies, and hence T_m^{IMU} , need to be corrected. The absolute frequency, ω , which is the frequency that would be measured by a stationary device is commonly associated to the encounter (intrinsic) frequency, ω_e , using the following equation (Faltinsen, 1993; Nielsen et al., 2021; Nelli et al., 2023):

$$\omega_e = \omega - \frac{\omega^2}{g} V \cos(\beta), \quad (4.8)$$

where V is the forward speed of the ship, g is gravitational acceleration, and β is the angle between the heading of the ship and the prevailing wave direction, ϑ (Nielsen and Jensen, 2011; Rossi, 2021). However, as the wave direction is unknown, the estimated mean wave direction from ERA5 reanalysis is used in its place (Sect. 4.2.5).

Equation 4.8 is based on linear wave theory, which suggests that for the initial assumption that navigation and sea-state conditions are stationary and ergodic (observer fixed in the medium), waves behave according to the dispersion relation (Campana et al., 2016):

$$\omega_e = \sqrt{gk \tanh(kd)} \quad (4.9)$$

where k is the wavenumber, d is the local water depth, and $kd \rightarrow \infty$ ($\tanh(kd) \rightarrow 1$) in deep water ($d > \frac{1}{2}\lambda$; Holthuijsen, 2007). However, as stated above, the ship was travelling with a forward speed, V , relative to the underlying ocean currents. Therefore, a new frame of reference needs to be considered where the frequency of a wave is measured in a fixed frame of reference i.e. fixed to a stationary bottom (Holthuijsen, 2007). In this context, linear wave theory thus relates the absolute frequency to the encounter (intrinsic) frequency by (Collins et al., 2017; Derkani et al., 2021):

$$\omega = \omega_e + kV \cos(\beta), \quad (4.10)$$

where $k = \frac{\omega^2}{g}$ in open water (Fox and Haskell, 2001). The difference between ω_e and ω therefore follows directly from the distortion of the moving ship, and the term kV may be noted as the Doppler shift. The wavenumber is related to the wavelength by:

$$k = \frac{2\pi}{\lambda}, \quad (4.11)$$

where the wavelength for an arbitrary depth is defined as:

$$\lambda = \frac{g(T_m^{IMU})^2}{2\pi} \tanh\left(\frac{2\pi d}{\lambda}\right). \quad (4.12)$$

For deep water and away from coastal regions, $\tanh\left(\frac{2\pi d}{\lambda}\right)$ tends to 1, and thus the deep-water wavelength becomes

$$\lambda = \frac{g(T_m^{IMU})^2}{2\pi}. \quad (4.13)$$

Therefore, from Equations 4.8 and 4.10, T_m^{IMU} can then be corrected with the following inverse relationship:

$$T_m^{IMU} = \frac{2\pi}{\omega_e + kV \cos(\vartheta)}. \quad (4.14)$$

4.2.3. Response amplitude operator

For this study, the RAO, $Z(\omega_e, \vartheta)$, was calculated by the project partner – the University of Melbourne. They estimated it for the S.A. Agulhas II using a Bayesian approach in order to acquire the best convergence between the measured (left-hand-side of Eq. 4.1) and the computed (right-hand-side) response spectrum. The RAO is an inherent feature of the ship and, if it is unknown, it can be estimated by solving the equation of motion of a solid body with several methodologies (Newman, 2018). These are executed in commercial and open-source software. The open-source NEMOH software package (Delhommeau et al., 2022) was shown to perform well against commercial software (Nelli et al., 2021). It is based on linear free-surface potential flow theory and can describe heave motion relatively well (Nelli et al., 2023). This process involves a parametric method based on analytical spectral shapes (see Nielsen et al., 2021; Nelli et al., 2023). The input conditions include the ship speed, geometry, the hull type, and the draught – the vertical distance between the waterline and the bottom on the hull (Fig. 4.4), which on the S.A. Agulhas II it is designed to be 7.65 m (Soal et al., 2015). However, the effective draught during the expeditions was unknown due to unspecified and variable loading conditions. Thus, a trial-by-error approach is applied to identify the draught for the RAO by reducing errors in the assessment of the estimated wave in relation to satellite observations. Since Nelli et al. (2023) found their optimal RAO to be 8.5 m for the Antarctic Circumnavigation Experiment, and considering the S.A. Agulhas II's design draught of 7.65 m, we initially started our approach with a draught of 7 m. However, these smaller draughts did not yield the best correlation with satellite observations. Consequently, we increased the RAOs to the upper end of the range and determined the optimal RAO for both seasons to be 15.5 m. Figure 4.5 denotes the sensitivity of the RAO to different loading conditions, with draughts ranging between 7 m to 15.5 m.

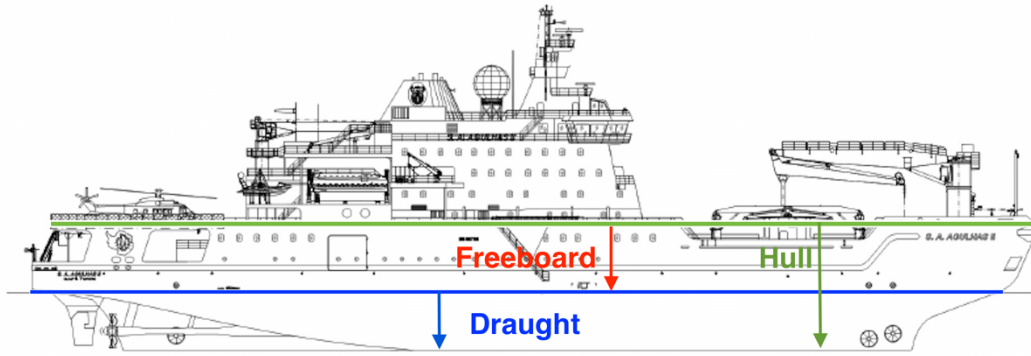


Figure 4.4. Schematic of the S.A. Agulhas II – by Soal et al. (2015). The hull of the ship lies below the green horizontal line. The design draught of the ship (7.65 m) is found below the waterline (blue line). The freeboard is found above the waterline.

The ship's forward speed impacts added mass and dampening (Faltinsen, 1993), and thus it is also required when computing the RAO. Although, a real-time speed adjustment is more desirable, there were missing data in the time series of the GPS position, and hence in the derived forward speed. Consequently, following the initial assumption that navigation and sea-state conditions are stationary and ergodic (Sect. 4.2.1), the ship's speed, during both the Winter and Spring Cruises, is taken as 0 m s^{-1} . In this regard, the effective RAO holds the combination of all unknowns, and thus may vary temporally from the theoretical design draft.

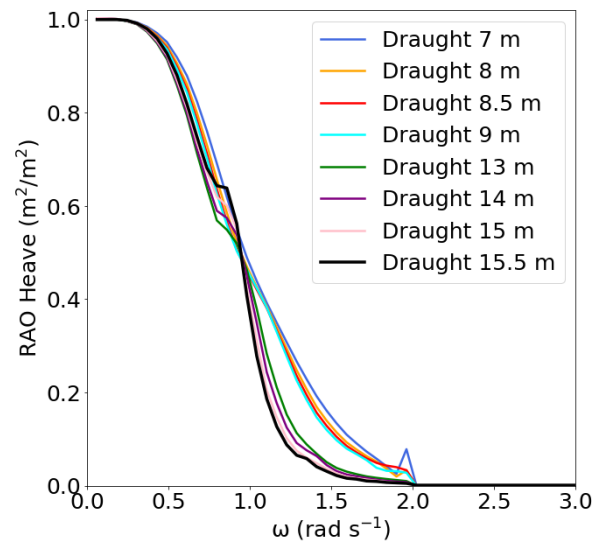


Figure 4.5. The response amplitude operator (RAO) for the heave, with variable drafts. The optimal draught for both seasons is 15.5 m and is denoted by the thick black line.

4.2.4. RAO calibration with satellite data

The response spectra $R(\omega_e)$ and RAO $Z(\omega_e, \vartheta)$ are related (Eq. 4.2) to estimate the significant wave height H_s^{IMU} and mean wave period T_m^{IMU} (Eq. 4.5 and 4.6). These wave parameters, for the period when ship was in the open ocean, are then compared to remotely sensed observations. Since altimeters do not directly measure the wave period, an algorithm proposed by Remya et al. (2010) is applied (see Salles, 2020). This algorithm is based on the wind speed and the significant wave height, which has been shown to have a high accuracy in both wind and swell generated seas (Remya et al., 2010). The remotely sensed observations (H_s^{sat} and T_m^{sat}) were obtained from altimeter sensors mounted on the Jason-2, Jason-3, Sentinel-3A, Sentinel-3B, Saral and Cryosat-2 satellites, and extracted from the Australian Ocean Data Network (see Ribal and Young, 2019). These data records are already quality controlled, calibrated and validated against all available buoy data (Ribal and Young 2019, 2020). Altimetry data provide accurate wave height information up until the start of the ice cover (Monteban et al., 2019). Therefore, only observations north of 55° S are used in order to remove any impacts of sea ice. However, while the reconstructed parameters have a regular time interval of 20 minutes, the satellite observations are scattered in time and space. In order for their significant wave heights to be compared, altimetry data with a spatial range of 0.5° centred on the ship's position, and a time window of ± 2 hours is considered to maintain the same transient states within the open-ocean wave field. However, for the Spring Cruise, only a few data points are available within these ranges. Subsequently, as an additional measure, the H_s^{IMU} are also compared with ERA5 reanalysis data to help determine the optimal RAO.

4.2.5. Environmental data

The reconstructed wave parameters at the ship's location and data from the SWIFT buoys were correlated with SIC data from the 3.125 km spatial resolution AMSR2 satellite product, and the synoptic atmospheric variables from ERA5 reanalysis, following methods outlined in Sections 2.2.2 and 3.2. Meteorological conditions were measured along the cruise track in both seasons. These atmospheric variables from the ship's scientific data system (SDS) were sent to the European Center for Medium-Range Weather Forecasts (ECMWF) to be assimilated into the ERA5 reanalyses. Thus, as shown in Fig. 4.6, the ERA5 data captured the temporal variability of the measured mean sea level pressure (mslp) and air temperature well, with small biases in the 10-m wind speed. Biases are systematically larger when ship data are not assimilated, as for instance demonstrated for an earlier experiment in Vichi et al. (2019). The ERA5 atmospheric conditions herein are not corrected further since they are used to demonstrate the relative changes in the surrounding environment rather than to quantify momentum transfers. The significant wave height (swells and waves) and mean wave direction data are also obtained from the ERA5 dataset with an hourly time interval. The mean wave direction data are interpolated to 20 minutes and co-located to the ship's position in order to correct for the Doppler shift (see Eq. 4.8).

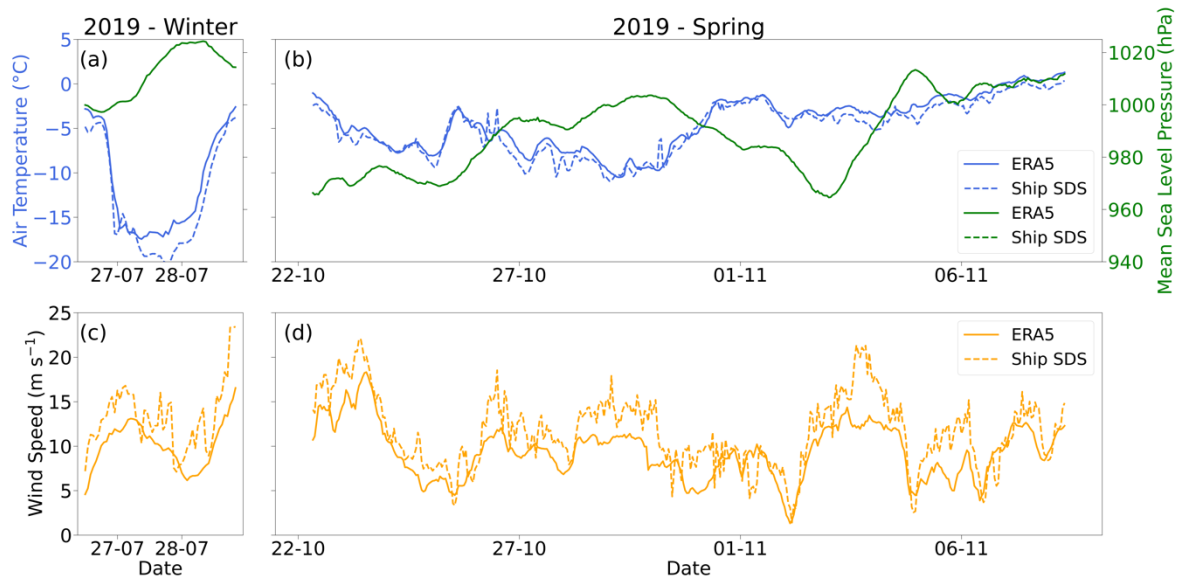


Figure 4.6. Assessment of ECMWF atmospheric analysis product (ERA5) with in-situ observations from the ship, during in-ice activities. Data presented here are the time series of **(a-b)** the air temperature (in °C) and mean sea level pressure (in hPa), and **(c-d)** the 10-m wind speed (in m s^{-1}). The ship's data (from the on-board scientific data system, SDS) are interpolated to an hourly time interval for comparison with the hourly ERA5 reanalysis data. Winter is shown on the left-hand-side, and spring is shown on the right-hand-side.

The remotely sensed 0 % SIC data have been used to estimate the ice edge, as in Chapters 2 and 3. However, since photographs of the local sea-ice conditions were gathered along the SWIFT buoys' trajectories and sea-ice observations along the ship's transects, it allows for an additional comparison between satellite remote sensing and in-situ observations. Ship-based observations were collected by Hepworth et al. (2023) using the Antarctic Sea Ice Processes and Climate (ASPeCt) protocol (Worby and Allison, 1999; ASPeCt, 2018), which is standardised, quantifiable, and internationally recognised as the method for collecting sea-ice characteristic and thickness observations. While this observational method may be limited due to the subjective biases of human interpretation of the sea-ice environment (Beitsch et al., 2014), it does allow for the determination of how well satellite products estimate the ice edge, made up of fragmented ice floes. It also provides a unique in-situ opportunity to determine the ability of remote sensing products in describing Antarctic MIZ extent.

4.2.6. Waves-in-ice buoys

Two drifting Surface Wave Instrument Float with Tracking (SWIFT) buoys were deployed and retrieved during both the Winter and Spring expeditions (Fig. 4.1; Wahlgren et al., 2023; Thomson et al., 2023). These floating buoys, named SWIFT 20 and SWIFT 21, were each fitted with a Geoforce GPS tracker, a wave-tracking IMU, meteorological stations (airmar WX-200 Ultrasonic weather stations) and a low-resolution GoPro camera ≈ 1 m above the ocean surface (Table C1; Wahlgren et al., 2023). They were deployed directly into the water, in between ice floes, and estimated the wave field by measuring the horizontal (GPS) and vertical (IMU) displacements of the buoy in the sea-ice cover (Thomson, 2012). The power spectral density is computed from the displacements by presuming a circular orbital motion (Thomson et al., 2018), which is predicted using linear wave theory for deep-water waves (Wahlgren et al., 2023), in a frame of reference moving with the current (i.e. Eq. 4.9;

Holthuijsen, 2007). The reliability of this is analysed using the check factor – the ratio between the measured horizontal and vertical displacements (Thomson et al., 2015). This ratio should equal to one, and significant variations denote that the measurements are unreliable. The significant wave height, H_s^{SWIFT} , and mean wave period, T_m^{SWIFT} , can then be derived using Eq. (4.5 and 4.6), respectively. The raw data were collected and pre-processed to an hourly sampling interval by Thomson et al. (2023).

During winter, SWIFT 20 was deployed ≈ 25 km south of the ice edge at 56.6° S, 0.0° E, on the 26 July 2019 at 22:58. SWIFT 21 was deployed ≈ 10 nm (18.52 km) south of SWIFT 20 at 57.07° S, 0.0° W, on 27 July at 01:19 (Ryan-Keogh and Vichi, 2022; Fig. 4.1 and 4.7) – the same location as polar-Iridium Surface Velocity Profiler number 1 (ISVP 1; Table 3.1). Both deployments were done from the side door of the environmental hangar and lowered via a slip line while being steadied by the attached float lines. Immediately after their release, the ship thrust away and thus pushed the buoys into the ice. SWIFT 20 and 21 continued to remain between 14-18 km apart as they drifted north-eastwards (Fig. 4.7a) until 28 July, when they were retrieved due to technical issues. The SWIFT buoys were deployed and remained in sea-ice conditions that consisted of interstitial frazil ice and 1-5 m pancake ice floes, around 40 cm thick (Wahlgren et al, 2023). During this short period in the ice, both SWIFT buoys measured significant wave heights (H_s^{SWIFT}) ranging between 0.5 and 1 m, while their mean wave period (T_m^{SWIFT}) ranged between 10 and 15 s (Fig. 4.8g and i, respectively). SWIFT 21 indicated slightly higher wave attenuation, since it was deployed further south into the ice cover (Fig. 4.7a), and had lower H_s^{SWIFT} and T_m^{SWIFT} values (Wahlgren et al., 2023). These measurements were taken at distances of < 200 km from the ice edge (Wahlgren et al., 2023).

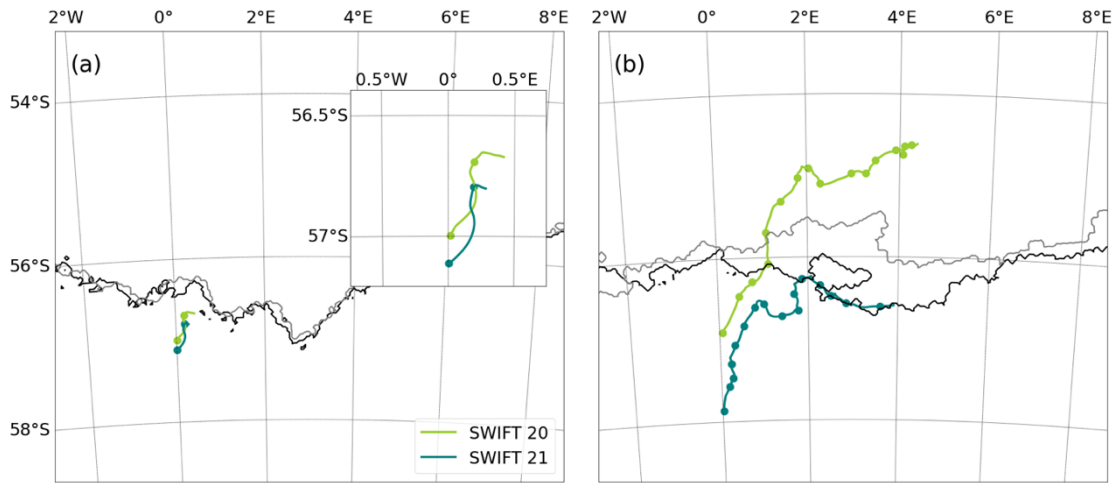


Figure 4.7. Trajectories of SWIFT 20 (light green) and SWIFT 21 (teal) in **(a)** winter and **(b)** spring. The inset in **(a)** shows closer view of the SWIFT buoys' winter trajectories. The corresponding coloured circles along the buoys' trajectories denote their daily positions. The black (grey) contour denotes the AMSR2 0 % sea-ice concentration contour on the day that the S.A. Agulhas II entered (exited) the MIZ. These dates are the same as in Fig. 4.1.

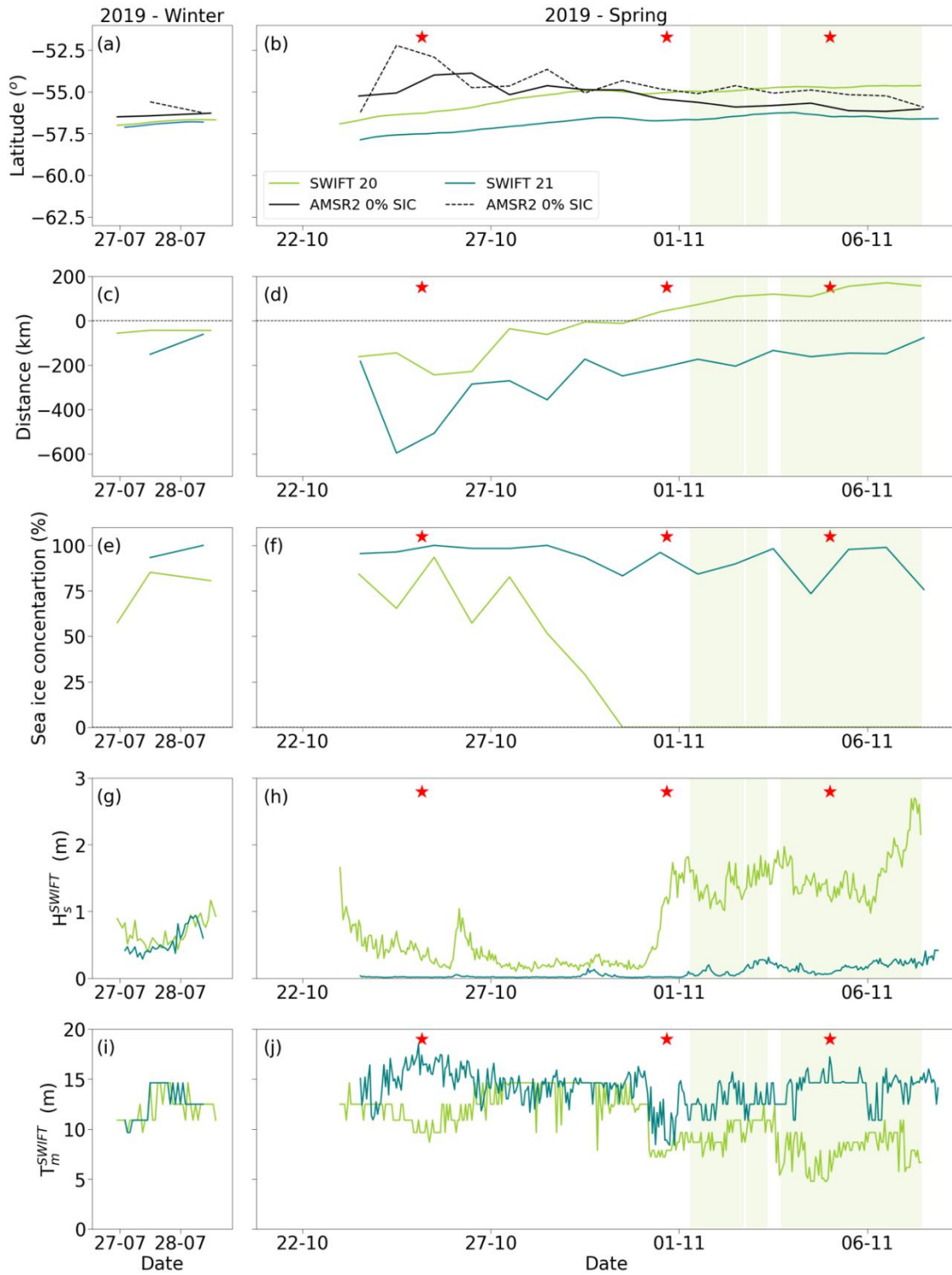


Figure 4.8. (a-b) Time series of the daily latitudinal location of SWIFT 20 (light green) and SWIFT 21 (teal) and the daily ice edge, defined as the $> 0\%$ SIC contour along the same longitude, for each day of the buoys' drift. The black line (dashed) corresponds to SWIFT 20 (SWIFT 21). (c-d) Time series of the relative distance between the buoys and the ice edge, for each day. The horizontal grey line denotes 0 km. (e-f) Time series of the AMSR2 SIC co-located to the buoys' position. (g-h) Time series of the measured significant wave height. (i-j) Time series of the measured mean wave period. The green shaded areas (right-hand-side) denote when SWIFT 20 left the ice cover ship (from on-board cameras). The Winter Cruise is shown on the left, and the Spring Cruise on the right. The red stars denote the spring cyclones previously analysed in Chapter 3.

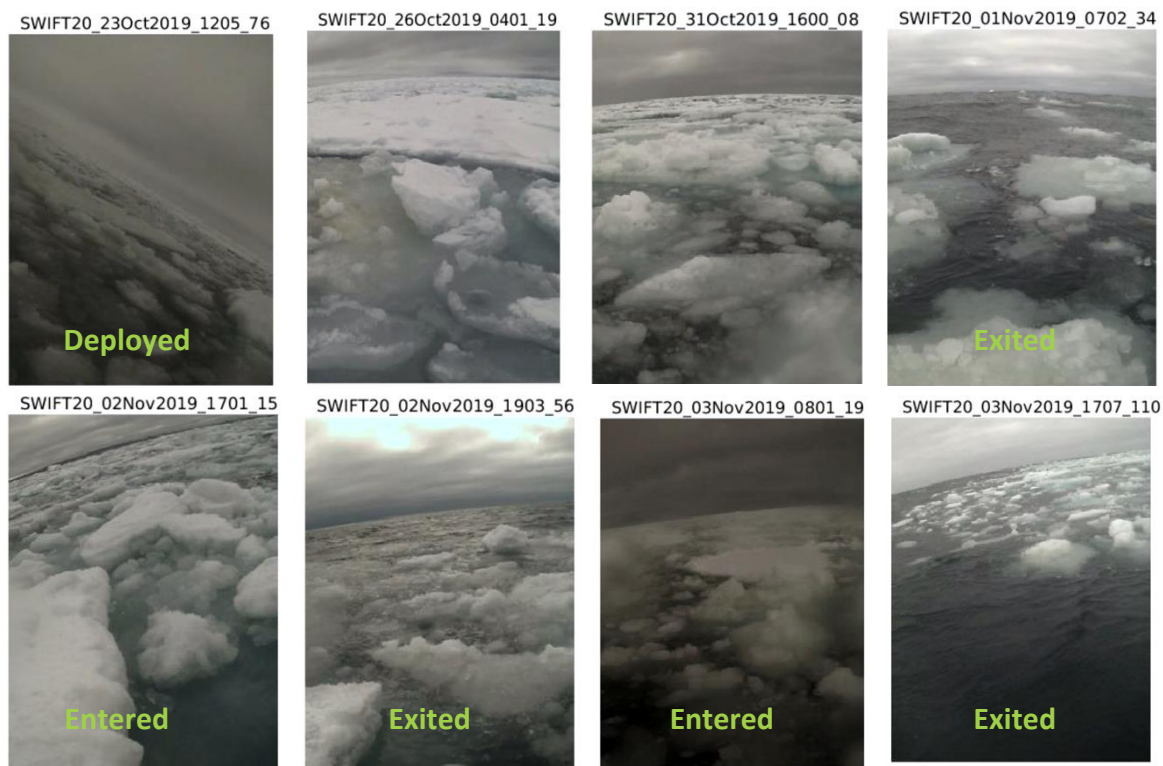


Figure 4.9. Low resolution photographs of sea-ice conditions from SWIFT 20's on-board GoPro camera (≈ 1 m above the surface) during spring – from deployment until it exited the sea-ice cover.

During the Spring cruise, SWIFT 20 was re-deployed on 22 October 2019 at 23:08, and at the same location as its winter deployment (56.6° S, 0.0° E), however this time it was over 100 km south from the ice edge (Figs. 4.7b and 4.8d; Wahlgren et al., 2023). Sea-ice conditions during deployment once again consisted of 1-5 m pancake ice floes with interstitial frazil ice (Wahlgren et al., 2023). SWIFT 21 was deployed on 23 October at 57.54° S, 0.0° W, ≈ 130 km south of SWIFT 20, and in consolidated ice conditions with minimal wave activity (Ryan-Keogh and Vichi, 2022; Wahlgren et al., 2023). Both SWIFT buoys were deployed using the same method as their winter deployments (Ryan-Keogh and Vichi, 2022) and drifted in a north-eastern direction (Fig. 4.7b).

According to its on-board camera, SWIFT 20 left the ice cover on 1 November, nine days into its deployment. It then proceeded to fluctuate in and out of the ice cover until 3 November, when it permanently exited (see Fig. 4.9). This has been additionally highlighted by the light-green shaded regions on the right-hand-side of Fig. 4.8. Subsequently, while the satellite-derived ice edge estimated that SWIFT 20 left the MIZ on 30 October, the mounted GoPro camera indicates the buoy returned twice on 2 and 3 November. Thus, the significant increase in H_S^{SWIFT} , between 31 October and 1 November, occurred while SWIFT 20 was still drifting within the ice cover. SWIFT 21, on the other hand, remained within the MIZ during its deployment (see Fig. 4.10; Wahlgren et al., 2023), and due to its further distance from the ice edge (Fig. 4.8d) surrounded by larger ice floes, it measured less wave energy at its location (Fig. 4.8h; Wahlgren et al., 2023).

