

Satellite studies of interactions of internal solitary waves with other ocean phenomena

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Programa Doutoral em Engenharia Geográfica

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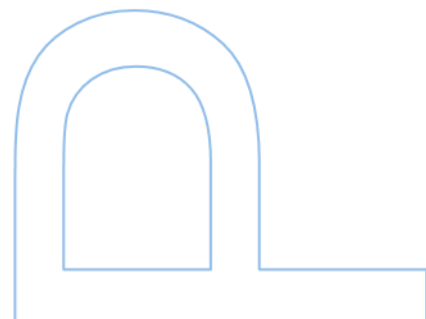
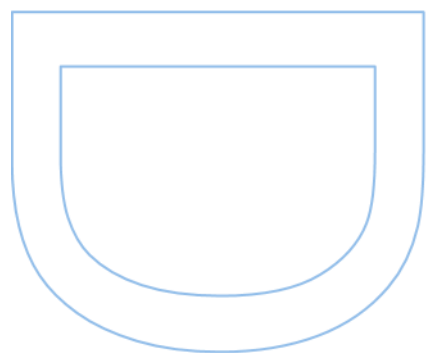
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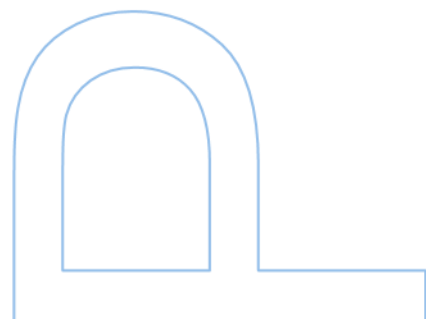
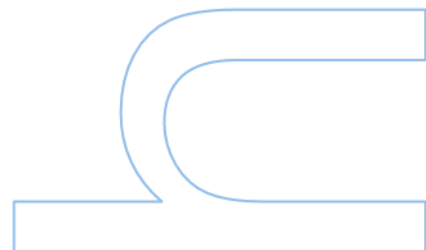
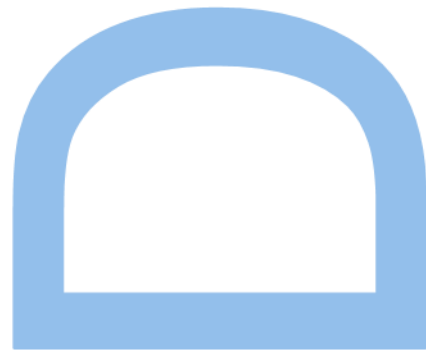
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Abstract

This thesis sets out to investigate sea surface manifestations produced by Internal Solitary Waves (ISWs) in the upper ocean with satellite sensors from Sentinel-3. During the investigation, intriguing observations of large scale ISWs in the Equatorial Pacific Ocean led to a systematic analysis of data which motivated the formulation of a hypothesis for the generation of these waves. Hence, the thesis is structured in two main chapters: in the first main chapter ([Chapter 2](#)) interactions between ISWs and smaller ocean scales (surface waves) are addressed, using mainly Synthetic Aperture Radar (SAR) altimetry and Ocean and Land Color Instrument (OLCI) images from Sentinel-3 for the study. In the second main chapter ([Chapter 3](#)) the interactions of mesoscales with ISWs are studied, using mainly OLCI and thermal images from the Sea and Land Surface Temperature Radiometer (SLSTR) of Sentinel-3. An introductory chapter presents the main challenges within a satellite oceanography context and the objectives of the thesis ([Chapter 1](#)), while a concluding chapter ([Chapter 4](#)) underlines the wider scope of the study: that ISWs in the upper ocean are relevant elements of the interaction between the ocean and atmosphere, with implications in the global climate system.

In [Chapter 2](#) the effects of surface wave breaking caused by ISWs on Significant Wave Height (SWH) are addressed. The SWH signal from the SAR altimeter in Sentinel-3A and -3B satellites is analyzed to study how ISWs are manifested in this parameter. In total 38 cases of ISW signatures are validated in OLCI and/or Sentinel-2 Multi-Spectral Imager (MSI), in two different regions: the tropical Atlantic Ocean off the Amazon shelf and the Banda Sea in the Indian Ocean. This comprises the period from 1 September 2016 until 29 May 2021, when cloud free scenes of Sentinel-3 OLCI acquired simultaneously with the SAR altimeter displayed signatures of large amplitude ISWs in SWH. Both unfocused SAR (UF-SAR 20 Hz) and fully focused SAR (FF-SAR 160 Hz) modes are analyzed, which are retracked in full range and over a reduced range of bins (truncation carried out dynamically ten gates away from the estimated epoch position). Additionally to the SWH signatures on scales of 1–3 km, which were described elsewhere, and are characterized by a strong decrease in radar power over the rough band of the ISWs followed by a small increase in the smooth band relative to the unperturbed ocean background, a second (but consistent) SWH signature is discovered. This SWH signature is noted in scales of about 20 km (i.e. ± 10 km) around the ISW crests. SWH is attenuated downstream of the ISW's crests, when compared to the upstream signal. The variance of the SWH signal is of the order of 10 – 20 %, depending on the retracking processing scheme or SAR product. The SWH signatures are unique in showing that the surface wave energy does not return to its unperturbed level after the passage of a solitary wave. We interpret this as surface wave energy dissipation caused by intense meter-scale surface wave breaking. This interpretation is consistent with evidence of wave-breaking in Sentinel-2 MSI images, which show white-capping, i.e. patchiness within the ISW crests characterized by radiance increase at all (visible) wavelengths. Modulation of breaking

waves owing to ISWs is demonstrated by estimates of the fraction of breaking waves in the presence of internal waves.

In [Chapter 3](#) a generation hypothesis for trains of ISWs that are observed within the cold tongue of the equatorial Pacific Ocean is presented and discussed. The presence of these ISWs in a zonal band from 5°S to 5°N and ranging from 210°E to 265°E is remarkable given the absence of steep bottom topography (and tides) that could explain their presence from classical theories. Instead, the origin of these trains of ISWs is consistent with a generation mechanism that involves buoyant gravity currents with sharp fronts detectable in Sentinel-3 (in the visible and thermal infrared). A 2D numerical model is successful to simulate the propagation of gravity currents into the stably stratified equatorial cold tongue, and the generation of trains of ISWs. The generation process is known as resonant generation in a transcritical regime, where energy is transferred from the gravity current to ISW trains. In total, 116 ISWs trains were identified during one full year (2020), with an average crest length of about 300 km, distance between wave crests of typically 1500 m, and modelled wave amplitude of about 10 m. Similar ISW trains were observed in Sentinel-3 images from 2019 to 2022. While 2019 was an El Niño year, 2020 and 2021 were La Niña years. The phenomenon is expected to be more visible during La Niña years because thermal fronts are stronger, with a more sharply bound equatorial cold tongue. The time scales inferred from the satellite images (average of about 28 hours) suggest the ISW trains are not short-lived, hence capable of energy transfer over large spatial scales.

Resumo

Esta tese tem como principal objetivo investigar as manifestações na superfície do mar produzidas pelas ondas internas solitárias (ISWs) na camada superficial do oceano com sensores do satélite Sentinel-3. Durante esta investigação, observações invulgares de ISWs de grande escala no Oceano Pacífico Equatorial levaram a uma análise sistemática de dados que motivou a formulação de uma hipótese para a geração destas ondas. Assim, a tese está estruturada em dois capítulos principais: no primeiro capítulo principal ([Capítulo 2](#)) são abordadas as interações entre ISWs e escalas oceânicas inferiores (ondas de superfície), utilizando principalmente altimetria de radar de abertura sintética (SAR) e imagens do instrumento de cor do oceano e da terra (OLCI) do Sentinel-3 para o estudo. No segundo capítulo principal ([Capítulo 3](#)) são estudadas as interações de mesoescalas com ISWs, utilizando principalmente OLCI e imagens térmicas do radiômetro da temperatura superficial do oceano e da terra (SLSTR) do Sentinel-3. Um capítulo introdutório apresenta os principais desafios no contexto da oceanografia por satélite e os objetivos da tese ([Capítulo 1](#)), enquanto um capítulo final ([Capítulo 4](#)) sublinha o propósito mais amplo deste estudo: que as ISWs na parte superior do oceano são elementos relevantes da interação entre o oceano e a atmosfera, com implicações no sistema climático global.

No [Capítulo 2](#) são abordados os efeitos da rebentação das ondas de superfície causadas por ISWs na Altura Significativa de Onda (SWH). O sinal SWH do altímetro SAR nos satélites Sentinel-3A e -3B é analisado para estudar como as ISWs se manifestam neste parâmetro. No total, 38 casos de assinaturas de ISWs são validados em OLCI e/ou Sentinel-2 Multi-Spectral Imager (MSI), em duas regiões diferentes: o Oceano Atlântico tropical ao largo da plataforma Amazônica e o Mar de Banda no Oceano Índico. Isto compreende o período desde 1 de setembro de 2016 até 29 de maio de 2021, quando imagens sem contaminações por nuvens do Sentinel-3 OLCI adquiridas simultaneamente com o altímetro SAR exibiram assinaturas de ISWs de grande amplitude em SWH. São analisados os modos “SAR não focados” (do inglês unfocused SAR, UF-SAR 20 Hz) e SAR totalmente focado (do inglês fully focused SAR, FF-SAR 160 Hz), que são amostrados em toda a faixa do alcance dinâmico do pulso retrodifundido do radar, e em uma faixa reduzida de bins (corte realizado dinamicamente a dez portas/posições de distância da posição estimada da época). Além das assinaturas SWH em escalas de 1 a 3 km que foram descritas nesse capítulo e noutros estudos, e que são caracterizadas por uma forte diminuição na potência do radar sobre a banda rugosa das ISWs seguida por um pequeno aumento na banda lisa em relação ao fundo oceânico imperturbado, uma assinatura de segunda ordem em SWH (mas consistente) é descoberta. Esta assinatura SWH é observada em escalas de cerca de 20 km (ou seja, ± 10 km) na direção aproximadamente “perpendicular” à crista da ISW. A SWH é atenuada a jusante das cristas da ISW, quando comparada ao sinal a montante. A variação do sinal SWH é da ordem de 10 – 20%, dependendo do processamento SAR. As assinaturas da SWH mostram que a energia das ondas superficiais não retorna ao seu nível imperturbado após a passagem de uma onda

solitária. Isso é interpretado como dissipação de energia das ondas superficiais causada pela intensa rebentação das ondas em escala métrica. Esta interpretação é consistente com a evidência de rebentação de onda nas imagens MSI do Sentinel-2, que mostram irregularidades dentro da crista da ISW caracterizada pelo aumento de radiância em todos os comprimentos de onda visíveis, resultantes da formação de espuma branca devido à rebentação. A hipótese da rebentação das ondas devido às ISWs é demonstrada por estimativas da fração de rebentação das ondas na presença de ondas internas.

No **Capítulo 3** é apresentada e discutida uma hipótese de geração para trens de ISWs (i.e., grupos de várias ISWs) que são observadas na língua fria do Oceano Pacífico equatorial. A presença destas ISWs numa banda zonal de 5°S a 5°N ao longo dos 210°E a 265°E é notável, dada a ausência de topografia de fundo íngreme (e marés) que poderia explicar a sua presença a partir de teorias clássicas. Em vez disso, a origem destes trens de ISWs é consistente com um mecanismo de geração que envolve correntes gravíticas flutuantes com frentes nítidas detetáveis no Sentinel-3 (no visível e infravermelho térmico). Um modelo numérico 2D é capaz de simular a propagação de correntes gravíticas na língua fria equatorial e a consequente geração de trens de ISWs. O processo de geração é conhecido como geração ressonante em regime transcrito, onde a energia é transferida da corrente gravítica para os trens de ISWs. No total, 116 ISWs foram identificadas durante um ano completo (2020), com comprimento médio de crista de cerca de 300 km, distância entre cristas de onda de tipicamente 1500 m e amplitude de onda, segundo o modelo, de cerca de 10 m. Trens de ISWs semelhantes foram observadas em imagens do Sentinel-3 de 2019 a 2022. Embora 2019 tenha sido um ano de El Niño, 2020 e 2021 foram anos de La Niña. Espera-se que o fenómeno seja mais frequente durante os anos de La Niña porque as frentes térmicas são mais fortes, com uma língua fria equatorial mais acentuada e extensa. As escalas de tempo inferidas a partir das imagens de satélite (média de cerca de 28 horas) sugerem que as ISWs não têm vida curta, sendo, portanto, capazes de transferir energia em grandes escalas espaciais.

Keywords

Buoyant gravity currents

Equatorial cold tongue

Internal Solitary Waves

Ocean science

Ocean phenomena and processes

Satellite Altimetry

Satellite remote sensing

Sentinel-2

Sentinel-3

Surface wave breaking

Surveying engineering

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*There is, one knows not what sweet mystery about this sea, whose
gently awful stirrings seem to speak of some hidden soul beneath.*

Herman Melville

How inappropriate to call this planet Earth, when clearly it is Ocean.

Arthur C. Clarke

I try all things; I achieve what I can.

Herman Melville, Moby Dick

Contents

Abstract.....	ii
Resumo	iv
Keywords.....	vi
Acknowledgements.....	vii
List of Figures.....	xiii
List of Tables	xvi
Acronyms	xvii
CHAPTER 1.....	1
General Introduction.....	1
1.1. Scope and Overview.....	1
1.2. Internal Solitary Waves and their interaction with other-scale phenomena	7
1.3. Remote Sensing and Satellite altimetry	11
1.4. Motivations and main goals	15
1.5. Thesis outline.....	17
CHAPTER 2.....	19
Prologue.....	19
EFFECTS OF SURFACE WAVE BREAKING CAUSED BY INTERNAL SOLITARY WAVES IN SAR ALTIMETER: SENTINEL-3 COPERNICUS PRODUCTS AND ADVANCED NEW PRODUCTS.....	21
2.1. Introduction.....	21
2.1.1. Oceanic Internal Waves, Their Surface Manifestations and Remote Sensing	21
2.1.2. Surface Wave Breaking Induced by Strong ISWs	24
2.2. Material and Methods.....	27
2.2.1. Fundamentals of satellite altimetry and its sea surface signatures of ISWs...	28
2.2.2. Sentinel-3 SAR Altimeter and Detection of ISWs	29
2.2.3. Ocean and Land Color Instrument (OLCI)	35
2.2.4. Characterization of Fraction of Wave Breaking in Sentinel-2 MultiSpectral Instrument (MSI).....	36
2.3. Results	37
2.3.1. Sentinel-3 Case Study Using Optical Imagery and Standard Copernicus SAR Altimeter Data	37

2.3.2. Results for Standard Copernicus SAR Data.....	38
2.3.3. A Case Study Using Optical Imagery and Advanced SAR Altimeter Products	41
2.3.4. Results for Advanced FFSAR Products.....	43
2.3.5. Consistency of Results for Different SAR Products.....	43
2.3.6. Image Processing for the investigation of wave breaking in Sentinel-2 images	44
2.4. Summary and Conclusions	48
CHAPTER 3.....	51
Prologue.....	51
INTERNAL SOLITARY WAVES WITHIN THE COLD TONGUE OF THE EQUATORIAL PACIFIC GENERATED BY BUOYANT GRAVITY CURRENTS.....	55
3.1. Introduction.....	55
3.2. Data and Methods	58
a) Satellite data.....	58
b) In situ data	59
c) Numerical modelling.....	59
3.3. Results	61
a) Case studies	62
b) Large scale picture	65
c) Prediction of internal wave linear phase speed in the study region	66
d) Statistics of wave observations.....	69
e) Model simulations	71
3.4. Discussion.....	73
a) Gravity current's energy considerations	73
b) Generation mechanism of ISWs.....	74
c) ISW spatial structure and age.....	75
d) Broader scope	77
3.5. Summary and Conclusion.....	78
CHAPTER 4.....	81
CONCLUSION	81
4.1. Concluding remarks.....	81

4.2. Wider Significance of Chapter 2.....	83
4.3. Wider Significance of Chapter 3.....	83
4.4. Future Work.....	85
4.5. Epilogue.....	89
REFERENCES	91

List of Figures

Figure 1.1 - A cover from MOBY DICK (by Herman Melville, original version from 1851) is shown here as a testimony to mankind's fascination for the oceans (cover from a 2023 online version edited by Dialogues Morlaix, available at www.dialoguesmorlaix.com).....	2
Figure 1.2 - Every two to seven years, the equatorial Pacific Ocean warms up (El Niño in left top panel) or cools down (La Niña in right top panel) at least 0.5°C, whereas one follows the other (eventually with some neutral conditions between them).....	3
Figure 1.3 - Classical time and space scales for selected ocean processes (adapted from Lampitt et al. 2010 and other authors as indicated in the lower-right inset).....	6
Figure 1.4 - A surface soliton is shown propagating in a narrow water channel, which was recreated in an experiment similar to that of Sir Scott Russell in 1834.....	8
Figure 1.5 - Illustration showing how ISWs (with amplitude a and phase speed C) feeding on energy from larger-scale phenomena (see top-left) may travel along ocean basins, while interacting with other-scale phenomena, and transferring energy into the lower end of the ocean energy cascade.	10
Figure 1.6 - Nominal spatial resolutions in Sentinel sensors used in this thesis (and related studies). Sentinels-1 to 3 (and their sensors) are colour coded in the top inset. Important synergies used in the studies present in Chapters 2 and 3 (and phenomena therein) are shown using red arrows. See text for more details.	12
Figure 1.7 - Typical parameters in classical satellite altimetry. Note how a measure of SSH may be obtained from a platform in space. However, altimetric signals resulting from surface echoes that are reflected back to the satellite (often termed waveforms) carry additional information.	14
Figure 1.8 - General and specific objectives of this thesis and addressed questions.	16
Figure P2.1 - OLCI image acquired on August 25, 2020, at 11:22 UTC (see inset for location approximately off the Amazon shelf-break)	20
Figure 2.1 - Photographs taken from the Research Vessel Charles Darwin showing sea surface manifestations of a large-amplitude ISW advancing towards the ship in the central region of the Bay of Biscay, which includes surface wave breaking (i.e., whitecaps).....	25
Figure 2.2 - (a) Quasi-true color image (Level-1b) acquired on 20 August 2019 at 13 h 33 m UTC by the OLCI sensor onboard the Sentinel-3A satellite (see inset for location)	26
Figure 2.3 - (a) Normalized SAR altimeter waveforms for an unperturbed (in relation to the ISWs' surface manifestations) sea surface (black) and for different SWHs.....	29
Figure 2.4 - (a) Quasi-true color image (Level-1b) acquired on 4 September 2020 at 13h01m UTC by the OLCI sensor onboard the Sentinel-3B satellite (see inset for location).	31

Figure 2.5 - Bar plots representing average significant wave heights (SWHs) for leading (i.e., front) and trailing (i.e., rear) sections (see text for details) of the ISWs for all cases in Table 2.1.....	40
Figure 2.6 - The rectangles depicted in red shades represent the ISW center at 3.95°N. From the center of ISW, it is considered ± 2 km to each side. To the left of it, we define the “SWH front” ahead of the ISW (“rough” in lighter blue using scales of about 10 km) and to the right, the “SWH rear” (“smooth” in darker blue using 10 km).	41
Figure 2.7 - As in Figure 2.6, but top panels of (a,b) shows along-track NRCSs (i.e., σ_0) measured by the SRAL altimeter (in synergy with the OLCI) with the SAR-mode retracking color-coded in (a) UFSAR, UFSAR 10 Bins and its large-scale mean and in (b) UFSAR MSS, UFSAR MSS 10 Bins and its large-scale mean (using a box running average of 3 km)	42
Figure 2.8 - Bar plots representing significant wave heights (SWHs) for leading (front) and trailing (rear) sections of the ISWs for all cases in Table S1 using different algorithms: (a) FF 10 Bins. (b) FF MSS 10 Bins. (c) UF 10 Bins. (d) UF MSS 10 Bins.	44
Figure 2.9 - (a) Quasi-true color image (Level-1b) acquired on 21 April 2019 at 01 h 31 m UTC by the OLCI sensor onboard the Sentinel-3B satellite	45
Figure 2.10 - (a) Zoom as shown in the black rectangle in Figure 2.9b corresponding to the NIR Band 8 of the Multispectral Instrument (MSI) onboard the Sentinel-2A satellite on 21 April 2019 at 04h25m UTC showing sea surface signatures of an ISW packet	46
Figure 2.11 - (a) W represents wave breaking as shown in Figure 2.10c but color-coded to highlight the ISWs’ different sections, in which we applied a threshold of 0.002 (see text for details). (b) Mean of wave breaking fraction, i.e., W calculated for each line of (a), which is also approximately along the ISWs’ crests.	47
Figure P3.1 – (a) Sentinel-3A OLCI image dated 18 June 2019 (acquired at 17:12 UTC) presented in quasi-true color and showing clear surface manifestations of ISWs propagating towards the southwest	51
Figure P3.2 - (a) Sentinel-3B OLCI image dated 27 April 2022 (acquired at 17:07 UTC) presented in quasi-true color and with evidence of ISWs propagating towards the south-southwest	53
Figure 3.1 – (a) Spatial distribution of ISWs documented in Jackson et al. 2012 (black dots) in the eastern tropical Pacific. Surface currents (arrows) and ocean fronts (dashed black lines) are displayed over a daily Multi-Scale Ultra High Resolution (MUR) SST image (dated 19 June 2020)	57
Figure 3.2 - Spatial distribution of the ISWs found in this study (magenta dots) for the year 2020. The black dots are the ISWs reported in Jackson et al. (2012). A MUR SST annual average image for the year 2020 is shown in the background.....	61

Figure 3.3 – (a) Sentinel-3B thermal infra-red image (Level 2 WST) dated 7 June 2020, acquired at 15:16:19 UTC (start time). The magenta lines and the red line are contours of the ISWs leading crests and the ocean front, respectively, also identified in (b).....	63
Figure 3.4 – (a) Sentinel-3A thermal infra-red image (Level 2 WST) dated 16 June 2020, acquired at 18:16:24 UTC (start time). The magenta and red lines are contours of the ISW leading crest and the ocean front identified in (b) and (c), respectively. The black rectangle corresponds to (d).....	64
Figure 3.5 - Sentinel-3 visible images (Level 1b OLCI) dated 15 June 2020 (right) and dated 16 June 2020 (left). The leading ISWs are marked in blue and green, respectively. Magenta dashed lines mark distances between points on the same ISW leading crest, and are used to estimate the average wave propagation speed.....	65
Figure 3.6 - Spatial distribution of the fronts (dotted lines) and ISW packets (solid lines), identified in the visible imagery of Sentinel-3, between 15 June and 22 June 2020. The colour for each day is identified in the upper part of the map.	67
Figure 3.7 – (a) Potential density and buoyancy frequency (N) profiles, respectively in blue and orange colors	68
Figure 3.8 - Phase speeds of linear long internal waves evaluated for different vertical profiles of horizontal velocity (U_1 and U_2 , eastward and northward, respectively)	69
Figure 3.9 – (a) Latitude distribution of the 116 ISWs selected from 2020 images in the eastern equatorial Pacific Ocean. (b) Distribution of the ISW wavelengths for the best 66 cases for the same period of time. (c) As in (b) for along-crest length scale.	70
Figure 3.10 - Numerical simulations of ISWs generated by a gravity current at a) $t=30$ min, b) $t=40$ min and c) $t=50$ min. In the six plots it is possible to see the ISW propagation from right to left.....	72
Figure 3.11 - Hovmöller (space–time) diagram for the surface horizontal velocity. The bright red leftward and downward leaning linear patterns show two consecutive ISWs. The blue coloured linear pattern, to the right of the ISWs, shows the gravity current.	73
Figure 4.1 - Multi-Scale Ultra High Resolution (MUR) SST image (dated 19 June 2020) in the equatorial Pacific band. The black dots are the ISWs identified in the visible imagery of Sentinel-3, between 15 June and 22 June 2020 (image adapted from Figure 3.6 of this thesis) and the locations of moorings and shipboard sampling (purple) described.....	86
Figure 4.2 - a) Multi-Scale Ultra High Resolution (MUR) SST image (dated 04 August 2023) in the Atlantic equatorial ocean. The black dashed rectangle corresponds to the location of panels b and c and the black dot is the ISW identified in the visible imagery of Sentinel-3 (panel b))	87
Figure 4.3 – SWOT KaRIN image showing SLAs, dated 10 July 2023 – PASS 021 of the 1-day repeat cycle orbit. Curtesy of Richard D. Ray.	88

List of Tables

Table 2.1 - List of Sentinel-3 cases exhibiting large-scale ISWs in optical imagery (OLCI) in synergy with the corresponding sea surface signatures in the SRAL altimeter 39

Table 2.2 - Mean of SWHs in front and rear of ISWs (see text for details) and differences between both (in percentage) for cases in the Banda Sea and Tropical West Atlantic.....40

Table 3.1 - Parameter values for each model run. Densities ρ are given in kg m^{-3} ; velocities u in m s^{-1} ; and depth scales, h , d and L in m 60

Acronyms

AATSR	Advanced Along-Track Scanning Radiometer
ARGO	Array for Real-Time Geostrophic Oceanography
BVP	Boundary Value Problem
CLS	Collect Localisation Satellites
ENSO	El Niño-Southern Oscillation
ESA	European Space Agency
EUC	Equatorial Undercurrent
FFSAR	Fully Focused SAR
ISW	Internal Solitary Wave
IW	Internal Wave
IT	Internal Tide
JPL	Jet Propulsion Laboratory
KaRIN	Ka-band Radar Interferometer
LEF	Leading Edge Front
MERIS	Medium Resolution Imaging Spectrometer
MSI	MultiSpectral Instrument
mss	Mean square slope
MSS	SARmultilookLS3 - adaptive retracking algorithm
MUR	Multiscale Ultrahigh Resolution
MWR	MicroWave Radiometer
NASA	National Aeronautics and Space Administration
NEC	North Equatorial Current
NECC	North Equatorial Counter Current
NEF	North Equatorial Front
NOAA	National Oceanic and Atmospheric Administration
NRCS	Normalized Radar Cross Section
NTC	Non-Time-Critical
OLCI	Ocean and Land Color Instrument
POD	Precise Orbit Determination
PRF	Pulse Repetition Frequency
PTR	Point Target Response
S2	Sentinel-2
S3	Sentinel-3
SAR	Synthetic Aperture Radar
SEC	South Equatorial Current
SEF	South Equatorial Front
SLA	Sea Level Anomaly
SLSTR	Sea and Land Surface Temperature Radiometer

SRAL	SAR altimeter
SSF	Small Scale Front
SWH	Significant Wave Height
SWOT	Surface Water and Ocean Topography
TAO	Tropical Atmosphere Ocean
TEF	Trailing Edge Front
TIV	Tropical Instability Vortex
TIW	Tropical Instability Wave
TOA	Top-Of-Atmosphere
U ₁₀	Wind speed at 10 meter height
UFSAR	Unfocused SAR
WEA	Warner et al. 2018

CHAPTER 1

General Introduction

1.1. Scope and Overview

Our home planet - Earth - is bonded to its atmosphere and oceans. They surround it in perpetual motions. For centuries we have been pursuing the details of their dynamics, and for centuries we have come to realize that we may pursue them for centuries more. Understanding them transcends curiosity and has long become a necessity, overgrowing, and likely to endure for times to come. History and recent scientific evidence confirms how the ocean and atmosphere have long been a part of our imagination and reality.

For instance, in **Figure 1.1**, a cover from the well-known novel *MOBY DICK* reminds us of the ocean's mysterious nature. Ironically, Melville writes that "When beholding the tranquil beauty and brilliancy of the ocean's skin, one forgets the tiger heart that pants beneath it; and would not willingly remember that this velvet paw but conceals a remorseless fang". Indeed, the surface may conceal what lies beneath it (in *MOBY DICK* referring to the giant white whale). A perfect analogy, perhaps, for the less well-known internal waves, which concealed in the ocean's interior may be many times larger than surface waves. A similar message of that in *Moby Dick* emerges today, but in science. For instance, the natural cycle called El Niño Southern Oscillation (or ENSO) provides growing evidence for a climate shift and extreme weather (Cai et al. 2023; Geng et al. 2023). Indeed, changes in sea surface temperature with ENSO (as illustrated in **Figure 1.2**) are reported to disrupt global weather patterns, and their future trends are a matter of strong debate, with scientists concerned they can have severe consequences to the planet (Wu et al. 2019; Jones 2022).

Perhaps one of the most immediate and yet intricate perceptions while studying the ocean is that coming from the sense of scales. Indeed, it cannot go unnoticed that the ocean moves, but with its many motions traversing multiple orders of magnitude in both time and space. Furthermore, even though they are ultimately powered by the Sun and the Moon, the details of how energy cascades - i.e. is transported or transferred - between the ocean's many motion scales is still an active area of research with great appeal in Ocean Sciences (Cao et al. 2019; Whalen et al. 2020; Taylor and Thompson 2023; Zhang et al. 2023). In a way, this thesis aims to contribute to these research efforts by investigating phenomena and processes thereof where energy transfers are directly between the submesoscale (assumed here between hundreds of meters and a few kilometers) and either mesoscale or small-scale phenomena.



Figure 1.1 - A cover from MOBY DICK (by Herman Melville, original version from 1851) is shown here as a testimony to mankind's fascination for the oceans (cover from a 2023 online version edited by Dialogues Morlaix, available at www.dialoguesmorlaix.com).

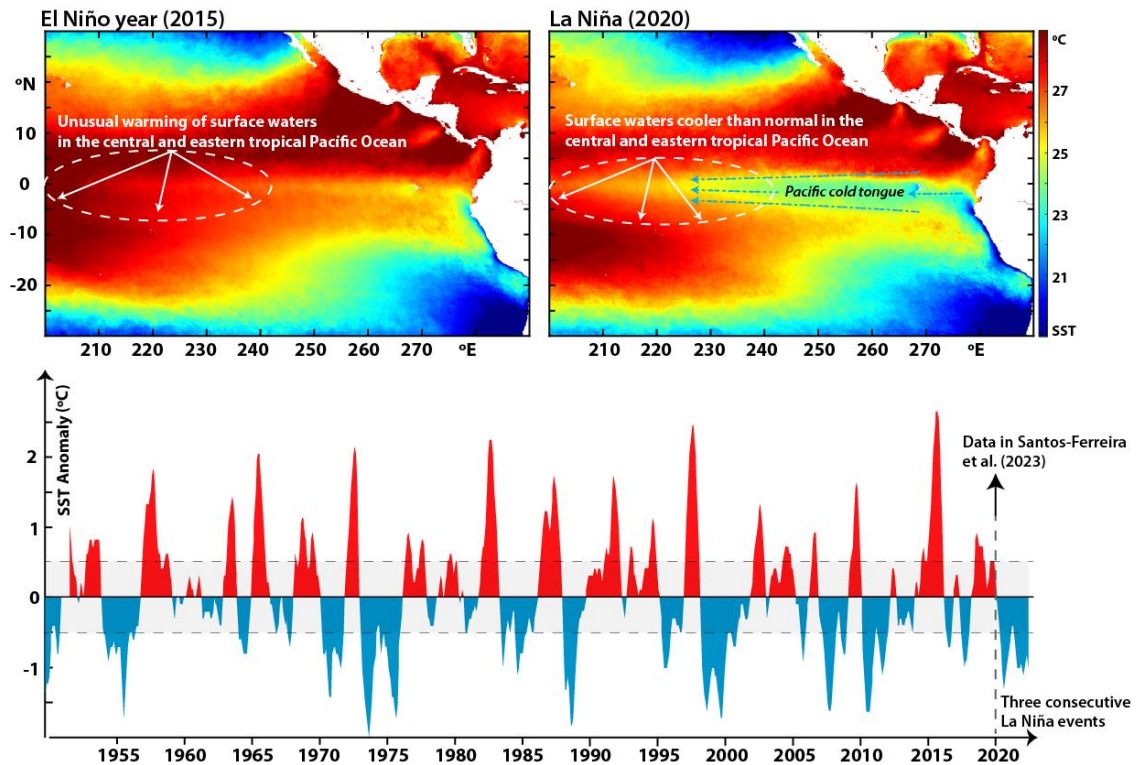


Figure 1.2 - Every two to seven years, the equatorial Pacific Ocean warms up (El Niño in left top panel) or cools down (La Niña in right top panel) at least 0.5°C , whereas one follows the other (eventually with some neutral conditions between them). These events are declared using the Oceanic Niño Index (ONI) using climatically averaged sea surface temperatures in the tropical Pacific Ocean (seen in a time series in the bottom panel, adapted from Leamon 2023).

The journey therefore begins by wondering what phenomena in the ocean can transport energy along great distances, while eventually transferring it from or to other phenomena with very different characteristic scales (potentially across several orders of magnitude in both time and space). However, the ocean behaves very differently when considering the horizontal and vertical directions, and hence the same naturally applies for energy transports and transfers in these two distinct directions. That is essentially a consequence of gravity, which works to make the ocean vertically stratified, and hence facilitates motions in the horizontal while attenuating them in the vertical. Consequently, phenomena (and processes thereof) are characteristically different when comparing the horizontal and vertical directions – often privileging one over the other and rarely dominating both. Or could it be that some phenomena in the ocean can (at least to some extent) accomplish both?

In a way, this means finding phenomena which, while propagating in the horizontal direction, can overcome gravity and participate in energy transfers in the vertical direction. Intuitively, that means having large vertical motions and vertical velocities - interestingly, in the ocean, the biggest of which are provided by waves. Wave motions are an important component of ocean dynamics (Gill 1982). Nonetheless, there is a great variety of wave phenomena, but if the purpose is that of finding large propagation distances and yet having significant contributions for the oceans' vertical motions, then that inevitably means Internal Waves. Internal Waves in the ocean are ubiquitous and may be regarded as an analogue of

the more familiar surface waves (Jackson et al. 2012). The only difference in this simplified view would be that rather than propagating along the sharp interface between air and water at the ocean surface, they use the much subtler density gradients along the ocean's pycnocline as their waveguide (see e.g. Gerkema and Zimmerman 2008).

A more comprehensive description of Internal Waves is provided in [section 1.2](#), but a few of their details are discussed next which add to the scope of the issues investigated in this thesis. In particular, a special class of Internal Waves, which are usually referred to as Internal Solitary Waves (ISWs, formally introduced in the next section), and are typically observed in satellite images and in situ data to have time and space scales that are the same as those of many ocean's processes. For instance, they can propagate across basin-scale distances and yet display a highly turbulent character in their propagation, meaning their energetics could naturally link the larger scales of global overturning circulation with the smaller scales in turbulent motion (Whalen et al. 2020). In addition, Internal Solitary Waves provide (by far) the ocean with its largest vertical velocities (up to 1 m s^{-1}) over vertical scales up to about one hundred meters (da Silva et al. 2015), meaning large vertical gradients in a significant stretch of the water column.

To clarify the extent over which (at least in principle) Internal Solitary Waves may interact with other ocean phenomena, we show in **Figure 1.3** an illustration of typical time and space scales (partially adapted from Lampitt et al. 2010) but added with an alternative view of our own making. This alternative view is designed to highlight how time and space scales in Internal Solitary Waves can potentially transport or transfer energy between them and other ocean scales. Before proceeding, however, we stress that selecting scales representative of ocean phenomena can sometimes be regarded as too subjective. This means other choices in **Figure 1.3** could equally be argued suitable, but these should lead to the same clarification of the problem, whereas inappropriate scales would only lead to contradictions.

A first look at **Figure 1.3** reveals that typical time and space scales (in a classical view) in Internal Solitary Waves fit interestingly in the middle of the plot, which in a way are already suggesting possible contributions in connecting the oceans' larger and smaller scales. However, it could be argued that their influence (in the sense of transporting and transferring energy) may in fact stretch farther in both time and space. Note that, given that they are wave phenomena, choosing time and space scales simply relies on characteristic scales of wave periods and wavelengths (respectively). Nonetheless, Internal Solitary Waves can be highly nonlinear (to be shown in the next section). This means that, in their (up to) basin-scale propagation distances, they transport and transfer not only energy (a defining characteristic of a wave) but also mass and momentum (Helfrich and Melville 2006). In this sense, two examples can illustrate how Internal Solitary Waves may influence or be influenced by other phenomena in the view of **Figure 1.3**.

On the one hand, these waves' well-known turbulent character dictates that mass and momentum eventually cascade into smaller scales of motion (eventually all the way down to turbulent scales). On the other hand, the ocean is naturally stratified and therefore any

process in the ocean able to disturb its inner stratification will act as a potential generation mechanism (in this regard see e.g. chapter 1 in Grimshaw 2002). This means that processes with scales larger than those of Internal Solitary Waves may impact the oceans' isopycnals, and hence feed internal vertical motions in the ocean stable stratification.

In light of this extended view, new criteria may be sought in **Figure 1.3** other than those traditional for wave phenomena (i.e. wave period and wavelength). For instance, their lifespan may equally be considered as a characteristic time and space measure of their ability to transport and transfer energy to or from other ocean phenomena. This extended view is shown in **Figure 1.3** by means of green arrows, which point to selected lifespans documented in the literature (see inset in top-left corner).

The message is clear. Internal Solitary Waves can be clearly perceived as a process that acts as a redistribution mechanism, by which energy is transported along great distances and potentially transferred between different phenomena. This view sets the stage for the investigations presented and discussed in this thesis, which aim to provide details of two processes involving Internal Solitary Waves and their energy transfers from/to larger/smaller-scale phenomena. The first presents satellite evidence documenting how Internal Solitary Waves and surface waves can interact to increase surface wave breaking (and hence turbulence and mixing), while the second investigates how Internal Waves may feed from basin-scale ocean circulations closely related with climate oscillation time-scales (respectively, [Chapter 2](#) and [Chapter 3](#) in this thesis).

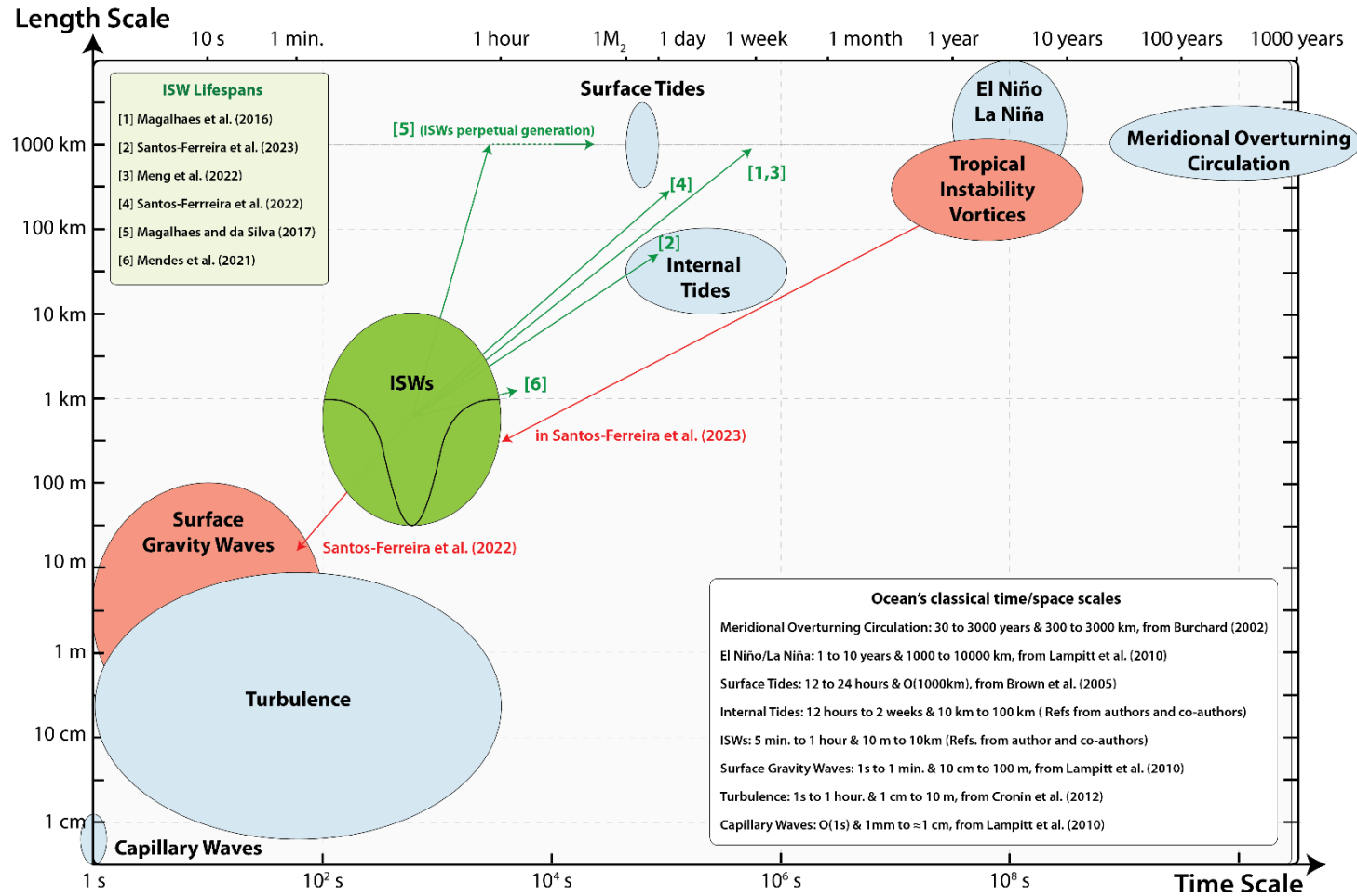


Figure 1.3 - Classical time and space scales for selected ocean processes (adapted from Lampitt et al. 2010 and other authors as indicated in the lower-right inset). Those for Internal Solitary Waves (ISWs) are highlighted together with an extended view of their scales considering alternative criteria (in green arrows according to the inset in the top-left). The arrows in red represent novel views for energy transfers involving ISWs presented in Chapters 2 and 3 in this thesis. Concerning reference [5] in the top-left corner, it is noted that these ISWs are not tidally generated but rather from steady currents flowing along bottom bathymetry. For this case, a distance scale is chosen according to their observed extent along the Patagonian shelf-break, while their time scale is still undetermined since they may be perpetually feeding from the steady flow (represented as the horizontal green-dashed arrow, see also Magalhaes and da Silva 2017b). See text for more details.

1.2. Internal Solitary Waves and their interaction with other-scale phenomena

Waves in nature can be found nearly everywhere (Narayanan and Swapan 2015), but the ocean environment is certainly amongst where their views excel the most. However, while surface waves are a familiar feature of the sea surface, the ocean environment is now well-known to exhibit an extensive array of other wave phenomena. Waves in the ocean can come in very different shapes and sizes, range across several orders of magnitudes in both time and space, and have very distinctive origins and dynamics (LeBlond and Mysak 1981), but perhaps none as counterintuitive as Internal Waves (IW's).

Indeed their sense of motion, and hence scales, cannot be grasped by means relying on visual references (as e.g. swell which can be timed from a cliff or at a distance), which make these less-familiar waves (i.e. IW's) much harder to conceptualise. An alternative way to introduce them may be simply to picture them as having their largest vertical amplitudes in the ocean's interior rather than e.g. at its surface (see e.g. chapter 1 in Gerkema and Zimmerman 2008). However, this general definition includes internal motions in the ocean covering a large frequency spectrum, which according to IW theory is set between f and N (i.e. the Coriolis and buoyancy or Brunt-Väisälä frequencies, respectively), and hence equally broad spectra in time and space - ranging from a few minutes and tens of meters to days and a hundred of kilometers (see e.g. Holloway 1983; Gerkema and Zimmerman 2008 and reviews by Jackson et al. 2012 and Grimshaw 2002).

There are, nonetheless, special classes of IW's which have been given significant attention in Ocean Sciences (see e.g. St. Laurent et al. 2012), the most important of which in the framework of this thesis being Internal Tides (IT's) and (even more so) Internal Solitary Waves (ISW's). The reference to IT's is more intuitive and hence easier to begin with. These IW's are simply the consequence of the barotropic surface tide meeting rough ocean bathymetry, which causes vertical motions at the tidal frequency that may then propagate along the ocean's pycnocline as interfacial waves (hence *Internal Tides*). In turn, IT's in the ocean are often observed to steepen - i.e. their initial nearly-sinusoidal shape quickly changes (over timescales shorter than or comparable to their periodicity) and their leading slopes become steeper while their trailing sections become flatter (see e.g. Figure 16 in Grimshaw et al. 2010, and see also chapters 7 and 8 in Gerkema and Zimmerman 2008). This process is often referred to as the disintegration of the IT into shorter-period IW's, which are often observed in satellite images (to be further discussed in the next chapter) and usually referred to as ISW's.

