



# Mélange et anomalies de surface de la mer dans le sillage inertiel des cyclones tropicaux : processus et contribution des données satellites micro-ondes

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# THESE DE DOCTORAT DE

## L'UNIVERSITE DE BRETAGNE OCCIDENTALE

ECOLE DOCTORALE N° 598

*Sciences de la Mer et du Littoral*

Spécialité : *Océanographie, physique et Environnement*

Par

**Clément COMBOT**

## Mixing and sea surface anomalies in the inertial wake of tropical cyclones

Processes and contribution of microwave satellite data

Thèse présentée et soutenue à Brest, le 7 février 2023

Unité de recherche : **Laboratoire d'océanographie physique et spatiale (LOPS)**, équipe satellite et interaction air-mer (SIAM), à IFREMER, Univ. Brest IUEM, CNRS, IRD, Plouzané, France.

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# General Introduction: Space odyssey into Tropical Cyclones.

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*Version Française*

*“Mon nom est légion car nous sommes nombreux.” [Marc 5,9](#).*

Cette citation biblique, souvent attribuée au diable, pourrait parfaitement introduire notre sujet d’étude, tant sa désignation peut fluctuer régionalement d’un bout à l’autre du globe et tant il arbore chaque année de multiples visages. Fantala, Pam, Patricia, Irma ou encore Florence, tous désignent un même phénomène climatique qui sillonne et ravage les zones autour des parallèles du Cancer et du Capricorne : les cyclones tropicaux, appelés aussi ouragan dans les bassins américains, typhons en Asie, et cyclones dans les bassins indiens et du pacifique sud.

Chaque année, c’est en moyenne 85 de ces tempêtes qui visitent les eaux chaudes des basses et moyennes latitudes [[Ramsay 2017](#)], parmi lesquels 55% atteindront le seuil d’intensité cyclonique de  $33 \text{ m.s}^{-1}$  (64 knots). Répartis sur la plupart des aires maritimes tropicales, ces évènements couvrent l’essentiel du calendrier avec des périodes et des pics d’activité qui varient selon l’hémisphère et les bassins concernés, septembre étant généralement le mois le plus prolifique et mai le plus indigent à l’échelle de notre planète. Ces temporalités sont dépendantes de la réunion des conditions favorables à la cyclogénèse, dont la saisonnalité des paramètres est propre à chacune des régions cycloniques. Véritables machines thermiques, ces systèmes alimentés par des mécanismes de convection profonde déploient en retour une énergie mécanique intense, avec une puissance comparable à celle de plusieurs centaines de réacteurs nucléaires. Si la plupart d’entre eux restent cantonnés (pour ne pas dire confinés) aux grands espaces offerts par l’océan, 40% passeront ou termineront malheureusement leurs courses dans des zones côtières, avec pour destination favorite, la Chine, le Japon, les Philippines et les Etats-Unis, qui accueillent chacune en moyenne plus de 3 cas par an [[Fudeyasu et al. 2014](#)]. Loin d’être anecdotiques, les cyclones sont une réelle préoccupation, tant ils font parties intégrantes du climat de ces régions et peuvent engendrer plusieurs milliards de dollars de dégâts matériels, à l’image de Florence en 2018. Plusieurs milliers de victimes peuvent malheureusement aussi s’y ajouter, notamment dans les pays les plus démunis comme le Bangladesh avec Gorky en 1991, la Birmanie en 2008 avec Nargis, ou encore plus récemment le Mozambique avec Eloise. La force de leurs vents et leur régularité annuelle font des cyclones les évènements extrêmes les plus dévastateurs de notre planète. Cette triste réalité n’est que renforcée par l’augmentation simultanée de l’activité humaine dans les zones côtières, comme c’est le cas aux États-Unis.

Outre cette dynamique atmosphérique, ces systèmes larges de plusieurs centaines de kilomètres ne se résument pas seulement à un champ de vent tridimensionnel, ils sont

accompagnés d'une pluviométrie très prononcée et organisée sur l'ensemble de l'amas nuageux, au-dessous duquel est associé un champ océanique perturbé. Sous l'exercice de ces forçages mécaniques, une multitude de processus se met ainsi en place à différentes échelles. Que ce soit à l'interface sous la forme d'états de mer complexes et sévères, d'un réhaussement local du niveau de mer (storm surge), de flux air/mer exacerbés, ou que ce soit plus en profondeur via l'excitation d'ondes inertielles, qui de concert avec des mécanismes de mélange turbulent, génèrent une réponse elle aussi tridimensionnelle, appelée le sillage froid. Ces perturbations corrélées au champ de vent sont d'ailleurs en grande partie responsables des dégâts causés par le cyclone, à hauteur de 70%, principalement sous l'action cumulée des pluies torrentielles et des états de mer. Plus qu'un phénomène atmosphérique, les cyclones sont des systèmes couplés qui interagissent en permanence avec l'océan. La complexité de leurs interactions et des perturbations qui en découlent, ont depuis longtemps attiré la curiosité et la vigilance de la communauté scientifique.

Si à l'aube des années 70, plusieurs études pionnières ont déjà apporté les premières descriptions théoriques du champ cyclonique et de sa réponse océanique, c'est à partir du dernier quart du XXème siècle que cette discipline et plus largement les sciences environnementales, connaissent un nouvel essor porté par les récentes avancées techniques. On observe alors nombre de progrès en matière de compréhension de ces systèmes, de suivis et de prévisions, ou encore d'analyses des marqueurs de la dynamique océanique, cela ayant été rendu possible grâce à l'amélioration des modèles et des capacités de calcul, mais aussi surtout grâce au développement important des méthodes d'observations à distance et in-situ. On peut citer à titre d'exemple les efforts portés sur les estimations de trajectoire, utiles à la prévention des dommages cycloniques. Nombre de questions subsistent toutefois à différents niveaux, notamment lorsqu'il s'agit d'observation et de modélisation des vents les plus intenses, ou de compréhension et de paramétrisation du coefficient de poussée  $C_d$ , un paramètre essentiel quant à la représentation des échanges air/mer et du forçage dans les modèles 3D, mais dont le comportement est toujours discuté pour les vents d'intensités cycloniques. En ce qui concerne les phases d'intensification, plusieurs interrogations persistent sur les processus exacts se déroulant au sein de l'inner-core et qui aboutissent à des évolutions plus ou moins rapides de l'intensité du cyclone. D'un point de vue océanique, on peut aussi citer les questionnements sur la modélisation des mécanismes de couplage, sur la représentation du champ de vagues, sur la paramétrisation de la réponse océanique (sillage), ainsi que sur l'effet long terme du forçage sur le contenu thermique et la circulation globale. Enfin, les impacts du changement climatique sur l'activité tropicale sont objets de nombreuses études récentes.

Si toutes ces différentes problématiques font effectivement cogiter la communauté scientifique, il y a une question d'un tout autre ordre qui est souvent revenue dans mon entourage, ou au cours de rencontres, lorsque j'évoque la nature de mes travaux : Pourquoi étudier les cyclones tropicaux à Brest, en Bretagne ? Il est vrai que malgré son emplacement propice aux intempéries et aux caprices de l'océan, ces systèmes exotiques sont bien étrangers de nos frontières, et même si les reliquats de certaines tempêtes (Extra-tropical system) peuvent y être de passage, la Bretagne, et plus largement la France métropolitaine, sont exempts de ce genre de phénomènes. Au-delà de cette simple considération géographique, les cyclones tropicaux demeurent une composante importante de notre machine climatique et des mouvements grandes échelles, ils sont essentiels à prendre en compte dans les produits globaux, qu'ils soient fournis par des opérationnels ou le fruit d'analyse de centre de recherche. Plusieurs motifs peuvent donc pousser un laboratoire comme le LOPS à étudier ces monstres marins. Outre sa localisation sur la pointe du diable, qui fait ironiquement écho à notre phrase d'introduction, on peut citer



deux raisons fondamentales : d'une part l'enclavement des territoires d'outre-mer français dans les bassins tropicaux, souvent sujets à ce type de catastrophes naturelles, et d'autre part les méthodes d'observations. Cette dernière raison, simple, est directement liée à notre sujet de thèse et constitue un élément de réponse à cette question que l'on m'a si souvent posée. Car lorsqu'il s'agit de scruter les cyclones, plusieurs problématiques se posent au sujet de l'échantillonnage spatial et temporel, la réelle interrogation devenant alors : comment les observe-t-on de manière systématique ?

Aux regards profanes, lorsque l'on mentionne le domaine de l'océanographie, ce monde se retrouve le plus souvent projeté à celui des mesures in-situ, aux bouées et autres flotteurs Argo, aux relevés hydrographiques réalisés lors d'expéditions maritimes, aux mesures de surface comme les plateformes et les radars côtiers, ou encore les ballons-sondes pour les mesures atmosphériques. Tout ceci à juste titre, car ces différents outils sont des données essentielles qui ont permis d'établir une vérité de terrain, nécessaire à la calibration de mesures plus indirectes. Cependant, tout comme l'atmosphère, l'océan ne saurait être uniquement résolu via ces moyens de mesures très localisés, les systèmes grandes échelles comme les cyclones demandent un échantillonnage spatial important, trop important pour pouvoir être entièrement couvert par ce genre de procédés. C'est dans cette optique que les sciences du climat se sont dotées d'une flotte redoutable, fer de lance des progrès amorcés à partir des années 70, et véritables acteurs d'une révolution dans l'étude des cyclones : les satellites.

Cette constellation qui gravite à plusieurs centaines, voire dizaine de milliers de kilomètres (ex : satellites géostationnaires) de la terre offre une vision unique de bon nombre de paramètres physiques, à une échelle plus globale et avec un échantillonnage spatial sans précédent, apportant ainsi aux observations existantes une nouvelle verticalité. Bien qu'un satellite soit une source intermittente d'information, du fait de sa rotation, l'échantillonnage temporel peut être pallié par l'utilisation concomitante de plusieurs instruments similaires, voir utilisés conjointement avec des mesures plus in-situ. Ces différentes approches ont permis, tout au long des années qui ont suivi, de cartographier un ensemble de paramètres atmosphériques et océaniques à l'échelle du globe, mais aussi de décrire les différents processus physiques opérant à des échelles diversifiées (essentiellement de la méso-échelle jusqu'à la grande échelle). L'utilisation de ces données spécifiques a ainsi apporté un immense bénéfice quant à l'élaboration d'outils essentiels à la communauté des sciences du climat, allant de produits globaux comme les prévisions, les analyses et réanalyses, au simple suivi des cyclones, dont la chasse a été grandement facilitée depuis le lancement des premiers satellites géostationnaires dans les années 70 et la prolifération d'instruments variés qui lui succède à partir des années 80. Il est d'ailleurs intéressant de constater qu'avant cette ère spatiale, certains systèmes pouvaient potentiellement échapper à notre vigilance, la plupart évoluant dans des eaux désertes de toute présence humaine et observationnelle. Ces performances sont bien-sûr le fait des systèmes de mesures embarqués, le satellite n'étant en lui-même qu'un navire orbitant autour des mers gravitationnelles de notre planète. Cette plateforme spatiale est d'autant plus intéressante qu'elle permet souvent la coexistence de plusieurs instruments à la fois, greffés sur son armature, apportant ainsi des prismes de lecture différents pour une même zone observée. Véritable tournant donc, les études pionnières qui en découlent ont permis de façonner la recherche sur les cyclones telle qu'on la connaît aujourd'hui, de par l'aspect novateur de cette observation et de par sa gamme d'instruments qui ne cesse de s'étoffer avec le temps.

Fort des propriétés de leur signal, ces instruments électromagnétiques opèrent dans un spectre large en fréquence, allant principalement des ondes micrométriques du domaine

de l'infrarouge (IR) aux ondes centimétriques dites micro-ondes (MW), chacune de ces tranches fréquentielles agissant comme un filtre qui module le champ de l'observation (les ondes nanométriques du spectre visible sont aussi monnaie courante). Selon la longueur d'onde utilisée, le capteur pourra ainsi analyser soit les différentes strates de l'atmosphère ou la pénétrer pour observer la surface des océans. Bien qu'étant une mesure indirecte des marqueurs géophysiques telle la température de surface océanique (SST), les satellites passent par le biais d'autres grandeurs, e.g la température de brillance, qui est caractéristique d'un rayonnement émis par un corps (océan, atmosphère selon la fréquence), ou la surface équivalente radar normalisée (appelé  $\sigma_0$ ) qui est une propriété qu'à la surface d'un corps (ici l'océan) à réfléchir un signal électromagnétique, il s'agit donc d'une description de la géométrie de l'interface océanique, une mesure de sa rugosité. Outre le fait d'être facilement documentées par l'outil satellite, ces deux variables présentent l'intérêt d'être intrinsèquement liées aux conditions de leur environnement, et sont donc d'excellents proxys d'un ou plusieurs paramètres géophysiques en fonction de la nature de l'observation et de la fenêtre spectrale employée. Ainsi les mesures de rugosité réalisées par les capteurs actifs sont vectrices d'informations sur l'intensité des vents, tandis que les températures de brillance océanique mesurées par les capteurs passifs radiométriques sont indicatrices de la SST. Cette dernière dépendance bien que directe dans le domaine de l'IR (approximation du corps noir) est beaucoup plus contrastée pour des longueurs d'ondes MW, où la rugosité et la salinité joue aussi un rôle dans l'intensité du signal émis, ce qui peut nécessiter plusieurs instruments ou capteurs multi-fréquentiel pour les dissocier, comme c'est le cas par exemple pour ASMR-2.

Parallèlement à ces propriétés spectrales, la polarisation vient aussi changer la réponse du signal, son amplitude, sa sensibilité et sa tendance face aux facteurs extérieurs (intensité de vent...) et sous-jacents à l'observation (angle d'incidence du signal, bruit...). La différence de signature mesurée par un capteur actif dit co-polarisé (le signal rétrodiffusé mesuré et celui émis ont la même polarisation) et cross-polarisé en est un parfait exemple. Quel que soit le type de capteurs finalement sollicité, les variables géophysiques sont ensuite déduites de ces paramètres bruts ( $T_b$ ,  $\sigma_0$ ) grâce à des lois d'inversion plus ou moins complexes (en fonction des dépendances observées), à l'image des Geophysical Model Function (GMF) utilisées pour le calcul des intensités de vent. A partir de ces algorithmes, les variables géophysiques peuvent être associées aux coordonnées géographiques échantillonnées le long de la trajectoire du satellite (aussi appelé produit de niveau L2), auxquelles d'autres processus peuvent s'ajouter, comme le quadrillage régulier des données (L3), puis le recoupement et l'interpolation de plusieurs observations sur une carte globale (L4), ceci afin de fournir des produits plus élaborés et utiles aux analyses. Bien sûr l'utilisation de telles données nécessite en amont la calibration des instruments à des valeurs de références, que seules des informations plus directes comme les mesures in-situ peuvent apporter, servant ainsi de vérité terrain. Les dernières décennies auront donc vu défiler leurs cohortes d'observations et d'intercomparaisons, qui ont permis au fil du temps d'affiner les calibrations et la prise en compte de paramètres de correction, comme ceux concernant l'atmosphère (effet d'absorption, diffusion aérosol, ionosphère...). Si les mesures satellites constituent donc un moyen indirect d'informations, elles sont quelque part une manifestation directe des interactions et des processus se déroulant à l'interface, à l'image des anomalies de réflectivité provoquées par la pluie, de l'augmentation des températures de brillance avec la prolifération des embruns et de l'écume lors des épisodes de vents forts, ou tout simplement de la rugosité, qui traduit des variations de pression exercées par les vents de surface. Ces fluctuations sont des empreintes de l'atmosphère sur l'interface et font partie intégrante du "chant" électromagnétique murmuré par l'océan, et entendu par ces voyageurs spatiaux, tantôt simple spectateur (capteur passif), tantôt

acteur (capteur actif).

Mis à part la dynamique de ce signal, les satellites sont aussi conditionnés par les propriétés de leurs orbites : inclinaison, altitude, répétitivité. Qu'il s'agisse d'étudier une région particulière ou des phénomènes à une fréquence régulière, elles sont le plus souvent prédéterminées par les spécificités de leurs missions et assujetties aux besoins de l'observation. TRMM et sa trajectoire équatoriale inclinée de  $35^\circ$  fut l'observatoire idéal de l'activité pluviométrique des régions tropicales. Les géostationnaires perchés à 36 000 km, permettent eux un suivi continu d'une même aire géographique, tandis que les appareils en orbite polaire offrent une couverture quasiment globale des océans avec une répétitivité plus ou moins longue. Les satellite sun-synchronous sont un cas particulier de ces derniers, ils affichent une orbite légèrement inclinée par rapport à l'axe de rotation terrestre, tout en offrant une exposition solaire constante pour un même point de passage (i.e. heure solaire fixe pour un même point de la surface terrestre). La grande majorité des instruments qui seront présentés dans cette thèse obéissent d'ailleurs à ce genre de trajectoire.

La télédétection est donc une alchimie délicate et savamment pensée, depuis la nature du signal jusqu'à la définition de la trajectoire et du capteur utilisé. Il n'est pas étonnant, compte tenu de ces propriétés, que cette mesure singulière ait trouvée toute sa place dans le suivi de ces phénomènes tropicaux, à la fois trop larges et difficile d'accès pour toute autre type de mesures (à l'exception des vols). Au cours de cette odyssée spatiale dans les cyclones, maintenant vieille de plus de 40 ans, les instruments opérant dans le domaine du visible et de l'infrarouge ont largement dominé la plupart des procédures opérationnelles (analyse d'intensité, suivi ...). Malgré les grandes forces de ces types de capteurs, cela reste une observation dépendante de la couverture nuageuse, et par conséquent une mesure des propriétés de la cime nuageuse des cyclones. Dans l'optique de notre thèse, nous nous tournerons donc principalement vers les capteurs micro-ondes de la tranche de 1 à 40 GHz (SHF), qui permettent l'observation de la surface océanique indépendamment de l'obstruction nuageuse. Si les signaux MW présentent une relation plus ambiguë avec la SST, ils contiennent en revanche plus d'information sur les autres paramètres de l'interface et sont donc particulièrement intéressants pour analyser la réponse océanique. Au fil des années, les capteurs de cette gamme de fréquences ont su s'imposer jusqu'à être incorporés dans nombre d'analyses, si bien qu'ils sont devenus incontournables dans le paysage satellitaire actuel. Les altimètres, les radiomètres MW, les diffusiomètres ou encore les instruments SAR en sont des exemples. La dernière décennie a été particulièrement fructueuse sur ce plan avec le passage à 6 altimètres simultanément en fonction, une nouvelle génération de radiomètres L-band s'intéressant à la fois à la salinité et à la mesure des vents intenses, ainsi que le développement de nouvelles méthodes d'inversion de vent employant l'imagerie SAR. Pour toutes ces raisons, les satellites MW de cette dernière décennie (2010s) seront au cœur de notre thèse.

En raison de l'intermittence temporelle caractéristique de cette observation, l'utilisation simultanée de plusieurs instruments sera nécessaire, d'autant plus que pour analyser le sillage des cyclones tropicaux, il nous faudra combiner différents types de mesures pour pouvoir caractériser d'une part le forçage (Part I), et d'autre part la réponse océanique (Part II). Si un mot pouvait résumer cette thèse ce serait donc celui de synergie : synergie entre les différentes observations, synergie entre les différentes variables physiques associées au champ cyclonique et à la réponse océanique, synergie dans les différentes méthodologies mises en place, et enfin synergie entre les différentes études qui ont servi de pilier à cette thèse, et que nous allons aborder dans un premier chapitre avec le con-

texte général. Ce premier chapitre introductif décrira les variables clés océaniques et les mécanismes responsables du mélange, avant de plonger dans les bassins d'études qui nous intéressent. Puis dans un second temps, nous analyserons les propriétés du champ cyclonique, afin de pouvoir ensuite aborder la réponse océanique à travers le sillage froid du cyclone. Au regard de ces éléments de contexte général, nous pourrions établir l'état de l'art et dégager la problématique de notre thèse. Enfin, la dernière section développera les moyens d'y répondre via un résumé des méthodologies déployées, en particulier celles basées sur les observations SAR et altimétrique, souvent négligées dans l'étude du sillage, et qui combinées, constituent un atout majeur pour notre thèse.

Ce long discours aux allures de panégyrique sert d'introduction générale. S'il met en exergue les observations satellites, c'est pour mieux amener l'idée centrale autour de l'utilisation combinée des capacités actuelles de télédétection. Dans les chapitres qui suivent l'introduction scientifique, nous démontrerons successivement les capacités récentes des instruments MW à décrire précisément le champ cyclonique, à prospecter la zone des vents maximums jusqu'alors inaccessible par les inversions de vent de surface, à guider et vérifier le profil complet des vents d'un modèle paramétrique de vortex cyclonique, ou encore à capturer la variabilité spatio-temporelle des signatures du sillage froid. La coexistence de ces différentes technologies de télédétection et des mesures in-situ Argo offre une vision inégalée du sillage quasi-inertiel, et permet une lecture plus approfondie des anomalies de surface et des interactions air/mer (estimation du Cd), grâce à l'utilisation de lois d'échelles développées en collaboration avec le Pr. Vladimir Kudryavtsev du laboratoire SOLab. En définitive, cette thèse souhaite promouvoir une approche multi-plateforme pour une meilleure représentation du sillage froid et de ses paramètres, mais aussi démontrer sa nécessité à travers plusieurs études de cas. Cette philosophie observationnelle est en accord avec les recommandations d'études précédentes, notamment l'enquête visionnaire publiée par Bill Gray en 1970, qui mentionnait déjà à l'époque l'importance de développer davantage les techniques de télédétection et de les combiner pour une meilleure lecture des propriétés des cyclones tropicaux. Cette affirmation trouve encore un écho aujourd'hui, avec les nombreuses observations micro-ondes émergentes qui ne sont pas encore pleinement exploitées ou incorporées dans des produits opérationnels, comme l'indique [Knaff et al. \[2021\]](#). Sur un ton plus fantaisiste, on peut résumer l'esprit de notre thèse à la manière de Tolkien dans *Le Seigneur des anneaux* : "une base multi-satellite pour les amener tous, et dans l'analyse du sillage les lier".

*“My name is Legion, for we are many.” [Marc 5,9](#).*

This biblical quote, most often assigned to the devil, could perfectly introduce our subject of study, as its names can fluctuate regionally and as it manifests itself through several events every year. Fantala, Pam, Patricia, Irma or Florence, all refer to the same climatic phenomenon that furrows and ravages the areas around the tropics of Cancer and Capricorn: the Tropical Cyclones (TCs), also called hurricanes, typhoons or simply cyclones in their respective basins.

Each year, an average of 85 of these storms visit the warm waters of the low and mid-latitudes, among which 55% will reach the cyclonic intensity threshold of  $33 \text{ m.s}^{-1}$  (64 knots). If they cover most of the calendar and tropical waters extent, the periods and the peaks of activity vary according to the hemisphere and basins, with September and May being respectively the most prolific and the poorest months in terms of global number of cases. These temporalities are dependent on the gathering of favourable conditions for the cyclogenesis, whose seasonality is specific to each cyclonic region. The vertically large clouds that rotate around the eye are fed by deep convection mechanisms, and part of this energy helps to fuel the wind field which displays a power comparable to several hundreds of nuclear reactors. If most of them remain confined to the open-ocean, 40% will pass or unfortunately end up in coastal areas, with China, Japan, the Philippines and the United States as their favourite destinations with at least 3 cases per year. Far from being sporadic, cyclones are a real part of the climate of those regions and can cause billions of dollars of material damage, as Florence in 2018. Unfortunately, several thousand victims can be added to this, particularly in the poorest countries such as Bangladesh with Gorky in 1991, Burma in 2008 with Nargis, or more recently Mozambique with Eloise. The strength of their winds and their annual regularity make cyclones the most devastating extreme events on our planet. This sad reality is only reinforced by the concurrent increase in human activity in coastal areas, as is the case in the USA.

In addition to their three-dimensional wind fields, these wide systems are associated with a very intense rainfall, organized over the entire cloud cluster below which lies a disturbed oceanic field. Under the wind stress forcing, a multitude of processes are set up at different scales, whether at the interface in the form of complex and severe sea states, local storm surge, enhanced air/sea exchanges, or whether at greater depths through the excitation of inertial waves and turbulent mixing mechanisms that generate a three-dimensional response, called the cold wake. The wave field generated is largely responsible for the landfall damage caused by the cyclone, up to 70% with the cumulative effect of extreme rainfalls. Tropical cyclones are more than just atmospheric phenomena and act as truly coupled systems that constantly interact with the oceans. The complexity of their interactions, their exchanges, as well as the associated responses, have long attracted the interest and vigilance of the scientific community.