Both SWIFT buoys were once again recovered on 7 November 2019, with SWIFT 20 drifting in the open ocean and SWIFT 21 drifting in sea ice consisting of 1-5 m pancakes and frazil ice (Wahlgren et al., 2023; their Fig. S3).



Figure 4.10. Low resolution photographs of sea-ice conditions from SWIFT 21's on-board GoPro camera (≈ 1 m above the surface) during spring – from deployment until it was retrieved.

4.3. Results and discussion

4.3.1. Calibration of the reconstructed significant wave height

Following the methods described in Sect. 4.2.3 and 4.2.4, several different RAOs were estimated for each season, using various draughts that ranged between 7-15.5 m. These were then applied to the measured response spectra in order to estimate the encounter wave energy spectra (Eq. 4.2), and hence, the H_S^{IMU} (Eq. 4.5). All estimated H_S^{IMU} values were calibrated with H_S^{sat} values, as shown in Fig. 4.11a-d. Additionally, for better clarification regarding the most optimal RAO, the estimated H_S^{IMU} values were also compared with those from ERA5 reanalysis, in the open-ocean.

The difference between H_S^{IMU} and H_S^{sat} is measured for the full duration of the expedition by the root mean square error (RMSE) and is shown as a function of the draught. In winter, the error for the initial draught of 7 m is 1.58 m and decreases to 1.34 m with a draught of 15.5 m (Fig. 4.11a). The corresponding correlation coefficient (r) exhibited very similar values for each draught (Fig. 4.11c), with the highest correlation of 0.63 at 15.5 m. The error metrics between the ERA5 significant wave heights and H_S^{IMU} indicated a larger RMSE and lower r values for all draughts. However, both values reach optimal conditions for the draught value of 15.5 m (i.e. minimum RMSE and maximum r). Therefore, using both the satellite altimeter data and ERA5 reanalysis data, the optimal RAO was taken for the 15.5-m draught. However, as shown in Fig. 4.11e, the standard deviation (light-blue shaded region) for all H_S^{IMU} values at each time step is extremely small, indicating little variance between all RAOs.

In spring, as there were only four satellite data values within the spatial and temporal ranges (see Sect. 4.2.4), a correlation is meaningless and only the RMSE is presented for reference. The RMSE for the initial draught of 7 m was 0.93 m, and decreased until a draught of 15.5 m with a minimum error of 0.67 m (Fig. 4.11b). The RMSE for ERA5 data also decreases with larger draughts. Its corresponding correlation coefficients exhibited similar values at ≈ 0.58 for each draught (Fig. 4.11d). For consistency, and given the few available data for calibration, the optimal RAO in spring was computed with the same draught value used in winter. However, similar to that of the Winter Cruise, the standard deviation at each time step is very small, as shown in Fig. 4.11f.

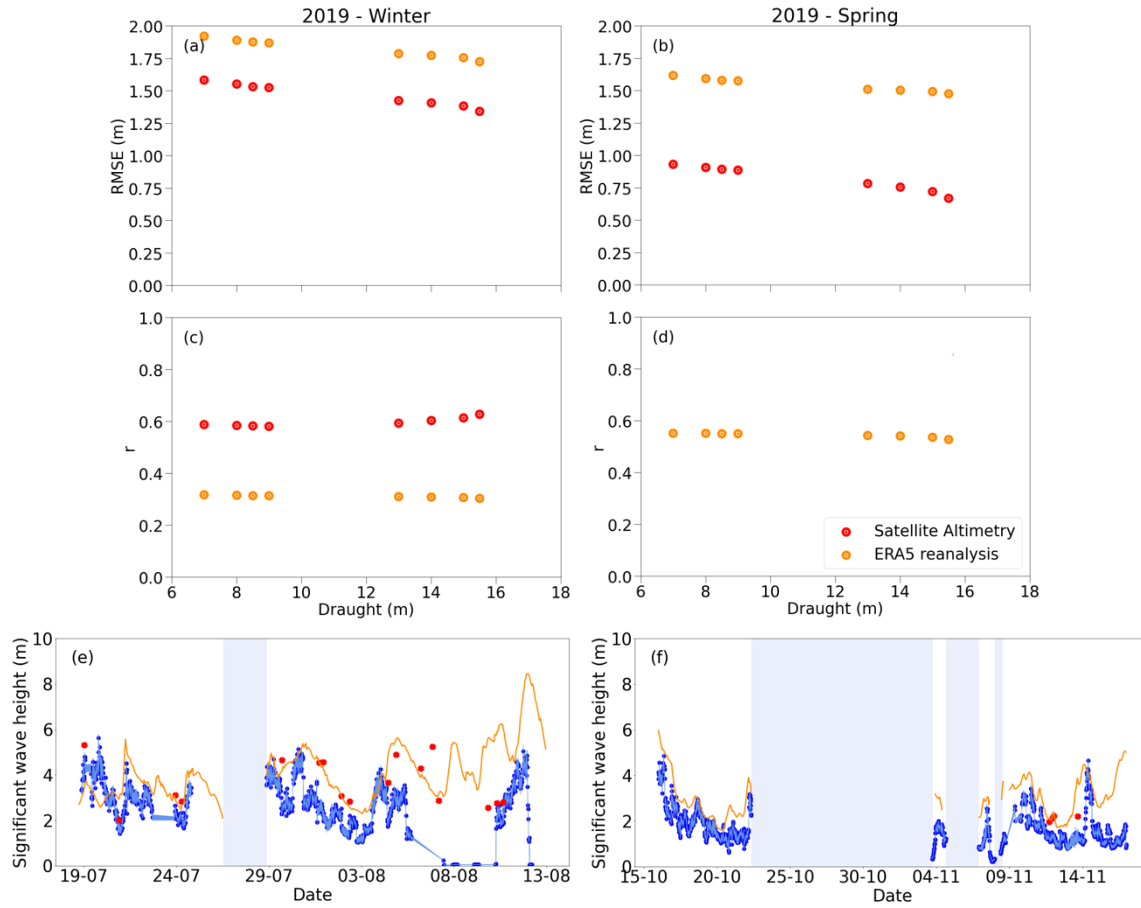


Figure 4.11. (a-b) Root Mean Square Error (RMSE) showing the differences between the significant wave height estimated from the ship's motion response and satellite observations (red points) and ERA5 reanalysis (orange points), as a function of the draught. (c-d) Correlation Coefficient (r) relative to (a-b). (e-f) The time series of the mean reconstructed significant wave height across all computations of the RAO (blue points), at each time step, with the standard deviation range (light-blue). The corresponding satellite and ERA5 data at the ship's location is also shown. The blue shaded regions denote the period when the ship was in ice. The Winter Cruise is shown on the left and the Spring Cruise is shown on the right.

4.3.2. Reconstructed waves parameters in the MIZ

During the Winter Cruise, the S.A. Agulhas II travelled south along the 0° E meridian until $\approx 58^\circ$ S (see Fig. 4.1). During the two days in the ice, a high-pressure cell persisted, with mslp reaching 1020 hPa (Fig. 4.12a). The 2-m air temperatures concurrently decreased from near melting point to $< -15^\circ\text{C}$. The ship then followed a similar route out of the MIZ. During this exit, air temperatures rapidly increased back to near melting point, as the mslp decreased slightly.

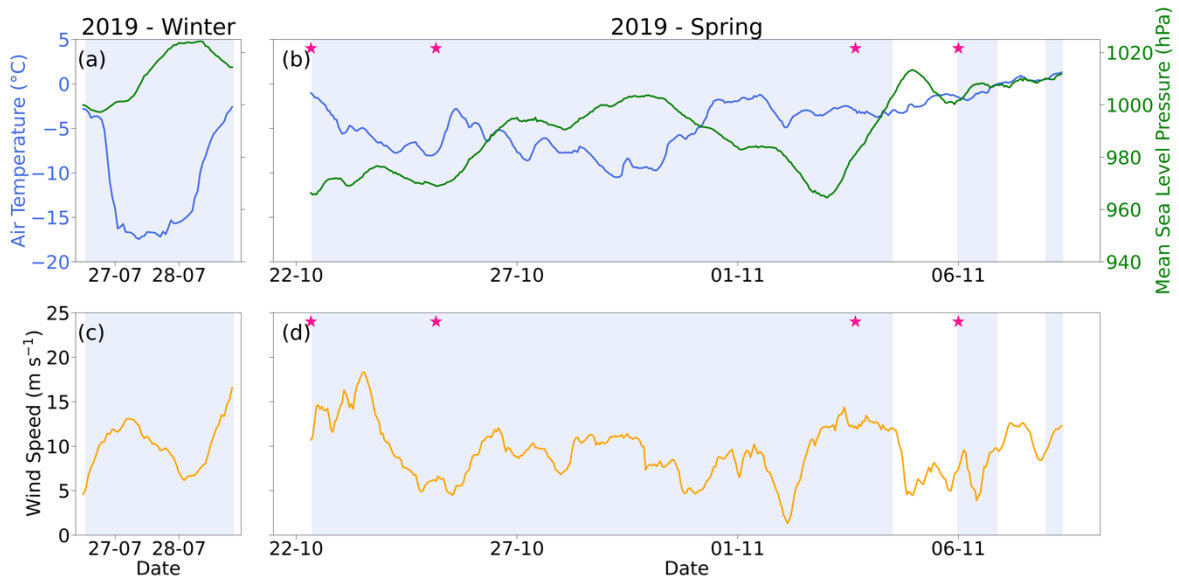


Figure 4.12. (a) Time series of the ERA5 2 m air temperature (blue) and mean sea level pressure (green) at the location of the S.A. Agulhas II, during winter. (b) Same as in (a) but for spring. (c-d) Time series of the corresponding ERA5 10-m wind speeds in each season. The blue shaded areas denote when the ship was in ice (Ryan-Keogh and Vichi, 2022). The pink stars denote when the cyclones, previously analysed in Chapter 3, were closest to the ship.

The winter sea-ice cover was an organised material consisting of several ice types, neatly arranged in space – smaller floes and frazil ice filled the space between the larger ice floes – creating an orderly scene (Tersigni et al., 2023). From image inspection (Fig. 4.13) and ASPeCt on-board thickness observations (Hepworth et al., 2020), the outer region of the MIZ (< 100 km) was characterised by young ice formations, which included grease ice, frazil ice and newly formed and mobile (unconsolidated) pancake-ice floes. Herein, the term ‘mobile’ is used frequently to convey the dynamic nature of unconsolidated sea ice in this region, where ice type is less related to the concentration value (see Chapter 2; Vichi, 2022). Rafting between individual pancake floes were quite apparent (e.g. Fig. 4.13 – left image). Between 100-200 km from the ice edge, sea-ice concentration was 100 % (Fig. 4.14e) and mature pancake-ice floes became the dominant ice type, with interstitial water and/or frazil ice. Consolidated ice conditions were later observed > 200 km from the ice edge (on 27 July) and were often characterised by leads of variable lengths (Tersigni et al., 2023).

The satellite-derived SIC varied greatly from the observed SIC on the first and last day of the analysed period (Fig. 4.14e). On 26 July, the AMSR2 satellite product underestimated the SIC, as the MIZ was observed to rapidly become 100 % covered (Hepworth et al., 2020), while still maintaining ≈ 1 m waves. This is because remote-sensing algorithms tend to underestimate the SIC in regions of thin ice, as well as close to the ice edge (Pang et al., 2018), where ice melt and growth conditions make up a large portion of the sea-ice regime (Agnew and Howell, 2003). On the other hand, the satellite product over-estimated the SIC on 28 July. This may likely have been a result of the rapid increase in air temperatures from -15°C to -5°C, between 27-28 July (Fig. 4.12a), where the onset of ice melt caused high variability in the ice emissivity (Kern et al., 2022). It also may have been a result of the changing winds, which caused a rapid change in the ice-cover geometry.

As the ship initially sailed southwards into the sea-ice cover, the reconstructed H_S^{IMU} rapidly decreased from 1 m to ≈ 0 m in consolidated conditions (Fig. 4.14g). However, while we may have expected the T_m^{IMU} to concurrently increase, the mean period instead decreased from 10 s to 1 s (Fig. 4.14i). The peak-wave period (T_p^{IMU} ; red points) on the other hand does show high values between 17-26 s, in addition to values of ≈ 1 s. This observed phenomenon could potentially be a result of the interactions between the ship and the more consolidated sea-ice conditions on 27 July (Figs. 4.13 and 4.14e), with the ship's voyage through the ice causing fractures. This fracturing in turn may have released energy causing the IMU aboard the ship to register and measure higher frequencies during these conditions. This contrasts with the mobile pancake-ice conditions when the ship entered the MIZ on 26 July (Fig. 4.13). An ad hoc filtering procedure may need to be developed to remove higher frequencies when the significant wave height is below a certain threshold. This would however require a much larger dataset to properly identify the threshold. Overall, the reduction in wave height and increase in high peak periods further from the ice edge, indicates that the ice cover acted as a low-pass filter, with a decreasing low cut-off frequency. In their study, Doble et al. (2015) reported a progressive reduction in significant wave height in the advancing pancake-ice region of the Weddell Sea, with a concurrent increase in the mean-wave periods.

As the ship began its return out of the ice cover, it travelled back through more mobile pancake conditions (Fig. 4.13). Subsequently, the T_m^{IMU} returned to a more realistic range, with values matching those measured by the SWIFT buoys (Fig. 4.14i). Additionally, the H_S^{IMU} rapidly increased to > 2 m. These larger waves can be attributed to the decreased distance from the ice edge as well as the increase in air temperatures (Fig. 4.12a), which together likely caused the reduction in SIC from 100 % to ≈ 70 % (see Fig. 4.13). Here, increased open water and thinner ice would have allowed the open-ocean waves to more freely propagate > 100 km into ice cover. Additionally, as seen in Fig. 4.12c, wind speeds increased as a low-pressure system was approaching the region, which would have caused increased wave energy near the ice edge to propagate into the ice cover (Thomson et al., 2019; Vichi et al., 2019). In contrast to the initial progression into the ice cover, the wave parameters also indicated more variance and scattering as the ship travelled northwards out of the ice cover. This is likely due to the mobile nature of pancake floes and their continuously evolving geometry/shape within the MIZ, as a result of deformation processes, which in turn impacts wave attenuation in the ice cover.



Figure 4.13. Daily photographs of the local sea-ice conditions during the Winter Cruise taken from the S.A. Agulhas II. The dates and time stamps for each photograph are shown.

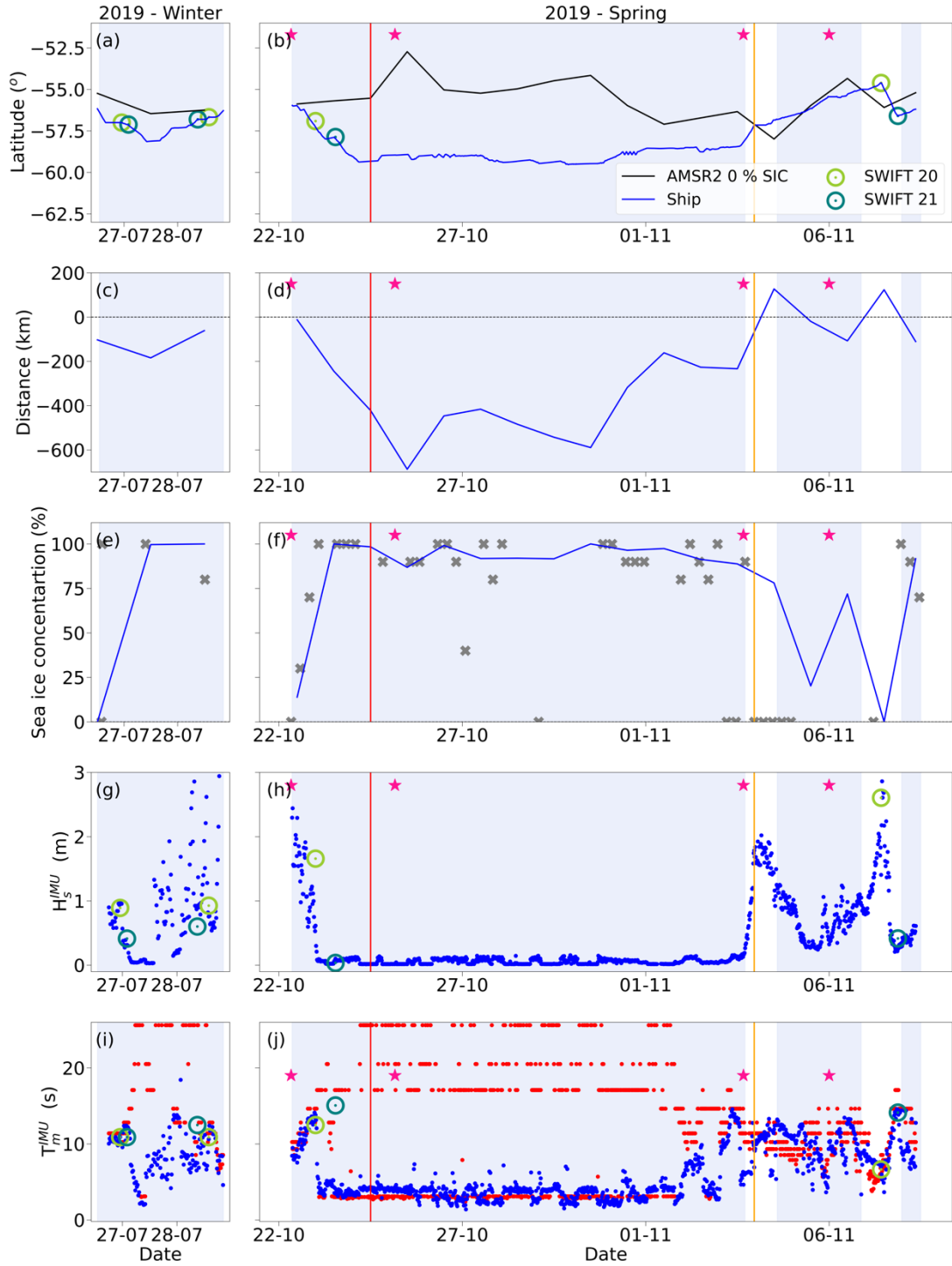


Figure 4.14. (a-b) Time series of the latitudinal location of the S.A Agulhas (blue line) and the daily ice edge, defined as the > 0 % SIC contour (black line), along the same longitude, for each day of the ship's transect. (c-d) Time series of the relative distance between the ship and the ice edge, for each day. The grey dashed line denotes 0 km. (e-f) Time series of the AMSR2 SIC co-located to the ship's position. The grey crosses denote the observed SIC, estimated to the nearest 10 % (Hepworth et al., 2020). (g-h) Time series of the reconstructed significant wave height measured along the ship's transect. (i-j) Time series of the reconstructed mean wave period, with the inclusion of the peak wave period (red points). The light green and teal circles denote the corresponding values measured by SWIFT 20 and SWIFT 21, respectively, during their deployment and retrieval. The blue shaded areas denote when the ship was in the ice cover (Ryan-Keogh and Vichi, 2022). The red vertical line denotes when the ship began travelling east. The orange vertical line denotes when ship travelled back west. The Winter Cruise is shown on the left, and the Spring Cruise on the right. The pink stars denote when the cyclones, previously analysed in Chapter 3, were closest to the ship.

During the Spring Cruise, the ship entered the MIZ on 22 October 2019, at 08:40, during storm conditions (Fig. 4.12b and d) i.e. < 970 hPa mslp, near-melting air temperatures and > 10 m s⁻¹ wind speeds. The ship once again travelled south along 0° E, but this time much deeper into the ice cover ($\approx 59^\circ$ S; Fig. 4.1). On 24 October, the ship then turned to travel eastwards, where it sailed predominantly > 400 km from the ice edge. After reaching 24° E on 4 November, the ship turned back westwards and began its final leg within the ice-edge region of the MIZ (< 100 km from ice edge). The ship finally left the MIZ on 8 November at 08:40, from the same location as its starting point. During this in-ice period, the mslp and 2-m air temperature fluctuated as passing cyclones caused the characteristic decrease in air pressure, while concurrently advecting warm-moist air onto the sea-ice cover (Fig. 4.12b). However, as detailed earlier in Section 3.4.1, the air temperatures in spring appeared to be less related to the local atmospheric conditions, as a result of the seasonal change in the surface heat balance.

The sea-ice cover during spring exhibited a much more variable composition (Tersigni et al., 2023). Grease ice and brash ice were the first ice types observed from the ship, and these conditions persisted for the first 150 km (Fig. 4.15; Tersigni et al, 2023). However, as the ship sailed deeper into the ice (> 150 -300 km), the sea-ice cover became more consistently dominated by first year ice with leads (Tersigni et al., 2023). The remotely sensed SICs from the daily AMSR2 data (Fig. 4.14f) were not too dissimilar from the observed ones, especially during the first five days (22-27 October). As the sea-ice cover progressively broke up and melted, the AMSR2 product, known for its low temporal resolution, encountered limitations in accurately determining the SIC at the locations of the ship. During this southward transect, the reconstructed H_S^{IMU} rapidly decreased from 3 m to ≈ 0 m (Fig. 4.14h), while the T_m^{IMU} increased from 6 s to 15 s (Fig. 4.14j). This concurrent lengthening of the mean-wave period and lengthening of the mean-floe size, as the ship sailed deeper into the ice cover, is an indication of the feedback nature of wave-ice interactions within the MIZ (e.g. Weeks and Ackley, 1986; Shen and Ackley, 1991; Doble and Wadhams, 2006; Meylan et al., 2014; Shen et al., 2001; Rogers et al., 2021).

As the ship proceeded to sail eastwards through the ice cover (from the red vertical lines in Fig. 4.14) primarily at distances > 400 km from the ice edge (Fig. 4.14d), large open-water fractions were initially observed (Tersigni et al., 2023). This may have been a result of the initial cyclone on 22 October, and the subsequent increase in air temperature and wind speed between 22-24 October (Fig. 4.12b and d). However, as first year ice remained the principal ice type along the first half of the eastward transect (0-12° E; 24-30 October), waves were notably attenuated, with H_S^{IMU} persisting at ≈ 0 m (see Section 1.1.2). However, from 23 October, the T_m^{IMU} remained remarkably small with values < 5 s. The peak-wave period (T_p^{IMU} ; red points) on the other hand shows high values between 17-26 s, in addition to values of ≈ 2 -3 s. This recurrence of this phenomenon in consolidated-ice conditions strongly suggests that the ship's on-board IMU was measuring a predominant higher-frequency signal, while navigating through the ice, that is likely not wave-related.

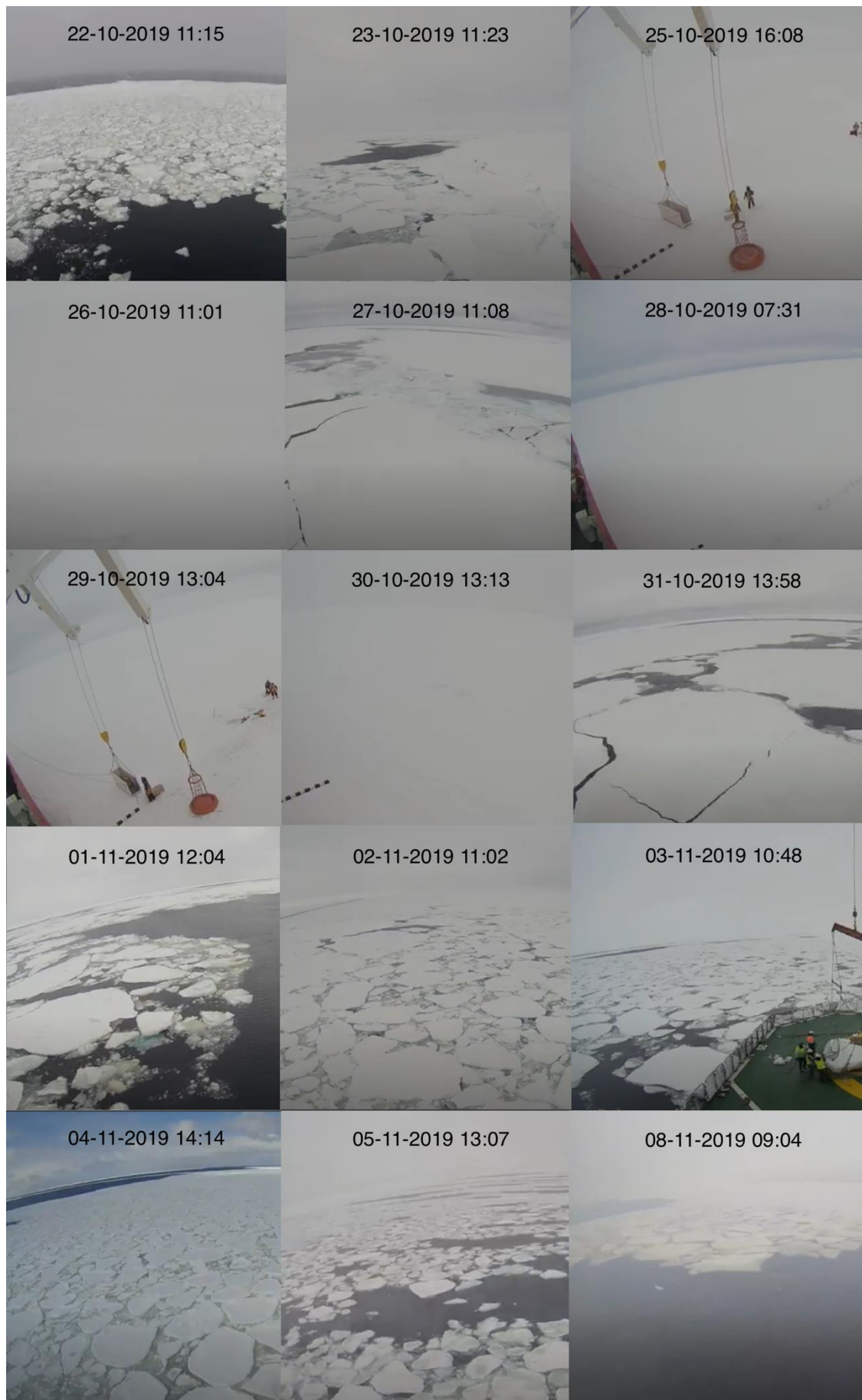


Figure 4.15. Daily photographs of the local sea-ice conditions during the Spring Cruise taken from the S.A. Agulhas II. The dates and time stamps for each photograph are denoted. Note, there are missing days.

Further east, the distance between the ship and the ice edge decreased to ≈ 200 km (Fig. 4.14d), as the ship sailed in a more north-east direction (Fig. 4.1b; Tersigni et al., 2023). The ice type changed to a more scattered combination of various-sized brash ice floes, new pancake formations, occasional leads and bands of open water reaching as wide as ≈ 10 km (Fig. 4.15; Hepworth et al., 2020; Tersigni et al., 2023). These fractured floes and changes in SIC would have been a direct result of the seasonal increase in air temperature (Sect. 3.5.1). However, while the ice cover was highly variable (SIC ranging between 0-100 %; Fig. 4.14f), the reconstructed H_S^{IMU} remained near 0 m until 3 November at 22:30. This can be attributed to the large (> 5 m) broken ice floes (Fig. 4.15), which were likely still able to attenuate the wave energy. After 1 November, the T_m^{IMU} began to increase once again. The ship's encounter with progressively more fragmented and pancake-like ice floes was likely the reason, as they spread out and limited the availability for large open-water fraction for new wind-waves to form. These floes would have consumed the high-frequency waves, resulting in the progression towards longer wavelengths and periods (Shen et al., 2001). Concurrently, these waves would have contributed to the further break-up of the ice cover (Zhang et al., 2015; Sect. 1.1.2).

On 4 November (orange vertical lines in Fig. 4.14), the ship began its final leg westwards through the MIZ, and quickly exited the MIZ during the passage of a cyclone (3 November). The loss of sea ice and increased cyclonic winds over the open ocean together resulted in a rapid rise in H_S^{IMU} , up to ≈ 2 m (Fig. 4.14h). These stormy sea-state conditions quickly subsided as the cyclone passed, and the ship re-entered the MIZ on 5 November. However, the waves still held enough energy to propagate into this region, comprised of smaller pancake-shaped brash ice (< 5 m) amid bands of open ocean (Fig. 4.15; Hepworth et al., 2020). Thus, while the waves did decay in the ice, they were not fully attenuated, and the H_S^{IMU} only reached a minimum of 0.5 m. As the ship once again left the ice cover – to retrieve SWIFT 20 – the H_S^{IMU} subsequently reached a maximum of 3 m, while the T_m^{IMU} concurrently decreased to 5 s. This is attributed to the last cyclone which passed near the ice edge on 6 November, causing rough sea-state conditions. However, immediately after this peak in wave height, the reconstructed H_S^{IMU} notably decreased, as the ship sailed southwards back into the ice to retrieve SWIFT 21.