ISW's have very distinctive dynamics and their nature greatly transcends the boundaries of the ocean environment - they have been extensively investigated theoretically, in laboratory and numerical experiments, and are found across countless fields in natural sciences (see e.g. Drazin and Kazanijian 1983). It is, however, very curious to note that their account was documented first in a water environment almost two centuries ago (Russell 1844). A Scottish naval engineer (Sir John Scott Russell) was conducting a series of experiments with boats in a shallow water canal. The experiments were designed to improve

boat performance and the setup involved horses pulling a boat along a narrow water channel. In that process - Sir Russell noticed - a surface wave would form with the moving boat (as seen in **Figure 1.4**), but that wave would detach and continue along the canal as the boat would come to rest. This wave was apparent in the form of a large solitary elevation - hence coined solitary wave (a name also following from theory and observations showing that solitary waves may have properties analogous to those of elementary particles, see discussions in Osborne and Burch 1980) - which continued its course along the channel for great distances apparently without change of form or speed.



Figure 1.4 - A surface soliton is shown propagating in a narrow water channel, which was recreated in an experiment similar to that of Sir Scott Russell in 1834. Note how the solitary wave propagates ahead of the boat as it decelerates and comes to a full stop. See more at <http://ramanujan25449.blogspot.com/2017/05/solitones.html>.

Since this remarkable discovery, the amount of research involving solitary wave theory has greatly increased. In the ocean environment, and despite surface solitary waves in the ocean being an important field of research in their own right (such as that reported by Sir Russel, see also Miles 1980), it is interfacial ISWs propagating along the ocean's inner pycnocline that have been gathering considerable attention and research efforts. Their first documented observations and measurements can be traced back for at least nearly 60 years, when Perry and Schimke (1965) measured internal motions (using mechanical bathythermographs) north of Sumatra that were later interpreted as resulting from ISWs. Henceforth, the literature describing their observations in the ocean grew considerably, as did their theoretical, laboratory, observational, and numerical studies (see e.g. reviews by Garrett and Munk 1979; Helfrich and Melville 2006; Jackson et al. 2012 and chapter 8 from Gerkema and Zimmerman 2008 and Grimshaw 2002).

Investigations regarding ISW dynamics and their interactions with other phenomena in numerous fields in Oceans Sciences have kept pace with these advances, including generation, propagation and dissipation processes, as well as impacts in underwater

acoustics, shear-stress in man-made structures, biogeochemical processes, climate, etc. - the list is extensive and continues to grow (see e.g. reviews by Woodson 2018 and Boegman and Stastna 2019). Therefore, it seems fair to question why ISWs are found to influence (and be influenced by) so many different phenomena and processes in the ocean.

On the one hand, the answer to why ISWs are able to influence a series of different ocean phenomena may lie (at least to some extent) in their time and space scales. Waves carry energy, and ISWs are no different. However, ISWs have been frequently found to propagate across basin-scale distances for several days (see e.g. Jackson 2007; Magalhaes et al. 2016; Jensen et al. 2020). In addition, ISWs can be highly nonlinear, meaning that they carry not only energy but also mass (and hence momentum – Lamb 1997). Together with their large propagation distances, and their many shelf-break hotspots (see Jackson 2007), this means an important redistribution mechanism for water masses between coastal and deep water environments (see e.g. discussion in Sandstrom and Elliott 1984 and Sandstrom et al. 1989). Particularly interesting for this thesis, is also the fact that ISWs provide the ocean with its largest vertical velocities (up to 1 m s^{-1}) and vertical excursions up to 200 meters (Huang et al. 2016) as they propagate along the ocean's pycnocline. Since they have a highly turbulent character, that means an important source for cascading energy into diapycnal mixing in a natural vertical boundary (separating subsurface and deeper-waters) where diapycnal mixing is naturally suppressed (Ferrari et al. 2016). It might be wondered, nonetheless, if the turbulent and mixing potential in ISWs may also reach other sections (i.e. other than intermediate sections at pycnocline depths) in the water column - to be explored in this thesis in [Chapter 2](#).

On the other hand, however, ISWs themselves may be influenced by other phenomena in their generation, propagation and dissipation stages (see review papers e.g. by Boegman and Stastna 2019). In the particular case of generation, which is of special interest for this thesis, an argument could be brought forward to help explain why ISWs are so frequently found nearly anywhere in the world's oceans (see Jackson 2007). The reason is that, provided suitable stratification exists, any phenomena (or any process thereof) capable of disturbing the ocean pycnocline is a potential ISW generator. Surface tides are a good example when considering large-scales in the ocean. They are barotropic and propagate around the world's oceans without significant vertical motions. In the presence, however, of bathymetry features such as canyons, promontories, ridges, seamounts, an internal vertical oscillating component is added in the water column. If suitable stratification exists, then that oscillation will propagate as an IT and eventually disintegrate into ISWs - which is in fact the case for numerous ISW hotspots found in both deep and coastal waters (see e.g. Jackson et al. 2012). Again, however, it might be wondered if other non-tidal phenomena with considerable large-scales may also play an important part in feeding ISWs - to be explored in this thesis in [Chapter 3](#).

These two complementary views that ISWs may play a part in the ocean's energy cascade - i.e. that they influence (hinting on smaller-scales) and are influenced by (hinting on larger-scales) other-scale phenomena - is illustrated in **Figure 1.5**. Another purpose in this

presentation is to acknowledge that energy being fed into ISWs will inevitably be redistributed elsewhere along their basin-scale propagation distances. A number of mechanisms and processes in this energy cascade (i.e. both in feeding ISWs and in their interactions and energy transfers towards smaller-scales) are already well-documented in the literature (see a few examples in **Figure 1.5**). Others, however, may exist, and it is the aim of this thesis to contribute to a better understanding of the oceans' energy cascade in light of processes involving ISWs (see **Figure 1.5**). Namely, **Chapter 2** presents evidence of enhanced surface wave breaking owing to large-scale ISWs in a number of hotspots in the world's oceans, and **Chapter 3** presents evidence that Tropical Instability Waves/ gravity currents may feed ISW motions in a large section of the tropical Pacific Ocean. In both accounts, these novel observations were only possible due to the unique views of satellite Remote Sensing and its ability to detect the sea surface signatures of ISWs.

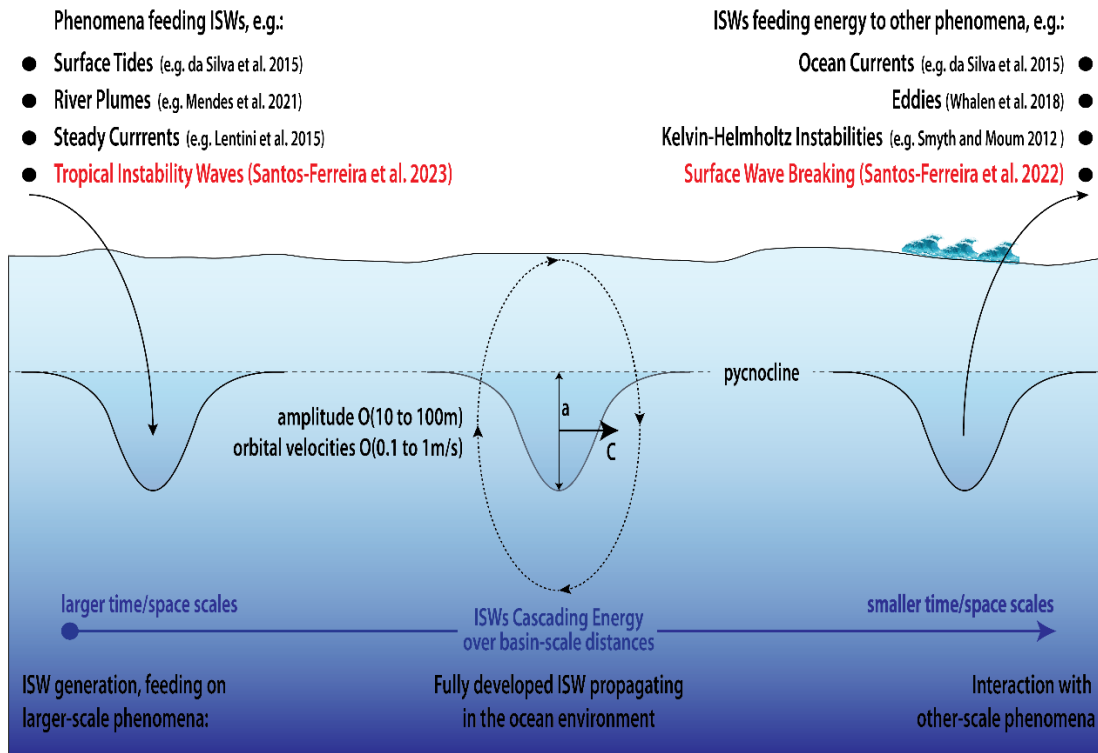


Figure 1.5 - Illustration showing how ISWs (with amplitude a and phase speed C) feeding on energy from larger-scale phenomena (see top-left) may travel along ocean basins, while interacting with other-scale phenomena, and transferring energy into the lower end of the ocean energy cascade. Two novelties presented in this thesis are highlighted in red: one relating to Tropical Instability Waves/ gravity currents that may help generate ISWs in the tropical Pacific Ocean; and another relating to enhanced surface wave breaking owing to ISWs.

1.3. Remote Sensing and Satellite altimetry

The ocean is simply too large to be monitored continuously by conventional instrumentation (meaning in water samplings), and hence remains largely under sampled (see e.g. Talley 2011; Cochran et al. 2019). Recently, in the last few decades (arguably since the 70s with Landsat 1, see e.g. <https://www.usgs.gov/landsat-missions>, and the NOAA satellites series for weather observations), satellite Remote Sensing has been complementing traditional in situ sampling to document unprecedented scales and surface properties in countless phenomena in the ocean environment (see e.g. reviews by Esaias et al. 1998; Roy et al. 2017 and Robinson 2010).

In its essence, Satellite Remote Sensing may be thought of as a technique to acquire information from a target at a distance - i.e. no direct contact between instrument and object is required. Different views regarding its intricacies may then follow from different authors depending on applications, platforms or fields of science, but the principles remain the same: i.e. electromagnetic radiation, and the laws it is governed by, are used to *sense* and measure physical properties from afar (see e.g. Robinson 2004). The details need not concern us here (may be found e.g. in text books by Joseph 2005; Emery and Camps 2017; Ulaby and Long 2014), but discussing how satellite Remote Sensing was paramount in acquiring the data and results shown in [Chapters 2](#) and [3](#) deserves some additional remarks.

First and foremost, however, it cannot go unnoticed that while for some communities satellite images showing evidence of ISWs may be a common feature, for others it may be hard to grasp how sensors flying hundreds of kilometers above the sea surface can identify waves propagating hundreds of meters below the sea surface. In this regard, clarification and some intuitiveness may come from acknowledging that it is not the waves, but rather their surface manifestations (or signatures), that are captured in satellite imagery, especially in surface (e.g. wave or roughness) fields which change from an unperturbed background owing to surface ISW-induced effects. The details of IW-Surface Wave interactions, and their signatures in the electromagnetic signals, are well documented in the literature (see e.g. Alpers 1985; da Silva et al. 1998, 2000; Jackson 2007; Magalhaes and da Silva 2017a; Santos-Ferreira et al. 2018, 2019), but are still a subject of considerable research (see Magalhaes et al. 2021, and references therein), and far from being fully understood.

Second, it is important to realise that surveying several years of data in remote regions such as the tropical Atlantic, Indian and Pacific oceans, while collecting evidence of ISWs and their interactions with other-scale phenomena, would be prohibitively out of reach for conventional in situ measurements. Satellite data is clearly the main source in these cases. However, imaging ISWs in light of their interactions with other ocean scales may not be straightforward. Namely, when dealing with satellite images for some time, it becomes clear that choosing the right sensor is key to capture the right phenomenon - especially since time and space resolutions as well as ground swaths need to comply with orbit, data acquisition and transmission and power restrictions (see e.g. chapters 3 and 4 in Robinson 2004).

The same applies to ISWs, but with a few interesting twists. First, as highlighted in the previous sections, ISWs' time and space scales are already observed to range across several

orders of magnitude - this is well-documented e.g. with periods and characteristic widths of a few minutes and hundreds of meters off the Portuguese coast (see e.g. Jeans and Sherwin 2001) to about an hour and several kilometers in the South China Sea (see e.g. Huang et al. 2016). Secondly, they interact with other-scale phenomena - which is the focus in this thesis - while propagating for distances of the order of a thousand kilometers (cf. **Figure 1.3**). Therefore, capturing these waves and the phenomena they interact with cannot be attained without a multiscale approach (as illustrated in **Figure 1.6**).

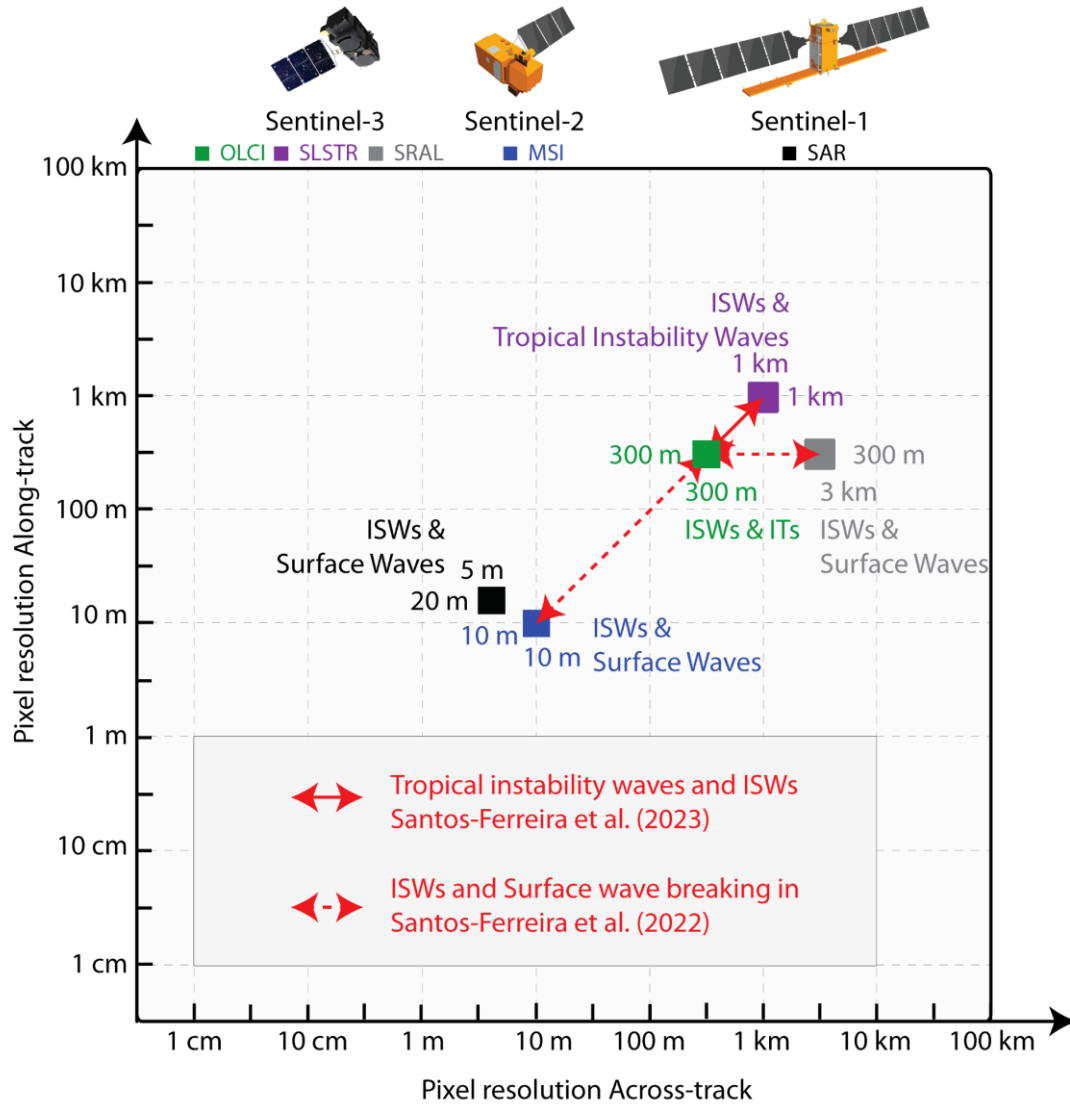


Figure 1.6 - Nominal spatial resolutions in Sentinel sensors used in this thesis (and related studies). Sentinels-1 to 3 (and their sensors) are colour coded in the top inset. Important synergies used in the studies present in **Chapters 2** and **3** (and phenomena therein) are shown using red arrows. See text for more details.

The aforementioned multiscale approach inevitably leads to satellite synergies. As highlighted in **Figure 1.6**, the investigations presented in this thesis strongly relied on the ability of satellite imagery to have enough resolution to capture surface signatures of the waves' individual details (e.g. from Ocean and Land Colour Instrument - OLCI or SAR

altimeter - SRAL in green and grey colours, respectively), but complemented with synergies (including other platforms) suited either for smaller or larger scale phenomena (e.g. MultiSpectral Instrument - MSI or Sea and Land Surface Temperature Radiometer – SLSTR, for surface waves or ocean thermal fronts, in blue and purple colours, respectively). More complex and sophisticated (Level 4) satellite data such as Multiscale Ultrahigh Resolution (MUR) SST, available from NASA Jet Propulsion Laboratory (JPL) PODAAC, also provided valuable information.

Figure 1.6 also highlights yet another important detail when considering data used during the investigations presented in this thesis. Indeed, observations from Sentinel-3 were in a way the key to the ISW interactions studies presented in [Chapters 2](#) and [3](#). For instance, results presented in [Chapter 2](#) were only possible owing to a large survey collecting case studies in which clear surface signatures of ISWs could be unambiguously identified in SAR altimetry. To the date of the study presented in [Chapter 2](#) and in Santos-Ferreira et al. 2022, only Sentinel-3 could offer that synergy, since it provides ocean colour data (OLCI) acquired simultaneously with the enhanced spatial resolutions of SAR altimetry. Similarly, results in [Chapter 3](#) benefited from another large survey in the tropical Pacific Ocean, only this time collecting examples in which ISWs were observed in the vicinities of the Pacific Cold Tongue. Again, unambiguous synergies between surface signatures of ISWs and SST can only be attained via Sentinel-3 using both OLCI and SLSTR.

In the particular case of [Chapter 2](#), it is briefly recalled next some of the details in SAR altimetry, which are of special importance in the framework of this thesis. Dedicated missions of satellite altimetry began in 1992 with the TOPEX/Poseidon mission with its fundamental principle being simply measuring time a microwave pulse needs to travel from the satellite to the sea surface and back, which translates to distance after a series of corrections have been accounted for (Frappart et al. 1999). Nonetheless, over the last three decades, the contents of the returned radar echo have been found to provide additional information about the sea surface (Stammer and Cazenave 2017) - including products e.g. with Sea Surface Heights (SSHs), Significant Wave Heights (SWHs) and wind speeds. More recently, research efforts have been dedicated to introducing SAR processing in radar altimeters, which essentially means upgrading their along-track resolution to a few hundred meters rather than a few kilometers (as in conventional altimeters).

A key problem in this transition is how to keep intercomparable time series in the altimetry data record with these two distinct acquisition geometries (i.e. conventional vs. SAR). One approach to this problem is to investigate how conventional and SAR altimeters compare when intentionally breaking down the assumption of a uniform footprint in the radar echo (known as the Brown assumption; Brown 1977). In turn, the simplest way to do that is to seek ISW sea surface signatures since it is well known that ISWs change dramatically the surface wave field and surface roughness (i.e. capillary-surface wind waves in the range of a few centimeters) in horizontal scales comparable with those of conventional and SAR altimetry (see schematic illustration in **Figure 1.7** below).

These investigations led to previous publications (Santos-Ferreira et al. 2018, 2019, 2022; Magalhaes et al. 2023) showing that ISWs produce measurable signatures in SAR altimeters. One parameter, however, was left without full interpretation in these studies. That parameter is SWH and a novelty presented in this thesis is its interpretation in light of surface wave breaking when modulated by ISWs (Chapter 2).

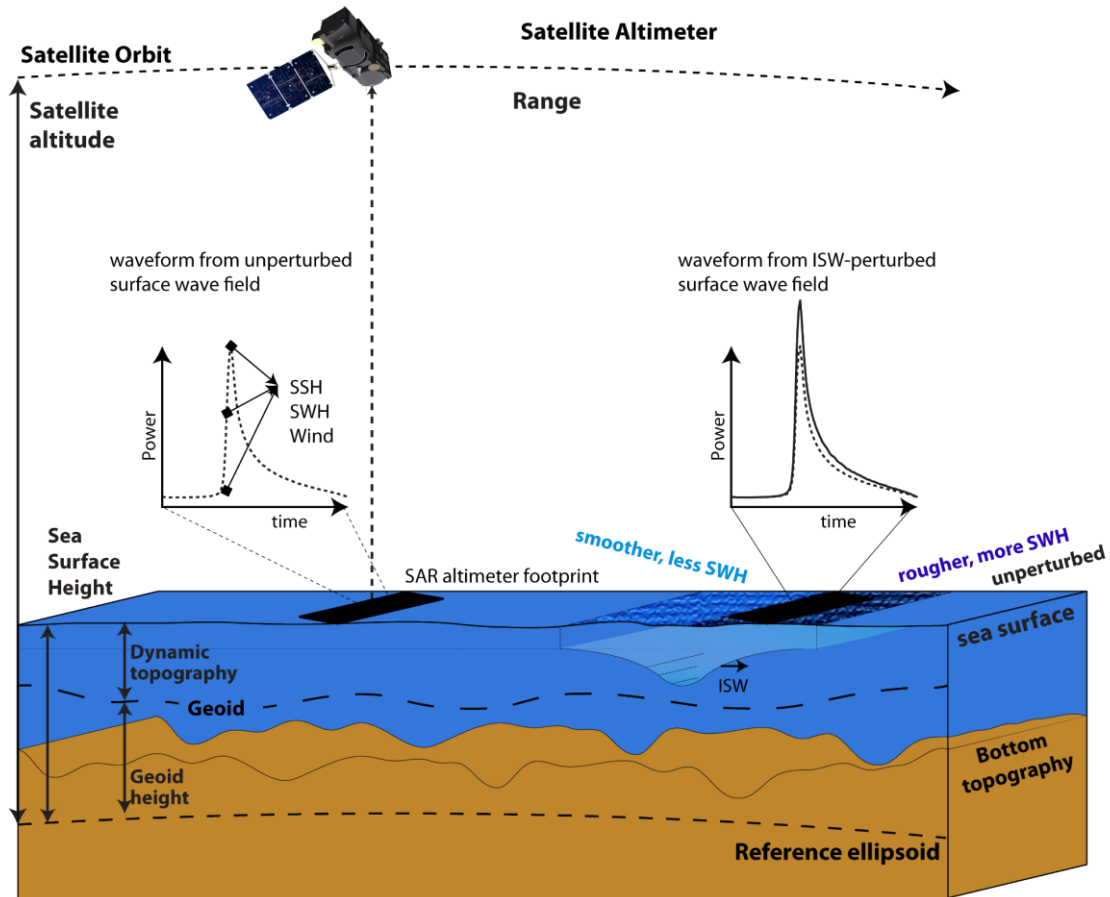


Figure 1.7 - Typical parameters in classical satellite altimetry. Note how a measure of SSH may be obtained from a platform in space. However, altimetric signals resulting from surface echoes that are reflected back to the satellite (often termed waveforms) carry additional information. These waveforms provide operational altimetry products such as SSH, SWH or wind speeds, which may change when affected by surface effects owing to ISWs (see text for more details).

In the particular case of Chapter 3, a few short notes are also in order. Despite having Sentinel-3 as its main data source (as mentioned earlier), contributing in fact for a one-year complete survey during a strong La Niña oscillation in the tropical Pacific Ocean, the ISWs imaged in OLCI and associated with thermal fronts in SLSTR were only part of the picture. To complete the puzzle, a view of the vertical structure in the water column was needed. That cannot be provided from satellite Remote Sensing (which can only *see* the surface or near-surface layers of the ocean), and hence the concept of synergy was widened to include in situ data and numerical simulations detailing the waves' propagation or generation dynamics. The in situ data included monthly mean climatological data provided by NOAA/OAR/ESRL PSD, available at Boulder, Colorado (see Chapter 3 for details), and Array

for Real-Time Geostrophic Oceanography (ARGO) floats when data were obtained close to a satellite observation.

1.4. Motivations and main goals

The investigations discussed in this thesis were mainly motivated by the improved observational capabilities in recently launched earth observation satellites, and possibilities thereof to learn about ocean processes, not restricting our views to their specific time and space scales, but rather expanding them into the larger frames of ocean dynamics. This means examining longer time series when compared with the past in satellite oceanography.

Nowadays we have access to new browsing tools as NASA's Worldview (<https://worldview.earthdata.nasa.gov/>) and the European Ocean Data Lab (<https://www.oceandatalab.com/home>) promoted by ESA, which allow us to visualize easily and interactively global and daily satellite images from different sensors.

Literature discussing how energy is transferred across the oceans' time and space scales can be regarded as being both classical (Munk 1966) and yet of considerable interest in the current state of the art (Warner and Moum 2019; Whalen et al. 2020). At the same time, however, the traditional view of the energy cascade has been recently acknowledged as having both up-scale and down-scale components linking the turbulent near-end of the ocean spectrum with the far-end of climate changes (see e.g. Taylor and Thompson 2023).

Nonetheless, a common view that consistently emerges from these studies is that IWs play a key role in both energy transports and transfers (Shroyer et al. 2010) - interestingly a view that comes forward from theoretical considerations (Munk 1966), in situ measurements (Shroyer et al. 2010), laboratory experiments (Mercier et al. 2012), and numerical models (Bourgault et al. 2016), but which still lacks the unique views of satellite remote sensing. This presents a research opportunity with added motivation from nowadays new generation earth observing sensors whose current trends are to combine synergetic views of large synoptic scales with the details of high spatial resolutions - and hence promising further insights into multiscale processes and their interactions, which has been a key goal since the initial stages of the investigations presented in this thesis.

The case is even more so for processes involving IWs. Historically, observations documenting this phenomenon can be traced back for more than a century (Gerkema and Zimmerman 2008). And for more than a century we have seen significant developments in the literature discussing their theoretical and numerical models as well as their in situ and laboratory views. However, in a way, it could be argued that it is satellite remote sensing that has been connecting the dots between them (e.g. Jackson et al. 2012; Zhang and Li 2022), and further to their importance in the overall dynamics of ocean processes (Taylor and Thompson 2023) mainly by revealing that they are ubiquitous in the world's oceans. Presently, an unprecedented suite of sensors is providing complementary views of ocean processes including ocean color and surface roughness, whose acquisitions are wide enough to capture the IWs propagation history across mesoscale distances with decameter or even meter-scale resolutions (in the latter case SAR imaging sensors). This means "instantaneous"

views of multiscale ocean processes, which in the particular case of IWs offer the prospect of revealing how they interact with both larger and smaller-scale phenomena.

Making the most of the “instantaneous” views of satellite remote sensing means working with synergies, a concept discussed here in more detail in the interest of motivations and goals behind the main results presented in this thesis. Historically satellite data was often regarded as a precious asset (with just a few satellites being used until the late 1990s). Therefore, their insights were complemented whenever possible, within a wide scope of synergy, with other data that included more satellite images or in situ or modelling data - the latter not necessarily simultaneous with the former. Presently, earth observation platforms such as Sentinel-3 allow exact synergies (i.e. different data acquired at the same time) with large fields of view of synoptic scales, which in turn may benefit from a systemic network of in situ observations and online models that cannot go unnoticed to reveal what remote sensing cannot. Therefore, it is worthwhile noting that despite having satellite observations as our main drive to study ISWs and their interaction with other ocean processes, the same goal of exhausting all other available data sources is equally used to uncover whatever details possible.

Consequently, this thesis has two main chapters where the general and specific objectives are addressed. In **Chapter 2** we specifically study signatures in SRAL altimeter (expanding earlier studies, as for example Santos-Ferreira et al. 2018; 2019) – see objective 1 in **Figure 1.8**. **Chapter 3** investigates ISWs in a region where a major hot-spot is observed but whose explanation of their existence there prompts for a new generation mechanism previously overlooked in the deep ocean – see objective 2 in **Figure 1.8**.

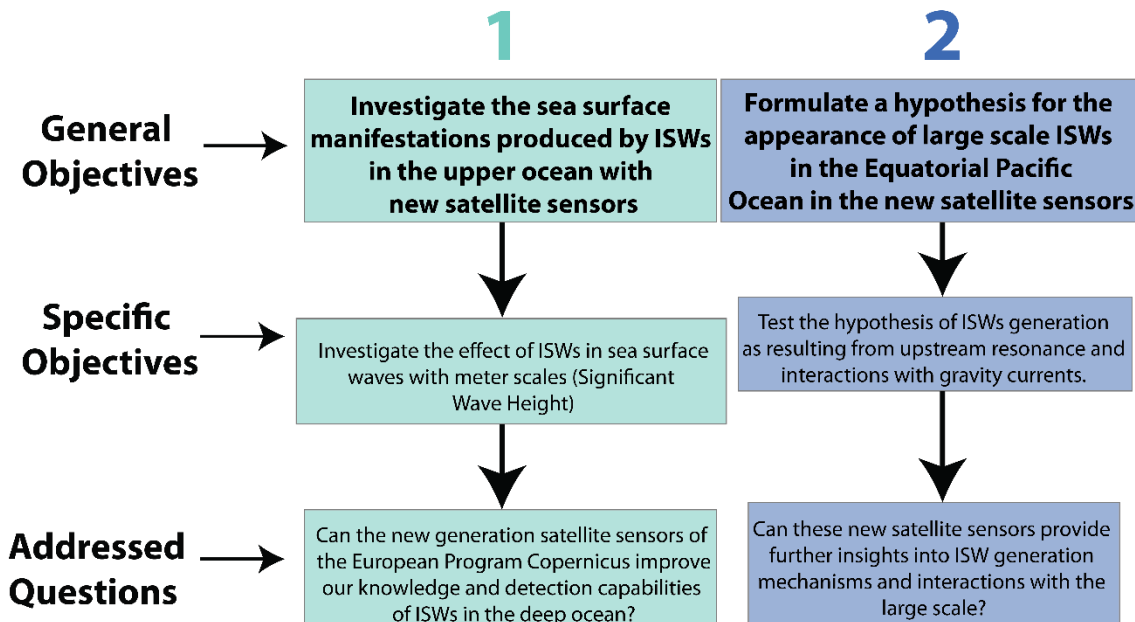


Figure 1.8 - General and specific objectives of this thesis and addressed questions.

1.5. Thesis outline

This thesis is divided into four chapters, including a general introduction, two research articles published in peer-reviewed indexed journals, and a last chapter with concluding remarks. The different chapters are organized as follows:

- **Chapter 1** consists of a general introduction that summarizes the most important concepts, definitions and properties related to ISWs. It begins with a scope and overview of the PhD thesis ([section 1.1](#)) which is meant to convey the general idea that, even though they happen in the ocean interior and are normally not known to the layperson, ISWs are an important redistribution mechanism of energy between different ocean phenomena at different scales. This section is followed by a brief historical description of how ISWs have been studied through time and how they can be originated by larger-scale ocean phenomena (as studied in **Chapter 3**), “feeding” energy to other smaller-scale phenomena (as studied in **Chapter 2**) – [section 1.2](#). In the next section (i.e. [section 1.3](#)) it is briefly described how satellite remote sensing and radar altimetry have been fundamental in the investigation of different phenomena at different scales, including ISWs, with an emphasis on the ESA’s Sentinel sensors used in this thesis. The first chapter ends with the motivations and main goals of the thesis ([section 1.4](#)) and a thesis outline ([section 1.5](#)).

- **Chapter 2** starts with a prologue that describes the historical overview and delineates the main chain of events that motivated the article that follows (Santos-Ferreira et al. 2022). The paper highlights the novelty of enhanced detection of surface wave breaking associated with large-scale ISWs in the Sentinel-3 SAR altimeter. The work is focused on exploring the details of SAR altimetry signals relating to SWH signatures of ISWs at small (1 – 3 km) and intermediate scales (± 10 km). It is shown for the first time that large-scale ISWs propagating in deep waters produce a distinct signature in the SWH record in scales of ± 10 km. This signature is characterized by a decay of SWH amplitudes downstream of the solitary wave crest, when compared to upstream values.

- **Chapter 3** starts also with a prologue that explains how the contents of the chapter originated. In a location where we initially believed there would be no conditions to detect ISWs with Sentinel-3 sensors, we first found a single case that was clear enough to motivate seeking for more events. The motivation arose from curiosity about how ISWs could appear in the equatorial Pacific cold tongue, a region where the tides are weak and there are practically no steep bottom topography features (Egbert and Erofeeva 2002). Despite the severe cloud coverage during most of the year, and with mainly OLCI and SLSTR images available (Sentinel-3), it was possible to formulate a hypothesis about how ISWs emerge, which allowed the writing of the article that follows (Santos-Ferreira et al. 2023). In the article, the characteristics of the ISWs were described in some detail, and a hypothesis for their generation mechanism is proposed.

- **Chapter 4** finally concludes this thesis by presenting some concluding remarks, a summary, and discussion of the most pertinent results presented in **Chapters 2** and **3**, evaluating the initial motivations and main goals stated in [section 1.4](#). As a conclusion of this thesis, it is important to emphasize the urgency of obtaining more observational data, since

few data are available at fine resolutions in the most remote regions of the oceans. Specifically, in [Chapter 3](#), we felt the need for SAR imaging data, which despite the many missions since the 1990s, there are practically no archived data from the equatorial Pacific. Dedicated in situ data to validate our hypothesis are therefore needed, if possible including high resolution radar image data. This chapter ends with some suggestions for future work.

CHAPTER 2

Prologue

The first set of results presented in this thesis uses SAR altimetry to document how ISWs interact with the surface wave field. These results were published in Santos-Ferreira et al. (2022) and are presented below. Before proceeding, however, a brief account is warranted describing how previous investigations inspired, and ultimately led to, a sequence of events leading to interpretation in satellite observations relating to surface breaking waves.

This account begins with a peculiar study region in the tropical Atlantic Ocean off the Amazon shelf (see **Figure P2.1** below for location), where ISWs can be frequently observed in satellite images (see also Magalhaes et al. 2016). Some of the reasons why this particular study region is unique are as follows:

- First, ISWs are large-scale and they can propagate for distances of the order of a thousand kilometers. The individual waves in each packet (especially the leading wave which is the largest) have typical wavelength of the order of 10 kilometers (see e.g. Figures 1 to 4 in Santos-Ferreira et al. 2018), and crest lengths extending for typically 200 km or more. This means more sampling area is available for the SAR altimeter (as explained in [section 1.3](#)) to distinguish different ISWs, their leading and trailing sections, and how they depart from an unperturbed background.
- Second, ISWS in this region propagate in a near-meridional direction, which is approximately aligned with SAR-altimetry ground tracks in Sentinel-3 (cf. black lines in **Figure P2.1a**). Consequently, the geometry of the SAR altimeter footprint is optimally aligned to sample the waves' rough and smooth sections (as explained previously in [section 1.3](#)).
- Finally, this region benefited from two international projects dedicated to investigate ISWs off the Amazon shelf-break (see Santos-Ferreira et al. 2018, 2019, 2022). This meant more research time and efforts could be made available in this particular study region and largely devoted to their sea surface signatures in satellite remote sensing (namely in the SAR altimeter onboard Sentinel-3).

Altogether, the outcome from having these unique conditions for ISWs off the Amazon shelf-break, were two investigations demonstrating that SAR altimetry (using Sentinel-3) can certainly detect ISWs owing to their surface signatures in different geophysical parameters such as radar backscatter (σ_0), Sea Level Anomaly (SLA) and Significant Wave height (SWH), see Santos-Ferreira et al. (2018, 2019). However, while these findings answered key questions (e.g., regarding SAR vs. conventional altimeters), they would soon raise others.

In particular, two puzzling features kept emerging in SWH, consistently associated with the transitions in ISWs' between their rough and smooth sections, as well as with their surroundings immediately ahead of and behind the wave packets (with typical lengths of ± 10 km). These are seen e.g. in Figure 4 in Santos-Ferreira et al. (2018), where a localised increase in SWH is seen in the leading rough section of the leading ISW, but where it is also

hinted that two distinct plateaus in SWH separate sections before and after the wave passage - i.e. one ahead (unperturbed by it), and one (significant lower) behind it. Clearly, something in the largest surface waves was being consistently affected by ISWs, but how could that be interpreted in light of the SAR altimetry data?

Part of the answer would come from a different sensor, still a SAR, but of the image kind (see also [section 1.3](#)). In a related study - also including the waves off the Amazon shelf-break - Magalhaes et al. (2021) showed that in SAR images, surface wave breaking can have appreciable contributions to backscatter. In that study, data from SAR altimetry was processed to reinforce this result, since it showed that SWH shows a similar signature, which result is reproduced here in **Figure P2.1** below. It shows a localized increase located at the leading sections of the ISW, which marks a decrease between averaged backscatter ahead and behind the ISW (see **Figure P2.1c**).

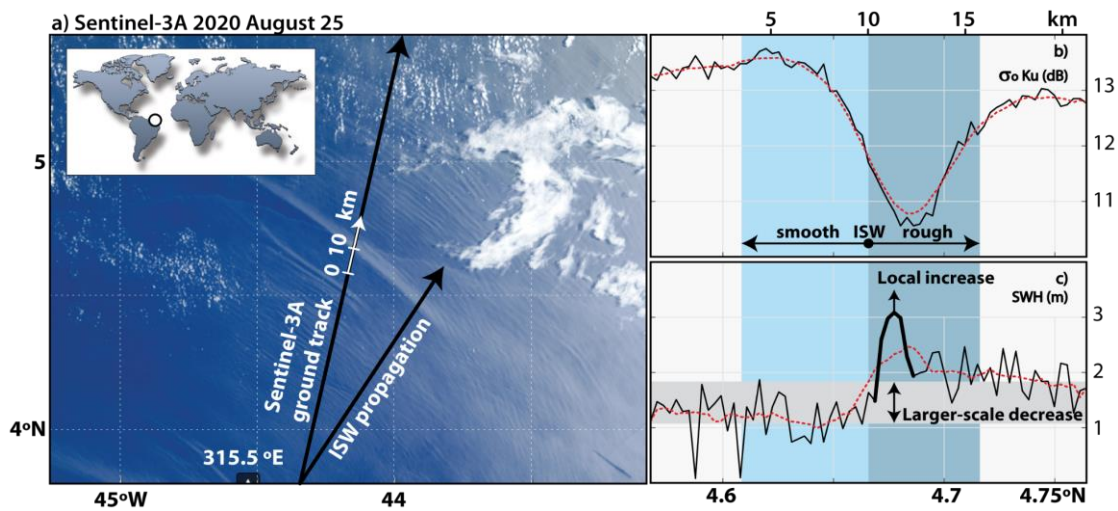


Figure P2.1 - OLCI image acquired on August 25, 2020, at 11:22 UTC (see inset for location approximately off the Amazon shelf-break). The black arrows indicate the ground track of the satellite and the ISW direction of propagation. (b) SRAL backscatter acquired in exact synergy with OLCI. (c) Same as (b) SWH. A grey envelope highlights a decrease in SWH when comparing sections ahead and behind the ISW. Blue shades distinguish the ISW's leading (rough) and trailing (smooth) sections. The red dashed lines are box moving averages.

It was this finding that motivated additional research efforts to investigate other large-scale ISWs (made possible via the Multi Sensor Synergy Study for Sentinel-3C/D) in other known hotspots (e.g. the Banda Sea in the Indian Ocean), and to explore in detail the contents of SAR altimetry signals relating to SWH. The results of that study (Santos-Ferreira et al. 2022) are shown next (in [Chapter 2](#)), which highlights the novelty in enhanced surface wave breaking (in Sentinel-3) associated with large-scale ISWs, and further corroborated with Sentinel-2 MSI images.

EFFECTS OF SURFACE WAVE BREAKING CAUSED BY INTERNAL SOLITARY WAVES IN SAR ALTIMETER: SENTINEL-3 COPERNICUS PRODUCTS AND ADVANCED NEW PRODUCTS

(Santos-Ferreira et al. 2022)

Abstract

We address surface wave breaking caused by oceanic Internal Solitary Waves (ISWs) and how ISWs are manifested in the SAR altimeter onboard Sentinel-3A and -3B satellites by means of their effects in Significant Wave Height (SWH). Two different regions of the ocean are selected, namely the tropical Atlantic Ocean off the Amazon shelf and the Banda Sea in the Indian Ocean, where there are scenes of Sentinel-3 OLCI acquired simultaneously with an along-track SAR mode altimeter, which include signatures of large amplitude ISWs. New data of unfocused SAR (UF-SAR 20 Hz) and fully focused SAR (FF-SAR 160 Hz) modes are analyzed, which are retracked in full range and over a reduced range of bins (truncation carried out dynamically ten gates away from the estimated epoch position). At first order, in scales of 1–3 km, a strong decrease in the normalized radar cross section (NRCS) over the rough part of the ISWs is observed followed by a small increase in the smooth part relative to the unperturbed ocean background. A second order ISW signature, in scales of 20 km, is noted: the SWH is attenuated after the passage of an ISW, considering length scales of about 10 km before and after the ISW crest. The SWH signatures are unique in showing that the surface wave energy does not return to its unperturbed level after the passage of an ISW, admittedly because intense meter-scale wave breaking results in surface wave energy dissipation. Furthermore, Sentinel-2 MSI images are analyzed and provide insights into this same phenomenon: white-capping resulting in a radiance increase at all (visible) wavelengths. Modulation of breaking waves owing to ISWs is demonstrated by estimates of the fraction of breaking waves in the presence of internal waves.

Index Terms - internal solitary waves (ISWs); wave breaking; delay-Doppler altimeter; radar backscatter; altimetry; SAR; Significant Wave Height (SWH); mean square slope (mss).

2.1. Introduction

2.1.1. Oceanic Internal Waves, Their Surface Manifestations and Remote Sensing

Oceanic internal waves are waves propagating in the interior of the ocean, usually along density interfaces such as the seasonal or permanent pycnocline. They are allowed to propagate when the ocean consists of layers of different densities, i.e., when the ocean is

stratified, and when there is a physical mechanism that can perturb the interface between the layers, which, in coastal areas, is most often tidal flow over shallow bottom topography gradients. In most cases, the internal waves are nonlinear and appear in wave packets consisting of several internal solitons and are hence often called Internal Solitary Waves (ISWs). We will adopt this terminology throughout this paper. The ISWs become detectable from space primarily via ISW-induced variation of the sea surface roughness, which is detectable by optical and radar sensors, and in particular by Synthetic Aperture Radars (SARs). Historically, ISWs have first been detected from space on photographs taken by astronauts, who have pointed their hand-held cameras into the sunglint region. The sunglint region is the sea surface area from which the sunlight is reflected at roughly the same angle as the camera is viewing the surface. Since 1972, ISWs have also been detected on sunglint images taken from satellites, such as the American Landsat-1, which was called at that time the Earth Resources Technology Satellite (Apel et al. 1975). However, optical images have their limitations in ISW studies, since they can be taken only during daytime when the sky is cloud-free or almost cloud-free and when certain criteria concerning the viewing geometry are fulfilled.

A breakthrough in ISW remote sensing from space came with the launch of the American Seasat satellite in 1978. The L-band (1.25 GHz) SAR on this satellite captured numerous ISWs (Fu and Holt 1982; Apel and Gonzalez 1983; Alpers and Salusti 1983). Since then, many more ISW images have been acquired from various satellites, such as the European ERS-1/ERS-2, Envisat, and Sentinel-1A/1B satellites, the Canadian Radarsat-1/2 satellites, the Japanese ALOS satellite, and the Chinese Gaofen-3 satellite. ISW sea surface signatures have also occasionally been detected in ocean color images (Kim et al. 2018; Zhang et al. 2020) and recently also in radar altimeter data (Zhang et al. 2020; Magalhães and da Silva 2017; Santos-Ferreira et al. 2018, 2019; Magalhães et al. 2021).

There are two fundamentally different radar-based sensors operating from orbiting satellites; they are side-looking imaging SARs, which look down at an oblique angle to the side of the satellite orbit, and nadir-looking instruments that are pointed directly down towards the Earth. Despite the fact that both types of sensors are active instruments, meaning that they emit their own microwave pulses and record the subsequent echoes, their data acquisition geometries are intrinsically different, and they survey different aspects of ocean dynamics.

On the one hand, imaging SARs are sensors which acquire two-dimensional measurements of the ocean's horizontal surface. They comprise very high spatial resolutions (up to the order of meters), in which backscattered radiation returning into the satellite direction is described, to a first order, by a Bragg scattering mechanism. Bragg scattering is particularly successful to account for sea surface signatures of small-scale phenomena via their changes in sea surface roughness mainly induced by wind wave and surface current variability. On the other hand, satellite altimeters are not imaging sensors. The recorded echoes are registered along a single "line" corresponding to the satellite's ground-track, and hence their data are usually referred to as along-track data. This "line" corresponds to the

satellite's flight direction (or azimuth) projected onto the earth's surface, and nominal spatial resolutions are of the order of kilometers—although new generation altimeters can reduce this to a few hundred meters in the along-track direction.