If at the dawn of the 70s, several pioneering studies have already provided the first theoretical descriptions of the cyclonic field and its oceanic response, it is from the following years that this discipline and more broadly the environmental sciences are experiencing a new growth driven by recent technical advances, such as the improvement of models and computing capacities, but also thanks to the significant development of remote and in-

situ observing systems. One example is the efforts made on trajectory estimates, which, despite a constant growth of coastal activities, have greatly contributed to prevent from cyclonic damages. However, many questions remain at different levels, especially when it comes to observe and model the most intense winds, or to parameterize the drag coefficient  $C_d$ , an essential parameter for the depiction of the air/sea exchanges and the forcing in 3D models, but whose behavior is still debated for cyclonic winds. More generally, the inner core processes that lead to rapid intensification or weakening of the TC remain quite challenging. From an oceanic point of view, we can also mention the issues on modelling the coupling mechanisms, the 2D wave field, or on the parameterization of the oceanic response (wake), as well as the long-term effect of the forcing on the heat content and the global circulation. Finally, the impacts of climate change on tropical activity are the subject of many recent studies.

If all these different issues are putting the scientific community in ebullition, there is a question of a completely different order that has often come up with my relatives when I mention the nature of my work: why do we study tropical cyclones in Brest, in Brittany? It is true that despite its stormy location in prey to the whims of the ocean, these exotic systems are complete strangers to our borders, and even if the remnants of certain storms (Extra-tropical system) may visit our lands, Brittany, and more broadly the metropolitan France, are exempt from this type of phenomenon. Beyond this simple geographical consideration, tropical cyclones remain an important component of our climate machine and large-scale movements, they are essential to consider in global products, whether provided by operational or the result analysis of research center analysis. Several reasons can therefore push a laboratory like LOPS to study these marine monsters: on the one hand, the isolation of French overseas territories in tropical basins, which often undergo this type of natural hazard, and on the other hand, the observation methods. Indeed, when it comes to TCs monitoring, several problems arise from a spatial and temporal point of view, the real question then becoming: how do we observe them in a systematic way?

To the uninitiated, the oceanography world is most of the time associated to in-situ measurements, buoys and other Argo floats, hydrographic surveys conducted during maritime expeditions, surface measurement systems such as platforms and coastal radars, or weather balloons for atmospheric measurements. If they are all essential data that have settled a ground truth necessary for the calibration of more indirect measurements, the global atmosphere and ocean can not be only solved by these very localized measurements. Large-scale systems such as cyclones do require larger spatial sampling than the one provided by classical instruments. It is in this perspective that the climate sciences have acquired a formidable fleet, spearheading of the progress made since the 1970s, and real actors of a revolution in the study of cyclones: the satellites.

This constellation at hundreds or even tens thousand kilometers high, offers a unique vision of many physical parameters on a global scale and with an unprecedented spatial sampling. Although a satellite is an intermittent source of information, the limited temporal sampling can be mitigated by the concomitant use of similar instruments, or even used jointly with in-situ measurements. These different approaches have allowed throughout the years to globally map a set of atmospheric and oceanic parameters, but also to describe the different physical processes operating at multiple scales (essentially mesoscale to large scale). The use of these specific data has thus brought a huge benefit in the development of essential tools to the climate science community, ranging from global products such as forecasts, analyses, and reanalyses, to the simple monitoring of cyclones,

whose hunting has been greatly eased since the launch of the first geostationary satellites in the 70s and the proliferation of various instruments that succeeded from the 80s. It is interesting to note that before this space era, some systems could potentially escape our vigilance, most of them evolving in waters deserted of by human and observational presence. These performances are of course due to the onboard measurement systems, the satellite itself being only a ship orbiting around the gravitational seas of our planet. This space platform is even more interesting as it allows the coexistence of several instruments at the same time, grafted on its frame, thus bringing different views for the same observed area. Being a real breakthrough, the pioneering studies that followed have helped to shape the modern tropical cyclone research, by the innovative aspect of its observation and by its range of various instruments that keeps on blooming.

Thanks to their signal properties, the onboard electromagnetic instruments cover a broad frequency spectrum ranging mainly from the micrometric waves of the infrared (IR) to centimetric waves called microwaves (MW), each frequency channel acting as a filter that modulates the observation frame (nanometric waves of the visible spectrum are also widespread). Depending on the wavelength used, the sensor will be able to analyse either the different atmospheric layers or penetrate it to look at the ocean surface. Although being an indirect measure of geophysical markers such as the sea surface temperature (SST), remote sensors derive other quantities: e.g the brightness temperature which is characteristic of a radiation emitted by a surface (ocean, atmosphere depending on the frequency), or the normalized radar cross section (NRCS) which is a description of the ocean interface geometry and roughness. In addition to being easily documented by the satellite, these two variables have the advantage of being intrinsically linked to the conditions of their environment, they are therefore excellent proxies of one or several geophysical parameters, depending on the nature of the observation and the spectral window used. Thus, the roughness measurements made by active sensors provide information primarily on wind intensity, while the oceanic brightness temperatures measured by passive radiometric sensors can be marker of the SST. This latter dependency is however much more ambiguous for MW wavelengths than IR ones (black body approximation), as sea surface roughness and salinity also play a role in the intensity of the emitted signal. Several instruments or multi-frequency sensors may then be needed to dissociate them, as it is the case for example with ASMR-2. In addition to these spectral properties, polarization may also affect the signal response, its amplitude, its sensitivity and its trend to environmental factors like wind stress and wind direction, or to instrument parameters like incidence angle, noise... . For instance, the co-polarized and cross-polarized NRCS do not behave similarly under an increasing wind stress.

Whatever the type of sensors finally used, the geophysical variables are deduced from these raw parameters ( $T_b$ ,  $\sigma_0$ ) thanks to some inversion laws like the Geophysical Model Function (GMF) used to derive wind intensities. From these algorithms, the geophysical variables can be provided along geographic coordinates sampled along the satellite swath (also called L2 level product) and other processes provide the regular gridding of the data (L3), followed by the merging and the interpolation of several observations on a global map (L4), to provide more sophisticated products for the analyses. Prior to operational applications, such indirect measurements require calibration to reference values that are usually provided by ground measurements, the so-called ground truth. Many intercomparisons have been made through the last decades to refine the calibrations and the corrections with regard to the atmosphere influence (absorption effect, aerosol scattering, ionosphere ...). If remote sensing is an allusive source of information, its observation is somehow a direct manifestation of the processes taking place at the interface, such as the

reflectivity anomalies triggered by rain impinging the specular ocean, or the brightness temperatures increase due to the ubiquity of sea spray and whitecaps under high winds regime. All these variations are the translation of the atmospheric imprints on the electromagnetic song whispered by the ocean and recorded by the satellites.

Apart from the signal dynamics, satellites are also constrained by their orbits: tilt angle, altitude, cycle of repeatability. Whether it is to study a particular region or phenomena at a regular frequency, they are predetermined by the observation needs. TRMM and its equatorial trajectory tilted by  $35^\circ$  was the ideal observatory of the rainfall activity in the tropical regions; the geostationary instruments with their 36 000 km orbit allow a continuous follow-up of a same geographical area, while the polar trajectories offer an almost global coverage of the oceans with a defined cycle, the sun-synchronous satellites being a special case since they display a polar orbit while offering a constant solar exposure for a same passage point (i.e. fixed solar time for the same point on the earth surface). The great majority of the instruments which will be presented in this thesis obey this kind of trajectory.

Remote sensing is a complex alchemy dependent on the orbit, the sensor type and spectral properties of the signal, but well-suited for the monitoring and the analysis of these moving tropical systems that are too large or too remote for ground observations. Since the 40 years-old beginning of the space odyssey into cyclones, satellites operating in the visible and infrared range have largely dominated most operational procedures (intensity analysis, monitoring ...). Despite the great strengths of these instruments, it remains an observation highly dependent on cloud coverage, and therefore primarily an observation of the TC top atmosphere. For the sake of our thesis, we will turn to microwave sensors from the 1-40 GHz band (called super high frequency band SHF), which grants observation of the ocean surface quite independently of the atmospheric obstruction. While MW signals have a more ambiguous relationship with SST, they contain more information about the interface parameters and are therefore particularly interesting for the ocean response analysis. Over the years, sensors in this frequency range have gained in popularity and have been incorporated into some operational analyses. They offer a wide choice of instruments such as altimeters, MW radiometers, scatterometers, and SARs. The last decade has been particularly prolific in that matter, with the increase to 6 altimeters simultaneously in operation, a new generation of L-band radiometers addressing both salinity and intense wind measurements, and the development of new wind inversion methods based on SAR imagery. Due to their capabilities and the recent breakthrough of some of their sensors, this thesis will focus on the MW satellites of the SMOS era (since 2010).

The simultaneous use of several instruments will be mandatory to compensate their temporal undersampling. Besides, to fully-analyse the tropical cyclone wake, we will combine different types of satellite data to be able to depict on the one hand the forcing (Part II), and on the other hand the oceanic response (Part III). If one word could synthetize this thesis, it would be synergy: synergy between the different observations, synergy between the different geophysical variables associated to the TC field and the oceanic response, synergy between the different methodologies used, and finally synergy between the different studies that provide guidance for the thesis, and that we will introduce in a first chapter with the general context (Part I). This first chapter will describe the key parameters and the mechanisms responsible for the oceanic turbulent mixing, before focusing to the tropical basins. Then, in a second step, we will draw our attention into the properties of the TC field. Thanks to these two descriptions we will be able to



address their coupling through the cold wake response. We will of course outline the recent progress made through a state-of-art, which will allow us to identify our problematic. Finally, the last section will introduce the methodologies deployed to address some of the shortcomings previously listed. In particular, the innovative ones based on under-track altimetry and SAR observations, often overlooked in the study of the TC wake, and which combined, constitute a major asset for our thesis. .

This long, panegyric-like speech serves as a general introduction, and if it highlights satellite instruments, it is to better convey the central idea of our study around the combined use of current remote sensing capabilities. In the chapters that follow the scientific introduction, we will successively demonstrate the recent capabilities of MW instruments to accurately depict the cyclonic field, to prospect the previously inaccessible maximum wind zone, to guide and monitor the complete wind profile of a parametric TC vortex model, or to capture the spatio-temporal variability of cold wake signatures. The coexistence of these different remote sensing technologies with in-situ Argo measurements provides an unparalleled view of the quasi-inertial wake, and allows for a more in-depth reading of surface anomalies and air/sea interactions (Cd estimation), thanks to the use of scaling laws developed in collaboration with Prof. Vladimir Kudryavtsev of the SOLab laboratory. Ultimately, this thesis wishes to promote a multi-platform approach for the scaling and the better depiction of the cold wake, but also to demonstrate its necessity through several case studies. This satellite observation philosophy is indeed in line with recommendations of previous reviews, including the visionary survey released by Bill Gray in 1970, which already at that time mentioned the importance of further developing remote sensing techniques and combining them for a better reading of the TC properties. This assertion still echoes today, with the many emerging micro-wave observations that are not yet fully exploited or incorporated into operational products as stated by [Knaiff et al. \[2021\]](#). In a more fanciful tone, we can sum up the spirit of our thesis in the manner of Tolkien in Lord of the Rings: "a multi-platform satellite base to bring them all, and in the analysis of the wake bind them."

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# **Part I**

## **Context, Issues and Thesis Stakes.**

# Scientific Introduction: A turbulent story of coupling

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As a fundamental element of our climate, the ocean is as ubiquitous as it is intrinsic to the life of tropical cyclones (TCs). It is with the ocean that TCs come to life and fade away as they flee warm waters, or quickly disappear as they leave their blue cradle to venture over land. While it may seem counter-intuitive to begin with the oceanic description before saying anything about TCs, it is their close relationship that characterises them and differentiates them from other low pressure systems. Unlike mid-latitude storms or polar cyclones, which are driven by baroclinic instabilities resulting from the horizontal temperature gradient of the atmosphere (called frontogenesis); TCs are born thanks to deep convective mechanisms that take advantage of the tropical oceanic heat (cyclogenesis), which will feed the storm with the latent heat release [Beucher 2010]. Our journey will therefore start as a hurricane, in the ocean, from where we will provide an overview of its overall pattern and structure, before focusing on tropical regions.



## 1.1 The Ocean Description.

### 1.1.1 Global Ocean structure

#### 1.1.1.1.a. General properties

Of all the fabrics that make up the Earth's cloak, it is the blue of the ocean that stands out and dominates the colour spectrum of the global landscape, giving it the nickname of the Blue Planet. 70% of the Earth is covered by a deep layer of salt water, unevenly distributed between the two hemispheres. This vast expanse of blue water plays a major role: it is an essential reservoir of heat, salinity and energy that fuels and regulates the entire climate [Wunsch and Ferrari 2004]. To stay with our first analogy, the climate machine (atmosphere, ocean, cryosphere) can be seen as a fabric made up of different layers, in which several patterns of different sizes interlock, intersect and interact with each other at different levels, some repeating themselves at shorter or longer intervals. Figure 1.1 shows the complexity of this entanglement of phenomena taking place at various spatial and temporal scales, ranging from evanescent micro-turbulence caused by, for example, the injection of air bubbles during the breaking of waves, to more perennial circulations on the scale of basins or the planet, such as jet streams [Cronin et al. 2019]. These different levels communicate via non-linear interactions, most often characterised by the dissipation of energy from the high to the small scale via the mixing process [Ferrari and Wunsch 2009], at the heart of the ocean's thermal and salinity equilibrium.

If the atmosphere and the ocean are the two dynamic fluid layers responsible for the observed climates, there is a strong asymmetry between these two media. With its density 1000 times higher and the particularity of its salinity, the ocean has a global stability in terms of stratification [Wunsch and Ferrari 2004] (apart for regions under strong local wind stress) and this despite the diurnal cycle. This property is associated with a much higher heat capacity. 2.5 m of water column is equivalent to the entire dry troposphere [Melville 1996], making the ocean the inertia of the climate system, and modulating in return the different exchanges and variations in the observed cycles [Cronin et al. 2019].

The ocean still remains a dynamic media. It displays an impressive network of circulations with surface currents under atmospheric and buoyant control, overlapping deeper and saltier flow of the Thermohaline system, whose connection is ensured by overturning motion [Emanuel 2001]. This global pattern conveys heat and mass flux, like the so-called ocean heat transport (OHT) from tropics to the pole, for which buoyancy fluxes play a leading role (OHT  $\sim 3$  PW) [Wunsch and Ferrari 2004]. Yet, the ocean circulation is mainly wind-driven, and a significant fraction of its induced mechanical energy is injected into the upper ocean through waves generation [Huang and Qiao 2010] and turbulent pro-

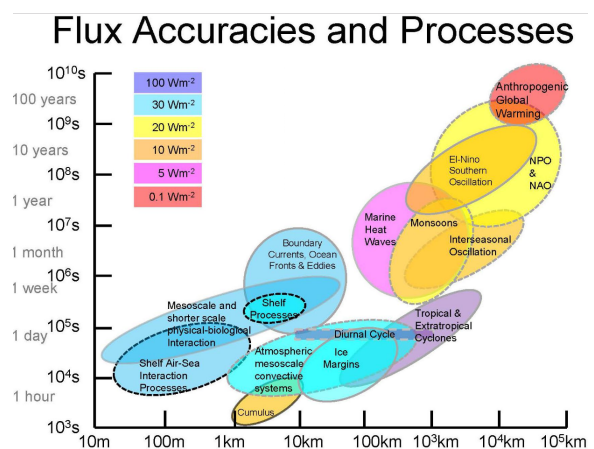


Figure 1.1: *Space-time scale of different atmospheric/oceanic processes. Extracted from Cronin et al. [2019].*



cesses that reach and deposit energy into the ocean interior, the remaining fraction going directly for the ocean circulation itself [Ferrari and Wunsch 2009]. At the surface, flows can be decomposed into several equilibria, such as the geostrophic flow, which reflects the balance between pressure and Coriolis force, or the Ekman flow, which phrases the net oceanic transport driven by a stable wind deflected by the Coriolis effect [Chereskin and Price 2001; Bôas et al. 2019] ... . The subtropical gyre and the coastal upwelling are examples.

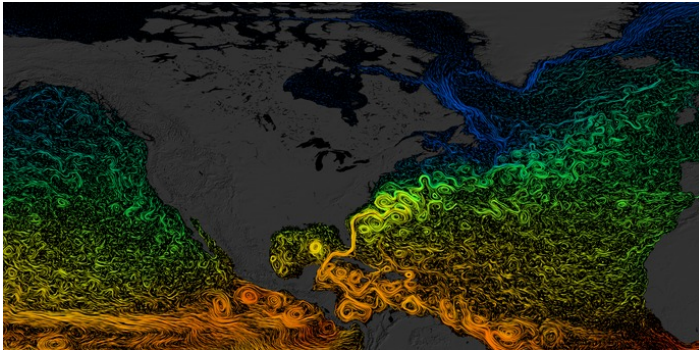


Figure 1.2: *Zoom on the oceanic surface circulation in the Northern Atlantic. Flow vectors are derived from a model used by JPL that ingests in-situ and altimeters data. Colors stand for SST. Extracted from JPL website.*

Among all these circulations, there is one of particular importance for our story, derived from the natural oscillations of waters due to inertia: the near-inertial current. Quite systemically, local wind events are sources of internal gravity waves with a near-inertial frequency [D’Asaro 1985; Alford 2003] which can even dominate the kinetic energy (KE) of the ocean during strong wind forcing as in the TCs [Firing 1997]. These waves are ubiquitous and a prominent source of turbulent kinetic energy for the ocean interior [Alford et al. 2016], necessary to

maintain its stratification and for the overturning circulation to occur [Wunsch and Ferrari 2004]. Whether directly or indirectly, the wind is thus undoubtedly a driving force behind oceanic motions, leaving a significant amount of potential energy (PE) through mixing and wind-driven currents. Shaped by these baroclinic instabilities, mesoscale eddies cover the ocean surface and account for 80 % of the ocean KE and at least half of its variability [Le Traon and Morrow 2001; Chelton et al. 2007, 2011]. A representation of this different surface circulations is shown in Figure 1.2.

This first very general description is not innocuous, as it introduces some main concepts and foreshadows the effects of TC on global circulation that we will depict in section 1.3.2, especially that related to the OHT [Emanuel 2001]. Although we have focused on the imprint of the atmospheric forcing on the ocean dynamic, it is important to recall that the SST distribution induced by heat fluxes actually fuels all the weather and wind systems which, in turn, inject momentum in the ocean and set the motions. Thereby, it is time to explore briefly the oceanic variables field and their leading mechanisms.

#### 1.1.1.b. Horizontal Pattern

Horizontal patterns of SST and SSS are paramount as they dictate heat and mass exchanges with atmosphere, but also within the ocean itself, as their fluctuations monitor the buoyancy fluxes. At the global scale, the overall distribution observed in Figure 1.3a phrases the different large-scale circulations set to balance the uneven zonal net income of radiative flux [Josey and Taylor 2019]. The SST mainly follows this meridional trend, with a progressive decrease of the temperature far from equator, as a direct consequence of solar radiation [Tomczak 2019]. Indeed, the spatial variability of the thermal field is partially dominated by surface fluxes [Alexander et al. 2000], which are also capped by the shortwave input (daily average  $100\text{--}300\text{ W.m}^{-2}$ , at maximum  $1000\text{ W.m}^{-2}$ ) [Josey and Taylor 2019]. Apart from the geostrophic eddies that advect materials and locally

impact the temperature field, some departures from this meridional trend can be found in specific regions under current regime. At their vicinities, turbulent fluxes are enhanced, sensible fluxes ( $10\text{-}20 \text{ W.m}^{-2}$ ) may reach  $100 \text{ W.m}^{-2}$  and latent heat flux ( $100 \text{ W.m}^{-2}$ ) can exceed several hundreds of  $\text{W.m}^{-2}$  [Josey and Taylor 2019]; this represents in winter a net loss of energy for warm western currents. Besides, near some western coasts of the continental shelf, strong upwellings occur due to the Ekman transport by global winds [Chereskin and Price 2001].

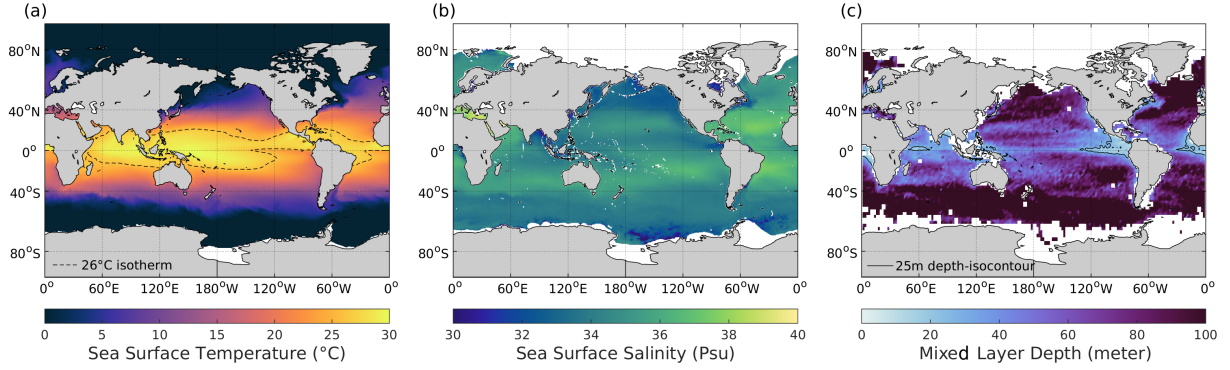


Figure 1.3: *Global distribution of the sea surface main parameters, averaged over the winter season of 2010-2018 period (DJFM North hemisphere / JJAS South Hemisphere) for: a) Sea Surface Temperature with the horizontal 26°C isotherm in dashed contour, b) Sea Surface Salinity, c) and mixing layer depth with the 25m depth isocontour as a reference point.*

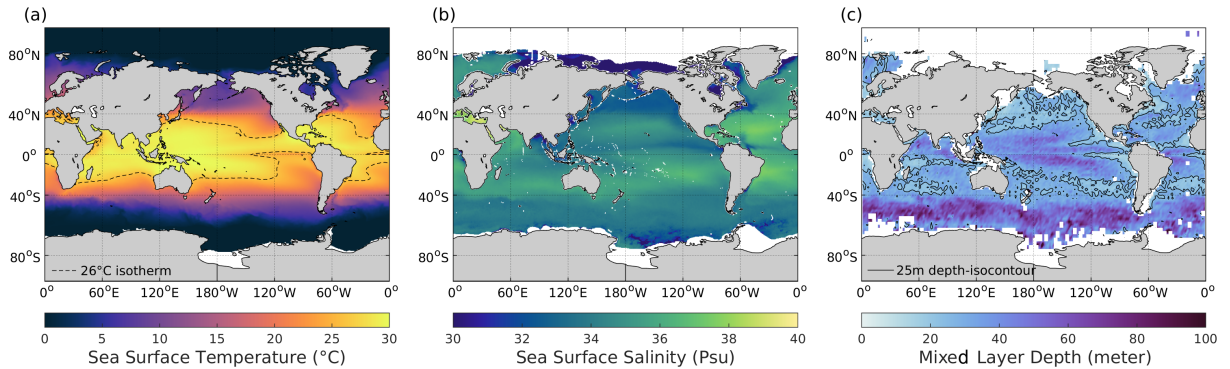


Figure 1.4: *Global distribution of the sea surface main parameters as in Figure 1.3, but averaged over the cyclonic season of 2010-2018 period (JJAS North hemisphere / DJFM South Hemisphere).*

Regarding SSS, it is strongly dependent on the freshwater flux which is expressed as the difference between the evaporative and precipitation fluxes (E-P). Its distribution follows this balance, at least for latitudes ranging up to  $60^\circ$  (high latitude, ice melt process), with maximum observed at the subtropics where evaporation is intense and precipitation are rare (diverging area) [Tomczak 2019]. On the opposite, the deep rain regions like the Equator and midlatitudes have weaker salinity (as observed in Figure 1.3b). If mechanisms monitoring SSS features are substantially similar for the same latitudes [Mignot

et al. 2012], some external sources of freshwater may disrupt this zonal distribution, such as the large river discharges like in the Bengal Bay or in the gulf of Amazon-Orinoco rivers, where persistent freshwater plumes are observed. At last, there is also a zonal disparity between the different basins, with the Atlantic being saltier than the Pacific.