4.3.3. Comparison between SWIFT buoys and reconstructed parameters

During winter, both SWIFT buoys were deployed in and continued to drift in pancake-ice conditions (Wahlgren et al., 2023), at distance < 200 km from the sea-ice edge (Fig. 4.8c). Therefore, because both SWIFT 20 and SWIFT 21 did not leave the classic winter MIZ conditions, they exhibited elevated significant wave heights (0.5 to 1 m) in comparison to the reconstructed estimates, from the ship's motion response, between 27-28 July (Fig. 4.16a and d). Therefore, compared to the ship, which sailed further into the ice cover, the waves at these distances were not fully attenuated. In comparison, Alberello et al., (2020) reported significant wave heights in the range of 0 and ≈ 1.5 m for a WIIOS buoy (see buoy B2 in Sect. 2.2.2.2) that drifted in pancake-ice conditions during an intense storm. These measurements were taken at distances of 100-200 km from the ice edge during a period of storm quiescence. However, their overall trends are consistent, noting that the SWIFT buoys do not show multiple peaks due to their lower temporal resolution of one hour.

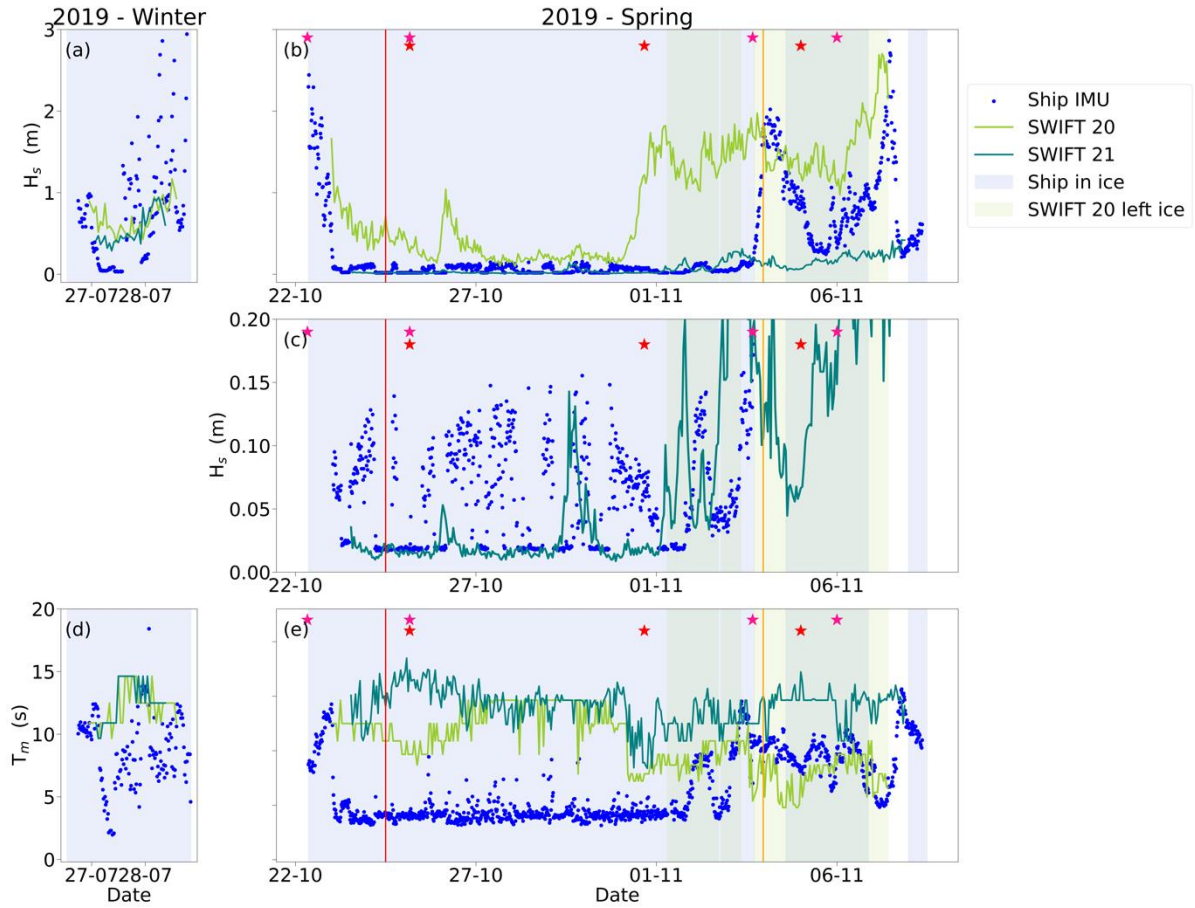


Figure 4.16. Time series of the significant wave height during **(a)** the 2019 Winter Cruise and **(b)** during the Spring Cruise, as estimated from the ship’s motion response (blue points), and measured from SWIFT 20 (light-green line) and SWIFT 21 (teal line). **(c)** Same as in **(b)** but showing a closer view of reconstructed wave height, and of SWIFT 21’s wave height. **(d-e)** Time series of the mean wave period during the Winter Cruise and Spring Cruise, respectively. The blue shaded areas denote when the ship was in the ice cover (Ryan-Keogh and Vichi, 2022). The green shaded areas (right-hand-side) denote when SWIFT 20 left the ice cover ship (from on-board cameras). The red vertical line denotes when the ship began travelling east. The orange vertical line denotes when ship travelled back west. The pink (red) stars denote when the cyclones, previously analysed in Chapter 3, were closest to the ship (SWIFT buoys).

During spring, SWIFT 20 once again was deployed in and continued to drift in pancake-ice conditions, where it initially exhibited a large reduction in its measured H_s^{SWIFT} from > 1.5 m to 0.5 m (Fig. 4.8h; Wahlgren et al., 2023). This is attributed to the return of calmer atmospheric and wave conditions following the passage of the cyclone on 22 November (see Fig. 4.12b). On the other hand, SWIFT 21 was deployed in consolidated-ice conditions, and therefore measured negligible waves-in-ice activity (Fig. 4.16c; Wahlgren et al., 2023). On 25 October, an ice-landing cyclone passed over the ice edge, advecting warm mid-latitude oceanic air onto the sea-ice cover, and generating large open-ocean waves ≈ 6 -7 m in height (Fig. 4.17b-c). These storm-generated waves maintained considerable energy, and reached SWIFT 20 on 26 October, at ≈ 200 km from the ice edge (Fig. 4.8d). However, further south at the locations of SWIFT 21 and the ship, these waves were almost fully attenuated (Fig. 4.16c-d). This is attributed to the dominance of more compact and thicker ice at their locations (Fig. 4.9; Tersigni et al., 2023). Following this cyclone, quiescent atmospheric conditions persisted until 30 October, and little waves-in-ice activity was measured by the SWIFT buoys and the ship.

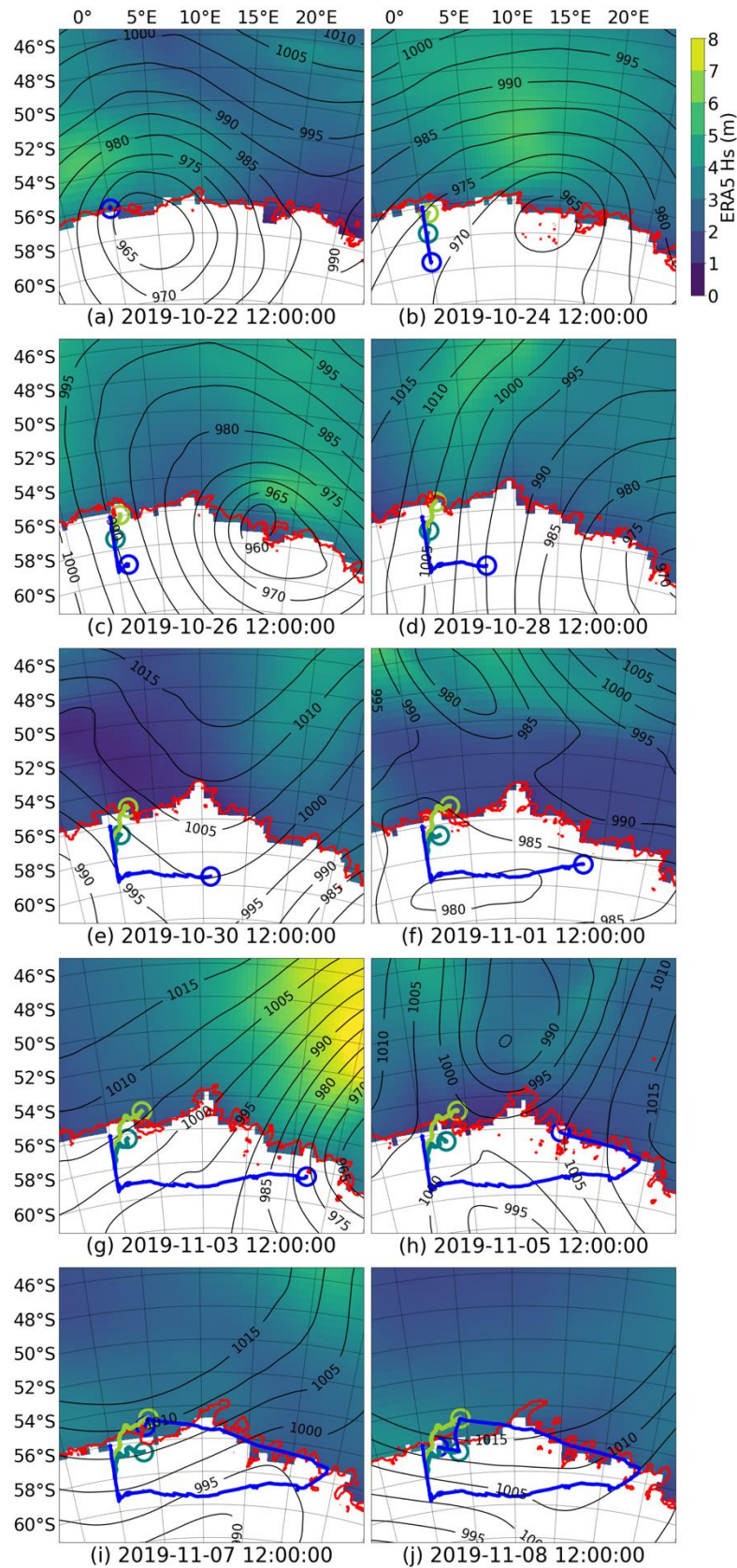


Figure 4.17. Evolution of the environmental conditions during the period when the S.A. Agulhas II travelled through the sea-ice cover. Each panel illustrates the ERA5 mean sea level pressure (black lines, interval 5 hPa), ERA5 significant wave height (coloured shading, interval 1 m), and the AMRS2 0 % SIC – the ice edge (red line). The ship track (blue line), and SWIFT 20 (light-green line) and SWIFT 21 (teal line) are shown from their starting position to their current position, denoted by the circles. All data are at 12:00 GMT.

On 31 October, another cyclone passed over the ice edge (red star; Fig. 4.16, Fig. 4.17e-f). Both SWIFT buoys recorded an increase in higher-frequency waves at their respective locations, since their H_S^{SWIFT} increased and their T_m^{SWIFT} decreased (Fig. 4.16b and d). However, as SWIFT 20 was drifting within the ice-edge region, comprised of small floes and brash ice (Fig. 4.9), the wave energy was far less attenuated (Wahlgren et al., 2023). Thus, its H_S^{SWIFT} reached > 1.5 m, while its T_m^{SWIFT} decreased to ≈ 6 s. On the other hand, since SWIFT 21 drifted further from the sea-ice edge (> 200 km; Fig. 4.8d), where it was surrounded by larger ice floes (Fig. 4.10; Wahlgren et al., 2023), it instead exhibited a small but notable increase in its wave height. Collectively, these findings suggest that the seasonal increase in air temperature, as discussed in Sect. 3.5.1, caused the ice cover to melt and break up. Consequently, the ice cover likely became more susceptible to further break-up events, initiated by storm-generated waves. As the ship was located $> 15^\circ$ east (> 850 km) from the SWIFT buoys (Fig. 4.17f-g), the same cyclone only impacted the ice cover surrounding the ship two days later (Fig. 4.16b). Furthermore, the cyclone had intensified as it reached the ship's position, with ERA5 significant wave heights reaching 8 m in the open ocean. Thus, the reconstructed H_S^{IMU} only peaked on 4 November. In relation to Vichi et al. (2019) and Chapter 2 (see Sect. 2.4), this further emphasises how synoptic cyclones in the Southern Ocean reshape the ice-covered ocean, in a clockwise direction. Their large influence is a result of the rapid variations that these cyclones cause on sea ice, when they cross the MIZ (e.g. Vichi et al., 2019).

As the ship then sailed back westwards, the ice cover became progressively more fragmented, with increased open-water fractions (Fig. 4.15). Subsequently, once the ice cover had broken up, its properties and behaviours would have been qualitatively different from those of the large-consolidated floes (Dumont et al., 2011), that were observed earlier in the expedition. Thus, the waves that propagated into the MIZ were weakly attenuated, as seen by the larger H_S^{IMU} estimates (Fig. 4.16b), and by SWIFT 21's gradual increase in H_S^{SWIFT} . These are consistent with results from Ardhuin et al. (2020), who showed that when the “break-up front” is displaced southwards and ice is fractured in proximity to a buoy (or ship), the attenuation is much weaker. Herein, we confirm that this is in response to the seasonal break-up and melting of the ice cover, where large floes are unable to reform after each cyclonic event, and their associated storm-generated waves. In agreement with Kousal et al. (2020) and Curry et al. (1995), we can see that this creates a positive feedback through the reduction in albedo and the presence of more open water, where waves can more easily propagate and further break-up the ice cover (Kohout et al., 2014; Ardhuin et al., 2020; Passerotti et al., 2022).

4.4. Summary and conclusions

This study employed two methods to explore the interactions between waves and sea ice within the Antarctic MIZ, during winter and spring of 2019. These included the reconstruction of wave parameters – significant wave height and mean wave period – from ship motion using the wave-buoy analogy (WBA), which were then supplemented with data obtained from two drifting SWIFT buoys (Thomson et al., 2023; Wahlgren et al., 2023). While the former has been more commonly applied to estimate open-ocean sea states (e.g. Nielsen and Jensen, 2011; Newman, 2018; Nielsen et al., 2021; Nelli et al., 2023), it provided the unique opportunity to measure waves in different types of ice conditions, and was not location restricted. The derived parameters were first calibrated with satellite altimetry and ERA5 reanalysis data in the open ocean before being analysed in the sea-ice cover.

Subsequently, this study is the first to implement the WBA technique in the Antarctic MIZ. The results show that the reconstructed wave parameters corresponded well with the SWIFT-buoy measurements at the time of their deployment (Fig. 4.16), and varied in response to the evolving MIZ, the distance from the ice edge, and to the passage of the cyclones.

During both seasons, larger significant wave heights and lower mean wave periods were measured by both the ship and SWIFT buoys when they were located closer to the ice edge (Figs. 4.8g-h and 4.14g-h). However, wave penetration was observed to reach > 200 km from the ice edge, when the ice cover was dominated by mobile floes. Conversely, when compact ice conditions prevailed at similar distances from the ice edge, the wave heights were almost negligible (Fig. 4.16a-b), and hence, the wave energy was almost completely attenuated. These phenomena are in agreement with prior literature (e.g. Meylan and Masson, 2006; Shen et al., 2001; Horvat et al., 2020; Alberello et al., 2022), which report a progression towards longer wavelengths and the presence of larger floes further from the sea-ice edge. Together, these highlight the interconnected relationship between waves and sea ice where: (i) waves decay at a rate governed by the nature of sea ice, and (ii) the propagation of waves into the ice cover leaves behind a wake of broken floes, which is known to promote sea-ice growth during winter when the total energy balance is dominated by losses (Li et al., 2021), and accelerate ice melt in spring when surface heat balance changes (Fig. 4.12b). However, unexpectedly, the ship's motion response during both seasons revealed higher-frequency waves (smaller T_m^{IMU} ; Fig. 4.14i-j), during the periods when the ship sailed through more consolidated-ice conditions. These conditions were observed at distances > 200 km from the ice edge in winter, and > 400 km during spring (Tersigni et al., 2023). While this was speculated to be a result of the interactions between the fracturing of the sea ice by the ship, it is important to note that delving deeper into this relationship is beyond the scope of this study and warrants further exploration and research.

The longer Spring Cruise provided an opportunity for a more in-depth analysis of the effects of passing cyclones on the seasonal MIZ. The results, during this melt season, indicate that there was a clear relationship with the passage of cyclones and the occurrence of increased wave energy in the interior of the MIZ (Fig. 4.17b and d). These storm-generated waves made the ice cover more mobile, and re-established the separate mosaic of individual floes that were previously cemented together (Fig. 4.15). This would have allowed the ice cover to become more susceptible for drift and deformation by winds and ocean currents (see Sect. 3.5.1). Furthermore, the thermal effect of the seasonal increase in air temperatures, likely caused the initial flexure and cracking of the sea ice (Sect. 3.5.1). Subsequently, the propagation of waves further contributed to the break-up and ultimate retreat of the ice cover, as large floes would not always have been able to reform after each storm (see Fig. 4.15). Together, this would have created a positive feedback through the reduction in albedo and the presence of more open water, where wave could freely propagate and break up the ice cover.

The wave data presented in this study further contribute to our understanding of the intricate air-ice interactions within the Antarctic MIZ, especially in during the spring-melt season. It is imperative that such high-resolution measurements become a routine on expeditions to the Antarctic during all seasons. It will contribute to a more complete sampling of sea ice in different regions across the Antarctic region, and thus provide necessary information to further elucidate the dynamics driving the Antarctic seasonal cycle and improve both models and remote sensing products.

Chapter 5

5. Key findings and outlook

In this thesis, sea-ice dynamics, within the Antarctic marginal ice zone (MIZ), were investigated in response to wind and wave forcing, with a particular focus on the effects of ice-landing cyclones. A large portion of the Southern Ocean, up to 4 % of the Earth's surface during winter (NOAA, 2020), is covered by seasonal sea ice in the Antarctic MIZ. This forms the foundation of a significantly complex system, as a result of a wide range of ongoing dynamic and thermodynamic processes. However, this region has seldom been directly sampled in winter due to its inaccessibility. Consequently, prediction models in the Southern Ocean are prone to significant inaccuracies. With this in mind, this thesis combines in-situ measurements from multiple buoy arrays; satellite remote sensing data; atmospheric reanalysis data; and reconstructed wave parameters, using the wave-buoy analogy (WBA), to describe in detail the interconnected and dynamical processes of sea ice at both the synoptic and sub-daily time scales, as well as across different seasons and regions in the Antarctic. The physical properties of sea ice differ significantly with changes in seasons and regions, suggesting that different processes govern the transfer of momentum from the atmosphere and ocean to the ice cover. An improved understanding of the dynamical processes between the atmosphere, ocean and sea ice in the MIZ still requires a significant amount of attention. This is necessary for enhancing our capabilities to monitor and predict its formation and retreat. Additionally, it ensures the more accurate parametrisation of the momentum transfers, especially since the Antarctic sea-ice extent has shown new minima in the past few years.

The overall aim of this thesis was therefore to enhance our knowledge of the dynamic response of sea ice to extra-tropical cyclones, and to waves propagating into the ice cover. By leveraging on autonomous buoy data deployed during multiple expeditions in the Atlantic and Indian Oceans sector, this thesis allowed for a better description and quantification of sea-ice dynamics within the MIZ, with respect to the initial research question: *“How does atmospheric and oceanic (waves and currents) forcing, and local ice conditions, affect sea-ice dynamics in the Antarctic MIZ, across several temporal and spatial scales?”*. Furthermore, insights from deficiencies and limitations in the research will aid future developments in sea-ice dynamics modelling and improvements of satellite observations' reliability.

Chapter 1 established the importance of Antarctic sea ice to our global climate. Acting as a natural barrier between the surface ocean and the below-freezing atmosphere, its existence regulates the transfer of wind and

wave energy, and constituents such as atmospheric gases, dust and aerosols. However, while these phenomena exhibit variability based on the distribution of various sea-ice characteristics, including type, thickness and extent, previous literature have demonstrated that's these exchange processes also regulate and govern the evolution of the Antarctic sea-ice cover (e.g. Eayrs and other, 2019a; Maksym, 2019; Matear et al., 2015; Pezza et al., 2012; Turner et al., 2017; Messori et al., 2017; Schemm, 2018; Schlosser et al., 2018; Vichi et al., 2019; Alberello et al., 2020). This is because extra-tropical and polar cyclones transport energy to the Antarctic, and in turn act to modulate the sea ice, both dynamically and thermodynamically. These dynamic processes involve the drift and deformation of sea ice by winds, waves, ocean currents, and internal ice stresses (Eq. 1.1), with the principal factor being the wind forcing. Therefore, as highlighted by **Chapter 1**, when examining sea ice and its characteristics it is crucial to understand its development in context with other components of the polar environment. However, majority of our scientific knowledge on sea-ice dynamics in the Antarctic has been derived from pack-ice conditions, which have a dissimilar spatial morphology to the MIZ. This is largely due to the scarcity of in-situ sea-ice observations, which have limited our understanding of these interactions, especially the impacts that extra-tropical cyclones have on the ice cover when crossing the Antarctic MIZ (Vichi et al., 2019).

Although the MIZ is widely recognised as the region where these ocean, atmosphere, and sea-ice processes are intricately interconnected, and where cyclonic events produce the most rapid and extreme changes (e.g. Andreas et al., 2010; Zhang et al., 2015; Vichi et al., 2019, Alberello et al., 2019); the considerable knowledge gap regarding the impacts of ice-landing cyclones represents a significant limitation in our current understanding of the complex dynamics at play in this region. Only recently has it been shown that the effects of winter extra-tropical cyclones extend into the ice-covered region (e.g. Vichi et al., 2019). Considering these limitations, **Chapter 2**, in response to the first and third objectives (Sect. 1.3), comprehensively investigated this transfer of momentum from the atmosphere to the ice cover. This examination utilised data from a four-month-long drift of a buoy that was deployed on a pancake-ice floe, within the Indian Ocean sector of the Southern Ocean, during the 2017 winter sea-ice expansion. The main findings indicated that surface winds, particularly the meridional winds that are associated with polar cyclones, were the main drivers of floe's drifts, and that they exerted a dominant physical control on ice drift in regions of > 80 % SIC, and at distances > 200 km from the satellite-derived ice edge. This chapter revealed that, in addition to this significant synoptic response of the sea ice, there was a strong inertial signature in the drift data, indicating a plausible correlation to the presence of polar cyclones. Subsequently, it was suggested that, although the mobility of sea ice is assumed to decrease in austral winter due to consolidation, the high mobility of the ice floe was maintained by heterogonous ice conditions like the ones observed at deployment. Therefore, it was found that the ice-tethered buoy did not leave the Antarctic MIZ for the entirety of its four-month drift despite going through multiple conditions of advance and retreat of the sea-ice edge. The quantitative results of this investigation subsequently raised questions around the capability of the concentration-based MIZ definition (15-80 % SIC) for describing sea-ice type in the Antarctic, which contributed to its revised definition proposed by Vichi (2022). This is due to the maintenance of the highly dynamic nature of the MIZ even when high SICs were observed from satellite records. In connection with Vichi et al. (2019), this was attributed to the intricate interactions between sea ice, polar cyclones and storm-generated waves, as outlined in **Chapter 1**.

Building upon these findings, **Chapter 3**, in response to the first four objectives (Sect. 1.3), was designed to further explore the mechanisms driving sea-ice dynamics within the Antarctic MIZ. In this chapter, two ensembles of drifting ice buoys, deployed in the north-eastern Weddell Sea region, were analysed with the aim of characterising the dynamics driving both sea-ice drift and deformation during the winter growth and the spring retreat seasons of 2019. The winter data, akin to the findings from **Chapter 2**, indicated a dynamic response of the sea ice to wind forcing, during the first few days of deployment. As such, the quantitative results revealed a dominant physical control of wind forcing on sea-ice drift at the synoptic scale. However, in contrast to the 2017 ice floe, the role of the ocean circulation was more evident. The ice floes in 2019 were closer to the core of the ACC, which was able to provide a steady source of significant energy to the sea ice. The subsequent enhancement of the wind-driven ice drift by the eastwards flowing ACC, led to a highly energetic and mobile ice cover that was characterised by free-drift conditions but without inertial oscillations in response to passing cyclones. This marked a stark contrast to the phenomena observed in **Chapter 2**, and was ascribed to the different regional sector and its relationship with the ACC. The spring measurements, right after deployment, initially suggest more pronounced internal ice stresses. Ice drift remained loosely related to wind forcing at the synoptic scale as observed for the prior winter ice cover. However, as increased air temperatures and wave activity caused the consolidated ice to melt and break-up into smaller ice floes, the drift became dominated by the initiation of inertial oscillations of sea ice, again in response to cyclonic winds as found in **Chapter 2**. In turn, this suggested that internal ice stresses began to reduce, and the fractured ice floes progressed to a more free-drift motion on the ocean's surface. Following this, **Chapter 3** revealed that the window of consolidated conditions, in which Antarctic sea ice is more likely to present stronger internal stresses, is limited to the September-October period that we were unable to sample.

Chapter 3 examined the deformation of sea ice during the measured seasons, again in relation to the passage of cyclones. The results demonstrated that, in the absence of significant internal ice stresses, sea-ice deformation during winter remained strongly related to the large-scale atmospheric forcing. In contrast, the counterintuitively less wind-driven ice-ocean system in spring – characterised by the presence of inertial oscillations – led to the deformation exhibiting a strong de-coupling to large-scale atmospheric forcing in the frequency domain. The energy spectrum of deformation was high at higher frequencies where wind energy decreased, despite our expectation that increasingly broken floes would be more susceptible to wind forcing, as seen by the winter ice cover. The role of the ocean and especially of sub-mesoscale eddies was therefore invoked, indicating how air-sea ice-ocean processes are interrelated in Antarctic sea ice. These eddies may have been the energy source contributing to the shape of the spring deformation spectra at the higher frequencies. Ultimately, a comparative study of existing deformation datasets from different regions around Antarctica, indicated that Antarctic sea-ice deformation is determined by the mechanics and type of the sea ice, i.e. the rheology, and the balance between external atmospheric and oceanic forcing. Thus, expanding on the **Chapter 1** and **2** findings, **Chapter 3** further highlights the need to understand the dynamics of sea ice in the context of the local marine environment. This observation arises from the demonstrated fact that the ice cover's response to external forcing is predominantly governed by seasonality (growth or melt) and location, and hence the rheology of sea ice.

Given the role played by the propagation of ocean waves in understanding the ice response to the synoptic atmosphere, **Chapter 4**, in response to the fifth objective (Sect. 1.3), delved into a deeper analysis of wave-ice

interactions by employing a new technique based on the ship using the Wave Buoy Analogy (WBA; Nielsen et al., 2021; Nelli et al., 2023). This investigation employed both waves-in-ice buoys and the WBA to identify the significant wave heights and mean wave periods within the Antarctic MIZ, in conjunction with sea-ice features observations during the 2019 winter and spring seasons. The results confirmed the expectations from theory, that wave-ice interactions are mutual (e.g. Herman, 2017; Bennetts et al., 2015; Squire and Montiel, 2016; Squire, 2020; Alberello et al., 2021; Montiel et al., 2022): waves alter sea-ice properties through dynamic and thermodynamics processes, while the waves are concurrently attenuated at a rate governed by the nature of sea ice. This chapter further confirmed that these interactions lead to the progression towards longer wavelengths, and the formation of larger ice floes. Nevertheless, wave penetration was still observed to reach > 200 km from the ice edge. This is indicative that these open-ocean waves have enough energy to propagate hundreds of kilometres into the ice cover, which agrees with previous literature (e.g. Stopa et al., 2018; Vichi et al., 2019; Kohout et al., 2014; Kohout et al., 2020).