Nonetheless, both radar-based sensors use microwaves as their sensing radiation. This means using wavelengths of the order of centimeters (e.g., Ku and C bands have nominal wavelengths of about 2 and 4 cm, respectively), which in turn interact with similar length scales of the sea surface (note that the sea surface is essentially opaque to microwaves). Therefore—to leading order—they are essentially sensitive to centimeter-scale waves, which are mostly wind-generated in the ocean and hence (in practice) an intrinsic feature of the sea surface. It is well known that wind waves will naturally interact and group in progressively larger waves (see e.g., Babanin 2011). When wind waves are continuously fetched by the sea surface (i.e., assuming there is at least a mild wind), these growing waves will continue to increase in amplitude until some equilibrium balance is reached—upon which they eventually break, and their energy (which had been accumulating from the wind) dissipates. This brings about an important question: Since breaking surface waves are an inherent part of the ocean wave field, how do they affect the radar signature of other oceanic phenomena, such as oceanic fronts, oceanic eddies, and internal waves?

In principle, isolating the contributions owing to surface breaking waves from other potential contributions could be beneficial for understanding and quantifying different radar backscattering mechanisms. This can be achieved by having a nominal background (i.e., sea surface unperturbed by current shears and wind variability) against which two contrasting sea surfaces could be compared: one having little (or no) surface breaking waves, and another having a significant amount of enhanced surface wave breaking. Furthermore, the corresponding horizontal scales should be sufficiently large to ensure distinct transitions in the remotely sensed radar ground footprints—preferably located in deep ocean conditions and away from the more dynamical coastal environments. Interestingly, these exact conditions can be naturally found in deep ocean conditions, e.g., in the Tropical Atlantic Ocean off the Amazon River mouth. Consequently, large-amplitude ISWs with horizontal wavelength scales around 1 to 10 km and crest lengths of more than 100 km, which typically include a rougher leading section (often with patches of enhanced surface breaking and whitecaps) and a smooth (slick-like) trailing section, are ideally suited to investigate enhanced surface wave breaking contributions to both satellite SAR altimeters and imaging sensors.

The paper is organized as follows: in [section 2.1.2](#), we address wave breaking in the context of ISWs' manifestations, e.g., quoting from a report written in the 19th century in which mariners describe observations of roughness bands in the Andaman Sea, which must have been sea surface signatures of strong ISWs, and then we present a typical photograph taken from a ship which shows sea surface signatures of ISWs containing breaking surface waves. All the data and methods used in this work are described in [section 2.2](#). In [section 2.3](#), we present new observations from SAR altimetry (from the Sentinel-3 mission) with clear evidence of Significant Wave Height (SWH) variations along the propagation paths of ISWs

in scales of about 20 km or more. New advanced SAR altimetry products (e.g., Fully Focused FFSAR) are presented and compared with standard Copernicus products. Evidence of wave breaking in high-resolution optical images owing to large ISWs is also presented in [section 2.3](#). Finally, in [section 2.4](#), we summarize our results and conclusions.

2.1.2. Surface Wave Breaking Induced by Strong ISWs

Wave breaking is a ubiquitous phenomenon in the ocean at high to medium wind speeds. It generates turbulence in the upper ocean layer and causes dissipation of wave energy. However, ocean surface waves can break also in the absence of wind, for instance when a swell shoals on a beach. This paper deals with breaking surface waves, which are not generated by the action of the wind, but by a resonant interaction with an ISW as will be discussed later in this section. There are numerous visual observations, in particular from ships, that ISWs are associated with bands of choppy waters containing surface breaking waves. An early description of this phenomenon can be found in the book of Maury 1861, which is quoted in Osborne and Burch (1980). This book contains reports about roughness bands or rippings, which are often observed from ships in the Andaman Sea: “The rippings are seen in calm weather approaching from a distance, and in the night their noise is heard a considerable time before they come near. They beat against the sides of a ship with great violence, and pass on, the spray sometimes coming on deck; and a small boat could not always resist the turbulence”. “The rippings consist of broken water, which makes a great noise when the ship is passing through the rippings in the night”. At that time, it was not known that the roughness bands are sea surface manifestations of ISWs. However, Perry and Schimke (1965), who made oceanographic surveys in the Andaman Sea north of Sumatra from a ship in 1964, identified the roughness bands or “rips” as sea surface signatures of “large-amplitude internal waves”. They observed roughness bands, which ranged from 200 to 800 m in width, stretched from horizon to horizon (approximately 30 km), and had a spacing of about 3200 m between the bands. The bands of choppy water contained short, steep, randomly oriented waves with heights of about 0.3 to 0.6 m. Each band stood out distinctly in an otherwise undisturbed sea (wind speed: 4 m/s). Moreover, Osborne and Burch (1980) observed roughness bands in the Andaman Sea during a campaign, which was designed to measure internal waves and the associated response of a drillship simultaneously, and attributed them to sea surface manifestations of ISWs. On one occasion, they observed six distinct roughness bands, each approximately 600 to 1200 m wide, with breaking waves about 1.8 m high. After the passage of the roughness band, the wave heights quickly decreased in wave height to less than 0.1 m, and the sea surface had the appearance of a “millpond”. The entire process repeated at approximately 40 min intervals.

Enhanced surface wave breaking associated with large-scale ISWs is now widely accepted, and in **Figure 2.1** we present a photograph taken from Research Vessel (R/V) Charles Darwin in the central region of the Bay of Biscay (approximately 46°N, 7°W), in which a band of strongly increased sea surface roughness containing surface breaking waves is visible. It is associated with a large-amplitude internal wave generated by a tidal internal

wave beam in deep water (New and da Silva 2002). It seems that the breaking waves have wavelengths in the order of meters. In these large-scale ISWs (with horizontal scales of the order of kilometers), enhanced effects in surface wave breaking may be imaged in high spatial resolution satellite images. **Figure 2.2** shows an image taken in the visible band by the Multispectral Instrument (MSI) onboard the Sentinel-2A satellite (resolution: 10 m) showing also sea surface signatures of an internal wave packet in the Banda Sea (in the Indonesian Seas) generated over a sill in the Ombai Strait in between the Alor and Atauro Islands, one of the major passages of the Indonesian Through Flow. In this case, whitecaps are visible (**Figure 2.2c**, and respective inset to the right) which are generated by breaking waves associated with the internal wave.

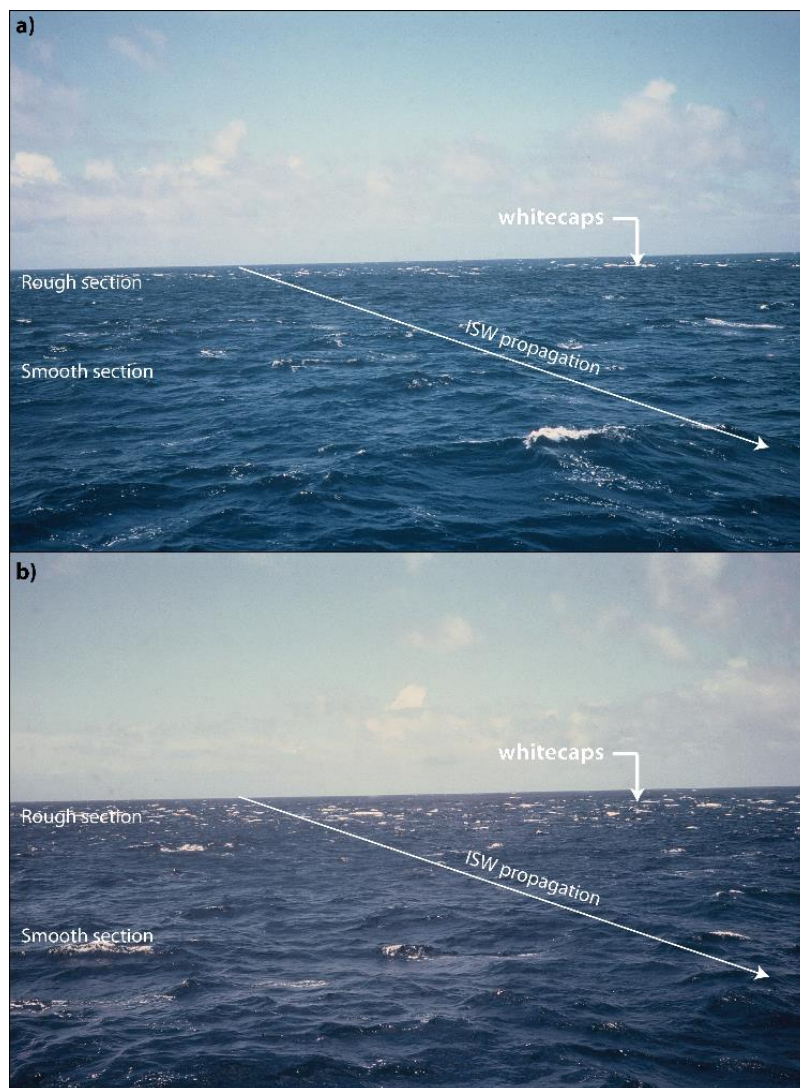


Figure 2.1 - Photographs taken from the Research Vessel Charles Darwin showing sea surface manifestations of a large-amplitude ISW advancing towards the ship in the central region of the Bay of Biscay, which includes surface wave breaking (i.e., whitecaps). Note that both panels (a,b) are just a few minutes apart, and wave breaking has moved towards the ship from (a) to (b). Courtesy: Adrian L. New from the National Oceanography Centre, Southampton.

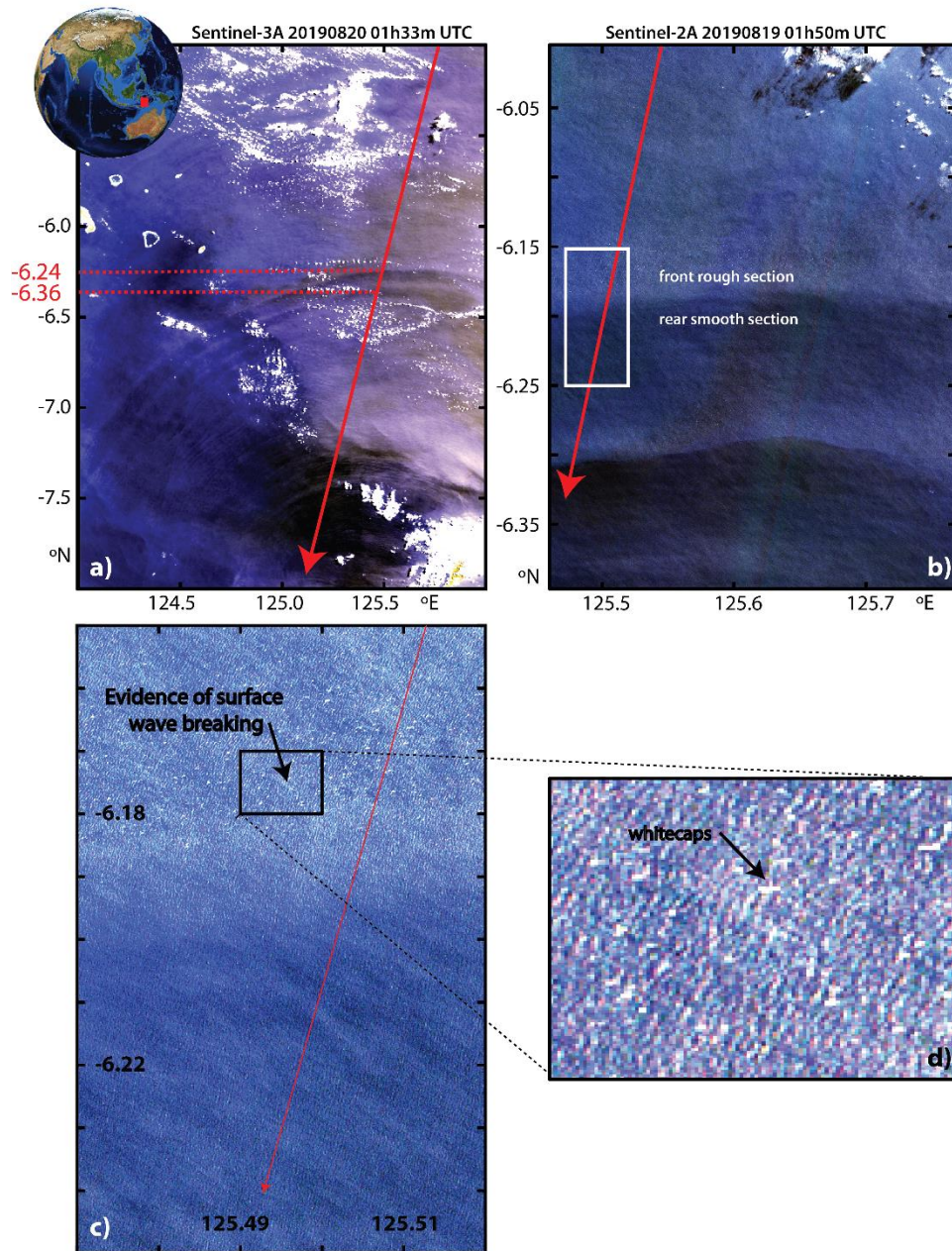


Figure 2.2 - (a) Quasi-true color image (Level-1b) acquired on 20 August 2019 at 01 h 33 m UTC by the OLCI sensor onboard the Sentinel-3A satellite (see inset for location). The red line represents the ground-track of the satellite, which in this case is close to the ISWs' direction of propagation. (b) Image taken in the visible band by the Multispectral Instrument (MSI) onboard the Sentinel 2A satellite on 19 August 2019 at 01 h 50 m UTC showing sea surface signatures of an ISW packet. Note the bright spots aligned in bands, which are evidence of whitecaps generated by breaking surface waves. The red line represents the ground-track of the satellite Sentinel-3, approximately one day after. (c) Zoom of the Sentinel-2 image in (b), in the location of the white rectangle. A larger zoom is also displayed in (d), representing the black rectangle in full resolution, and showing whitecaps.

As stated before, we attribute the roughness bands associated with ISWs as caused by a resonant interaction between surface waves and the internal waves. Such interaction has been studied theoretically (see e.g. Philips 1973; Lewis et al. 1974; Craig et al. 2012; Kodaira et al. 2016; Taklo and Choi 2020) and has been verified in laboratory experiments (Lewis et

al. 1974; Kodaira et al. 2016). Resonance occurs when the group velocity of the surface waves ($C_{\text{group}}^{\text{SW}}$) matches the phase speed of the internal waves ($C_{\text{phase}}^{\text{iW}}$):

$$C_{\text{group}}^{\text{SW}} = C_{\text{phase}}^{\text{iW}} \quad (2.1)$$

While Lewis et al. 1974 studied this interaction for a monochromatic internal wave, Kodaira et al. 2016 studied it for an ISW. The slowly varying surface currents associated with the ISW interact with the surface waves, and those waves which are in resonance or near-resonance with the varying surface current field steepen and give rise to the appearance of short-wave packets on the sea surface. According to the theory of Kodaira et al. 2016, the wave packet is located above the front half of the ISW, and propagates with a speed close to that of the ISW underneath, which is in agreement with observations. Applying **equation 2.1** and using the dispersion relation for deep water waves, the resonant surface wave has a wavelength of:

$$\lambda_{\text{resonant}}^{\text{SW}} = (8\pi/g) (C_{\text{phase}}^{\text{iW}})^2 \quad (2.2)$$

For a typical phase speed of 2 m/s for an ISW, we obtain $\lambda_{\text{resonant}}^{\text{SW}} = 10$ m, which is a typical wavelength of an “intermediate-scale wave”. It has long been suspected that these waves are strongly modulated by internal waves (Lyzenga and Bennett 1988; Plant et al. 2010), and that their breaking on the sea surface contributes to the high radar backscatter signatures of internal waves. When these surface waves break, they generate turbulence in the upper surface layers and dissipate energy. The breaking stage is maintained by replenishment of energy from the ISW via the resonant interaction. It is often assumed that surface wave breaking occurs when the phase velocity equals the orbital velocity of the wave, which results in $H \geq \lambda/2\pi$, where H is the surface wave height and λ its wavelength. However, Ochi and Tsai (1983) propose the condition $H \geq 0.020gT^2$, where g is the acceleration due to gravity and T the surface wave period, which results from laboratory measurements with random waves. In terms of wavelength the condition reads:

$$H \geq 0.142 \lambda^{\text{SW}} \quad (2.3)$$

Note that **equation 2.3** is consistent with Stokes (1880) criterion for surface breaking waves. Therefore, for a typical wavelength λ^{SW} of 10 m, we obtain for the height of the breaking surface wave $H = 1.5$ m. Wave heights of this order, but slightly larger (1.8 m) have been measured in roughness bands associated with an ISW in the Andaman Sea by Osborne and Burch (1980), where the speed of the ISW was 2.2 m/s and thus the wavelength of the resonating surface wave $\lambda_{\text{resonant}}^{\text{SW}} = 12.4$ m (which would yield $H = 1.76$ m). As will be shown in **section 2.3**, we have obtained similar values in our analysis of Sentinel-3 altimetry data.

2.2. Material and Methods

In previous work it has been discussed how surface wave breaking (resulting from ISWs) is thought to affect the radar backscatter in imaging SARs (see e.g. Kudryavtsev et al. 2014; Magalhaes et al. 2021). Since SAR altimeters, with enhanced spatial resolution, also use active

microwaves to sense surface roughness, it is reasonable to assume that surface wave breaking will make a measurable signature in their radar returned echoes – even though they have a different nadir-looking acquisition geometry and operate in the more energetic part of the microwave spectra.

2.2.1. Fundamentals of satellite altimetry and its sea surface signatures of ISWs

Before proceeding to investigate possible contributions from surface wave breaking in satellite altimeters, it is important to explain the fundamentals of satellite altimetry and the mechanisms that allow it to detect the sea surface manifestations of ISWs. In essence, satellite altimeters use nadir-pointed radar pulses (i.e. microwave pulses emitted directly downwards in the vertical from the satellite) to “illuminate” the sea surface and record their subsequent microwave echoes. The altimeter echoes are recorded along the satellite’s flight (or azimuth or along-track) direction, meaning data is recorded and related to a single footprint on the ground, and along one-dimensional paths (i.e. the satellite’s ground-track) rather than in a two-dimensional image format.

The primary and most traditional measurement made by an altimeter is the two-way travel time from radar echoes received from the backscatter radiation, which of course translates to a distance after a series of very accurate corrections in microwave propagation speeds are accounted for along the two-way travelled distance through the Earth’s atmosphere. These distances can then be referenced to a mean sea level to yield e.g. the well-known Sea Level Anomalies (SLAs) or sea level height anomalies, once the Mean Sea Surface is used.

The altimeter’s echoes that are reflected backwards to the satellite antenna are recorded onboard over a period of time, meaning that not only times but also the power and shapes of the backscattered radar intensities can be used to estimate geophysical properties of the sea surface. These echoes are usually termed waveforms, and each waveform corresponds to an emitted pulse which returns to the sensor after reflection at the sea surface. For the ocean, surface waveforms typically follow a well-known pattern. An illustrative example is shown in **Figure 2.3a**, which is seen to begin at the sensor’s noise floor (preceding the echo’s arrival, also denominated thermal noise), then raising rapidly as the emitted pulse returns to the sensor (often termed the waveform’s leading slope), and finally decreasing along a trailing edge. Well-established algorithms (called retracking algorithms) are used to fit the recorded waveforms to those expected from theory and hence provide estimates for several geophysical quantities that are typically used in ocean sciences (Brown 1977; Dinardo et al. 2015). The Normalized Radar Cross Section (NRCS), which is a measure of the power received in the radar backscatter depends mostly on the small-scale sea surface roughness (about three times the radar wavelength, i.e. greater than about 7 cm for Ku-Band), and relates closely to the waveforms’ maximum power. The Significant Wave Height (SWH), which is a statistical measure of the highest third of the waves of the local wave spectrum, relates to the waveforms’ leading slopes (see **Figure 2.3a**).

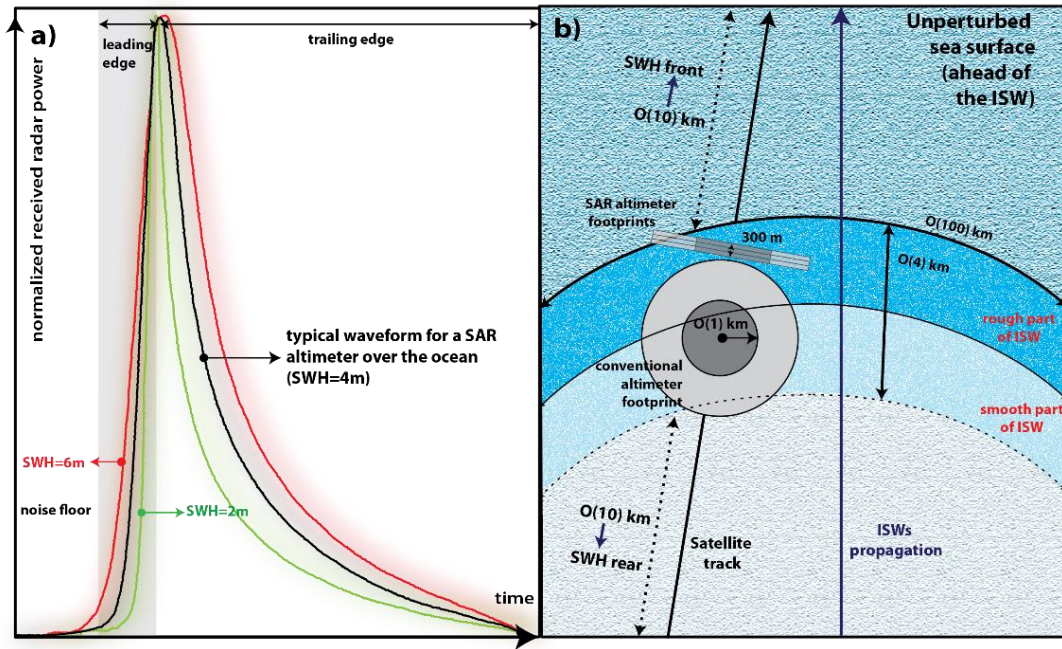


Figure 2.3 - (a) Normalized SAR altimeter waveforms for an unperturbed (in relation to the ISWs' surface manifestations) sea surface (black) and for different SWHs (green and red). (b) Schematic representation of a typical ISW and footprints of a conventional (circular shape) and a SAR altimeter (SRAL, rectangular shapes). Dark gray represents illuminated areas corresponding to maximum backscattered power, and light gray the decaying part of the echo due to off-axis antenna gain attenuation.

It is well known that Internal Tides (ITs, i.e. IWs of tidal period with horizontal length scales of order of tens to hundreds of kilometers) can be successfully mapped using sea surface height data from satellite altimeters (see e.g. Ray and Mitchum 1996). Magalhaes and da Silva (2017a) have shown that conventional satellite altimetry can also detect the sea surface roughness signatures of large-scale ISWs (i.e. with horizontal scales of about 10 km), in which the waveforms were found to change significantly. In particular, the high-sampling rate (20 Hz) of the Jason-2/3 altimeters made it possible to detect several short-period signatures in the NRCS (or σ_0) data, which resulted from ISWs.

2.2.2. Sentinel-3 SAR Altimeter and Detection of ISWs

New generation satellite altimeters, which use SAR processing in the along-track direction to simulate larger antenna apertures, increase the along-track ground resolution. They are commonly known as SAR altimeters (e.g. those flying onboard ESA's Cryosat-2 and Sentinel-3 missions and more recently on the Copernicus Sentinel-6 Michael Freilich satellite). Nonetheless, this SAR processing, which uses the echoes' Doppler Effect induced by the satellite motion, can only be used in the sensor's direction of flight – i.e. in the along-track direction. This means that, in practice, the ground resolutions of these more advanced altimeters are in fact sharpened to about 300 meters in the along-track direction (nominal along-track resolution), but still remain the same as those of conventional altimeters in the across-track direction (i.e. more than 1 kilometer as illustrated in Figure 2.3b).

The reduced along-track footprint of SAR (or Delay/Doppler) altimeters relative to conventional altimeters is achieved through a technique developed in the 1990s (see Raney 1998), by which return radar echoes are recorded in both range bins (time domain) and Doppler frequency shift. Although the reflections from directly nadir have no Doppler shift, those reflections from a little in front of the sub-satellite point are shifted to a slightly higher frequency and those behind to a lower frequency. This is made possible thanks to the SAR altimeter capability of operating at high pulse repetition frequency (PRF) (at higher rate than conventional altimeters) which ensures the needed “pulse-to-pulse” radar waves coherence for delay/Doppler signal processing.

Depending on how ISWs crests are oriented relative to the altimeter ground track, this means that the altimeter’s footprint (i.e. the area of the illuminated sea surface) will be distributed differently between the unperturbed background (i.e. unaffected by ISWs) and the ISWs sea surface manifestation patterns (as shown in **Figure 2.3b**). Note that individual waveforms can come from just one of these characteristic footprints, which may or may not exhibit strong contrasts relative to the unperturbed background according to the orientation of the sharpened along-track footprint in relation to the ISW crests and troughs. If the sharpened along-track direction of the satellite is aligned with the waves’ direction of propagation it is likely that large-scale ISWs produce a clear signature in the SAR radar altimeter. This is the case shown in [section 2.3](#) for the Banda Sea, and to some extent is also the case for the Tropical Atlantic Ocean cases shown in this paper.

Santos-Ferreira et al. (2018, 2019) have already documented how the sea surface roughness signatures of large-scale ISWs can affect the enhanced resolution waveforms in SAR altimeters. They used optical imagery from Sentinel-3 (from the Ocean and Land Color Instrument - OLCI) to visually inspect and validate the sea surface manifestations of ISWs that could be unambiguously identified (in synergy) with the SAR altimeter data, which is acquired simultaneously from the SAR altimeter (i.e. the SRAL) onboard the Sentinel-3 satellites. It was shown that the SRAL footprint is small enough (about 300 meters in the along-track direction) and it can capture radar power fluctuations over consecutive ISW crests and troughs, and hence over rough and slick surface patterns arrayed in parallel bands with scales of a few kilometers.

The morphology of SAR waveforms is also significantly altered in the presence of ISWs – including SWH estimates, since the leading slopes (in SAR waveforms) feature a variable steepness between the internal waves’ leading and trailing sections (see Figures 3 and 4 in Santos-Ferreira et al. 2018). This feature is relevant in the framework of this paper because SAR waveforms are sensitive to SWH along an ISW section (crest and trough), and appear to be sensitive to enhanced surface waves within a given signature of ISWs. Hence, re-tracked geophysical parameters such as σ_0 or SWH are suited to investigate possible contributions in the altimetry echoes resulting from enhanced surface wave breaking.

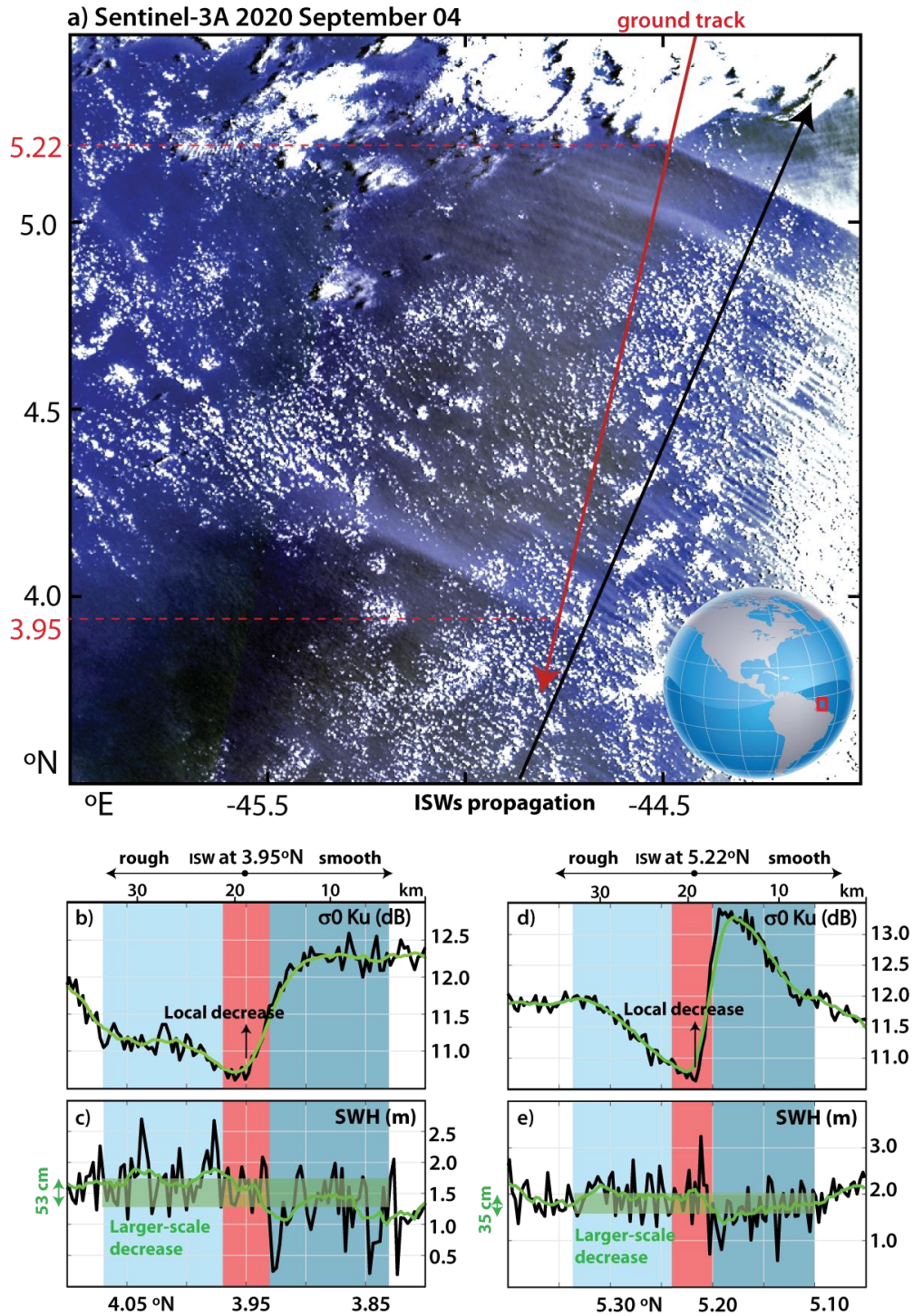


Figure 2.4 - (a) Quasi-true color image (Level-1b) acquired on 4 September 2020 at 13h01m UTC by the OLCI sensor onboard the Sentinel-3B satellite (see inset for location). The red line represents the ground-track of the satellite, which in this case is practically parallel to the ISWs' direction of propagation. (b) Along-track 20 Hz NRCS (i.e., σ_0) measured by the SRAL altimeter showing variations of approximately 2 dB in the leading ISW around 3.95°N. (c) SWH 20 Hz for the same ISW showing a larger-scale variation of 0.53 m when comparing the section in front of the leading wave with that in the rear. (d) Same as (b) for the leading ISW around 5.22°N, with variations of approximately 3 dB (e) Same as (c) showing a larger-scale variation of 0.35 m. Recall that the ISW propagates northwards, hence from right to left in these panels.

2.2.2.1. Sentinel-3 Unfocused SAR (UFSAR)

In this paper we explore the radar altimeter system on board Sentinel-3 satellites in unfocused SAR mode (UFSAR) and Fully-Focused SAR mode (FFSAR) over the open ocean. This system is composed by the SAR Radar Altimeter (SRAL) together with the MicroWave Radiometer (MWR) and Precise Orbit Determination (POD) sub-systems, for ocean topography measurements.

As implemented in the Sentinel-3 mission, the SRAL emits patterns of 64 coherent Ku-band (13.575GHz, bandwidth=320MHz used) pulses in (closed) bursts at a PRF of approximately 18 kHz, enclosed by two C-band (5.41GHz, bandwidth=290MHz used) pulses to provide ionospheric bias correction (see e.g., Donlon et al. 2012; Boy et al. 2017). The SAR altimetry processing first consists in coherently combining pulses from a burst to generate Doppler beams that are further steered to different surface samples location on the ground. A set of Doppler echoes originating from different bursts are then gathered at each surface sample (typically 212 looks) to form a stack that can subsequently be incoherently averaged to produce a multi-looked echo (also termed SAR-mode echo) with a much higher signal-to-noise ratio (see Boy et al. 2017, and references therein for more details). The multi-looked echo has a beam-limited illumination pattern in the along-track direction while maintaining the pulse-limited form in the across-track direction. Hence, a sharpened along-track spatial resolution (around 300 m for Sentinel-3 SRAL) is achieved whereas in the across-track direction, it is still limited to the diameter of the pulse-limited circle (see **Figure 2.3b**). A mean least square estimator (retracking algorithm denominated SAMOSA, see Ray et al. 2014 for details) is adjusted to each multi-looked echo, and hence it is possible to retrieve geophysical parameters such as range, SWH and σ_0 that is inferred from the amplitude of the fitted waveform. The shape of the multi-looked echo has a steeper leading edge than waveforms from conventional radar altimeters, and thus the range is determined more precisely (within less than 1.3 cm at 2 m SWH in SAR mode on board Sentinel-3, see Raynal et al. 2016).

Sentinel-3 SRAL Non-Time-Critical (NTC) Marine Level 2 products are used in the first part of this study. The data is distributed by EUMETSAT through Copernicus Online Data services (<http://archive.eumetsat.int/usc/>). The product contains “enhanced measurement” data files which include radar waveforms corrected for Automatic Gain Control (AGC), Sea Level Anomaly (SLA), SWH, σ_0 and other auxiliary variables at 20 Hz. Wind speed at 10 meters height (U_{10}) is provided in this product at 1 Hz.

2.2.2.2. Sentinel-3 Fully Focused SAR (FFSAR)

The fully-focused SAR technique is a novel altimetry approach inherited from SAR imagery systems which, unlike SAR mode processing, optimizes the focusing of the radar pulses along the satellite ground track to achieve the highest possible resolution in azimuth. The FFSAR concept has been originally designed for altimetry by Egido and Smith (2017) and first used with Cryosat-2 data demonstrating its improvement over UFSAR data on many water surfaces. The focusing technique consists in a coherent combination of the pulse echoes that are acquired over the illumination time of a scatterer on the surface. This

processing involves a fine alignment of the echoes in phase and range to make them coherent to focus on the point target, then a summation to produce a single-looked waveform at that position with the highest along-track resolution (up to 50 cm corresponding to the theoretical limit of half the antenna length). Successive singled-look echoes are thereby generated along the track, spaced approximately 40 cm apart according to the pulse repetition interval. This unprecedented gain in resolution is of particular interest for probing water surfaces within highly heterogeneous scenes such as inland waters, sea-ice leads and coastal areas. Over rough homogeneous surfaces like the ocean, the speckle noise is particularly present in the backscattered signal, which masks any low-amplitude signals. Consecutive singled-look echoes are thus summed incoherently to reduce the speckle and form a (less-noisy) multi-looked waveform, at the expense, however, of a loss in resolution. The number N_{sl} of single looks that are summed into one multi-looked waveform gives the posting rate P in Hertz, defined as $P = \frac{PRF}{N_{sl}}$. The noise reduction brought by the FFSAR multi-looking is much more effective than for the unfocused SAR methods (Egido and Smith 2017), allowing higher along-track resolution to be used, while ensuring high precision altimeter measurements. The present study seeks to exploit such potential FFSAR improvements to better resolve small-scale sea surface features like ISWs, that conventional altimetry fails to adequately capture. In this respect, FFSAR data were processed at a 160-Hz posting rate (equivalent to a 40 m spacing between measurements on ground) then assessed to determine whether ISW-induced signals can benefit from higher sampling rates.

Despite this improvement, one should note that the lacunar sampling of the Sentinel-3 closed-bursts chronogram introduces replicas in the azimuth FFSAR point target response (PTR), which alters its along-track resolution (Egido and Smith 2017). This makes it difficult to separate real scatterer signals from replicas once they are superposed. However, for long wavelength waves of a few kilometers like ISWs, the impact of the replicas contamination coming from forward or backward waves might not be that important since replicas appear every 90 m on average and rapidly falls off in amplitude for increasing range.

Besides that, it has been shown that the orbital wave velocities are widening the azimuth point target response of the UFSAR and more particularly the FFSAR processing (Reale et al. 2020; Buchhaupt et al. 2021), resulting in the degradation of their native along-track resolution. This further leads to a SWH over-estimation (about 10-15 cm at maximum) if SAR altimeter retracking model does not account for this sea-state parameter.

2.2.2.3. Retracking of SAR Altimeter Waveforms

Geophysical parameters are estimated from Sentinel-3 FFSAR 160-Hz and UFSAR 20-Hz waveforms, generated over the Banda Sea and Tropical West Atlantic areas, using a common retracking algorithm. The retracking consists of fitting at best (in an optimization sense) the altimeter waveforms with a model dependent on geophysical parameters determined here by Levenberg Marquardt optimization. In the SAR mode, the ocean retracking model was analytically formulated by Ray et al. (2014) translating the

conventional altimetry Brown model (see Brown 1977) to SAR altimetry. This model expressed in the time domain is based on the convolution of three terms (probability density function, impulse response function, and the flat sea surface response) which is quite time-consuming. A faster computational approach replacing the time domain convolution by the frequential domain multiplication was introduced by Buchhaupt (2019).

A similar frequential model retracking algorithm as the one developed in Buchhaupt (2019), called *SARmultilookLS3*, was used in this study estimating the classical ocean parameters as the range, SWH, and σ_0 . One fundamental assumption of the associated waveform model is that of considering a constant backscattering over the total area illuminated by the radar. However, this assumption is not always valid, depending on the sea surface roughness that is present within the altimeter footprint. One way to improve the model is to introduce a mean square slope (mss) parameter, to account for the dependency of σ_0 with respect to the viewing angle, using the geometrical optics theory as it is conducted in some published works (see Dinardo et al. 2021; Tourain et al. 2021). Such an adaptive retracking algorithm that additionally estimates a constant mss was developed and applied in this study. It is called *SARmultilookLS3_MSS*, abbreviated as MSS throughout the rest of the paper.

Another possible source of error in the waveform modeling is the σ_0 heterogeneity that is characteristic of ISWs (where high and low values are observed on different sides of the radar footprint of the same ISW front). This is particularly the case when the wave does not propagate in the same direction as the satellite track, resulting in variations of the mss within the radar footprint (that is not accounted for in the model) and potentially leading to biases on the geophysical parameters. To mitigate these errors, a proposed solution is to reduce the range window used in the retracking, i.e., estimate the parameters only on a portion of the waveform around the leading edge. The altimeter footprint size is in turn reduced in the cross-track direction, making the constant mss assumption more likely to be valid, at the expense, however, of an increase in the measurement noise. Based on that, two different approaches were considered: one using the full range of the waveforms as it is classically performed in operational SAR-mode processing and the other one by retracking the waveforms over a reduced range of bins (more precisely, a truncation is carried out dynamically ten gates away from the estimated epoch position), called *10 bins processing*.

2.2.2.4. Characterization of ISW's SWH Signatures

SWH along-track profiles are examined in this paper at two different scales. First, an algorithm (described in Santos-Ferreira et al. 2019) is used to locate the approximate center of the ISWs, which is depicted at the location of the largest gradients of the SWH signal, usually at the transition between a local maximum and the immediately following local minimum. The algorithm first detects the ISW in σ_0 , and in this work we perform the identification of the ISW center manually, i.e., based on our best interpretation of the SWH signal. Secondly, we consider the ISW as the feature $\pm 0.02^\circ$ of latitude to each side of the ISW center. This represents approximately ± 2 km in the along-track direction to each side of the

ISW center, hence corresponding to an approximate wavelength of 4 km (note that in this paper we select ISW fronts perpendicular to the satellite tracks). This wavelength is consistent with the altimeter profiles we show and analyze in [section 2.3.](#), being also the average wavelength reported in Magalhaes et al. (2016), who used SAR images to study ISWs off the Amazon shelf. An illustration of this “small-scale” region is shown in **Figure 2.4b,c** in red shaded rectangles.

The following step consists of defining a larger-scale region of influence of the ISW, so that we can examine larger variations in SWH, which in this paper are hypothesized to be associated with surface wave breaking owing to the ISWs (see [sections 2.3.](#) and [2.4.](#)). This “larger-scale” is defined as $\pm 0.10^\circ$ of latitude to the front and to the rear of the “small-scale” ISW, described above. It is illustrated in shaded blue colors in **Figure 2.4b–e**. A box averaging filter is applied along the SWH signal to produce a smooth profile (see green lines in **Figure 2.4**). This box width of the averaging filter is defined as 10 cells along-track, which is typically the ISW wavelength. The filter width is chosen in a way such that, while still capturing the full backscattering power variations across an ISW front (i.e., crest to trough), it is sensitive to large-scale SWH variations of the order of ± 10 km. As will be shown in [section 2.3.](#), with this filter the SWH variation is still observed at shorter scales (of 1–4 km), while becoming a less noisy and hence clearer signal at larger scales (± 10 km). We call this the “larger-scale” signature of SWH, which is used to characterize the SWH decrease to the rear of the ISWs. In practice, we use a single average value to characterize the SWH in front of the ISWs (corresponding to approximately 10 km in front of the ISW), and another single average value (corresponding to the same length) to characterize the average SWH to the rear of the ISW. The results are shown in the form of bar graphics in [section 2.3.](#)

2.2.3. Ocean and Land Color Instrument (OLCI)

In this paper we use Level-1b OLCI optical products from Top-Of-Atmosphere (TOA) radiometric measurements provided by the ESA-Copernicus Open Access Hub, which were corrected, calibrated, and spectrally characterized. These products are quality controlled and ortho-geo-located (with latitude and longitude coordinates) with a resolution of approximately 300 m at nadir. Quasi-true color images covering 1270 km width on the ground in the across-track direction are analyzed to interpret oceanic ISWs near the air-sea interface. The Level-1b data are appropriate to scrutinize oceanic internal waves visible in the sun glint patterns (direct specular reflection of the Sun in the ocean). Note that the OLCI strip is not centered in relation to the spacecraft’s nadir, but with a westward inclination. As the radar altimeter is aimed at nadir, the Sentinel-3 track is displaced about three quarters of the image to the right (assuming the North pointing vertically upwards). It appears that, in many cases, the sun glint band is close to the satellite track, and that is why it is easier to visualize ISWs (Jackson 2007). We further note the exact temporal correspondence between OLCI and SRAL.

2.2.4. Characterization of Fraction of Wave Breaking in Sentinel-2 MultiSpectral Instrument (MSI)

The MultiSpectral Instrument (MSI) onboard Sentinel-2 measures the Ocean's reflected radiance in 13 spectral bands from Visible and Near Infrared wavelengths (VNIR) to Short Wave Infrared (SWIR) (see e.g., <https://sentinels.copernicus.eu/web/sentinel/technicalguides/sentinel-2-msi/msi-instrument>, accessed on 10 May 2021). In this work, Band 8 that operates at centered wavelength 832.8 nm is used to retrieve an estimate of the wave breaking fraction at the sea surface. This band, as well as Bands 2, 3, and 4 that operate in visible wavelengths have spatial resolutions of 10 m, which is adequate to detect wind waves with scales of the order of tens of meters as well as to detect whitecaps and foams with subpixel resolution owing to their large reflectance in relation to the reflectance of sea water unaffected by breaking waves. We use a similar approach to Kubryakov et al. (2021), who studied a similar problem with the Landsat-8 satellite. It is important to note that while Kubryakov et al. (2021) developed their method for measuring the fraction of wave breaking in relation to wind conditions, here we are only interested in observing the contrast of the fraction of breaking in relation to background conditions unaffected by ISWs.