From a temporal point of view, the observed seasonal variability of the SST is largely explained by the seasonal variation of the net surface heat flux, although its importance is mitigated according to the location (moderate effect in Tropics for example, see Fig 1.3a & 1.4a). On the opposite, SSS is not very sensitive to these seasonal fluctuations (Fig 1.3b & 1.4b) as the evaporative and rainfall regions remain similar. Yet, monsoon activity and river plumes shifts are important vectors of temporal variation in salinity [Tomczak 2019]. While the oceanic surface properties are often governed by its SST and dominated by the heat flux, some areas like river discharge or heavy precipitation areas (ITCZ) are under the freshwater flux regime [Cronin and Sprintall 2001]. These salient regions are of particular interest, as this input of freshwater changes the buoyancy of the ocean surface layer and so the vertical structure [Anderson et al. 1996], but also impacts both the ocean dynamic (enhanced air/sea flux) and its response to surface stress [Balaguru et al. 2012; Reul et al. 2014] such as tropical cyclone winds.

Until now, we have just mentioned the general properties of the ocean and reduced it to its surface. Still, the vertical structure is of critical importance, as it dictates the stability of the water column and its propensity to change, thereby influencing the heat and salinity budget. Apart from surface fluxes, another process governs the SST variability and acts on a larger vertical scale: the entrainment flux induced by the turbulent mixing [Anderson et al. 1996]. This mechanism mostly driven by local winds [Price 1981] is not only at the heart of this thesis but plays a major role in the observed SST and SSS, whose balance integrates both surface and vertical processes (Fig 1.5).

### 1.1.1.c. Vertical structure and Mixing

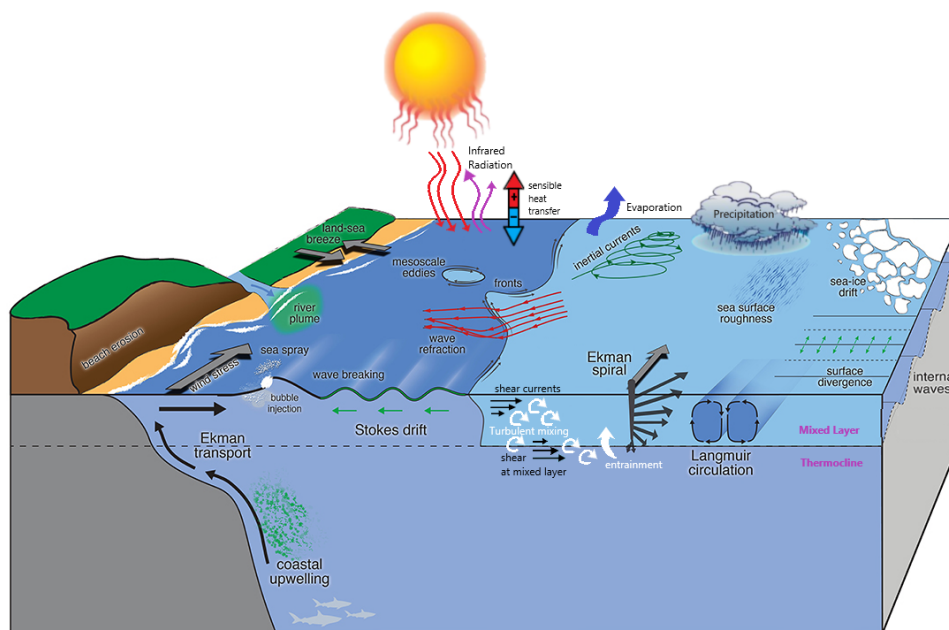


Figure 1.5: *Sketch of processes and interactions occurring in the ocean. Extracted and modified from [Cronin et al. 2019].*

The vertical structure is the backbone of the ocean, it depicts its stratification and its stability to external forces, but also informs on its heat content and on the potential energy (PE) available for the oceanic flow. In addition to the large-scale circulation and the mesoscale eddies that we have already addressed, the oceanic structure is plagued by multiple scale stress (see Fig 1.5), involving 1D (ex: heat flux), 2D (ex: horizontal advection) and 3D (ex: Ekman pumping) processes [Toba, Y. 2010]. They act at different levels on the vertical structure and influence SST and SSS fields. For instance, heat and mass fluxes change the surface buoyancy and cause forced convection, advective fluxes transport nonlinearly ocean materials, while the turbulent mixing enhances entrainment flux through shear production and cools the surface waters [Toba, Y. 2010]. This last mechanism is mainly wind-driven (directly or indirectly) and an important source of local change in ocean structure under storm regime [Price 1981]. It includes several processes: (i) the wind-induced shear currents, acting on a vertical extent equal to the Ekman depth (10-100 m) [Rudnick 2001]. (ii) Different wave-like phenomena: going from breaking waves [Melville 1996] that participate both to the air-sea flux (sea spray, air bubbles) and the turbulent mixing [Melville 1996; Rollano et al. 2019] on a short penetration depth [Huang and Qiao 2010], to the Langmuir circulation, an Ekman-Stoke motion [McWilliams et al. 2014] that transports efficiently horizontal momentum downward and assists wind-driven shear [Kukulka et al. 2009, 2010] on vertical scale of  $O(10)$  meters [Polton and Belcher 2007]; (iii) Excitation of near-inertial waves (NIW) of several hundred meters of vertical extension, they imply strong vertical displacement inside the vertical column and are a major driver of the upper ocean mixing [Price et al. 1994; Firing 1997; Ferrari and Wunsch 2009] due to their prominent vertical shear with an energy peak occurring between 100-400m [Alford et al. 2016].

Whatever the mechanisms involved, they all contribute to shape the oceanic structure according to their depth of penetration, thus stratifying the water column into 3-4 layers (see Fig 1.6) [Sprintall and Cronin 2001] of different dynamics [Moum and Smyth 2001], whose existence and depth depend on season and location, as some of the external stresses are spatially localized and vary seasonally [Wijesekera and Boyd 2001].

At the surface, the net heat flux and the entrainment flux from the turbulent mixing create a well-mixed layer of several tens of meters depth, homogeneous in temperature and salinity: the **mixing layer** (hereafter ML). This highly turbulent interface mediates the exchange of mass, heat and momentum between ocean and atmosphere and ensures the communication between the surface forcing and the deeper ocean [Sprintall and Cronin 2001]. As the SST, the mixing layer depth (MLD) has a well-documented seasonal cycle, due to the fluctuation of the heat flux and the storm activity [de Boyer Montégut et al. 2004]. During summer, the strong heating tends to stratify the vertical structure, it induces a thinner layer around 20-50 m (Fig 1.4c), while in winter, the heat loss and the intense storm activity (particularly at midlatitudes) act together to cool and deepen the ML to 60-150 m (Fig 1.4c). Similarly, but at shorter time scale, diurnal cycle also affects the ML with period of heating and stratification at daytime, followed by

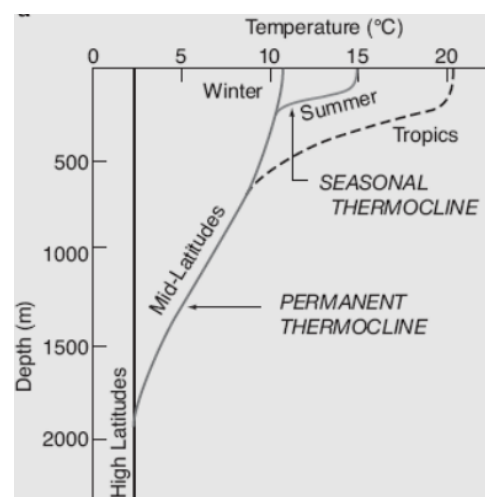


Figure 1.6: *Sketch and description of a vertical temperature profile, as regard to location and season.*



period of cooling and convection at night [Wijesekera and Boyd 2001]. If the diurnal warming can go up to 2°-3° difference for the very first meter of ML under low wind speed and strong insolation [Price et al. 1986], the data we use applies model corrections to provide a bulk SST that no longer contains diurnal effects [Gentemann et al. 2003]. It is to this estimate, also called the foundation SST founding, that we will refer throughout the thesis (see Fig A.1 Annex A).

Below this turbulent surface layer lies a stably stratified region, free from direct heat flux influence (as 90% are absorbed in the first 20 m), with a sharp temperature decrease and a gradient of salinity: the **Pycnocline** (O(1000-2000 m)). Its erosion through turbulent mixing processes, participates in the cooling of ML with the entrainment of its cold waters. When the ML starts to shrink in spring, the pycnocline region can be segmented into two slopes of different steepness, the seasonal and the permanent pycnocline. In the Figure 1.6 they are referred according to the thermal pattern, i.e thermocline, as the density structure is generally controlled by the temperature gradient. Located directly below the ML temperature jump, the **seasonal thermocline** (hereafter  $ST_h$ ) is impacted by the wind stress forcing and sharpened by the induced currents and NIW to create the steepest part of the ocean profile, extending over several hundreds of meters [Sprintall and Cronin 2001]. Of course, this process only occurs for thin ML. In fall, its deepening progressively erases this sharp region until it joins the top of the permanent thermocline, which is stable at scale of decades [Rudnick 2001]. Both ML and  $ST_h$  belong to what is called the "**Upper ocean**", i.e the turbulent oceanic layer under wind forcing. On the contrary, the **permanent thermocline** represents the boundary free from surface turbulence, preventing the deep isobaric layer (abyss) from atmospheric forcing [Kantha and Clayson 2002]. As those two underlying layers are driven by other mechanisms out of scope of the analysis, we will only focus on the upper ocean structure for the rest of the study.

In some cases, it happens that the pycnocline is not only controlled by its thermal properties, but also by its salinity (Halocline). As mentioned in 1.1.1b, the freshwater fluxes are dominating the upper ocean thermodynamic in some particular regions, such as river discharge and rainfall areas like near the Equator [de Boyer Montégut et al. 2004; Liu et al. 2009]. This advection of warm and fresh water results in the trapping of warm but saltier water from the surface into a subsurface bolus called "**barrier layer**". Through the heat flux and turbulent mixing, the freshwater lens above develops into a homogeneous ML, while the barrier layer, of similar temperature, displays a strong gradient of salinity (halocline) [Anderson et al. 1996]. This process dissociates the temperature profile from the density one, which is now first described by its salinity. It also strongly strengthens the vertical stratification with the existence of two distinct gradient regions (barrier layer & thermocline). As a direct consequence, the mixed layer is decoupled from the thermocline and the turbulent mixing at its base is inhibited by the rise of warm water from the subsurface bolus, thus preventing its cooling [Anderson et al. 1996]. Hence, the barrier layer can considerably modifies the properties of the ocean and the resulting air/sea fluxes [Balaguru et al. 2012; Reul et al. 2014].

As a multilayered column, the stability of the ocean is thus ensured by both its vertical distribution of temperature and salinity, on which the density profile depends. This latter is expressed through the UNESCO seawater state equation:

$$\rho = \rho_0[1 - \alpha(T - T_0) + \beta(S - S_0)] \quad (1.1)$$

with  $\alpha = 2.7 \times 10^{-4} \text{ } ^\circ\text{C}^{-1}$  and  $\beta = 7.6 \times 10^{-4} \text{ psu}^{-1}$  the thermal and salinity expansion coefficients respectively,  $\rho$ , T and S, the density, temperature and salinity at given depth,

and the reference values  $\rho_0 = 1025 \text{ kg.m}^{-3}$ ,  $T_0 = 22^\circ$  and  $S_0 = 36$  psu. From the vertical density profile, a main stratification metric can be derived, the Brunt-Väisälä parameter:

$$N^2 = \frac{-g}{\rho_0} \frac{\delta \rho}{\delta z} \simeq -g \left[ \alpha \frac{\delta T}{\delta z} + \beta \frac{\delta S}{\delta z} \right] \quad (1.2)$$

It is a description of the water column stability to vertical movements, it reflects oscillations within the structure and is equated to a frequency. The knowledge of the stratification also allows the estimation of the phase speed of the induced internal motions (section 1.3.1). Broadly speaking, this stratification parameter is critical for the upper ocean dynamic, for which few key equations allow to synthesize the variation of the surface field.

#### 1.1.1.d. Upper ocean budget

As we have seen, the upper ocean is under turbulence regime forced by the wind stress. ML properties are shaped by both heat and entrainment flux, while  $ST_h$  only refers to the latter [Price 1981; Alexander et al. 2000]. The deepening of MLD and the cooling induced by turbulent mixing stem from the erosion of the thermocline through shear-driven entrainment flux [Price 1981; Moum and Smyth 2001], monitored by the combination of two dominant mechanisms, the Kelvin-Helmholtz instabilities and the excitation of NIW. They both actively participate to the downward transport of wind momentum from ML to  $ST_h$ . The wind-induced shear currents generate KH instabilities at ML base, through the shape of 1-m thick billows [Rudnick 2001]. For a transient local wind, the wind stress curl makes the current oscillates at near-inertial frequency, and generates NIW with long vertical scale ( $\lambda \sim 100$  m) [D'Asaro 1985] that enhance the shear at ML base and penetrate inside the  $ST_h$  for mixing. This process actually occurs when the mechanical mixing exceeds the buoyancy flux input, and the Richardson number (Ri) reaches its critical value [Price et al. 1986]. This dimensionless parameter depicts the ratio between stratification and shear and is expressed as :

$$Ri = \frac{\frac{g}{\rho} \frac{\delta \rho}{\delta z}}{\left( \frac{\delta u}{\delta z} \right)^2} \quad (1.3)$$

When it approaches its critical value <sup>1</sup>, the turbulent mixing is fully set up, and it erodes  $ST_h$  and deepens the ML [Price et al. 1986]. The ML processes have been quite extensively analysed through different models, whether in the specific case of a cyclonic forcing or not. If some of them may focus on certain aspects of the ocean dynamics (NIW, for Geisler [1970]), most of the differences come from the parameters chosen and their modelling. Some models were solely based on 1D mechanisms [Pollard et al. 1973], while others choose to incorporate advective terms [Elsberry et al. 1976] or even diffusion for some numerical solutions ([Vincent et al. 2012a]). Although 1D processes are dominant for most regions, areas of strong advection or salinity gradient [Soloviev et al. 2015] are very sensitive to 3D processes, as well as for certain wind conditions (slow TCs see 1.3 [Vincent et al. 2012a; Jullien et al. 2012]). Thereby, three main terms are usually used to depict the heat and salinity budget (see equation 1.4 & 1.5), while the MLD lays on both advection and entrainment. They can be expressed according to Price [1981] :

$$\frac{\delta SST}{\delta t} = \frac{Q}{h_m} - U \cdot \nabla SST + W_e \frac{\Delta T}{h_m} \quad (1.4)$$

$$\frac{\delta SSS}{\delta t} = \frac{SSS \times F}{h_m} - U \cdot \nabla SSS + W_e \frac{\Delta S}{h_m} \quad (1.5)$$

$$\frac{\delta h_m}{\delta t} = -\nabla \cdot (h_m U) + W_e \quad (1.6)$$

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<sup>1</sup>stratified shear flow stability:  $Ri > Ric$  with  $Ric = 0.25$

with  $h_m$  the MLD,  $W_e$  the entrainment velocity,  $\Delta T$  and  $\Delta S$  the temperature and salinity gradient at ML base,  $F$  the mass flux ( $F = E - P$ ) and  $Q$  the heat flux expressed as  $Q = SW + LW + Ql + Qs$  (SW: shortwave radiation, LW: longwave radiation,  $Qs$  and  $Ql$ : the sensible and latent flux). The two turbulent heat fluxes can be parameterized as:  $Ql \propto U_{10} (q_{10} - q_{sea})$  and  $Qs \propto U_{10} (T_{10} - SST)$  with  $q_{10}$  and  $T_{10}$  the relative humidity and the temperature of atmosphere at 10 m. These equations are part of a three-layer model, for which the different balances are described. Here only the surface heat and salinity budget is presented. Part of the solution derived to resolve the TC wake in this thesis arises from this model assumption (see section 1.5.4).

This kind of model tries to represent the disturbances related to turbulent mixing within the ML and thermocline. Whether for analysis or to initiate models, the monitoring of the different surface parameters is essential to understand and analyse the oceanic response. In this regard, the surface dynamic and the vertical motions associated to ocean heat content (OHC) changes can be integrated through the measurement of a main ocean variable that we have not yet mentioned: the sea surface height (hereafter SSH). It represents the height of the ocean surface relative to a reference level. Once adjusted to the marine geoid <sup>2</sup> (see Fig 1.9), it phrases the contribution of many dynamic processes of different time and space scales [Calman 1987]:

$$\eta = \bar{\eta} + \eta_{bar} + \eta_{tide} + \eta_{meso} + \eta_{waves} + \eta' \quad (1.7)$$

where  $\bar{\eta}$  is related to mean currents,  $\eta_{bar}$  is the inverse barometer response to atmospheric pressure,  $\eta_{tide}$  and  $\eta_{waves}$  are respectively due to tides and surface waves,  $\eta_{meso}$  stands for the mesoscale features, and  $\eta'$  is other time-dependent contribution (e.g., due to wind). In equation 1.7,  $\eta$  is assimilated to the dynamic fraction of the sea surface height, also called the absolute dynamic topography (ADT). Throughout the thesis, we will refer to this variable (ADT) whenever we mention SSH. Operational products (L2/L3/L4), focusing on the ocean circulation, remove  $\eta_{bar}$  and some high-frequency contribution ( $\eta_{tide}$ ,  $\eta_{waves}$ ) thanks to dedicated models [Level et al. 2017]. Equation 1.7 is then written only in terms of mesoscale and turbulent activities (e.g. wind, heat flux...), their influences on the sea surface height can be described in two different ways. (i) a steric component that translates the isopycnals displacement (baroclinic motion) due to changes in ocean heat content, generally attributed to the effect of heat fluxes and turbulent mixing [Sharma and MM 2014]. (ii) a non-steric component, mainly due to the surface transports (barotropic motion) and to the geostrophic balance of the different surface circulations [Sharma and MM 2014]. In addition of being vital to map ocean currents, SSH is thus a good integrator of the vertical structure processes [Goni et al. 2009; Shay 2009; Mei et al. 2013]. Its temporal evolution complete indeed the upper ocean budget along with SST and SSS. Bunch of models have focused on this variable to depict the vertical motions linked to mixing processes, and are based on NIW propagation theory like in Geisler [1970]. The description of the associated mathematical formalism will follow later in section 1.5.4.

Before looking at our area of study, i.e tropical basins, we will now briefly introduce the measurement methods that are used to monitor these different sea surface variables.

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<sup>2</sup>an equipotential surface closest to the mean sea surface if the ocean was motionless



## 1.1.2 Observation of oceanic variables

### 1.1.2a Radiometer

1978 was a turning point in oceanographic remote sensing, with the first mission entirely dedicated to sea surface microwave observations: SEASAT. The same year, another satellite is sent with the AVHRR (Advanced Very High Resolution Radiometer) technology on board [Cracknell 2018]. These two missions have in common to use one of the most exploited tools for the measurement of the sea surface emissivity: the radiometers. These passive instruments measure, from the radiation emitted by a surface, the brightness temperature ( $T_b$ ), which is the temperature of a black body that <sup>3</sup> would emit the same amount of radiation as the targeted surface in a specified spectral band [Martin 2013]. Thereby  $T_b$  is not the physical surface temperature but it is of great interest since it depends directly on it:  $T_b = e_{(\theta,\nu)} \times SST$ , with  $e$  the emissivity. When  $e_{(\theta,\nu)}$  gets closer to 1, the surface behaves like a black body and  $T_b$  is only linked on the SST properties. This relationship is strongly dependent on the frequency channel of the observation. Such condition is fulfilled for the IR wavelength. Instruments operating in this frequency range observe a direct relationship with a very high resolution ( $O(km)$ ), which explains why efforts were initially focused on this type of measurement [Cracknell 2018]. However, they are extremely sensitive to the atmospheric effects. Many gases have absorption peaks in the infrared realm, so the transmittance of the atmosphere at these intervals is very low [Wentz et al. 2000]. IR observation of the SST is thus only possible in the two spectral windows for which the atmosphere is nearly transparent, i.e. 3-4  $\mu m$  and 10-14  $\mu m$  [Yu 2009]. Corrections are still necessary to remove attenuation, scattering by the atmosphere and aerosols, and polarization issues due to ionospheric effect. Several channels are therefore often used for this purpose in order to retrieve  $T_b$  [Bates and Smith 1985]. In addition to noise sources, the acquisition of sea surface data with IR imagers can only be done in cloud-free conditions [Wentz et al. 2000].

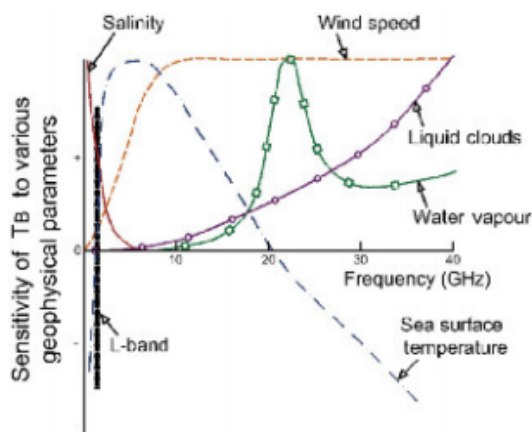


Figure 1.7: *sensitivity of MW brightness temperature to geophysical parameters and atmospheric effect*

From this perspective, the microwave (MW) part of the electromagnetic spectrum is relevant. These so-called centimetric waves are insensitive to cloud obstruction, and the atmosphere is mostly transparent at these wavelengths (transmissivity  $\sim 1$ ) especially for the low frequency MW of 1-10 GHz [Wentz et al. 2000], for which water vapor and rain effects are the weakest (strong impact of rain for highest frequency) [Martin 2013]. MW instruments are therefore very interesting tools to probe the ocean, especially during cyclonic conditions to observe through the cloud cover [Bates and Smith 1985; Stramma et al. 1986]. These properties can be nevertheless hindered by a coarser resolution ( $\sim 40$  km [Reul et al. 2020]) and a much more ambiguous relationship with the SST. At these frequencies, the ocean does not behave as a black body, the emissivity is not

only depending on temperature but on the full dielectric properties (salinity included), and on surface roughness [Njoku et al. 2000]. Besides, the e-m (H/V polarization) and

<sup>3</sup>a body that absorbs all the energy received

optical properties (incidence angle) of the instrument also modulate the emissivity level [Dinnat et al. 2002] and induce some fluctuations of the sensitivity to geophysical parameters. Despite this more complex relationship, the MW signature of the ocean contains much more information about surface properties, measurable by satellite (SST, SSS, U10)[Toba, Y. 2010]. The sensitivity of the brightness temperature to these variables depends strongly on the frequency chosen in the MW domain (Fig 1.7), so their signatures can be dissociated from each other depending on the use of a specific channel or on the combination of several channels and polarizations, as it is the case for SST retrieval. The channels of 7 and 11 Ghz [Gentemann et al. 2010]) are for instance employed to strip temperature from other contributions such as wind stress, considered spurious in this situation. Through the emissivity, other quantities like SSS and U10 can be retrieved, since they take part in the variation of the signal [Meissner and Wentz 2012; Reul et al. 2020].

If the SST now benefits from several decades of observations since SEASAT in 1978, and a global monitoring since the late 90s [Cracknell 2018], the remote sensing of salinity is quite recent with the launch of SMOS in 2009. Its measurement is essentially done through the 1.4 Ghz channel, which is particularly sensitive to its variations and has the advantage of being prominent compared to other signatures [Njoku et al. 2000](Fig 1.7). However, there is a strong asymmetry in the mapping of these quantities at the global scale and for operational use. While temperature follows a meridional distribution of several tens of degrees  $[-2^{\circ}; 40^{\circ}]$  with temporal variations that can reach several degrees [Tomczak 2019], 90% of the open ocean salinity is contained between  $[34, 36]$  psu [Wunsch 2015], with small variations essentially located between 0.1 and 0.2 psu. To achieve a daily coverage, it would require a brightness temperature precision of 0.1K [Reul et al. 2020] when the instruments accuracy is around 1 K for salinity measurements (0.5 K for SST [Wentz et al. 2000]). Therefore, the global SST data can be provided at daily resolution, whereas SSS is given through 10-day or monthly resolution products. These limitations must be taken into account when observing the TC-induced response and its temporal frame (see section 1.3.1).

### 1.1.2.b Altimeter

The year 2018 marked the 25th anniversary of operational radar altimetry, 25 years of continuous and valuable data to the oceanographic community. While the instruments onboard the GEOS-3 (1975), SEASAT (1978) and GEOSAT-1 (1985) satellites have demonstrated the richness of ocean surface topography measurements, it is with TOPEX/Poseidon (1992) that altimetry has reached a new level [Morrow and Le Traon 2012]. These pioneering instruments have made it possible to document global ocean circulation, whether associated with large current systems or mesoscale activity, for wavelengths higher than 150-200 km in average [Fu et al. 2010; Pujol et al. 2016]. They have quickly become a major part of the satellite landscape, with the singular nature of their measurements which can not be performed by any other type of instrument.