The presence of these propagating waves is known to determine ice type and foster the growth of pancake ice during the austral winter months (Doble and Wadhams, 2006; Petrich and Eicken, 2017; Wadhams et al., 2018; Alberello et al., 2019). Consequently, with the northwards expansion of freezing air temperatures occurring during the autumn and winter months, the sea-ice cover concurrently extends northwards under the influence of solar, atmospheric and oceanic forcing. Given the understanding of the interconnected relationship between waves and sea ice, **Chapter 4** substantiates the suggestion from **Chapter 2** that instead of a local transition from smaller pancake ice floes into a consolidated frozen pack ice, as previously reported by Doble and Wadhams (2006), there is rather a northwards movement of a frontal band of unconsolidated sea ice. In turn, this leads to the maintenance of the highly dynamic nature of the MIZ even at high ice concentrations. Furthermore, **Chapter 4** demonstrated that during the spring months, the same wave-ice interactions contributed to the acceleration of sea-ice melt as air temperatures increased. This thermal effect would have initiated the cracking and thinning of the ice cover, through lateral and bottom ice melt. As successive cyclones then passed over the ice cover, their storm-generated waves further facilitated the break-up of the ice cover, as large ice floes were able to reform after each cyclone. Consequently, as the ice cover became more mobile and dynamic during the progression of the 2019 spring season, the ice floes became increasingly more susceptible for drift and deformation by winds and ocean currents. This connection is notably demonstrated by **Chapter 3**, where the initial rise in drift kinematics of the 2019 Spring buoys corresponded with (i) the dates of the second and third cyclones (31 October and 5 November), (ii) when air temperatures began oscillating near melting point, and (iii) when increase in waves-in-ice activity was measured by both the SWIFT buoys and the reconstructed wave parameters. Therefore, this study showed there is a clear connection with the role that waves play on the evolution of the MIZ extent, and the seasonal cycle of the ice cover – in response to the seasonally changing air temperatures.

During the course of this thesis, several knowledge gaps and potential opportunities have emerged. As a summary, the most noteworthy findings, derived from each aspect of the investigation, are outlined below.

In combination, **Chapters 2-4** proposed that, together with the seasonal changes in air temperatures, the propagation of waves affect the local dynamics of ice floes, and how they respond to wind and oceanic forcing at both the synoptic and inertial scale. Pancake-ice floes therefore remain extremely mobile and in free drift if there

is adequate wave energy preventing them from freezing together. Conversely, in regions further from the ice edge, consolidated ice incurs higher internal stresses. This hinders its mobility and response to external forcing. However, as shown in **Chapter 4**, several ice types can co-exist simultaneously, highlighting the complexity of ice-ocean-atmosphere interactions. Additionally, **Chapter 3** indicated that during three different winter deployments, governed by pancake-ice conditions, deformation rates were largely varied for similar spatial scales. This was attributed to the varying waves-in-ice conditions occurring during the drift of the three winter-buoy arrays. Subsequently, this led to a dissimilar sea-ice rheology and a varied resistance to deformation by winds and ocean currents. This showcases the imperative to develop more accurate rheology descriptions that account for the varied ice types present in different seasons and regions around Antarctica. This need is particularly true for the MIZ, since majority of our scientific knowledge on sea-ice dynamics has historically been derived from pack-ice conditions. This is crucial for improving the representation of sea-ice dynamics in the MIZ, where ice is in free drift and under the influence of cyclone induced inertial motion, and waves that can extend up to > 200 km from the ice edge.

In **Chapters 2 and 3**, wind forcing was confirmed to be the primary driver of ice drift, with the momentum transfer from winds at the lower (multi-day) frequencies being identified as the primary effect of atmospheric forcing and the initiation of inertial oscillations of sea ice as the secondary effect. The secondary effect greatly influences ice drift at shorter timescales and causes deviations of the ice drift from a straight-line path. However, it remains unclear how periods of cyclonic activity (5-20 days) may drive the generation of inertial oscillations. This uncertainty arises from instances where power intensifications at the inertial frequency occurred outside of the dates of higher cyclonic activity. Consequently, it was suggested that the penetration of storm-generated waves may also have contributed in initiating the observed inertial oscillations in the ice drift. Alternatively, these waves could have facilitated in allowing the underlying geostrophic currents to keep the weaker oscillations during periods of quiescence by breaking the consolidated ice into smaller floes, or by maintaining a more fluid sea-ice type throughout the extended Antarctic MIZ. Therefore, considering that inertial oscillations arise from the combined ice-ocean system, it is imperative to investigate the role played by the geostrophic currents, in conjunction with examining the impact of atmospheric forcing i.e. cyclonic events. As highlighted by **Chapter 3**, the lack of oceanic observations in the Antarctic MIZ has and will limit our understanding of the highly-interconnected drivers of sea-ice dynamics. This leads to the recommendation for in-situ observations of underlying currents, or better robust methods to determine under-ice geostrophic currents from remote sensing platforms.

Unlike the lack of ocean-currents observations, the amount of research on wave-ice interactions, in the Antarctic MIZ, has grown significantly over the past few years (see Sect. 4.1). However, it is still clear that the understanding of these interactions is incomplete. Therefore, the continued collection of waves-in-ice data is imperative for understanding sea-ice characteristics, and hence, how they concurrently attenuate the wave energy. While there are several approaches for this data collection, **Chapter 4's** results indicated that the WBA is a valuable source of data in regions where measurements based on buoy arrays or satellite observations are challenging, i.e. the polar latitudes. This is particularly noteworthy since on-board meteorological and sea-ice observations can be concurrently acquired. While this technique can be intended as alternative, this study suggests that, by relying on the limited and publicly available information, this method can be applied to previous Antarctic

sea-ice expeditions where ship motion data in sea ice are available. This approach can yield an increased number of datasets of the basic wave parameters, such as the significant wave height and mean wave period. These datasets can contribute to the validation of satellite observations and aid in the development of models in the MIZ. However, it is important to highlight that the uncharacteristically small mean wave periods, during the 2019 winter and spring first-year ice conditions, will need to be further explored. This phenomenon could potentially indicate a new source of high-frequency activity within the sea-ice cover itself, possibly resulting from release of pressure during ice fracturing.

Based on the collective findings, the present concentration-based definitions of the MIZ being a narrow translational and temporary region bordering the open ocean is not in agreement with the comprehensive analysis provided in the thesis. While this definition allows for the large-scale description of the sea-ice zone on synoptic timescales, it does not directly capture or represent wave-ice or wind-ice coupled interactions. There is an expectation that when sea ice attains 100 % coverage from space it becomes consolidated and subject to strong internal stresses. However, in **Chapter 2**, pancake-frazil ice conditions were maintained throughout the four-months of the ice-tethered buoy, even during 100 % SIC, as derived from satellite remote sensing products. This is because the highly dynamic nature of the MIZ is maintained despite the high sea-ice concentrations observed from space. This is due to the complex interactions between sea ice, polar cyclones and storm-generated waves. Therefore, the concentration-based MIZ definition, where the SICs range between 15-80 %, is inadequate to describe the sea-ice type in the Antarctic MIZ.

This thesis stresses the necessity for a more comprehensive definition of the MIZ that better integrates the interconnected processes among the atmosphere, ocean, and the sea-ice cover. The observations presented here demonstrate how intense polar cyclones govern sea-ice dynamics in the MIZ, thereby influencing the configuration of the MIZ itself. However, it also reveals the limited availability of data around Antarctica, including buoy data and oceanic data. Consequently, there is a likelihood that we are likely still not observing all interconnected processes pertaining to sea-ice dynamics in the Antarctic MIZ. This limitation hinders our ability to make definitive assertions about sea-ice rheology, the ice cover's response to external forcing, and subsequently a more accurate representation of the Antarctic MIZ. This is crucial for present and future models and for improving global climate predictions.

Chapter 6

Appendix A

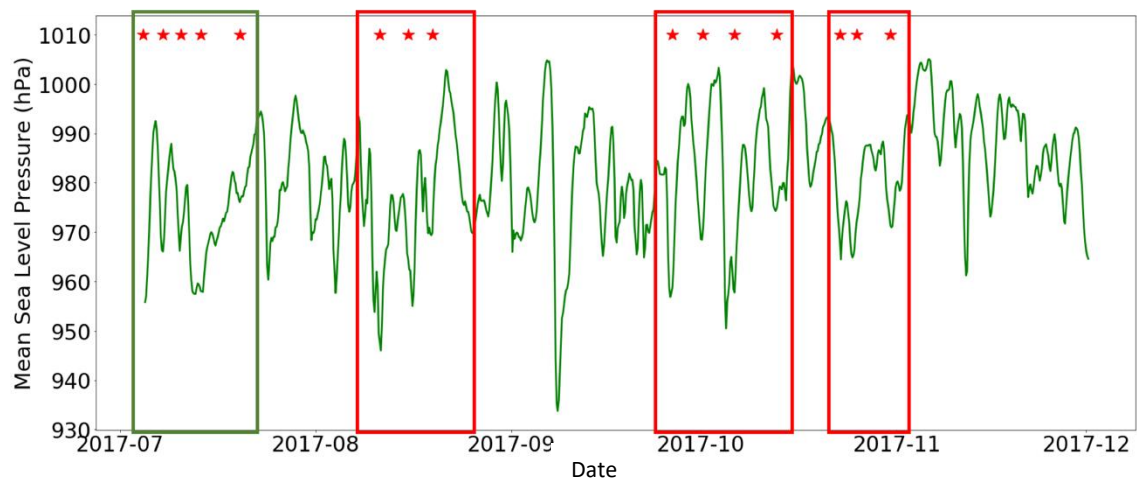
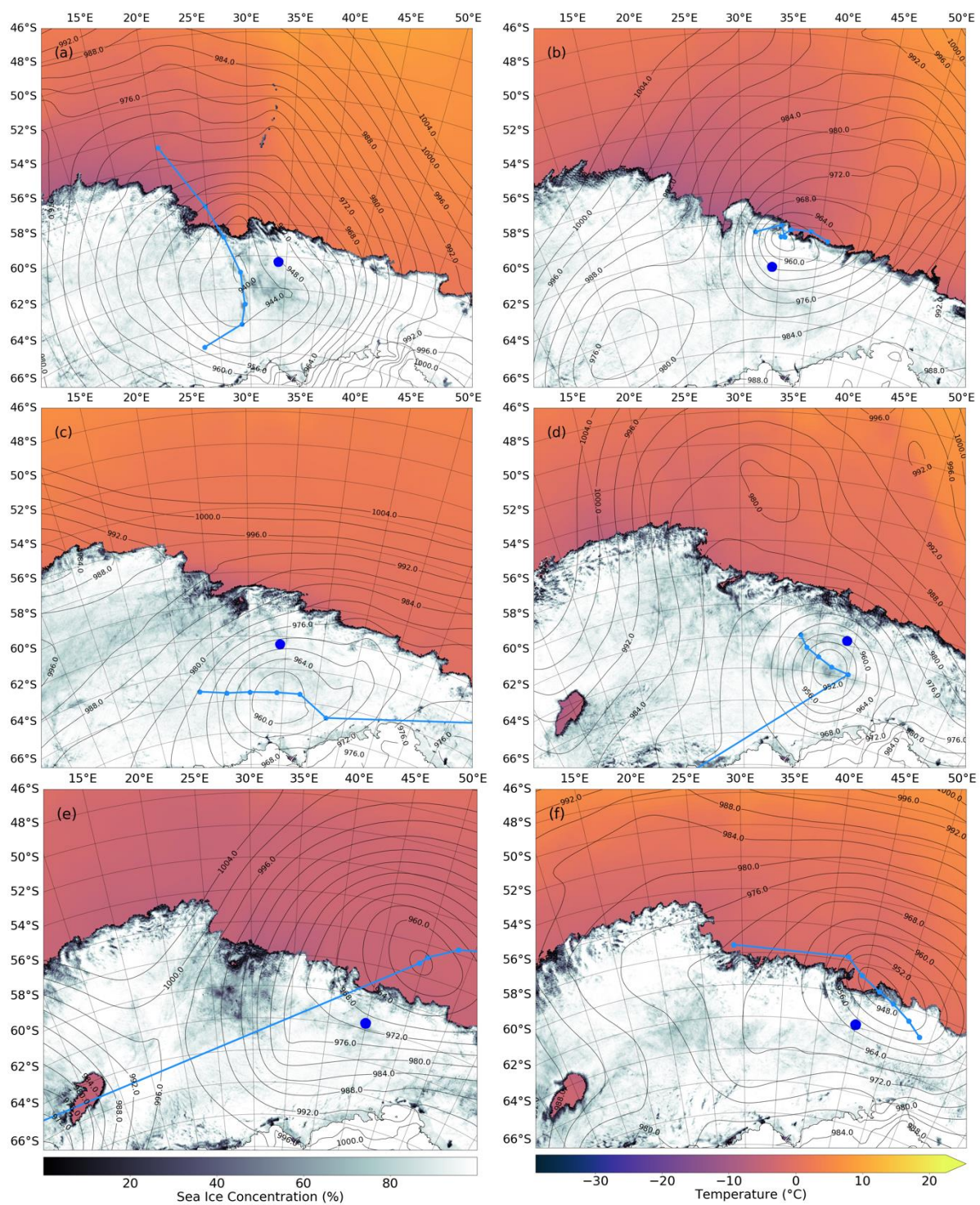


Figure A1. Time series of the mean sea level pressure at the location of the buoy. The red star symbols indicate the time when the storm cores were closest to the buoys' location. The red boxes denote the three sets of cyclones analysed in Table 2.1. The green box denotes the five storms previously examined by Vichi and others (2019) and Alberello and others (2020).



Continued ...

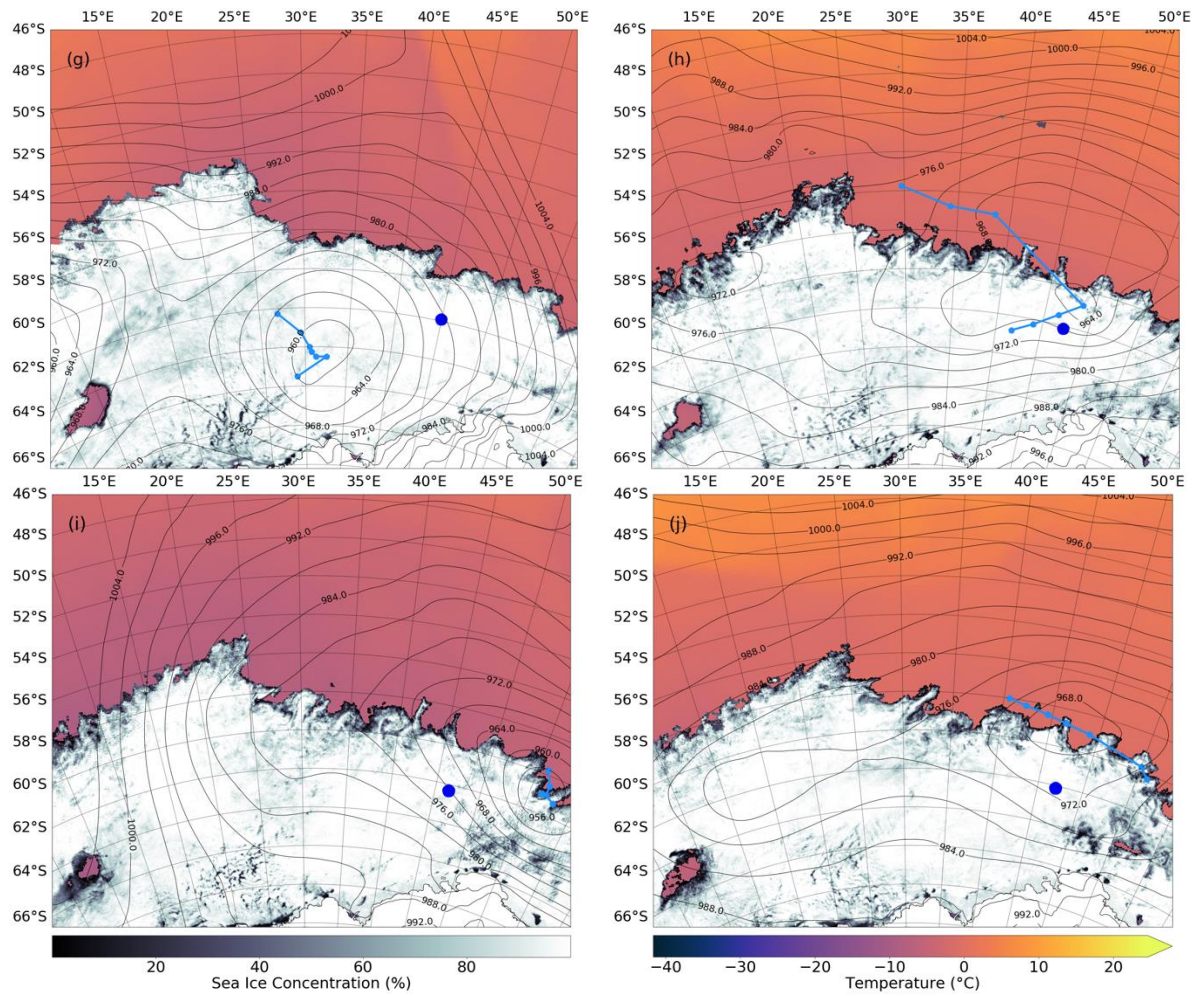


Figure A2. Maps of the ERA5 mean sea level pressure (mslp) and 2-m temperature fields of when each cyclone was closest to the buoy – **(a)** Cyclone A1 on 11 August at 04:00, **(b)** Cyclone A2 on 15 August at 16:00, **(c)** Cyclone A3 on 19 August at 12:00, **(d)** Cyclone B1 on 26 September at 12:00, **(e)** Cyclone B2 on 1 October at 08:00, **(f)** Cyclone B3 on 6 October at 08:00, **(g)** Cyclone B4 on 12 October at 12:00, **(h)** Cyclone C1 on 23 October at 00:00, **(i)** Cyclone C2 on 25 October at 16:00, and **(j)** Cyclone C3 on 31 October at 00:00. The grey shadings show the AMSR2 ice concentration for the corresponding day. The dark-blue circle denotes the buoy for the corresponding time of when each cyclone was closest to it. The light-blue line denotes the trajectory (in a north-west to south-east or south-west to north-east migration) of each cyclone core from 12 hours before to 12 hour after each cyclone was closest to the buoy. The light-blue markers denote the four-hourly position of each cyclone's core.

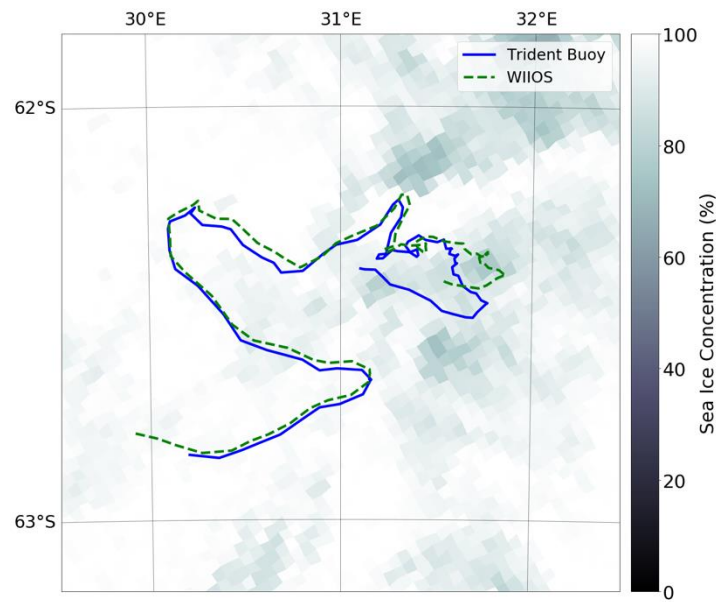


Figure A3. The trajectories of the trident buoy (blue line) and the WIiOS (green line). The WIiOS was deployed at 12:00 on 4 July and drifted until 16:00 on 19 July. The Trident buoy was deployed 12 hours after the WIiOS at 00:00 on 5 July and is only shown until 16:00 on 19 July.

Appendix B

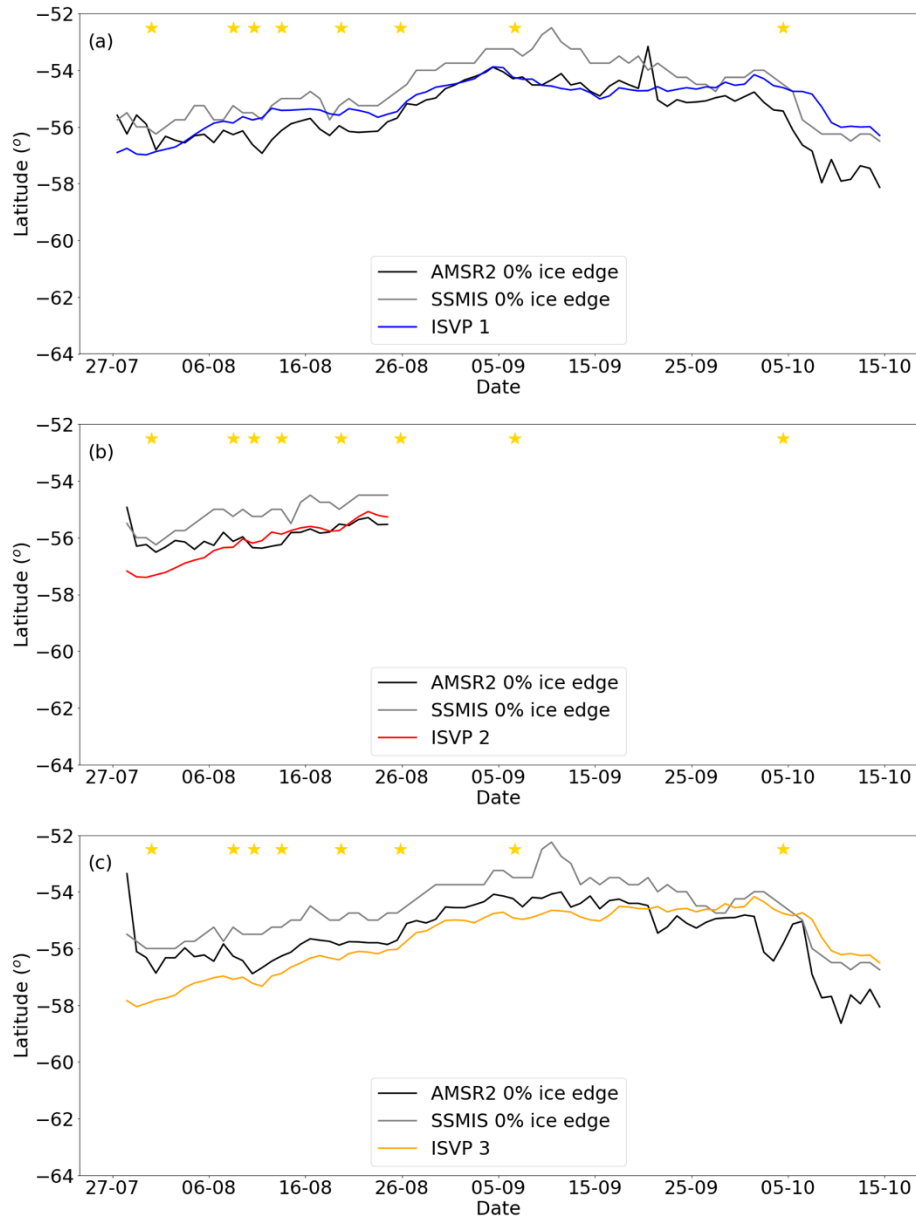


Figure B1. (a) Latitudinal location of ISVP 1 and the sea ice edge, defined as the > 0 % sea ice concentration contour, along the same longitude, for each day of the buoys' trajectories. The black line denotes the latitude of the AMSR2 ice edge. The grey line is from the SSMIS sea ice product. (b) Same as in (a) but for ISVP 2. (c) Same as in (a) but for ISVP 3. The yellow star symbols indicate the time when the winter cyclones were closest to the buoys' location.

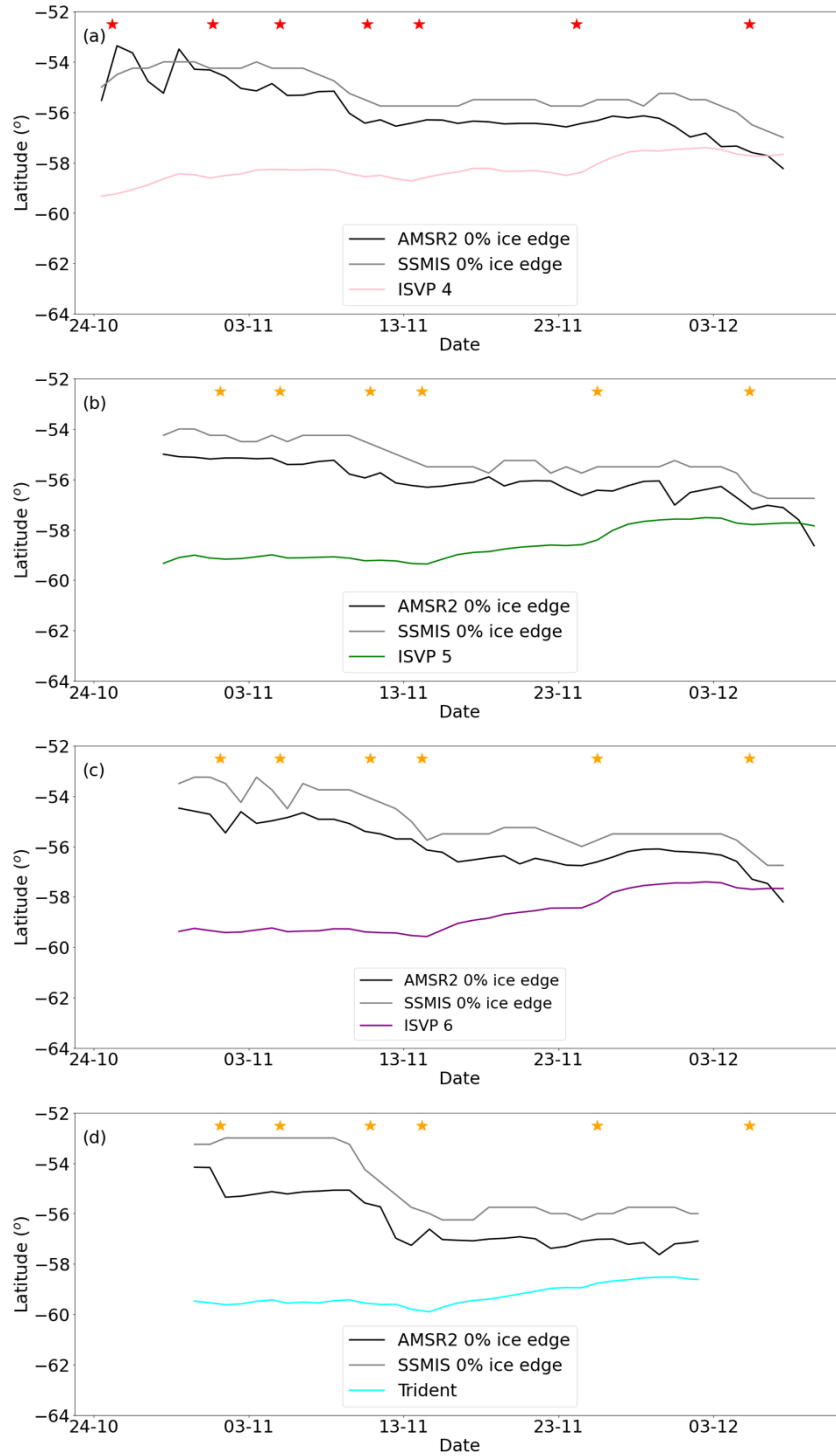


Figure B2. (a) Latitudinal location of ISVP 4 and the sea ice edge, defined as the $> 0\%$ sea ice concentration contour, along the same longitude, for each day of the buoys' trajectories. The black line denotes the latitude of the AMSR2 ice edge. The grey line is from the SSMIS sea ice product. (b) Same as in (a) but for ISVP 5. (c) Same as in (a) but for ISVP 6. (d) Same as in (a) but for the Trident. The red stars denote when the spring cyclones were closest to ISVP 4. The orange stars denote when the same spring cyclones were closest to the main spring cluster $\approx 5^\circ$ east of ISVP 4.