The method of whitecap fraction retrieval includes the main steps described in Kubryakov et al. (2021), which in summary consists of: considering only azimuthal angles from 30° to 70° to avoid sun glint (Step 1); excluding cloud affected pixels using a suitable threshold in the visible bands (Bands 2, 3, and 4) based on visual inspection (Step 2); calculating the whitecap affected pixels using a similar method to Kubryakov et al. (2021) based on reflectance (Step 3); identifying “outlier” pixels, considered as whitecap affected in comparison to the background by high-passed reflectance in Band 8, by applying a Wiener-2d adaptive noise-removal filter (Lim 1990) based on the computation of the standard deviation against the local mean (Step 4 in Kubryakov et al. (2021) here adapted with a window size of 60 x 60 pixels). With this step, we exclude large-scale variations in the background owing to small sun glint modulations in the internal wave field. Essentially, positive extreme radiance values correspond to whitecaps. For more details, please refer to Kubryakov et al. 2021. The final step (Step 5 in Kubryakov et al. 2021) consists of choosing a suitable threshold to account for “whitecap-affected” pixels or active breakers. Again, we follow these authors closely, who explain that the effective reflectance of the foam in the NIR band (Band 8) is approximately 0.22, and the reflectance of active breakers is 0.55 (see also Koepke 1984; Frouin et al. 1996). After the Wiener filtering in Step 3 (see above) and choosing a threshold based on manual inspection of the image in question (see section 2.3.6), which is 0.002, we retain those pixels with a foam ratio greater than $0.002/0.22$, i.e. 1%, or active breakers (whitecaps) with a ratio of more than 0.3%. This means that the chosen threshold should only retain those pixels with a foam ratio greater than 1% or active breakers' ratio greater than 0.3%. Please note that this corresponds to $0.01 \times 100 \text{ m}^2 = 1 \text{ m}^2$ of a foam-covered area in a pixel, or approximately 0.3 m^2 covered by active breakers. As stated in Kubryakov et al. (2021), this method does not allow differentiation between foams and active breakers, but since the former likely originate from the latter, it is not important in this work. In this

paper, we are only aiming to demonstrate that there exists wave breaking over the ISW crests in the region of study, namely the Banda Sea (see [section 2.3.6](#)).

2.3. Results

2.3.1. Sentinel-3 Case Study Using Optical Imagery and Standard Copernicus SAR Altimeter Data

Figure 2.4a shows a red-green-blue (RGB) image composite from an OLCI acquisition (i.e., a quasi-true color image from Sentinel-3B) where large-scale ISWs with characteristic width scales of about 10 km are seen propagating offshore the South American continent in an almost meridional direction. In this case, the along-track direction of the SRAL ground-track coincides closely with the waves' propagation direction—which we recall is simultaneously collecting data in synergy with the OLCI optical sensor. This means that the increased ground resolution resulting from the SAR processing in the along-track direction is conveniently aligned to measure the variations in the ISWs' sea surface roughness sections (very similar to the schematics presented in **Figure 2.3b**). Note that the increased along-track resolution of about 300 m (for the UFSAR mode) will fit within the ISWs' rough and smooth sections, and hence we expect to see a transition within the sea surface roughness signatures that are of the order of a few km. In other words, a significant number of individual waveforms will be acquired over the waves' crest and trough sections. Furthermore, this acquisition geometry allows for possible contributions from enhanced surface wave breaking (expected at the leading sections of these large-scale ISWs) in the altimetry record to be further analyzed.

Two transects are shown in **Figure 2.4b–e** for variations of NRCS (σ_0) and SWH (top and bottom panels, respectively), all related to the same along-track sections of the two ISWs packets highlighted in **Figure 2.4a** (see red labels at 5.22°N and 3.95°N). The results are consistent with the previous analysis in Santos-Ferreira et al. (2018, 2019) for large-scale ISWs and their SAR altimetry sea surface signatures. In the present cases, the σ_0 variations corresponding to the rough and smooth sections of the sea surface affected by ISW manifestations vary by approximately 2 dB and 3 dB, respectively (top panels in **Figure 2.4b,d**, respectively), when comparing with an unperturbed section immediately preceding the ISW (i.e., ahead of the wave and hence unaffected by it). This asymmetry between the ISW rough and smooth sections may be related to small-scale roughness (of the order of three times the ku-band radar wavelengths), but may also contain an additional radar backscatter contribution from enhanced surface wave breaking which frequently features in the ISWs leading sections (as illustrated in **Figures 2.1 and 2.2**).

Similar examples were presented in Santos-Ferreira et al. (2018) and Magalhaes et al. (2021) which already showed that SWHs from SAR altimetry have unambiguous surface signatures in the presence of ISWs. We therefore proceed in a similar manner but investigate it further for surface wave breaking. According to our results in **Figure 2.4c,e**, some sharp and localized increases of the order of 1 m can be seen over the rough (leading) sections of the ISWs—which is precisely where whitecaps are expected to occur (see also **Figures 2.1**

and 2.2), meaning the local maxima for surface wave heights has probably been reached (i.e., the waves there cannot grow any larger and eventually break). A detailed analysis reveals another larger-scale effect in SWH that is the focus of this paper, related to both the ISW and its immediate surroundings (of the order of ± 10 km to the front and to the rear of its center). It is highlighted in **Figure 2.4c,e** (bottom panels for different ISWs' latitudes) by a green rectangle which shows that SWH decreases on average by about 53 cm and 35 cm, respectively, when comparing the sections ahead and behind the ISW, whereas the transition appears clearly related to the sharp decrease in σ_0 (labeled "Local decrease" in **Figure 2.4b,d**). This is related to the expected enhanced surface wave breaking. We suggest that these are two separate features observed in SWH, i.e., the "Local increase" in σ_0 and the "Larger-scale decrease", which are in fact related via the internal wave—surface wave interaction.

It is therefore important to evaluate whether these characteristic signals in SWH visible in **Figure 2.4c,e** are isolated events, or if they are instead typical features of large ISWs that include enhanced surface wave breaking in their sea surface manifestations. To investigate this, we again rely on the synergetic acquisitions in Sentinel-3 between the optical imagery (OLCI) and its SAR altimeter (SRAL). In particular, OLCI imagery was used to survey other large-scale ISWs in this study region as well as in another region known for large-scale ISWs (the Banda Sea of Indonesia). We note in passing, however, that regions such as the South China Sea where very large ISWs are known to be associated with breaking waves (see e.g., Zhang et al. 2020 and Figure 8 in Liu et al. 2008) cannot (a priori) be used in this analysis since we need a proper geometry (as that in **Figure 2.3b**) between the ISWs and the satellite ground track—whereas in the South China Sea, the ISWs' propagation and the altimeter track are nearly perpendicular, rendering the enhanced ground resolution of the SRAL essentially useless. The ISWs propagating in a nearly meridional direction in the Banda Sea (near Indonesia) fulfil the optimum geometry for SRAL observation of ISWs, and in this paper, we focus our search there (and in our current study region) for additional cases showing admittedly evidence of enhanced surface breaking in the SRAL altimeter.

2.3.2. Results for Standard Copernicus SAR Data

We now present in **Table 2.1** results of the SWH average cross-sections of ISWs observed (simultaneously) in OLCI acquisitions and in σ_0 . We used 38 examples in total, from 23 different dates (i.e., sometimes more than one ISW was encountered on the same day), being 10 in the Banda Sea and 13 in the Tropical West Atlantic. To calculate the SWH average cross-sections of ISWs, we used the average value within ± 10 km before and after the ISWs' center regions (depicted in light red shades in **Figure 2.4b–e**, which are about 4 km in length), which we named SWH rear and SWH front, respectively. Please see [section 2.2.2.4](#) for more details.

Table 2.1 - List of Sentinel-3 cases exhibiting large-scale ISWs in optical imagery (OLCI) in synergy with the corresponding sea surface signatures in the SRAL altimeter. For each case, a respective measure of Significant Wave Height (SWH) is given for leading (front) and trailing (rear) sections of the ISWs (as in Figure 2.4). Differences between these characteristic values ($\Delta\text{SWH} = \text{SWH}_{\text{front}} - \text{SWH}_{\text{rear}}$) are also shown. For each acquisition, representative estimates are also shown for the surface wave field (including direction and wave period—from <https://earth.nullschool.net/>, accessed on 10 May 2021). All cases can be accessed in their corresponding permalinks from OceanDataLab (<https://www.oceandatalab.com/home>, accessed on 10 May 2021).

I N D I A N O C E A N - B A N D A S E A	DATE	REL. ORBIT	CYCLE	PASS	ISW (°E; °N)	SWH FRONT (m)	SWH REAR (m)	DELTA SWH (m)	PEAK WAVE DIRECTION		PERMALINK
									°TN	Period (s)	
T R O P I C A L W E S T A T I A N C	27.10.2018	188A	37	376	125,51; -6,12	0,40518	0,23571	0,16947	120	4,2	https://odl.bzh/4Ef5VZFC
					125,27; -7,23	0,46247	0,16659	0,29588	105	4,2	https://odl.bzh/KrMExcZ1
	25.03.2019	245B	23	490	125,93; -6,34	0,50846	0,45982	0,04864	340	4,6	https://odl.bzh/WzP5Ezm1
	21.04.2019	245B	24	490	125,97; -6,15	1,2725	1,0296	0,26201	115	5,1	https://odl.bzh/b8LxyW55
					125,72; -7,28	0,62323	0,31226	0,31097	90	5,1	https://odl.bzh/b8LxyW55
	10.06.2019	188B	26	376	124,85; -6,53	0,92049	0,42788	0,49261	100	6,3	https://odl.bzh/LF-Vxh__
	20.08.2019	188A	48	376	125,68; -5,32	1,2486	1,2242	0,0244	120	6	https://odl.bzh/HSaOiMnJ
					125,49; -6,24	1,0881	0,78976	0,29834	110	6,1	https://odl.bzh/HSaOiMnJ
	03.09.2019	245B	29	490	125,93; -6,37	1,1701	1,0148	0,1553	110	5,1	https://odl.bzh/JQMuqlcm
	16.09.2019	188A	49	376	125,45; -6,28	1,6424	1,2135	0,4289	100	5,4	https://odl.bzh/9_5wRYcZ
					125,23; -7,34	1,0898	0,7775	0,3123	80	5,5	https://odl.bzh/9_5wRYcZ
	30.09.2019	245B	30	490	125,95; -6,27	1,0215	0,53449	0,48701	115	5,1	https://odl.bzh/MtaLtlvy
					125,70; -7,33	0,67391	0,59438	0,07953	85	5	https://odl.bzh/MtaLtlvy
	02.05.2021	188A	71	376	125,47; -6,21	1,5645	1,25	0,3145	120	4,8	https://odl.bzh/bmwz6-sl
					125,22; -7,38	0,67488	0,59038	0,0845	90	4,5	https://odl.bzh/bmwz6-sl
S E A	29.05.2021	188A	72	376	125,54; -6,00	1,8899	1,456	0,4339			https://odl.bzh/UHooCZrC
					125,28; -7,17	0,87386	0,85046	0,0234			https://odl.bzh/UHooCZrC
	01.09.2016	152A	8	304	43,01; 5,55	1,2279	0,92231	0,30559	120	8,4	https://sentinelshare.page.link/uvbx
	07.10.2017	095A	23	190	43,96; 5,52	1,9382	1,7371	0,2011	130	8,6	https://odl.bzh/VdEggv5S
	**11.10.2017	152A	23	304	43,30; 4,28	2,3694	1,8705	0,4989	110	9,6	https://odl.bzh/Lpqi_JIT
	27.09.2018	152A	36	304	43,19; 4,75	2,0583	1,9157	0,1426	85	10,9	https://odl.bzh/Na_TJmBh
	24.10.2018	152A	37	304	42,99; 5,71	1,815	1,5135	0,3015	340	11,2	https://odl.bzh/JOAzPnPS
	23.08.2019	095B	29	190	44,56; 4,88	1,1651	0,87815	0,28695	110	9,1	https://odl.bzh/6NeM1WHF
					44,70; 4,3	1,0569	0,97263	0,08427	110	9,1	https://odl.bzh/6NeM1WHF
	20.11.2019	209B	32	418	42,24; 6,94	2,4213	2,1723	0,249	45	9,2	https://odl.bzh/KfnQ6nXW
					42,50; 5,78	2,0414	1,9034	0,138	45	9,2	https://odl.bzh/KfnQ6nXW
	06.07.2020	152A	60	304	43,02; 5,49	2,0593	1,3832	0,6761	100	8	https://odl.bzh/Cj11Y7Fp
					43,31; 4,25	1,5908	1,5369	0,0539	100	8	https://odl.bzh/Cj11Y7Fp
	12.08.2020	152B	42	304	43,47; 5,56	1,9343	1,6387	0,2956	125	9,3	https://odl.bzh/w_Bt-Fwr
	21.08.2020	038A	62	76	45,18; 4,2	1,7181	1,16	0,5581	105	9,2	https://odl.bzh/ifCVN1l5
					45,46; 2,98	1,3141	0,92823	0,38587	105	9,1	https://odl.bzh/ifCVN1l5
	04.09.2020	095B	43	188	44,49; 5,22	1,9822	1,6322	0,35	115	9,1	https://odl.bzh/zJX2S9l3
					44,77; 3,95	1,7456	1,2196	0,526	110	8,8	https://odl.bzh/I10mN45U
					45,05; 2,74	1,5703	1,4927	0,0776	105	8,7	https://odl.bzh/I10mN45U
	18.10.2020	095A	64	190	44,05; 5,11	1,1499	0,98153	0,16837	100	8,1	https://odl.bzh/U5tfVGM-
					44,30; 3,98	1,5645	1,2352	0,3293	105	7,9	https://odl.bzh/bSfaqEFH
					44,62; 2,6	1,3232	1,1173	0,2059	95	7,6	https://odl.bzh/bSfaqEFH
	03.12.2020	366A	65	732	45,37; 7,55	1,5067	1,4067	0,1	25	14,6	https://odl.bzh/rLepYH4m

According to the results in **Table 2.1**, which are more clearly appreciated in **Figure 2.5**, the SWH estimates retrieved from the SAR altimeter show that a decrease is consistently seen between the waves' leading and trailing sections, which, on average, are of about 25% of the unperturbed SWHs (i.e., ahead of the ISW) in the Banda Sea and 17% in the Tropical West Atlantic (see also **Table 2.2**). We also note in passing that these changes in SWH hold for different wind conditions, ranging from weak to moderate (even strong) winds and with different orientations, which adds to the robustness of these results.

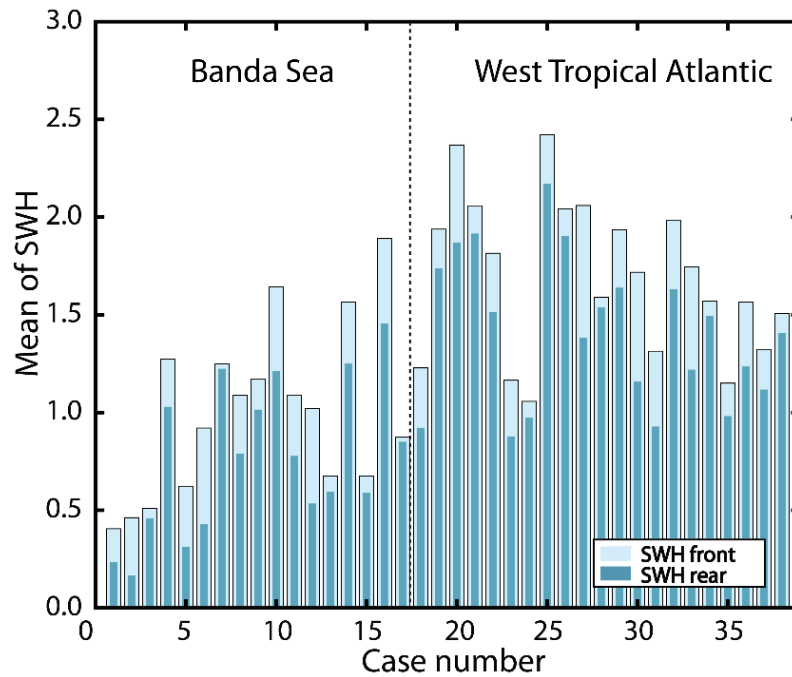


Figure 2.5 - Bar plots representing average significant wave heights (SWHs) for leading (i.e., front) and trailing (i.e., rear) sections (see text for details) of the ISWs for all cases in Table 2.1.

Table 2.2 - Mean of SWHs in front and rear of ISWs (see text for details) and differences between both (in percentage) for cases in the Banda Sea and Tropical West Atlantic. In the bottom row, the same method is used for all Copernicus UFSAR cases (which are not exactly the same as those considered in two rows before, see text for more details).

Products	Banda Sea			Tropical West Atlantic		
	<SWH> (m)	<SWH> (m)	< Δ SWH> (%)	<SWH> (m)	<SWH> (m)	< Δ SWH> (%)
FF 10 Bins	1,06	0,88	16	1,67	1,49	12
FF mss 10 Bins	1,15	1,01	12	1,79	1,61	10
UF 10 bins	1,04	0,90	13	1,75	1,46	17
UF mss 10 bins	1,10	1,00	10	1,80	1,60	11
UF copernicus	1,01	0,76	25	1,69	1,41	17

We therefore feel confident to propose as a possible interpretation for the observed decrease in SWH to the rear sections of an ISW as being the result of ISWs in the local surface wave field, particularly the ISW-induced enhanced surface breaking. Since SWH is observed in the SAR altimeter to decrease as we move from unperturbed sections ahead of the ISW to sections behind the ISW (i.e., after the wave has passed), it is possible that a significant part of the surface wave spectrum is damped as a consequence of the passage of an ISW (in

particular for the highest third of surface waves). This could be the result of additional surface wave breaking, triggered by the passing ISW, since this would cause the highest waves to increase beyond their limiting amplitude, break, and dissipate wave energy. In this case, the highest waves in the wave spectrum would undergo significant damping and cause a decrease in SWH in the transition between the ISWs' rough and smooth sections—which we again recall is where whitecaps are consistently observed in optical imagery.

2.3.3. A Case Study Using Optical Imagery and Advanced SAR Altimeter Products

Figures 2.6 and 2.7 concern the same data scene as in Figure 2.4 but regarding advanced SAR products, which are FFSAR and UFSAR with different retracker, respectively (the retracker schemes are briefly described in section 2.2.2.3). To calculate the SWH average cross sections of ISWs, we used the average value within 10 km before and after the ISW center region (as conducted in Figure 2.4, which are about 4 km in length, but using the advanced SAR altimeter products described in section 2.2.2.3). We coin the names “SWH-rear” and “SWH-front”, respectively.

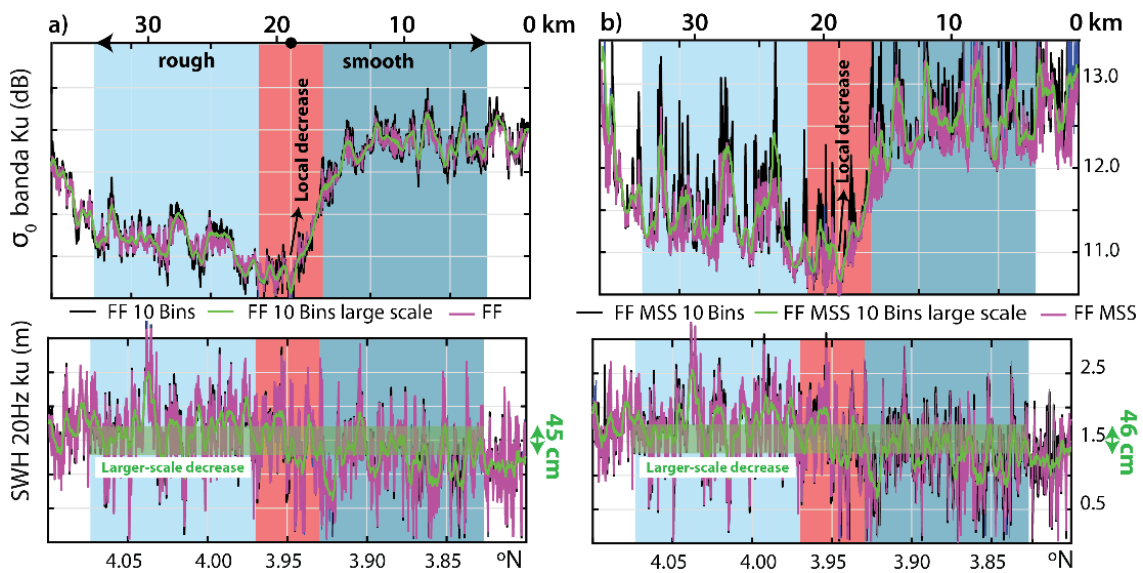


Figure 2.6 - The rectangles depicted in red shades represent the ISW center at 3.95°N. From the center of ISW, it is considered ± 2 km to each side. To the left of it, we define the “SWH front” ahead of the ISW (“rough” in lighter blue using scales of about 10 km) and to the right, the “SWH rear” (“smooth” in darker blue using 10 km). Top panels show along-track NRCs (i.e., σ_0) measured by the SRAL altimeter (in synergy with the OLCI) with the SAR-mode retracking color-coded in (a) FFSAR, FFSAR 10 Bin and its large-scale mean and in (b) FFSAR MSS, FFSAR MSS 10 Bins and its large-scale mean (using a box running average with approximately 3 km). Bottom panels show SWHs' large-scale variations of 0.45 m in (a) and 0.46 m in (b), identified in the green shaded rectangles, when comparing the section in front of the primary ISW with that in the rear.

The list used to evaluate the percentage difference between the front and rear of ISWs in these new algorithms is in Supplementary Materials S1 (see **Table S1**), being similar to **Table 2.1**. General remarks can be made regarding those advanced SAR processing results:

- Consistent results for all products: from Copernicus products, FF and UF full bins to FF and UF 10 bins processing schemes.
- SWHs from the Tropical Atlantic Ocean are higher than for the Banda Sea, which is understandable since in the Banda Sea, swell waves are practically absent; note that the average wave periods are 9.2 and 5.1 s, respectively.
- As expected, the adaptive MSS algorithms allow for better capturing of σ_0 variations along the track, the 10 bin processing being a little more sensitive to heterogeneous sea surfaces.
- It is suggested however that the 10 bin processing has little impact on the SWH results for these applications and study regions.
- It is also suggested that the FFSAR processing (both schemes applied by CLS) has little impact on SWH differences in this particular application (Internal Solitary Waves). It is as if by lowering the radar footprint, the processing is much less impacted by heterogeneities on the sea surface, making the MSS and 10 bin processing almost useless.

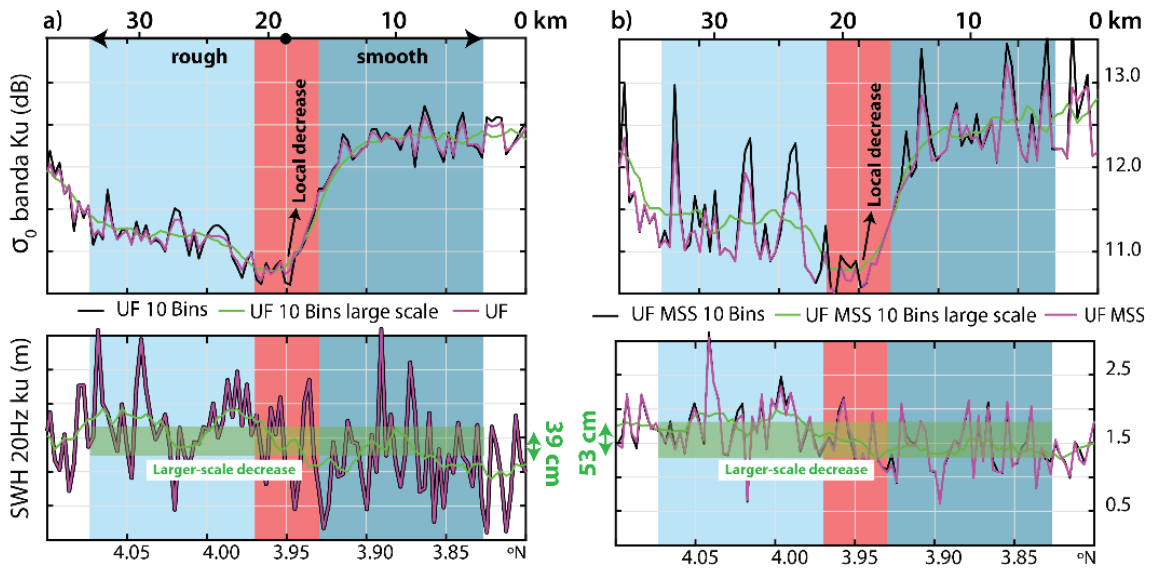


Figure 2.7 - As in Figure 2.6, but top panels of (a,b) shows along-track NRCs (i.e., σ_0) measured by the SRAL altimeter (in synergy with the OLCI) with the SAR-mode retracking color-coded in (a) UFSAR, UFSAR 10 Bins and its large-scale mean and in (b) UFSAR MSS, UFSAR MSS 10 Bins and its large-scale mean (using a box running average of 3 km). Bottom panels show SWHs' large-scale variations of 0.39 m in (a) and 0.53 m in (b), identified in the green shaded rectangles, when comparing the section in front of the primary ISW with that in the rear for the same products.

2.3.4. Results for Advanced FFSAR Products

More statistical analyses of the SWH variations between the front and rear of ISWs are presented in **Figure 2.8**, which are measured by the different unfocused and focused SAR processing configurations. Globally, the same results as those obtained in the case study of [section 2.3.3](#) are observed. In particular, it is confirmed that the adaptive MSS processing scheme tends to reduce the SWH attenuation in UFSAR, whereas it does not seem to impact the FF-SAR measurements, suggesting that the narrow footprint of the FFSAR makes the model less sensitive to the roughness inhomogeneities of the surface.

2.3.5. Consistency of Results for Different SAR Products

Table 2.2 demonstrates the consistency of results for all SAR products processed in this paper for SWHs. While the average attenuation in SWH is in the range 10–13% for the UFSAR 10 bin processing scheme that compares with 25% for the standard Copernicus products in the Banda Sea, for the West Tropical Atlantic off the Amazon river mouth, the range is between 11 and 17% in comparison with the 17% of the Copernicus products. Furthermore, in the FFSAR mode (and recalling that the processing is all using the 10 bin method), one can see that for the Banda Sea, the range is 12–16% in comparison with the 25% of the Copernicus standard processing, while for the Western Tropical Atlantic, the range in the FFSAR mode is 10–12% in comparison with the 17% in the standard UFSAR Copernicus processing scheme. We note that the adaptive MSS processing scheme (FF MSS 10 Bins and UF MSS 10 Bins in **Table 2.2**) gives systematically slightly smaller SWH attenuations than the *SARmultilookLS3* ocean retracking algorithm (FF 10 Bins and UF 10 Bins in **Table 2.2**). All this said, the standing feature from **Table 2.2** is the consistency of the average SWH attenuation for all products studied in this paper.

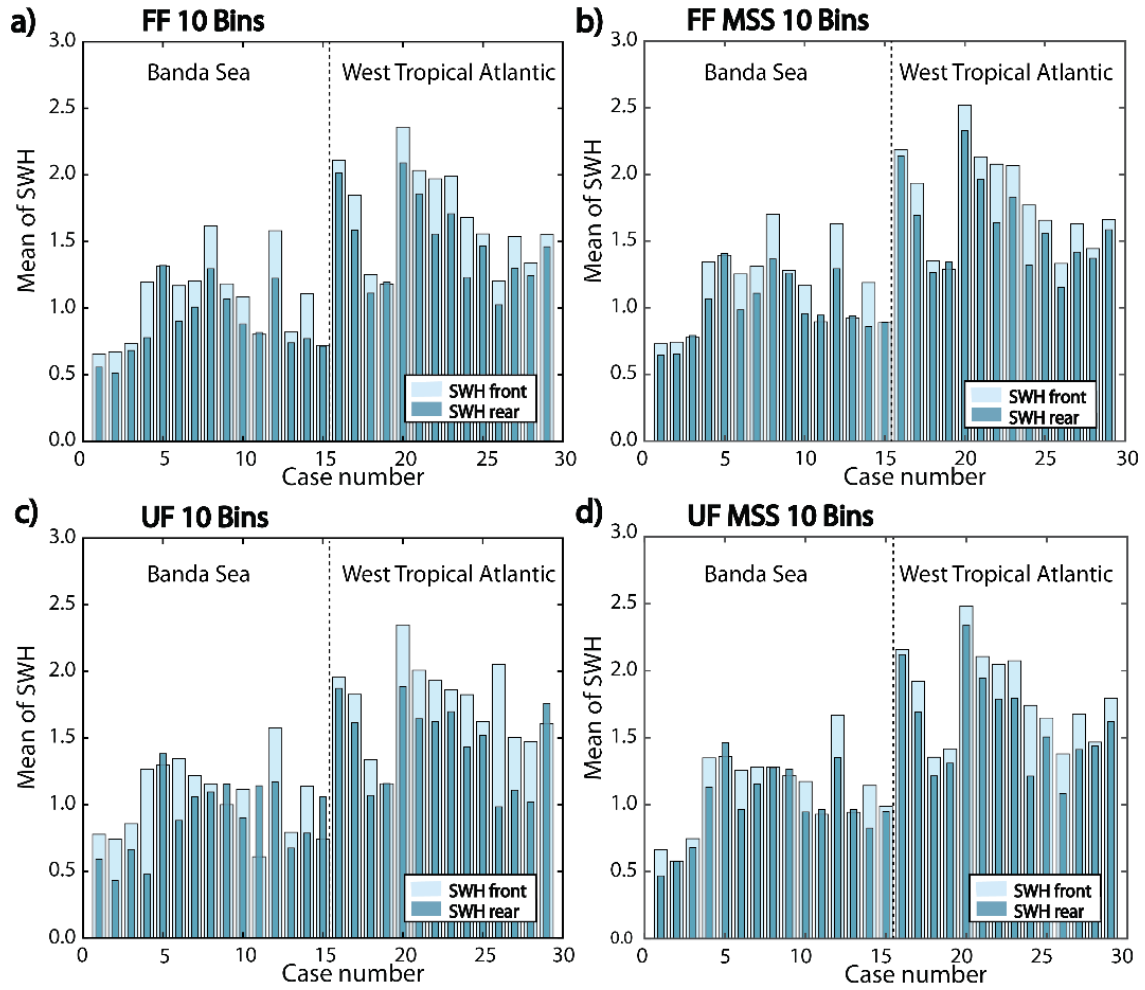


Figure 2.8 - Bar plots representing significant wave heights (SWHs) for leading (front) and trailing (rear) sections of the ISWs for all cases in Table S1 using different algorithms: (a) FF 10 Bins. (b) FF MSS 10 Bins. (c) UF 10 Bins. (d) UF MSS 10 Bins.

2.3.6. Image Processing for the investigation of wave breaking in Sentinel-2 images

In this section, we calculate the percentage of whitecaps (or a fraction of wave breaking) in a specific MSI image of the Sentinel-2 satellite. The image was selected because there is synergy with the Sentinel-3B image shown in **Figure 2.9a**, which is from the Indian Ocean, in the Banda Sea, acquired on 21 April 2019. ISWs are clearly seen at -7.21°N and at -7.28°N in the Sentinel-2 and Sentinel 3B images, respectively, which were acquired with approximately a 3 h difference (see **Figure 2.9a, b**).

We follow the methods published in Kubryakov et al. (2021), which were applied for Landsat satellite data but can be adjusted to the MSI (see [section 2.2.4](#)). Here we select an image section (corresponding to the black rectangle in **Figure 2.9b**), in the near-infrared NIR band (Band 8), that corresponds to the center wavelength of 842 nm in the Sentinel-2 MSI and apply the methods described in [section 2.2.4](#).

Firstly, we verify the azimuthal angles that were found to be between 30° and 70° , in this specific case at around 35° . Band 10 is used to detect and discard high cirrus clouds, meaning that cirrus clouds are filtered with a threshold of 0.003. In this case, all pixels satisfy this

criterion, and the image section is hence unaffected by high cirrus clouds. Then, we apply a wiener filter (see **Figure 2.10**), with a window size of 60 pixels (see [section 2.2.4](#) and Kubryakov et al. 2021). This is a low-pass filter that is usually applied to an intensity image that has been degraded by constant power additive noise. The wave breaking fraction (W) is the difference between original Band 8 (NIR) and this filtered image (see **Figure 2.10**).

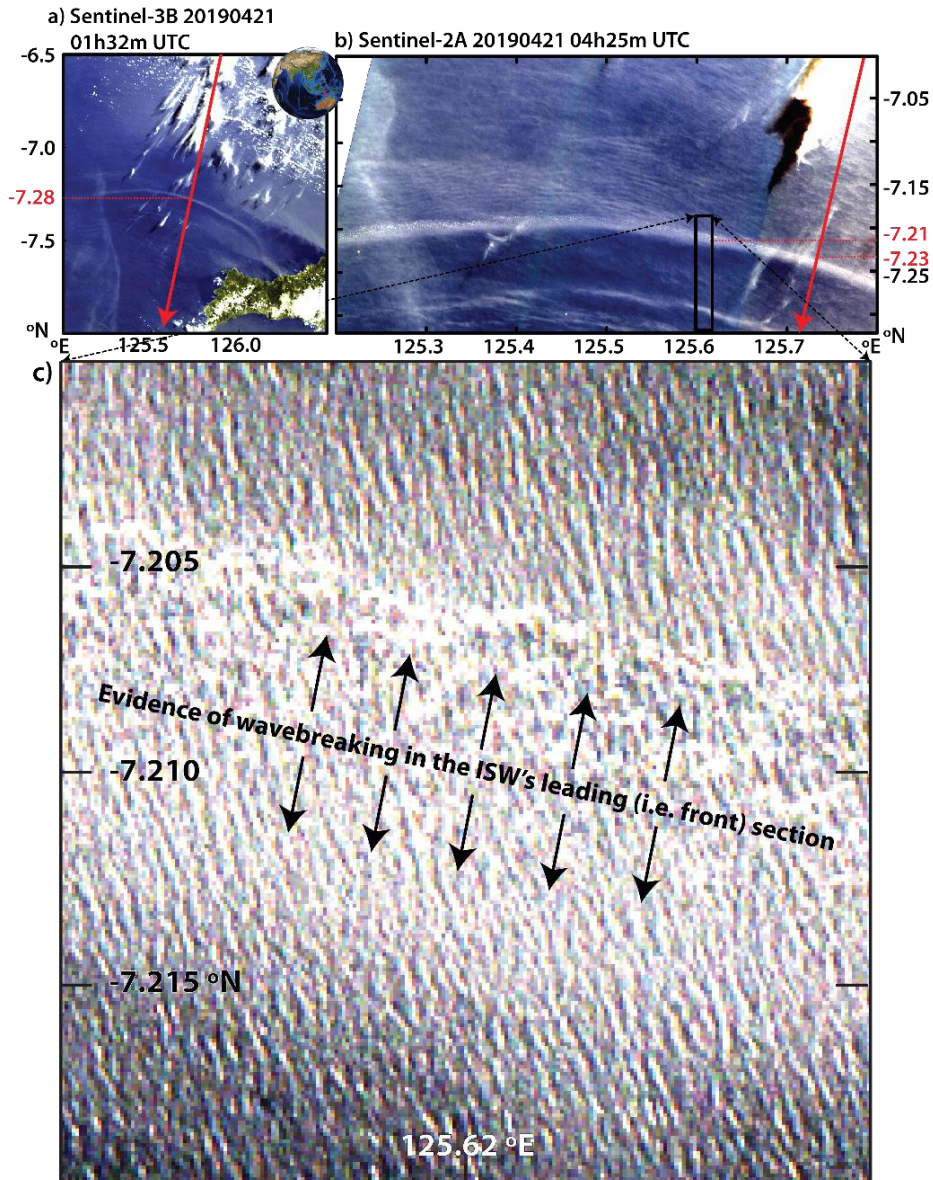


Figure 2.9 - (a) Quasi-true color image (Level-1b) acquired on 21 April 2019 at 01 h 31 m UTC by the OLCI sensor onboard the Sentinel-3B satellite. The red line represents the ground-track of the satellite, which in this case is practically parallel to the ISWs' direction of propagation. (b) Image taken in the visible band by the Multispectral Instrument (MSI) onboard the Sentinel-2A satellite on 21 April 2019 at 04:25 UTC showing sea surface signatures of the same ISW packet seen in (a) approximately 3 h later. The red line also represents the ground-track of the satellite Sentinel-3 in the same day but at a different time. (c) Zoomed section of the leading ISW seen in (b). Note the bright spots aligned in bands, which are evidence of whitecaps generated by breaking surface waves.

As explained in Kubryakov et al. (2021) and in [section 2.2.4](#), a suitable threshold to obtain the fraction of breaking is 0.002. The result is presented in **Figure 2.11**. Pixels where no breaking is detected are colored in white in **Figure 2.11a**, whereas the colored scale applies to the actual number of fractions of breaking. To comprehend the amount of breaking that is caused by the presence of the ISWs, we plot in **Figure 2.11b** an average profile along the ISW propagation direction. It can be seen that the fraction of wave breaking varies along the ISW direction, from 2.5×10^{-3} to more than 7.5×10^{-3} . An average estimate of the fraction of breaking in a region unaffected by the ISW (ahead of the ISW packet) is 3.0×10^{-3} (not shown). This means that wave breaking is enhanced by at least two-fold over the sections of the ISW where surface wave breaking occurs. These results are important in the context of this work because they show by an independent method that wave breaking occurs for those ISWs typically analyzed in the altimeter records in the previous sections. We note in passing that the wave breaking fractions found here are in line with similar values shown in Kubryakov et al. (2021), who computed W values of the order of $7\text{--}10 \times 10^{-3}$ for large ISWs near the Strait of Gibraltar (see their Figure 9b).

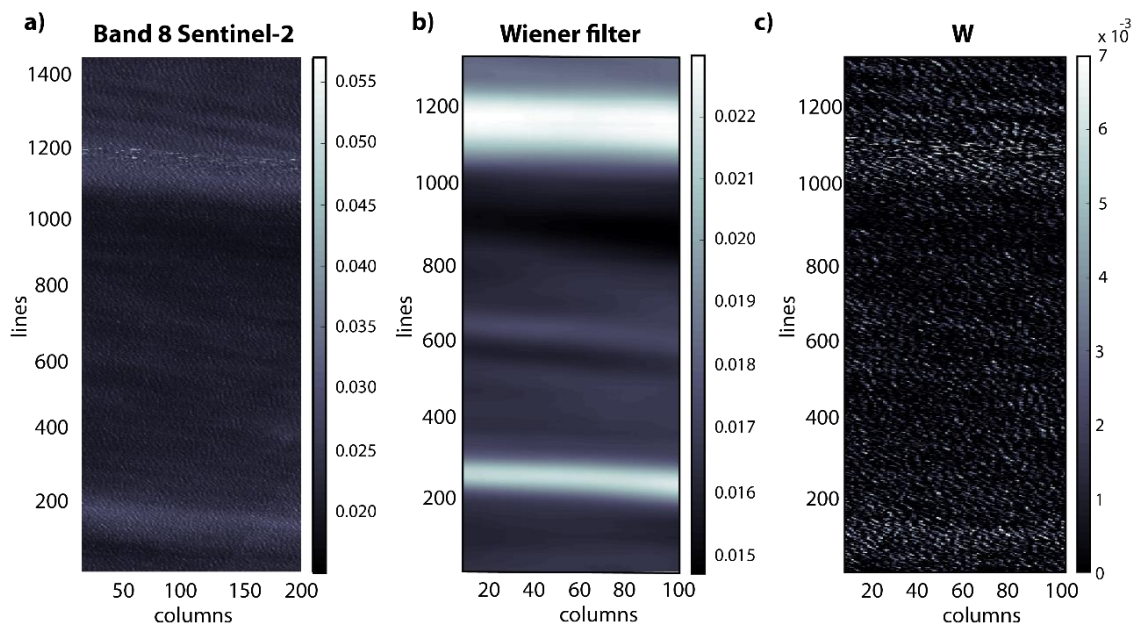


Figure 2.10 - (a) Zoom as shown in the black rectangle in Figure 2.9b corresponding to the NIR Band 8 of the Multispectral Instrument (MSI) onboard the Sentinel-2A satellite on 21 April 2019 at 04h25m UTC showing sea surface signatures of an ISW packet. (b) Wiener filter applied to the image in part (a) – see text for details. (c) The image labeled W represents the wave breaking fraction, calculated by subtracting part (b) from (a).

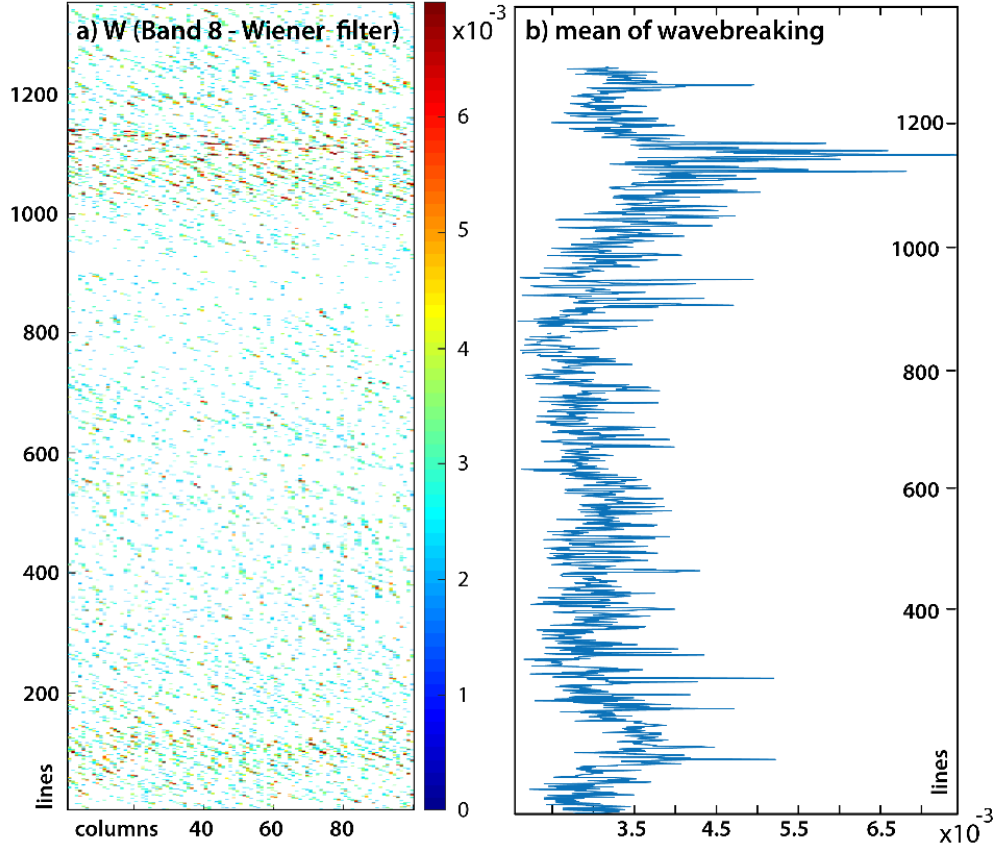


Figure 2.11 - (a) W represents wave breaking as shown in Figure 2.10c but color-coded to highlight the ISWs' different sections, in which we applied a threshold of 0.002 (see text for details). (b) Mean of wave breaking fraction, i.e., W calculated for each line of (a), which is also approximately along the ISWs' crests.

The same authors also estimate that the ISWs in their study enhance the fraction of breaking, i.e., whitecap enhancement, by a factor of 3 for a wind speed of about 10 m/s. In our study case of **Figures 2.10** and **2.11**, the wind speed estimate is about 7 to 8 m/s, and the whitecap enhancement is at least two-fold and up to 7.5×10^{-3} . Therefore, after the analysis explained in this section, we are confident to say that significant wave breaking occurs for typical ISWs in the Banda Sea. The same procedure is more difficult to apply to the Tropical West Atlantic, but we are confident that the wave breaking effect there is strong and similar to the Banda Sea.

We finally note that the case demonstrated in this section, dated 21 April 2019, is of course a synergy case, presented in **Table 2.1** ([section 2.3.2](#)). The average SWH evaluated 10 km in front of the ISW located at 125.72°E , -7.28°N is 0.62 m, while the same average (in scales of 10 km) evaluated behind the ISW is of 0.31 m. In this case, the relative SWH decrease, presumably owing to breaking surface waves, is 50%. We note however that, in such low wave height scenarios, the radar altimeter SWH measurements are probably less accurate and should be taken with caution.

2.4. Summary and Conclusions

This paper focuses on satellite remote sensing of Internal Solitary Waves (ISWs) by the new Synthetic Aperture Radar (SAR) altimeter on board Sentinel-3. We have pointed out that observations show that the sea surface roughness bands generated by ISWs often contain breaking waves having wavelengths of the order of meters and decameters (i.e., intermediate-scale waves, see **Figure 2.1**). Sometimes, such roughness bands containing breaking intermediate-scale waves have been observed when the sea surface was glassy with little or no surface winds. Investigations for oblique-looking satellite SAR imagers already show that non-polarized scattering mechanisms have to be invoked to explain radar backscattering from breaking waves adequately (see e.g., Kudryavtsev et al. 2003, 2005, 2013, 2014, 2019) in the presence of inhomogeneous surface currents such as those produced near the surface by ISWs. These effects were hypothesized as the suppression of small-scale wind ripples by turbulence generated by breaking waves and were also recently confirmed by direct wave tank measurements described in Ermakov et al. 2020.