Unlike the radiometers we have just presented, altimeters are active sensors, as they send several short pulses and measure the power of the reflected signal. The general principle of the instrument is based on the travel time between the transmitted and received signals to obtain the satellite range (i.e. the distance from the satellite to the surface) and deduce the sea surface height (SSH). The good reading of the signal is still necessary to interpret properly its characteristics, as the shape (waveform) and magnitude of the return echo actually provide a complete picture of the surface dynamics, including additional geophysical variables such as wave heights and wind speed [Chelton et al. 2001].

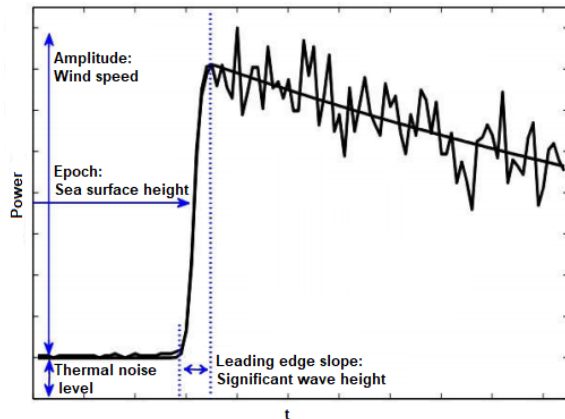


Figure 1.8: *Sketch of the temporal evolution of the reflected power of altimeter signal. It shows how the different geophysical variables are extracted from the altimeter return echo.*

The Figure 1.8 features the subtleties of altimeter measurement, with the time evolution of the backscattered power received by the sensor. The waveform is modulated by the different elements encountered at the sea surface by the transmitted signal, namely a first return related to the meeting between the leading edge of the pulse with the top-most crests of the waves, followed by a gradual rise of the reflected energy as the footprint increases. A maximum is reached once the pulse met the lowest wave trough. The significant wave height is then determined from the leading edge slope, while the wind speed intensity is related to the maximum amplitude of the return echo (gives the backscatter coefficient, see 1.2.3). At last, the temporal values (epoch) associated to the mid-point of the signal leading edge gives the satellite range [Chelton et al. 2001].

The absolute dynamic topography (ADT) is deduced from this measurement and from the precise knowledge of the orbit altitude with respect to the earth ellipsoid and the geoid shape (see Fig 1.9). These two reference information are obtained with great accuracy (orbit < 1 cm errors) by auxiliary on-board sensors like DORIS in ESA altimeters and satellite gravity missions like GOCE to meet the requirements for examining the ocean variability [Abdalla 2021], generally in the order of  $\vartheta(10cm)$ . Some corrections are also made to the interactions between the signal and certain atmospheric layers (dry/wet troposphere, ionosphere) through models (e.g. ECMWF), on-board radiometric observations, or thanks to the dual-frequency capability of most of current instruments [Abdalla 2021].

Altimetry is a critical tool for operational oceanography, the spatial sampling of its modern instruments allows to obtain a fine resolution of approximately 7 km along the track [Level et al. 2017]. However, altimetry is a diamond tip, and its observation is constrained by the narrowness of its swath resulting from its nadir-looking approach. Several instruments are thereby used to complete the description of the ocean circulation, leading to a compromise on the resolution of the observed structures. Another limiting factor comes from rain contamination as the Ku band, widely used in altimetry, is particularly sensitive to its effects. Yet most of these sensors used C-band (much less but still affected) in dual mode with Ku band and can help to mitigate rain contamination [Quilfen et al. 1998, 2010].

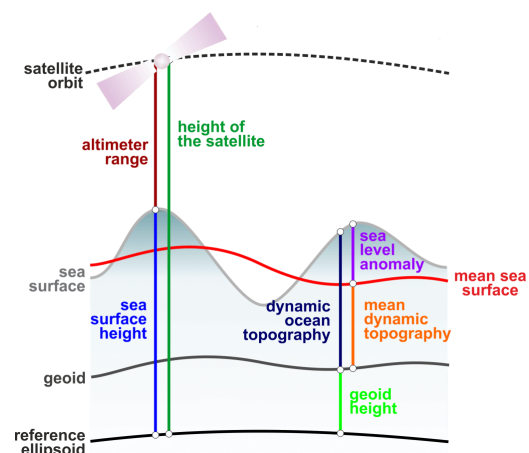


Figure 1.9: *Description of the quantities related to sea surface heights .*

### 1.1.2.c Argo

While some characteristic depths, useful for estimating the heat content of the upper layer, can be deduced from relationships based on sea level measurements [Leipper and Volgenau 1972; Shay et al. 2000; Goni et al. 2009], the use of in-situ data, or climatologies derived from them, is necessary to know the ocean stratification.

Such observations can come from the large Argo fleet, composed of about 4000 floats dispersed in all oceans. They are a valuable source of information when it comes to describing the vertical profile of temperature and salinity over a depth of generally up to 2000 metres [Freeland 2010]. These measurements are taken after a 10-day cycle, during which the instrument drifts and gradually submerges, changing its buoyancy, and then records the vertical values of temperature and salinity as it rises to the surface, where it can transmit the data via a satellite antenna. These data are collected and processed by the Global Data Assembly Centers (GDACs) which, after an automatic quality control, distribute a near-real time version and then, within six months, a reanalysed product with adjusted variables, called delayed mode [Freeland 2010]. These higher quality profiles can be used directly for a detailed analysis of a specific area or integrated into a climatology for a global description of the mean ocean state. The in-situ analysis system (ISAS) developed by IFREMER in collaboration with the CORIOLIS data centre is an example of a monthly gridded field in temperature and salinity, based on a time average of several years and with a vertical resolution of 152 levels distributed between 0 and 2000 metres and a horizontal resolution of  $0.5^\circ$  [Gaillard et al. 2016]. This global product is updated every two to three years with the new measurements (ISAS13, ISAS15, ISAS17...).

Whether it is the SST, SSH or ISAS global climate products, they all have to deal with the irregular sampling of the measurements they incorporate into their analyses. In order to map the data into a regular grid, most of them use an optimal interpolation method, which attempts to minimise the variance of the errors associated with the measurement and those introduced by smoothing and filtering the data, making the best use of the spatial and temporal sampling of the instruments [Bretherton et al. 1976; Reynolds and Smith 1994; Pujol et al. 2016]. The interpolation parameters of course differ from one product to another, depending on the nature of the observations and the geophysical variables represented. This method therefore makes a compromise between the sampling capabilities of the instruments and the physical field we want to depict, which determines the resolution of the product [Pujol et al. 2016].

In this thesis, several levels of products will be used to characterise the different geophysical variables needed to describe the ocean. REMSS and CMEMS daily gridded products will be used for SST and SSH respectively, as well as individual swaths from a collection of altimeters also processed by CMEMS. For the vertical structure information, we will use ISAS-15 climatology (based on the 2002-2015 time window) and individual profiles, both provided by IFREMER services, and with a preference for delayed versions.

To conclude on the observations related to ocean variables, examples of individual vertical profiles made at different latitudes by the same type of sensor as Argo are shown in Figure 1.10, for the temperature. It is from this type of profile that we can deduce the different stratification parameters (depth of the layers, brunt-Vaisala, ...). The Figure 1.10 shows the heterogeneity of the MLDs and the thermocline gradients between geographical areas, and reflect the different dynamics inherent to these regions.

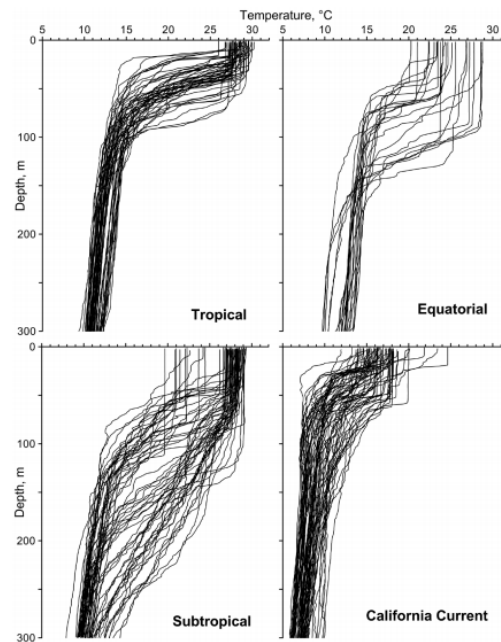


Figure 1.10: CTD Temperature profiles in four regional areas of the Eastern Pacific: Tropical, Equatorial, Subtropical and the California current. Extracted from [Fiedler 2010].

### 1.1.3 Tropical basins dynamic

The tropical ocean is characterized by a thin, warm surface layer of more than  $26^{\circ}\text{C}$ , located above a sharp temperature and density gradient [Fiedler 2010] (Fig 1.10). Its boundary fluctuates around  $25^{\circ}$  of latitudes (see Fig 1.11) according to the seasonal variations and oscillations of the Intertropical Convergence Zone (ITCZ), and can extend to the East China Sea during the summer for the western Pacific basin [Beucher 2010]. As the tropics benefit from a net heat gain throughout the year [Josey and Taylor 2019], the mixed layer undergoes less deepening than in the more northern regions and the seasonal thermocline persists through the winter, albeit altered. Some permanent BL features are also observed [de Boyer Montégut et al. 2004; Liu et al. 2009] due to the combination of large river discharges and phenomena typical of these regions, such as monsoon or equatorial upwelling [Mignot et al. 2012]. The mixing occurring in this highly stratified zone of the tropical ocean, actively participates in the maintenance of the meridional overturning circulation [Scott and Marotzke 2002].

According to the seasons, the basins at these low latitudes can provide a favourable environment for the development of intense cyclonic systems. The required climatological conditions, as described by E. Palmen [1948], include a sufficiently warm and deep water column (i.e.  $> 26^{\circ}\text{C}$  over 50 m) to supply the cyclone with latent heat release (Fig 1.11), and high relative humidity in the lower and middle troposphere to enhance condensation and reduce surface pressure. The formation of such a vortex, however, requires an initial influx of convective systems into an atmosphere that is sufficiently unstable to allow intense convection and trigger large-scale convergence [Vincent 2011]. This is why TCs are most often initiated by pre-existing tropical disturbances, developed in monsoon troughs, or formed by easterly waves [Goldenberg et al. 2001] or cluster detachments at the margin of the ITCZ [Beucher 2010]. Several hundred of these structures are generated each year when thermodynamic conditions are favourable, i.e. during the cyclone season, but only



a handful of them manage to develop into Tropical systems. Gray [1998] emphasised the importance of dynamical factors in cyclogenesis, such as the presence of divergent flows in the upper troposphere (e.g. jet streams), low vertical shear and, above all, a relative and planetary <sup>4</sup> vorticities associated to the cluster to help the development of the cyclonic circulation (see 1.2). Once these conditions are met, the disturbance will gradually develop into a warm low pressure core and become a tropical depression. As the circulation organizes and intensifies, this system will then be able to pass through the stages of a tropical storm and finally a tropical cyclone [Beucher 2010].

The RSMC and TCWC are responsible for diagnosing cyclone activity. They monitor and predict the evolution of tropical systems, by providing warnings, forecast and operational analyses of the parameters of cyclones for their assigned regions [Knapp et al. 2010]. These analyses, called Best-tracks, are a smoothed representation of the TC life history [Landsea and Franklin 2013], i.e. they provide 6-hourly estimates of location, intensity and other main parameters (e.g. wind radii) for each tropical system. Initiated from near-real-time observations, they are revisited after the TC seasons to take benefit of all available measurements that are incorporated into the analysis process [Knapp et al. 2010]. The typical tropical cyclone tracks are shown for our database in the Figure 1.12. They are derived from the analyses of the NHC and the JTWC, two other operational centres that provide TCs history for all basins [Sampson and Schrader 2000]. The spatial division of the basins follows the NHC/JTWC convention.

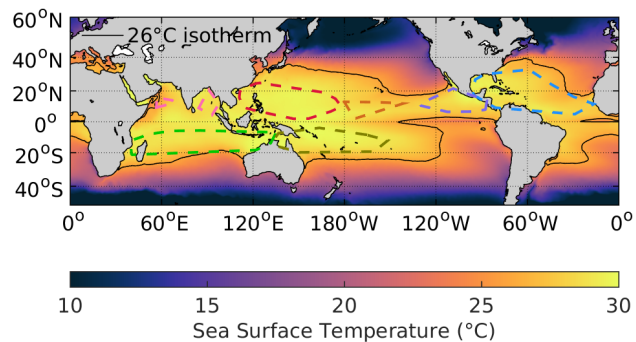


Figure 1.11: Same as Figure 1.4 but with the formation areas of tropical cyclones.

Most tropical regions have TCs, with the exception of the south-east Pacific and the South Atlantic, where strong upwelling or shear situations prevent the formation of tropical systems. The cyclone activity is unevenly distributed across the seven remaining areas, with annual averages being strongly basin-dependent [Fudeyasu et al. 2014; Ramsay 2017]. Western Pacific (WP) is by far the most active, with an average of 30 cases per year [Knapp et al. 2010] and no apparent season (unlike the other basins), although there is a peak in activity during the summer period [Ramsay 2017]. Each basin has its own characteristics which in turn modulate the TC properties. The East Pacific has the highest density of cases due to its narrowness [Bell et al. 2013], while the Atlantic and its great width allows for several extratropical transitions [Vincent 2011]. The confined seas of the North Indian region and the open ocean of the Central Pacific experience very few cases [Ramsay 2017]. The Northern Hemisphere accounts for 70% of the total cyclone activity, with the remaining 30% distributed between the South Indian and Pacific oceans [Ramsay 2017]. The annual number of cases per basin is given for our database in the Figure 1.13.

These variations in the distribution of cyclones result from the different variabilities that affect cyclonic regions on different time scales (e.g. in Atlantic, Foltz et al. [2019] and Kossin and Vimont [2007]), the most well-known of which is that introduced by the

<sup>4</sup>TCs generally form at latitudes above 5 degrees



interannual el nino/la nina episodes (weakening/strengthening of hurricanes activity in Atlantic). Shorter timescale phenomena like the Madden-Julian Oscillation, or longer such as the Pacific Decadal and Atlantic Multidecadal Oscillations (PDO & AMO), also affect the distribution of cases to a greater or lesser degree, depending on the magnitude of the variations [Foltz et al. 2019]. These variabilities often alter the sst and the vertical wind shear conditions within the tropical basin, which in turn affect cyclones activity and their intensification [Kossin 2017]. The time interval covered by our database, 2010-2018, is representative of the different known variabilities.

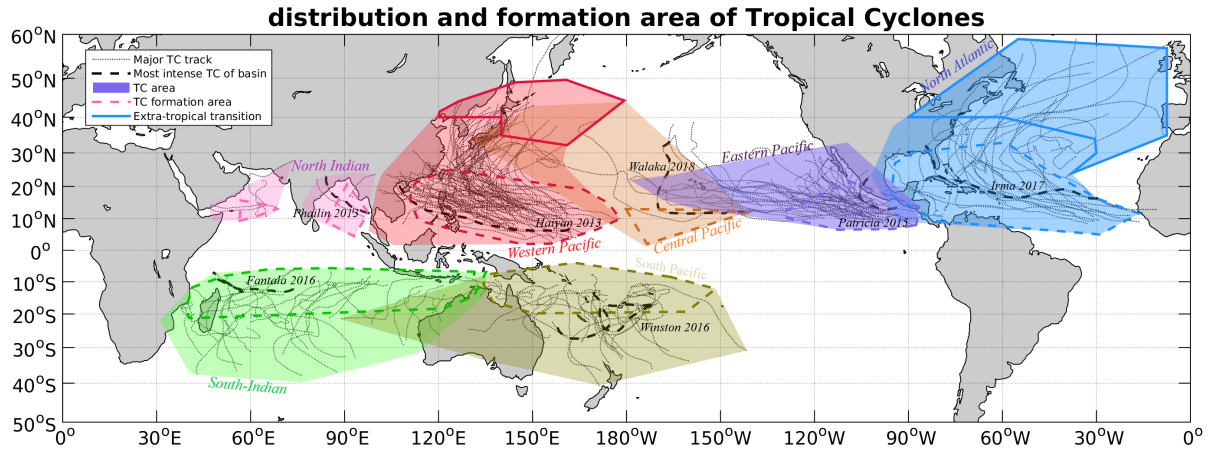


Figure 1.12: Global map of cyclone activity for the period 2010 to 2018 according to the regional classification, and with all the tracks of systems that exceeded the intensity threshold of  $33 \text{ m.s}^{-1}$ . The formation zones (dashed contour) are dissociated from the global area covered by tropical cyclones motion (thick contour). In our period and for each basin, the most intense case is specified (thick dotted line).

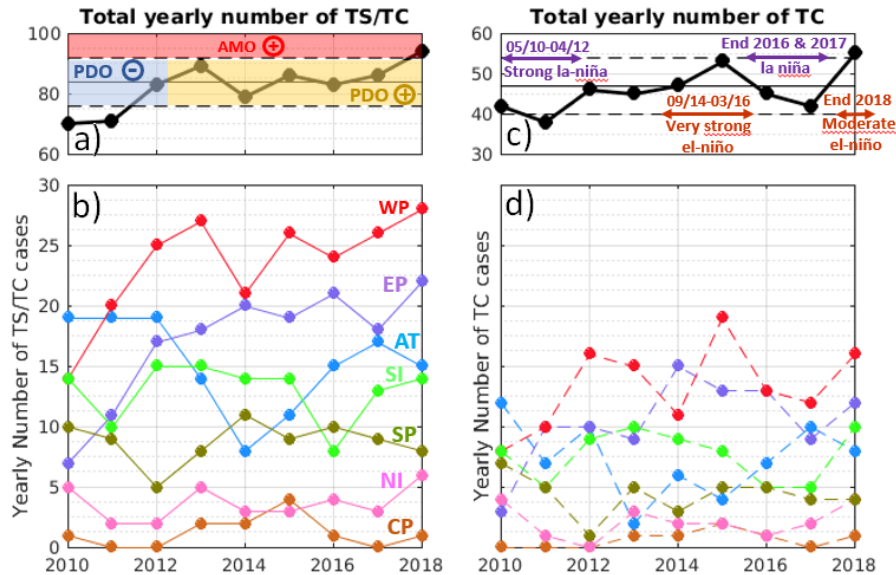


Figure 1.13: Evolution of the annual number of storms and tropical cyclones, total (a, c) and split by basin (b). The upper panels display the time interval spanned by the different long (a) and short (c) variabilities that impact the tropical basins. The graph d) represents only the annual number of cyclonic intensity systems ( $> 33 \text{ m.s}^{-1}$ ) per basin, the colour scheme follows that used to define the different basins in the Figure 1.12.

## 1.2 Tropical Cyclone Description.

### 1.2.1 Tropical cyclone structure

What was only at beginning an atmospheric whisper is now a deep convective and low-pressure system furrowing the warm waters of the tropical basin. Several hundred tropical systems have actually passed through during our sample period (2010-2018). These include typhoons Atsani (2015) and Trami (2018), major hurricanes Igor (2010), Irma and José (2017) in the Atlantic; Lane (2018) and Blanca (2015) in the eastern Pacific, as well as the Three Brothers episode (2015), in which no less than three cyclones (Jimena, Ignacio and Kilo) occurred simultaneously. Each of these cases is of particular interest and will be discussed recurrently in our manuscript, so a summary of their trajectories and some key parameters is presented in the Appendix A. Throughout our story, it is mostly the example of Hector that will illustrate our remarks. A major hurricane during the 2018 season, Hector lasted nearly 20 days and represents the perfect midget storm of Eastern pacific, with, moreover, phases of rapid intensification that make its analysis particularly difficult by operational products to monitor[Velden et al. 2006, 2017; Leroux et al. 2018], a gap that our analysis could fill.

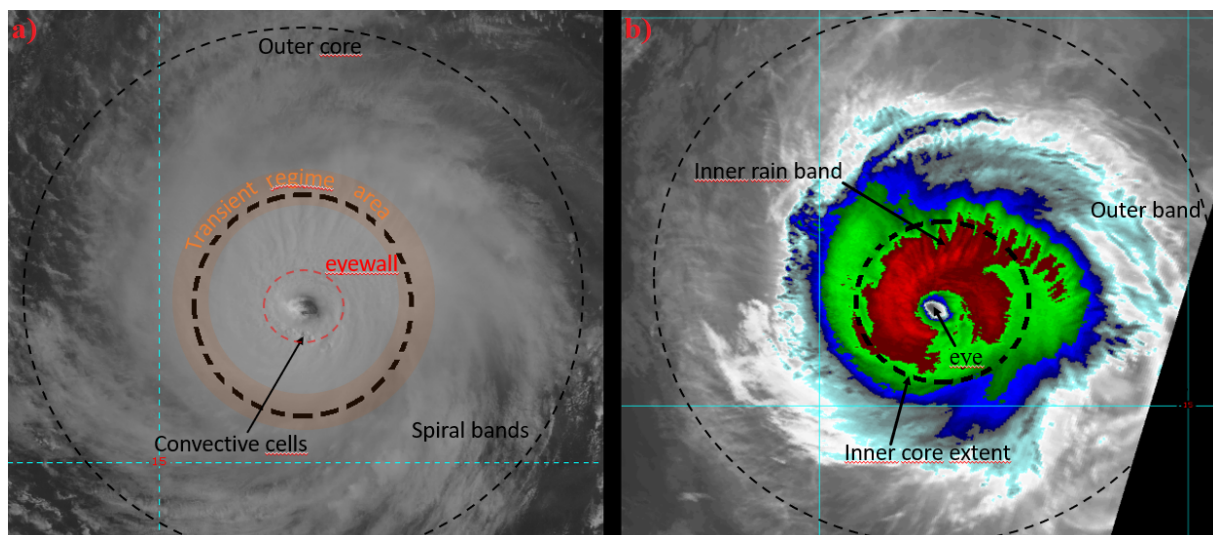


Figure 1.14: *Horizontal view of the top clouds structure of hurricane Hector, the 7th august 2018 around 18h45. Characteristic patterns are indicated in two subplots: a) Image in visible channel from geostationary imagery GOES-15, b) enhanced infrared image from MODIS/VIIRS interpolated in 1km mercator grid. Extracted from RAMMB.*

The Figure 1.14 shows two satellite acquisitions made during its passage, the first in the visible channel where we can observe the nebulosity (Fig 1.14 a), the second in the infrared, which provides first indications on the cloud structure of the cyclone (Fig 1.14 b). The image in the visible channel shows convective clouds organized in spirals orbiting around a quiescent eye, around which a more compact cloudy mass surrounds it [Houze 2010]. This dense cluster is called the eyewall, it is in its meanders that the deepest convection occurs [Gray and Shea 1973] and that the most prominent cumulonimbus are found [Jorgensen 1984a,b], reaching the top of the troposphere (represented in the horizontal sketch Fig 1.15) as shown by the colder brightness temperature observed in Figure 1.14b (inverted scale, red=coldest, blue=coolest). A few bright, bulging spots are also noticeable in its vicinity, characteristic of intense rain cells [Didlake and Houze 2013a]. The surrounding region, still under deep convection regime, describes what is

called the **inner core** [Gray and Shea 1973]. Outside, at the periphery, are the rainbands that build the **outer core** [Weatherford and Gray 1988a], they exhibit less organised convection and a smaller vertical extent, and are mainly formed by stratiform rainfalls with some convective cells embedded ([Didlake and Houze 2013b], Fig 1.16). Above, a canopy of cirrus clouds overlies the whole [Velden et al. 2006]. This entire cloud system stems from the secondary circulation, which provides the necessary energy by releasing latent heat [Holland and Merrill 1984].