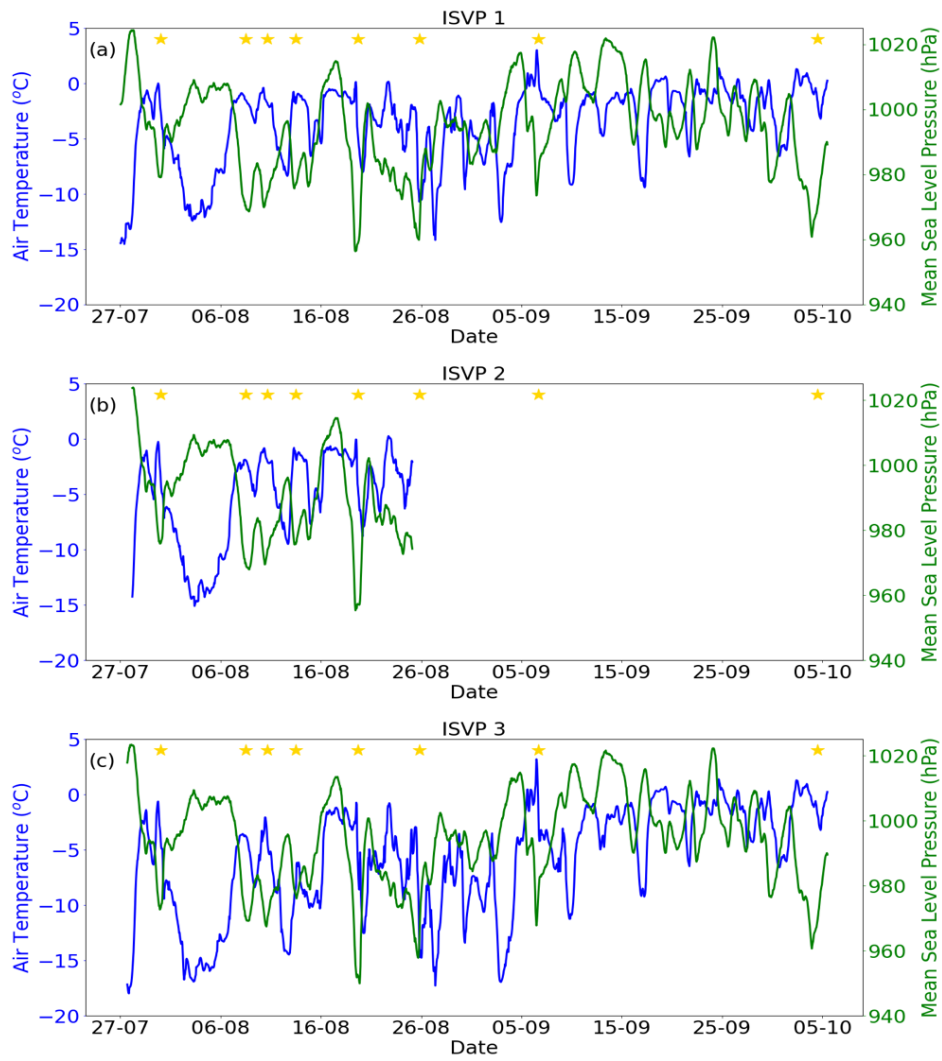


Figure B3. (a) Time series of the ERA5 2 m air temperature (blue) and mean sea level pressure (green) at the location of ISVP 1, during winter. (b) Same as in (a) but for ISVP 2. (c) Same as in (a) but for ISVP 3. The yellow stars denote when the cyclones were closest to the Winter buoys' location.

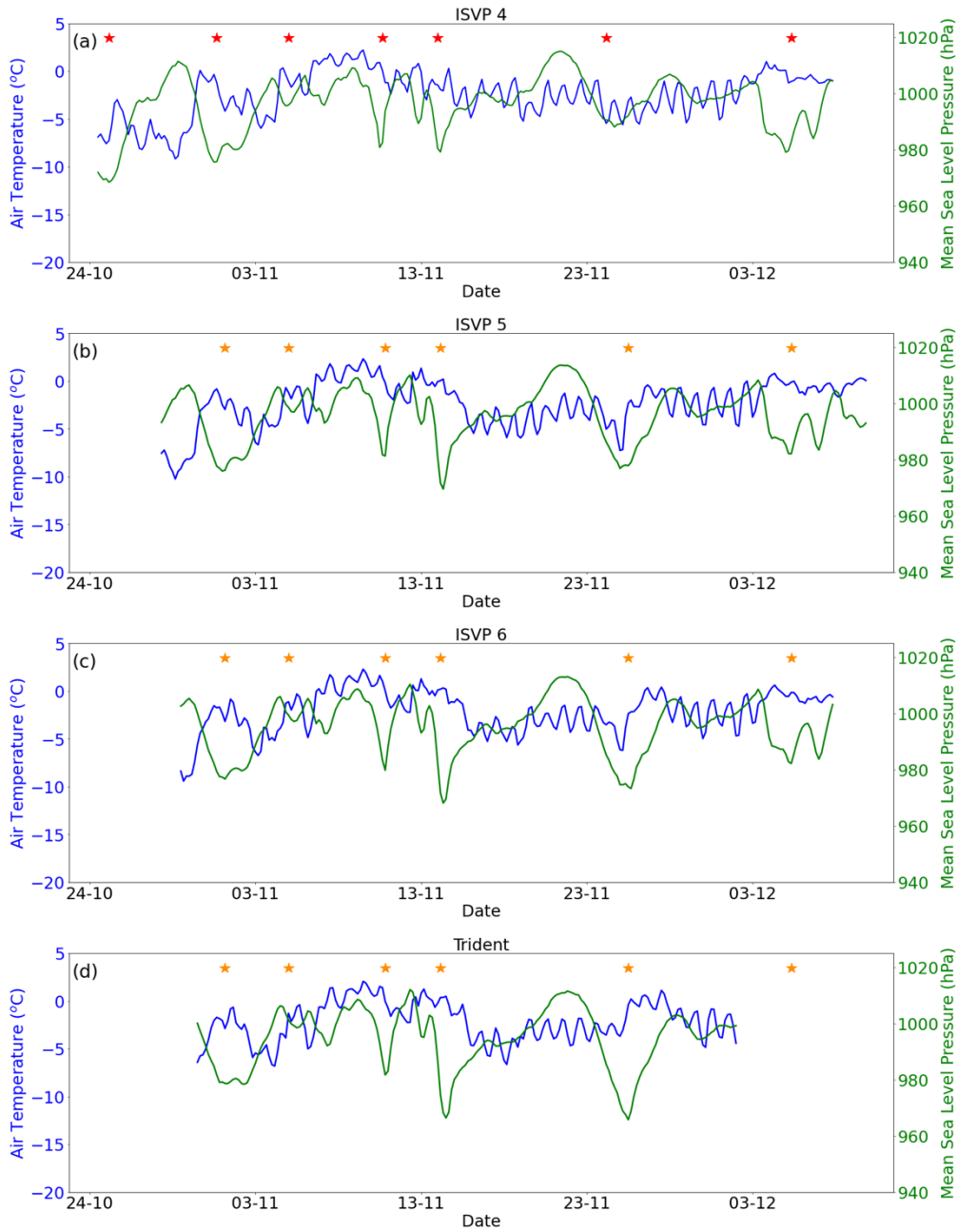


Figure B4. (a) Time series of the ERA5 2 m temperature (blue) and mean sea level pressure (green) at the location of ISVP 4, during spring. (b) Same as in (a) but for ISVP 5. (c) Same as in (a) but for ISVP 6. (d) Same as in (a) but for the Trident. The red stars denote when the cyclones were closest to the eastwards main cluster. The orange stars denote when the same cyclones were closest to ISVP 4 as it was deployed $\approx 5^\circ$ west of the main spring cluster.

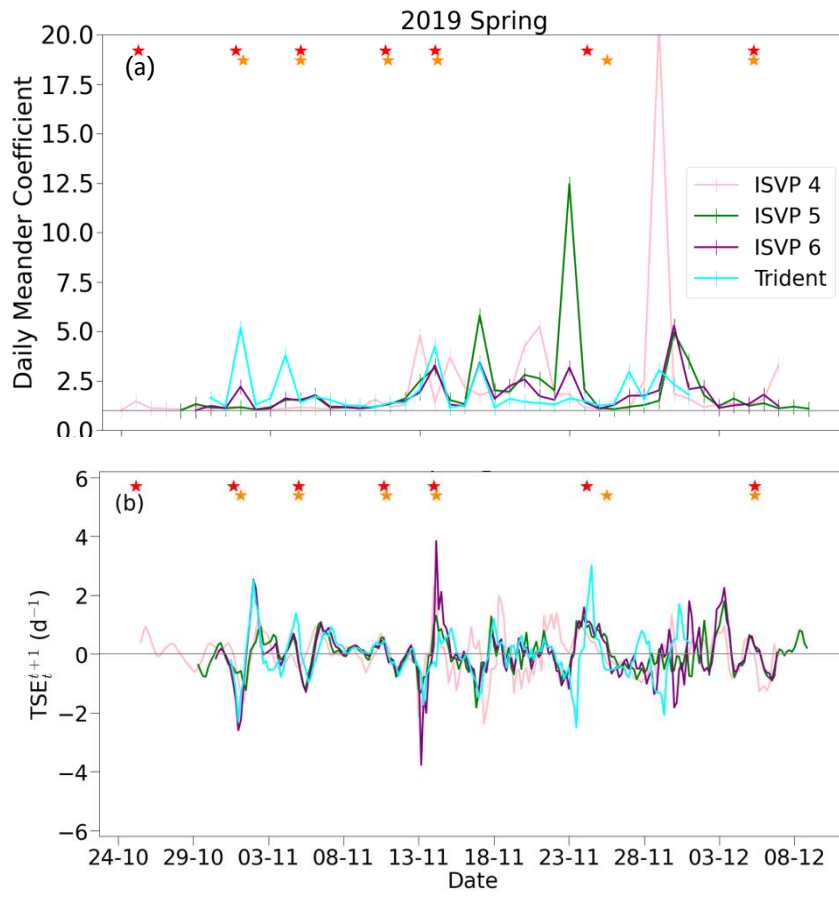


Figure B5. (a) Time series of the daily discrete meander coefficient computed using the Trident's four-hourly time step for all Spring buoys. The vertical markers are indicative of the daily time interval and the horizontal grey line at $M=1$ denotes straight-line drift. **(b)** Time series of the TSE for the duration of the buoys' deployment computed using the Trident's four-hourly time step for all Spring buoys. The red stars denote when the spring cyclones were closest to ISVP 4. The orange stars denote when the same spring cyclones were closest to the main spring cluster, $\approx 5^\circ$ east of ISVP 4.

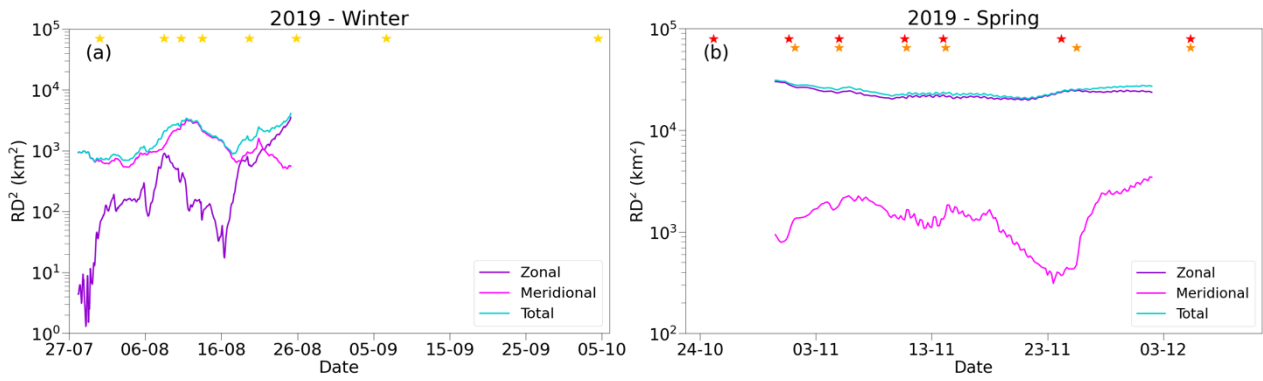


Figure B6. (a) Time series of the relative (two-particle) dispersion of the Winter buoys when all buoys were transmitting (≈ 30 days for both seasons). (b) Same as (a) but for the Spring buoys. It must be noted that the scaling of the y axis is different between the seasons. The yellow stars denote when the winter cyclones were closest to the Winter buoys' location. The red stars denote when the spring cyclones were closest to ISVP 4. The orange stars denote when the same spring cyclones were closest to the main spring cluster, $\approx 5^\circ$ east of ISVP 4.

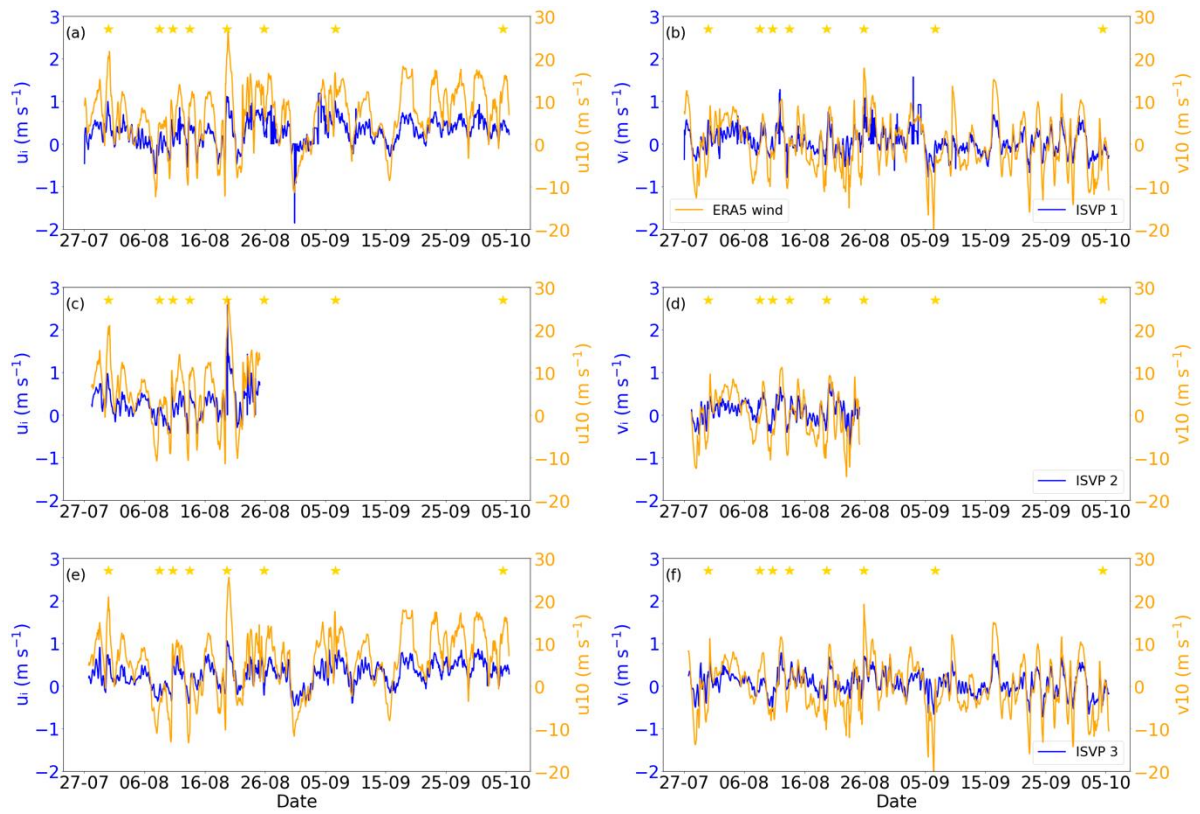


Figure B7. Velocity components of the Winter buoys (on the left axis) and 10 m wind from the ERA5 reanalyses (on the right axis): the zonal and meridional component of ISVP 1 **(a-b)**, ISVP 2 **(c-d)**, and of ISVP 3 **(e-f)**. The yellow stars denote when the winter cyclones were closest to the Winter buoys' location.

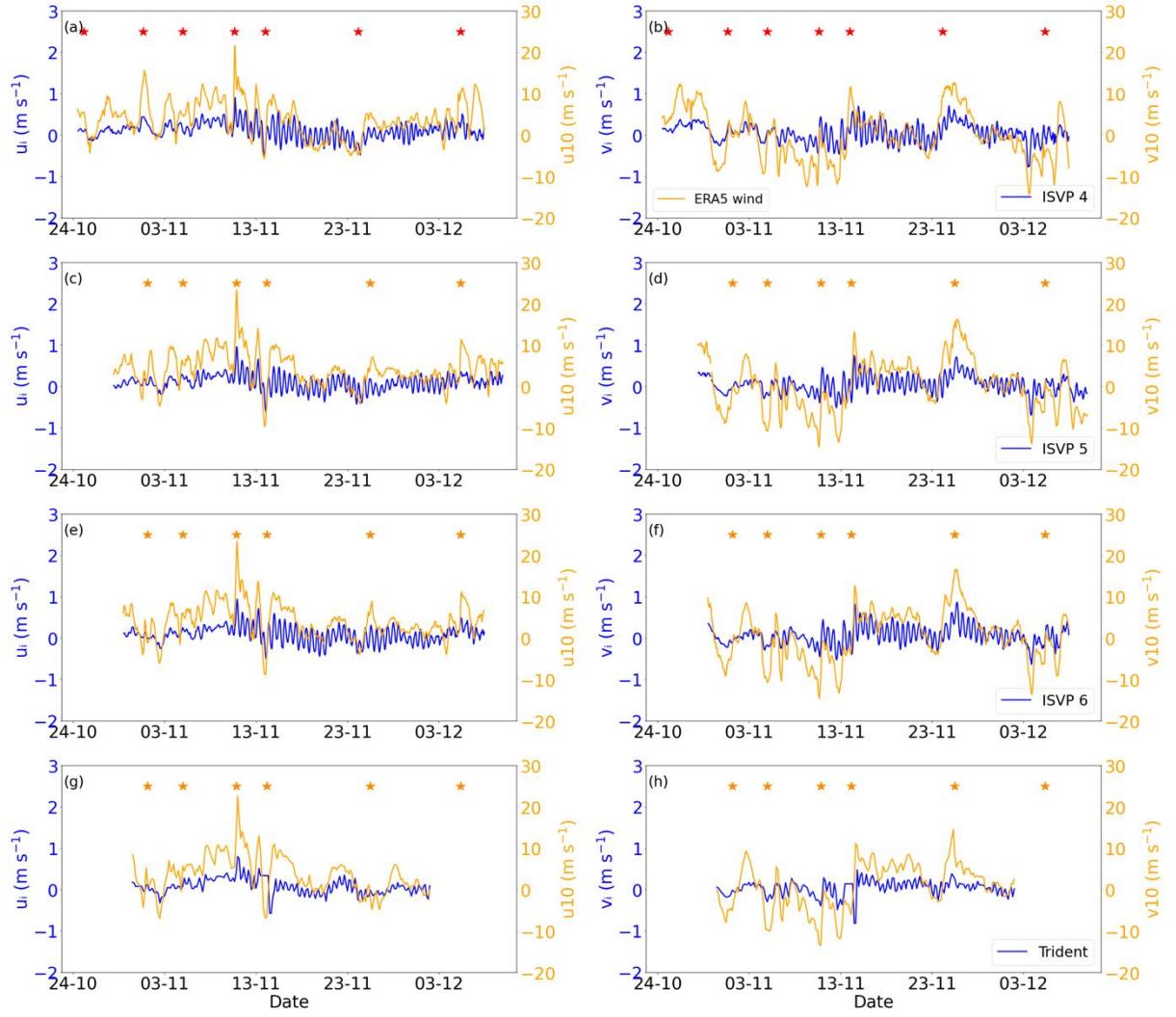


Figure B8. Velocity components of the Spring buoys (on the left axis) and 10 m wind from the ERA5 reanalyses (on the right axis): the zonal and meridional component of ISVP 4 (a-b), ISVP 5 (c-d), ISVP 6 (e-f), and of the Trident (g-h). The red stars denote when the spring cyclones were closest to ISVP 4. The orange stars denote when the same spring cyclones were closest to the main spring cluster $\approx 5^\circ$ east of ISVP 4.

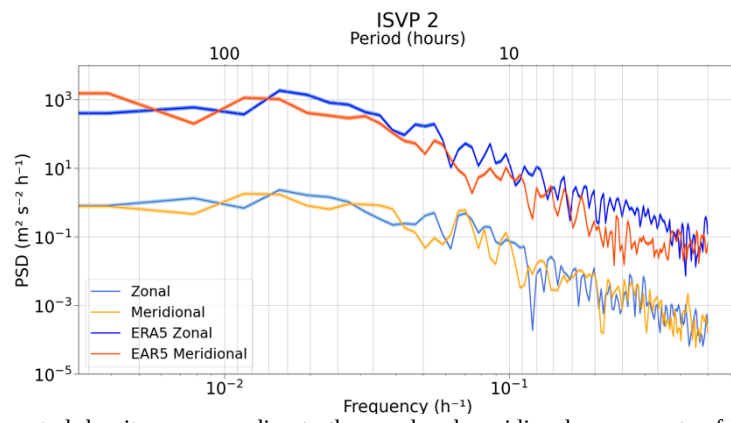


Figure B9. Power spectral density corresponding to the zonal and meridional components of ISVP 2, in winter, and its corresponding ERA5 wind velocity.

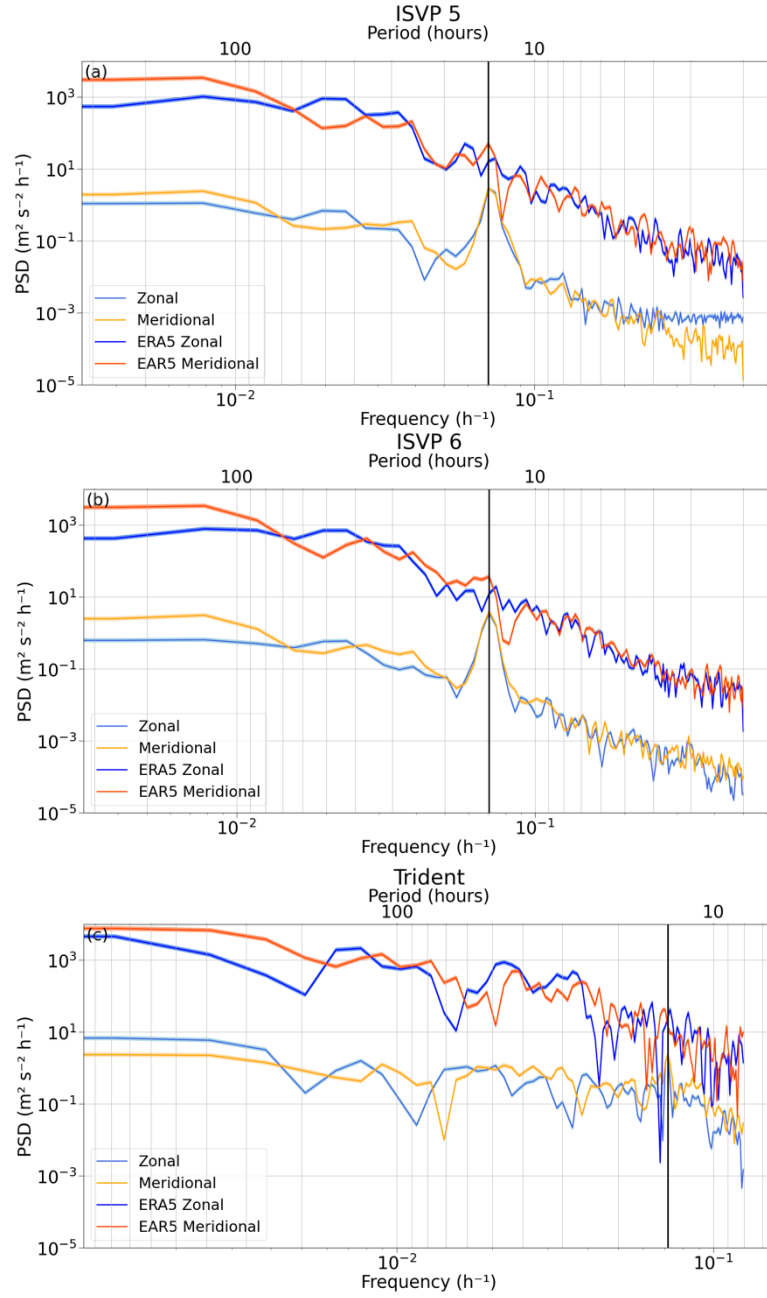


Figure B10. Power spectral density corresponding to the zonal and meridional components of **(a)** ISVP 5, in spring, and its corresponding ERA5 wind velocity. **(b)** Same as in **(a)** but for ISVP 6. **(c)** Same as in **(a)** but for the Trident. The black vertical line indicates the peak associated with inertial oscillations at 14.22 hours for both ISVPs and 13.92 hours for the Trident.

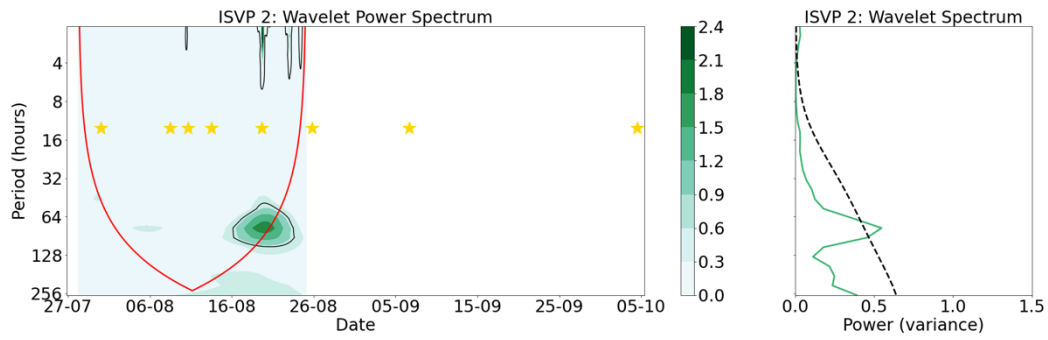


Figure B11. The wavelet power spectrum with time (left panel) and wavelet spectrum (right panel) of the filtered velocity magnitude spectrum of ISVP 2, during winter. The red line indicates the cone of influence (left panel), the black contours (left panel) and black dashed lines (right panel) indicate the 95 % significance level. The yellow stars denote when the cyclones were closest to the Winter buoys' location, and are found at the theoretical inertial frequency, defined by the Coriolis parameter.

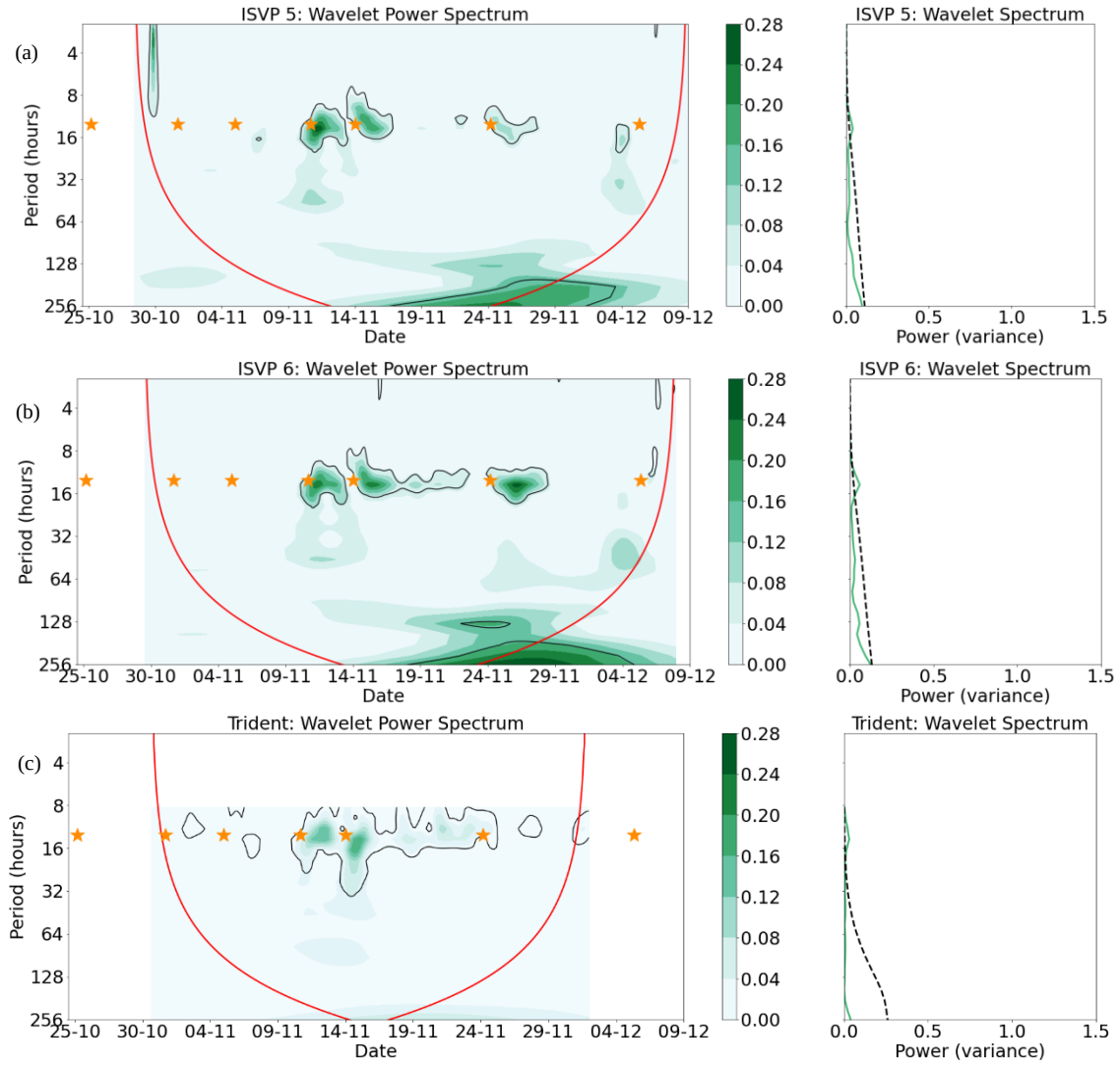


Figure B12. (a) The wavelet power spectrum with time (left panel) and wavelet spectrum (right panel) of the filtered velocity magnitude spectrum of ISVP 5, during spring. (b) Same as (a) but for ISVP 6. (c) Same as (a) but for the Trident buoy. The red line indicates the cone of influence (left panel), the black contours (left panel) and black dashed lines (right panel) indicate the 95 % significance level. The orange stars denote when the same spring cyclones were closest to the main spring cluster $\approx 5^\circ$ east of ISVP 4, and are found at the theoretical inertial frequency, determined by the Coriolis parameter.