While σ_0 estimates relate more to small-scale roughness (elements with length scales greater than about three times the incident electromagnetic wavelength) rather than to some measure of the local wave spectrum (and hence with surface wave breaking), a more suitable parameter to detect surface wave breaking could be SWH. As said before, this can be retrieved from satellite altimetry via the waveforms' leading slopes. Note that SWH is, by design, a statistical parameter that conveys a measure of the significant wave height (traditionally the highest third of the surface wave field). Therefore, if the highest waves in the local (i.e., altimeter footprint) surface wave spectrum change—say by resonance with an ISW propagating below—then a measurable variation is expected to appear in the altimeter's SWH estimates. More importantly, we expect a signal in SWH when these surface waves grow beyond their maximum equilibrium point and break, causing wave energy dissipation behind the wave breaking band of the ISW surface manifestation.

Along-track SWH estimates from SAR altimeters revealed a sudden reduction in surface wave height owing to the presence of ISWs in the deep ocean. Recalling that energy in surface waves is proportional to the surface wave height squared, we propose an explanation for the decrease in wave energy after the passage of a large solitary wave: that the evidence of wave dissipation observed in the altimeter is due to breaking waves provoked by an internal wave—a surface wave interaction, presumably owing itself to resonance. The former argument is clearly consistent with the high resolution Multi-Spectral Instrument (MSI) Sentinel-2 shown in **Figure 2.9**, which reveals the existence of a region of enhanced meter-scale waves and whitecaps. We take this feature as evidence of strong wave breaking owing to a resonant interaction. This is a new and important result in oceanography, since ISWs may propagate at basin-scale distances (see e.g., Magalhaes et al. 2016), meaning that they could cause a significant amount of enhanced surface turbulence along several hundreds of kilometers.

Furthermore, despite the altimeter evidence in this study concerning only large-scale ISWs, it has been shown that smaller-scale ISWs' propagation over the shelf also feature

enhanced surface breaking—this was shown e.g., using aerial LIDAR measurements of intermediate-scale ISWs (of the order a few hundreds of meters) in their dissipation stages off the west coast of Portugal (Magalhaes et al. 2013). Being a ubiquitous feature in both deep-ocean and coastal environments, ISWs may play a more efficient role than previously acknowledged in atmosphere-ocean interactions, owing to enhanced surface wave breaking. While energy input into the wave field by wind wave growth mechanisms may be enhanced by ISW resonant effects, it is eventually balanced by sinking terms owing to wave breaking, which generates turbulence and injects air bubbles into the upper water layer (Magalhaes et al. 2013), typically across depth scales of the same order as the surface wave heights (as shown by Wang and Wijesekera 2018). We suggest that the breaking of surface waves owing to large amplitude ISWs may play a key role in the dynamics of the upper ocean, i.e., in the exchange of momentum, heat, and gases between the atmosphere and the ocean.

Energy in the surface wave field is slowly accumulated under the wind action over thousands of wave periods (Babanin 2011), being redistributed by nonlinear wave-wave interactions (Wang and Wijesekera 2018). Interestingly, all that energy can be quickly cascaded down to turbulent scales and dissipation—owing to enhanced surface wave breaking from an ISW—within very short time scales (typically of the order of a single internal wave period, about 40 min). This is an important aspect about ISWs previously undocumented, provided there is a minimum amount of surface wave breaking, which can be caused when resonance occurs between ISWs and surface waves. When the ISW phase speed is large enough (e.g., in deep ocean regions), energy can be transferred between the ISW and surface waves with meter-scales, the latter being able to grow in amplitude and eventually break dissipating significant amounts of energy. Alternatively, or even simultaneously, it could also be possible that the presence of ISWs shifts the surface wave spectrum towards smaller wavelengths. A similar feature, but ascribed to other ocean phenomenon, was described in (Ardhuin et al. 2017), who show a large attenuation in SWHs (measured by Jason-3) in the vicinity of the Gulf Stream off the coast of North America (see their Figure 3b). These authors noted large gradients in SWH on scales of 10 km as the altimeter footprints illuminate waves inside and outside the Gulf Stream by a factor of 2. Because of the current-wave interaction, it is likely that these large gradients in relatively short spatial scales may have to do with wave breaking and surface wave energy dissipation in a similar manner to the phenomena reported in this paper.

Satellite altimetry—including new-generation and more sophisticated missions (such as Sentinel-6 and the upcoming SWOT mission)—could prove a useful tool to understand and investigate these questions better. Note that, unlike imaging SARs, which have a limited number of acquisitions per orbit (owing to power limitations), satellite altimeters operate continuously and could (at least in principle) be used to map ocean internal waves on a routine and global basis and hence help quantify their turbulent dissipation more effectively.

Regarding the altimeter SAR processing, the consistency of the results between the different processing was demonstrated; both UFSAR and FFSAR (in full and reduced bins mode) are impacted by ISWs' signatures in SWH and σ_0 . It does seem, however, more

valuable and relevant to make use of higher sampling altimeter processing as the fully focused SAR may better detect small scale features (which would provide additionally a better gain in resolution as long as the vertical wave motion does not alter it that much). It must also be emphasized that further improvements would be made possible with the Sentinel-6 Open-burst operating mode (i.e., interleaved) which will not only reduce the noise of the altimeter measurements, but also significantly improve the focusing precision of the FFSAR measurements (an important step forward compared to the Cryosat-2 and Sentinel-3 missions which are currently impacted by sidelobes in the along-track point target response caused by the azimuth replicas of the Closed-burst mode).

Ultimately, surface wave breaking owing to ISWs add to the turbulent mixing and dissipation caused by breaking internal waves, both phenomena driving vertical transport of heat, carbon dioxide, and other climatologically important tracers in the ocean and at the air-sea interface. Consequently, ISWs play an important role in shaping the ocean circulation and distributions of heat and carbon within the climate system (see e.g., the review in Whalen et al. 2020). The relative importance of surface wave breaking due to large ISWs in the deep ocean remains to be accounted for.

Supplementary Materials: The following supporting information can be downloaded at: <https://www.mdpi.com/article/10.3390/rs14030587/s1>, Table S1: List of Sentinel-3 cases in SAR-mode retracking approach exhibiting large-scale ISWs in optical imagery (OLCI) in synergy with the corresponding sea surface signatures in the SRAL altimeter. Note that cases are referring to the Banda Sea in the Indian Ocean before 27 September 2018, and the Tropical Western Atlantic after that, and that some dates include two or three ISWs, which were captured in the same image (i.e., relative orbit and cycle).

Author Contributions: Conceptualization, Adriana M. Santos-Ferreira (A.M.S.-F.); Data curation, A.M.S.-F. and Samira Amraoui (S.A.); Formal analysis, A.M.S.-F.; Funding acquisition, José Carlos Bastos da Silva (J.C.B.d.S.) and Franck Borde (F. B.); Investigation, A.M.S.-F., J.C.B.d.S., Thomas Moreau (T.M.) and Claire Maraldi (C.M.); Methodology, A.M.S.-F., J.C.B.d.S. and T.M.; Project administration, F.B.; Software, A.M.S.-F. and S.A.; Supervision, J.C.B.d.S.; Validation, François Boy and Nicolas Picot (N.P.); Visualization, J.M.M.; Writing of original draft, A.M.S.-F. and Jorge Manuel Magalhães (J.M.M.); Writing—review and editing, J.C.B.d.S., T.M., C.M., François Boy and N.P. All authors have read and agreed to the published version of the manuscript.

CHAPTER 3

Prologue

In the same fashion as for the previous chapter, a few notes are discussed before presenting the results concerning the ISWs documented in the cold tongue of the Pacific Ocean, which help frame this chapter in the overall structure of this thesis and provide additional context for the sequence of events leading to the results that follow.

As in the previous case (i.e. the contribution to surface wave breaking from ISWs), the key data allowing for this study were again acquired via S3 and its ability to quickly survey case studies in online interactive interfaces (e.g. Ocean Virtual Laboratory). In fact, it is noteworthy that the first observation found in the Pacific Ocean was simply a case of fortuitous luck. Approximately midway through this PhD, a pursuit for novelty led to a series of efforts to document high frequency phenomena in remote regions where only satellite capabilities were available. The Pacific Ocean was clearly a top candidate (given the little or no availability of in situ data) and a systematic search was undertaken with special emphasis on the equatorial region. The results were initially disappointing, with effectively no cases worthwhile investigating, until a single case emerged with clear and distinctive surface manifestations of an ISW packet (see **Figure P3.1**).

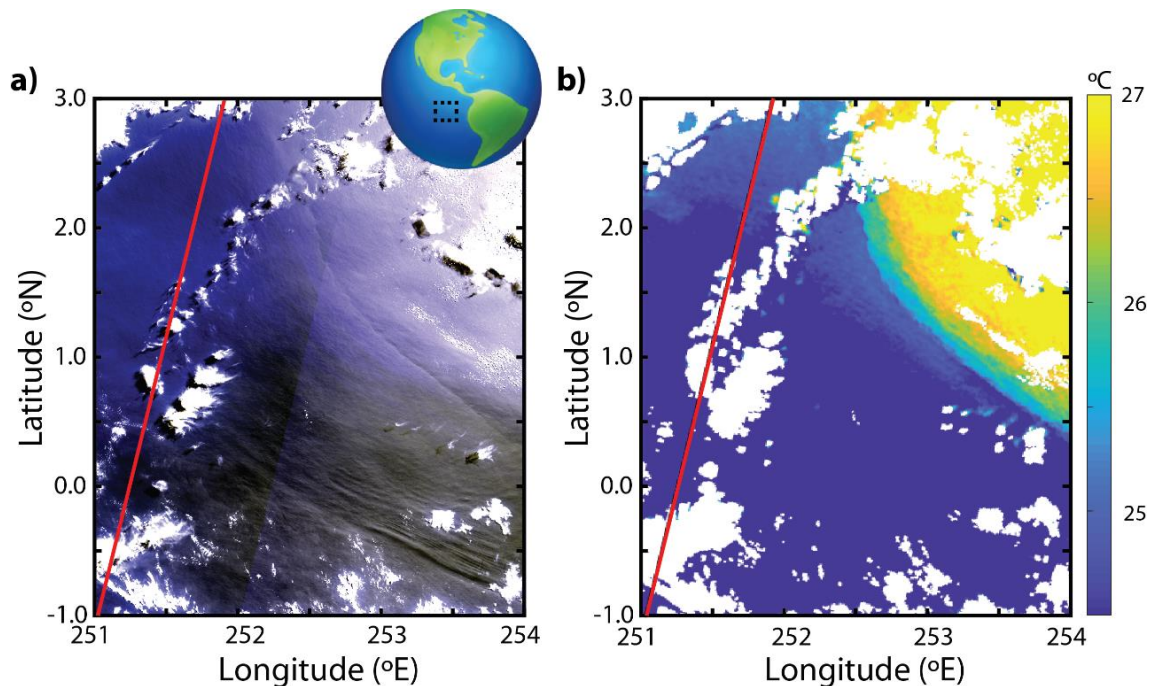


Figure P3.1 – (a) Sentinel-3A OLCI image dated 18 June 2019 (acquired at 17:12 UTC) presented in quasi-true color and showing clear surface manifestations of ISWs propagating towards the southwest. The red line represents the Sentinel-3A ground track. (b) Sentinel-3A thermal infra-red image (Level 2 WST) in exact synergy with the left panel. Note how a thermal front appears closely related with the surface manifestations of the ISWs observed in the OLCI image.

The events that followed are easily summarized. A literature review in that region quickly came across an ISW global survey documented in Jackson (2007), which despite indicating previous observations in the equatorial Pacific, had no mention of their characteristics, origin or dynamics. Aside from this study, these waves had apparently never been documented before. This stimulated our curiosity even further and led to an early collection of other cases in the year of 2020 (a La Niña year, as we would later discover), and an attempt to investigate their sea surface signatures in S3 SAR altimeter - at the time something following naturally from the previous study in the Banda Sea and looking to expand evidence of ISW-induced surface wave breaking. Nonetheless, the ISWs in this region have amplitudes likely in the range of about 10 meters (see below [section 3.4](#) in Santos-Ferreira et al. 2023), meaning they are too small to leave a clear measurable signature in the SAR altimeter (see **Figure P3.2**). Interestingly, this turned our attention to focus not on how energy in these ISWs could cascade down to smaller scales, but rather on how such a significant number of their observations could be explained so far away from any forcing tides and bathymetry (Egbert and Erofeeva 2002). Again, it was S3 providing valuable insights via simultaneous synergies between OLCI and SST (from SLSTR), revealing a clear link between the ISWs and thermal fronts (from Tropical Instability Waves - TIWs, see right panel in **Figure P3.1**) originating ISWs within the Pacific cold tongue. Despite being observable in most years, these ISWs are more frequent in La Niña years. For instance, **Figure P3.2** shows an image found later in 2022 (corresponding also to a La Niña year) where it is possible to see what appears to be the initial stages where ISWs emerge from a gravity current, itself probably originating from a TIW (Warner et al. 2018).

The synergies provided by S3 allowed a meaningful investigation into the waves generation mechanism, but in this regard a small note is warranted. The issue relates to previous Earth observation satellites that had already provided the scientific community with equal imaging (and synergy) capabilities - e.g. ENVISAT launched in 2002 also has an imaging spectrometer (MERIS) with similar resolutions of about 300 meters, as well as a radiometer (AATSR) for precise SST data (with pixel resolution of 1 km x 1 km). Why is it then, that these waves were not documented and explained before?

The answer is twofold. First, as discussed previously, a systematic approach of considerable persistence was needed to find a first case and from there a comprehensive collection with sufficient data to investigate the waves' dynamics. Second, and most importantly, even if provided with enough data, a key piece of the puzzle would still be missing. The satellite images may provide the horizontal structure of the ISWs and hint about their origin from gravity currents, but the vertical structure therein needed additional data. It was (fundamentally) the work done by Sally Warner and co-authors (Warner et al. 2018) that delivered the observations with the detail needed to characterize the density gradients, vertical structure, and the gravity current's dynamics over the cold tongue of the Pacific Ocean, which were the basis of the modelling efforts described in section 3 of Santos-Ferreira et al. (2023). Without these measurements, a plausible view of these waves' generation mechanism would simply remain too ambiguous.

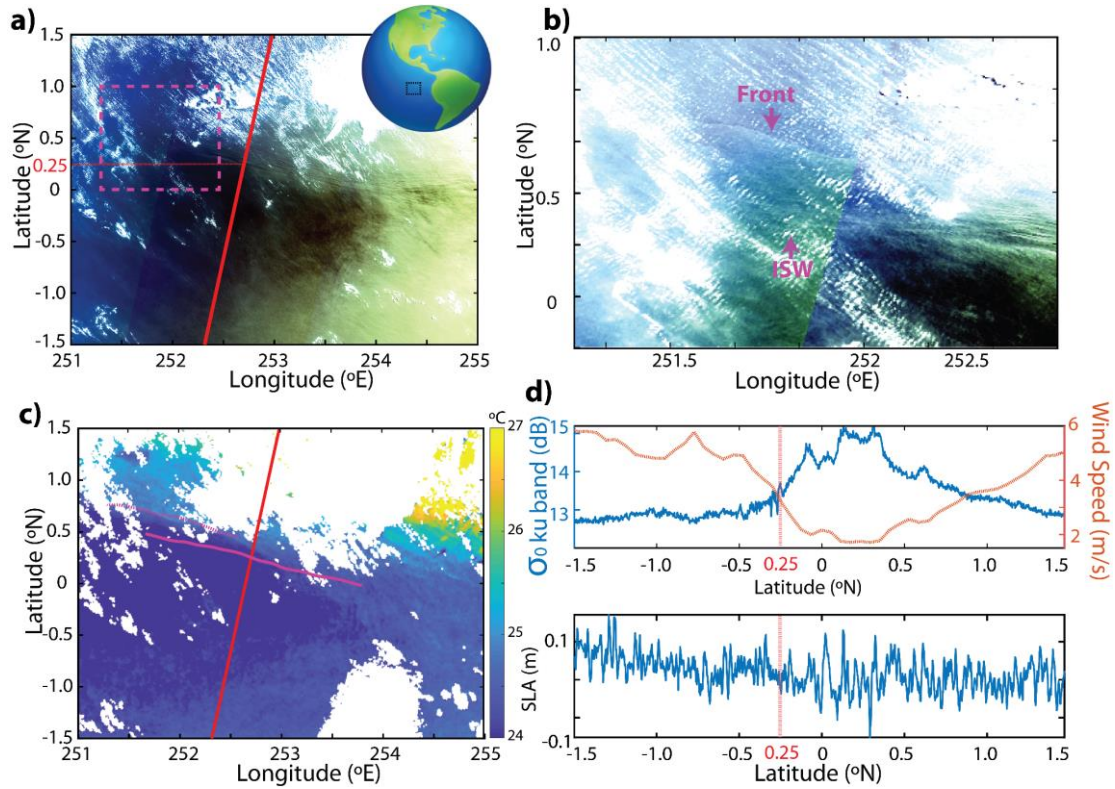


Figure P3.2 - (a) Sentinel-3B OLCI image dated 27 April 2022 (acquired at 17:07 UTC) presented in quasi-true color and with evidence of ISWs propagating towards the south-southwest. The red line represents the Sentinel-3B ground track and red dashed line the latitude of selected ISWs. The square in a magenta dashed line corresponds to a zoom shown in (b). **(b)** zoom showing details of the ISWS and where it is possible to note a front preceding their location by about ten kilometers. **(c)** Sentinel-3B thermal infrared image (Level 2 WST) acquired in exact synergy with top-left panel. The magenta in full and dashed lines are contours of ISWs and the ocean front visually identified in (b), respectively. **(d)** Selected Sentinel-3B SRAL altimeter parameters. In the upper plot, in blue, is the backscatter standard Level-2 SAR Ku-band and in dashed brown the wind speed (ms⁻¹). In the lower plot is the corresponding SLA (meters).

A final note follows concerning that generation mechanism. Literature reviews regarding IWs systematically point out to the barotropic tide as their main source, from which a series of generation mechanisms can be found at work in either deep or coastal waters (see e.g. Jackson et al. 2012). Cases documenting ISWs originating from processes involving phenomena other than tides are in much less number, and only recently it has been suggested that their contributions to global ocean energetics may be larger than previously thought. For instance, waves generated from river plumes have been described in Nash and Moum (2005) and in Mendes et al. (2021), and from ocean steady currents e.g. in Lentini et al. (2016) and in Magalhaes and da Silva (2017b). In light of discussions in the earlier sections of this thesis (namely those concerning energies transfers in the ocean from/to IWs), it is again stressed here that in these cases the ISWs are being forced not by tides, whereas the latter case shows evidence of a direct link between the large ocean scales of quasi-steady currents and the submesoscales of typical IW motions. In light of this recent evidence, it can

be wondered how much more energy in the ocean can follow similar pathways and what are the mechanisms behind them.

This is of direct relevance for the results that follow (published in Santos-Ferreira et al. 2023) given that they too provide evidence of another non-tidal ISW mechanism forced by the large-scales of the ocean - in this case when the cold tongue in the Pacific Ocean appears and interacts with westward moving TIWs, which is more frequent during La Niña years. In a way, from a dynamical point of view, the problem can be regarded as a case of a gravity current (or perhaps more intuitively a density current), in which warmer and fresher water advances towards colder and saltier water almost like a surge, whereas in the process ISWs are generated at a deep layer of water (the pycnocline within the cold tongue). It is if a three-layer system is released by a lock-exchange in a laboratory experiment, where the less dense water would advance over the denser adjacent layer, generating internal waves in the deeper pycnocline. It is interesting to note, that gravity currents in the ocean have been investigated (see e.g. Yoder et al. 1994; Johnson 1996; Nash and Moum 2005; Warner et al. 2018; Mendes et al. 2021), but their observations in the real ocean are rarely reported - especially in such large scales as those reported here (see e.g. Figure 6 in Santos-Ferreira et al. 2023, in this thesis **Figure 3.6**), which adds insight to the view that non-tidal ISWs may play an important role in global ocean energy budgets.

INTERNAL SOLITARY WAVES WITHIN THE COLD TONGUE OF THE EQUATORIAL PACIFIC GENERATED BY BUOYANT GRAVITY CURRENTS

(Santos Ferreira et al. 2023)

Abstract

The equatorial cold tongue in the Pacific Ocean has been intensely studied during the last decades as it plays an important role in air-sea interactions and climate issues. Recently, Warner et al. (2018) revealed gravity currents apparently originating in Tropical Instability Waves. Both phenomena have strong dissipation rates, and were considered to play a significant role in cascading energy from the mesoscale to smaller horizontal scales, as well as to vertical scales less than one meter. Here, we present Sentinel-3 satellite observations of Internal Solitary Waves (ISWs) in the Pacific cold tongue near the equator, in a zonal band stretching from 210° to 265°, away from any steep bottom topography. Within this band these waves propagate in multiple directions. Some of the waves' characteristics, such as the distance between wave crests, crest lengths and time scales, are estimated from satellite observations. In total we identify 116 ISW trains during one full year (2020), with typical distances between crests of 1500 m and crest lengths of hundreds of km. These ISW trains appear to be generated by buoyant gravity currents having sharp fronts detectable in thermal infrared satellite images. A 2D numerical model confirms that resonantly generated nonlinear internal waves with amplitudes of $O(10)$ m may be continuously initiated at the fronts of advancing gravity currents.

Significant Statement

Satellite imagery reveals the repeated occurrence of internal solitary waves in the near equatorial region of the East Pacific, despite the absence of topography. These waves appear to be resonantly generated over the sheared Equatorial Undercurrent by gravity currents that propagate as frontal zones of 1000 km scale tropical instability waves, providing a physical link with viscous mixing scales.

3.1. Introduction

Intense ocean fronts along the eastern equatorial Pacific, observed in a zonal band from 5°S to 5°N, have been frequently reported in literature. For example, Warner et al. (2018) (hereafter WEA) presented observations and model simulations of two major fronts in this region, the North and South Equatorial Fronts (NEF and SEF, respectively) that bound the equatorial cold tongue (**Figure 3.1**). These fronts are frequently populated with eddies that

develop into Tropical Instability Waves (TIWs) possessing sharp gradients in surface temperature, surface salinity, and ocean colour. They can be observed in satellite Sea Surface Temperature (SST) images, especially during La Niña years, as will be presented below.

During La Niña and ENSO-neutral years, TIWs with wavelengths of about 1000 km are present during the boreal autumn and winter, owing to the intensification of the trade winds that shape the upper equatorial circulation (Willett et al. 2006) into westward-flowing North and South Equatorial Currents (NEC, SEC) in between which lies the eastward-flowing North Equatorial Counter Current (NECC). Beneath the SEC, the subsurface Equatorial Undercurrent (EUC) flows eastward, with a core centered at about 100 m at 220°E that shoals from west to east. The upper currents (especially the NEC and NECC) may become unstable owing to strong horizontal shear between the westward-flowing NEC and SEC and the eastward-flowing NECC and EUC, molding the NEF into a wave-like pattern of cold crests (or cusps) and warm troughs, usually confined southward of 5°N down to the equator. Typical TIW periods peak at 17 and 33 days (Lyman et al. 2007, Willett et al. 2006). The sharpening of the NEF and SEF is well documented, e.g. in thermal imagery from satellites (WEA; Caltabiano et al. 2005), but not much is known about the physical processes involved in their intensification in late boreal summer and autumn, or about their subsequent dissipation, typically at the end of boreal winter.

Associated with TIWs are coherent anticyclonic Tropical Instability Vortices (TIVs) that slowly propagate westward at quasi-steady phase-velocities, dissipating in the western Pacific after a couple of months (**Figure 3.1**). While TIVs are highly coherent throughout their life-span, they can exchange up to 50% of their water mass with the surroundings after one complete vortex rotation, which typically lasts 20 days (see Dutrieux et al. 2008). This exchange presumably reflects mixing with surrounding waters, leading to a buoyant surge overshooting the cold/salty equatorial water (WEA) and, as revealed here, disintegrating into internal solitary waves (ISWs).

Sharp small-scale fronts (SSFs) often appear in the presence of TIWs (see WEA) both along their leading and trailing edge. Leading and trailing edge fronts are labeled LEF and TEF respectively (**Figure 3.1**). For example, Yoder et al. (1994) and Archer et al. (1997) observed such fronts near 220°E between 2°N and 4°N. They characterized these fronts, which were visible from ships, aircrafts, and satellites, as being “remarkably straight and narrow”. Johnson (1996) also encountered a sharp front at 2.1°N, 220°E along the western (leading) edge of a TIW cold cusp that exhibited downward velocities close to 1 cm s⁻¹ and a horizontal momentum balance between pressure gradient force and advection. This front was interpreted as part of a propagating gravity current made of warm water overriding the colder water (i.e. equatorial cold tongue), consistent with a non-rotating gravity-flow theory of subducting dissipative fronts (Johnson 1996). It has been suggested that some of the front’s energy loss was through turbulent dissipation, similarly to the SSFs reported by WEA, but a considerable portion must have gone through some other mechanisms, possibly in the form of radiating interfacial internal waves (Johnson 1996). However, the theory used by Johnson (1996) was inadequate to test this hypothesis. The main goal of the present study is to

examine the possibility that small-scale fronts could radiate internal wave energy in the Equatorial Pacific.

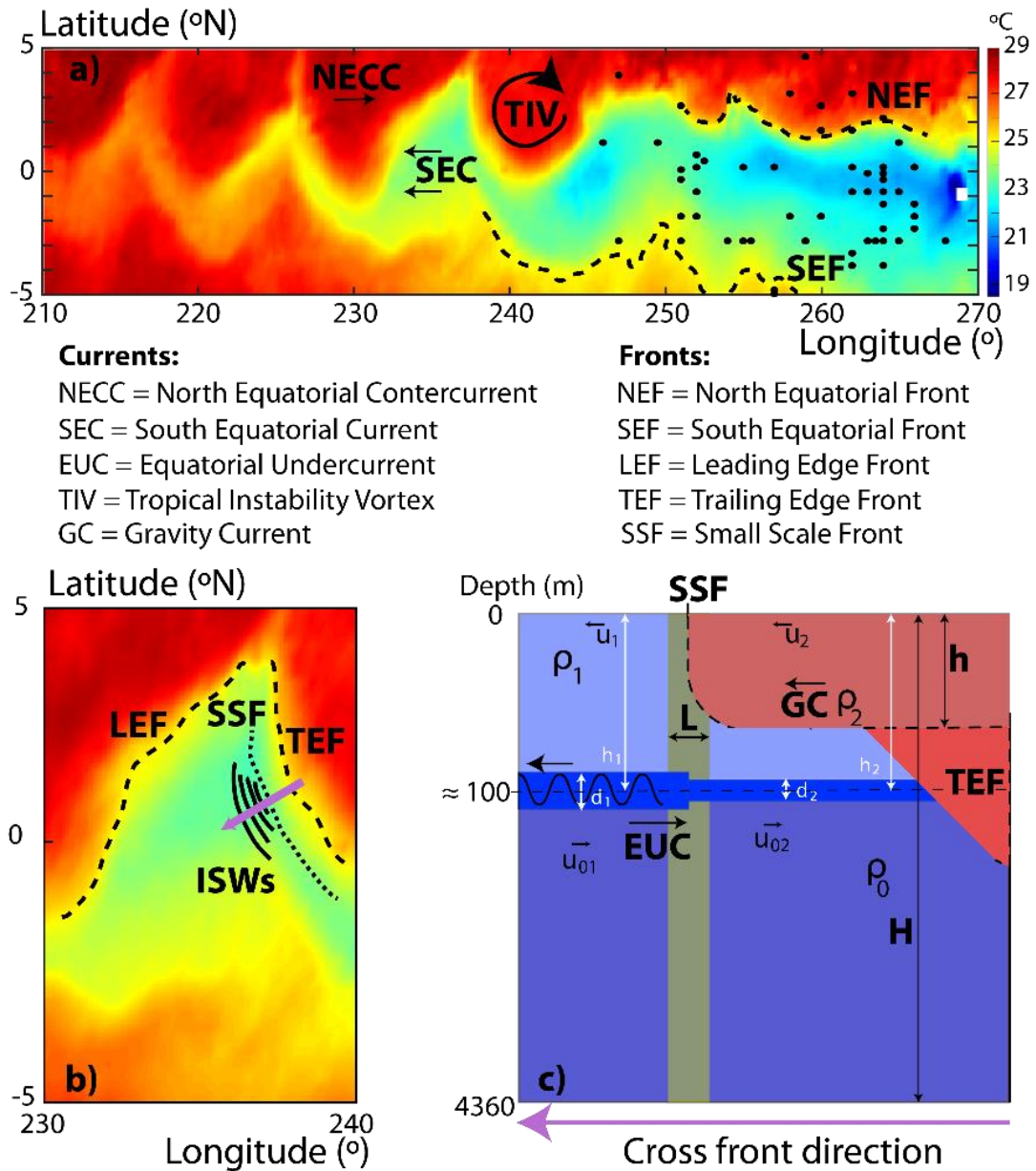


Figure 3.1 – (a) Spatial distribution of ISWs documented in Jackson et al. 2012 (black dots) in the eastern tropical Pacific. Surface currents (arrows) and ocean fronts (dashed black lines) are displayed over a daily Multiscale Ultrahigh Resolution (MUR) SST image (dated 19 June 2020). (b) Zoom of image section in (a) between 230°E and 240°E showing the SSF (dotted line) generated by the TEF and the (hypothetical) subsequent generation of ISWs (solid lines). (c) A schematic vertical section with a gravity current propagating southwestward of the TEF with buoyant upper layer of thickness h , density ρ_2 and initial velocity u_2 . The gravity current propagates over colder and saltier water in the equatorial cold tongue with density ρ_1 . The reference density underneath the thermocline is ρ_0 . Under the gravity current there is an initial velocity u_{02} and upstream an initial velocity u_{01} where a train of ISWs is represented. The direction of the gravity currents is shown by purple arrows.

3.2. Data and Methods

We use combined satellite image data, in situ data and numerical modelling to investigate the characteristics and origin of ISWs in the zonal band of the equatorial Pacific Ocean (from 5°S to 5°N) that are mostly visible inside the surface cold tongue. Combining SST satellite images with visible images affected by sun glint allows for the observation of both ocean fronts and ISW surface manifestations owing to gradients in surface roughness. Currents within the ISWs are characterized by convergent and divergent flow at the sea surface propagating in phase with the internal wave's subsurface waveform at the main pycnocline. These convergence/divergence surface bands cause variations in sea surface roughness at meter and sub-meter spatial scales that appear as distinctive bright features in optical sun glint imagery (Santos-Ferreira et al. 2022). A high resolution nonlinear and non-hydrostatic numerical model is used to complement satellite observations and estimate some ISW characteristics not retrievable from satellite data. The model also provides insights into the mechanisms of wave generation.

a) Satellite data

Our observation method consists of visual recognition of ISWs in Level-1b Ocean and Land Colour Instrument (OLCI) optical imagery from Top-Of-Atmosphere (TOA) radiometric measurements provided by the European Space Agency's (ESA)-Copernicus Open Access Hub. These images are geometrically corrected, calibrated, and provided as optical multichannel band images. The images are also ortho-geolocated, with geographic coordinates provided for each image pixel with a nominal spatial resolution of 300 m at nadir. This optical sensor on-board of Sentinel-3A and Sentinel-3B satellites provides quasi-true colour images over a swath width of 1270 km, covering most of the equatorial Pacific on a daily basis. The OLCI swath is not centered at nadir in order to mitigate the effects of sun glint patterns (direct specular reflection of the Sun in the ocean) but has a westward inclination. However, these sun glint patterns allow easier visibility of oceanic internal waves.

We use Sea and Land Surface Temperature Radiometer (SLSTR) products from Sentinel-3 satellites, namely the Level-2 Water Surface Temperature (WST) product providing SST. The spatial resolution of the SLSTR is 1 km, covering a swath width of 1400 km, and the sensing data are obtained simultaneously, i.e. synergistically, with an accuracy of 0.2 K, because these instruments (OLCI and SLSTR) are on the same platform.

The SEF and NEF have a surface thermal signature, visible only when cloud coverage is reduced. Obviously, cloud cover is very frequent around the equator, where cloudiness over the ocean is around 72% (King et al. 2013). Hence, we complement the SLSTR observations with the Multiscale Ultrahigh Resolution (MUR) SST image data available daily from NASA JPL Physical Oceanography DAAC (<https://doi.org/10.5067/GHM25-4FJ42>). MUR imagery is a global, gap-free, gridded, daily SST dataset created by merging multiple Level-2 satellite SST datasets. Data are available from 2002 to present (Chin et al. 2017).

A satellite image dataset has been assembled with images showing fronts and ISWs for the year 2020 during which a moderate to strong La Niña event emerged. The statistical analysis (section 3.3d) is conducted on data between 230°E and 265°E, and between 5°S and 5°N. For 2020, nineteen relative orbits of Sentinel-3A and nineteen relative orbits of Sentinel-3B are chosen for scrutiny. These 38 orbits correspond to, respectively, 12 and 13 satellite cycles (an orbit cycle is the time taken for the satellite to pass over the same geographic point, being 27 days for Sentinel-3A and Sentinel-3B). Sharp fronts and ISWs are more commonly observed between April and August, as a result of greater cloud cover during other periods of the year. The ISWs that are visible (with distinguishable amplitude and large enough crest-to-crest distance -- for short, hereafter referred to as ISW 'wavelength') are consistent with a generation by oceanic fronts. By "consistency", we mean crest orientations nearly parallel to the oceanic fronts visible in the same image and within reasonable distances from these fronts (typically on the order of 10-100 km, but sometimes much larger distances).

The phase speed of the ISWs is estimated from satellite data. For this, images on consecutive days are geolocated in the same projection and with the same resolution, overlapping them side by side. In these images, the ISWs leading crests are determined. Their average distance was calculated assuming a set of seven representative points along the first image. We determine the minimum distance of each of these points to points on the wave front on the following day. Since we know the exact times of the satellite images' acquisition, a reasonable estimate of the propagation speed of the ISW train is obtained by dividing the average distance between the two sets of points by the elapsed time. This method provides good results in previous studies (see e.g. da Silva et al. 2011; Tensubam et al. 2021), as long as wave fronts do not change shape significantly in the time elapsed between the images.

b) In situ data

To characterize density stratification in the ocean, we use monthly mean climatological data provided by NOAA/OAR/ESRL PSD, Boulder, Colorado, available online at <https://www.cpc.ncep.noaa.gov/products/GODAS/>. The dataset has a global spatial coverage, with resolution of 0.333° in latitude and 1.0° in longitude, and 40 vertical levels from 5 m to 4478 m depth. Climatological stratification profiles are used to compute the linear phase speed of internal waves propagating in the study region under the influence of climatological shear currents, such as the EUC (Behringer and Xue 2004). A Matlab code provided by Oregon State University tools in <https://blogs.oregonstate.edu/salty/matlab-toolsto-solve-the-viscous-taylor-goldstein-equation-for-both-instabilities-and-waves/> is used to solve the inviscid Taylor-Goldstein equation, providing theoretical estimates of the linear internal wave phase speed (Miles 1961; Smyth et al. 2011 and Lian et al. 2020).

c) Numerical modelling

Numerical simulations are performed to examine whether a gravity current with the hydrographic conditions typical of the study region reported by WEA (0°, 220°E and 0.75°N,

250°E) would be capable of generating ISWs. We use the nonhydrostatic and nonlinear 2-D model of Bourgault and Kelley (2004) that solves the non-rotating Euler equations in a vertical (x-z) plane. This model has previously been used in a similar configuration to simulate ISWs arising from gravity currents, but in different regions than WEA (Bourgault et al. 2016; Mendes et al. 2021).

The model domain is a two-dimensional ocean that extends laterally over a distance of 2000 km and vertically over a depth of 4360 m, representative of the mean depth in the region (Smith and Sandwell 1997). The model domain is bounded by a free surface, the ocean bottom, and open boundaries at $x = \pm 1000$ km. The Arakawa-C grid has 2296 points horizontally and 214 points vertically. The horizontal resolution decreases with distance from the central part of the domain ($-10 \text{ km} < x < 10 \text{ km}$). The smallest horizontal grid spacing (near the center of the domain) is $\Delta x = 10$ m. This resolution exponentially decreases to $\Delta x = \pm 10$ km towards the open boundaries. The open boundaries are located far away (± 1000 km) from the region of interest to minimize the influence of open boundary conditions. At the open boundary, far away from the central region of interest, Neumann conditions are taken, such that during the computation time (1 hour) any waves reflecting from these boundaries have not yet returned to this region. The vertical grid spacing is $\Delta z = 1$ m in the upper 150 m and progressively decreases to $\Delta z = 370$ m at the bottom. The duration of the model integration is 1 hour and the model output is every 5 seconds for all variables analyzed in this manuscript.

The model is initialized with a density field that mimics the gravity current observed by WEA. The initial density field is separated in three water masses with different densities: two surface layers (ρ_1, ρ_2) are separated by a front of width (L) centered at $x = 0$ m overlying a denser bottom layer (ρ_0) (**Figure 3.1c**). The surface layers are separated from the bottom layer by a pycnocline of half-thickness d_1, d_2 centered at a depth of h_1, h_2 , respectively. The vertical density distribution within the pycnocline follows a hyperbolic tangent shape where $\rho_i, d_i, h_i, (i=1,2)$ and ρ_0 determine its steepness. The same function is used to create the front, except that L , instead of d_i , determines the horizontal gradient of horizontal velocities and densities. Two model simulations are setup to represent idealizations of the observations made by WEA (see parameters in **Table 3.1**). In **section 3.3e**, Run 2 will be taken as reference, as it is the simulation that best fits the gravity current average depth, hence best fits the observations on November 2, 2008 (WEA).

Table 3.1 - Parameter values for each model run. Densities ρ are given in kg m^{-3} ; velocities u in m s^{-1} ; and depth scales, h, d and L in m.

Run	ρ_0	ρ_1	ρ_2	u_{01}	u_{02}	u_1	u_2	h_1	h_2	d_1	d_2	L
1	1026.5	1023.5	1023.3	0.5	0.502	-0.14	-0.36	40	40	15	10	50
2	1026.5	1023.5	1023.3	0.5	0.506	-0.14	-0.36	100	100	15	10	50

The velocity field at time $t = 0$ is defined in a similar manner to the density field described above. There are two surface velocities u_1, u_2 , respectively associated with the surface layers of density ρ_1, ρ_2 , and the horizontal velocity distribution across the front follows a hyperbolic tangent function with u_i, L, x as parameters. However, the velocity field of the deeper (ρ_0) layer comprises two components u_{01}, u_{02} (see **Table 3.1**). The velocity field u_{01} occurs below the surface layer with density ρ_1 , which is on the left, southwestward side of the front. The velocity u_{02} , below the surface layer with density ρ_2 , is initially calculated to balance the transport on each side of the front. For more details about the model set-up, the reader is referred to Bourgault and Kelley (2004) and Bourgault et al. (2016), and references therein.

3.3. Results

Satellite imagery in 2020 shows that in the eastern equatorial Pacific ISWs cluster around the equatorial waveguide (between 5°S and 5°N). This occurs preferentially within the cold tongue, where TIWs are most energetic (see **Figure 3.2**). These ISWs are found in 103 Level-1b Sentinel-3 OLCI images, selected by visual inspection of cloud-free images (for both Sentinel-3A and Sentinel-3B satellites). In each image, the centers of the largest ISWs are selected, marked in **Figure 3.2** (magenta dots). In 13 images we find two or more centers for different ISW trains. Each “center” is defined as the middle of the crest of the first soliton of each ISW train. The same map shows ISWs (black dots) registered in Jackson et al. (2012) using the method described in section 3.2 (see also **Figure 3.1a** and Jackson 2007). These authors made a global map of ISWs, based on a comprehensive set of Moderate Resolution Imaging Spectroradiometer (MODIS) images, where they could identify the oceanic ISWs during a period of time greater than one year (from August 2002 until May 2004, corresponding to a mild El-Niño period). However, we stress that no generation mechanism was attributed by these authors to the ISWs along the equatorial Pacific band.

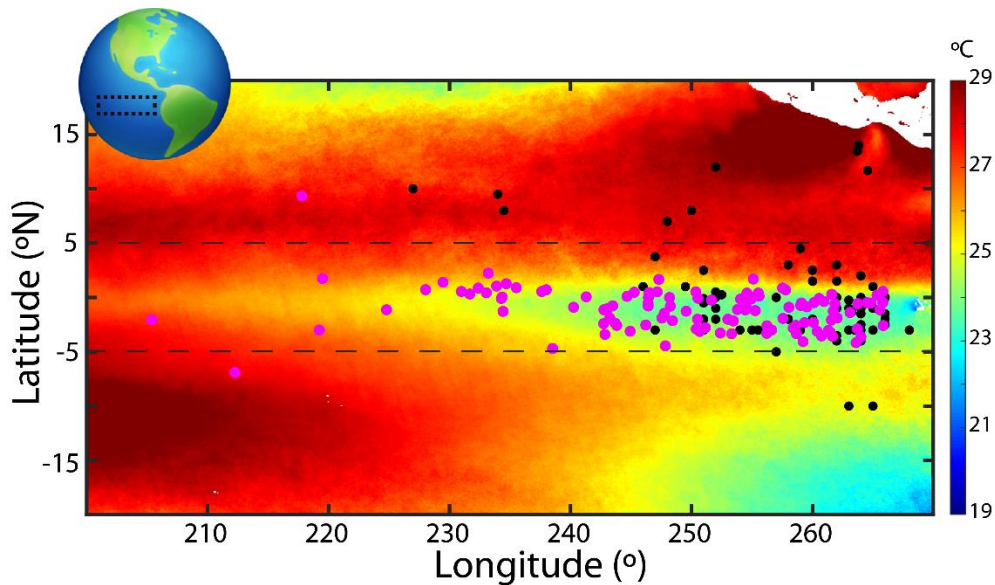


Figure 3.2 - Spatial distribution of the ISWs found in this study (magenta dots) for the year 2020. The black dots are the ISWs reported in Jackson et al. (2012). A MUR SST annual average image for the year 2020 is shown in the background.

a) Case studies

Two case studies provide insight into the generation mechanism of ISWs and allow to estimate their average phase speed for particular days. In Case 1, multiple fronts and ISWs are identified propagating in different directions, the common factor being that ISWs and thermal fronts are aligned in a quasi-parallel direction. This suggests a link between the thermal fronts and the generation of ISWs. In Case 2, an estimation of the average propagation speed of ISWs (admittedly also linked to the thermal fronts) is made, based on observations separated in time by approximately one day.

Fronts and ISW signatures are identified first in visible satellite images. All ocean fronts are confirmed in thermal imagery, while most ISWs do not have a thermal signature due to reduced spatial resolution of both SLSTR (1 km) and MUR.

CASE 1) 7 JUNE 2020

Oceanic fronts are ubiquitous in both SST and quasi-true colour images along the equator in a zonal band from 5°S to 5°N. **Figure 3.3a**, from 7 June 2020, shows a SLSTR SST colourcoded image where a vast region of cold water is surrounded by warmer water to the north and south. A front can be distinguished in the northeastern part of the region, oriented along the northwest-southeast direction (red line). It is similar to the SSFs documented by WEA. The SST varies by at least 1°C across the front and occurs within one pixel of the image, in somewhat less than 1 km (which is the resolution of the SLSTR). The same pattern (“front”) is observed and indicated in the quasi-true colour OLCI image (**Figure 3.3b**). In the same figure, internal wave features are identified and labelled “ISWs”. One of these ISW crests has a similar orientation as the thermal front.

Joint analysis of **Figures 3.3a and 3.3b** (SLSTR and OLCI) facilitates interpretation of various other ISWs leading crests (magenta lines). One of these is amplified in the zoom shown in **Figure 3.3c**, where several ISW crests are visible and rank ordered in both image brightness and wavelength. According to nonlinear wave theory, the rank ordered wavelengths of these wave fronts is an indication of the nonlinear character of the wave train. Hence, we refer to these waves as ISWs, as is common in literature ([Magalhaes et al. 2021](#)). This nonlinearity is confirmed in the image transect taken across the wave fronts (black line in **Figure 3.3c**), which is displayed in radiance levels along the wave profile (**Figure 3.3d**). The first two waves exhibit larger radiance amplitudes and wavelengths, typical of nonlinear dispersive waves described in solitary wave-like theories (e.g. KdV theory, see [Osborne and Burch 1980](#)). In this example, the approximate wavelength of the first ISW is more than 1800 m.