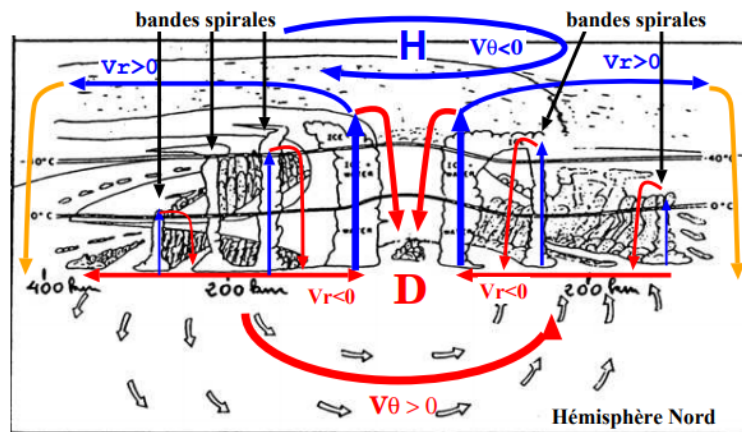


Figure 1.15: *Vertical Sketch of a tropical cyclone. D stands for the centre of low surface pressure, H the centre of high pressure in the upper troposphere (100-200 hPa).  $V_\theta$  represents the so-called "primary" circulation of a cyclone;  $V_r$  represents the so-called "secondary" circulation of a cyclone.. Extracted from Beucher [2010].*

The circulation itself is set by the central depression accompanied by a warm anomaly in the eye of the cyclone [E.Palmen 1948] (+5/10° [Vincent 2011]). As it rises, it lowers the density of the air column and maintains the surface low [Riehl 1950]. This pressure difference results in strong inflow of moist air heading towards the center of the storm [Gray and Shea 1973; Holland and Merrill 1984]. There, the warm, moist air rises through intense convection mechanisms, a huge amount of energy is displayed with the latent heat release, that keeps on feeding the whole thermal machine [Emanuel 2004]. At the top, the ascending flows diverge into an anticyclonic circulation, the dry air eventually cools down and subsides into the free troposphere [Emanuel and Rotunno 2011]. As it descends, the air warms up and joins again the converging flow, thus completing the loop, which will continue to be fed as long as the TC remains on warm waters [E.Palmen 1948]. This process constitutes the main branch of the thermodynamic circulation within the cyclone, and is accompanied by several cells of upward and downward motions like illustrated in Figure 1.14. At the upflow location, intense rainfall occur [Didlake and Houze 2013a], as presented in Figure 1.16 a & b, while subsidence areas (downward flux) are characterized by rain-free region, like the "moat", located after the deep convective area of the eyewall [Chou et al. 2013]. These two images illustrate, through active and passive measurements, the heterogeneity of cloud structures and associated rainfall signatures within the cyclone (89 Ghz channel is suited for precipitation due to its sensitivity to cloud liquid water [Neeck et al. 2010; Imaoka et al. 2010]). They show an area of intense and convective precipitation, strongly organized by the vortex circulation (inner-core), and a more convection-free region with large rain-free area and some stratiform rain and squall lines [Chou et al. 2013; Didlake and Houze 2013b; Yang et al. 2018].

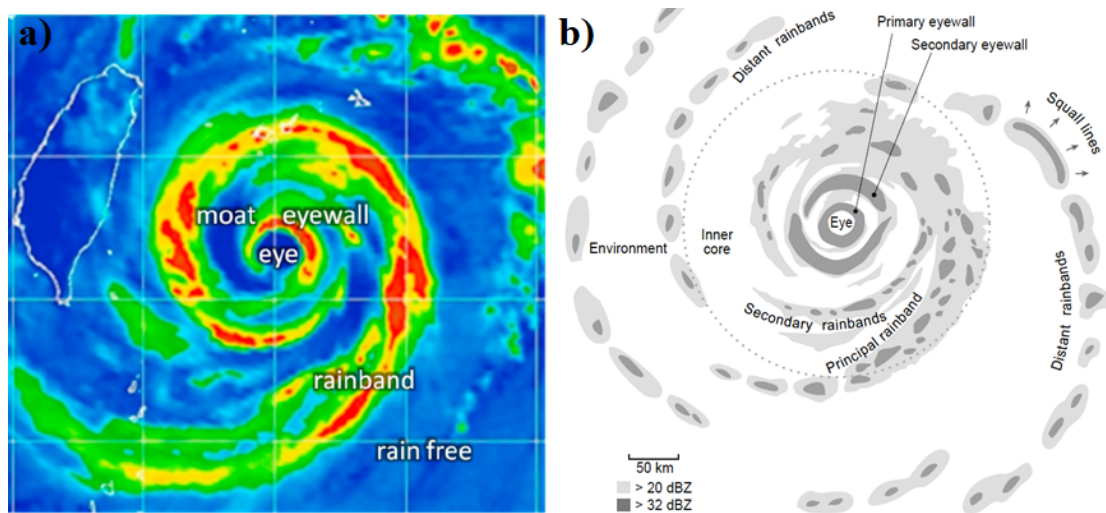


Figure 1.16: *Depiction of the activity and signature of the eyewall and rainbands. a) Example of microwave image of 89GHz brightness temperature from eyewall and rainbands. b) sketch of TC cloud structure and the basic reflectivity observed. Extracted from [Chou et al. 2013] and [Houze 2010].*

If this cloud system is fueled by the evaporative flux from warm tropical ocean and initiated by pressure instability, there is another essential parameter for the cyclonic circulation to occur: the vorticity [E.Palmen 1948]. As we saw in 1.1.3, a pre-existing cluster evolving in an unstable troposphere is necessary for cyclogenesis, providing strong baroclinic instabilities through the latent heat release that increases the relative vorticity ( $\zeta$ ) [Gray and Shea 1973]. Coupled with the planetary one, it drastically reduces the radius of deformation  $L_R = \frac{NH}{\zeta + f}$  to  $O(10\text{km})$ , making the system geostrophically adjusted [Beucher 2010]. As a result, the strong winds converging on the TC eye depression rotate counterclockwise (Northern hemisphere) due to the Coriolis effect (Fig 1.15). This tangential circulation is known as the primary circulation, and depicts the TC wind field. Along the full vertical column, the strongest winds are usually observed in the first few hundred meters [Powell et al. 2003] (300-500m), free from the effect of surface friction. The primary and secondary circulations represent two different aspects of the motions inside the cyclone, the latter allowing the existence of the former and the maintenance of the structure and its vorticity [Emanuel 1995]. They are bounded together via the radial distribution of a variable inherent to the dynamics of the cyclone: the angular momentum (AM) [Merrill 1984]. It can be expressed as follow:

$$\underbrace{M}_{\text{absolute ang. momentum}} = \underbrace{rV}_{\text{relative}} + \underbrace{1/2fr^2}_{\text{earth}} \quad (1.8)$$

The evolution of this quantity offers another reading of the low level circulation. Without external constraint, the angular momentum is a conservative variable [Riehl 1963], which translates into an increase in speed for small radii to satisfy conservation. At the surface, this transfer between planetary (i.e. earth: EAM) and relative angular momentum (RAM) is not lossless [Holland and Merrill 1984]. While  $M$  is maximum at the edge of the storm, where  $r$  is the largest and the RAM is 0, there is a loss of AAM (absolute angular momentum) during the inward motion of air parcels to TC center, due to the frictional forces applied by the surface layer [Chavas et al. 2015]. The remaining momentum is converted from planetary to the primary storm circulation up to the radius of maximum wind ( $R_{\text{max}}$ ), where the flow ascends and move outward aloft, like we have just described



with the Figure 1.15. This partial conservation of AAM connects the dynamics of the wind circulation to the different thermodynamic regimes in place [Chavas et al. 2015].

These different regimes are again associated to the inner core and the outer core. Albeit being part of the same entity, they are thermodynamically distinct. The first one joints the relative TC vorticity to the intense vertical motions of the eyewall [Emanuel 2004] and the second to the free troposphere influence (radiative cooling) [Emanuel and Rotunno 2011]. The parameterization of these two equilibria allows the surface wind field to be described [Chavas et al. 2015] and will be detailed with the introduction of the wind model in section 1.5.5 and chapter 4.1. The two regions are also independent of each other in terms of their properties and temporal fluctuations [Weatherford and Gray 1988a], the evolution of the inner-core can hardly be inferred from outer region. Three criteria are generally used to describe the structure of the surface field: the intensity, the size and the strength as defined by Merrill [1984]. The intensity is descriptive of the inner core, it is determined by the maximum wind speed ( $V_{max}$ ) and its associated radius ( $R_{max}$ ) [Gray and Shea 1973]. For a long time, size was equated with the radius of the maximum winds, the forcing was then primarily prescribed by the area of the strongest intensity. Since the advent of wind retrieval from spaceborne MW instruments, special attention has been paid to the size and strength parameters in order to improve the description of the TC forcing, which is needed in risk analysis, as wind waves are for example sensitive to the sustained area of strong winds. Size has been since associated with distant radii from the outer core such as the radius of 15 ( $R_{15}$ ) or 17 ( $R_{17}$ )  $m.s^{-1}$  [Chavas and Emanuel 2010; Chan and Chan 2012], or the radius of the outermost-closed isobar (ROCI) [Merrill 1984; Knaff et al. 2014]. As for strength, its definition fluctuates: as the averaged wind speed from  $R_{max}$  to gale radius [Merrill 1984], or the averaged wind from  $1^{\circ}$ - $2.5^{\circ}$  latitude radius from TC center,  $1^{\circ}$  radius being an approximate boundary used to define the extent of the inner core [Weatherford and Gray 1988a]. Of course, the boundary between these two regions of the cyclone is not so strict in reality, a transition zone exists between the deep-convective realm and the outer region. These three parameters nevertheless obey different mechanisms and are poorly correlated, the couple  $V_{max}/R_{max}$  and the outer structure follow unrelated variations along the TC lifetime (i.e intrastorm variability) [Chavas and Lin 2016]. In contrast, cross-comparison between different TC events (interstorm variability) showed some covariance between the deep structure ( $R_{max}$ ) and external radii size, meaning that large systems also tend to have larger eyewalls [Chavas and Lin 2016]. Similarly, some relations are found between intensity, size and strength when the TC phase (intensification ...) is taken into account [Weatherford and Gray 1988b; Kimball and Mulekar 2004; Chan and Chan 2013, 2015b].

Like the ocean we described earlier, the cyclone is a mini-system in its own right involving a complex range of processes that set its structure and circulation. The different variables describing the shape of the wind field respond to different dynamics and modes of variability (Interstorm and intrastorm) [Chavas and Lin 2016]. Several factors dictate intensity, strength and size mutations and have been listed in Chan and Chan [2013] and Chan and Chan [2015b]. They involve angular momentum transport [Merrill 1984; Chan and Chan 2013], planetary vorticity [Chan and Chan 2014; Knaff et al. 2014], SST [Lin et al. 2015], initial vortex size [Chan and Chan 2013, 2014; Chavas et al. 2016], environmental parameters (humidity, shear) [Hill and Lackmann 2009; Maclay et al. 2008], entropy flux [Xu and Wang 2010] and intensification [Weatherford and Gray 1988b; Knaff et al. 2014] ... . Capturing the different variabilities of the wind field is thus not straightforward. Several parametric models have tried to represent the complete surface wind profile with a collection of simple parameters, we will now see some of them.

### 1.2.2 Tropical Cyclone surface wind field

Along our story, the primary circulation has been essentially reduced to the surface field. If this low-level circulation emanates from the 3D storm [Riehl 1950], the forcing and exchanges of momentum are done through its surface wind stress. Hence from now, we will only see TCs as a 2D field as we mainly focus on air/sea interaction and ocean mixing.

Let us start by considering the circulation as perfectly axisymmetric around the eye center, the 2D field can thus be described via its radial wind distribution [Riehl 1963]. This profile can be decomposed into two main trends described by a Rankine vortex [Deppermann, S.J. 1947]. An ascending part in solid body rotation, representative of the winds going from the eye to the maximum radius of the winds located in the eyewall [Holland 1980], where the convection is the deepest, at a distance of about 20-50km. This part corresponds to a very sudden rise of the winds and can thus be approximated by a linear behavior. Then, a smoother descending part follows a hyperbolic trend, that describes the progressive fall of the winds from the eyewall to the TC extent. The description of these two sections by the Rankine model is expressed as follows:

$$V(r) = V_{max} \begin{cases} r/R_{max}, & \text{for } r \leq R_{max} \\ R_{max}/r, & \text{for } r > R_{max} \end{cases} \quad (1.9)$$

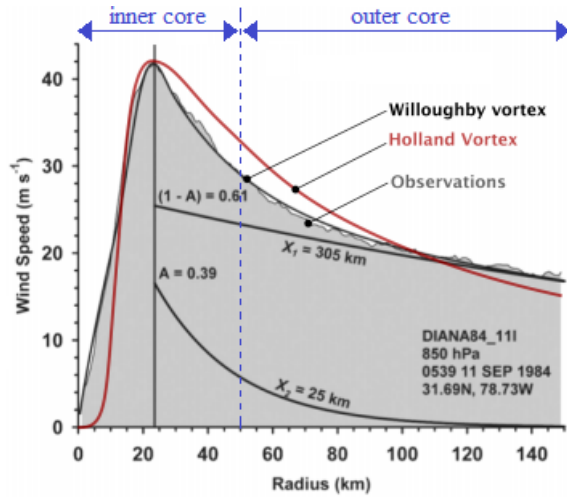


Figure 1.17: *Radial transect of TC from SFMR flight and two parametric models. Extracted from [Willoughby and Rahn 2004].*

Of course, this model just helps us to sketch the main pattern of the radial wind profile, but it is certainly too simplistic for representing the eyewall slope. As we have just seen, the eyewall is a region of few tens of kilometers, that concentrates the strong TC winds and makes the transition between the upward and downward trend of the wind profile. This region is thus characterized by a smooth shoulder-shaped curve (Fig 1.17), unlike the steep peak usually given by a Rankine vortex (see Deppermann, S.J. [1947]) Other models, Rankine-like, are specially dedicated to the parameterization of the cyclonic wind field and try to reproduce the different parts of the profile [Lei and Chen 2005], including the transition area around the eyewall.

The Willoughby model [Willoughby et al. 2006] is also a vortex based on empirical analysis of the TC curve, that was statistically fitted with observations and provide solutions for the three regions described (see Fig 1.17):

$$V(r) = \begin{cases} V_{in} = V_{max}(r/R_{max})^n, & \text{for } r \leq R1 \\ V_{out} = V_{max}[(1-A)e^{-(r-R_{max})/X1} + Ae^{-(r-R_{max})/X2}], & \text{for } r \geq R2 \\ V_{wall} = (1-w)V_{out} + wV_{in}, & \text{for } R1 \leq r \leq R2 \end{cases} \quad (1.10)$$



where  $n$  describes the growth of the winds in the eye;  $X1$  and  $X2$  are two folding lengths of the winds decay, the  $A$  sets proportion between the two exponential terms of  $V_{out}$  and  $w$  is a weight function that ensures the smooth connection between the winds in and out of the eye (Fig 1.17). The rising part is still written as a Rankine with a power law almost equal to 1. The descending part of the winds follows a decay composed of two exponential terms, both of which serve to describe the two gradients generally observed, i.e. the fast decrease after the eyewall ( $X2$ ), and the soft decay along the outer core ( $X1$ ) [Willoughby et al. 2006].

This three-part description can also be seen from a momentum perspective. In the first few kilometres, deep inside the cyclonic vortex, the winds are under cyclostrophic (Pressure+Centrifugal) balance until a radius of about  $\lambda_R$ , after which the Coriolis force starts to become strong and in equilibrium with the two others forces [Beucher 2010]. This intermediate state is suited to depict the fast decay following the eyewall. Once past this strong gradient, far from the highly constrained core rotation, the winds are under geostrophic equilibrium (Coriolis + Pressure) and observe a slow decay. These different balances neglect the friction acting on the boundary layer, and are more representative of the circulation aloft (at flight-level) [Knaff et al. 2011b]. The intermediate state, called gradient wind equilibrium, has often been used to depict the full radial distribution of the wind field [Holland 1980], as is the case in the widely used Holland model (see Fig 1.17). This is a gradient wind balance-like model, also based on an exponential decay law of the surface pressure [Schloemer 1954]. It can be expressed as:

$$P(r) = P_c + (P_n - P_c)e^{-A/r^B} \quad (1.11)$$

$$V_g(r) = \sqrt{ab(P_n - P_c)\frac{e^{-A/r^B}}{\rho r^B} + (rf/2)^2} - rf/2 \quad (1.12)$$

$B$  is a scaling parameter that monitors the shape of the profile and  $A$  its location, they are linked through the relation  $R_{max} = A^{1/B}$ . This Holland-B parameter has been modified several times to better describe the slope around the eyewall, it is even estimated by operational to quickly probe wind-pressure relationship [Knaff et al. 2011b], as this model originally infer  $V_{max}$  from the pressure gradient [Holland 1980]. In addition to the radial description of cyclonic winds, this model incorporates several aspects of the 2D wind field. It introduces an inflow angle to reproduce the surface circulation, that slightly changes the inclination of the wind radii vectors. Besides, this model breaks with the axisymmetry approximation, it incorporates the influence of the translation speed in the TC flow [Jelesnianski 1966], which is the main driver of the wind field asymmetry at first order, at least for the inner region [Knaff et al. 2007, 2016]. The total wind field is then expressed as the sum of the cyclonic wind field and the moving compo-

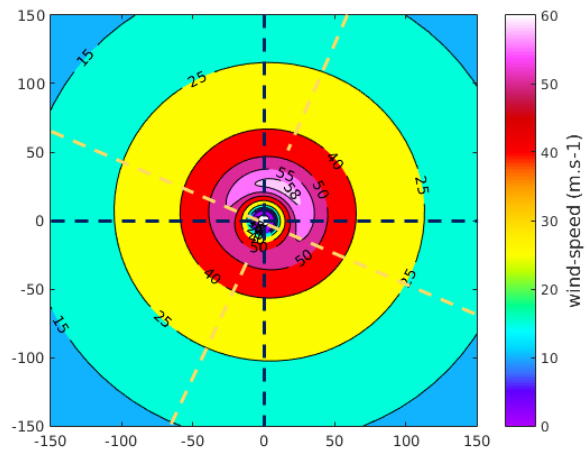


Figure 1.18: *surface wind-field derived from Holland model for Hector at the same time as Figure 1.14*

nent from the environmental advection [Pan et al. 2016]. Depending on the orientation of the wind field and the heading direction of the storm, this contribution may increase or decrease the measured wind values. This effect is also dependent on the distance from TC center, following the Jelesnianski [1966] formula, with enhanced activity at Rmax region:

$$V_{mov} = V_{fm} \frac{R_{max} \times R}{R_{max}^2 + R^2}. \quad (1.13)$$

with  $V_{mov}$  the component from TC motion,  $V_{fm}$  the vector of the translation speed. As we go along the outer radii ( $R \gg R_{max}$ ), the asymmetries advected by the storm motion fade. The Figure 1.18 shows the wind field derived from the Holland model for the Hector case at the same date as the previous satellite picture. At this stage, Hector was moving westward with a slight deflection to the North ( $150^\circ$ ) and had a fast translation speed which induces a strong polarisation of its maximum winds to the right of its displacement. This crescent shape is characteristic of the asymmetries present in the inner core, the strongest winds are generally observed to the right(/left) of the cyclone track (North/South hemisphere), where the wind vectors associated with the rotating system are perfectly aligned with the translation vector [Knaff and Harper 2010; Uhlhorn et al. 2014; Klotz and Jiang 2017]. Thus, the surface wind field is not solely based on its structure (Rmax, outer radii) and intensity ( $V_{max}$ ) parameters, but also by its translation speed which affects these parameters azimuthally. Several operational analyses discretize the TC into 4 quadrants to represent this distribution. They essentially follow two conventions: Geographic (black-dashed line in fig 1.18), like in the Best-Track, the reference frame is set according to the North, all cyclones are split into quadrants of the same orientation (NE, SE, SW, NW). Cyclonic, the reference frame is adjusted according to the heading direction and is thus specific to the storm (yellow-dashed line in fig 1.18), each quadrant is shifted according to the direction angle (FR: front-right, RR: rear-right, RL, FL) [Kudryavtsev et al. 2015]. While the former eases the cross-comparisons, the latter is particularly relevant for studying TC-induced ocean response. This convention will always be specified throughout our study.

Through this succession of parametric models, we were able to progressively address the different variables inherent to the primary circulation of the cyclone, from its radial profile to its 2D aspect. Several parametric models have been developed, whether they are semi-empirical [Holland 1980; Holland et al. 2010; Wood et al. 2013], climatological [Knaff et al. 2007, 2016] with different constants depending on the basins concerned [Knaff et al. 2018], or whether they are based on statistical fits [Willoughby et al. 2006] or on pre-existing models such as Rankine-like [Riehl 1963; Rappin et al. 2013] or Holland-like vortex. They are all conditioned by their starting hypotheses and approaches, the Holland model with its gradient wind equilibrium is much more representative of very intense and compact TCs because of the rapid decay of its winds [Willoughby and Rahn 2004], or by the observations from which they have benefited, but also and above all by the parameters on which they depend. Most of the parametric solutions discussed here are Rmax-based models [Knaff et al. 2015]. This radius is difficult to infer from most of the observations, and its accuracy can have a drastic impact on the model. Therefore, the last decade has seen the emergence of new models forced by the information of an external radius (like R17, R15), easily documented by most observations [Herndon et al. 2010; Knaff et al. 2021], one of which will be studied in Chapter 4. Regardless of the parametric solution chosen, they all require prior input. In addition, to probe the vital parameters of the TC forcing, satellite observations allow a global monitoring and unparalleled coverage of the large-scale storm circulation, necessary for operational forecasting and analysis [Rappaport et al. 2009], and to investigate the wind field and its different properties.

### 1.2.3 Tropical cyclone wind field observation

Due to the transient and cloudy nature of tropical systems, near-surface cyclonic conditions were for a long time confined to local measurements on land and on ships [Truchelut et al. 2013]. While they allowed the first maps of extreme surface winds to be made and the first theoretical models to be developed, they were clearly deficient for global monitoring and record analyses (especially of the Pacific basins), or for the development of climatology and 2D wind field mapping. Although the 1940s saw the advent of the first aircraft reconnaissance flight, which are still a source of reference for others observations today [Sapp et al. 2019], their sampling was too sparse, focused on coastal events, and localised to specific basin. This is still the case, but efforts are actually on-going to test dedicated long-range small uncrewed, remotely controlled aircrafts (drones). Prior to satellite era, an undefined but substantial number of cyclones could therefore not be recorded, especially in predominantly open ocean basins due to the scarcity of observations, such as in the eastern Pacific [Truchelut et al. 2013]. With the arrival of geostationary instruments in the mid-1960s and their proliferation in the 1970s, satellites started a revolution in tropical cyclone analysis, which has continued with the multiplication of MW instruments and new measurement techniques [Knaff et al. 2021].

Several methods have then been developed to estimate the TC parameters in terms of structure, intensity and to derive a relatively large fraction of its global wind field, depending on the resolution and the nature of the measurement [Herndon et al. 2010; Knaff and Harper 2010]. While all satellite-derived wind information stem from indirect measurements, the techniques used can be divided into two main groups: the extraction of parameters directly inferred from the surface wind field retrieval (called direct method), and the indirect estimation of these parameters by atmospheric observations of the cloud system [Knaff et al. 2021]. This second category includes all the techniques based on IR top-cloud [Mueller et al. 2006; Velden et al. 2006; Kossin et al. 2007; Knaff et al. 2014] or at level-flight measurements [Demuth et al. 2004; Knaff et al. 2015] that do not retrieve surface wind field, but attempt through regression and empirical laws [Kossin et al. 2007; Knaff et al. 2014], parametric vortex [Mueller et al. 2006; Knaff et al. 2007, 2016] or cloud pattern recognition [Velden et al. 2006; Olander and Velden 2007], to deduce intensity and reconstruct the structure of the cyclone with wind radii estimations [Knaff et al. 2011a]. The implementation of these different approaches requires the extraction of several criteria from the satellite scenes. Infrared techniques, for example, focus on the distribution of brightness temperatures along the cloud structure [Mueller et al. 2006] and in certain patterns such as the eye size [Kossin et al. 2007; Lajoie and Walsh 2008], to develop empirical laws on  $R_{max}$ ,  $V_{max}$ , wind radii [Demuth et al. 2006] or any other reference radius that will then allow, by means of a parametric model, to complete the structure [Knaff et al. 2016]. In this respect, the Rankine-like vortex is widely used to extrapolate the outer radii at each quadrant, due to its simplicity and malleability [Mueller et al. 2006; Knaff et al. 2007]. Similarly, MW sounders guess the TC intensity with regression laws documented by the atmospheric temperature anomalies of different vertical levels [Demuth et al. 2004, 2006], retrieved from their multi-frequency channels. By collecting criteria from different altitudes, wind radii can also be derived from statistical procedure. Other indirect methods include the cloud drift winds from visible geostationary imagery [Dunion et al. 2002], which tracks low-level clouds to deduce the outer circulation only (cirrus canopy issues). These indirect methods have long dominated the analysis and reanalysis of TC vitals in operational product like BTK, in parallel with in-situ measurements (platforms, buoys, coastal radar...) and reconnaissance flights when available [Knaff et al. 2016; Sampson et al. 2017]. In their absence, parameter estimations are strongly dependent on these spaceborne instruments [Sampson and Knaff 2015]. For instance, TC intensity is mainly

constrained by the Dvorak analysis, which associates a T-number to each visible/IR scenes through cloud pattern recognition, that can be further related to an intensity [Velden et al. 2006]. Recent efforts have also been put in combined products to provide guidance for analysis and forecasting, Algorithm like SATCON [Herndon et al. 2012; Velden and Herndon 2020] blends MW sounders, pressure-wind relation, and different IR-technics for intensity estimates, such as the advanced Dvorak analysis (ADT) [Olander and Velden 2007; Velden et al. 2017] which also used 89 GHz MW imagers to depict eye size [Wimmers and Velden 2007, 2016]. Under this impetus, a multiplatform analysis even gathers those atmospheric observations with MW surface wind measurements [Knaff et al. 2011a].