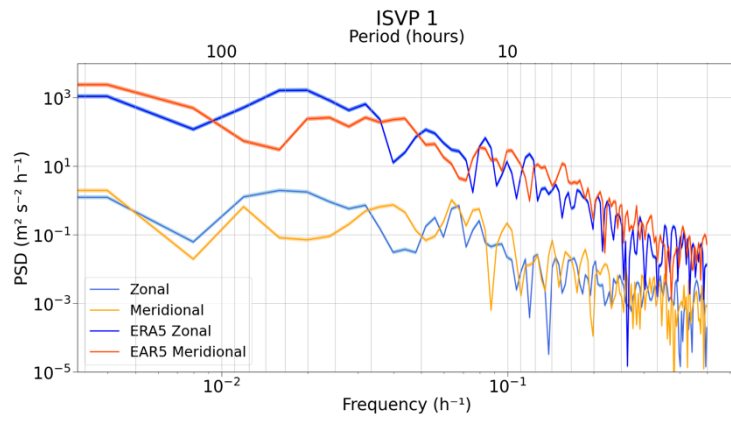


Figure B13. Power spectral density corresponding to the first 10 days of the zonal and meridional components of ISVP 1, in winter, and its corresponding ERA5 wind velocity. The lighter shadings indicate to the 95 % confidence interval for each corresponding ice and wind velocity component.

Appendix C

Table C1: Summary of autonomous buoys

Year	Buoy type	Temporal resolution	Deployment method	Survivability	Measured Variables
2017 Winter	Trident Sensors Helix Beacon	4 hours	Deployed by hand by three people, lowered by crane from the ship to the ice on a basket cradle on a pancake floe.	Dependent on staying fixed to the floe (i.e. non-floating) and the battery life.	Global positioning system (GPS) time, GPS position (latitude and longitude).



2019 Winter	polar Iridium Surface Velocity Profilers (MetOcean model)	30 minutes (ISVP 1) and 1 hour (ISVP 2, ISVP 3)	Deployed by hand by three people, lowered by crane from the ship to the ice on a basket cradle on a pancake floe.	Dependent on battery life.	GPS time, GPS position (latitude and longitude), air temperature, ice temperature, barometric pressure.
-------------	--	---	---	-------------------------------	---



2019 Spring	Standard Iridium Surface Velocity Profilers (Pacific Gyre Model)	1 hour	Deployed by hand by three people, lowered by crane from the ship to the ice on a basket cradle on consolidated ice.	Dependent on battery life.	GPS time, GPS position (latitude and longitude), air temperature, barometric pressure.
-------------	--	--------	---	-------------------------------	---



Specifically
designed frames
were built around
these buoys to allow
them to stand firmly
on the ice floes
without damaging
the non-polar
battery.

2019 Spring	Trident Sensors Helix Beacon	4 hours	Deployed by hand by three people, lowered by crane from the ship to the ice on a basket cradle on consolidated ice.	Dependent on staying fixed to the floe (i.e. non-floating) and the battery life.	GPS time, GPS position (latitude and longitude), air temperature.
-------------	------------------------------------	---------	---	--	---



2019 Winter and Spring	Surface Wave Instrument Float with Tracking buoys	1 hour	Deployed from the side door of the ship's environmental hanger and lowered via a slip line while being steadied by the attached float lines.	Dependent on battery life.	GPS time, GPS position, air temperature, air pressure, wind speed, wind direction, wave spectra, peak wave period, significant wave height, low-resolution GoPro videos.
------------------------	--	--------	--	----------------------------	--

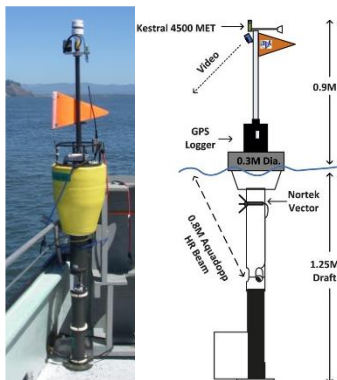


Fig. 1 of Thomson (2012).

References

- Aaboe, S., Down, E.J., and Eastwood, S.: Validation report for the global sea-ice edge and type product, OSI-403-c & EUMETSAT, Norway, 38 pp, 2021.
- Ackley, S.F., Xie, H., and Tichenor, E.A.: Ocean heat flux under Antarctic sea ice in the Bellingshausen and Amundsen Seas: two case studies. *Annals of Glaciology*, 56(69), 200–210. <https://doi.org/10.3189/2015AoG69A890>, 2015.
- Agnew, T., and Howell, S.: The use of operational ice charts for evaluating passive microwave ice concentration data. *Atmosphere-ocean*, 41(4), 317–331. <https://doi.org/10.3137/ao.410405>, 2003.
- Aksamit, N. O., Scharien, R. K., Hutchings, J. K., and Lukovich, J. V.: A quasi-objective single-buoy approach for understanding Lagrangian coherent structures and sea ice dynamics, *The Cryosphere*, 17, 1545–1566, <https://doi.org/10.5194/tc-17-1545-2023>, 2023.
- Alberello, A., Onorato, M., Bennetts, L., Vichi, M., Eayrs, C., MacHutchon, K., and Toffoli, A.: Brief communication: Pancake ice floe size distribution during the winter expansion of the Antarctic marginal ice zone, *The Cryosphere*, 13, 41–48, <https://doi.org/10.5194/tc-13-41-2019>, 2019.
- Alberello, A., Bennetts, L., Heil, P., Eayrs, C., Vichi, M., MacHutchon, K., Onorato, M., and Toffoli, A.: Drift of pancake ice floes in the winter Antarctic marginal ice zone during polar cyclones. *Journal of Geophysical Research: Oceans*, 125(3), p.e2019JC015418. <https://doi.org/10.1029/2019JC015418>, 2020.
- Alberello, A., Bennetts, L.G., Onorato, M., MacHutchon, K., Eayrs, C., Ntamba, B.N., Benetazzo, A., Bergamasco, F., Nelli, F., Pattani, R., Clarke, H., Tersigni, I., and Toffoli, A.: Three-dimensional imaging of waves and floes in the marginal ice zone during a cyclone. *Nature Communication*, 13, 4590. <https://doi.org/10.1038/s41467-022-32036-2>, 2022.
- Allison, I.: The East Antarctic sea ice zone: ice characteristics and drift. *Geojournal*, 18(1), 102–115. <https://doi.org/10.1007/BF00722394>, 1989.

Allison, I.: Physical processes determining the Antarctic sea ice environment. *Australian Journal of Physics*, 50(4), 759-771. <https://doi.org/10.1071/P96113>, 1997.

Allison, I., Brandt, R.E., and Warren, S.G.: East Antarctic sea ice: Albedo, thickness distribution, and snow cover. *Journal of Geophysical Research: Oceans*, 98(C7), 12417-12429. <https://doi.org/10.1029/93JC00648>, 1993.

Andreas, E.L., and Ackley, S.F.: On the differences in ablation seasons of Arctic and Antarctic sea ice. *Journal of Atmospheric Sciences*, 39(2), 440-447. [https://doi.org/10.1175/1520-0469\(1982\)039<0440:OTDIAS>2.0.CO;2](https://doi.org/10.1175/1520-0469(1982)039<0440:OTDIAS>2.0.CO;2), 1982.

Andreas, E.L., Horst, T.W., Grachev, A.A., Persson, P.O.G., Fairall, C.W., Guest, P.S., and Jordan, R.E.: Parametrizing turbulent exchange over summer sea ice and the marginal ice zone. *Quarterly Journal of the Royal Meteorological Society*, 136(649), 927-943. <https://doi.org/10.1002/qj.618>, 2010.

Antarctic Sea-Ice Processes and Climate (ASPeCt): <http://aspect.antarctica.gov.au/>, last access: 5 November 2023.

Ardhuin, F., Aksenov, Y., Benetazzo, A., Bertino, L., Brandt, P., Caubet, E., Chapron, B., Collard, F., Cravatte, S., Delouis, J.-M., Dias, F., Dibarboure, G., Gaultier, L., Johannessen, J., Korosov, A., Manucharyan, G., Menemenlis, D., Menendez, M., Monnier, G., Mouche, A., Noguier, F., Nurser, G., Rampal, P., Reniers, A., Rodriguez, E., Stopa, J., Tison, C., Ubelmann, C., van Sebille, E., and Xie, J.: Measuring currents, ice drift, and waves from space: the Sea surface KInematics Multiscale monitoring (SKIM) concept, *Ocean Science*, 14, 337–354, <https://doi.org/10.5194/os-14-337-2018>, 2018.

Ardhuin, F., Otero, M., Merrifield, S., Grouazel, A., and Terrill, E.: Ice breakup controls dissipation of wind waves across Southern Ocean sea ice. *Geophysical Research Letters*, 47(13), p.e2020GL087699. <https://doi.org/10.1029/2020GL087699>, 2020.

Arndt, S.: Sea-ice surface properties and their impact on the under-ice light field from remote sensing data and in-situ measurements, Doctoral dissertation, University of Bremen, 2016.

Ashton, G.D.: Thin ice growth. *Water Resources Research*, 25(3), 564-566. <https://doi.org/10.1029/WR025i003p00564>, 1989.

Babanin, A. V.: Breaking and dissipation of ocean surface waves (p. 480). Cambridge University Press, United Kingdom. 2011.

Babanin, A.V.: Ocean Waves in Large-Scale Air-Sea Weather and Climate Systems. *Journal of Geophysical Research: Oceans*, 128(3), p.e2023JC019633. <https://doi.org/10.1029/2023JC019633>, 2023.

Babanin, A.V., Rogers, W.E., De Camargo, R., Doble, M., Durrant, T., Filchuk, K., Ewans, K., Hemer, M., Janssen, T., Kelly-Gerreyn, B., and Machutchon, K.: Waves and swells in high wind and extreme fetches, measurements in the Southern Ocean. *Frontiers in Marine Science*, 6, 361. <https://doi.org/10.3389/fmars.2019.00361>, 2019.

Bartels-Rausch, T.: Ten things we need to know about ice and snow. *Nature*, 494(7435), 27-29. <https://doi.org/10.1038/494027a>, 2013.

Bateson, A. W., Feltham, D. L., Schröder, D., Hosekova, L., Ridley, J. K., and Aksenov, Y.: Impact of sea ice floe size distribution on seasonal fragmentation and melt of Arctic sea ice, *The Cryosphere*, 14, 403–428, <https://doi.org/10.5194/tc-14-403-2020>, 2020.

Beitsch, A., Kern, S., and Kaleschke, L.: Comparison of SSM/I and AMSR-E sea ice concentrations with ASPeCt ship observations around Antarctica. *IEEE Transactions on Geoscience and Remote Sensing*, 53(4), 1985-1996. doi: 10.1109/TGRS.2014.2351497, 2014.

Bennetts, L. G., O'Farrell, S., and Uotila, P.: Brief communication: Impacts of ocean-wave-induced breakup of Antarctic sea ice via thermodynamics in a stand-alone version of the CICE sea-ice model, *The Cryosphere*, 11, 1035–1040, <https://doi.org/10.5194/tc-11-1035-2017>, 2017.

Bennetts, L.G., and Williams, T.D.: Water wave transmission by an array of floating discs. *Proceedings of the Royal Society A: Mathematical, Physical and Engineering Sciences*, 471(2173), 20140698. <https://doi.org/10.1098/rspa.2014.0698>, 2015.

Biddle, L.C., and Swart, S.: The observed seasonal cycle of submesoscale processes in the Antarctic marginal ice zone. *Journal of Geophysical Research: Oceans*, 125(6), p.e2019JC015587. <https://doi.org/10.1029/2019JC015587>, 2020.

Bouchat, A., Hutter, N., Chanut, J., Dupont, F., Dukhovskoy, D., Garric, G., Lee, Y.J., Lemieux, J.F., Lique, C., Losch, M., Maslowski, W., Myers, P.G., Olason, E., Rampal, P., Rasmussen, T., Talandier, C., Tremblay, B., and Wang, Q.: Sea ice rheology experiment (sirex): 1. scaling and statistical properties of sea-ice deformation fields. *Journal of Geophysical Research: Oceans*, 127(4):e2021JC017667, <https://doi.org/10.1029/2021JC017666>, 2022.

Boutin, G., Ardhuin, F., Dumont, D., Sévigny, C., Girard-Ardhuin, F., and Accensi, M.: Floe size effect on wave-ice interactions: Possible effects, implementation in wave model, and evaluation. *Journal of Geophysical Research: Oceans*, 123(7), 4779-4805. <https://doi.org/10.1029/2017JC013622>, 2018.

Boutin, G., Williams, T., Rampal, P., Olason, E., and Lique, C.: Wave–sea-ice interactions in a brittle rheological framework, *The Cryosphere*, 15, 431–457, <https://doi.org/10.5194/tc-15-431-2021>, 2021.

Brouwer, J., Fraser, A. D., Murphy, D. J., Wongpan, P., Alberello, A., Kohout, A., Horvat, C., Wotherspoon, S., Massom, R. A., Cartwright, J., and Williams, G. D.: Altimetric observation of wave attenuation through the Antarctic marginal ice zone using ICESat-2, *The Cryosphere*, 16, 2325–2353, <https://doi.org/10.5194/tc-16-2325-2022>, 2022.

Campana, J., Terrill, E.J., and de Paolo, T.: A new inversion method to obtain upper-ocean current-depth profiles using X-band observations of deep-water waves. *Journal of Atmospheric and Oceanic Technology*, 34(5), 957-970. <https://doi.org/10.1175/JTECH-D-16-0120.1>, 2016.

Campbell, A.J.: Stereo imaging techniques for three-dimensional surface wave measurement: applications and improvements, Doctoral dissertation, University of Wisconsin, <http://digital.library.wisc.edu/1793/69701>, 2014.

Cavalieri, D. J., Parkinson, C. L., Gloersen, P., and Zwally, H. J.: Sea Ice Concentrations from Nimbus-7 SMMR and DMSP SSM/I-SSMIS Passive Microwave Data, Version 1, NASA National Snow and Ice Data Center Distributed Active Archive Center Boulder, Colorado USA [data set], <https://doi.org/10.5067/8GQ8LZQVL0VL>, 1996.

Cole, S.T., Toole, J.M., Lele, R., Timmermans, M.L., Gallaher, S.G., Stanton, T.P., Shaw, W.J., Hwang, B., Maksym, T., Wilkinson, J.P., and Ortiz, M.: Ice and ocean velocity in the Arctic marginal ice zone: Ice roughness and momentum transfer. *Elem Sci Anth*, 5, 55. <https://doi.org/10.1525/elementa.241>, 2017.

Collins, C.O., Rogers, W.E., and Lund, B.: An investigation into the dispersion of ocean surface waves in sea ice. *Ocean Dynamics*, 67, 263–280. <https://doi.org/10.1007/s10236-016-1021-4>, 2017.

Colony, R., and Thorndike, A.S.: The horizontal coherency of the motion of summer Arctic sea ice. *Journal of Physical Oceanography*, 10(8), 1281-1289. [https://doi.org/10.1175/1520-0485\(1980\)010<1281:THCOTM>2.0.CO;2](https://doi.org/10.1175/1520-0485(1980)010<1281:THCOTM>2.0.CO;2), 1980.

Colony, R., and Thorndike, A.S.: An estimate of the mean field of Arctic sea ice motion. *Journal of Geophysical Research: Oceans*, 89(C6), 10623-10629. <https://doi.org/10.1029/JC089iC06p10623>, 1984.

Coon, M. D., Maykut, G.A., Pritchard, R. S., Rothrock, D.A., and Thorndike, A. S.: Modeling the pack ice as an elastic-plastic material. *AIDJEX Bull.* 24, 1-105. 1974.

Cooper, V.T., Roach, L.A., Thomson, J., Brenner, S.D., Smith, M.M., Meylan, M.H., and Bitz, C.M.: Wind waves in sea ice of the western Arctic and a global coupled wave-ice model. *Philosophical Transactions of the Royal Society A*, 380(2235), 20210258. <https://doi.org/10.1098/rsta.2021.0258>, 2022.

Copernicus Climate Change Service (C3S).: ERA5: Fifth generation of ECMWF atmospheric reanalyses of the global climate. Reading, UK: Copernicus Climate Change Service Climate Data Store (CDS). Retrieved from <https://cds.climate.copernicus.eu/cdsapp#!/home>, 2017.

Copernicus Marine Service Information (CMEMS).: ERA5: Global Total Surface and 15m Current (COPERNICUS-GLOBCURRENT) from Altimetric Geostrophic Current and Modeled Ekman Current Reprocessing. Reading, UK: Copernicus Marine Data Store (MDS). Retrieved from <https://marine.copernicus.eu/>, 2014.

Curry, J.A., Schramm, J.L., and Ebert, E.E.: Sea ice-albedo climate feedback mechanism. *Journal of Climate*, 8(2), 240-247. [https://doi.org/10.1175/1520-0442\(1995\)008<0240:SIACFM>2.0.CO;2](https://doi.org/10.1175/1520-0442(1995)008<0240:SIACFM>2.0.CO;2), 1995.

Delhommeau, G., Babarit, A., Singh, J., Guével, P., Daubisse, J.C., and Kurnia, P.: Nemoh [code software], <https://gitlab.com/lheea/Nemoh>, 2022.

Derkani, M. H., Alberello, A., Nelli, F., Bennetts, L. G., Hessner, K. G., MacHutchon, K., Reichert, K., Aouf, L., Khan, S., and Toffoli, A.: Wind, waves, and surface currents in the Southern Ocean: observations from the Antarctic Circumnavigation Expedition, *Earth Syst. Sci. Data*, 13, 1189–1209, <https://doi.org/10.5194/essd-13-1189-2021>, 2021.

Derkani, M.H., Hessner, K., Zieger, S., Nelli, F., Alberello, A., and Toffoli, A.: May. Evaluation of the numerical wave model (WaveWatch III) for wave simulation in the Antarctic marginal ice zone. In *EGU General Assembly Conference Abstracts*, p. 12247. <https://doi.org/10.5194/egusphere-egu2020-12247>, 2020.

de Vos, M., Barnes, M., Biddle, L.C., Swart, S., Ramjukadh C., and Vichi M.: Evaluating numerical and free-drift forecasts of sea ice drift during a Southern Ocean research expedition: An operational perspective, *Journal of Operational Oceanography*, 15(3), 187-203, <https://doi.org/10.1080/1755876X.2021.1883293>, 2022.

de Vos, M., Ramjukadh, C., de Villiers, M., Lyttle, C., Womack, A., and Vichi, M.: Polar Iridium Surface Velocity Profilers (p-iSVP), and standard Iridium Surface Velocity Profilers (iSVP) during SCALE 2019 Winter and Spring Cruises, Zenodo [data set], <https://doi.org/10.5281/zenodo.7954779>, 2023.

de Vos, M., Ramjukadh, C., Liesker, C.G., de Villiers, M., and Lyttle, C.: Southern Ocean Marginal Ice Zone Photographs from a 2019 Winter Research Cruise (SA Agulhas II). Pangea [data set]. <https://doi.org/10.1594/PANGAEA.905497>, 2019.

Dierking, W., Stern, H.L., and Hutchings, J.K.: Estimating statistical errors in retrievals of ice velocity and deformation parameters from satellite images and buoy arrays. *The Cryosphere*, 14(9), 2999-3016. <https://doi.org/10.5194/tc-14-2999-2020>, 2020.

Doble, M.J.: Simulating pancake and frazil ice growth in the Weddell Sea: A process model from freezing to consolidation. *Journal of Geophysical Research: Oceans*, 114(C9). <https://doi.org/10.1029/2008JC004935>, 2009.

Doble, M.J., and Bidlot, J.R., 2013. Wave buoy measurements at the Antarctic sea ice edge compared with an enhanced ECMWF WAM: Progress towards global waves-in-ice modelling. *Ocean Modelling*, 70, 166-173. <https://doi.org/10.1016/j.ocemod.2013.05.012>, 2013.

Doble, M.J., Coon, M.D., and Wadhams, P.: Pancake ice formation in the Weddell Sea. *Journal of Geophysical Research: Oceans*, 108(C7). <https://doi.org/10.1029/2002JC001373>, 2003.

Doble, M.J., and Wadhams, P.: Dynamical contrasts between pancake and pack ice, investigated with a drifting buoy array. *Journal of Geophysical Research: Oceans*, 111(C11), C11S24. <https://doi.org/10.1029/2005JC003320>, 2006.

Dobrynin, M., Murawsky, J., and Yang, S.: Evolution of the global wind wave climate in CMIP5 experiments. *Geophysical Research Letters*, 39(18). <https://doi.org/10.1029/2012GL052843>, 2012.

Dolatshah, A., Nelli, F., Bennetts, L.G., Alberello, A., Meylan, M.H., Monty, J.P., and Toffoli, A.: Hydroelastic interactions between water waves and floating freshwater ice. *Physics of Fluids*, 30(9). <https://doi.org/10.1063/1.5050262>, 2018.

Dumont, D., Kohout, A., and Bertino, L.: A wave-based model for the marginal ice zone including a floe breaking parameterization. *Journal of Geophysical Research: Oceans*, 116(C4). <https://doi.org/10.1029/2010JC006682>, 2011.

Eayrs, C., Faller, D., and Holland, D.M.: Mechanisms driving the asymmetric seasonal cycle of Antarctic Sea Ice in the CESM Large Ensemble. *Annals of Glaciology*, 61(82), 171-180. <https://doi.org/10.1017/aog.2020.26>, 2020.

Eayrs, C., Holland, D., Francis, D., Wagner, T., Kumar, R., and Li, X.: Understanding the seasonal cycle of Antarctic Sea Ice extent in the context of longer-term variability. *Reviews of Geophysics*, 57(3), 1037–1064. <https://doi.org/10.1029/2018RG000631>, 2019a.

Eayrs, C., Holland, D., Mojica, J., Vichi, M., Alberello, A., Bekker, A., Bennetts, L., De Jong, E., Joubert, W., Machutchon, K., Messori, G., Onorato, M., Saunders, C., Skatulla, S., and Toffoli, A.: SA Agulhas II Winter 2017 Cruise: Waves In Ice Observation Systems (WIIOS), Ver. 1, Australian Antarctic Data Centre [data set], doi:10.26179/5cc934992f065, 2019b.

Eayrs, C., Li, X., Raphael, M.N., and Holland, D.M.: Rapid decline in Antarctic sea ice in recent years hints at future change. *Nature Geoscience*, 14(7), 460-464. <https://doi.org/10.1038/s41561-021-00768-3>, 2021.

Ekman, V.W.: On the influence of the earth's rotation on ocean-currents. 1905.

Enrile, F., Besio, G., Stocchino, A., and Magaldi, M.G.: Influence of initial conditions on absolute and relative dispersion in semi-enclosed basins. *Plos one*, 14(7), p.e0217073. <https://doi.org/10.1371/journal.pone.0217073>, 2019.

Faltinsen, O.: Sea loads on ships and offshore structures (Vol. 1). Cambridge university press, United Kingdom. 1993.

Farooq, U., Rack, W., McDonald, A., and Howell, S.: Representation of sea ice regimes in the Western Ross Sea, Antarctica, based on satellite imagery and AMPS wind data. *Climate Dynamics*, 60(1-2), 227-238. <https://doi.org/10.1007/s00382-022-06319-9>, 2023.

Feltham, D.L.: Granular flow in the marginal ice zone. *Philosophical Transactions of the Royal Society A: Mathematical, Physical and Engineering Sciences*, 363(1832), 1677-1700. <https://doi.org/10.1098/rsta.2005.1601>, 2005.

Feltham, D.L.: Sea ice rheology. *Annual Review of Fluid Mechanics*, 40(1):91–112, <https://doi.org/10.1146/annurev.fluid.40.111406.102151>, 2008.

Figuerro, A., Peña, E., Rodríguez, A., Maciñeira, E., Rabuñal, J., Sande, J., Costa, F., Castro, A., and Maciñeira, E.: Inertial Measurement Unit to Determine Moored Vessels Movements. In *Proceedings of the 6th International Conference on the Application of Physical Modelling in Coastal and Port Engineering and Science COASTLAB 2016*. 2016.

Flato, G.M.: Sea-ice modelling. Mass balance of the cryosphere, 367-392. 2003.

Fox, C., and Haskell, T.G.: Ocean wave speed in the Antarctic marginal ice zone. *Annals of Glaciology*, 33, 350-354. <https://doi.org/10.3189/172756401781818941>, 2001.

Geiger, C.A., Ackley, S.F., and Hibler III, W.D.: Sea ice drift and deformation processes in the western Weddell Sea. *Antarctic sea ice: physical processes, interactions and variability*, 74, 141-160. <https://doi.org/10.1029/AR074p0141>, 1998.

Geiger, C.A., and Drinkwater, M.R.: Coincident buoy-and SAR-derived surface fluxes in the western Weddell Sea during Ice Station Weddell 1992. *Journal of Geophysical Research: Oceans*, 110(C4). <https://doi.org/10.1029/2003JC002112>, 2005.

Geiger, C.A., and Perovich, D.K.: Springtime ice motion in the western Antarctic Peninsula region. *Deep Sea Research Part II: Topical Studies in Oceanography*, 55(3-4), 338-350. <https://doi.org/10.1016/j.dsr2.2007.11.008>, 2008.

Gimbert, F., Jourdain, N.C., Marsan, D., Weiss, J., and Barnier, B.: Recent mechanical weakening of the Arctic sea ice cover as revealed from larger inertial oscillations. *Journal of Geophysical Research: Oceans*, 117(C11). <https://doi.org/10.1029/2011JC007633>, 2012a.

Gimbert, F., Marsan, D., Weiss, J., Jourdain, N.C. and Barnier, B.: Sea ice inertial oscillations in the Arctic Basin. *The Cryosphere*, 6(5), 1187-1201. <https://doi.org/10.5194/tc-6-1187-2012>, 2012b.

Girard, L., Weiss, J., Molines, J.M., Barnier, B., and Bouillon, S.: Evaluation of high-resolution sea ice models on the basis of statistical and scaling properties of Arctic sea ice drift and deformation. *Journal of Geophysical Research: Oceans*, 114(C8). <https://doi.org/10.1029/2008JC005182>, 2009.

Glover, D.M., Jenkins, W.J., and Doney, S.C.: Modeling methods for marine science. Cambridge University Press. <https://doi.org/10.1017/CBO9780511975721>, 2011.

Gordon, A.L.: Seasonality of Southern Ocean sea ice. *Journal of Geophysical Research: Oceans*, 86(C5), 4193-4197. <https://doi.org/10.1029/JC086iC05p04193>, 1981.

Goosse, H., and Zunz, V.: Decadal trends in the Antarctic sea ice extent ultimately controlled by ice–ocean feedback. *The Cryosphere*, 453–470. <https://doi.org/10.5194/tc-8-453-2014>, 2014.

Grieger, J., Leckebusch, G.C., Raible, C.C., Rudeva, I., and Simmonds, I.: Subantarctic cyclones identified by 14 tracking methods, and their role for moisture transports into the continent. *Tellus A: Dynamic Meteorology and Oceanography*, 70(1), 1454808. <https://doi.org/10.1080/16000870.2018.1454808>, 2018.

Haller, G.: Lagrangian coherent structures. *Annual review of fluid mechanics*, 47, 137-162. <https://doi.org/10.1002/9783527639748.ch3>, 2015.

Handcock, M. S. and Raphael, M. N.: Modeling the annual cycle of daily Antarctic sea ice extent, *The Cryosphere*, 14, 2159–2172, <https://doi.org/10.5194/tc-14-2159-2020>, 2020.