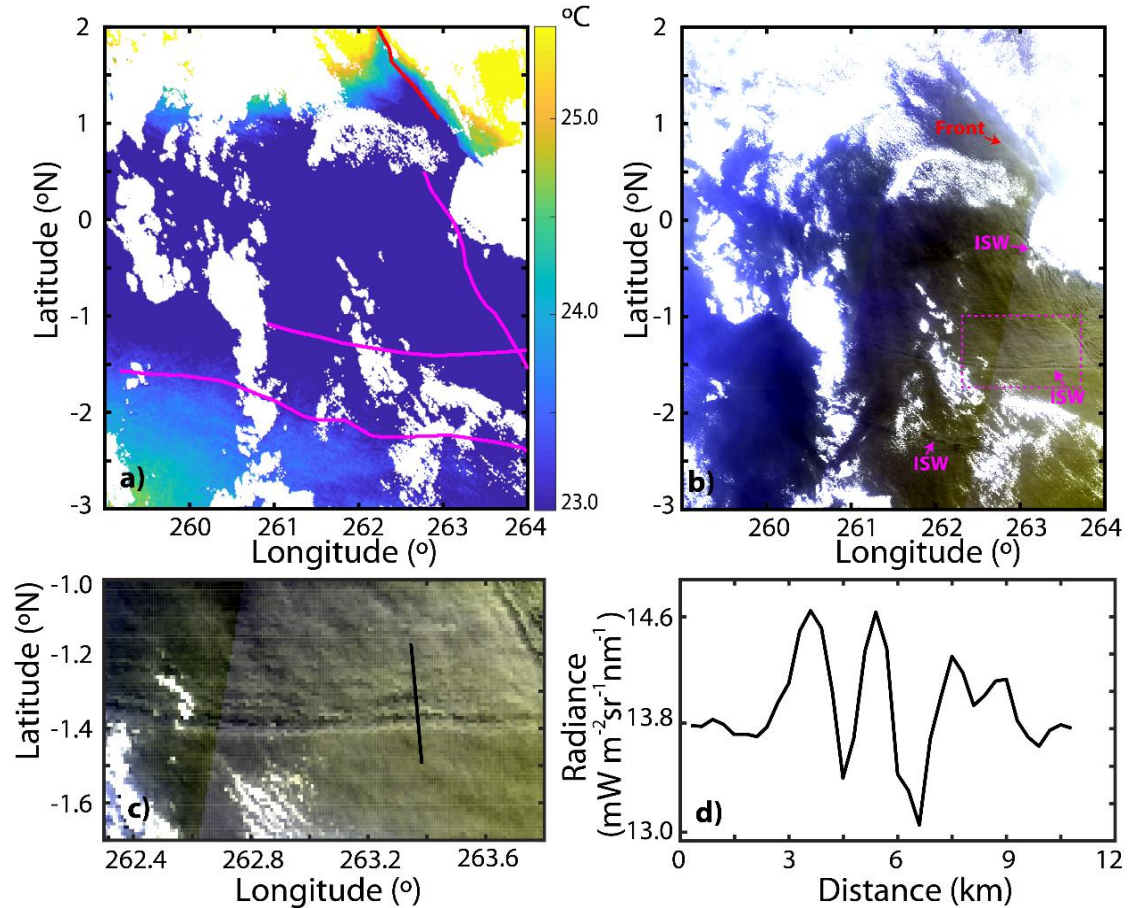


Figure 3.3 – (a) Sentinel-3B thermal infra-red image (Level 2 WST) dated 7 June 2020, acquired at 15:16:19 UTC (start time). The magenta lines and the red line are contours of the ISWs leading crests and the ocean front, respectively, also identified in (b). (b) Sentinel-3B visible image (Level 1b OLCI) simultaneous with a) and presented in quasi-true colour. The magenta arrows indicate the ISWs and the red arrow the front. (c) Zoom of image section (magenta rectangle) in (b) between 1.7°S - 1°S and 262.3°E - 263.8°E showing surface manifestations of ISWs owing to sun glint. The black line, perpendicular to the ISW crests indicates an image transect, shown in (d). (d) Radiance image transect (red band, 665 nm). The distance along the horizontal axis increases from south to north.

CASE 2) 15-16 JUNE 2020

Figure 3.4a gives a synergistic view of fronts (red lines) and ISWs (magenta lines) found on 15 and 16 June 2020. These are shown in visible images in Figures 3.4c and 3.4b, respectively, where the same ISW and front are visible. Figure 3.4a is a SLSTR SST colour coded image dated 16 June 2020 where a vast region of cold water is bounded by warmer water masses in the east and the northwest that are identified in Figures 3.4b and 3.4c. Several fronts can be distinguished in the eastern half and in the northwest in Figure 3.4a. The ISW leading crest in the eastern half (also visible in Figure 3.3a) is oriented along the northwest-southeast direction, while the front in the northwest is oriented along the northeast-southwest direction. Both are similar to the SSFs documented by WEA, along the TEF and the LEF, respectively (Figure 3.1). The same patterns are observed in the quasi-true colour OLCI (Figure 3.4b, acquired at the same time as Figure 3.4a). A front and ISWs are

also visible on 15 June 2020 (**Figure 3.4c**). **Figures 3.4d** and **3.4e** depict a zoom of **Figures 3.4a** and **3.4b** respectively, between 1.80-2.30°N and 236-236.8°E (see dashed black rectangles in each figure). The thermal gradient in **Figure 3.4d** coincides with small structure wave-like features in **Figure 3.4e**. We will discuss these patterns and possible interpretations in the [section 3.4](#).

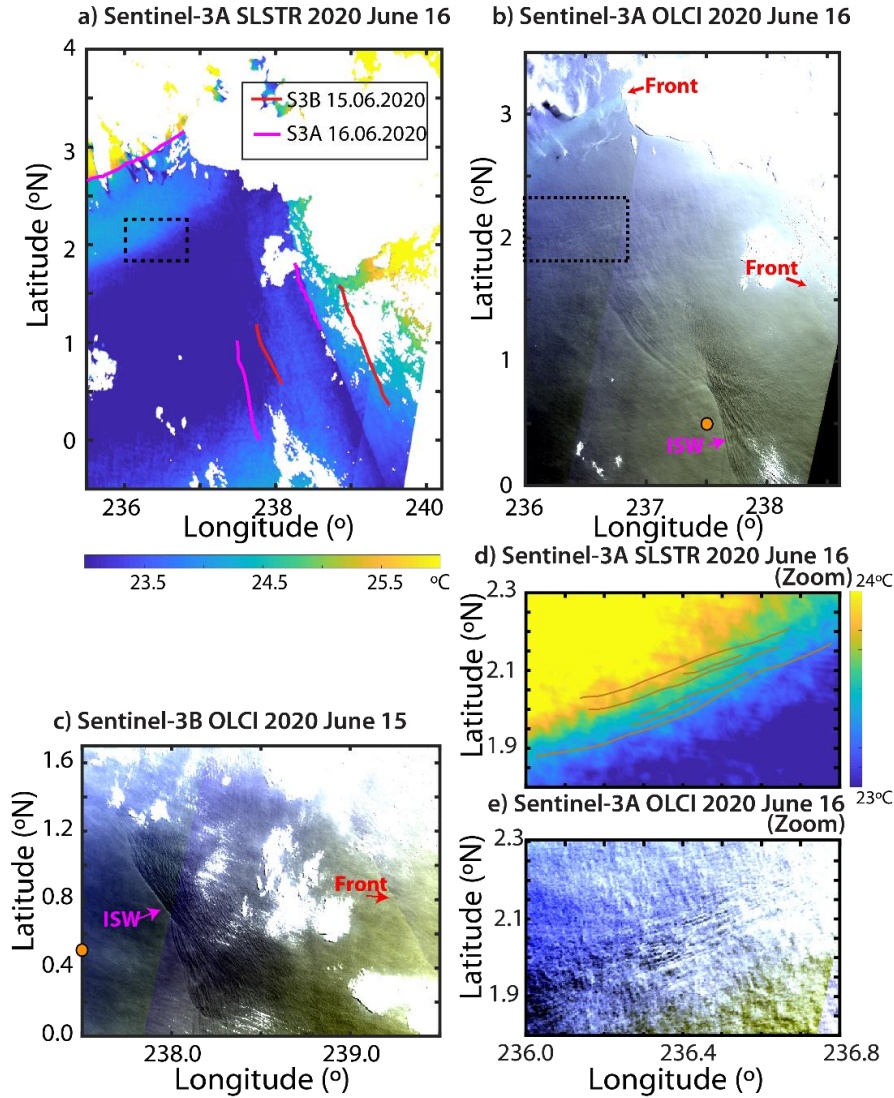


Figure 3.4 – (a) Sentinel-3A thermal infra-red image (Level 2 WST) dated 16 June 2020, acquired at 18:16:24 UTC (start time). The magenta and red lines are contours of the ISW leading crest and the ocean front identified in (b) and (c), respectively. The black rectangle corresponds to (d). (b) Sentinel-3A visible image (Level 1b OLCI) simultaneous with (a) and presented in quasi-true colour. The magenta coloured arrows indicate the ISWs and the red arrow indicates the front that is visible in the image. The black rectangle corresponds to (e). (c) Sentinel-3B visible image (Level1b OLCI) dated 15 June 2020, acquired at 18:03:08 UTC (start time) and presented in quasi-true colour (same label codes as in (a) and (b)). (d) Zoom of image section indicated by black rectangle in panel (a), between 1.80 - 2.30°N and 236° - 236.8°E, showing the SST gradient across the front (colour bars do not span the same range). The brown lines are manifestations of roughness that are seen in panels (b) and (e). (e) Zoom of image section indicated by black rectangle in (b) over 1.80 - 2.30°N and 236° - 236.8°E showing roughness linear features due to sun glint. The orange points in (b) and (c) correspond to the location of climatology data used in [section 3.3c](#).

Geo-locating and overlapping images from June 15 and 16 (**Figure 3.5**) reveals the location of the ISW leading crest (blue and green lines respectively), as well as a ‘corrugated pattern’ to the east of these fronts. The latter are believed to represent internal waves in the wake of the leading ISW. Based on the average distance between these ISW crests and the elapsed time, the ISW phase speed is estimated to be $0.50 \pm 0.03 \text{ m s}^{-1}$.

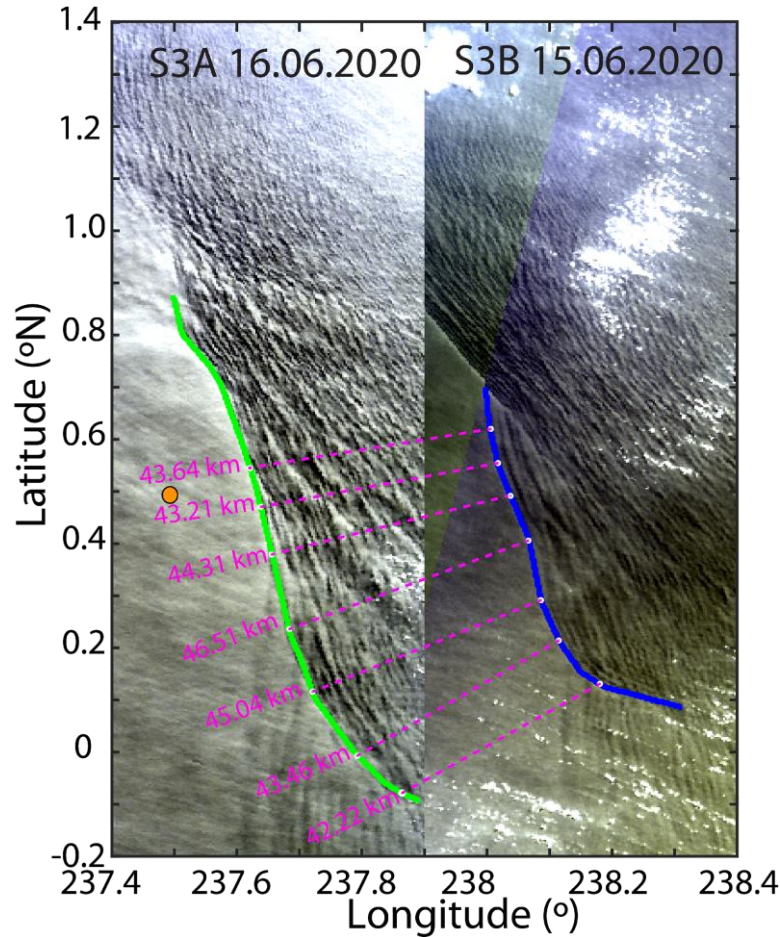


Figure 3.5 - Sentinel-3 visible images (Level 1b OLCI) dated 15 June 2020 (right) and dated 16 June 2020 (left). The leading ISWs are marked in blue and green, respectively. Magenta dashed lines mark distances between points on the same ISW leading crest, and are used to estimate the average wave propagation speed. Note that images are the same as in Figures 3.4b and 3.4c. The two images are superposed on a georeferenced map. The orange dot gives the location of climatology data used in [section 3.3c](#).

b) Large scale picture

Using **Figures 3.4** and **3.5** and other visible images which were partly cloud-free in the following 6 days, many ocean fronts and ISWs’ leading crests are identified and plotted onto a MUR SST composite image (**Figure 3.6**). This mode of display gives insight into the zonal and lateral extent and frequency of occurrence of both ocean fronts and ISWs. Dotted and solid lines indicate (sharp) thermal fronts and ISWs respectively, different colours representing different days. The MUR SST image in the background, from 19 June 2020, shows that the orientation of ISWs follow small scale fronts, which in turn appear to follow

those of the main TEF and LEF fronts referred to by WEA. Twelve visible images are used during 8 consecutive days, in the period 15-22 June 2020, extending along the equator over at least 45° in longitude. We identify 26 wave trains and 17 sharp fronts. Nine ISWs could be unambiguously associated with a specific front, leading to nine ISW-front pairs. The map in **Figure 3.6** is relevant because it is rare to find simultaneously ISWs and ocean fronts during several consecutive days owing to cloud coverage and sun glint. It is interesting to note the neat cusps and troughs of the TIWs delineating many of the sharp fronts (dotted lines), particularly to the north of the equator. Note however that the fronts that were found on 22 June 2020 are somewhat displaced to the west in comparison to the MUR SST image. This is attributed to the time elapsed between the MUR SST image and the OLCI/SLSTR.

c) Prediction of internal wave linear phase speed in the study region

Climatological data are used to estimate the linear phase speeds for the internal waves described in [sections 3.3a](#) and [3.3b](#) based on a standard Boundary Value Problem (BVP) with appropriate boundary conditions of the form:

$$(U_i - c_i) \left(\frac{d^2}{dz^2} - k^2 \right) \varphi - \frac{d^2 U_i}{dz^2} \varphi + \frac{N^2}{U_i - c_i} \varphi = 0, \quad (3.1a)$$

$$\varphi(0) = \varphi(-H) = 0, \quad (3.1b)$$

In this Taylor-Goldstein **equation (3.1a)** (Miles 1961), $\varphi(z)$ represents the (vertical velocity) modal structure, z the vertical coordinate, H the local depth and k the wavenumber. The phase speed c_i and horizontal currents $U_i(z)$ are given in East ($i=1$) or North ($i=2$) direction. Vertical profiles of Brunt-Väisälä frequency $N(z)$ and currents $U_i(z)$ are determined from climatology (Behringer and Xue 2004) at locations where the waves are observed (**Figure 3.6**). However, note that those climatology profiles are long term averages, and particular vertical profiles may significantly exceed these.

For the case study exhibited in [section 3.3a](#), **equation (3.1a)**, subject to boundary conditions **(3.1b)**, yields an estimate $c_1 = -0.5 \text{ m s}^{-1}$ for waves propagating to the west. This value agrees with the estimate obtained from the satellite images in [section 3.3a](#), case 2 (using a wavelength of 600 m, considering the examples in that subsection). **Figure 3.7** displays the vertical profiles of $N(z)$ and eastward directed $U_1(z)$ used in this calculation at 237.5°E and 0.5°N. The U_1 -profile clearly shows easterly currents centered at about 100 m deep, which could interact significantly with the wave at that latitude. This current is the well-known EUC, a narrow 400 km wide and fast undercurrent reaching speeds up to 1.5 m s^{-1} (Stellema et al. 2022). Disregarding these shear currents, the BVP yields a phase speed of about $\pm 0.76 \text{ m s}^{-1}$ (the $\pm$ sign corresponding to eastward/westward wave propagation respectively). The eastward directed EUC reduces the westward propagating phase speed to -0.5 m s^{-1} in the Earth's reference frame, showing the importance of the background flow on wave speed.

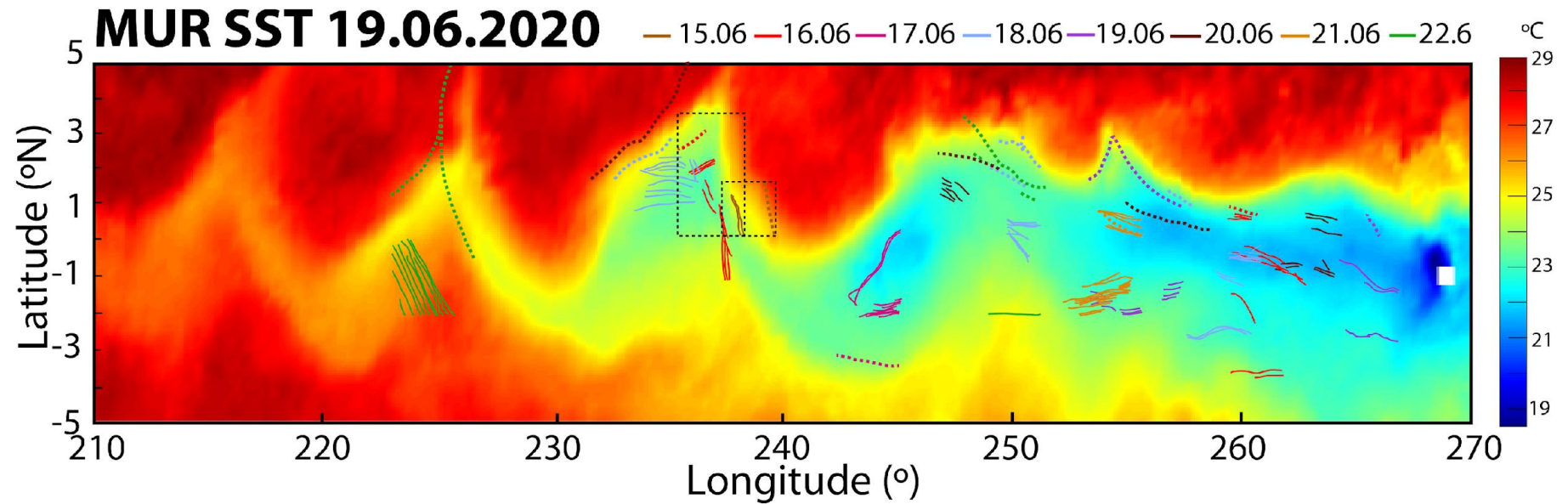


Figure 3.6 - Spatial distribution of the fronts (dotted lines) and ISW packets (solid lines), identified in the visible imagery of Sentinel-3, between 15 June and 22 June 2020. The colour for each day is identified in the upper part of the map. The dashed dark rectangles show the location of the images given in Figures 3.4 and 3.5. A MUR SST image, dated 19 June 2020, is shown in the background.

The impact of the EUC is further illustrated in **Figure 3.8**, where “west to east” etc., refer to the internal wave propagation direction. Here, phase speeds, evaluated from **equations (3.1a) and (3.1b)**, are plotted between 5°S and 5°N along 237.5°E, where images reveal strong ISWs (**Figure 3.6**). Due to shear in zonal currents, phase speed c_1 varies by as much as 67% for zonally propagating waves. Waves propagating in meridional direction are less affected by shear in meridional currents, with meridional phase speeds varying by 28% only. The main variation of the former occurs precisely where the EUC maximum is expected, between 2°S and 1°N (Behringer and Xue 2004). This “Doppler-like” effect is typical for internal waves propagating within strong quasi-steady ocean currents (see e.g., da Silva et al. 2011).

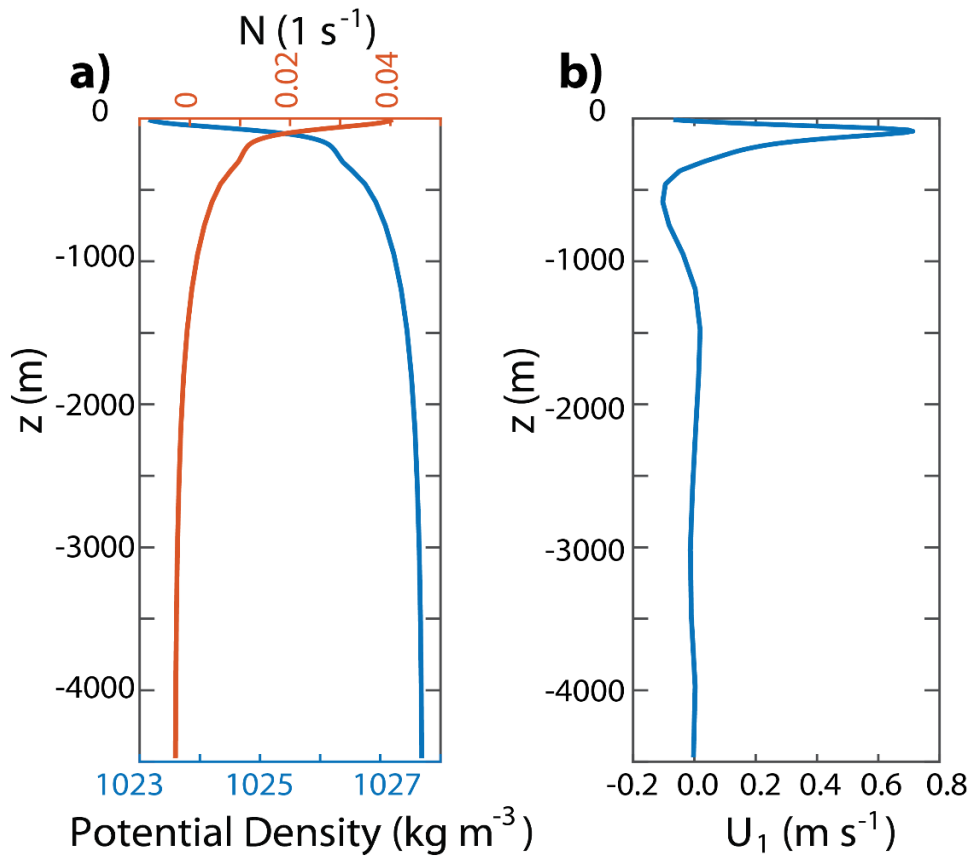


Figure 3.7 – (a) Potential density and buoyancy frequency (N) profiles, respectively in blue and orange colors. (b) U_1 velocity profile. All profiles are from monthly mean climatological data available at the location and time of the examples given in Figures 3.4 and 3.5 (i.e. June 2020).

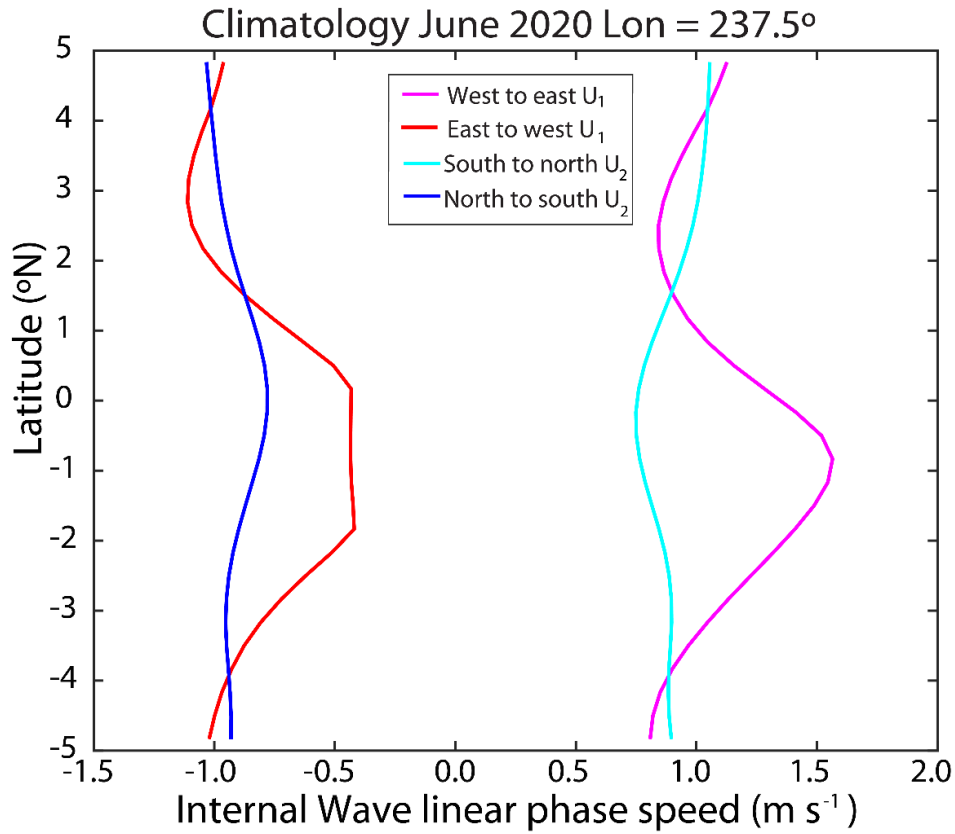


Figure 3.8 - Phase speeds of linear long internal waves evaluated for different vertical profiles of horizontal velocity (U_1 and U_2 , eastward and northward, respectively). The profiles of U_1 , U_2 and density have been taken between 5°S and 5°N at the fixed longitude of 237.5°E .

d) *Statistics of wave observations*

To determine the distance between subsequent crests and crest lengths we selected the “best” 66 cases of ISWs in the images. Here “best” relates to features that are most visible and/or largest, such that lengths can be measured with sufficient confidence. ISWs are most frequently observed from 5°S to 1°N (**Figure 3.9a**). This may partly reflect the decrease in cloud cover inside this latitude band, and the occultation of ocean features by clouds outside this band. The most frequent ISW wavelength estimates in this sample are in the range between 900 m and 1800 m, but waves with wavelengths larger than 4 km are also observed. Along-crest lengths are frequently in the range between 200 and 300 km, as depicted in **Figure 3.9c**. But ISW crests longer than 500 km are also found and are amongst the longest observed in the ocean (da Silva et al. 2011; Magalhaes et al. 2016; Tensubam et al. 2021).

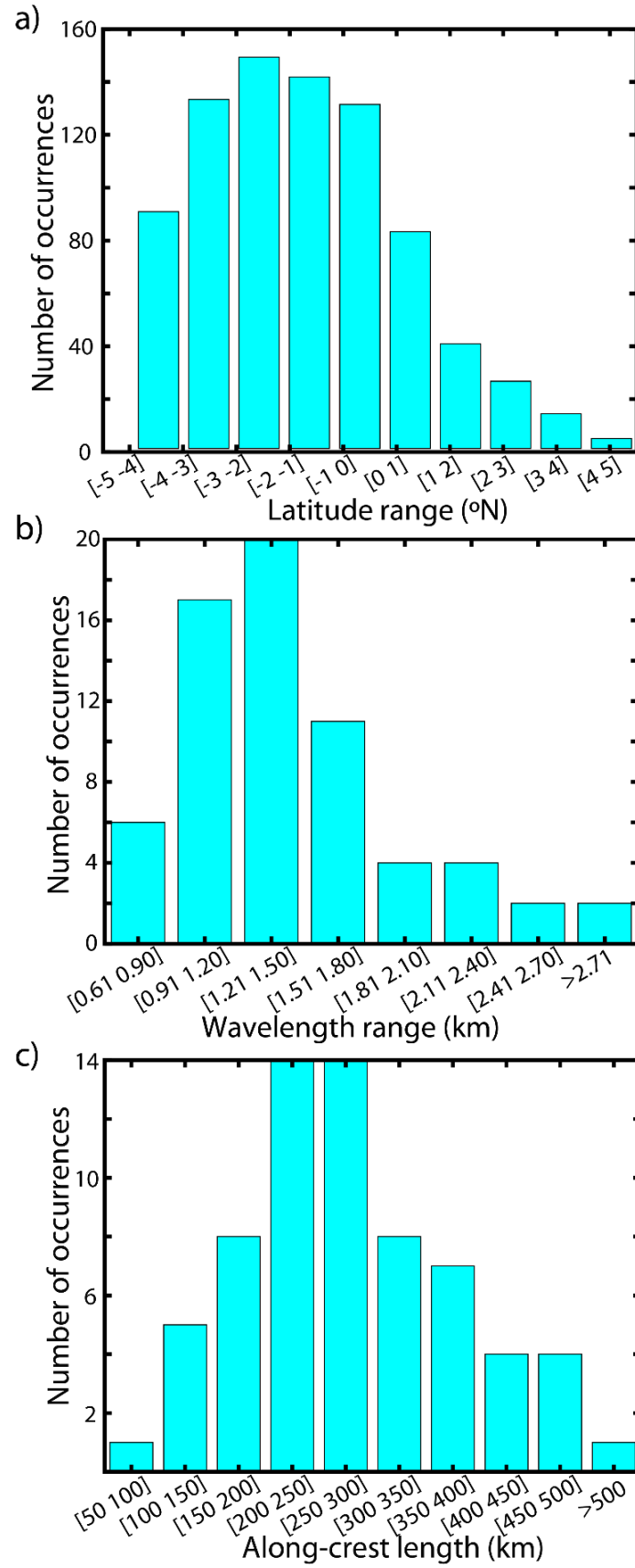


Figure 3.9 – (a) Latitude distribution of the 116 ISWs selected from 2020 images in the eastern equatorial Pacific Ocean. (b) Distribution of the ISW wavelengths for the best 66 cases for the same period of time. (c) As in (b) for along-crest length scale.

e) Model simulations

The initial conditions and parameter values used for the simulations are based on the observations of gravity currents shown by WEA. They involve fitting hyperbolic tangent functions to observational profiles of density and horizontal velocity. Two model runs are forced with different pycnocline depths, keeping other parameters approximately the same (**Table 3.1**). Here we present and discuss only one of the runs (Run 2) which reproduces reasonably well the measurements of November 2008, described by WEA; results from Run 1 are briefly mentioned but not displayed.

The initial conditions are intended to produce a gravitational adjustment to a lock-exchange configuration confined within the top 100 m (ambient surface layer, as sketched in **Figure 3.1c**). The reason for this choice of upper layer thickness is two-fold: firstly, this is the typical depth of the main pycnocline in the study region, and secondly, it leads to buoyant gravity currents with thicknesses typically observed by WEA, particularly for the measurements of November 2008. This setting corresponds to run 2 (see **Table 3.1**), when the model is adjusted to a three-layer setting. After gravitational adjustment the gravity current propagates at a constant speed, with an average thickness of 40 m.

Numerical model results (**Figure 3.10**) show typical crest-to-trough amplitudes of 10 m and maximum horizontal and vertical velocities of approximately $u \approx -0.3 \text{ m s}^{-1}$ and $w \approx \pm 0.04 \text{ m s}^{-1}$. These results may be interpreted as ISWs that propagate to the west (left side in **Figure 3.10**) into the equatorial cold tongue waters along the thermocline there, which is typically centered at 100 m depth. Furthermore, wavelengths of $\sim 1500 \text{ m}$ estimated from the model are consistent with the range of values measured in satellite images (**Figure 3.9b**).

The velocity sections portrayed in **Figure 3.10** represent the essential phases in the ISW generation and evolution. The propagation of the gravity current (see labels F1, F2 and F3 and dashed white line), the waves' detachment from the front and evolution (labels W1, W2 and W3 and dashed magenta line), and the propagation of the ISWs as pycnocline depressions upstream of the front (labels W'1, W'2 and W'3 and dashed magenta line), are all identified in 20 minutes of model simulation. Using the maximum horizontal velocity at the surface as indicators of soliton centers, W' and W refer to two consecutive ISWs generated at the gravity current's front (labeled F). Indices "1", "2" and "3" stand for three different times of model output, at respectively 30 min, 40 min and 50 min. The direction of wave propagation is indicated by horizontal black arrows. The wave amplitude and propagation speed increase through nonlinear dispersion. The wave is subsequently released from the front location when its phase speed significantly exceeds the front speed. If the front propagation is steady (for some period of time at least) the generation sequence is repeated after the complete detachment of the preceding solitary wave. In this process, energy is transferred from the gravity current to ISW trains.

A Hovmöller (space-time, time increasing downward) plot of the wave-induced horizontal velocity at the surface shows upstream-propagating ISWs, i.e. flowing in the opposite direction to the EUC (**Figure 3.11**). These appear as red colored bands of negative velocity (well depicted by contour $u = -0.3 \text{ m s}^{-1}$) that slope down and to the left. Those

distinct patterns of the ISWs in the model detach from the slower steady gravity current front. This is clearly seen in the blue colored “wedge” pattern that leans downward and to the left, contoured by the $u = -0.5 \text{ m s}^{-1}$ isoline. Average propagation speeds inferred from the slopes of the -0.3 and -0.5 m s^{-1} isolines in the Hovmöller diagram are -1.53 m s^{-1} and -0.48 m s^{-1} for the ISWs and gravity current, respectively (for run 1, nonlinear propagation speeds are -1.02 m s^{-1} and -0.40 m s^{-1} , respectively). Note that, during the time evolution in the model, the solitary waves appear rank ordered, meaning that amplitude and velocity perturbations are stronger for the leading ISW, and subsequently getting weaker for waves generated afterwards. This rank order, is observed in most of the satellite images we have analyzed, and is consistent with the findings in White and Helfrich (2012) for Type-II regime, i.e. gravity currents that generate upstream undular bores (see also their Figure 6).

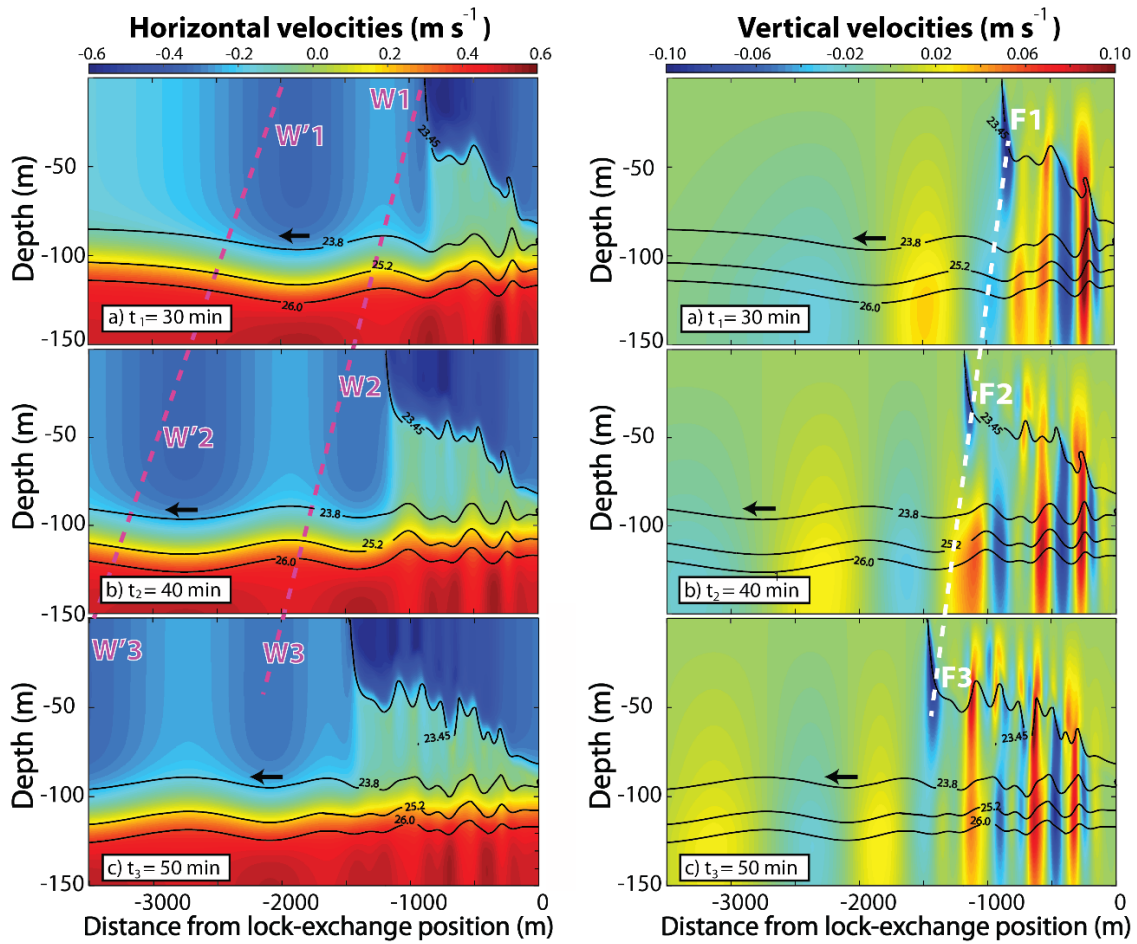


Figure 3.10 - Numerical simulations of ISWs generated by a gravity current at a) $t=30 \text{ min}$, b) $t=40 \text{ min}$ and c) $t=50 \text{ min}$. In the six plots it is possible to see the ISW propagation from right to left. The colours represent the horizontal velocities (left) and vertical velocities (right), while the black lines are density anomalies, σ . The magenta dashed lines (left panels), namely $W'1$, $W'2$, $W'3$ and $W1$, $W2$ and $W3$ are showing the propagation of consecutive ISWs. The white dashed lines (right panels) are representing the propagation of the gravity current.

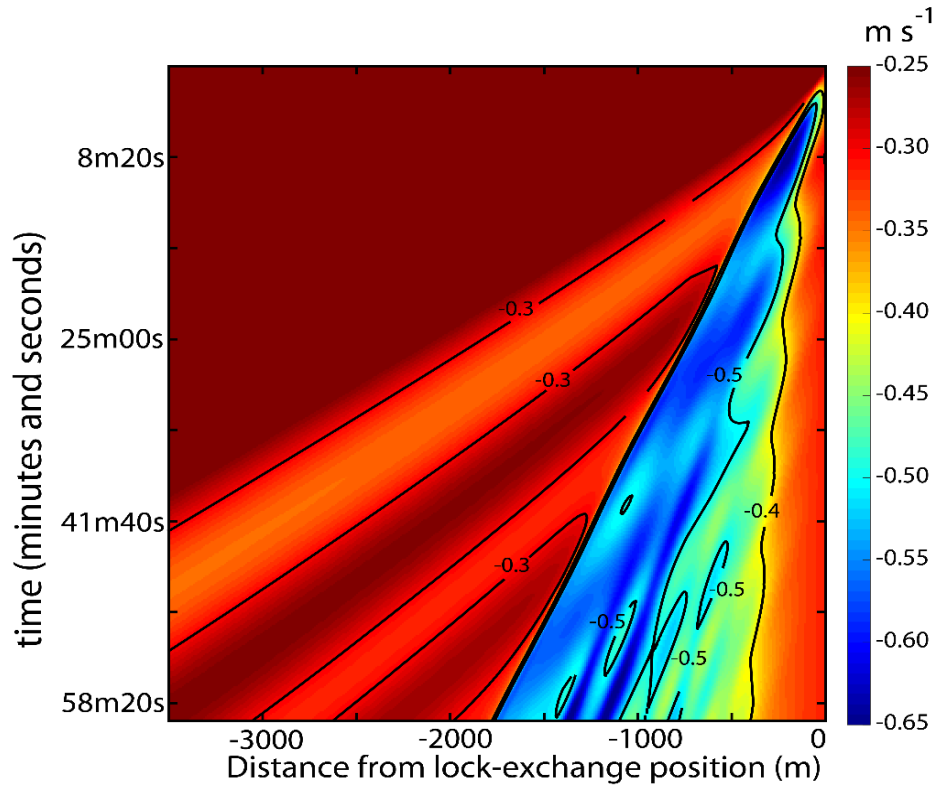


Figure 3.11 - Hovmöller (space-time) diagram for the surface horizontal velocity. The bright red leftward and downward leaning linear patterns show two consecutive ISWs. The blue coloured linear pattern, to the right of the ISWs, shows the gravity current.

3.4. Discussion

We found that long-crested ISWs (300 km or more) are visible in Sentinel-3 OLCI true colour images owing to strain on sea surface waves exerted by internal wave motions propagating along the main pycnocline of the Pacific Ocean between 5°S and 5°N. Idealized numerical simulations suggest that these ISWs are generated by gravity currents, themselves originating from large-scale fronts associated with TIWs (the NEF and SEF, that bound the equatorial cold tongue). Here, we connect these findings, proposing that these ISWs are upstream-propagating disturbances generated by gravity currents that propagate into the ambient stratification.

a) Gravity current's energy considerations

The propagation of the gravity currents observed in satellite images is subject to energy conservation constraints (WEA, following Simpson 1997). A gravity current, or “density current”, with thickness h and homogeneous density ρ_2 , holds a different potential energy than an adjacent (ambient) water mass with density ρ_1 . The potential energy difference can be written as $PE = (\rho_1 - \rho_2)g(h/2)$, where g is the acceleration due to gravity. When the system relaxes to equilibrium, assuming complete conversion from potential into kinetic energy $KE = (1/2)\rho_2 c_0^2$, that is $KE = PE$, the gravity current velocity is given by $c_0 = \sqrt{g'h}$, where $g' = g[(\rho_1 - \rho_2)/\rho_2]$ is the reduced gravity.

Based on our model simulation shown in [section 3.3e](#), itself based on the gravity current measured in November 2008 by WEA, we estimate h referring to the outcropping $1023.45 \text{ kg m}^{-3}$ isopycnal, at the front. This gives $h = 40 \text{ m}$, in good agreement with $h = 37.9 \text{ m}$ found by WEA. This estimate of h approximates the depth over which horizontal velocities u are homogeneous within representative regions before and after the passage of the gravity current front ($\Delta x \approx \pm 1000 \text{ m}$). The model simulation, averaged from $z = -h$ to $z = 0$, yields densities $\rho_1 = 1023.50 \text{ kg m}^{-3}$ and $\rho_2 = 1023.31 \text{ kg m}^{-3}$ on either side of the gravity current front. In case of complete PE to KE conversion, the resulting front velocity will be $c_0 = -0.27 \text{ m s}^{-1}$, the minus sign indicating leftward motion in **Figures 3.1c** and **3.10**.

However, taking the modelled front velocity u_{f0} , and subtracting the similarly averaged background flow field u_{bk} , yields the actual front velocity u_f in a reference frame moving with the background current:

$$u_f = u_{f0} - u_{bk} \quad (3.2)$$

Using the model, $u_{f0} = -0.48 \text{ m s}^{-1}$ and $u_{bk} = -0.35 \text{ m s}^{-1}$, the westward speed of the gravity current in the frame of reference of the background flow is $u_f = -0.13 \text{ m s}^{-1}$. The absolute value of u_f is smaller than that of c_0 , as expected, meaning that the energy transfer from PE to KE is not complete. The ratio of the front speed to c_0 may be viewed as a Froude number $Fr_h = |u_f/c_0| = 0.48$. This value is smaller than that reported by WEA, which is 0.68 for the November 2008 gravity current. WEA pointed out that this value indicates dissipation of energy owing to turbulence, and showed dissipation rates one thousand times greater in the gravity current compared to upstream waters. We further suggest that part of the potential energy is transferred to ISWs generated by the gravity current.

b) Generation mechanism of ISWs

White and Helfrich (2012) showed that gravity currents may transfer energy to upstream propagating ISWs, i.e. ISWs moving ahead of and in the same direction as the gravity current. This is consistent with hydraulic theory and significant to the problem of internal wave generation. These authors demonstrated an analogy with upstream resonance in two-layer flow over variable topography, where gravity currents “act as an obstacle to the flow” and are as effective to generate internal waves upstream as a topographic obstacle. Here we explore the hypothesis that the SSFs and associated gravity currents observed in the study region could generate interfacial waves with some degree of nonlinearity and propagate upstream of the gravity current.