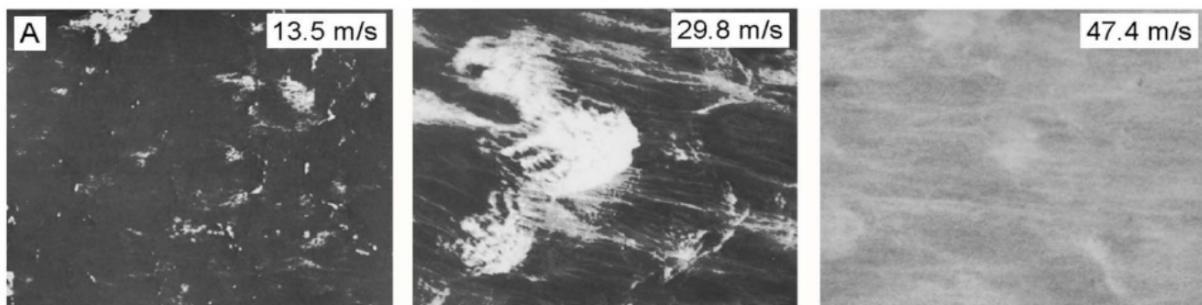


Figure 1.19: *White caps and streaks under medium to extreme wind speed conditions from low-level air reconnaissance flights. Extracted from [Holthuijsen et al. 2012].*

Over the last two decades, the satellite landscape has been mainly marked by the rise of wind-retrieval MW instruments and their progressive intrusion into operational analyses, with the arrival of scatterometers such as Quikscat for TC center fix [Brennan et al. 2009], its peer ASCAT, and more recently SMAP and SMOS radiometers [JTWc 2017; Bushnell et al. 2018], especially to estimate for wind radii in different quadrants of the TC. The unique ability of these imaging instruments to witness the wind imprints at the surface, is essentially based on the relative transparency of the atmosphere at these wavelengths, and the sensitivity of the interface emissivity and reflectivity properties under high wind forcing conditions [Katsaros et al. 2002]. In this regard, radiometric measurements of the ocean are notably receptive to wind speed fluctuations, for frequency higher than 6 GHz (see Fig 1.7). Since emissivity depends on both the dielectric and geometric components of the sea surface [Njoku et al. 2000], the brightness temperature can be separated into two terms:  $Tb_{flat}$  and  $Tb_{rough}$  [Dinnat et al. 2003]. In order to measure and isolate this second component, most instruments use several channels in the ranges most sensitive to wind variations, mainly the frequencies around 7, 11, 19 and 37 GHz (as GMI, Windsat, AMSR-E, AMSR-2). Recently, new L-band instruments have emerged with capability to retrieve both SSS and U10: SMOS [Yueh et al. 2010; Reul et al. 2016] and SMAP [Meissner et al. 2017]. If the sensitivity of  $Tb$  to wind decreases at this wavelength, the one associated with SST is even lower. Using contrast ( $Tb$  difference between smooth and rough ocean) of both polarisations and ancillary observations of temperature and salinity, wind intensities can be estimated with an accuracy of a few knots [Reul et al. 2016]. The advantage of this new kind of radiometers is twofold. On the one hand, the radio-brightness contrast varies quasi-linearly with wind intensity and its sensitivity is heightened for winds above the hurricane-force threshold [Reul et al. 2012]. This increase in sensitivity is mediated by the excess emissivity of the expanding foam coverage, due to the appearance of whitecap and streaks (see Fig 1.19), which partially dominates the MW response of the ocean under extreme wind conditions and for the 1-37



GHz band [Droppleman 1970]. Knowledge of the geometric and dielectric properties of the foam layer [Anguelova and Gaiser 2012] is therefore necessary to be taken into account in the emissivity model [Reul and Chapron 2003] in order to correctly estimate the wind intensity [Reul et al. 2016; Zabolotskikh et al. 2016; Meissner et al. 2017]. On the other hand, while the signal sensitivity is stronger at C-band [Uhlhorn and Black 2003], the rain signature is much weaker for the L-band than any other MW spectral bands (two orders magnitude less than C-band) [Reul et al. 2012]. For shorter wavelengths, rain is one of the main limitations to the accurate inversion of surface winds, as it involves different mechanisms that interfere with the signal (absorption, emission, scattering) [Wentz et al. 2000; Alpers et al. 2016] and is often difficult to differentiate from the wind signature itself, especially when using a single-frequency sensor [Quilfen et al. 1998].

Some multi-frequency radiometers, such as AMSR-2 [Zabolotskikh et al. 2014], are equipped with a high-frequency 89 GHz channel, which is sensitive to hydrometeors and useful for mitigating the effects of rain when coupled with lower bands [Imaoka et al. 2010]. Likewise, the C-band radiometer on-board reconnaissance flights, SFMR, uses six different frequencies ([4.5,7.2] GHz) to investigate surface wind field and the rain effects at high resolution <sup>5</sup>. Reul et al. [2012] showed that these effects were negligible for the 1.4GHz band up to 45-50 mm.h<sup>-1</sup>, later confirmed by direct comparison between SMAP and SFMR for different rain rates [Meissner et al. 2017]. These statements have been somewhat softened for two situations: i) in case of extreme precipitation (>60 mm.h<sup>-1</sup>), L-band may be weakly affected [Reul et al. 2012, 2016], though the size of these cells is generally much smaller than the actual resolution of the satellite footprint [Begum and Otung 2009; Reul et al. 2012] and wavelength [Uhlhorn and Black 2003], ii) in case of heavy precipitation at low wind speeds [Tang et al. 2013; Reul et al. 2016], for which the variability of Tb is smaller and therefore more easily contaminated by rain. Under this regime, an unmixed fresh lens may also occur and change the local SSS inducing errors in the radio-brightness contrast and the wind-retrieval. Thereupon, quality of the L-band product is more erratic in region of sharp SSS variability like river plumes, and under light winds (<12 m.s<sup>-1</sup>) due to tenuous signal-to-noise ratio coupled with this low Tb sensibility [Reul et al. 2020]. Apart from these caveats, radiometers are particularly interesting tools when it comes to explore the TC wind field, several studies have proven their capability to handle winds up to 70 m.s<sup>-1</sup> [Meissner et al. 2017], but this ability can be hampered by their low resolutions (30-50 km), possibly preventing them from accessing the strongest winds and limiting their descriptions of the inner-core to larger systems [Reul et al. 2017; Zhang et al. 2021].

While C/Xu band radiometers have better overall accuracy ( $\leq 1\text{ m.s}^{-1}$ ) [Meissner and Wentz 2009], L-band high winds are not degraded by rain and provide stronger guidance. To conclude on radiometric properties, the Figure 1.20 shows a composite image made from SMAP, SMOS and AMSR-2 acquisitions to illustrate the TC wind field story of the “three brothers” (Ignacio, Jimena and Kilo, Reul et al. [2017]). It demonstrates and synthesizes the capabilities of these instruments to monitor and describe a significant fraction of the cyclonic wind field. In addition to providing a broad spatial and temporal sampling, the different circulations are well captured as well as part of the strong winds of the inner core up to 45 m.s<sup>-1</sup>, which represents from 70 to 85% of the maximum wind speed observed for these TC cases. However, the eye and eyewall regions cannot be resolved. It is also important to note that radiometers deal with wind as a scalar variable only, the observation of the vector field requires another type of measurement.

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<sup>5</sup>This source of data will be used as ground truth in Chapter 3, more details will be provided

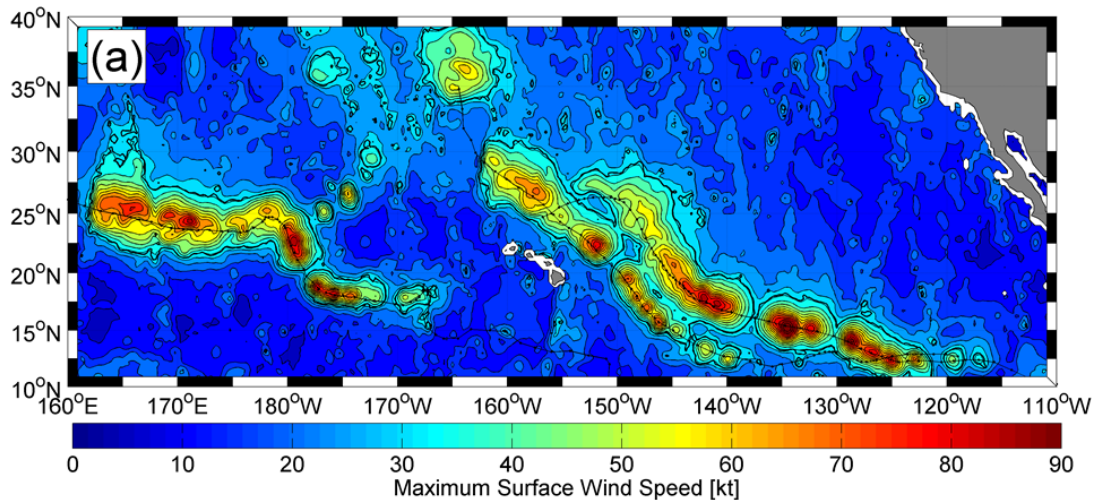


Figure 1.20: *Blended surface wind fields derived from the combination of radiometers SMAP/SMOS and AMSR2, along the track of Ignacio, Jimena and Kilo in 2015.*

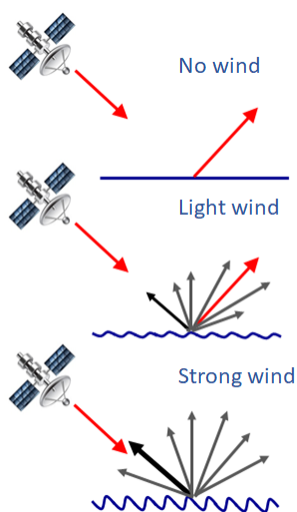


Figure 1.21: *Bragg-scattering of MW pulse. As the wind increases, the reflection is more diffuse and more signal is backscattered to the sensor.*

Complementary to MW radiometer surface winds observation, there are the active sensors like scatterometers. Unlike previous instruments, they can infer both intensity and direction of the ocean surface wind vector, thanks to the nature of their observations and their multi-looking approach (watch similar area at different azimuth angles) [Naderi et al. 1991]. Instead of gauging the natural emission from the ocean, scatterometers send a MW pulse and measure the intensity of the backscattered signal from the sea surface [Jones et al. 1982; Naderi et al. 1991]. The reflection is strongly dependent on the geometry of the interface, the ocean roughness, and is largely controlled through interactions with structures of similar size to the wavelength of the pulse, i.e capillary-gravity waves and breakers, generated by the atmospheric forcing to characterise this overall directional sea surface roughness [Wright 1966; Moore and Fung 1979]. Gravity-capillary waves and breakers essentially trace the stresses exerted by the wind, their growth and number proportional to its intensity. The stronger the forcing is, the rougher the ocean becomes, more in-

teractions occur with gravity-capillary waves and breakers which enhance the backscattering through resonant bragg and non-resonant mechanisms for low to moderate incident angle [Phillips 1988], like illustrated in the Fig 1.21.

The return signals received by the sensor are translated in terms of radar cross section [Naderi et al. 1991]. It depicts the equivalent area seen by the radar which received the incident energy and scattered a radiation equal to the energy received by the radar [Martin 2013]. This variable is generally normalized by the resolution cells to deal with homogeneous areas, it is the so-called sigma naught ( $\sigma_0$ ), it is expressed in decibels (db) and represents a measure of the sea surface roughness [Martin 2013].  $\sigma_0$  can be seen as the benchmark variable for active instrument, equivalent to the brightness temperatures for passive remote sensing. If the sea surface roughness is a proxy of wind stress, the



power of the backscattered signal is not only dependent on its intensity but also on its direction and on the incidence angle of the signal [Phillips 1988]. The empirical law that retrieves the wind speed from  $\sigma_0$  is called the geophysical model function (GMF) and is established for a given frequency and polarisation (like CMOD5 for C-band in VV polarization, Hersbach et al. [2007]). Most deployed scatterometers operate at C (Ascat-A/B) or Ku (Quickscat) band, they provide an average resolution of 12.5 to 25 km which can possibly allow to depict the eye and the approximate area of the strong winds [Brennan et al. 2009; Herndon et al. 2010]. While scatterometers provide better resolution than radiometers and do not limit the retrieved parameters to scalar values, their inversions in terms of wind field can be compromised by multiple interfering effects: rain, large gradients within a particular cell, and also the possible saturation of the backscattered signals, especially co-polarized ones, for intense winds [Quilfen et al. 1998].

Like for radiometers, rain is a limiting factor. Rain can involve a series of heterogeneous mechanisms with antagonistic effects [Alpers et al. 2016] that depend on wind intensity and on rain rates [Tournadre and Quilfen 2003; Chou et al. 2013]. Rain is thus an important parameter to deal with to ensure the quality of the wind estimates [Quilfen et al. 2010]. Note, impacts on the surface can compete directly, through ring waves generation, scattering of splash products, and downdraft winds associated to rain cells, with the signature of wind forcing on sea surface roughness, and affects the reflectivity more broadly via signal interactions with airborne raindrops (termed volume scattering and attenuation) [Alpers et al. 2016]. Tropical depressions are an example of low intensity systems that can be the seat of severe rainfall, and for which the backscattered signal can be completely overwhelmed by the rain signature [Brennan et al. 2009]. However, the acuteness of these effects remains strongly related to the instrumental frequency [Quilfen et al. 2010], the C-band being indeed ten times less impacted than the Ku band measurements [Quilfen et al. 1998; Tournadre and Quilfen 2003]. They also react differently, while the Ku-band is strongly affected by raindrops interaction [Tournadre and Quilfen 2003], the C-band remains insensitive and is more impacted by surface scatterers [Nie and Long 2007]. A general pattern can be still observed with an enhancement of the NRCS at low winds, an attenuation of the signal at high winds and compensated effect at moderate winds<sup>6</sup> [Tournadre and Quilfen 2003; Chou et al. 2013]. Several methods are used to cope with rain, such as the use of dual-frequency sensors to isolate different contributions to sea surface roughness [Quilfen et al. 2006], auxiliary observations of rain reflectivity to flag contaminated cells [Fulton et al. 1998; Wimmers and Velden 2007; Huffman et al. 2019], or to guide rain models, that tend to mitigate its effects in wind inversion algorithms [Nie and Long 2007; Chou et al. 2013]. Measurement of strong winds within tropical cyclones can thus remain quite elusive, since precipitation intensity often follows wind intensity [Rodgers and Adler 1981; Lin et al. 2015], in addition to signal saturation for co-polarized measurements. Extreme rainfall ( $>50 \text{ mm.h}^{-1}$ ) are often recurrent near the eyewall of intense TCs, for which even C-band instrument experience volume scattering and attenuation by raindrops [Alpers et al. 2016]. Yet, scatterometers, thanks to their wide swath configurations, remain a valuable asset for operational analyses, especially for wind closed-contour radius estimations, and thus, eye tracking [Brennan et al. 2009; Sampson et al. 2018]. Often reported, Ascat-A/B/C C-band VV instruments often display very good overall performance in the estimation of wind vectors ( $2 \text{ m.s}^{-1}$  and  $20^\circ$  errors) up to  $25 \text{ m.s}^{-1}$  [Bentamy et al. 2008; Verspeek et al. 2010; Chou et al. 2013].

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<sup>6</sup>For low-to-medium rain rates. For high rain rates, scattering predominates [Alpers et al. 2016]

### 1.2.4 Operational capabilities and convention.

The satellite constellation is very protean and is enriched over the years by a multitude of heterogeneous instruments (such as CYGNSS, [Morris and Ruf \[2017\]](#)) that constantly push the limits of observations and multi-modal analysis of this virtual multi-satellite constellation. We have only mentioned here the most used sensors for our study, in order to give a general picture of space oceanography and to share the complexity of its observations. Among them, MW sensors stand out as all-weather condition tools, allowing to monitor surface variables independently of cloud cover and with a relatively reduced sensitivity to rain (especially for bands ranging from 1Ghz-8Ghz). They will be our most valuable assets throughout the different chapters. Indirect methods, i.e. based on top-atmosphere measurements (like IR methods [\[Olander and Velden 2007\]](#)), will not be used in this thesis although they are hidden behind the Best-Tracks analyses that we will use and which are mainly based on their performances. As we have seen in section 1.2.2, the cyclonic wind field is defined by its intensity, its size but also by the gradient of its winds. One of the objectives of these observations is to be able to inform this profile from a set of synthetic parameters, which the Best-Tracks do very well.

|                                     |                                            | Vmax | Rmax | R33 (R64kt) | R25 (R50kt) | R17 (R34kt) |
|-------------------------------------|--------------------------------------------|------|------|-------------|-------------|-------------|
| Satellite-based Indirect methods    | Dvorak                                     | ●    | ●    | ●           | ●           | ●           |
|                                     | ADT                                        | ●    | ●    | ●           | ●           | ●           |
|                                     | IR TECHNIQS                                | ●    | ●    | ●           | ●           | ●           |
|                                     | MW SOUNDERS                                | ●    | ●    | ●           | ●           | ●           |
|                                     | CLOUD DRIFT VECTORS                        | ●    | ●    | ●           | ●           | ●           |
|                                     | SATCON                                     | ●    | ●    | ●           | ●           | ●           |
|                                     | BTK                                        | ●    | ●    | ●           | ●           | ●           |
| Surface wind-derived remote sensing | MW IMAGER RADIOMETER                       | ●    | ●    | ●           | ●           | ●           |
|                                     | SCATTEROMETER                              | ●    | ●    | ●           | ●           | ●           |
|                                     | SYNTHETIC APERTURE RADAR                   | ●    | ●    | ●           | ●           | ●           |
| Others                              | SFMR                                       | ●    | ●    | ●           | ●           | ●           |
|                                     | IN-SITU<br>(land station, platform, buoys) | ●    | ●    | ●           | ●           | ●           |

Figure 1.22: *Synthesis of the different capabilities of operational products and instruments to retrieve TC vitals. Labels stand for: green=available, blue=mostly available, orange=limited or large errors, red=unavailable.*

The figure 1.22 summarizes the contrasting operational capabilities of the instruments, to extract the main TC vitals that we will use in this manuscript. While the three characteristic wind radii received later interest than the historical BTK parameters (Vmax trajectory), they have become established in most forecasts and analyses and are the subject of reanalysis within the BTK [\[Knaff et al. 2007; Sampson and Knaff 2015; Sampson et al. 2017\]](#). Since the last decade, these radii are observable by most of the existing sensors and benefit from a large diversity of measurements [\[Herndon et al. 2010; Knaff et al. 2021\]](#). However, the performance decreases with the intensity of the wind radii [\[Landsea and Franklin 2013\]](#), as for winds higher than 25m.s<sup>-1</sup>, scatterometers are no longer involved in the procedure due to saturation issues [\[Brennan et al. 2009\]](#). R25 and

R33 are then strongly dependent (especially in open sea) on other satellite methods of higher frequency (IR / Visible, MW sounders), which can significantly increase the associated uncertainties [Landsea and Franklin 2013], like R17 in the absence of scatterometer measurements [Sampson et al. 2018]. To this end, L-band radiometers have been recently incorporated into BTK [Bushnell et al. 2018] to help existing capabilities provide better quality wind radii, and can even support Vmax estimates for broad systems [Reul et al. 2017; Knaff et al. 2021].

The intensity itself relies almost exclusively on top atmosphere observations (see fig 1.22) via Dvorak analyses [Velden et al. 2006; Velden and Herndon 2020]. Some conflicting situations are known to strongly affect their evaluations, such as eye definition issues that can hamper cloud recognition patterns [Velden et al. 2006] or jeopardize classic scheme of pressure-wind relationships [Demuth et al. 2004; Kossin 2015], whether caused by cirrus cover [Olander and Velden 2007], by the absence of a well-defined structure (the occurrence of an eye is generally around 28 m.s<sup>-1</sup>, Vigh et al. [2012]) or either by the presence of a misleading eye (like pinhole eye) or eyewall pattern that disturbs the analysis [Olander and Velden 2007; Vigh et al. 2012; Kossin 2015]. These geometric singularities can be the cause of drastic TC evolution, such as rapid intensification/weakening situations<sup>7</sup> or as ERC episodes (eyewall replacement cycle), during which two eyewalls coexist and evolve simultaneously before merging to give a larger inner core [Sitkowski et al. 2011]. They correspond to situations of high variability with abrupt structural and intensity changes occurring in several rapid phases [Maclay et al. 2008; Kossin and Sitkowski 2012]. The Dvorak procedure is limited for such sharp evolution as it follows some constraints that allow for a maximum change in intensity over a certain period of time [Sangster and Landsea 2020], these two mechanisms are thus a pitfall for prediction or analysis [Kossin and DeMaria 2016; Leroux et al. 2018]. Apart from these limiting events, Vmax is a parameter well identified by a set of instruments, unlike Rmax (Fig 1.22).

Indeed, the maximum wind radius does not benefit from the robust and long-time experience methodology of the Vmax, which is not deduced from an observation around the strong convective area but derives from relations naturally established with the cloud and Tb architecture, or from its relation with the pressure [Kossin 2015; Velden and Herndon 2020]. As mentioned by Kossin et al. [2007] and more recently by Knaff et al. [2021], while the outer core structure is dissected by a wide variety of observations, the inner core remains routinely accessible only to low-level aircraft reconnaissance, the emerging capabilities of L/C band radiometers and scatterometers can support, but still for high and/or low intensity cyclones [Reul et al. 2016]. Given the scarcity of surface information into the eyewall, operators turn to regression laws, using predictors based on the distribution of Tb [Mueller et al. 2006], on the eye size when it is clear [Kossin et al. 2007], or climatological relationships including Vmax, latitude [Quiring et al. 2011; Knaff et al. 2016] or external radii such as R50 [Takagi and Wu 2016]. These very indirect methods have the advantage of almost systematically providing an estimate, but they are for the most part very basin-dependent [Knaff et al. 2007; Quiring et al. 2011], uncorrelated with the Vmax estimation methods, suffer from large or poorly-estimated errors, and are not reanalyzed in BTK [Knaff et al. 2021].

In the face of these turmoils, synthetic aperture radar (SAR) offers a high resolution MW view capable of probing the TC inner core, but its exploitation for this purpose has long remained embryonic and marginal [Katsaros et al. 2000; Dagestad et al. 2013]. Several efforts have been made since then, and several of these sensors are now part of the

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<sup>7</sup>an intensity change of at least 15 m.s<sup>-1</sup> in a 24h-period [Kaplan and DeMaria 2003]

ocean remote sensing arsenal [Freeman et al. 2019], but they are still not ingested into tropical cyclone-oriented analysis products [Knaff et al. 2021]. Like scatterometers, SARs are active instruments that measure the return echo of the backscattered signal [Martin 2013]. They are called synthetic aperture because of their ability to simulate a very large antenna aperture using the Doppler shift induced by the relative motion of a small effective aperture moving along the scene of interest [Katsaros et al. 2002]. This technique allows SAR to observe mesoscale ocean features [Johannessen et al. 1996; Chapron et al. 2001; Wang et al. 2019] and atmospheric structure footprints [Thompson et al. 1983; Alpers and Brummer 1994; Foster 2005] on its surface at a very fine scale, while overcoming physical constraints and antenna sizing issues. Newer versions interpret not only the co-polarized signal, but also the depolarized <sup>8</sup> portion with what is called a dual-polarized (e.g VV+VH) or full polarimetric (HH,VV,HV,VH) instrument [Martin 2013]. When the transmitted and received echoes are in orthogonal polarization, it is called a cross-polarized signal. The notable interest of this signal is the tenuous interaction it has with breaking waves [Phillips 1988], it has been shown that the cross-polarized  $\sigma_0$ , unlike copolarized one, is strongly correlated with the rough sea foam emissivity of L-band radiometers [Zhao et al. 2018], which illustrates the progressive control of wave breaking on both emissive and reflective properties of the ocean surface under hurricane force winds [Zhang and Perrie 2012; Reul et al. 2012]. More information will be given about the stakes of this measurement in section 1.5.3, while the two articles in Chapter 3 and in Annex B will provide more details about SAR instruments and capabilities.

| Wind speeds                            |                     |                   |               |                |                 |
|----------------------------------------|---------------------|-------------------|---------------|----------------|-----------------|
| (for 1-minute maximum sustained winds) |                     |                   |               |                |                 |
| Tropical Cyclones                      | Category            | Convention<br>m/s | knots (kn)    | mph            | km/h            |
|                                        | Five                | $\geq 70$ m/s     | $\geq 137$ kn | $\geq 157$ mph | $\geq 252$ km/h |
|                                        | Four                | 58–70 m/s         | 113–136 kn    | 130–156 mph    | 209–251 km/h    |
|                                        | Three               | 50–58 m/s         | 96–112 kn     | 111–129 mph    | 178–208 km/h    |
|                                        | Two                 | 43–49 m/s         | 83–95 kn      | 96–110 mph     | 154–177 km/h    |
|                                        | One                 | 33–42 m/s         | 64–82 kn      | 74–95 mph      | 119–153 km/h    |
|                                        | Tropical storm      | 18–32 m/s         | 34–63 kn      | 39–73 mph      | 63–118 km/h     |
|                                        | Tropical depression | $\leq 17$ m/s     | $\leq 33$ kn  | $\leq 38$ mph  | $\leq 62$ km/h  |

Figure 1.23: *Summary of intensity categories of tropical systems according to different unit conventions.*

To conclude this section, it is important to clarify the conventions governing wind analysis. Actually, its nature is inherent to its temporal definition, both remote and surface instruments have a different sampling or scanning time, which represents an average value for a specific resolution cell. This value can stand for a mean made over a given time period, such as the 10-min winds, or can be a maximum averaged value collected over a longer observation time. For instance, the 1-min maximum sustained winds correspond to the maximum 1-min winds observed in a 10 min period. These two temporal definitions are the most commonly used, although the mean definition may vary from one centre to another (Indian centers use a 6-min definition), it is generally accepted that low-to-medium resolution MW instruments measure wind speeds that correspond to 10-min values. In contrast, the widely used Saffir-Simpson scale (see Fig 1.23) and most

<sup>8</sup>scattered in different polarization (e.g. H-polarized signal is backscattered in V-polarized state).

of the operational methods provide a 1-min maximum sustained winds. It is to this scale that we will mostly refer to, as Best-Tracks from ATCF are based on. For the units we will prefer  $\text{m.s}^{-1}$  to knots for intensities and km to nautical miles for distances. At last, whatever the measurement sources, they are all standardized and adjusted to provide a 10 meter height wind value, the so-called u10.