Hausmann, U., Sallée, J.B., Jourdain, N.C., Mathiot, P., Rousset, C., Madec, G., Deshayes, J., and Hatterman, T.: The role of tides in ocean-ice shelf interactions in the southwestern Weddell Sea. *Journal of Geophysical Research: Oceans*, 125, e2019JC015847. <https://doi.org/10.1029/2019JC015847>, 2020.

Hedges, T.S.: An empirical modification to linear wave theory. *Proceedings of the Institution of Civil Engineers*, 61(3), pp.575-579. <https://doi.org/10.1680/iicep.1976.3408>, 1976.

Heil, P.: Sea-ice growth, drift and deformation off Eastern Antarctica, Doctoral dissertation, University of Tasmania, 1999.

Heil, P., Fowler, C.W., Maslanik, J.A., Emery, W.J., and Allison, I.: A comparison of East Antarctic sea-ice motion derived using drifting buoys and remote sensing. *Annals of Glaciology*, 33, 139-144. <https://doi.org/10.3189/172756401781818374>, 2001.

Heil, P., and Hibler, W.D.: Modeling the high-frequency component of Arctic sea ice drift and deformation. *Journal of Physical Oceanography*, 32(11), 3039-3057. [https://doi.org/10.1175/1520-0485\(2002\)032<3039:MTHFCO>2.0.CO;2](https://doi.org/10.1175/1520-0485(2002)032<3039:MTHFCO>2.0.CO;2), 2002.

Heil, P., Hutchings, J.K., Worby, A.P., Johansson, M., Launiainen, J., Haas, C., and Hibler III, W.D.: Tidal forcing on sea ice drift and deformation in the western Weddell Sea in early austral summer, 2004. *Deep Sea Research Part II: Topical Studies in Oceanography*, 55(8-9), 943-962. <https://doi.org/10.1016/j.dsr2.2007.12.026>, 2008.

Heil, P., Massom, R.A., Allison, I., Worby, A.P., and Lytle, V.I.: Role of off-shelf to on-shelf transitions for East Antarctic sea ice dynamics during spring 2003. *Journal of Geophysical Research: Oceans*, 114(C9). <https://doi.org/10.1029/2008JC004873>, 2009.

Heil, P., Massom, R.A., Allison, I., and Worby, A.P.: Physical attributes of sea ice kinematics during spring 2007 off East Antarctica. *Deep Sea Research Part II: Topical Studies in Oceanography*, 58(9-10), 1158-1171. <https://doi.org/10.1016/j.dsr2.2010.12.004>, 2011.

Heorton, H.D., Feltham, D.L., and Hunt, J.C.: The response of the sea ice edge to atmospheric and oceanic jet formation. *Journal of Physical Oceanography*, 44(9), 2292-2316. <https://doi.org/10.1175/JPO-D-13-0184.1>, 2014.

Heorton, H.D., Tsamados, M., Cole, S.T., Ferreira, A.M., Berbellini, A., Fox, M., and Armitage, T.W.: Retrieving sea ice drag coefficients and turning angles from in-situ and satellite observations using an inverse modeling framework. *Journal of Geophysical Research: Oceans*, 124(8), 6388-6413. <https://doi.org/10.1029/2018JC014881>, 2019.

Hepworth, E., Vichi, M., van Zuydam, A., Taylor, N.C., Bossau, J., Engelbrecht, M., and Aarskog, T.: Sea Ice Observations in the Antarctic Marginal Ice Zone during Winter 2019. *Pangea* [dataset], <https://doi.org/10.1594/PANGAEA.921759>, 2020.

Herman, A.: Wave-induced stress and breaking of sea ice in a coupled hydrodynamic discrete-element wave-ice model, *The Cryosphere*, 11, 2711–2725, <https://doi.org/10.5194/tc-11-2711-2017>, 2017.

- Herman, A.: Wave-induced surge motion and collisions of sea ice floes: Finite-floe-size effects. *Journal of Geophysical Research: Oceans*, 123, 7472–7494. <https://doi.org/10.1029/2018JC014500>, 2018.
- Hibler, W.D.: A dynamic thermodynamic sea ice model. *Journal of physical oceanography*, 9(4), 815–846. [https://doi.org/10.1175/1520-0485\(1979\)009<0815:ADTSIM>2.0.CO;2](https://doi.org/10.1175/1520-0485(1979)009<0815:ADTSIM>2.0.CO;2), 1979.
- Hibler, W.D.: Ice Dynamics. In: Untersteiner, N. (eds) The Geophysics of Sea Ice. NATO ASI Series. Springer, Boston, Massachusetts. https://doi.org/10.1007/978-1-4899-5352-0_10, 1986.
- Hibler III, W.D., Heil, P., and Lyle, V.I.: On simulating high-frequency variability in Antarctic sea-ice dynamics models. *Annals of Glaciology*, 27, 443–448. <https://doi.org/10.3189/1998AoG27-1-443-448>, 1998.
- Hibler, W.D., Hutchings, J.K. and Ip, C.F.: Sea ice arching and multiple flow states of Arctic pack ice. *Annals of Glaciology*, 44, 339–344. <https://doi.org/10.3189/172756406781811448>, 2006.
- Hibler, W.D., Weeks, W.F., Kovacs, A., and Ackley, S.F.: Differential sea-ice drift. I. Spatial and temporal variations in sea-ice deformation. *Journal of Glaciology*, 13(69), 437–455. <https://doi.org/10.3189/S0022143000023212>, 1974.
- Hobbs, W.R., Massom, R., Stammerjohn, S., Reid, P., Williams, G., and Meier, W.: A review of recent changes in Southern Ocean sea ice, their drivers and forcings. *Global and Planetary Change*, 143, 228–250. <https://doi.org/10.1016/j.gloplacha.2016.06.008>, 2016.
- Holthuijsen, L.H.: Waves in oceanic and coastal waters (1st Ed.). Cambridge university press, United Kingdom. 2007.
- Hopkins, M.A., Tuhkuri, J., and Lensu, M.: Rafting and ridging of thin ice sheets. *Journal of Geophysical Research: Oceans*, 104(C6), 13605–13613. <https://doi.org/10.1029/1999JC900031>, 1999.
- Horvat, C.: Floes, the marginal ice zone and coupled wave-sea-ice feedbacks. *Philosophical Transactions of the Royal Society A*, 380(2235), p.20210252. <https://doi.org/10.1098/rsta.2021.0252>, 2022.
- Hoskins, B.J., and Hodges, K.I.: A new perspective on southern hemisphere storm tracks. *Journal of Climate*, 18(20), 4108–4129. <https://doi.org/10.1175/JCLI3570.1>, 2005.
- Hunkins, K.: The oceanic boundary layer and stress beneath a drifting ice floe. *Journal of Geophysical Research*, 80(24), 3425–3433. <https://doi.org/10.1029/JC080i024p03425>, 1975.

Hutchings, J.K., and Hibler III, W.D.: Small-scale sea ice deformation in the Beaufort Sea seasonal ice zone. *Journal of Geophysical Research: Oceans*, 113(C8). <https://doi.org/10.1029/2006JC003971>, 2008.

Hutchings, J.K., Roberts, A., Geiger, C.A., and Richter-Menge, J.: Spatial and temporal characterization of sea ice deformation. *Annals of Glaciology*, 52(57), 360-368. <https://doi.org/10.3189/172756411795931769>, 2011.

Irving, D., Simmonds, I., and Keay, K.: Mesoscale cyclone activity over the ice-free Southern Ocean: 1999–2008. *Journal of Climate*, 23(20), 5404–5420. <https://doi.org/10.1175/2010JCLI3628.1>, 2010.

Itkin, P., Spreen, G., Cheng, B., Doble, M., Girard-Arduin, F., Haapala, J., Hughes, N., Kaleschke, L., Nicolaus, M., and Wilkinson, J. Thin ice and storms: Sea ice deformation from *buoy arrays deployed during N-ICE 2015*. *Journal of Geophysical Research: Oceans*, 122(6), 4661–4674. <https://doi.org/10.1002/2016JC012403>, 2017.

Johnson, S., Audh, R., De Jager, W., Matlakala, B., Vichi, M., Womack, A., and Rampai, T.: Physical and morphological properties of first-year Antarctic sea ice in the spring marginal ice zone of the Atlantic-Indian sector. *Journal of Glaciology*, 1-14. <https://doi.org/10.1017/jog.2023.21>, 2023.

Jones, D.A., and Simmonds, I.: A climatology of Southern Hemisphere extratropical cyclones. *Climate Dynamics*, 9, 131–145. <https://doi.org/10.1007/BF00209750>, 1993.

Kamphuis, J.W.: Introduction to Coastal Engineering and Management 3rd Edition, Advanced Series on Ocean Engineering: Volume 48, Singapore: World Scientific. 7, 145–161. <https://doi.org/10.1142/11491>, 2020.

Kawaguchi, Y., Itoh, M., Fukamachi, Y., Moriya, E., Onodera, J., Kikuchi, T., and Harada, N.: Year-round observations of sea ice drift and near-inertial internal waves in the Northwind Abyssal Plain, Arctic Ocean. *Polar Science*, 21, 212–223. <https://doi.org/10.1016/j.polar.2019.01.004>, 2019.

Kern, S., Lavergne, T., Notz, D., Pedersen, L. T., Tonboe, R. T., Saldo, R., and Sørensen, A. M.: Satellite passive microwave sea-ice concentration data set intercomparison: closed ice and ship-based observations, *The Cryosphere*, 13, 3261–3307, <https://doi.org/10.5194/tc-13-3261-2019>, 2019.

Kimura, N.: Sea ice motion in response to surface wind and ocean current in the Southern Ocean. *Journal of the Meteorological Society of Japan*. Ser. II, 82(4), 1223–1231. <https://doi.org/10.2151/jmsj.2004.1223>, 2004.

Kimura, N., and Wakatsuchi, M.: Relationship between sea ice motion and geostrophic wind in the Northern Hemisphere. *Geophysical Research Letters*, 27(22), 3735–3738. <https://doi.org/10.1029/2000GL011495>, 2000.

King, J.C., Marshall, G.J., Colwell, S., Arndt, S., Allen-Sader, C., and Phillips, T.: The Performance of the ERA-Interim and ERA5 Atmospheric Reanalyses Over Weddell Sea Pack Ice. *Journal of Geophysical Research: Oceans*, 127(9), p.e2022JC018805. <https://doi.org/10.1029/2022JC018805>, 2022.

Kohout, A.L., Williams, M.J.M, Dean, S.M., and Meylan, M.H.: Storm-induced sea ice breakup and the implications for ice extent. *Nature*, 509(7502), 604–607. 10.1038/nature13262, 2014.

Kohout, A.L., Smith, M., Roach L.A., Williams, G., Montiel, F., and Williams, M.J.M.: Observations of exponential wave attenuation in Antarctic sea ice during the PIPERS campaign. *Annals of Glaciology*, 1–14. <https://doi.org/10.1017/aog.2020.36>, 2020.

Kok, M., Hol, J.D., and Schön, T.B.: Using inertial sensors for position and orientation estimation. *Now foundations and trends*. <https://doi.org/10.48550/arXiv.1704.06053>, 2017.

Kottmeier, C., Olf, J., Frieden, W., and Roth, R.: Wind forcing and ice motion in the Weddell Sea region, *Journal of Geophysical Research: Atmospheres*, 97, 20, 373–20, 383. <https://doi.org/10.1029/92JD02171>, 1992.

Kousal, J., Voermans, J.J., Liu, Q., Heil, P., and Babanin, A.V.: A Two-Part Model for Wave-Sea Ice Interaction: Attenuation and Break-Up. *Journal of Geophysical Research: Oceans*, 127(5), p.e2022JC018571. <https://doi.org/10.1029/2022JC018571>, 2022.

Kwok, R., Pang, S.S., and Kacimi, S.: Sea ice drift in the Southern Ocean: Regional patterns, variability, and trends. *Elem Sci Anth*, 5. <https://doi.org/10.1525/elementa.226>, 2017.

LaCasce, J.: Statistics from Lagrangian observations. *Progress in Oceanography*, 77(1), 1-29. <https://doi.org/10.1016/j.pocean.2008.02.002>, 2008.

Langhorne, P.J., Squire, V.A., Fox, C., and Haskell, T.G.: Break-up of sea ice by ocean waves. *Annals of Glaciology*, 27, 438-442. <https://doi.org/10.3189/S0260305500017869>, 1998.

Lammert, A., Brümmer, B., and Kaleschke, L.: Observation of cyclone-induced inertial sea ice oscillation in Fram Strait. *Geophysical research letters*, 36, 10. <https://doi.org/10.1029/2009GL037197>, 2009.

Lavergne, T., and Down, E.J.: 2022: Validation report for the global sea-ice drift climate data record, OSI-455 & EUMETSAT, Norway, 36 pp. doi: 10.15770/EUM_SAF_OSI_0012, 2022.

Lei, R., Gui, D., Hutchings, J.K., Heil, P., and Li, N.: Annual cycles of sea ice motion and deformation derived from buoy measurements in the western Arctic Ocean over two ice seasons. *Journal of Geophysical Research: Oceans*, 125(6), p.e2019JC015310. <https://doi.org/10.1029/2019JC015310>, 2020.

Lei, R., Hoppmann, M., Cheng, B., Zuo, G., Gui, D., Cai, Q., Belter, H.J., and Yang, W.: Seasonal changes in sea ice kinematics and deformation in the Pacific sector of the Arctic Ocean in 2018/19. *The Cryosphere*, 15(3), 1321-1341. <https://doi.org/10.5194/tc-15-1321-2021>, 2021.

Lemieux, J., Bouillon, S., Dupont F., Flato, G., Losch, M., Rampal, P., Tremblay, L., Vancoppenolle, M., and Williams, T.: Sea Ice Analysis and Forecasting. Edition 1. Edited by T. Carrieres, M. Buehner, J. Lemieux and L.T. Pedersen. Cambridge University Press, United Kingdom. doi: 10.1017/9781108277600. 2017.

Leppäranta, M.: The drift of sea ice, Edition 2, Springer Berlin, Heidelberg. <https://doi.org/10.1007/978-3-642-04683-4>, 2011.

Leppäranta, M., and Omstedt, A.: Dynamic coupling of sea ice and water for an ice field with free boundaries. *Tellus A: Dynamic Meteorology and Oceanography*, 42(4), 482-495. <https://doi.org/10.3402/tellusa.v42i4.11892>, 1990.

Li, J., Babanin, A.V., Liu, Q., Voermans, J.J., Heil, P., and Tang, Y.: Effects of wave-induced sea ice break-up and mixing in a high-resolution coupled ice-ocean model. *Journal of Marine Science and Engineering*, 9(4), 365. <https://doi.org/10.3390/jmse9040365>, 2021.

Li, M., Liu, J., Wang, Z., Wang, H., Zhang, Z., Zhang, L., and Yang, Q.: Assessment of sea surface wind from NWP reanalyses and satellites in the Southern Ocean, *Journal of Atmospheric and Oceanic Technology*, 30, 1842–1853, <https://doi.org/10.1175/JTECH-D-12-00240.1>, 2013.

Lindsay, R. W.: Ice deformation near SHEBA, *Journal of Geophysical Research: Oceans*, 107(C10), 8042, <https://doi.org/10.1029/2000JC000445>, 2002.

Lindsay, R.W., and Stern, H.L.: The RADARSAT geophysical processor system: Quality of sea ice trajectory and deformation estimates. *Journal of Atmospheric and Oceanic Technology*, 20(9), 1333-1347. [https://doi.org/10.1175/1520-0426\(2003\)020<1333:TRGPSQ>2.0.CO;2](https://doi.org/10.1175/1520-0426(2003)020<1333:TRGPSQ>2.0.CO;2), 2003.

Lindsay R.W., Zhang, J., and Rothrock D. A.: Sea-ice deformation rates from satellite measurements and in a model, *Atmosphere-Ocean*, 41:1, 35-47. <https://doi.org/10.3137/ao.410103>, 2003.

Liu, P.C., and Miller, G.S.: Wavelet transforms and ocean current data analysis. *Journal of atmospheric and oceanic technology*, 13(5), 1090-1099. [https://doi.org/10.1175/1520-0426\(1996\)013<1090:WTAOCD>2.0.CO;2](https://doi.org/10.1175/1520-0426(1996)013<1090:WTAOCD>2.0.CO;2), 1996.

Liu, A.K., and Mollo-Christensen, E.: Wave propagation in a solid ice pack. *Journal of physical oceanography*, 18(11), 1702-1712. [https://doi.org/10.1175/1520-0485\(1988\)018<1702:WPIASI>2.0.CO;2](https://doi.org/10.1175/1520-0485(1988)018<1702:WPIASI>2.0.CO;2), 1988.

Lu, F., Konecny, M., Chen, M., and Reznik, T.: A Barotropic Tide Model for Global Ocean Based on Rotated Spherical Longitude-Latitude Grids. *Water*, 13(19), 2670. <https://doi.org/10.3390/w13192670>, 2021.

Lukovich, J.V., Babb, D.G., and Barber, D.G.: On the scaling laws derived from ice beacon trajectories in the southern Beaufort Sea during the International Polar Year-Circumpolar Flaw Lead study, 2007–2008. *Journal of Geophysical Research: Oceans*, 116(C9). <https://doi.org/10.1029/2011JC007049>, 2011.

Lukovich, J.V., Hutchings, J.K., and Barber, D.G.: On sea ice dynamical regimes in the Arctic Ocean. *Annals of Glaciology*, 56(69), 323-331. <https://doi.org/10.3189/2015AoG69A606>, 2015.

Lukovich, J.V., Geiger, C.A., and Barber, D.G.: Method to characterize directional changes in Arctic sea ice drift and associated deformation due to synoptic atmospheric variations using Lagrangian dispersion statistics. *The Cryosphere*, 11(4), 1707-1731. <https://doi.org/10.5194/tc-11-1707-2017>, 2017.

Lukovich, J.V., Jafarikhasragh, S., Tefs, A., Myers, P.G., Sydor, K., Wong, K., Stroeve, J.C., Stadnyk, T.A., Babb, D., and Barber, D.G. A baseline evaluation of oceanographic and sea ice conditions in the Hudson Bay Complex during 2016–2018. *Elementa: Science of the Anthropocene*, 9(1), 00128. <https://doi.org/10.1525/elementa.2020.00128>, 2021.

Lund, B., Graber, H.C., Persson, P.O.G., Smith, M., Doble, M., Thomson, J., and Wadhams, P.: Arctic sea ice drift measured by shipboard marine radar. *Journal of Geophysical Research: Oceans*, 123(6), 4298-4321. <https://doi.org/10.1029/2018JC013769>, 2018.

Machutchon, K., Vichi, M., Eayrs, C., Alberello, A., Babanin, A., Bekker, A., Bennetts, L., Holland, D., De Jong, E., Joubert, W., Messori, G., Mojica, J., Onorato, M., Saunders, C., Skatulla, S., and Toffoli, A.: SA Agulhas II Winter 2017 Cruise: ice edge drifters, Ver. 1, Australian Antarctic Data Centre [data set]. <https://doi.org/10.26179/5cc937513bd6d>, 2019.

Maksym, T.: Arctic and Antarctic Sea ice change: contrasts, commonalities, and causes. *Annual Review of Marine Science*, 11, 187-213. <https://doi.org/10.1146/annurev-marine-010816-060610>, 2019.

- Marquart, R., Bogaers, A., Skatulla, S., Alberello, A., Toffoli, A., Schwarz, C., and Vichi, M.: A computational fluid dynamics model for the small-scale dynamics of wave, ice floe and interstitial grease ice interaction. *Fluids*, 6(5), 176. <https://doi.org/10.3390/fluids6050176>. 2021.
- Marsan, D., Stern, H., Lindsay, R., and Weiss, J.: Scale dependence and localization of the deformation of Arctic sea ice. *Physical review letters*, 93(17), 178501. <https://doi.org/10.1103/PhysRevLett.93.178501>, 2004.
- Martin, J.E.: Mid-latitude atmospheric dynamics: a first course. John Wiley & Sons. 2013.
- Martin, P.A., and Dalrymple, R.A.: On amphidromic points. Proceedings of the Royal Society of London. Series A, *Mathematical and Physical Sciences*, 444(1920), 91-104. <https://doi.org/10.1098/rspa.1994.0006>, 1994.
- Martin, S., and Thorndike, A.S.: Dispersion of sea ice in the Bering Sea. *Journal of Geophysical Research: Oceans*, 90(C4), 7223-7226. <https://doi.org/10.1029/JC090iC04p07223>, 1985.
- Martin, T.: Arctic sea ice dynamics: drift and ridging in numerical models and observations (Vol. 563). Alfred-Wegener-Institut für Polar-und Meeresforschung.: Arctic sea ice dynamics: drift and ridging in numerical models and observations (Vol. 563). Alfred-Wegener-Institut für Polar-und Meeresforschung. 2007.
- Massom, R.A., Scambos, T.A., Bennetts, L.G., Reid, P., Squire, V.A., and Stammerjohn, S.E.: Antarctic ice shelf disintegration triggered by sea ice loss and ocean swell. *Nature*, 558(7710), pp.383-389. <https://doi.org/10.1038/s41586-018-0212-1>, 2018.
- Massom, R.A., and Stammerjohn, S.E.: Antarctic sea ice change and variability—physical and ecological implications. *Polar Science*, 4(2), 149–186. [10.1016/j.polar.2010.05.001](https://doi.org/10.1016/j.polar.2010.05.001), 2010.
- Matear, R.J., O’Kane, T.J., Risbey, J.S., and Chamberlain, M.: Sources of heterogeneous variability and trends in Antarctic sea-ice. *Nature communications*, 6(1), 8656. <https://doi.org/10.1038/ncomms9656>, 2015.
- Mayhew, S.: A dictionary of geography. OUP Oxford, United Kingdom. 2015.
- Maxwell, J.B. and Barrie, L.A.: Atmospheric and climatic change in the Arctic and Antarctic. *Ambio*, 42-49. 1989.
- McPhee, M.G.: An analysis of pack ice drift in summer. Sea ice processes and models, 62-75. 1980.

McPhee, M.G.: Analysis and prediction of short-term ice drift. 110(1): 94-100. <https://doi.org/10.1115/1.3257130>, 1988.

McPhee, M.G.: Sea Ice. Edition 3. Edited by David N. Thomas. Chichester: John Wiley and Sons Ltd. 2017.

Meehl, G.A., Arblaster, J.M., Chung, C.T., Holland, M.M., DuVivier, A., Thompson, L., Yang, D., and Bitz, C.M.: Sustained ocean changes contributed to sudden Antarctic sea ice retreat in late 2016. *Nature communications*, 10(1), 14. <https://doi.org/10.1038/s41467-018-07865-9>, 2019.

Meier, W.N., Fetterer, F., Stewart, J.S., and Helfrich, S.: How do sea-ice concentrations from operational data compare with passive microwave estimates? Implications for improved model evaluations and forecasting. *Annals of Glaciology*, 56(69), pp.332-340. <https://doi.org/10.3189/2015AoG69A694>, 2015.

Meier, W.N., and Stewart, J.S.: Assessing uncertainties in sea ice extent climate indicators. *Environmental Research Letters*, 14(3), p.035005. <https://doi.org/10.1088/1748-9326/aaf52c>, 2019.

Melsheimer, C., Spreen, G., Ye, Y., and Shokr, M.: First results of Antarctic sea ice type retrieval from active and passive microwave remote sensing data, *The Cryosphere*, 17, 105–126, <https://doi.org/10.5194/tc-17-105-2023>, 2023.

Meredith, M.P., and Brandon, M.A.: Oceanography and sea ice in the Southern Ocean. Ch. 8 in *Sea ice*, 216-238. <https://doi.org/10.1002/9781118778371.ch8>, 2017.

Messori, G., Woods, C., and Caballero, R.: On the drivers of wintertime temperature extremes in the High Arctic. *Journal of Climate*, 31(4), 1597–1618. <https://doi.org/10.1175/JCLI-D-17-0386.1>, 2017.

Meylan, M.H., Bennetts, L.G., and Kohout, A.L.: In-situ measurements and analysis of ocean waves in the Antarctic marginal ice zone. *Geophysical Research Letters*, 41(14), 5046-5051. <https://doi.org/10.1002/2014GL060809>, 2014.

Monteban, D., Johnsen, H., and Lubbad, R.: Spatiotemporal Observations of Wave Dispersion Within Sea Ice Using Sentinel-1 SAR TOPS Mode. *Journal of Geophysical Research: Oceans*, 124(12), 8522-8537. <https://doi.org/10.1029/2019JC015311>, 2019.

Montiel, F., Kohout, A.L., and Roach, L.A.: Physical drivers of ocean wave attenuation in the marginal ice zone. *Journal of Physical Oceanography*, 52(5), 889-906. <https://doi.org/10.1175/JPO-D-21-0240.1>, 2022.

Montiel, F., Squire, V.A., and Bennetts, L.G.: Attenuation and directional spreading of ocean wave spectra in the marginal ice zone. *Journal of Fluid Mechanics*, 790, 492-522. <https://doi.org/10.1017/jfm.2016.21>, 2016.

Montiel, F., and Squire, V.A.: Modelling wave-induced sea ice break-up in the marginal ice zone. *Proceedings of the Royal Society A: Mathematical, Physical and Engineering Sciences*, 473(2206), 20170258. <https://doi.org/10.1098/rspa.2017.0258>, 2017.

Montiel, F., Squire, V.A., Doble, M., Thomson, J., and Wadhams, P.: Attenuation and directional spreading of ocean waves during a storm event in the autumn Beaufort Sea marginal ice zone. *Journal of Geophysical Research: Oceans*, 123(8), 5912-5932. <https://doi.org/10.1029/2018JC013763>, 2018.

Murphy, B.F., and Simmonds, I.: An analysis of strong wind events simulated in a GCM near Casey in the Antarctic. *Monthly weather review*, 121(2), 522-534. [https://doi.org/10.1175/1520-0493\(1993\)121<0522:AAOSWE>2.0.CO;2](https://doi.org/10.1175/1520-0493(1993)121<0522:AAOSWE>2.0.CO;2), 1993.

Nakayama, Y., Ohshima, K.I., and Fukamachi, Y.: Enhancement of sea ice drift due to the dynamical interaction between sea ice and a coastal ocean. *Journal of Physical Oceanography*, 42(1), 179–192. <https://doi.org/10.1175/JPO-D-11-018.1>, 2012.

Nansen, F.: Oceanography of the North Polar Basin. *Scientific Results*, 3(9), 1–427. 1902.

Nelli, F., Derkani, M.H., Alberello, A., and Toffoli, A.: A satellite altimetry data assimilation approach to optimise sea state estimates from vessel motion. *Applied Ocean Research*, 132, p.103479. <https://doi.org/10.1016/j.apor.2023.103479>, 2023.

Nelli, F., Van Zuydam, A., Pferdekamper, K., Alberello, A., Derkani, M., Bekker, A., and Toffoli, A.: Reconstructing sea-states in the Southern Ocean using ship motion data. In International Conference on Offshore Mechanics and Arctic Engineering, 85161, V006T06A027. *American Society of Mechanical Engineers*. <https://doi.org/10.1016/j.apor.2023.103479>, 2021.

Newman, J.N.: Marine hydrodynamics (p. 448). The MIT press. Cambridge. 2018.

Nielsen, U.D., and Dietz, J.: Estimation of sea state parameters by the wave buoy analogy with comparisons to third generation spectral wave models. *Ocean Engineering*, 216, 107781. <https://doi.org/10.1016/j.oceaneng.2020.107781>, 2020.

Nielsen, U.D., and Jensen, J.J.: A novel approach for navigational guidance of ships using onboard monitoring systems. *Ocean Engineering*, 38(2-3), 444-455. <https://doi.org/10.1016/j.oceaneng.2010.11.024>, 2011.

NSIDC. SMMR, SSM/I, and SSMIS Sensors: <https://nsidc.org/data/nsidc-0051/versions/1>, last access 27 February 2023.

Oikkonen, A., Haapala, J., Lensu, M., Karvonen, J., and Itkin, P.: Small-scale sea ice deformation during N-ICE 2015: From compact pack ice to marginal ice zone. *Journal of Geophysical Research: Oceans*, 122(6), 5105-5120. <https://doi.org/10.1002/2016JC012387>, 2017.

Orsi, A.H., Whitworth III, T., and Nowlin Jr, W.D.: On the meridional extent and fronts of the Antarctic Circumpolar Current. *Deep Sea Research Part I: Oceanographic Research Papers*, 42(5), 641-673. [https://doi.org/10.1016/0967-0637\(95\)00021-W](https://doi.org/10.1016/0967-0637(95)00021-W), 1995.