White and Helfrich (2012) used nonlinear hydraulic theory and numerical modelling to classify gravity currents and their upstream flow dynamics. They found a wide range of parameters for which energy exchange between the front and upstream ISWs is significant. This range of parameters was interpreted by these authors as a *transcritical* regime, generating nonlinear internal waves and upstream bores by resonant transfer of energy (see also White and Helfrich 2008). However, ISW generation by gravity currents is more intricate than that by flow over topography since the front is a free boundary that may interact with

the internal waves. The problem is analogous to a dam break setting, in which a volume of fluid with density ρ_2 is released into a two-layer system, where the density and the thickness of the lower layer are, respectively, ρ_0 and d_0 , and the density and thickness of the upper layer are, respectively, ρ_1 and $H-d_0$, where H is total depth. In the model of White and Helfrich (2012), the “lock-exchange” release generates a propagating gravity current with front velocity u_{f0} , and approximate thickness h . (Note that their setting simulates the collapse of denser fluid at the bottom, for comparison with experiments; our model settings are reversed, because we wish to simulate a buoyant gravity current at the surface). The relative density difference between the gravity current and the ambient water is described by parameter $S \equiv (\rho_0 - \rho_1)/(\rho_0 - \rho_2)$. Defining the reduced gravity as $g^* = g(\rho_0 - \rho_1)/\rho_0$, we have $S = 0.94$ and $g^* = 0.03 \text{ m s}^{-2}$ for our model Run 2. Note that White and Helfrich (2012) scaled all lengths by H , and scaled all velocities by $(g^*H)^{1/2}$. In their model, the linear phase speed in non-dimensional units yields then $c_0/\sqrt{g^*H} \approx 0.17$, or an equivalent two-layer system with their parameter $d_0 \approx 0.03$. In our numerical experiment $U = u_{f0}/\sqrt{g^*H} \approx 0.48/10.95 = 0.05$. In the model of White and Helfrich (2012), in the (U, S) parameter plane for $d_0 = 0.1$ (their Figure 14a), our Run 2 falls into the lower right corner, for Type II resonant wave generation. Although White and Helfrich (2012) did their classification for gravity currents with lower limit $d_0 = 0.1$, these authors note that the size of the parameter plane that permits internal wave generation upstream of the gravity current front (their Type II class) increases for smaller d_0 and larger S . The case being considered in the equatorial Pacific Ocean, with $d_0 \approx 0.03$, would certainly fall within the Type II class (see White and Helfrich 2012). In this case, internal waves are forced at the front, their amplitude grows, increasing speed through nonlinear dispersion, and consequently are released from the front location when their phase speed exceeds the speed of the propagating front. This sequence is repeated when one solitary wave has completely detached from the front location, which is precisely the result in our numerical simulation (Figure 3.10). Therefore, the model results reported in this manuscript are consistent with resonant generation of ISWs, by which energy is transferred from the gravity current to the ISWs. In fact, White and Helfrich (2012) propose that, for ambient parameters similar to our case, as much as 40% of the total energy of the gravity current is used to generate internal waves.

c) ISW spatial structure and age

Concerning the observability of the ISWs in satellite sun glint images (i.e. OLCI true color), we remark that horizontal strain rates at the sea surface typical of internal waves vary from 10^{-4} to 10^{-2} s^{-1} (Pinkel 1979, Mendes et al. 2021). The former value is typical of internal waves generated by river plumes in coastal waters, while the latter is typical for strongly nonlinear ISWs of large amplitude. In fact, fields of surface current divergence/converge at kilometer and sub-kilometer scales of the order of 10^{-5} s^{-1} have been shown to produce observable contrasts in satellite imagery of the same kind (Kudryavtsev et al. 2012; see e.g. their Figure 6b). The model surface velocity field $u(x)$ for our case study yields a strain rate of about $\partial u/\partial x \approx 2 \times 10^{-4} \text{ s}^{-1}$. This value is similar to that for ISWs observed in the Messina

Strait (Mediterranean Sea) which produced strong radar signatures in satellite images (see Alpers and Salusti 1983). ISWs with strain rates of such a magnitude can be detected in sun glint images (Mendes et al. 2021).

As summarized in **Figure 3.6**, in the East Pacific ISWs co-exist with sharp fronts all along the equator from the Galapagos Islands (270°E) to the West, up to at least 220°E. Although **Figure 3.6** only reports an 8-day sequence of surface manifestations of these phenomena, selected owing to particularly good environmental conditions (little cloud cover), we speculate that similar features occurred in all seasons in 2020. Note that for the 8-day period in question, we identify 26 ISW trains, 17 sharp fronts, of which 9 pairs could be clearly related. The age of the ISW trains at the moment of observation may be roughly estimated from the distance between front and leading ISW crest. Assuming an average relative propagation speed of 1 m s^{-1} between ISW and front velocity (the relative speed suggested by the numerical model result), the ISW age is estimated to be between 4 and 55 hours. This yields an average age of 28 hours for the 9 observed pairs of fronts-ISWs. These values exceed the front's life-span from 4 to 11 hours estimated by WEA, because ISW energy dissipation is less vigorous than energy dissipation at fronts.

But, as seen in [section 3.3](#), the SSFs reported by WEA and Yoder et al. (1994) are different from the main thermal signatures north of 1.5°N (west of 262°E) in **Figure 3.3a**, which are larger and broader, and are associated with TIWs and in this case possibly with a TIV. The SSFs are interpreted as buoyant gravity currents, from which ISWs are generated. The distribution of internal wave trains is nearly uniform in latitude. The fewer number of cases found to the north of the equator is due to excessive cloud coverage in the Intertropical Convergence Zone, an area where trade winds converge into a zonal band (Willett et al. 2006). Along the zonal direction, ISWs appear less frequent to the west of 230°E (**Figure 3.6**) where fronts get weaker along the equator.

The ISW phase speeds within the cold tongue are quite variable. We have seen in [section 3.3c](#) that the linear phase speeds may vary by as much as 67% for waves propagating in the zonal direction and 28% for waves propagating in the meridional direction. The above estimates were made for internal wavelengths of 600 m. ISWs that propagate with a prevalent zonal component may propagate slower when they propagate against the EUC, or faster when they propagate in the same direction as the EUC. For the same reason, ISWs may propagate slower when they are generated by gravity currents originating from the TEF, or faster when generated by gravity currents originating from the LEF. We have also seen in [section 3.3d](#) (**Figure 3.9b**) that wavelengths can reach 1500 m, or even higher values. The Taylor-Goldstein **equation (3.1a)** is rather sensitive to internal wavelengths, and we next briefly describe phase speed mismatches between the numerical model used in this work, the Taylor-Goldstein **equation (3.1a)** for different wavelengths, and observations.

The phase speed derived from **equation 3.1a** is based on assuming a wavelength of 600 m, as observed in [section 3.3a](#), case 2 and on climatological ambient conditions. Rounded to one significant figure, the resulting phase speed is identical to the observation (-0.5 m s^{-1} in both cases). This suggests that the ISWs are in fact characterized by rather small amplitudes

that make them behave more or less like linear waves. Otherwise, given the dependency of the phase speed on wave amplitude for ISWs, one would expect the linear phase speed to be significantly smaller than the observed phase speed. It's either this or that the ISWs have indeed large amplitudes but then the climatological ambient conditions do not reflect well the instantaneous conditions.

The mismatch between the observation and the numerical model for the phase speed (-0.5 m s^{-1} vs. -1.5 m s^{-1} , respectively) and for the wavelength (600 m vs. 1500 m) is also presumably indicative that the ambient conditions used for model initialization, taken as representative of November 2008, do not represent well the conditions of June 2020 (i.e. Case study 2). For example, reducing the depth of the pycnocline from $h_1 = 100 \text{ m}$ to $h_1 = 40 \text{ m}$ (see **Table 3.1** for details) reduces the phase speed from -1.5 m s^{-1} to -1.0 m s^{-1} and the wavelength to about 1000 m. Other parameter combinations could well lead to the desired values of observed phase speed and wavelength, but it would require a more systematic numerical study to fully explore the sensitivity of the results to the parameter space. Nevertheless, for the purpose of this observational study, the numerical model used here provides a reasonable physically-based ground to propose a general mechanism for the origin of the observed ISWs keeping in mind that the details may not be accurate.

Together with the buoyant gravity currents previously known (WEA; Archer et al. 1997; Johnson 1996; Yoder et al. 1994), which are strongly dissipative, these newly found ISWs may be relevant in the energy budget and turbulent dissipation in the upper equatorial Pacific. As quantified by WEA from shear probes on Chameleon, turbulent dissipation rates of fronts associated with gravity currents were 1000 times greater within the warm and buoyant gravity current layer, compared to that in the cold tongue water underneath. The ISWs reported in this paper may add considerable dissipation into even smaller scales.

d) Broader scope

The existence of the cold tongue in the Equatorial Pacific requires a balance between the intense solar heating, the upwelling of cold waters from Ekman divergence near the surface, and the subsurface mixing at the base of the thermocline with colder middle/lower thermocline water. The latter results in cooling of the upper thermocline, and subsequent cooling of surface waters in the cold tongue. While Liu et al. (2016) suggest that many of the mixing events within the cold tongue are caused by lower Richardson numbers resulting from TIWs' shear instabilities and the EUC, some of the mixing at the upper layer of the thermocline may originate from the deep cycle layer. The "deep cycle" layer is a 24-hour cycle of band-limited vertical turbulent motions that extend below the mixed layer at night. It mixes stratified waters generated by intense solar heating during daytime, together inducing a surface buoyancy flux (see e.g. Moum et al. 2009; Moum et al. 2022a; Moum et al. 2022b). Holmes and Thomas (2015) revealed that TIWs modulate the turbulent heat flux estimated by Moum et al. (2009). Because of such shear driven turbulence, stratification tends to weaken. Vertical mixing within the EUC core is also modulated by TIWs (Moum et al. 2009). The former study reveals that TIW horizontal strain may alter the zonal shear over the

EUC core through horizontal vortex stretching, and accounts for some cooling implied in the turbulent heat flux estimated in the latter study. In this paper we add another element to the energy cascade from large to small scales and mixing: the generation of long-crested (typically 300 km) ISWs by relatively long-lived buoyant gravity currents, themselves originating from TIW dynamics (reported by WEA, and references therein). Travelling along the thermocline in the Equatorial Pacific cold tongue, these ISWs are subject to strong vertical velocity shears, leading to conditions of “marginal instability” or “marginally stability” (Miles 1961). Consequently, these ISWs may be prone to transient growth instabilities, enhancing mixing in a similar manner as K-H instabilities and wave breaking measured over the continental shelf (Moum et al. 2003). Some of the satellite images analyzed in this study (not shown) are consistent with shorter-scale wave-like features within large ISWs. Passaggia et al. (2018) argue that transient growth instabilities formed in the solitary waves may grow nonlinearly into “*optimal disturbance packets*”. These packets may in turn generate large K-H instabilities on the forward leaning surface of an ISW that break into turbulence in the rear face of the ISW and downstream of the wave (Carr et al. 2017). All these processes, including the strong dissipation rates associated with the gravity currents measured by WEA, may in the end impact El Niño Southern Oscillation (ENSO) irregularity and predictability (Holmes et al. 2019; Warner and Moum 2019). The above-mentioned measurements, interpretations and our speculations suggest a complex relationship between very different time and space scales. The instabilities’ spatial scales range over at least three orders of magnitude, from TIWs of typically 10^3 km to horizontal scales less than 1 km within ISWs. This relationship would in turn influence coupled ocean-atmosphere phenomena ranging over seasonal and interannual timescales within the equatorial Pacific, possibly with global-scale ramifications.

3.5. Summary and Conclusion

We present a hypothesis for the generation mechanism of trains of ISWs that are observed in the equatorial Pacific Ocean, within a zonal band from 5°S to 5°N; these waves were partly catalogued in a world map of ISWs observed by satellite (Jackson 2007), but to date had not been explained. The presence of these ISWs in this zonal band from 210°E to 265°E is remarkable given the absence of steep bottom topography (and tides) that could explain their presence from classical theories (Egbert and Erofeeva 2002). In this study, we provide estimates and model-based predictions of some ISW characteristics such as their wavelength, crest length, typical amplitude, as well as age. In total, we identified 116 ISW trains during one full year (2020), with an average crest length of 300 km, typical wavelengths of 1500 m, and modelled amplitude of 10 m. The origin of these trains of ISWs is consistent with a generation mechanism that involves buoyant gravity currents with sharp fronts detectable in satellite imagery (visible and thermal infrared from Sentinel-3). The propagation of gravity currents into stably-stratified environments is known to generate internal waves (e.g. Nash and Moum 2005; Bourgault et al. 2016; Wang et al. 2019; Mendes et al. 2021), but reports of such phenomena in the open ocean are scarce, if existent at all (WEA; Lavrova and Mityagina 2017). Here we report indications that such internal wave

generation mechanism is ubiquitous along the equatorial Pacific, and we suggest that ISWs may have been overlooked in many other frontal regions in the deep ocean. Energetic TIWs have been linked to buoyant gravity currents in WEA, which propagate within the equatorial cold tongue that is strongly stratified, hence providing conditions for the propagation of ISWs initiated by the gravity currents.

Although ISW generation by gravity currents has previously been simulated for specific conditions (White and Helfrich 2008, 2012), our knowledge of the process is insufficient in open-ocean environments. As such, the use of a numerical model has been helpful for interpreting and complementing the incomplete satellite observations, as well as for making physically-based predictions about the properties of the internal wave field. The model appears to simulate the generation of ISWs by resonant interactions under the surface front of the gravity current. The wave amplitude and propagation speed increase through nonlinear dispersion. The wave is subsequently released from the front location when its phase speed significantly exceeds the front speed. If the front propagation is steady (for some period of time at least) the generation sequence is repeated after the complete detachment of the preceding solitary wave. In this process, energy is transferred from the gravity current to ISW trains. The model results are consistent with the satellite observations of internal wave wavelength and phase speed, and the steady propagation of sharp fronts. Energy considerations based on model runs, themselves based on measurements by WEA, show that there is incomplete conversion of potential energy into kinetic energy near the gravity current front. This result is consistent with findings by WEA, which reported that dissipative losses occur owing to turbulence within the gravity current forehead. This potential energy may also transfer to internal waves generated at the front.

Similar ISW trains emanating from sharp fronts were observed in satellite images from 2019 to 2022. While 2019 was an El Niño year, 2020 and 2021 were La Niña years. We expect the surface manifestations of the phenomena reported in this paper to be more visible during La Niña years because thermal fronts are stronger, with a more sharply bound equatorial cold tongue. Perhaps it is no coincidence that the satellite (MODIS) observations reported in Jackson (2007) for the period August-2002 to May-2004 (an El-Niño period) extend less to the west and are fewer than our own observations shown in **Figure 3.2**. Furthermore, it is stressed that satellite observations of the sharp fronts and ISWs are compromised during part of the year because of severe cloud cover (typically after September). For that reason, we avoid a seasonal characterization of the ISW features in this study.

The association reported here of sharp ocean fronts and gravity currents with the generation of large scale ISWs in the open ocean may turn out to be more common than previously thought. New observation techniques with finer resolutions, either by means of remote sensing or field programs, may uncover new insights and quantifications of the cascade of turbulent energy from large to small and smaller scales. For this reason, it is suggested that a comprehensive oceanographic campaign, conducted in the study region, may clarify the issues discussed in this paper. Furthermore, the development of a more complete nonhydrostatic and nonlinear numerical model that could simulate the entire

sequence from the large-scale TIWs formation, to gravity currents and down to fine-scale ISWs would become an invaluable tool to provide a more comprehensive understanding of these complex interacting phenomena.

Author Contributions: Conceptualization, Adriana M. Santos-Ferreira (A.M.S.-F.) and José Carlos Bastos da Silva (J.C.B.d.S.); Data curation, A.M.S.-F.; Formal analysis, A.M.S.-F.; Investigation, A.M.S.-F.; Methodology, A.M.S.-F. and J.C.B.d.S.; Numerical modelling, Bruno St. Denis (B S.-D.) and Daniel Bourgault (D.B.); Supervision, J.C.B.d.S. and Leo Maas (L.M.); Visualization, A.M.S.-F.; Writing of original draft, A.M.S.-F. and J.C.B.d.S.; Writing—review and editing, A.M.S.-F., L.M., D.B. All authors have read and agreed to the published version of the manuscript.

CHAPTER 4

CONCLUSION

4.1. Concluding remarks

Surface wave breaking caused by propagating ISWs in the deep ocean is reported in [Chapter 2](#), with S3 satellite observations that demonstrate how these large-scale ISWs are manifested in the SAR altimeter by means of their effects in Significant Wave Height (SWH). In more detail, unfocused SAR (UF-SAR 20 Hz) and fully focused SAR (FF-SAR 160 Hz) modes were analyzed, which were retracked in full range and over a reduced range of bins (truncation carried out dynamically ten gates away from the estimated epoch position). In both these modes ISWs display a clear signature at small along-track scales (1-3 km), as already demonstrated in [Santos-Ferreira et al. \(2018\)](#), but most importantly, a second order SWH signature on scales of 20 km: the SWH is attenuated downstream (or after the passage) of an ISW, considering length scales of about ± 10 km, i.e. upstream and downstream (or, before and after the passage) of the ISW leading crest.

This is a newly reported result in radar altimetry, which is validated owing to exact synergy with the OLCI instrument in the S3 mission that allows unambiguous detection of ISW crests. The interpretation of the wave breaking associated with the ISWs, which was based on MSI onboard S2, may be explained by resonant interaction between ISWs and surface waves, a condition that occurs in deep waters typically for meter-scale surface waves. When the phase speeds of ISWs are close or equal to the group speeds of surface waves, energy can be effectively transferred between the ISWs and the surface waves with meter-scales, causing strong amplification of the surface waves leading to eventual breaking, and consequently dissipating significant amounts of energy. This energy dissipation produces turbulence and mixing in the wake (i.e. downstream) of the ISW crest and may dampen not only the amplitude of meter-scale surface waves (observed in the SWH record), but also small-scale wind ripples (and hence observed in the σ_0 record of a radar altimeter). Both these signatures are clearly seen in [Figures P2.1b,c](#) of the prologue to [Chapter 2](#). This latter effect, seen in part b) of [Figure P2.1](#), is consistent with laboratory experiments in a wave tank ([Ermakov et al. 2020](#)).

The outcome of synergetic observations with Sentinel-3 (SAR altimeter and OLCI) and Sentinel-2 (MSI) reveals that the breaking of meter-scale surface waves owing to (resonant) interactions with large-amplitude ISWs plays a significant role in the dynamics of the upper ocean, meaning exchange of momentum, heat, and gases between the atmosphere and the ocean. The general and specific objectives in part 1 of this thesis (see [Figure 1.8](#) in [section 1.4](#)) were hence addressed successfully, and the answer to the question “if the new generation satellite sensors of the Copernicus program could improve our knowledge and detection capabilities of ISWs in deep ocean?”, is therefore a clear “yes”.

In the second part of the thesis, a hypothesis that sharp ocean fronts and associated gravity currents (revealed in Warner et al. 2018) observed within the Pacific cold tongue could generate ISWs, which propagate upstream of the gravity currents, is tested by means of Sentinel-3 observations (OLCI and SLSTR) and a numerical model. This was proposed in objective 2 of the general and specific objectives in [section 1.4](#) (see also **Figure 1.8**). Before continuing, a note is warranted to explain why here we did not use the SAR altimeter to document the ISWs. Early on in the investigation, it became apparent that the ISWs in the study region were not clearly seen in the SAR altimeter record. Later, after the model analysis, the reason became more clear: the amplitudes of the ISWs within the cold tongue were estimated to be about 10 m, which are at the lower limit of detection of the SAR altimeter sea level anomalies in comparison with noise in that region. Indeed, in **Figure P3.2** of the prologue to [Chapter 3](#), the reader can verify that the ISWs practically exhibit no signature both in SLA and in σ_0 (with the remarkable exception of two small peaks in σ_0 at the location of the ISWs visible in OLCI quasi-true color images, in **Figure P3.2d**).

The view defended in this thesis, and indeed proposed in Santos-Ferreira et al. (2023), is that the satellite observations are consistent with a transcritical regime where the ISWs are generated by resonant transfer of energy from the buoyant gravity current moving at the surface with a certain phase velocity. As proposed in White and Helfrich (2012), an analogy can be made with upstream resonance in two-layer flow over variable topography, where gravity currents “act as an obstacle to the flow” and would be as effective to generate ISWs upstream as a topographic obstacle. Interestingly, **Figure 1.4** presented in [section 1.2](#) has also some similarity to the analogy described above: in that case a boat would act as the obstacle, which, moving at a speed within the transcritical regime, generates the surface solitary wave that is seen in the picture (**Figure 1.4**). And why do we say that the satellite observations are consistent with this view? Because the ISW trains are always visible propagating upstream of the SSFs and the observed gravity currents (in SLSTR), with no exception. Although this is not a conclusive proof of the generation mechanism, it is consistent with existing theory (White and Helfrich 2012). Despite being not frequently reported in the ocean, particularly in deep waters, the above mentioned generation mechanism is more common in coastal waters (e.g. Nash and Moum 2005; Mendes et al. 2021; Wang et al. 2019), and even more frequently reported in the atmosphere (e.g. the morning glory waves in the Gulf of Carpentaria in Rottman and Simpson 1989; da Silva and Magalhaes 2009).

While the general and specific objectives have been addressed consistently and convincingly in this second part of the thesis (see [section 1.4](#) and **Figure 1.8**), the answer to the question that was formulated there is subtler. In Effect, Sentinel-3 has been essential to address this problem because of the exact synergy between OLCI and SLSTR. Nevertheless, one could argue that ESA’s ENVISAT mission also had visible/thermal sensors operating at similar spatial and spectral resolutions, and exact synergy. Hence, one could ask another question: “*why similar work to this was not developed by the time of ENVISAT mission, 2002 – 2012?*” The reason is perhaps twofold: 1) The quantity of image data that were necessary to visualize is very large, but nowadays more easily accessed with *online* tools such as in Ocean

Virtual Laboratory and 2) The existence of gravity currents in the study region was not documented by that time, as it had to wait for Warner et al. (2018). It was this latter research work that provided the foundations of our hypothesis. Furthermore, by the time of the ENVISAT mission the theoretical foundations for the generation mechanism of ISWs by gravity currents (White and Helfrich 2012), were not yet finished.

Moreover, there was no funding available to conduct a satellite investigation of the required magnitude, which was only possible owing to the FCT's PhD Grant awarded to the author of this thesis.

4.2. Wider Significance of Chapter 2

Chapter 2 addressed surface wave breaking caused by oceanic ISWs and some of the surface manifestations in the SAR altimeter onboard the Sentinel-3A and -3B satellites by means of their effects on Significant Wave Height (SWH). From a technical viewpoint it was concluded that both unfocused SAR (UFSAR) and fully focused SAR (FFSAR) modes are capable of detecting perturbations in SWH owing to surface wave breaking. The prospect of future SAR altimeter missions with FFSAR capabilities, combined with techniques such as retracking over a reduced range of bins (truncation carried out dynamically some number of gates away from the estimated epoch position), may improve further the sensitivity of the SWH signal as well as of other SAR altimeter parameters.

From a wider scientific perspective, surface wave breaking owing to ISWs add to the turbulent mixing and dissipation caused by breaking internal waves (deeper in the pycnocline), both phenomena driving vertical transport of heat, carbon dioxide, and other climatologically important tracers in the ocean and at the air-sea interface. Consequently, ISWs play an important role in shaping the ocean circulation and distributions of heat and carbon within the climate system (see e.g., the review in Whalen et al. 2020). The relative importance of surface wave breaking due to large ISWs in the deep ocean remains to be accounted for Satellite synergy (namely SAR altimeters and high-resolution visible imagery) is expected to enhance our knowledge of this phenomenon.

4.3. Wider Significance of Chapter 3

Oceanic thermal anomalies closely related to the El Niño Southern Oscillation (ENSO) include maxima in SST in the eastern and central equatorial Pacific that suppress or completely obliterate a phenomenon known as the Equatorial Cold Tongue (ECT). The ECT is an immense “band” of colder surface water when compared to tropical warmer water to the south and north of it that extends along the equator, spanning from roughly 270°E to 210°E. Although being “cold” in comparison to adjacent surface waters, it is characterized by a net heat uptake from the atmosphere. The simplest and traditional explanation of the elevated El Niño's SSTs along the eastern and central Pacific waters is attributed to anomalous weakening of the trade winds (of the Southern Hemisphere) with inherent weakening of the equatorial upwelling (Bjerknes 1969). This allows the formation of a deeper

thermocline in the eastern Pacific, SST warming and the onset of an El Niño event. Nevertheless, recent research has been showing that there may be a lot more to this story (Moum et al. 2013; Moum et al. 2022a).

While SST over the rest of the global ocean has warmed in response to global atmospheric heating (including the anthropogenic component), the Pacific cold tongue shows no significant warming, and it may even have cooled during the period 2000 – 2012, known as “the hiatus” in atmospheric warming (Seager et al. 2019). In the ECT of the Pacific, maintaining cool SSTs in the presence of intense solar heating requires a combination of subsurface cooling and vertical advection of heat transporting surface heat downwards. In other words, since there is a net heat gain from above the surface, the heat must go somewhere deeper so that this ECT is colder than surrounding tropical waters.

Ocean mixing at the base of the thermocline, where warm waters from the surface layers are mixed with colder layers from below the thermocline occurs because of current shear (see e.g. Moum et al. 2013; Liu et al. 2016). The equatorial current system is a highly sheared environment where surface currents, namely the South Equatorial Current (SEC), normally flows westwards but gradually change direction downwards until it reverses completely when it reaches the depth of the Equatorial Under Current (EUC), which is a strong narrow undercurrent whose core lies at a typical depth of 100 m and flows eastward practically all along the equatorial Pacific. Such vertical shear is very efficient in mixing water in the vertical even in the presence of strong stratification, as is the case within the cold tongue (Warner and Moum 2019), and hence, cool water from below the thermocline encounters warmer surface layer water, causing mixing and cooling it down.

The question here is then: “*What do ISWs have to do with this mixing, and in what way may they affect the climate?*”. ISWs also produce strong shear near the thermocline (see e.g. Osborne and Burch 1980), and as discussed above, this shear adds to an already strongly sheared environment near the base of the thermocline, but also within the core of the EUC. The environment where these ISWs propagate is characterized by strong stratified waters, but because of the vertical shear owing to the strong EUC flowing underneath the SEC, the system is close to “marginal instability” (Miles 1961; Smyth et al. 2013). The ISWs studied in [Chapter 3](#) may tip shear even further to become unstable, causing overturns and hence stronger mixing.

Warner and Moum (2019) have shown that mixing – or lack thereof – contributes to cooling the sea surface during transitions to La Niña events or heating the sea surface during transitions to peak El Niños. Mixing therefore acts as a positive feedback mechanism to the development of El Niños and La Niñas. The ISWs studied in [Chapter 3](#) are likely to have a significant impact in those feedback mechanisms, especially in cooling the sea surface during transitions to La Niña years, as ISWs appear more frequently during La Niña, when the ECT is fully developed during a full year (Santos-Ferreira et al. 2023). It may therefore be speculated that the ISWs studied here may enhance heat uptake from the atmosphere to depths below the thermocline, where this heat may then be an important element in ocean circulation and climate variability.

So, putting the question in another way: *“why would this additional exchange mechanism be good to establish a milder climate, in particular to slow down global warming?”* The new findings in Santos-Ferreira et al. (2023) add to the knowledge of mixing owing to ISWs within the cold tongue, and thus to vertical heat transfer in a critical region of the ocean for climate, which may motivate further research to understand better how the uptake of heat from the atmosphere within the cold tongue may slow down (at least for some period) anthropogenic climate change, such as global warming.

The broader scope addressed in this section is nicely summarized in Warner and Moun (2019) who propose a feedback mechanism for ENSO in which subsurface mixing reduces equatorial SST by cooling from below, thereby strengthening atmospheric stability and enhancing further the high air pressure in the eastern Pacific and causing a larger zonal pressure gradient along the equator. This in turn strengthens easterly winds (Bjerknes 1969), and enhance zonal surface currents (SEC) and upper ocean shear, bringing the current system nearer to instability, and thus reinforcing mixing. We add another important element to this feedback mechanism: ISWs further enhance mixing because they are prone to transient growth instabilities owing to overturns and wave breaking (Passaggia et al. 2018), hence entering the feedback cycle. All these processes may impact ENSO variability and predictability (Holmes et al. 2019).

4.4. Future Work

As suggested in the final section of Chapter 3, and published in Santos-Ferreira et al. (2023), new observation techniques with finer resolutions, either by means of remote sensing or field programs, may uncover more details of the internal waves studied within the Pacific Cold Tongue and bring new insights into the cascade of turbulent energy from large to smaller scales. Interestingly, just one month after Santos-Ferreira et al. (2023) was published, the author of this thesis found by chance that there is a project to study internal waves in the equatorial Pacific funded by the National Science Foundation of the United States. According to the Investigators of that project, the general aim is to study turbulent mixing processes near the equator (within about $\pm 5^\circ$ from the equator, but mostly near the longitude of 220°E) that may be associated with TIW fronts through their potential for creating or modifying internal waves (see <https://www.mod.ucsd.edu/tropical-instability-waves>). This follows the remote sensing and modelling findings addressed in Chapter 3. The team of investigators of this project, led by Caitlin Whalen of the University of Washington (Applied Physics Laboratory), are using a combination of yearlong moored observations and two dedicated cruises which will certainly increase the knowledge of these processes. Although their chosen location for in situ data collection at 220°E is not where we found a maximum number of ISWs in our published study (see Figure 3.2 and Figure 4.1), there are Tropical Atmosphere Ocean (TAO) buoys that can complement measurements in their project. The instruments at this location (0° , 220°E) have been deployed continuously since September 2005, and the location has probably been preferred because this is the only record long

enough to begin an assessment of interannual variability in studies of El Niño and La Niña (see Warner and Moum 2019). The reason why there are less ISWs observations at 220° is probably related to the cold tongue getting “weaker” to the west, in comparison to eastern longitudes. When their project is concluded, we anticipate a better understanding of the phenomena described in Chapter 3. It may even be possible to confirm or dismiss the generation mechanism proposed in this thesis for the ISWs observed with Sentinel-3.

Furthermore, it may be possible that radar observations of the equatorial cold tongue (Sentinel-1 SAR vignettes and/or Surface Water and Ocean Topography Ka-band Radar Interferometer - SWOT KaRIn σ_0 images) reveal more details of the internal waves discussed in Chapter 3, since radar is little affected by clouds and operates at night time.

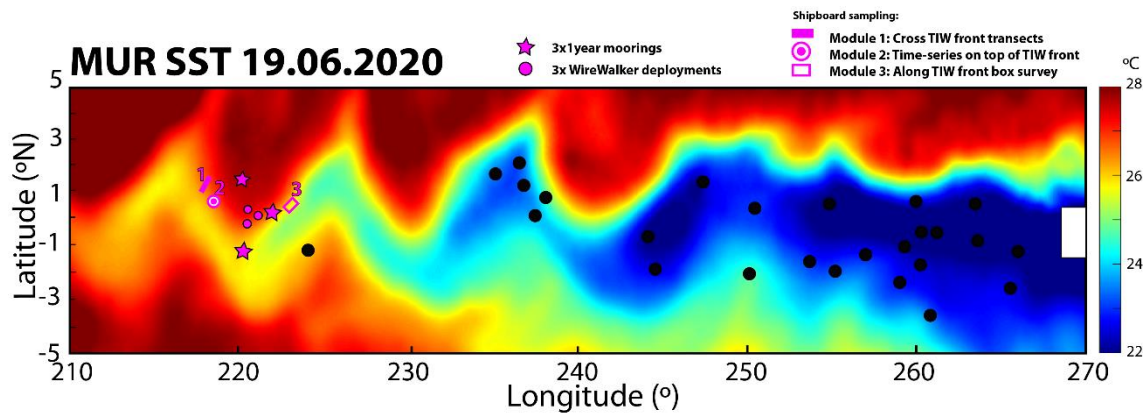


Figure 4.1 - Multiscale Ultrahigh Resolution (MUR) SST image (dated 19 June 2020) in the equatorial Pacific band. The black dots are the ISWs identified in the visible imagery of Sentinel-3, between 15 June and 22 June 2020 (image adapted from Figure 3.6 of this thesis) and the locations of moorings and shipboard sampling (purple) described in the proposal that is available online at <https://www.mod.ucsd.edu/tropical-instability-waves>.

A brief search of ISWs in Sentinel-3 for the region of the Atlantic Cold Tongue, with the same methodology adopted in Chapter 3, showed that the Atlantic cold tongue also holds signatures of surface manifestations of ISWs in OLCI. **Figure 4.2** shows an example of a recent case of an OLCI/SLSTR synergy pair in the equator, dated 4 August 2023, where the ISW in the image seems to emerge from a sharp front. This suggests a similar generation mechanism to that proposed in Chapter 3 (and in Santos-Ferreira et al. 2023), and adds circumstantial evidence to the findings reported in this thesis. The matter disclosed in **Figure 4.2** would be equally relevant in climate studies. Indeed, many studies have shown that the Atlantic cold tongue is a region of climatic relevance for the surrounding continents (Okumura and Xie 2004; Hummels et al. 2013), with some similarities with the Pacific equatorial ocean. Furthermore, as well as in the Pacific cold tongue, there was only a slight small warming (if any) of the Atlantic cold tongue (see section 4.3 and Nnamchi et al. 2020) during the satellite observation era. Hence, it is proposed that similar studies to the one in this thesis within the Atlantic Cold Tongue may become beneficial to a better understanding of climate issues.

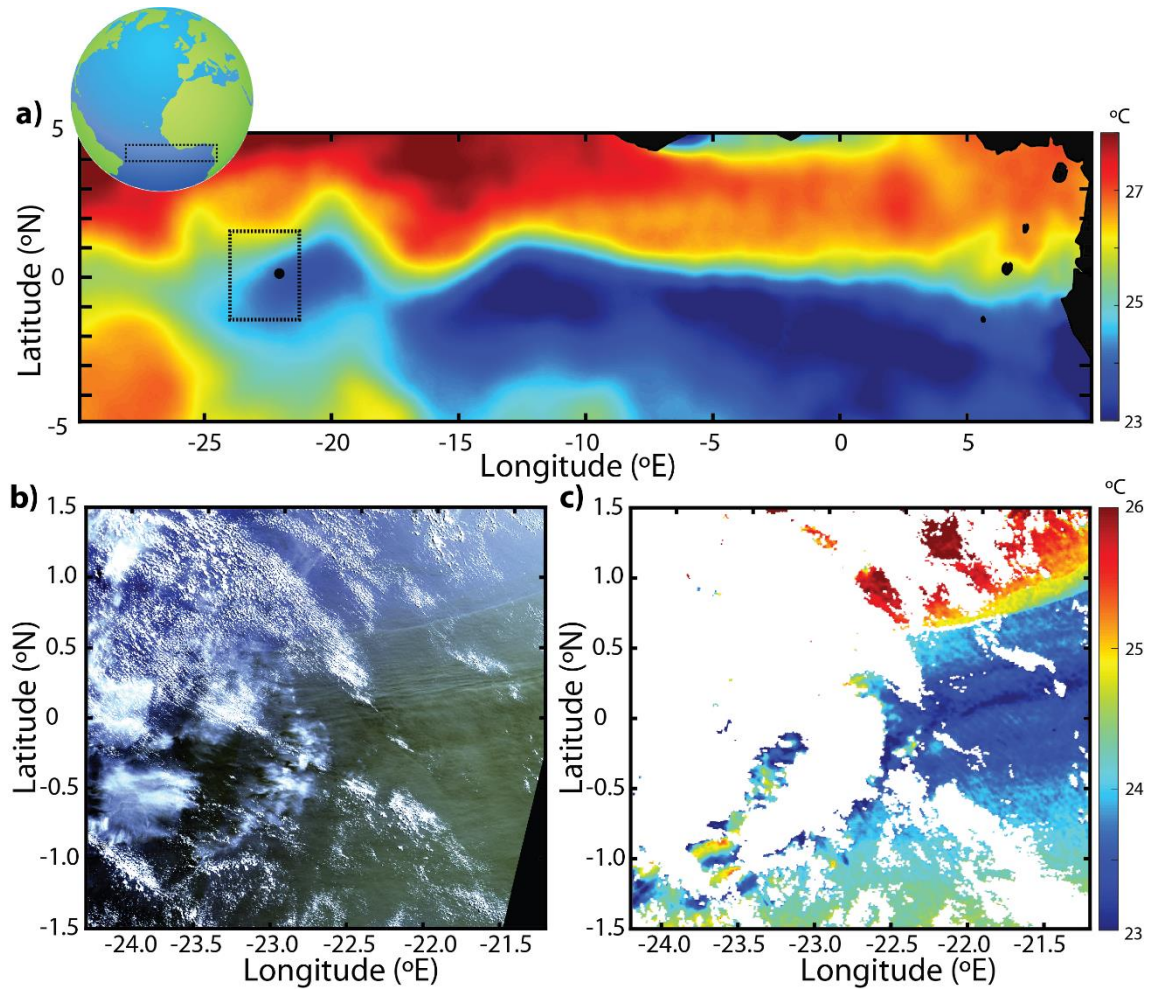


Figure 4.2 - a) MUR SST image (dated 04 August 2023) in the Atlantic equatorial ocean. The black dashed rectangle corresponds to the location of panels b and c and the black dot is the ISW identified in the visible imagery of Sentinel-3 (panel b)). **b)** Sentinel-3B visible image (Level 1b OLCI) acquired at 11:34:39 UTC (start time) presented in quasi-true colour. **c)** Sentinel-3B thermal infra-red image (Level 2 WST) simultaneous with panel b.

A note is warranted about the new satellite mission SWOT. This mission comprises wide swath altimetry technology, and covers most of the world's ocean with repeated high-resolution elevation measurements, providing “images” of the ocean's surface topography as well as radar backscatter σ_0 . Sea level height is measured by a wide-swath radar interferometer operating in Ka band. An innovative instrument called KaRIn developed by NASA's JPL), produces coincident images (in space and time) of SSH and radar power, which may be used not only to measure internal tidal waves, but also ISWs. This appears as a step forward for small scale ocean research and ISW detection.

The proposed project for this thesis initially aimed at establishing a method to estimate, within reasonable bounds of error, the upper layer current perturbations induced by ISWs (see IGARSS conference paper Santos-Ferreira and da Silva 2020). Surface elevation anomalies associated with ISWs are measurable in S3 (see Santos-Ferreira et al. 2018) and it is possible to retrieve the wave amplitude integrating the satellite's along-track surface elevation gradient, in a similar manner proposed in Pinkel (2000). Validation of this method

could have been possible in 2020, during a field campaign funded by international project AMAZOMIX, led by Dr. Ariane Koch-Larrouy from the French Laboratory of Space Geophysical and Oceanographic Studies (LEGOS) in collaboration with FCUP. However, due to the covid pandemic, the campaign was postponed several times and in the end, some results have not been shared at the time of writing. The proposed method allows determination of ISW amplitudes and surface currents, provided one has knowledge of the density stratification.

As the method could not have been validated with in situ data from AMAZOMIX, the author continued with the work presented in Chapter 2, as a contingency plan. Nevertheless, the new satellite mission SWOT can provide an alternative data set to cross-validate this methodology, if S3's SLAs can be co-located with KaRIn. **Figure 4.3** shows an image from KaRIn revealing ISWs in the Banda Sea (that were studied in Chapter 2) as “an image” of surface heights. The parameter shown is SLA and units are in cm, meaning that these ISWs can have surface amplitudes of 20-25 cm.

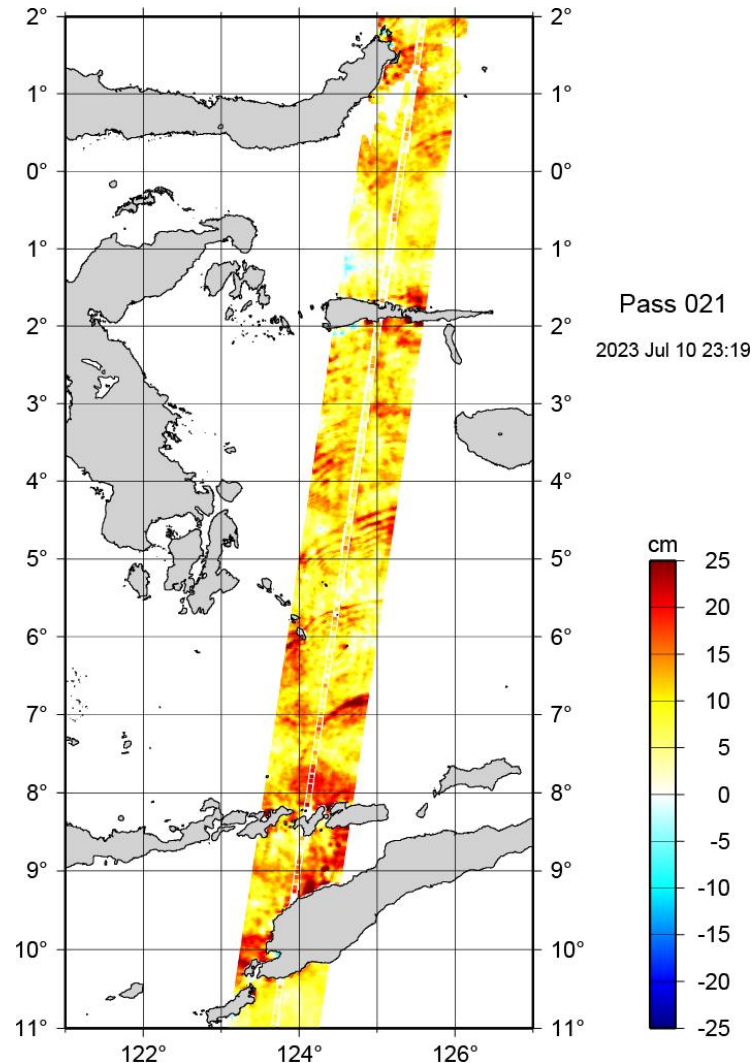


Figure 4.3 – SWOT KaRIN image showing SLAs, dated 10 July 2023 – PASS 021 of the 1-day repeat cycle orbit. Curtesy of Richard D. Ray.

4.5. Epilogue

During the review phase of Santos-Ferreira et al. (2023) concerns were raised about the longevity of the ISWs and gravity currents observed in the satellite images. In this concluding paragraph those concerns are addressed. In fact, since Sentinel-3 image data was used in Chapter 3, and we were using mainly OLCI visible bands to detect ISWs, we were restricted by a single overpass (image) per day in a given swath. Although we attempted to find a sequence of images “following” the same feature (ISW and/or SSF), this was only possible in one single occasion (for the year 2020), which is shown in Figure 3.4. The reason for such scarcity is due to severe cloud coverage year-round in the study region. Strictly speaking, only in the case of 15-16 June 2020 (Figures 3.4 and 3.5) can it be said that the observed ISW train lasted at least 24 hours. That *age* is natural for a train of “medium-sized” ISWs in the ocean, as it is well known that large ISWs trains can propagate for a few days in the deep ocean, without severe attenuation (see e.g. da Silva et al. 2011; Magalhaes et al. 2016). The controversy regarding their age arises because some of the ISWs in Chapter 3 have been observed propagating with a strong westward component nearly at the equator, where they would be opposing the fast-flowing underneath EUC. Admittedly, this fact would cause strong dissipation, as these waves might be unstable to shear instabilities above the core of Equatorial Undercurrent. In fact, Moum et al. (1992) and Hebert et al. (1992) measured interfacial internal waves at the equator with the same configuration (i.e. propagating westward over the core of the fast eastward-flowing EUC), and reported dissipation rates so high that, for a particular wave event, the total wave energy would have been dissipated within a single buoyancy period (order of one hour or less). While the average *age* of the waves in Chapter 3 was estimated to be 28 hours, independent reports suggest *ages* within an order of magnitude below this figure. Here it is pointed out that the waves measured in Moum et al. (1992) and Hebert et al. (1992) have different characteristics and origins. The waves reported in Moum et al. (1992) and Hebert et al. (1992) have much shorter wavelengths (150-250 m) and are presumably generated by the wind. In fact their waves were reported to exist mainly during the night (wave potential energy greater by a factor of 100 at night), and when there was a sustained westward wind forcing. The waves reported in those two papers, more than 30 years ago, had amplitudes of the order of one meter, with the exception of the single event reported in Hebert et al. (1992), which had the maximum peak-to-peak amplitude of approximately 20 m and occurred at 50 m depth. In view of the discussion above, we feel that Santos-Ferreira et al. (2023) presents and discusses different internal waves, and there should not be ambiguity between those two kinds of internal waves, with different characteristics and origins.

This thesis highlights the importance of remote sensing satellites to study the ocean, and the role of satellite sensors with high spatial resolution to observe small ocean scales. It has now been recognised that to comprehend ocean processes, mixing and transport, it is necessary to know and quantify the small scale ocean phenomena (e.g. Whalen et al. 2020; Mahadevan 2019). Nevertheless, the temporal resolutions provided by satellite missions, including synergies and tandem missions, are still not optimized to observe the ocean at

small scales, and are limited (with the exception of visible wavelengths) to surface observations. This underlines the need for more field campaigns and continuous acquisition of in situ data to be able to understand and quantify vertical mixing and the stratification dynamics of the ocean.