## 1.3 In the Wake of TC.

We have just addressed separately the two main components of our subject of study. On the one hand, the ocean, with a set of geophysical variables that describe its surface properties and vertical stratification (see 1.1.1). On the other hand, the surface cyclonic field, defined by its radial wind distribution, its asymmetry, and its advection along the tropical and subtropical basins (see 1.2). We also discussed the processes in charge of the pre-cyclonic ocean structure (1.1.3), which we saw to be dependent on seasonal mean surface fluxes, local disturbances and the various variabilities that are at work and which in turn affect the cyclonic activity. These different processes are indeed exacerbated during the passage of a cyclone and contribute to intense oceanic mixing, visible in remote sensing through the different markers of these dynamics.

This imprint left along the track is called the cold wake and represents the typical oceanic response to a tropical cyclone. It can be split into two phases [Price et al. 1994]. The first, called the forced stage, corresponds to the effective residence time of the cyclone and represents the local, direct and short time-scale answer to the wind stress field ( $\sim 1$  day). After its passage and during few inertial periods, a non-local response to the wind stress curl, called the relaxation stage, is set in the shape of a spreading wake of near-inertial waves that radiated out the energy injected from the storm. This two-stage response is related to the vertical mixing process and is followed by a long recovery period, where the inertial motions decay and the ocean gradually returns to climatological conditions (restratification). Both mixing and restratification will be described in the two following subsections.

### 1.3.1 Mixing process

As a tropical cyclone is an exceptionally strong and localised forcing, its surface stress and curl triggers strong mixing with multi-scale disturbances that excite different modes of ocean variability (e.g. barotropic/baroclinic motions) [Price et al. 1994; Ginis 2002], the energy transfer is mainly driven by turbulent mechanisms.

At the interface, it usually results in severe sea states of several metres ( $H_s$  of 20 m was measured for hurricane Julia in Kudryavtsev et al. [2015]) and a local storm surge that move with the low-pressure field. Numerical [Jelesnianski 1966; Irish et al. 2008; Moon et al. 2003] and analytical [Young 1988; Kudryavtsev et al. 2015] models have diagnosed them from the forcing conditions, as a function of the intensity ( $V_{max}$ ), size ( $R_{max}$ ) and forward speed motion of the storm ( $V_{fm}$ ) ([Young and Vиноth 2013; Irish et al. 2008; NHC 2016]). Indeed, the properties of the wave field (wavelength, steepness, symmetry...) are strongly determined by the translation speed of the cyclone and the wind stress pattern (resonance between storm and wave motion). A phenomenon of trapped waves and reduced waves can emanate from the right front and left rear quadrant respectively, where the waves benefit from an extended fetch on one side and a reduced fetch on the other due to the synchronous and opposite movement of the wind field [Bowyer and MacAfee 2005; Young and Vиноth 2013; Kudryavtsev et al. 2015]. In addition, as presented in Figure, different sea state systems can be observed in the cyclonic field from the interaction between the previously generated swell and the wind sea [Holthuijsen et al. 2012; Kudryavtsev et al. 2015], for which microwave altimetry is particularly suitable despite its sampling and rain contamination shortcomings [Quilfen et al. 2018]. As such, several altimetry-based studies (on-board a satellite Young and Vиноth [2013] or an aircraft

Black et al. [2007]; Hwang [2016]) have highlighted the complex and highly asymmetric sea states deployed by a moving cyclone [Kudryavtsev et al. 2015; Yurovskaya et al. 2022]. These different sea states (cross-swells,...) are likely to have an impact on momentum fluxes and lead to asymmetry in the surface roughness of the wind sea [Black et al. 2007; Holthuijsen et al. 2012], as waves, and in particular breaking waves with the generation of foam and spray, transform and modify the interface properties [Andreas and Emanuel 2001; Kudryavtsev 2006]. They are also vectors of ocean mixing [Melville 1996; Wunsch and Ferrari 2004], although their effects seem to be limited and confined to the very upper part of the mixing layer during the passage of a cyclone [Stoney et al. 2017], and are therefore rarely modelled in the ocean response. On the other hand, the swell ahead the storm could disrupt the ocean structure earlier [Stoney et al. 2017], due to its longer wavelength and thus deeper penetration into the mixed layer [Huang and Qiao 2010], but to what extent remains an open question <sup>9</sup> (See the discussion on the Trami case).

Within the ocean structure, the tropical cyclone feeds other turbulent mechanisms, of which wind-driven mixing remains generally the dominant process in the oceanic response [Price 1981], with several persistent anomalies injected along the water column in the aftermath of the wind-field. This set of geophysical and biochemical signatures takes the form of a cold surface trough, elongated along the trajectory and similar in width to the size of the cyclone. This singular footprint, characteristic of a TC, remains intense for some inertial periods during which strong ML currents occur and propagate horizontally and downward, leaving a complex, oscillating hyperbolic-like wake around the surface trough. This cold wake stands out strongly from the different variabilities observed, as well as from the mesoscale activities present in the tropical and subtropical basins. It is characterised by a cooling of several degrees [Price 1981; Bender et al. 1993; Lloyd and Vecchi 2011; Vincent et al. 2012a], usually accompanied by a deepening of the mixed layer by several tens of metres [Pan and Sun 2013; Reul et al. 2014], an inertial currents of the order of  $1 \text{ m.s}^{-1}$  [Sanford et al. 1987; Firing 1997] and a surface trough of tens of centimetres [Ginis 2002], as well as an increase in salinity [Grotsky et al. 2012] and chlorophyll content [Lin et al. 2003; Babin et al. 2004]. These upheavals are associated with deeper subsurface anomalies in the seasonal thermocline [Price 1981; Price et al. 2008], sometimes at greater depth than the winter climatological mixing layer [Jansen et al. 2010; Vincent et al. 2013], which reflect the internal displacement of the isopycnets induced by the entrainment flux.

The surface cooling was the first response actually observed, with substantial evidence of local temperature anomalies provided by the hydrographic surveys of Fisher [1958] and Leipper [1967]. They were followed by numerous studies that corroborated their observations by combining the analyses provided by numerical models [Bender et al. 1993] with in situ measurements, such as buoys and moorings Price [1981], or more circumstantial observations like air-deployed instruments (AXCP/AXBT <sup>10</sup>) [Sanford et al. 1987; Price et al. 1994; Shay and Elsberry 1987]. In the meantime, satellite observations finally provided a comprehensive map of surface cooling with first infrared sensors (after storm passage) [Stramma et al. 1986; Cornillon et al. 1987] and then with microwave measurements (regardless of the cloud situation [Wentz et al. 2000]). Tropical cyclones thus generally impact the ocean surface temperature by creating cold wakes as fingerprints of the ocean interior responses. Note, the response of the ocean's surface salinity to the passage of a tropical cyclone, and the processes involved, are much less known. Along the

<sup>9</sup>the dissipation rate of wave-induced turbulence is function of wave steepness over its wavelength [Huang and Qiao 2010]

<sup>10</sup>air-expandable current profiler/bathythermographic

induced cooling, this surface salinity response can be a potential indicator of salinity's control. For some storms, tropical cyclones induce surface salinity increases associated with reduced mixing and cooling [Reul et al. 2021].

### 1.3.2 Restratification: long term effect

Following the disturbances induced by the passage of a TC on the oceanic vertical structure, several mechanisms subsequently occur to restratify the ocean by removing the thermal anomalies within the vertical column (see Fig 1.24) in order to return to climatological conditions [Dare and McBride 2011; Mrvaljevic et al. 2013]. It implies 1D, 2D and 3D processes that can be summarized as [Haney et al. 2012]:

- ① Surface Flux : heat radiative flux. (1D) [Price et al. 2008].
- ② Ekman Buoyancy flux. (2D) [Thomas and Lee 2005].
- ③ Vertical Mixing Layer Eddies (MLE) induced by geostrophic adjustment. (3D) [Fox-Kemper et al. 2008]

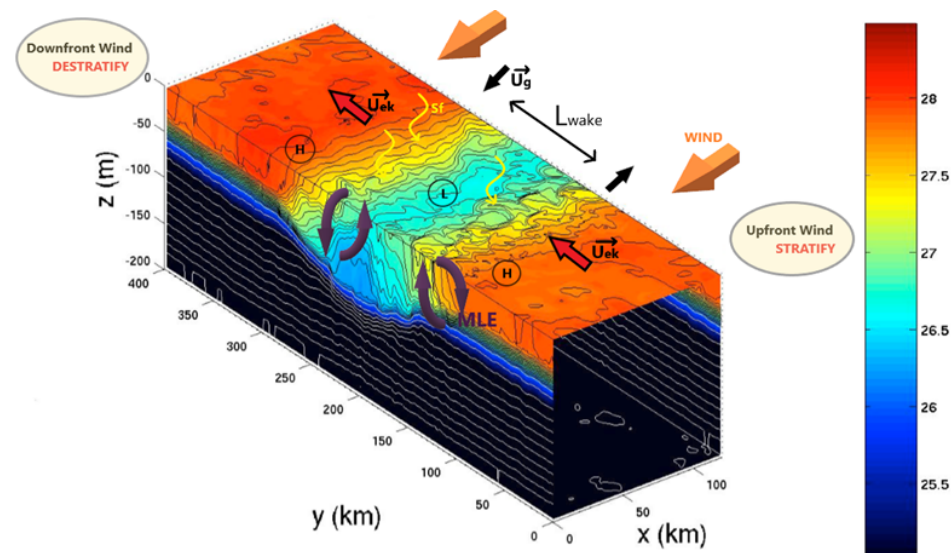


Figure 1.24: *Idealized Hurricane wake restratification, with one- (air-sea heat flux), two- (Ekman buoyancy fluxes) and three-dimensional (Mixed-layer eddies) processes represented. Extracted and modified from [Haney et al. 2012].*

These three mechanisms can set up a few days afterwards the cyclone, each acting in a different way on the wake with a specific depth of penetration [Haney et al. 2012]. The first to take action in the recovery is the surface flux, with the net input of solar heating, no longer counterbalanced by the others sources of exchange (latent, sensible, longwave radiation) due to the strong weakening of the air/sea flux induced by the cooling [Price et al. 2008]. The shortwave radiation operates directly after the departure of the clouds shield, depending on the size and the translation speed, this can generally take up to 2-3 days maximum (match with the beginning of the cooling damping) [Vincent et al. 2012a]. Once the cyclone winds regime has passed, the global winds take over [Price et al. 2008] and begin to act on either side of the wake through Ekman transport (see Fig 1.24) [Thomas and Lee 2005]. Last mechanisms to occur are the MLE ③, that stem from baroclinic instabilities generated during the geostrophic adjustment of the wake fronts

which take place one week after the TC disturbance [Fox-Kemper et al. 2008], leading to an overturning motion on either side of the wake.

While many studies have ventured into the processes at work in the cooling and wake generation (see 1.3.1), few have focused on the mechanisms involved in the restratification. Among the few studies that have addressed this issue, surface flux [Price et al. 2008; Vincent et al. 2012a; Jullien et al. 2012] have been emphasised in the recovery of surface anomalies up to 80% [Price et al. 2008; Vincent et al. 2012a] and seem to explain a large part of the evolution of the SSTA observed by satellites when a 1D model is used [Price et al. 2008]. In addition to this major contribution, the advective forces have been also noticed to act substantially in the warming of cold waters [Jullien et al. 2012], especially for TCs with strong forcing (high WPI: large and intense TCs) [Vincent et al. 2012a], for which the enhanced lateral advection displaced warm waters outward the wake and are potential source of heat afterward. On the other hand, the importance of the 3D processes have been poorly investigated. Haney et al. [2012] have emphasised the essential collaboration of these processes in the recovery of surface and subsurface anomalies, and the key role of MLE in the redistribution of water masses. Indeed, as shown by Vincent et al. [2013], subsurface warm anomalies are on average located in a layer at 30-100 m depth, for certain forcing conditions that also depend on the pre-storm MLD, cold anomalies can go up to an appreciable depth of more than 40 m (as it is the case for the José example). The photic (10-30m, [Mrvaljevic et al. 2013]) and Ekman layers are generally thinner, through their combined action they stratify the surface layer which cap the mixing above the subsurface bolus, so the deeper anomalies are out of reach for the 1D and 2D processes without an additional component. Although the MLE do not change the buoyancy of the water masses, they do redistribute them and make the deep anomalies accessible to the effect of surface heat and transport flux. In turn, the Ekman buoyancy flux also impacts the 3D process activity. When the wind blows uniformly along a rather symmetrical wake, the associated cross-transport inevitably creates an asymmetry between the two fronts: on the left side (upfront wind) the warm waters are conveyed on the cold waters and stratify the surface layer. It reduces the instability and thus slowdown the overturning motion. In contrast, on the right side the dense waters are carried on light waters increasing the instability and the MLE activity, resulting in faster removal of deep anomalies on the downfront wind side. The 2D processes modulate in some way the deep restratification activity, the only one able to act on the bolus, while ① and ② remain the dominant recovery processes within the first meters, ③ is the leading term for the subsurface. Their weights in the total restratification may also fluctuate according to the wake properties (width, cooling, deepening), a deeper wake will obviously rely on 3D processes, while a larger or narrower wake will respectively depend on 1D or 2D processes.

Despite the simultaneous action of these mechanisms, the signature of the forcing in the vertical column is long-lasting, the restratification is actually one of the TC long term effects, especially since the persistence of the aftermath is uneven along the upper ocean structure, between the SSTA, the wake cooling and the warm anomaly. Dare and McBride [2011] showed that the total recovery period could be split into two trends: a rapid decay of the cooling called e-folding ( $SSTA/e$ , Fig 1.25a), and a longer decrease to climatological conditions (see Fig 1.25b).



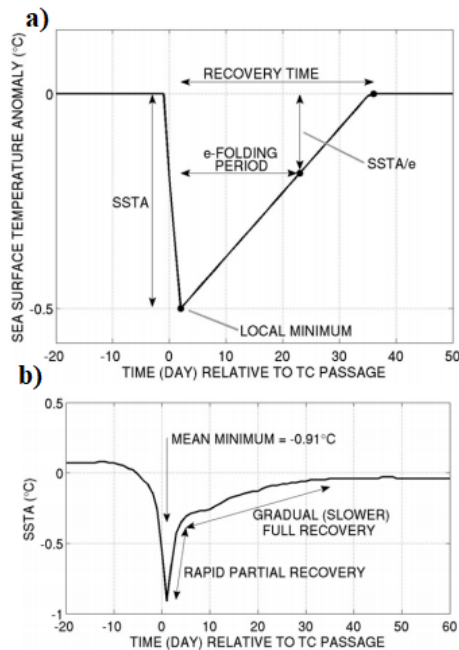


Figure 1.25: *Characteristic time period of the cold wake: a) Description of the e-folding and recovery time, b) Real mean-SSTA evolution left by TCs. Extracted from [Dare and McBride 2011].*

much more versatile among the cases [Mei and Pasquero 2013] and is generally contained in a large time window between 20 to 40 days.

If this duration can exceed the TC lifetime, it is even longer as one looks at the restoration of the upper ocean structure [Pasquero and Emanuel 2008]. For instance, Mrvaljevic et al. [2013] observed an e-folding time twice longer for the column cooling than for the surface anomalies. Traces of this cooling are suspected to remain for up to 2-3 months after the cyclone passage, and several months for warm anomalies ( $>6$ ) as indicated in Figure 1.26 based on the SSHA signature. As mentioned in 1.2.2, SSH is sensitive to heat content and is an excellent integral of the processes that took place in the upper ocean, so its signature is much more persistent. Evolution of OHC in the aftermath of TC and observation of the warm bolus is thus possible by long-term monitoring from altimetry.

Several studies have quantified these time scales for the SST. While the persistence of the signature could already be observed via in-situ observations, satellites have brought a direct impact in the monitoring of these anomalies. Most of the references on this subject date back to the beginning of the century with the first estimates of the temporal scale of the restoration [Hart et al. 2007]. Most of them agree on an e-folding period between 5 to 20 days, as demonstrated by the case study from Price et al. [2008] on two TCs (Fabian & Frances) and corroborated by quantitative studies released by Hart et al. [2007], Jansen et al. [2010], Dare and McBride [2011] and Mei and Pasquero [2013] with values ranging from 1 to 2 weeks, which are highly consistent with the model and AMSR-E estimates ( $\sim 15$  days) from Vincent et al. [2012a]. With the exception of the weakest cooling and TC events [Dare and McBride 2011], the e-folding time is found statistically invariant to the forcing parameters and remains relatively stable from one system to another [Dare and McBride 2011; Mei and Pasquero 2013]. In contrast, the period of full recovery is

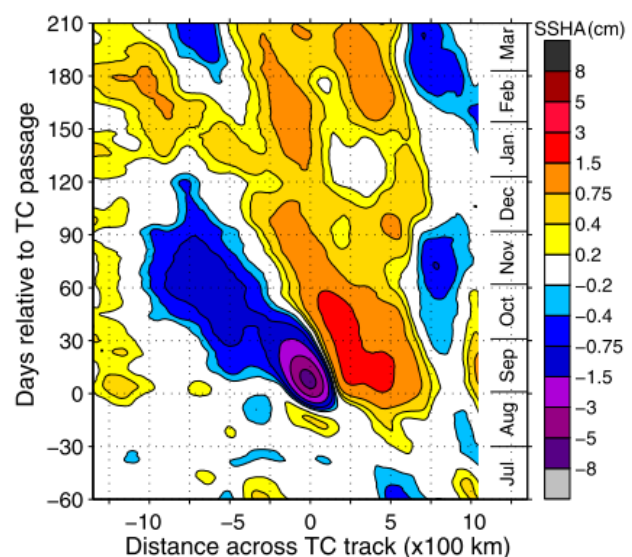


Figure 1.26: *Temporal evolution of along-track-averaged composite SSHA (centimeters) associated with the passage of major TCs in the Northern Hemisphere [Mei et al. 2013].*



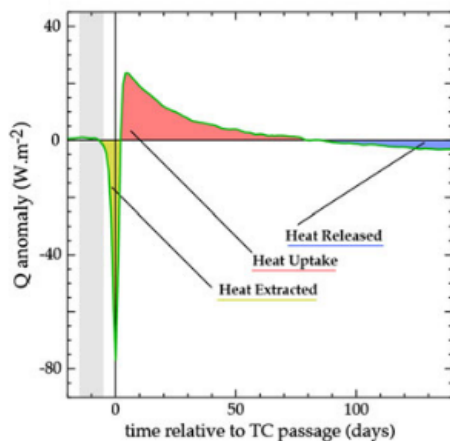


Figure 1.27: *Summarizing sketch of the ocean heat exchange in the TC wake for different time period. Extracted from [Vincent et al. 2013]*

Indeed, this evolution can be described in three phases, detailed in the Figure 1.27 with a first very brief phase, where intense mixing cools and deepens the mixing layer, during which the cyclone also extracts energy through intense air-sea fluxes including latent heat release. The change in heat content is associated with a negative peak in heat flux (OHE: ocean heat extracted) and is related to the time scale of the highest negative SSHA in Figure 1.26. This is followed by the long period of restoration, in which the ocean receives energy from solar heating and air-sea fluxes. During this stage, the energy received by the ocean to restore from the cooling is considered as similar as the energy injected in the subsurface bolus and is translated in terms of Ocean Heat Uptake (OHU), which is the quantity sought to estimate the warm anomalies induced by the TC forcing. To what extent these anomalies persist over time or dissipate is still a matter of debate. In their paper, Haney et al. [2012] tried to parameterize the respective recovery times for each process, and drew two conclusions: 1) the individual times are globally all greater than those observed for the SST, which is a further argument for the existing synergy between these processes, 2) even combined, the time scales associated with the removal of the warm anomalies are still multi-annual, implying that a significant fraction would survive beyond winter. Pasquero and Emanuel [2008] even suggested that at least 1/3 of the warm anomalies last one year, yet, the model they used did not include the seasonal cycle. Once the cold wake is dissipated, part of the warm anomalies are known to re-emerge and are associated to an ocean heat release to the atmosphere (OHR) due to the deepening of the ML in wintertime. The survival of the bolus strongly relies on its depth and on the climatological conditions (winter MLD) [Jansen et al. 2010], itself dependent on the latitude. To subsist, the warm anomalies must be located under the winter MLD (as illustrated by Figure 1.28a & b), preferentially in the permanent thermocline for subtropical latitudes, as the seasonal thermocline is gradually eroded during winter. As this bolus involves a density change of the vertical structure, the remaining fraction is suspected to participate to the thermohaline circulation via meridional ocean heat transport [Emanuel 2001].

Over the last 20 years, several studies have attempted to assess the OHU to investigate on the potential long-term effect of warm anomalies on the global circulation. As suggested earlier, there are essentially two major methods of estimating the OHU. The most common way is to estimate the cooling in the wake, which requires information on the SST before and after the TC, pre- & post-storm MLD, and the horizontal distribution of the cooling (width, and cyclone track length). Most of those information are derived from combination of climatology product with ocean model and (MW/IR or both) radiometric observation of SST. The second method uses altimeter observations, like in Mei et al. [2013], and bypasses the need of in-situ information, yet it requires a very long follow-up of the anomalies (8 months). Once the cooling is dissipated and the related trough is gone ( $\sim 4$  months Fig 1.27), the vertical structure is left with a net heat input and an associated positive SSHA that increases several months after the cyclone passage (5-7 months, Fig 1.27).