Overland, J., Turner, J., Francis, J., Gillett, N., Marshall, G. and Tjernström, M.: The Arctic and Antarctic: Two faces of climate change. *Eos, Transactions American Geophysical Union*, 89(19), 177-178. <https://doi.org/10.1002/joc.1406>, 2008.

Pang, X., Pu, J., Zhao, X., Ji, Q., Qu, M., and Cheng, Z.: Comparison between AMSR2 sea ice concentration products and pseudo-ship observations of the Arctic and Antarctic sea ice edge on cloud-free days. *Remote Sensing*, 10(2), 317. <https://doi.org/10.3390/rs10020317>, 2018.

Parkinson, C. L. and Cavalieri, D. J.: Antarctic sea ice variability and trends, 1979–2010, *The Cryosphere*, 6, 871–880, <https://doi.org/10.5194/tc-6-871-2012>, 2012.

Papritz, L., Pfahl, S., Rudeva, I., Simmonds, I., Sodemann, H., and Wernli, H.: The role of extratropical cyclones and fronts for Southern Ocean freshwater fluxes. *Journal of Climate*, 27(16), 6205-6224. <https://doi.org/10.1175/JCLI-D-13-00409.1>, 2014.

Passerotti, G., Bennetts, L.G., von Bock und Polach, F., Alberello, A., Puolakka, O., Dolatshah, A., Monbaliu, J., and Toffoli, A.: Interactions between irregular wave fields and sea ice: A physical model for wave attenuation and ice breakup in an ice tank. *Journal of Physical Oceanography*, 52(7), 1431-1446. <https://doi.org/10.1175/JPO-D-21-0238.1>. 2022.

Passow, U., and Carlson, C.A.: The biological pump in a high co₂ world. *Marine Ecology Progress Series*, 470, 249–271, 2012.

Peixoto, J.P., and Oort, A.H.: The physics of climate. American Institute of Physics, New York, 1992.

Petrich, C., and Eicken, H.: Sea Ice. Edition 3. Edited by David N. Thomas. Chichester: John Wiley and Sons Ltd. 2017.

Pezza, A.B., Rashid, H.A., and Simmonds, I.: Climate links and recent extremes in Antarctic sea ice, high-latitude cyclones, Southern Annular Mode and ENSO. *Climate dynamics*, 38, 57-73. <https://doi.org/10.1007/s00382-011-1044-y>, 2012.

Pope, J. O., Holland, P. R., Orr, A., Marshall, G. J., and Phillips, T.: The impacts of El Niño on the observed sea ice budget of West Antarctica. *Geophysical Research Letters*, 44, 6200–6208. <https://doi.org/10.1002/2017GL073414>, 2016.

Purich, A., and Doddridge, E.W.: Record low Antarctic sea ice coverage indicates a new sea ice state. *Commun Earth Environ* 4, 314. <https://doi.org/10.1038/s43247-023-00961-9>, 2023.

Qihe, L.: Integration of vibration acceleration signal based on labview. *Journal of physics: conference series*, 1345(4), 042067. *IOP Publishing*. doi: 10.1088/1742-6596/1345/4/042067, 2019.

Rampal, P., Weiss, J., Marsan, D., Lindsay, R., and Stern, H.: Scaling properties of sea ice deformation from buoy dispersion analysis. *Journal of Geophysical Research: Oceans*, 113(C3). <https://doi.org/10.1029/2007JC004143>, 2008.

Rampal, P., Weiss, J., and Marsan, D.: Positive trend in the mean speed and deformation rate of Arctic sea ice, 1979–2007. *Journal of Geophysical Research: Oceans*, 114(C5). <https://doi.org/10.1029/2008JC005066>, 2009.

Rapley, C.: First observations of the interaction of ocean swell with sea ice using satellite radar altimeter data. *Nature*, 307, 150–152. <https://doi.org/10.1038/307150a0>, 1984.

Remya G., Kumar, R., Basu, S., and Sarkar, A: Altimeter-derived ocean wave period using genetic algorithm. *IEEE Geoscience and Remote Sensing Letters*, 8(2), 354–358, doi: 10.1109/LGRS.2010.2075911, 2010.

Reiser, F., Willmes, S., Hausmann, U., and Heinemann, G.: Predominant Sea Ice Fracture Zones Around Antarctica and Their Relation to Bathymetric Features. *Geophysical Research Letters*, 46(21), 12117-12124. <https://doi.org/10.1029/2019GL084624>, 2019.

Ribal, A., and Young, I.R.: 33 years of globally calibrated wave height and wind speed data based on altimeter observations. *Scientific data*, 6(1), 77. <https://doi.org/10.1038/s41597-019-0083-9>, 2019.

Rio, M.H., Mulet, S., and Picot, N.: Beyond GOCE for the ocean circulation estimate: Synergetic use of altimetry, gravimetry, and in-situ data provides new insight into geostrophic and Ekman currents. *Geophysical Research Letters*, 41(24), 8918-8925. <https://doi.org/10.1002/2014GL061773>, 2014.

- Roach, L.A., Horvat, C., Dean, S.M., and Bitz, C.M.: An emergent sea ice floe size distribution in a global coupled ocean-sea ice model. *Journal of Geophysical Research: Oceans*, 123(6), 4322-4337. <https://doi.org/10.1029/2017JC013692>, 2018.
- Robin, G.D.Q.: Ocean waves and pack ice. *Polar Record*, 11(73), 389-393. <https://doi.org/10.1017/S003224740005350X>, 1963.
- Rogers, W.E., Meylan, M.H., and Kohout, A.L.: Estimates of spectral wave attenuation in Antarctic sea ice, using model/data inversion. *Cold Regions Science and Technology*, 182, p.103198. <https://doi.org/10.1016/j.coldregions.2020.103198>, 2021.
- Rolph, R.J., Feltham, D.L., and Schröder, D.: Changes of the Arctic marginal ice zone during the satellite era. *The Cryosphere*, 14(6), 1971-1984. <https://doi.org/10.5194/tc-14-1971-2020>, 2020.
- Rossi, G.B., Cannata, A., Iengo, A., Migliaccio, M., Nardone, G., Piscopo, V., and Zambianchi, E.: Measurement of sea waves. *Sensors*, 22(1), 78. <https://doi.org/10.3390/s22010078>, 2021.
- Ryan-Keogh, T. and Vichi, M.: SCALE-WIN19 & SCALE-SPR19 Cruise Report, Southern Ocean Seasonal Cycle Experiment, South Africa, 339 pp, <https://doi.org/10.5281/zenodo.5906324>, 2022.
- Salles, T.: RADWave – Wave analysis from Altimeter data, GitHub [code], <https://radwave.readthedocs.io/en/latest/index.html>, 2020.
- Schemm, S.: Regional trends in weather systems help explain Antarctic sea ice trends. *Geophysical Research Letters*, 45(14), 7165-7175. <https://doi.org/10.1029/2018GL079109>, 2018.
- Schlosser, E., Haumann, F.A., and Raphael, M.N.: Atmospheric influences on the anomalous 2016 Antarctic sea ice decay. *The Cryosphere*, 12(3), 1103-1119. <https://doi.org/10.5194/tc-12-1103-2018>, 2018.
- Schweiger, A.J., and Zhang, J.: Accuracy of short-term sea ice drift forecasts using a coupled ice-ocean model. *Journal of Geophysical Research: Oceans*, 120(12), 7827-7841. <https://doi.org/10.1002/2015JC011273>, 2015.
- Screen, J.A.: Sudden increase in Antarctic sea ice: Fact or artifact? *Geophysical Research Letters*, 38(13). <https://doi.org/10.1029/2011GL047553>, 2011.
- Shaw, T.A., Baldwin, M., Barnes, E.A., Caballero, R., Garfinkel, C.I., Hwang, Y.T., Li, C., O'gorman, P.A., Rivière, G., Simpson, I.R. and Voigt, A.: Storm track processes and the opposing influences of climate change. *Nature Geoscience*, 9(9), 656-664. <https://doi.org/10.1038/ngeo2783>, 2016.

Shen, H.H.: Wave-in-ice: theoretical bases and field observations. *Philosophical Transactions of the Royal Society A*, 380(2235), p.20210254. <https://doi.org/10.1098/rsta.2021.0254>, 2022.

Shen, H.H., and Ackley, S.F.A.: A one-dimensional model for wave-induced ice-floe collisions. *Annals of Glaciology*, 15(1), 87–95. <https://doi.org/10.3189/1991AoG15-1-87-95>, 1991.

Shen, H.H., Ackley, S.F., and Hopkins, M.A.: A conceptual model for pancake-ice formation in a wave field. *Annals of Glaciology*, 33, 361-367. <https://doi.org/10.3189/172756401781818239>, 2001.

Shen, H.H., Perrie, W., Hu., Y., and He., Y.: Remote sensing of waves propagating in the marginal ice zone by SAR. *Journal of Geophysical Research: Oceans*, 123(1), 189-200. <https://doi.org/10.1002/2017JC013148>, 2018.

Shestov, A.: Wave attenuation through pancake ice. Measurements from a moving vessel in the Barents Sea in 2017. In *Proceedings of the International Conference on Port and Ocean Engineering Under Arctic Conditions*. 2019.

Simmonds, I., 2015. Comparing and contrasting the behaviour of Arctic and Antarctic sea ice over the 35 year period 1979-2013. *Annals of Glaciology*, 56(69), 18-28. <https://doi.org/10.3189/2015AoG69A909>, 2015.

Skatulla, S., Audh, R. R., Cook, A., Hepworth, E., Johnson, S., Lupascu, D. C., MacHutchon, K., Marquart, R., Mielke, T., Omatuku, E., Paul, F., Rampai, T., Schröder, J., Schwarz, C., and Vichi, M.: Physical and mechanical properties of winter first-year ice in the Antarctic marginal ice zone along the Good Hope Line, *The Cryosphere*, 16, 2899–2925, <https://doi.org/10.5194/tc-16-2899-2022>, 2022.

Spreen, G., Kaleschke, L., and Heygster, G.: Sea ice remote sensing using AMSR-E 89-GHz channels. *Journal of Geophysical Research: Oceans*, 113, C2. <https://doi.org/10.1029/2005JC003384>, 2008.

Soal, K., Bienert, J. and Bekker, A.: Operational modal analysis on the polar supply and research vessel the SA Agulhas II. In *The 6th International Operational Modal Analysis Conference*. 2015.

Squire, V.A.: Of ocean waves and sea-ice revisited. *Cold Regions Science and Technology*, 49(2), 110-133. <https://doi.org/10.1016/j.coldregions.2007.04.007>, 2007.

Squire, V.A.: A fresh look at how ocean waves and sea ice interact. *Philosophical Transactions of the Royal Society A: Mathematical, Physical and Engineering Sciences*, 376(2129), 20170342. <https://doi.org/10.1098/rsta.2017.0342>, 2018.

Squire, V.A.: Ocean wave interactions with sea ice: A reappraisal. *Annual Review of Fluid Mechanics*, 52, 37-60. <https://doi.org/10.1146/annurev-fluid-010719-060301>, 2020.

Squire, V.A., and Montiel, F.: Evolution of directional wave spectra in the marginal ice zone: a new model tested with legacy data. *Journal of Physical Oceanography*, 46(10), 3121-3137. <https://doi.org/10.1175/JPO-D-16-0118.1>, 2016.

Squire, V.A., Dugan, J.P., Wadhams, P., Rottier, P.J., and Liu, A.K.: Of ocean waves and sea ice. *Annual Review of Fluid Mechanics*, 27(1), 115–168. <https://doi.org/10.1146/annurev.fl.27.010195.000555>, 1995.

Stern, H. L., and Lindsay, R. W.: Spatial scaling of Arctic sea ice deformation, *Journal of Geophysical Research: Oceans*, 114, C10017. <https://doi.org/10.1029/2009JC005380>, 2009.

Stopa, J.E., Sutherland, P. and Ardhuin, F.: Strong and highly variable push of ocean waves on Southern Ocean sea ice. *Proceedings of the National Academy of Sciences*, 115(23), 5861-5865. <https://doi.org/10.1073/pnas.1802011115>, 2018.

Strong, C., Foster, D., Cherkaev, E., Eisenman, I., and Golden, K. M.: On the definition of marginal ice zone width. *Journal of Atmospheric and Oceanic Technology*, 34(7), 1565–1584. <https://doi.org/10.1175/JTECH-D-16-0171.1>, 2017.

Stroeve, J.C., Crawford, A.D., and Stammerjohn, S., 2016. Using timing of ice retreat to predict timing of fall freeze-up in the Arctic. *Geophysical Research Letters*, 43(12), 6332-6340. <https://doi.org/10.1002/2016GL069314>, 2016.

Swart, S., du Plessis, M.D., Thompson, A.F., Biddle, L.C., Giddy, I., Linders, T., Mohrmann, M., and Nicholson, S.A.: Submesoscale fronts in the Antarctic marginal ice zone and their response to wind forcing. *Geophysical Research Letters*, 47(6), p.e2019GL086649. <https://doi.org/10.1029/2019GL086649>, 2020.

Tandon, N.F., Kushner, P.J., Docquier, D., Wettstein, J.J., and Li, C.: Reassessing sea ice drift and its relationship to long-term Arctic sea ice loss in coupled climate models. *Journal of Geophysical Research: Oceans*, 123(6), 4338–4359. <https://doi.org/10.1029/2017JC013697>, 2018.

Taylor, G.I.: Diffusion by continuous movements. *Proceedings of the London Mathematical Society*, 2(1), 196–212. <https://doi.org/10.1112/plms/s2-20.1.196>, 1922.

Taylor, N.C., Bekker, A., Kruger., Ship motion measurements on the SA Agulhas II - Winter Cruise 2019, SUNScholarData [data set]. <https://doi.org/10.25413/sun.24324799>, 2023a.

Taylor, NC, Bekker, A, Kruger, K. Ship motion measurements on the SA Agulhas II - Spring Cruise 2019, SUNScholarData [data set]. <https://doi.org/10.25413/sun.24325055>, 2023b.

Tersigni, I., Alberello, A., Messori, G., Vichi, M., Onorato, M., and Toffoli, A.: High-resolution thermal imaging in the Antarctic marginal ice zone: Skin temperature heterogeneity and effects on heat fluxes. *Earth and Space Science*. <https://doi.org/10.1029/2023EA003078> 2023.

Thomson, J.: Wave breaking dissipation observed with SWIFT drifters. *Journal of Atmospheric and Oceanic Technology*. 29(12), 1866-1882. <https://doi.org/10.1175/JTECH-D-12-00018.1>, 2012.

Thomson, J.: Wave propagation in the marginal ice zone: connections and feedback mechanisms within the air-ice-ocean system. *Philosophical Transactions of the Royal Society A*, 380(2235), p.20210251. <https://doi.org/10.1098/rsta.2021.0251>, 2022.

Thomson, J., Biddle, L.C, Wahlgren, S., and Swart, S.: Direct observations of wave-sea ice interactions in the Antarctic Marginal Ice Zone, Zenodo [data set], <https://doi.org/10.5281/zenodo.7845764>, 2023.

Thomson, J., Moulton, M., de Klerk, A., Talbert, J., Guerra, M., Kastner, S., Smith, M., Schwendeman, M., Zippel, S., and Nylund, S.: A new version of the SWIFT platform for waves, currents, and turbulence in the ocean surface layer. In 2019 IEEE/OES twelfth current, waves and turbulence measurement (CWTMW). 1-7. doi: 10.1109/CWTM43797.2019.8955299, 2019.

Thomson, J., Talbert, J., de Klerk, A., Brown, A., Schwendeman, M., Goldsmith, J., Thomas, J., Olfe, C., Cameron, G., Meinig, C.: Biofouling effects on the response of a wave measurement buoy in deep water. *Journal of Atmospheric and Oceanic Technology*, 32(6), 1281–1286. <https://doi.org/10.1175/JTECH-D-15-0029.1>, 2015.

Thorndike, A.S.: Kinematics of sea ice. In The geophysics of sea ice (pp. 489-549). Boston, MA: Springer US. 1986.

Thorndike, A.S., and Colony, R.: Sea ice motion in response to geostrophic winds. *Journal of Geophysical Research: Oceans*, 87(C8), 5845-5852. <https://doi.org/10.1029/JC087iC08p05845>, 1982.

Thorndike, A.S., Rothrock, D.A., Maykut, G.A. and Colony, R.: The thickness distribution of sea ice. *Journal of Geophysical Research*, 80(33), 4501-4513. <https://doi.org/10.1029/JC080i033p04501>, 1975.

Toffoli, A., Bennetts, L.G., Meylan, M.H., Cavaliere, C., Alberello, A., Elsnab, J., and Monty, J.P.: Sea ice floes dissipate the energy of steep ocean waves. *Geophysical Research Letters*, 42(20), 8547-8554. <https://doi.org/10.1002/2015GL065937>, 2015.

- Torrence, C., and Compo, G.P.: A practical guide to wavelet analysis. *Bulletin of the American Meteorological society*, 79(1), 61-78. [https://doi.org/10.1175/1520-0477\(1998\)079<0061:APGTWA>2.0.CO;2](https://doi.org/10.1175/1520-0477(1998)079<0061:APGTWA>2.0.CO;2), 1998.
- Turner, J., Comiso, J.C., Marshall, G.J., Lachlan-Cope, T.A., Bracegirdle, T., Maksym, T., Meredith, M.P., Wang, Z., and Orr, A.: Non-annular atmospheric circulation change induced by stratospheric ozone depletion and its role in the recent increase of Antarctic sea ice extent. *Geophysical research letters*, 36(8). <https://doi.org/10.1029/2009GL037524>, 2009.
- Turner, J., Hosking, J.S., Phillips, T., and Marshall, G.J.: Temporal and spatial evolution of the Antarctic sea ice prior to the September 2012 record maximum extent. *Geophysical Research Letters*, 40(22), 5894-5898. <https://doi.org/10.1002/2013GL058371>, 2013.
- Turner, J., Hosking, J.S., Bracegirdle, T.J., Marshall, G.J., and Phillips, T.: Recent changes in Antarctic sea ice. *Philosophical Transactions of the Royal Society A: Mathematical, Physical and Engineering Sciences*, 373(2045), 20140163. <https://doi.org/10.1098/rsta.2014.0163>, 2015.
- Turner, J., Hosking, J. S., Marshall, G. J., Phillips, T., and Bracegirdle, T. J.: Antarctic sea ice increase consistent with intrinsic variability of the Amundsen Sea Low. *Climate Dynamics*, 46, 2391– 2402. <https://doi.org/10.1007/s00382-015-2708-9>, 2016.
- Uotila, J., Vihma, T. and Launiainen, J.: Response of the Weddell Sea pack ice to wind forcing. *Journal of Geophysical Research: Oceans*, 105(C1), 1135-1151. <https://doi.org/10.1029/1999JC900265>, 2000.
- Uotila, P., Vihma, T., Pezza, A.B., Simmonds, I., Keay, K., and Lynch AH.: Relationships between Antarctic cyclones and surface conditions as derived from high-resolution numerical weather prediction data. *Journal of Geophysical Research: Atmospheres* 116. <https://doi.org/10.1029/2010JD015358>, 2011.
- Van Den Broeke, M.R.: The semi-annual oscillation and Antarctic climate. Part 1: Influence on near surface temperatures (1957–79). *Antarctic Science*, 10(2), 175-183. <https://doi.org/10.1017/S0954102098000248>, 1998.
- Vaughan, D.G., Comiso, J.C., Allison, I., Carrasco, J., Kaser, G., Kwok, R., Mote, P., Murray, T., Paul, F., Ren, J., and Rignot, E.: Observations: cryosphere. *Climate change*, 2103, 317-382. 2013.
- Vichi, M.: An indicator of sea ice variability for the Antarctic marginal ice zone, *The Cryosphere*, 16, 4087–4106, <https://doi.org/10.5194/tc-16-4087-2022>, 2022.

Vichi, M., Eayrs, C., Alberello, A., Bekker, A., Bennetts, L., Holland, D., de Jong, E., Joubert, W., MacHutchon, K., Messori, G., and Mojica, J.F. Effects of an explosive polar cyclone crossing the Antarctic marginal ice zone. *Geophysical Research Letters*, 46(11), 5948-5958. <https://doi.org/10.1029/2019GL082457>, 2019.

Vihma, T., and Launiainen, J.: Ice drift in the Weddell Sea in 1990-1991 as tracked by a satellite buoy. *Journal of Geophysical Research: Oceans*, 98, 14471-14485. <https://doi.org/10.1029/93JC00649>, 1993.

Vihma, T., Launiainen, J., and Uotila, J.: Weddell Sea ice drift: Kinematics and wind forcing. *Journal of Geophysical Research: Oceans*, 101(C8), 18279-18296. <https://doi.org/10.1029/96JC01441>, 1996.

Voermans, J.J., Babanin, A.V., Thomson, J., Smith, M.M., and Shen, H.H.: Wave attenuation by sea ice turbulence. *Geophysical Research Letters*, 46(12), 6796-6803. <https://doi.org/10.1029/2019GL082945>, 2019.

Wadhams, P.: Ice in the Ocean. CRC Press. London. <https://doi.org/10.1201/9781482283082>, 2000.

Wadhams, P.: The seasonal ice zone. In N. Untersteiner (Ed.), The geophysics of sea ice. Boston, MA: Springer US. 825–991. doi: 10.1007/978-1-4899-5352-0_15, 1986.

Wadhams, P.: The underside of Arctic sea ice imaged by sidescan sonar. *Nature*, 333(6169), 161-164. <https://doi.org/10.1038/333161a0>, 1988.

Wadhams, P., Aulicino, G., Parmiggiani, F., Persson, P.O.G., and Holt, B.: Pancake ice thickness mapping in the Beaufort Sea from wave dispersion observed in SAR imagery. *Journal of Geophysical Research: Oceans*, 123(3), 2213-2237. <https://doi.org/10.1002/2017JC013003>, 2018.

Wadhams, P., Gill, A.E., and Linden, P.F.: Transects by submarine of the East Greenland polar front. *Deep Sea Research Part A. Oceanographic Research Papers*, 26(12), 1311-1327. [https://doi.org/10.1016/0198-0149\(79\)90001-3](https://doi.org/10.1016/0198-0149(79)90001-3), 1979.

Wahlgren, S., Thomson, J., Biddle, L.C., and Swart, S.: Direct Observations of Wave-Sea Ice Interactions in the Antarctic Marginal Ice Zone. *Journal of Geophysical Research: Oceans*, 128, e2023JC019948. <https://doi.org/10.1029/2023JC019948>, 2023.

Wang, S., Bailey, D., Lindsay, K., Moore, J. K., and Holland, M.: Impact of sea ice on the marine iron cycle and phytoplankton productivity, *Biogeosciences*, 11, 4713–4731, <https://doi.org/10.5194/bg-11-4713-2014>, 2014.

- Wang, Q., Danilov, S., Mu, L., Sidorenko, D., and Wekerle, C.: Lasting impact of winds on Arctic sea ice through the ocean's memory, *The Cryosphere*, 15, 4703–4725, <https://doi.org/10.5194/tc-15-4703-2021>, 2021.
- Wang, R., and Shen, H.H.: Gravity waves propagating into an ice-covered ocean: A viscoelastic model. *Journal of Geophysical Research: Oceans*, 115(C6). <https://doi.org/10.1029/2009JC005591>, 2010.
- Waseda, T., Alberello, A., Nose, T., Toyota, T., Kodaira, T., and Fujiwara, Y.: Observation of anomalous spectral downshifting of waves in the Okhotsk Sea Marginal Ice Zone. *Philosophical Transactions of the Royal Society A*, 380(2235), p.20210256. <https://doi.org/10.1098/rsta.2021.0256>, 2022.
- Wassermann, S., Schmitt, C., Kottmeier, C. and Simmonds, I.: Coincident vortices in Antarctic wind fields and sea ice motion. *Geophysical research letters*, 33(15). <https://doi.org/10.1029/2006GL026005>, 2006.
- Watkins, A., and Simmonds, I.: Relationships between Antarctic sea-ice concentration, wind stress and temperature temporal variability, and their changes with distance from the coast. *Annals of Glaciology*, 27, 409–412. <https://doi.org/10.3189/1998AoG27-1-409-412>, 1998.
- Weeks, W.F.S., and Ackley, S.F.: The growth, structure, and properties of sea ice. In N. Untersteiner (Ed.), *The geophysics of sea ice*: 146, edited by: Thomas, D.N., and Dieckmann, G.S., Springer Berlin, Heidelberg, Germany 9–164. <https://doi.org/10.1002/9781444317145.ch2>, 1986.
- Wei, L., and Qin, T.: Characteristics of cyclone climatology and variability in the Southern Ocean. *Acta Oceanologica Sinica*, 35(7), 59–67. <https://doi.org/10.1007/s13131-016-0913-y>, 2016.
- Weiss, J.: Sea Ice Deformation. In *Drift, Deformation, and Fracture of Sea Ice*. Springer Briefs in Earth Sciences. Springer, Dordrecht, 31–51. https://doi.org/10.1007/978-94-007-6202-2_3, 2013.
- Weiss, J., and Dansereau, V.: Linking scales in sea ice mechanics. *Philosophical Transactions of the Royal Society A: Mathematical, Physical and Engineering Sciences*, 375(2086), 20150352. <https://doi.org/10.1098/rsta.2015.0352>, 2017.
- Weiss, J.: Drift, deformation, and fracture of sea ice: a perspective across scales. Dordrecht: Springer. 1–83, <https://doi.org/10.1007/978-94-007-6202-2>, 2013.
- Weiss, J., and Marsan, D.: Scale properties of sea ice deformation and fracturing. *Comptes Rendus Physique*, 5(7), 735–751. <https://doi.org/10.1016/j.crhy.2004.09.005>, 2004.

- Wilkinson, J.P., and Wadhams, P.: A salt flux model for salinity change through ice production in the Greenland Sea, and its relationship to winter convection. *Journal of Geophysical Research: Oceans*, 108(C5). <https://doi.org/10.1029/2001JC001099>, 2003.
- Woods, C. and Caballero, R.: The Role of Moist Intrusions in Winter Arctic Warming and Sea Ice Decline. *Journal of Climate*, 29(12), 4473–4485. <https://doi.org/10.1175/JCLI-D-15-0773.1>, 2016.
- Womack, A., Verrinder, R., Vichi, M., Alberello, A., MacHutchon, K., Skatulla, S., and Toffoli, A.: An ice-tethered, non-floating Trident Sensors Helix Beacon during SCALE 2019 Spring Cruise, Zenodo [data set], <https://doi.org/10.5281/zenodo.7954841>, 2023.
- Worby, A.P., and Allison, I: Cooperative research centre for the Antarctic and Southern Ocean environment: A Technique for Making Ship-Based Observations of Antarctic Sea Ice Thickness and Characteristics; Part I: Observational Technique and Results. *Antarctic CRCI*, 14, 6-23. 1999.
- World Meteorological Organization. Guide to wave analysis and forecasting. Secretariat of the World Meteorological Organization, 1998.
- Young, I.R., Zieger, S. and Babanin, A.V.: Global trends in wind speed and wave height. *Science*, 332(6028), 451-455. doi: 10.1126/science.1197219, 2011.
- Yuan, X.: High-wind-speed evaluation in the Southern Ocean, *Journal of Geophysical Research: Oceans*, 109, <https://doi.org/10.1029/2003JD004179>, 2004.
- Yuan, X., Patoux, J. and Li, C.: Satellite-based midlatitude cyclone statistics over the Southern Ocean: 2. Tracks and surface fluxes. *Journal of Geophysical Research: Atmospheres*, 114(D4). <https://doi.org/10.1029/2008JD010874>, 2009.
- Zhang, J., Schweiger, A., Steele, M. and Stern, H.: Sea ice floe size distribution in the marginal ice zone: Theory and numerical experiments. *Journal of Geophysical Research: Oceans*, 120(5), 3484-3498. <https://doi.org/10.1002/2015JC010770>, 2015.
- Zieger, S., Babanin, A.V., Rogers, W.E., and Young, I.R.: Observation-based source terms in the third-generation wave model WAVEWATCH, *Ocean Modelling*, 96, 2–25, <https://doi.org/10.1016/j.ocemod.2015.07.014>, 2015.
- Zwally, H.J., Comiso, J.C., Parkinson, C.L., Campbell, W.J. and Carsey, F.D.: Antarctic sea ice, 1973-1976: Satellite passive-microwave observations (Vol. 459). Scientific and Technical Information Branch, National Aeronautics and Space Administration. 1983.

Zwally, H.J., Comiso, J.C., Parkinson, C.L., Cavalieri, D.J., and Gloersen, P., 2002. Variability of Antarctic sea ice 1979–1998. *Journal of Geophysical Research: Oceans*, 107(C5), 1-9. <https://doi.org/10.1029/2000JC000733>, 2022.