REFERENCES

- Alpers, W., 1985: Theory of radar imaging of internal waves. *Nature*, **314**, 245–247, <https://doi.org/10.1038/314245a0>.
- Alpers, W., and E. Salusti, 1983: Scylla and Charybdis observed from space. *J. Geophys. Res. Oceans*, **88**, 1800–1808, <https://doi.org/10.1029/JC088iC03p01800>.
- Apel, J. R., H. M. Byrne, J. R. Proni, and R. L. Charnell, 1975: Observations of oceanic internal and surface waves from the Earth Resources Technology Satellite. *J. Geophys. Res.*, **80**(6), 865–881, <https://doi.org/10.1029/jc080i006p00865>.
- Apel, J. R., and F. I. Gonzalez, 1983: Nonlinear features of internal waves off Baja California as observed from the SEASAT imaging radar. *J. Geophys. Res. Oceans*, **88**(C7), 4459–4466, <https://doi.org/10.1029/JC088iC07p04459>.
- Archer, D., J. Aiken, W. Balch, D. Barber, J. Dunne, P. Flament, and Coauthors, 1997: A meeting place of great ocean currents: shipboard observations of a convergent front at 2° N in the Pacific. *Deep-Sea Res. II: Top. Stud. Oceanogr.*, **44**, 1827–1849, [https://doi.org/10.1016/S0967-0645\(97\)00031-3](https://doi.org/10.1016/S0967-0645(97)00031-3).
- Ardhuin, F., S. T. Gille, D. Menemenlis, C. B. Rocha, N. Rasche, B. Chapron, J. Gula, and J. Molemaker, 2017: Small-scale open ocean currents have large effects on wind wave heights. *J. Geophys. Res. Oceans*, **122**, 4500–4517, <https://doi.org/10.1002/2016JC012413>.
- Babanin, A., 2011: *Breaking and Dissipation of Ocean Surface Waves*. Cambridge University Press, <https://doi.org/10.1017/CBO9780511736162>.
- Behringer D. W., and Xue Y., 2004: Evaluation of the global ocean data assimilation system at NCEP: The Pacific Ocean. Eighth symposium on integrated observing and assimilation systems for atmosphere, oceans, and land surface, *AMS 84th annual meeting*, Washington State Convention and Trade Center, Seattle, Washington, 11–15.
- Bjerknes, J., 1969: Atmospheric teleconnections from the equatorial Pacific. *Mon. Weather Rev.*, **97**(3), 163–172, [https://doi.org/10.1175/1520-0493\(1969\)097<0163:ATFTEP>2.3.CO;2](https://doi.org/10.1175/1520-0493(1969)097<0163:ATFTEP>2.3.CO;2).
- Boegman, L., and M. Stastna, 2019: Sediment resuspension and transport by internal solitary waves. *Annu. Rev. Fluid Mech.*, **51**, 129–154, <https://doi.org/10.1146/annurev-fluid-122316-045049>.
- Bourgault, D., and D. E. Kelley, 2004: A laterally averaged nonhydrostatic ocean model. *J. Atmos. Ocean. Technol.*, **21**, 1910–1924, <https://doi.org/10.1175/JTECH-1674.1>.

- Bourgault, D., P. Galbraith, and C. Chavanne, 2016: Generation of internal solitary waves by frontally forced intrusions in geophysical flows. *Nat. Commun.*, **7**, 13606, <https://doi.org/10.1038/ncomms13606>.
- Boy, F., J. D. Desjonquères, N. Picot, T. Moreau, and M. Raynal, 2017 : CryoSat-2 SAR Mode Over Oceans: Processing Methods, Global Assessment, and Benefits. *IEEE Trans. Geosci. Remote Sens.*, **55**, 148-158, <https://doi.org/10.1109/TGRS.2016.2601958>.
- Brown, G., 1977: The average impulse response of a rough surface and its applications. *IEEE Trans. Antennas Propag.*, **25**, 67–74, <https://doi.org/10.1109/TAP.1977.1141536>.
- Buchhaupt, C., L. Fenoglio, M. Becker, and J. Kusche, 2021: Impact of vertical water particle motions on focused SAR altimetry. *Adv. Space Res.*, **68** (2), 853-874, <https://doi.org/10.1016/j.asr.2020.07.015>.
- Buchhaupt, C., 2019: Model Improvement for SAR Altimetry. Ph.D. Thesis, Schriftenreihe Fachrichtung Geodäsie der Technischen Universität Darmstadt.
- Caltabiano, A. C. V., I. S. Robinson, and L. P. Pezzi, 2005: Multi-year satellite observations of instability waves in the Tropical Atlantic Ocean. *Ocean Sci.*, **1**, 97-112, <https://doi.org/10.5194/os-1-97-2005>.
- Cai, W., B. Ng, T. Geng, F. Jia, L. Wu, G. Wang, Y. Liu, B. Gan, K. Yang, A. Santoso, X. Lin, Z. Li, Y. Liu, Y. Yang, F.-F. Jin, M. Collins, and M. J. McPhaden, 2023: Anthropogenic impacts on twentieth-century ENSO variability changes. *Nat. Rev. Earth Environ.*, **4**, 407-418, <https://doi.org/10.1038/s43017-023-00427-8>.
- Cao, H., Z. Jing, B. Fox-Kemper, T. Yan, and Y. Qi, 2019: Scale transition from geostrophic motions to internal waves in the northern South China Sea. *J. Geophys. Res. Oceans*, **124**, 9364-9383, <https://doi.org/10.1029/2019JC015575>.
- Carr, M., J. Franklin, S. E. King, P. A. Davies, J. Grue, and D. G. Dritschel, 2017: The characteristics of billows generated by internal solitary waves. *J. Fluid Mech.*, **812**, 541-577, <https://doi.org/10.1017/jfm.2016.823>.
- Chin, T. M., J. Vazquez-Cuervo, and E. M. Armstrong, 2017: A multi-scale high-resolution analysis of global sea surface temperature. *Remote Sens. Environ.*, **200**, 154-169, <https://doi.org/10.1016/j.rse.2017.07.029>.
- Cochran, J. K., H. J. Bokuniewicz, and P. L. Yager, 2019: *Encyclopedia of ocean sciences*. Academic Press, ISBN: 9780128130827.
- Craig, W., P. Guyenne, and C. Sulem, 2012: The surface signature of internal waves. *J. Fluid Mech.*, **710**, 277–303, <https://doi.org/10.1017/jfm.2012.364>.

- Da Silva, J. C. B., S. A. Ermakov, I. S. Robinson, D. R. G. Jeans, and S. V. Kijashko, 1998: Role of surface films in ERS SAR signatures of internal waves on the shelf: 1. Short-period internal waves. *J. Geophys. Res. Oceans*, **103**, 8009–8031, <https://doi.org/10.1029/97JC02725>.
- Da Silva, J. C. B., S. A. Ermakov, and I. S. Robinson, 2000: Role of surface films in ERS SAR signatures of internal waves on the shelf. 3. Mode transitions. *J. Geophys. Res.*, **105** (C10), 24089–24104, <http://dx.doi.org/10.1029/2000JC900053>.
- Da Silva, J. C., and J. M. Magalhaes, 2009: Satellite observations of large atmospheric gravity waves in the Mozambique Channel. *Int. J. Remote Sens.*, **30**(5), 1161–1182, <https://doi.org/10.1080/01431160802448943>.
- Da Silva, J. C. B., A. L. New, and J. Magalhaes, 2011: On the structure and propagation of internal solitary waves generated at the Mascarene Plateau in the Indian Ocean. *Deep Sea Res. Part I Oceanogr. Res. Pap.*, **58**, 229–240, <https://doi.org/10.1016/j.dsr.2010.12.003>.
- Da Silva, J. C. B., M. C. Buijsman, and J. Magalhaes, 2015: Internal waves on the upstream side of a large sill of the Mascarene Ridge: a comprehensive view of their generation mechanisms and evolution. *Deep-Sea Res. I: Oceanogr. Res. Pap.*, **99**, 87–104, <https://doi.org/10.1016/j.dsr.2015.01.002>.
- Dinardo, S., B. Lucas, and J. Benveniste, 2015: Sentinel-3 STM SAR ocean retracking algorithm and SAMOSA model. *IEEE Int. Geosci. Remote Sens. Symp. (IGARSS)*, Milan, Italy, 26–31 July 2015, 5320–5323, <https://doi.org/10.1109/IGARSS.2015.7327036>.
- Dinardo, S., L. Fenoglio-Marc, M. Becker, R. Scharroo, M. J. Fernandes, J. Staneva, and Coauthors, 2021: A RIP-based SAR retracker and its application in North East Atlantic with Sentinel-3. *Adv. Space Res.*, **68** (2), 892–929, <https://doi.org/10.1016/j.asr.2020.06.004>.
- Donlon, C., B. Berruti, A. Buongiorno, M. H. Ferreira, P. Féménias, J. Frerick, and Coauthors, 2012: The Global Monitoring for Environment and Security (GMES) Sentinel-3 mission. *Remote Sens. Environ.*, **120**, 37–57, <https://doi.org/10.1016/j.rse.2011.07.024>.
- Drazin, R., and R. Kazanijian, 1983: Strategy Implementation: Organizing for Internal Diversification. *Proc. Acad. Manage.*, **1983**(1), 7–11. Briarcliff Manor, NY 10510: Academy of Management, <https://doi.org/10.5465/ambpp.1983.4976308>.
- Dutrieux, P., C. E. Menkes, J. Vialard, P. Flament, and B. Blanke, 2008: Lagrangian Study of Tropical Instability Vortices in the Atlantic. *J. Phys. Oceanogr.*, **38**, 400–417, <https://doi.org/10.1175/2007JPO3763.1>.
- Egido, A., and W. Smith, 2017: Fully Focused SAR Altimetry: Theory and Applications. *IEEE Trans. Geosci. Remote Sens.*, **55**, 392–406, <https://doi.org/10.1109/TGRS.2016.2607122>.

Egbert, G. D., and S. Y. Erofeeva, 2002: Efficient inverse modelling of barotropic ocean tides. *J. Ocean. Atmos. Technol.*, **19**, 183–204, [https://doi.org/10.1175/1520-0426\(2002\)019o0183:EIMOBO42.0.CO;2](https://doi.org/10.1175/1520-0426(2002)019o0183:EIMOBO42.0.CO;2).

Emery, W., and A. Camps, 2017: *Introduction to satellite remote sensing: atmosphere, ocean, land and cryosphere applications*. Elsevier, ISBN: 9780128092590.

Esaias, W. E., M. R. Abbott, I. Barton, O. B. Brown, J. W. Campbell, K. L. Carder, D. K. Clark, R. H. Evans, F. E. Hoge, H. R. Gordon, W. M. Balch, R. Letelier, and P. J. Minnett, 1998: An overview of MODIS capabilities for ocean science observations. *IEEE Trans. Geosci. Remote Sens.*, **36** (4), 1250–1265, <https://doi.org/10.1109/36.701076>.

Ermakov, S. A., V. A. Dobrokhotov, I. A. Sergievskaya, and I. A. Kapustin, 2020: Suppression of wind ripples and microwave backscattering due to turbulence generated by breaking surface waves. *Remote Sens.*, **12**(21), 3618, <https://doi.org/10.3390/rs12213618>.

Ferrari, R., A. Mashayek, T. J. McDougall, M. Nikurashin, and J. M. Campin, 2016: Turning ocean mixing upside down. *J. Phys. Oceanogr.*, **46**(7), 2239–2261, <https://doi.org/10.1175/JPO-D-15-0244.1>.

Frappart, F., D. Blumstein, A. Cazenave, G. Ramillien, F. Birol, R. Morrow, and F. Rémy, 1999: *Satellite altimetry: Principles and applications in Earth sciences*. In *Wiley Encyclopedia of Electrical and Electronics Engineering*. Wiley: Hoboken, NJ, USApp. 1–25.

Frouin, R., M. Schwindling, and P.Y. Deschamps, 1996: Spectral reflectance of sea foam in the visible and near-infrared: in situ measurements and remote sensing implications. *J. Geophys. Res. Oceans*, **101** (C6), 14361–14371, <https://doi.org/10.1029/96JC00629>.

Fu, L. -L., and B. Holt, 1982: *Seasat Views Oceans and Sea Ice with Synthetic-Aperture Radar*. California Institute of Technology, Jet Propulsion Laboratory, 81.

Garrett, C., and W. Munk, 1979: Internal waves in the ocean. *Annu. Rev. Fluid Mech.*, **11**(1), 339–369, <https://doi.org/10.1146/annurev.fl.11.010179.002011>.

Geng, T., F. Jia, W. Cai, L. Wu, B. Gan, Z. Jing, S. Li, and M. J. McPhaden, 2023: Increased occurrences of consecutive La Niña events under global warming. *Nature*, **619**(7971), 774–781, <https://doi.org/10.1038/s41586-023-06236-9>.

Gerkema, T., and J. T. F. Zimmerman, 2008: *An introduction to internal waves*. Lecture Notes, Royal NIOZ, Texel, 207, 207.

Guill, E. A., 1982: *Atmosphere – Ocean Dynamics*. International Geophysics Series, 30, 122, ISBN: 9780080570525.

Grimshaw, R., 2002: *Environmental stratified flows* (No. 3). Springer Science & Business Media, ISBN: 9781475776683.

Grimshaw, R., E. Pelinovsky, T. Taipova, and A. Sergeeva, 2010: Rogue internal waves in the ocean: Long wave model. *Eur. Phys. J. Spec. Top.*, **185**(1), 195-208, <https://doi.org/10.1140/epjst/e2010-01249-4>.

Hebert, D., D. R. Caldwell, C. A. Paulson, and J. N. Moum, 1992: Turbulence and internal waves at the equator. Part II: Details of a single event. *J. Phys. Oceanogr.*, **22**(11), 1346–1356, [https://doi.org/10.1175/1520-0485\(1992\)022<1346:TAIWAT>2.0.CO;2](https://doi.org/10.1175/1520-0485(1992)022<1346:TAIWAT>2.0.CO;2).

Helfrich, K. R., and W. K. Melville, 2006: Long nonlinear internal waves. *Annu. Rev. Fluid Mech.*, **38**, 395-425, <https://doi.org/10.1146/annurev.fluid.38.050304.092129>.

Holloway, G., 1983: A conjecture relating oceanic internal waves and small-scale processes. *Atmos. - Ocean*, **21**(1), 107-122, <https://doi.org/10.1080/07055900.1983.9649159>.

Holmes, R. M., and L. N. Thomas, 2015: The modulation of equatorial turbulence by tropical instability waves in a regional ocean model. *J. Phys. Oceanogr.*, **45**, 1155-1173, <https://doi.org/10.1175/JPO-D-14-0209.1>.

Holmes, R. M., S. McGregor, A. Santoso, and M. H. England, 2019: Contribution of tropical instability waves to ENSO irregularity. *Climate Dyn.*, **52**, 1837-1855, <http://dx.doi.org/10.1007/s00382-018-4217-0>.

Huang, X., Z. Chen, W. Zhao, Z. Zhang, C. Zhou, Q. Yang, and J. Tian, 2016: An extreme internal solitary wave event observed in the northern South China Sea. *Sci. Rep.*, **6**(1), 30041, <https://doi.org/10.1038/srep30041>.

Hummels, R., M. Dengler, and B. Boulès, 2013: Seasonal and regional variability of upper ocean diapycnal heat flux in the Atlantic cold tongue. *Prog. Oceanogr.*, **111**, 52-74, <https://doi.org/10.1016/j.pocean.2012.11.001>.

Jackson, C., 2007: Internal wave detection using the moderate resolution imaging spectroradiometer (MODIS). *J. Geophys. Res. Oceans*, **112**, <https://doi.org/10.1029/2007JC004220>.

Jackson, C., J. C. B. da Silva, and G. Jeans, 2012: The generation of nonlinear internal waves. *Oceanogr.*, **25**, 108-123, <https://doi.org/10.5670/oceanog.2012.46>.

Jeans, D. R. G., and T. J. Sherwin, 2001: The variability of strongly non-linear solitary internal waves observed during an upwelling season on the Portuguese shelf. *Cont. Shelf Res.*, **21**(16-17), 1855-1878, [https://doi.org/10.1016/S0278-4343\(01\)00026-7](https://doi.org/10.1016/S0278-4343(01)00026-7).

- Jensen, T. G., J. M. Magalhaes, H. W. Wijesekera, M. C. Buijsman, R. Helber, and J. Richman, 2020: Numerical modelling of tidally generated internal wave radiation from the Andaman Sea into the Bay of Bengal. *Deep Sea Res. Part II Top. Stud. Oceanogr.*, **172**, 104710, <https://doi.org/10.1016/j.dsr2.2019.104710>.
- Johnson, E. S., 1996: A convergent instability wave front in the central tropical Pacific. *Deep-Sea Res. II: Top. Stud. Oceanogr.*, **43**, 753-778, [http://doi.org/10.1016/0967-0645\(96\)00034-3](http://doi.org/10.1016/0967-0645(96)00034-3).
- Jones, N., 2022: Rare 'triple' La Niña climate event looks likely. *Nature*, **607**(21), 10-1038, <https://doi.org/10.1038/d41586-022-01668-1>.
- Joseph, G., 2005: *Fundamentals of remote sensing*. Universities Press, ISBN: 9788173715358.
- Kim, H., Y.-B. Son, and Y.-H. Jo, 2018: Hourly Observed internal waves by geostationary ocean color imagery in the East/Japan Sea. *J. Atmos. Ocean. Technol.*, **35**(3), 609-617, <https://doi.org/10.1175/JTECH-D-17-0049.1>.
- King, M. D., S. Platnick, W. P. Menzel, S. A. Ackerman, and P. A. Hubanks, 2013: Spatial and Temporal Distribution of Clouds Observed by MODIS Onboard the Terra and Aqua Satellites. *IEEE Trans. Geosci. Remote Sens.*, **51**, 3826-3852, <https://doi.org/10.1109/TGRS.2012.2227333>.
- Kodaira, T., T. Waseda, M. Miyaya, and W. Choi, 2016: Internal solitary waves in a two-fluid system with a free surface. *J. Fluid Mech.*, **804**, 201–223, <https://doi.org/10.1017/jfm.2016.510>.
- Koepke, P., 1984: Effective reflectance of oceanic whitecaps. *Appl. Opt.*, **23**, 1816–1824, <https://doi.org/10.1364/AO.23.001816>.
- Kubryakov, A. A., V. N. Kudryavtsev, and S. V. Stanichny, 2021: Application of Landsat imagery for the investigation of wavebreaking. *Remote Sens. Environ.*, **253**, 112144, <https://doi.org/10.1016/j.rse.2020.112144>.
- Kudryavtsev, V. N., D. Hauser, G. Caudal, and B. Chapron, 2003: A semi-empirical model of the normalized radar cross-section of the sea surface 1. Background model. *J. Geophys. Res. Oceans*, **108**, FET-2, <https://doi.org/10.1029/2001JC001003>.
- Kudryavtsev, V.; D. Akimov, J. A. Johannessen, and B. Chapron, 2005: On radar imaging of current features: 1. Model and comparison with observations. *J. Geophys. Res. Oceans*, **110**, C07016, <https://doi.org/10.1029/2004JC002505>.
- Kudryavtsev, V., A. Myasoedov, B. Chapron, J. A. Johannessen, and F. Collard, 2012: Imaging mesoscale upper ocean dynamics using synthetic aperture radar and optical data. *J. Geophys. Res. Oceans*, **117**, <https://doi.org/10.1029/2011JC007492>.

Kudryavtsev, V., B. Chapron, A. G. Myasoedov, F. Collard, and J. A. Johannessen, 2013: On dual co-polarized SAR measurements of the ocean surface. *IEEE Geosci. Remote. Sens.*, **10**, 761-765, <https://doi.org/10.1109/LGRS.2012.2222341>.

Kudryavtsev, V., I. Kozlov, B. Chapron, and J. A. Johannessen, 2014: Quad-polarization SAR features of ocean currents. *J. Geophys. Res. Oceans*, **119**, 6046-6065, <https://doi.org/10.1002/2014JC010173>.

Kudryavtsev, V., S. Fan, B. Zhang, A. Mouche, and B. Chapron, 2019: On Quad-Polarized SAR measurements of the ocean surface. *IEEE Trans. Geosci. Remote Sens.*, **57**, 8362-8370, <https://doi.org/10.1109/TGRS.2019.2920750>.

Lamb, K.G., 1997: Particle transport by nonbreaking, solitary internal waves. *J. Geophys. Res. Oceans*, **102**(C8), 18641-18660, <https://doi.org/10.1029/97JC00441>.

Lampitt, R.S., P. Favali, C. R. Barnes, M. J. Church, M. F. Cronin, K. L. Hill, and Coauthors, 2010: In situ sustained Eulerian observatories. *Proceedings of the OceanObs'09: Sustained Ocean Observations and Information for Society*, conference 2, **1**, 395-404, <https://doi.org/10.5270/OceanObs09.pp.27>.

Lavrova, O., and M. Mityagina, 2017: Satellite Survey of Internal Waves in the Black and Caspian Seas. *Remote Sens.*, **9**, 892, <https://doi.org/10.3390/rs9090892>.

Leamon, R. J., 2023: The triple-dip La Niña of 2020–22: updates to the correlation of ENSO with the termination of solar cycles. *Front. Earth Sci.*, **11**, 1204191, <https://doi.org/10.3389/feart.2023.1204191>.

Lentini, C. A., J. M. Magalhaes, J. C. da Silva, and J. A. Lorenzzetti, 2016: Transcritical flow and generation of internal solitary waves off the Amazon River: Synthetic aperture radar observations and interpretation. *Oceanogr.*, **29**(4), 187-195, <https://doi.org/10.5670/oceanog.2016.88>.

LeBlond, P. H., and L. A. Mysak, 1981: *Waves in the Ocean*. Elsevier, ISBN: 9780080879772.

Lewis, J. E., B. M. Lake, and D. R. S. Ko, 1974: On the interaction of internal waves and surface gravity waves. *J. Fluid Mech.*, **63**, 773–800, <https://doi.org/10.1017/S00222112074002199>.

Lian, Q., W. D. Smyth, and Z. Liu, 2020: Numerical computation of instabilities and internal waves from in situ measurements via the viscous Taylor–Goldstein problem. *J. Atmos. Ocean. Technol.*, **37**, 759-776, <https://doi.org/10.1175/JTECH-D-19-0155.1>.

Lim, Jae S., 1990: *Two-Dimensional Signal and Image Processing*. Englewood Cliffs, 548 (equations 9.26, 9.27, and 9.29).

Liu, A. K., C.-T. Liu, H. Ming-Kuang, Q. Zheng, C.-R. Ho, and Coauthors, 2008: *Satellite Remote Sensing of South China Sea*. Edition: 1st. Publisher: Tingmao Publish Company Editors: Antony K. Liu, Chung-Ru Ho and Cho-Teng Liu. ISBN: 978-986-122-948-5.

Liu, C., A. Köhl, Z. Liu, F. Wang, and D. Stammer, 2016: Deep-reaching thermocline mixing in the equatorial pacific cold tongue. *Nat. Commun.*, **7**, 1-11, <https://doi.org/10.1038/ncomms11576>.

Lyman, J. M., G. C. Johnson, and W.S. Kessler, 2007: Distinct 17- and 33-Day Tropical Instability Waves in Subsurface Observations. *J. Phys. Oceanogr.*, **37**, 855–872, <https://doi.org/10.1175/JPO3023.1>.

Lyzenga, D. R., and J. R. Bennett, 1988: Full-spectrum modeling of SAR internal wave signatures. *J. Geophys. Res.*, **93**, 12, 345–12,354, <https://doi.org/10.1029/JC093iC10p12345>.

Magalhaes, J. M., J. C. B. da Silva, M. Batista, L. Gostiaux, T. Gerkema, A. L. New, and D. R. G. Jeans, 2013: On the detectability of internal waves by an imaging lidar. *Geophys. Res. Lett.*, **40**, 3429– 3434, <https://doi.org/10.1002/glr.50669>.

Magalhães, J. M., J. C. B. da Silva, M. C. Buijsman, and C. A. E. Garcia, 2016: Effect of the North Equatorial Counter Current on the generation and propagation of internal solitary waves off the Amazon shelf (SAR observations). *Ocean Sci.*, **12**, 243-255, <https://doi.org/10.5194/os-12-243-2016>.

Magalhaes, J. M., and J. C. B. da Silva, 2017a: Satellite altimetry observations of large-scale internal solitary waves. *IEEE Trans. Geosci. Remote Sens.*, **14**, 534-538, <https://doi.org/10.1109/LGRS.2017.2655621>.

Magalhaes, J. M., and J. C. da Silva, 2017b: Internal waves along the malvinas current: Evidence of transcritical generation in satellite imagery. *Oceanogr.*, **30**(3), <https://doi.org/10.5670/oceanog.2017.319>.

Magalhaes, J. M., W. Alpers, A. M. Santos-Ferreira, and J. C. B. da Silva, 2021: Surface wave breaking caused by internal solitary waves: Effects on radar backscattering measured by SAR and radar altimeter. *Oceanogr.*, **34**, 166-176, <https://doi.org/10.5670/oceanog.2021.203>.

Magalhaes, J. M.; I. G. Lapa, A. M. Santos-Ferreira, J. C. B. da Silva, and Coauthors, 2023: Using a Tandem Flight Configuration between Sentinel-6 and Jason-3 to Compare SAR and Conventional Altimeters in Sea Surface Signatures of Internal Solitary Waves. *Remote Sens.*, **15**, 392, <https://doi.org/10.3390/rs15020392>.

Mahadevan A., 2019: *Submesoscale Processes*. J. K. Cochran, H. J. Bokuniewicz, P. L. Yager (Eds.) Encyclopedia of Ocean Sciences, 3rd Edition, **1**, 35-41, Elsevier, ISBN: 978-0-12-813081-0.

Maury, M.F., 1861: *The Physical Geography of the Sea and Its Meteorology*. Harper: New York, NY, USA.

Mendes, R., J. C. B. da Silva, J. M. Magalhaes, B. St-Denis, D. Bourgault, J. Pinto, and J. M. Dias, 2021: On the generation of internal waves by river plumes in subcritical initial conditions. *Sci. Rep.*, **11**, 1-12, <https://doi.org/10.1038/s41598-021-81464-5>.

Mercier, M. J., M. Mathur, L. Gostiaux, T. Gerkema, J. M. Magalhães, J. C. Da Silva, and T. Dauxois, 2012: Soliton generation by internal tidal beams impinging on a pycnocline: laboratory experiments. *J. Fluid Mech.*, **704**, 37-60, <https://doi.org/10.1017/jfm.2012.191>.

Miles, J., 1961: On the stability of heterogeneous shear flows. *J. Fluid Mech.*, **10**, 496-508, <https://doi.org/10.1017/S0022112061000305>.

Miles, J. W., 1980: Solitary waves. *Annu. Rev. Fluid Mech.*, **12**(1), 11-43, <https://doi.org/10.1146/annurev.fl.12.010180.000303>.

Moum, J. N., D. Hebert, C. A. Paulson, and D. R. Caldwell, 1992: Turbulence and internal waves at the equator. Part I: Statistics from towed thermistors and a microstructure profiler. *J. Phys. Oceanogr.*, **22**(11), 1330-1345, [https://doi.org/10.1175/1520-0485\(1992\)022<1330:TAIWAT>2.0.CO;2](https://doi.org/10.1175/1520-0485(1992)022<1330:TAIWAT>2.0.CO;2).

Moum, J. N., D. M. Farmer, W. D. Smyth, L. Armi, and S. Vagle, 2003: Structure and generation of turbulence at interfaces strained by internal solitary waves propagating shoreward over the continental shelf. *J. Phys. Oceanogr.*, **33**, 2093-2112, [https://doi.org/10.1175/1520-0485\(2003\)033<2093:SAGOTA>2.0.CO;2](https://doi.org/10.1175/1520-0485(2003)033<2093:SAGOTA>2.0.CO;2).

Moum, J. N., R. C. Lien, A. Perlin, J. D. Nash, M. C. Gregg, and P. J. Wiles, 2009: Sea surface cooling at the equator by subsurface mixing in tropical instability waves. *Nat. Geosci.*, **2**, 761-765. <https://doi.org/10.1038/ngeo657>.

Moum, J. N., A. Perlin, J. D. Nash, and M. J. McPhaden, 2013: Seasonal sea surface cooling in the equatorial Pacific cold tongue controlled by ocean mixing. *Nature*, **500**(7460), 64-67, <https://doi.org/10.1038/nature12363>.

Moum, J. N., A. Natarov, K. J. Richards, E. L. Shroyer, and W. D. Smyth, 2022a: Mixing in equatorial oceans. *Ocean Mixing*, Meredith, M., A. N. Garabato, Elsevier, 257-273.

Moum, J. N., K. G. Hughes, E. L. Shroyer, W. D. Smyth, D. Cherian, S. J. Warner, B. Bourlès, P. Brandt, and M. Dengler, 2022b: Deep cycle turbulence in Atlantic and Pacific cold tongues. *Geophys. Res. Lett.*, **49**, e2021GL097345, <https://doi.org/10.1029/2021GL097345>.

Munk, W. H., 1966: Abyssal recipes. *Deep Sea Res. Oceanogr. Abstr.*, **13**, 707-730, [https://doi.org/10.1016/0011-7471\(66\)90602-4](https://doi.org/10.1016/0011-7471(66)90602-4).

Narayanan, A. S., and K. S. Swapan, 2015: *Waves and oscillations in nature: an introduction*. CRC Press, ISBN: 9781466590939.

Nnamchi, H. C., M. Latif, N. S. Keenlyside, and W. Park, 2020: A satellite era warming hole in the equatorial Atlantic Ocean. *J. Geophys. Res. Oceans*, **125**(4), e2019JC01583, <https://doi.org/10.1029/2019JC015834>.

Nash, J. D., and J. N. Moum, 2005: River plumes as a source of large-amplitude internal waves in the coastal ocean. *Nature*, **437**, 400-403, <https://doi.org/10.1038/nature03936>.

New, A. L., and J. C. B. da Silva, 2002: Remote-sensing evidence for the local generation of internal soliton packets in the central Bay of Biscay. *Deep Sea Res.*, **49**, 915-934, [https://doi.org/10.1016/S0967-0637\(01\)00082-6](https://doi.org/10.1016/S0967-0637(01)00082-6).

Ochi, M.K., and C.-H. Tsai, 1983: Prediction of the occurrence of breaking waves in deep water. *J. Phys. Oceanogr.*, **13**, 2008-2019, [https://doi.org/10.1175/1520-0485\(1983\)013<2008:POOOBW>2.0.CO;2](https://doi.org/10.1175/1520-0485(1983)013<2008:POOOBW>2.0.CO;2).

Okumura, Y., and S. P. Xie, 2004: Interaction of the Atlantic equatorial cold tongue and the African monsoon. *J. Clim.*, **17**(18), 3589-3602, [https://doi.org/10.1175/1520-0442\(2004\)017<3589:IOTAEC>2.0.CO;2](https://doi.org/10.1175/1520-0442(2004)017<3589:IOTAEC>2.0.CO;2).

Osborne, A. R., and T. L. Burch, 1980. Internal solitons in the Andaman Sea. *Science*, **208**, 451-460, <https://doi.org/10.1126/science.208.4443.451>.

Passaglia, P. Y., K. R. Helfrich, and B. L. White, 2018: Optimal transient growth in thin-interface internal solitary waves. *J. Fluid Mech.*, **840**, 342-378, <https://doi.org/10.1017/jfm.2018.19>.

Perry, B. R., and G. R. Schimke, 1965: Large-amplitude internal waves observed off the north-west coast of Sumatra. *J. Geophys. Res.*, **70**, 2319-2324, <https://doi.org/10.1029/JZ070i010p02319>.

Pinkel, R., 1979: Observations of strongly nonlinear internal motion in the open sea using a range-gated Doppler sonar. *J. Phys. Oceanogr.*, **9**, 675-686, [https://doi.org/10.1175/1520-0485\(1979\)009<0675:OOSNIM>2.0.CO;2](https://doi.org/10.1175/1520-0485(1979)009<0675:OOSNIM>2.0.CO;2).

Pinkel, R., 2000: Internal Solitary Waves in the Warm Pool of the Western Equatorial Pacific. *J. Phys. Oceanogr.*, **30**, 2906-2926, [https://doi.org/10.1175/1520-0485\(2001\)031<2906:ISWITW>2.0.CO;2](https://doi.org/10.1175/1520-0485(2001)031<2906:ISWITW>2.0.CO;2).

Plant, W. J., W. C. Keller, and K. Hayes, 2010: Chatham, G. Normalized radar cross section of the sea for backscatter: 1. Mean levels. *J. Geophys. Res.*, **115**, C09032, <https://doi.org/10.1029/2009JC006078>.

- Raney, R. K., 1998: The Delay/Doppler Radar Altimeter. *IEEE Trans. Geosci. Remote Sens.*, **36**, 1578-1588, <https://doi.org/10.1109/36.718861>.
- Ray, R. D., and G.T. Mitchum, 1996: Surface manifestation of internal tides generated near Hawaii. *Geophys. Lett.*, **23**, 2101–2104, [https://doi.org/10.1016/S0079-6611\(97\)00025-6](https://doi.org/10.1016/S0079-6611(97)00025-6).
- Ray, C., C. Martin-Pluig, M. P. Clarizia, G. Ruffini, S. Dinardo, C. Gommenginger, and J. Benveniste, 2014: SAR altimeter backscattered waveform model. *IEEE Trans. Geosci. Remote Sens.*, **53**, 911-919, <http://dx.doi.org/10.1109/TGRS.2014.2330423>.
- Raynal, M., S. Labroue, M. Orszynowicz, N. Picot, A. Guillot, and F. Boy, 2016: Sentinel-6A Delay Doppler Processing: Assessment over the Global Ocean. SAR Altimetry Workshop, La Rochelle, France, October 31.
- Reale, F., E. P. Carratelli, A. Di Leo, and F. Dentale, 2020: Wave orbital velocity effects on radar doppler altimeter for sea monitoring. *J. Mar. Sci. Eng.*, **8**, 447, <https://doi.org/10.3390/jmse8060447>.
- Robinson, I. S., 2004: *Measuring the oceans from space: the principles and methods of satellite oceanography*. Springer Science & Business Media, ISBN: 9783540426479.
- Robinson, I. S., 2010: *Discovering the Ocean from Space: The unique applications of satellite oceanography*. Springer Science & Business Media, ISBN: 9783540244301.
- Rottman, J. W., and J. E. Simpson, 1989: The formation of internal bores in the atmosphere: A laboratory model. *Q. J. R. Meteorol. Soc.*, **115**(488), 941-963, <https://doi.org/10.1002/qj.49711548809>.
- Roy, P. S., M. D. Behera, and S. K. Srivastav, 2017: Satellite remote sensing: sensors, applications and techniques. *P. Natl. A. Sci. India A.*, **87**, 465-472, <https://doi.org/10.1007/s40010-017-0428-8>.
- Russell, J. S., 1844: *Report on waves*. in Report of the 14th Meeting of the British Association for the Advancement of Science. John Murray, London, pp. 311–390.
- Sandstrom, H., and J. A. Elliott, 1984: Internal tide and solitons on the Scotian Shelf: A nutrient pump at work. *J. Geophys. Res. Oceans*, **89**(C4), 6415-6426, <https://doi.org/10.1029/JC089iC04p06415>.
- Sandstrom, H., J. A. Elliot, and N. A. Cchrane, 1989: Observing groups of solitary internal waves and turbulence with BATFISH and echo-sounder. *J. Phys. Oceanogr.*, **19**(7), 987-997, [https://doi.org/10.1175/1520-0485\(1989\)019<0987:OGOSIW>2.0.CO;2](https://doi.org/10.1175/1520-0485(1989)019<0987:OGOSIW>2.0.CO;2).

Santos-Ferreira, A. M., J. C. B. da Silva, and J. M. Magalhaes, 2018: SAR Mode altimetry observations of internal solitary waves in the Tropical Ocean Part 1: Case studies. *Remote Sens.*, **10**, 644, <https://doi.org/10.3390/rs10040644>.

Santos-Ferreira, A. M., J. C. B. da Silva, J. M. Magalhaes, and M. Srokosz, 2019: SAR mode altimetry observations of internal solitary waves in the Tropical Ocean Part 2: A method of detection. *Remote Sens.*, **11**(11), 1339, <https://doi.org/10.3390/rs11111339>.

Santos-Ferreira, A. M., and J. C. B. da Silva, 2020: Can We Retrieve Internal Soliton Amplitudes in the Ocean with Sar Altimetry? What Would This Be Good For? *IGARSS 2020-2020 IEEE Int. Geosci. Remote Sens. Symp.*, 3525-3528, <https://doi.org/10.1109/IGARSS39084.2020.9323963>.

Santos-Ferreira, A. M., J. C. B. da Silva, J. M. Magalhaes, S. Amraoui, T. Moreau, C. Maraldi, F. Borde, N. Picot, and F. Borde, 2022: Effects of Surface Wave Breaking Caused by Internal Solitary Waves in SAR Altimeter: Sentinel-3 Copernicus Products and Advanced New Products. *Remote Sens.*, **14**, 587, <https://doi.org/10.3390/rs14030587>.

Santos-Ferreira, A. M., J. C. B. Silva, B. St-Denis, D. Bourgault, and L. R. M. Maas, 2023: Internal Solitary Waves within the Cold Tongue of the Equatorial Pacific generated by buoyant gravity currents. *J. Phys. Oceanogr.*, **53**(10), 2419-2434, <https://doi.org/10.1175/JPO-D-22-0165.1>.

Seager, R., M. Cane, N. Henderson, D. E. Lee, R. Abernathey, and H. Zhang, 2019: Strengthening tropical Pacific zonal sea surface temperature gradient consistent with rising greenhouse gases. *Nat. Clim. Change*, **9**, 517–522, <https://doi.org/10.1038/s41558-019-0505-x>.

Shroyer, E. L., J. N. Moum, and J. D. Nash, 2010: Energy transformations and dissipation of nonlinear internal waves over New Jersey's continental shelf. *Nonlin. Processes Geophys.*, **17**(4), 345-360, <https://doi.org/10.5194/npg-17-345-2010>.

Simpson, J. E., 1997: *Gravity currents: In the environment and the laboratory*. Cambridge University Press, 245 pp.

Smith, W. H. and D. T. Sandwell, 1997. Global sea floor topography from satellite altimetry and ship depth soundings. *Science*, **277**(5334), pp.1956-1962.

Smyth, W. D., J. N. Moum, and J. D. Nash, 2011: Narrowband oscillations in the upper equatorial ocean, Part II: Properties of shear instabilities. *J. Phys. Oceanogr.*, **41**, 412–428, <https://doi.org/10.1175/2010JPO4451.1>.

Smyth, W. D., and J. N. Moum, 2012: Ocean mixing by Kelvin-Helmholtz instability. *Oceanogr.*, **25**(2), 140-149, <https://doi.org/10.5670/oceanog.2012.49>.

- Smyth, W. D., J. N. Moum, L. Li, and S. A. Thorpe, 2013: Diurnal shear instability, the descent of the surface shear layer, and the deep cycle of equatorial turbulence. *J. Phys. Oceanogr.*, **43**(11), 2432-2455, <https://doi.org/10.1175/JPO-D-13-089.1>.
- Stammer, D., and A. Cazenave, 2017: *Satellite altimetry over oceans and land surfaces*. CRC press, <https://doi.org/10.1201/9781315151779>.
- Stellema, A., A. Sen Gupta, A. Taschetto, and M. Feng, 2022: Pacific Equatorial Undercurrent: mean state, sources, and future changes across models. *Front. Clim.*, **138**, 4:933091, <https://doi.org/10.3389/fclim.2022.933091>.
- St. Laurent, L., M.H. Alford, and T. Paluszkiwicz, 2012: An introduction to the special issue on internal waves. *Oceanogr.*, **25**(2), 15–19, <http://dx.doi.org/10.5670/oceanog.2012.37>.
- Stokes, G. G., 1880: Supplement to a paper on the theory of oscillatory waves. *Math. Phys. Papers*, **1**, 314–326, <https://doi.org/10.1017/CBO9780511702242.016>.
- Taklo, T., and W. Choi, 2020: Group resonant interactions between surface and internal gravity waves in a two-layer system. *J. Fluid Mech.*, **892**, A14, <https://doi.org/10.1017/jfm.2020.180>.
- Talley, L. D., 2011: *Descriptive physical oceanography: an introduction*. Academic press, ISBN: 9780080939117.
- Taylor, J. R., and A. F. Thompson, 2023: Submesoscale dynamics in the upper ocean. *Annu. Rev. Fluid Mech.*, **55**, 103-127, <https://doi.org/10.1146/annurev-fluid-031422-095147>.
- Tensubam, C. M., N. J. Raju, M. K. Dash, and H. Barskar, 2021: Estimation of internal solitary wave propagation speed in the Andaman Sea using multi-satellite images. *Rem. Sens. Environ.*, **252**, 112123, <https://doi.org/10.1016/j.rse.2020.112123>.
- Tourain, C., F. Piras, A. Ollivier, D. Hauser, J. C. Poisson, F. Boy, P. Thibaut, L. Hermozo, and C. Tison, 2021: Benefits of the Adaptive Algorithm for Retracking Altimeter Nadir Echoes: Results from Simulations and CFOSAT/SWIM Observations. *IEEE Trans. Geosci. Remote Sens.*, **59**, 9927-9940, <https://doi.org/10.1109/TGRS.2021.3064236>.
- Ulaby, F. T., and D.G. Long, 2014: *Microwave Radar and Radiometric Remote Sensing*. The University of Michigan Press, Ann Arbor, 984, ISBN: 9780472119356.
- Wang, D. W., and H. W. Wijesekera, 2018: Observations of breaking waves and energy dissipation in modulated wave groups. *J. Phys. Oceanogr.*, **48**(12), 2937-2948, <https://doi.org/10.1175/JPO-D-17-0224.1>.

- Wang, C., X. Wang, and J. C. B. da Silva, 2019: Studies of Internal Waves in the Strait of Georgia Based on Remote Sensing Images. *Remote Sens.*, **11**, 96, <https://doi.org/10.3390/rs11010096>.
- Warner, S. J., R. M. Holmes, E. H. M. Hawkins, M. Hoecker-Martínez, A. C. Savage, and J. N. Moum, 2018: Buoyant gravity currents released from tropical instability waves. *J. Phys. Oceanogr.*, **48**, 361-382, <https://doi.org/10.1175/JPO-D-17-0144.1>.
- Warner, S. J., and J. N. Moum, 2019: Feedback of mixing to ENSO phase change. *Geophys. Res. Lett.*, **46**, 13920-13927, <https://doi.org/10.1029/2019GL085415>.
- Whalen, C.B., J.A. MacKinnon, and L.D. Talley, 2018. Large-scale impacts of the mesoscale environment on mixing from wind-driven internal waves. *Nature Geosci.*, **11**, 842-847, <https://doi.org/10.1038/s41561-018-0213-6>.
- Whalen, C. B., C. de Lavergne, A. C. N. Garabato, J. M. Klymak, J.A. Mackinnon, and K. L. Sheen, 2020: Internal wave-driven mixing: governing processes and consequences for climate. *Nat. Rev. Earth Environ.*, **1**(11), 606-621, <https://doi.org/10.1038/s43017-020-0097-z>.
- White, B. L., and K. R. Helfrich, 2008: Gravity currents and internal waves in a stratified fluid. *J. Fluid Mech.*, **616**, 327-356, <https://doi.org/10.1017/S0022112008003984>.
- White, B. L., and K. R. Helfrich, 2012: A general description of a gravity current front propagating in a two-layer stratified fluid. *J. Fluid Mech.*, **711**, 545-575, <https://doi.org/10.1017/jfm.2012.409>.
- Willett, C. S., R. R. Leben, and M. F. Lavín, 2006: Eddies and tropical instability waves in the eastern tropical Pacific: A review. *Prog. Oceanogr.*, **69**, 218-238, <https://doi.org/10.1016/j.pocean.2006.03.010>.
- Woodson, C. B., 2018: The fate and impact of internal waves in nearshore ecosystems. *Annu. Rev. Mar. Sci.*, **10**, 421-441, <https://doi.org/10.1146/annurev-marine-121916-063619>.
- Wu, X., Y. M. Okumura, and P. N. DiNezio, 2019: What controls the duration of El Niño and La Niña events? *J. Clim.*, **32**(18), 5941-5965, <https://doi.org/10.1175/JCLI-D-18-0681.1>.
- Yoder, J. A., S. G. Ackleson, R. T. Barber, P. Flament, and W. M. Balch, 1994: A line in the sea. *Nature*, **371**, 689-692, <https://doi.org/10.1038/371689a0>.
- Zhang, X., J. Zhang, J. Meng, Ch. Fan, and J. Wang, 2020: Observation of internal waves with OLCI and SRAL on board Sentinel-3. *Acta Oceanol. Sin.*, **39**(3): 56-62, <https://doi.org/10.1007/s13131-019-1510-7>.

Zhang, X., and X. Li, 2022: Satellite data-driven and knowledge-informed machine learning model for estimating global internal solitary wave speed. *Remote Sens Environ.*, **283**, 113328, <https://doi.org/10.1016/j.rse.2022.113328>.

Zhang, Z., Y. Liu, B. Qiu, Y. Luo, W. Cai, Q. Yuan, Y. Liu, and Coauthors, 2023: Submesoscale inverse energy cascade enhances Southern Ocean eddy heat transport. *Nat. Commun.*, **14**(1), 1335, <https://doi.org/10.1038/s41467-023-36991-2>.