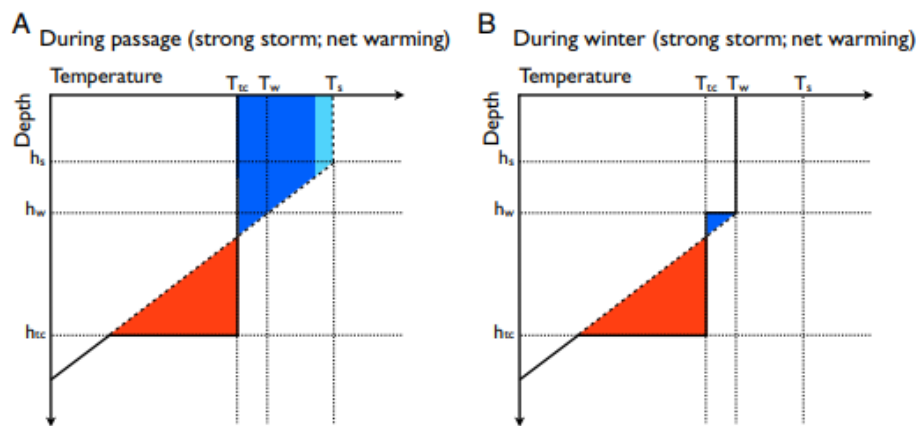


Figure 1.28: *Schematic diagram for the change in the vertical temperature profile for the case of strong storms with net warming: (A) immediately after the TC passage and (B) after the winter season. Dashed lines show the climatological condition, and solid lines show the situation with the effect of the TC.  $h_s$ , climatological mixed-layer depth during the summer season;  $h_{tc}$ , TC-induced mixing-layer depth;  $h_w$ , climatological mixed-layer depth during the winter season. Temperature follows the same convention. Extracted from [Mei et al. 2013].*

It is this positive signature, once averaged cross-track, that allows the estimation of the OHU by altimetry. Since the first analysis conducted by Emanuel [2001], the OHU estimates have considerably evolved, going from a maximum value of 1.4 PW to 0.117 PW. When compared to the total heat transport from tropics to polar regions (2.9 PW [Wunsch and Ferrari 2004]), the contribution of the cyclone activity changes from significant to negligible. This major discrepancy stems from the assumptions considered by the different approaches. For instance, Emanuel [2001] overestimated the wake width by fixing it to a constant value and considered, like Sriver et al. [2008], that the whole warm anomalies persist after winter. In contrast, Jansen et al. [2010] estimated that only a fraction of the total heat is permanently deposited below the ML ( $\sim 1/4$ ). At last, while most studies consider the cooling area equal to the warming one, Vincent et al. [2013] also take into account the energy extracted by the cyclone, whose loss is compensated by the OHU partition (Fig 1.29). These different considerations have led to a re-evaluation of the TC influence, and if there are still discrepancies between studies, they seem to point towards a low to moderate input from TCs to the total OHT and concentrate around a smaller range of values [0.1-0.5 PW] (see table 1.1).

These estimates should also be treated with caution because of their large errors, often of the same order of magnitude as the value itself (see table 1.1). Most of these studies rely heavily on models and climatological data rather than in-situ profiles. Besides, L4 gridded products used to provide daily field of SSH are suffering from large correlation window of similar time and space scale than the wake trench and its e-folding time period, which can smooth out the actual signature of the cyclone. As for the partition of the remaining energy in the meridional transport, Jansen et al. [2010]; Vincent et al. [2013] suggest that it can be split into two components: the equatorward and the poleward transport. A third part may also play a role, and even reduce the effective OHT to 73 TW [Vincent et al. 2013] due to latitudinal compensations. (Fig 1.29).

Table 1.1: OHU estimated from different analyses.

| Analysis<br>& Time window          | Forcing<br>& ocean field<br>& vertical profile                                                         | OHU (PW)             | % errors |
|------------------------------------|--------------------------------------------------------------------------------------------------------|----------------------|----------|
| [Emanuel 2001]<br>1996             | Coupled model( $\tau$ )<br>model(SSTA)/TOPEX-Poseidon(SSHA)<br>climatology(initialize model)           | $1.4 \pm 0.7$        | 50%      |
| [Srifer et al. 2008]<br>1998-2005  | Quikscat/TRMM-TMI (+3 day mean)<br>TRMM-TMI (SSTA) (+3 day mean)<br>Climatology (Pre & Post-storm MLD) | $0.480 \pm 0.1^{11}$ | 21%      |
| [Jansen et al. 2010]<br>1998-2005  | IBTrACS<br>TRMM-TMI/L4 SSH<br>climatology/model                                                        | $0.15 \pm 0.15$      | 100%     |
| [Vincent et al. 2013]<br>1983-2007 | IBTrACS/Willoughby<br>coupled model<br>model                                                           | 0.117                | X        |
| [Mei et al. 2013]<br>1993-2009     | IBTrACS<br>L4 SSH<br>$\emptyset$                                                                       | $0.32 \pm 0.15$      | 47%      |
| [Buetti et al. 2014]<br>2004-2005  | Rankine/IBTrACS<br>model<br>model                                                                      | $0.13 \pm 0.09$      | 70%      |

Thereby, if the disturbances induced by the cyclone forcing may not impact as significantly as expected the global circulation, TCs trigger a strong localized ocean response in the shape of a cold wake and disrupt the interface through intense mixing, enhanced air/sea flux and wave breaking with the production of sea sprays and foams that modify properties of the airflow. They can also perturb tropical activity at basin scale with far-reaching swells, and by impacting subsequent storm events or atmospheric cells [Buetti et al. 2014].

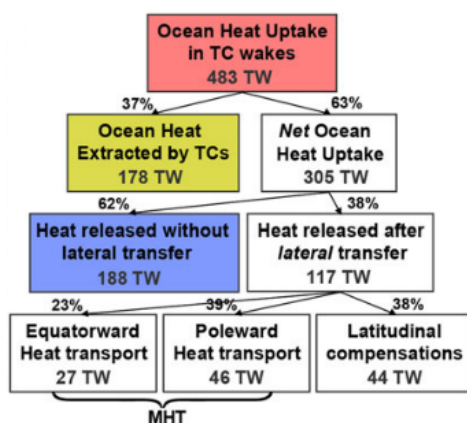


Figure 1.29: Partition of the total OHU, extracted from [Vincent et al. 2013]

Through the description of the cold wake and the intricacies of the anomalies lifetimes, it is fascinating to see how much the ocean remembers the passage of the cyclone, and fossils its imprint in its structure for several months, as a witness of the strength of these climatic systems. If remote sensing is suited to monitor such long-term response, the various studies we have discussed have not taken full advantage of the potential of the current constellation of MW and in-situ instruments to infer both TC vitals and cold wake anomalies. It is time to address the different caveats and shortcomings related to the portrayal of TC forcing and its ocean response, and see how the thesis can provide some answers.

## 1.4 State of Art and Issues.

During this journey into the impetuous realm of cyclones, we admired all the richness of oceanic processes and ventured into the tropical basins, where we got lost in the intricacies of the TC wind field before getting caught up in its cold wake and drifting for many months at the pace of re-stratification (like the writing of this thesis). Through these many descriptions, we have tried to establish a fairly detailed picture of the different actors in the oceanic response to TC forcing. A number of studies with accumulated observations sets the context of this thesis and the scientific heritage it benefits from. These passed efforts already highlighted the difficulty to fully understand all involved processes controlling the characteristics of TC wakes. Indeed, it requires efforts and often some luck to gather the necessary information to help perform thorough analysis.

In this regard, it is quite instructive to trace the various approaches of the most emblematic studies, and their assumptions and information methods used to interpret the oceanic footprint of TCs. From a closer examination, we can summarise our literature survey in two main trends that have shaped the TC wake analysis, both constrained by the advances and resources at their time. The first school benefited from the emergence of satellite observations. Starting in the 1970s until the 1990s, most efforts proposed studies, based on a few well-supplied examples of in-situ observations and satellite images [Price 1981; Shay et al. 1992; Ginis and Sutyrin 1995], to synthesise the dominant mechanisms governing the oceanic response with analytical and relatively simple numerical models. While these studies did provide first key insights, they were constrained by the limited number of MW observations capable of penetrating the cloud layer, and therefore the absence of satellite observations of the surface wind.

In the following decades, a second school emerges taking advantage of the progressive proliferation of observations and numerical developments, to more directly improve operational forecasts, analysis and re-analysis, using more involved 3D coupled models. The studies benefited from observational data, but also the development of climatologies and L4-product data to overcome the still existing lack of spatial and temporal sampling. These products allowed the global monitoring and mapping of the various TC wake markers. A number of quantitative studies were then carried out by combining several of these solutions, to investigate the wake properties (amplitude, depth, OHU ...) , the influence of the different processes and the coupling mechanisms [Bender et al. 1993; Vincent et al. 2012a; Jullien et al. 2012].

These two main approaches were of course complementary with their share of discoveries. More surprisingly, the recent emergence of new remote sensing technology (1.1.2 & 1.2.3) and the ever-increasing number of instruments did not necessarily trigger new studies to fully take advantage of combining the new capabilities of the MW sensors. Naturally, there is always a certain latency before a new instrument can be ingested into an operational product (ex: SMOS  $\sim$  10 years, [Bushnell et al. 2018]), but rare are dedicated efforts to possibly bridge the gap between the earlier parametric studies with the current satellite constellation. The actual virtual constellation of observing systems should indeed address some of the different shortcomings we mentioned along this introduction. We will list here the main issues related to the wake analysis, to identify the main questions addressed in the present thesis. Although ideally looking for exhaustivity, this is certainly a personal subjective interpretation of a young PhD student addressing a very large subject.

### 1.4.1 Observation of the oceanic response :

In section 1.1, we described the ocean as a complex, multi-layered medium on which processes at different time and space scales interact at the interface through dynamic markers. The vertical structure provides stability, modulates exchanges and influences the oceanic response. When a TC disrupts this structure, a range of perturbations acts to generate the resulting cold wake. This wake and associated mesoscale contributions can be sampled at different points of the cyclone track. Analysis of observations must thus take into account the associated spatial and temporal scales. For instance, once recovering a geostrophic balance, the ocean properties are advected at a rate of about 0.1 degree per day and must be considered in the spatial sampling. Although the oceanic signature of the TC remains singular, with its convergent/divergent anomaly patterns, and its succession of troughs along the track generated by near-inertial waves and surface Ekman transport/pumping, observations of these anomalies can be subject to several complications.

This first concern applies to the use of L4 multi-sensor products to track the expected anomalies, especially for SSHA. Used extensively in mixing or long-term analysis, these products ease the monitoring and provide global coverage by merging altimeter observations [Traon et al. 1995]. The constellation has well evolved from 3 to 6 instruments since 2011, but the nature of the measurements (nadir-looking) impose quite long ( $\sim 2$ -4 weeks) correlation windows to perform optimal interpolation, especially at low latitudes (see fig 1.30a). These scales are similar to those of the cooling and e-folding period and can therefore lead to significant smoothing of the signature or the creation of artefacts. A large number of cases are often used to offset for these effects [Jansen et al. 2010] (see table 1.1). With regard to the oceanic structure, pre-storm conditions are generally inferred from climatology of the global field (see table 1.1), which is convenient to study long-term trends based on a reference mean state averaged over several years [Gaillard et al. 2016]. For our purpose, section 1.1.3 highlighted the different variabilities that affect the seasonal activity of TCs. Overall, the last decade is under the same phase of decadal/multidecadal regime (see Fig 1.13), although two-three sharp episodes of ENSO occurred and may have influenced the climatological conditions [Gaillard et al. 2016].

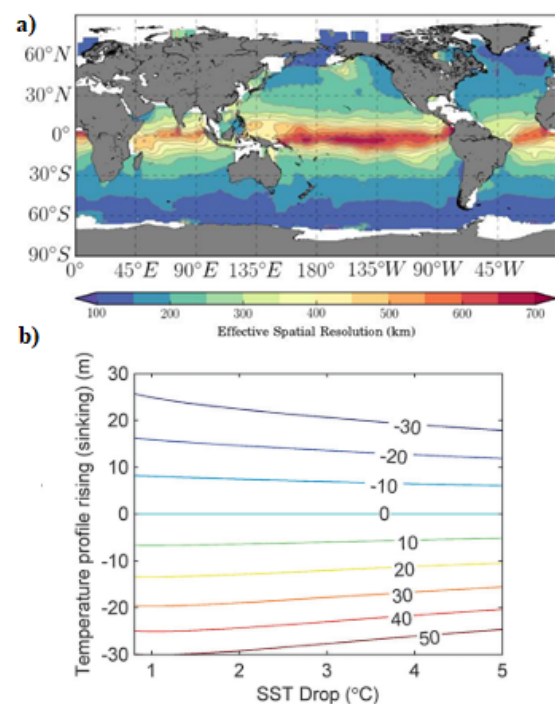


Figure 1.30: *Shortcomings of global ocean product. a) The zonal effective spatial resolution of L4 altimetry product (Tropical basin  $\sim 300$ km wavelength). b) Effect of the temperature profile rising (positive) and sinking (negative) by a pre-existing eddy on mixed layer deepening estimate (contour) for different typhoon induced SST drop.*

In addition, mesoscale eddies are also known to act on the pre-storm conditions by uplifting or pushing isotherm downward, to change the interaction between TC and the ML and to prescribe the ocean answer [Shay et al. 2000; Shay 2009; Chiang et al. 2011]. Several documented cases have shown the enhancement of TC intensity from warm core



eddies and the reduced associated cooling [Shay et al. 2000]. Such events are not unusual and even seem to be frequent in the path of intense TCs [Lin et al. 2008] and can not be informed by a climatology. Pan and Sun [2013] have shown the bias that occur in the estimation of the deepening as a result of the rise or sinking of the temperature profile by a pre-existing eddy, when using a pre-storm condition defined by climatology (see Fig 1.30b). These two caveats raise several questions:

- Can the current collection of altimeters and Argo floats provide both good sampling along the cyclone track and fine portrayal of the pre-cyclonic conditions?
- How much does the use of these data improve wake analysis and the scaling of TC induced anomalies ?

### 1.4.2 Parametrisation of the oceanic response :

The combination of variables describing the wake of the cyclone, its forcing and the oceanic conditions preceding its passage, allows to analyze the variations of its response for different cases and situations. The amplitude of the oceanic answer is already known to exhibit no dependence on a single parameter, but on a set of factors [D’Asaro et al. 2014; Vincent et al. 2012a] including: Vmax, Size, translation speed, stratification, and the presence of a BL.

Quantitative studies have demonstrated this non-linear behavior of temperature anomalies for any individual TC characteristic or environmental parameter, as shown in the Figure 1.31a with intensity, or with stratification and size in the Figure 1.31b. This last panel taken from Mei et al. [2015], describes the magnitude and temporal evolution of the SSTA for two distinct regions of the western Pacific basin, the China Sea and the open ocean. The former area is characterized by a much steeper seasonal thermocline, with TCs producing more intense cooling than in the open ocean for the same size distribution. Apart from Mei et al. [2013], these aspects have often been overlooked for the SSHA in favour of the assessment of the heat uptake (section 1.3.2), though Geisler [1970]; Greatbatch [1984]; Shay and Chang [1997]; Ginis and Sutyrin [1995] have demonstrated, through analytical and numerical models, the influence of the forcing parameters on the surface depression associated to the TC passage. Yet, so far, no systematic study has thoroughly investigated the fluctuations of the resulting wake sea surface height anomalies. In general, along-track altimetry data (L2/L3 product) are indeed seldom used in the analysis of the TC-induced trenches.

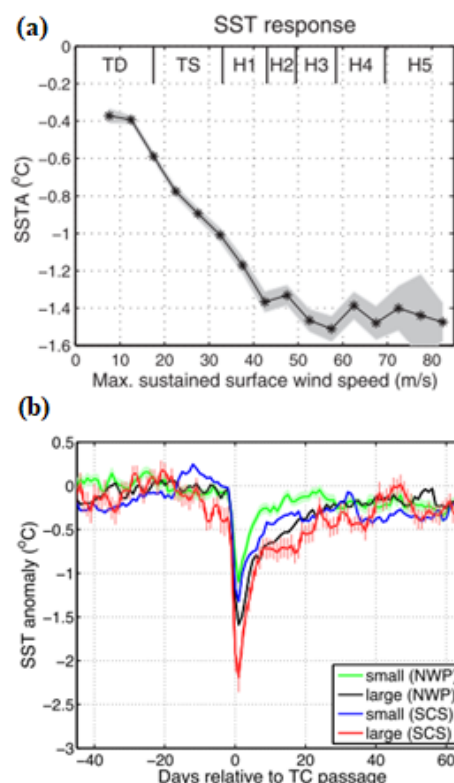


Figure 1.31: *Dependence of the SSTA: a) according to maximum wind speed, b) according to size, ocean background and time.*

Most often, the use of altimetry focuses on the heat balance within the wake, e.g. [Emanuel \[2001\]](#). Consequently, the studies of that time have never exploited along track altimeter data to probe NIW, probably due to a too limited number of instruments before 1992 and because of the large errors associated to the sea level measurements, making difficult the remote sensing of SSHA [[Bosch et al. 2014](#)]. This thesis therefore aims to investigate this overlooked approach in the field of near inertial wake([[Ginis 2002](#)]). Ideally, the sea level anomalies can directly be derived using the very short repeat cycle of the Jason altimeter series ( $\sim 10$  days, like in [Emanuel \[2001\]](#)), with one pre-storm track subtracted from the successive post-storm data. However, this brute-force method is not feasible for all altimeters, possibly having much longer repeat periods (i.e. S3A  $\sim 28$  days). A more specific methodology must thus be considered to take maximum advantage of the use of the direct information from along-track altimeter data.

While a scaling of the induced SSHA has never been achieved, some simple diagnosis of the SSTA were already proposed by [Greatbatch \[1984\]](#) and more recently with [Mei and Pasquero \[2013\]](#), based on some analytical solutions of the ML deepening caused by the entrainment from the shear stress only. [Vincent et al. \[2012b\]](#) proposed for the first time a scaling of the thermal response. They found a regression law that relates both the atmospheric forcing with the ocean background parameters, through the computation of the wind power (WPI) and the cooling inhibition (CI) index, two numbers that describe respectively the strength of the TC forcing integrated over its lifetime (combination of  $V_{max}/V_{fm}$ ), and the potential energy input required to disrupt and cool the upper ocean. If this bivariate fit accounts for the main trend, some large errors remain for cooling larger than  $3^{\circ}\text{O}(1^{\circ}\text{C})$ , [[Vincent et al. 2012b](#)]). Besides, the calculation of these two metrics is not so straightforward and was largely derived from numerical model and climatology information.

Here, we want to advocate a more direct approach, fully based on the remote sensing and in-situ capabilities to scale both anomalies. In addition to the issues related to the establishment of a multi-platform observation, the simultaneous description of these two responses also raises some questions:

- **How to link the description of the near-inertial wake with the mixing dynamic into an analytical framework that depicts both answers ?**
- **What are the main assumptions to take into account for our analysis ?**
- **From this analytical solution, how to derive the scaling of the anomalies?**

Most of the analyses performed on the near-inertial wake have focused on the baroclinic response of the TC signature, considered to be predominant [[Geisler 1970](#); [Ginis 2002](#)].

- **Can these observational data along with the analytical framework highlight the dominance of this response?**
- **To what extent does this contribution explain the variability of the oceanic answer?**

At last, this framework requires to retrieve fine sea level anomalies from direct measurements:

- **What methodology should we use to extract SSHA from along-track altimeter ?**

### 1.4.3 Tropical Cyclone Forcing issues:

Another key constraint to fully interpret the wake properties and scale its amplitude, is to rely on accurate description of the atmosphere forcing. In section 1.2, we described the structure of tropical cyclones, the surface wind field and the remote sensing measurements, where we discussed about issues to estimate cyclonic winds ( $>33 \text{ m.s}^{-1}$ ).

Importantly, information on the inner core is most often based on indirect methods, using Vis/IR sensors or sounders [Knaff et al. 2016], due to the lack of resolution or to the saturation of MW instruments. While  $V_{\max}$  can now enjoy a long line of well-established procedures with low uncertainties ( $\sim 5 \text{ m.s}^{-1}$ , [Torn and Snyder 2012; Landsea and Franklin 2013]), except for situations of fast evolution [Leroux et al. 2018] and eye replacement cycle,  $R_{\max}$  remains an elusive parameter for many observations (section 1.2.3).  $R_{\max}$  estimation is therefore very challenging to accurately and routinely assess [Knaff et al. 2021]. The few techniques used, suffer from significant errors, or are strongly constrained either by the geometry of the cyclone, or by the sensor resolution. No re-analysis is therefore conducted in the BTKs for this parameter [Knaff et al. 2021], and the uncertainties related to its estimation have never been really quantified by any studies that have assessed BTK errors ([Torn and Snyder 2012; Landsea and Franklin 2013]).

There are also strong uncertainties surrounding the precise determination of wind radii like  $R_{25}$  and  $R_{33}$ , especially when they rely solely on satellite observation. In their study Landsea and Franklin [2013] mentions how this accuracy is a function of the level of information ingested in the BTK procedure. Subsequently, largest wind radii uncertainties strongly rise when neither coastal measurements nor aircraft reconnaissance are available. For these situations, a slight decrease of the intensity accuracy is also noted ( $7\text{-}8 \text{ m.s}^{-1}$ ), especially for the most intense cyclones. For TC-wake analysis, mainly carried out in the deep ocean, far from other measurement systems, the largest uncertainties are thus expected. Fortunately, referring to Figure 1.20, we discussed in the section 1.2.2 that the current collection of MW radiometers (1-40 GHz) can be relevant means to document the outer core region of a TC and by extension to characterise the size of the phenomenon (useful to give a spatial framework to the wake). For the inner core, the Figure 1.32 shows a transect through the region of maximum winds, estimated using sentinel-1A for the Hector case. With a peak wind estimated at  $64 \text{ m.s}^{-1}$ , the wind retrieval algorithm benefits from both the

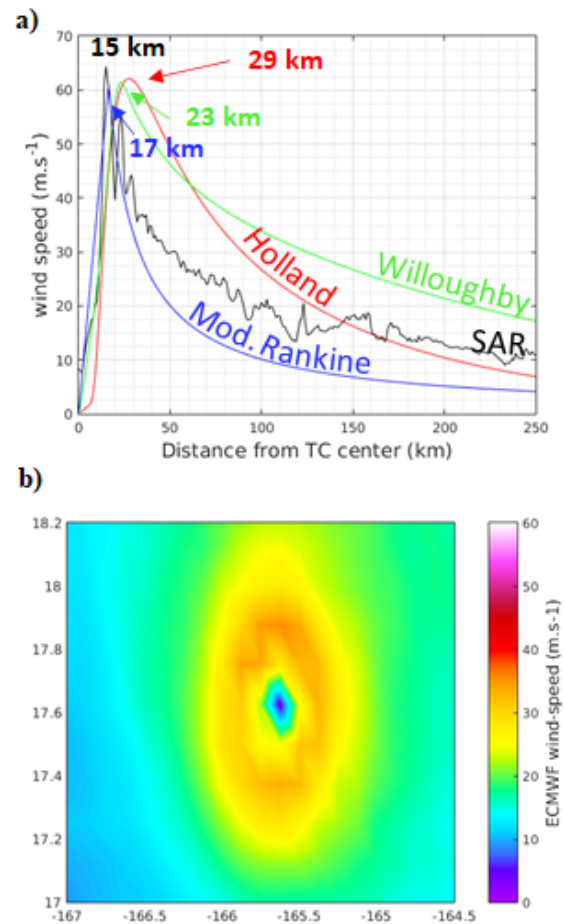


Figure 1.32: *Hector wind field: a) comparison between SAR and parametric models of the maximum wind transect, b) concomitant 2D wind field from ECMWF forecast.*

very high-resolution capability of the SAR instrument, as well as from the sensitivity of the combined co- and cross-polarized back-scatter signals. Today, SAR instruments are thus unique to map and probe the inner core of a TC. Still, such unique remote-sensed capabilities have still never been extensively exploited for a wide range of TC wind categories. For instance, [D'Asaro et al. \[2014\]](#) reported a SAR-derived surface wind-field for a single case study. Indeed, the implementation of a dedicated methodology to fully exploit SAR measurements is still quite challenging. Accordingly, the contribution of SAR measurements to help the analysis of the oceanic response to TC passages, has never been fully demonstrated or even evaluated.

Obviously, it must be recalled the relatively limited temporal sampling associated with SAR observing systems. Efforts shall thus be directed to advance the combined use of observations. Ideally, using SAR observations, strategies are to be derived to, statistically and/or analytically, provide robust estimates of the inner-core  $R_{max}$ , from  $V_{max}$  estimates and the outer-core information provided by low- and medium-resolution satellites.

In that prospect, two approaches are generally used by the different studies, the parametric wind models and/or operational products such as forecasts or analyses. As we saw in [1.2.2](#), parametric models are an idealized solution of the wind profile based on simplified assumptions, easy to set up and depending on a few parameters derivable from observations or from another data source such as best-tracks. They fall within a very specific context and may be representative of a short range of cases (like Holland model with some intense cases [[Willoughby and Rahn 2004](#)]). Operational products offer a more complete, coupled solution (forecast), and ingest available observations at regular times, but they may be limited by a rather coarse resolution that prevents any examination of the inner core (analysis and reanalysis), by the variability of wind intensity which can lead to important discrepancies (forecast), or by an underestimation of strong winds [[Magnusson et al. 2019](#)] (forecast, analysis, reanalysis). The parameterisation of momentum exchange at ocean surface is paramount in the determination of maximum winds, which is generally expressed through the drag coefficient  $C_d$ . There are many doubts about its behaviour in cyclonic winds (see next section [1.4.4](#)), notably on the rate of the trend, which undermine its parametrisation and would be a major cause of the underestimation of strong winds.

The inner core remains difficult to handle even by these methods. Besides, if some ambiguous parameters such as  $R_{max}$  guide the model, their large uncertainties can have negative feedback on its performance. This is for example the case with Hector, where the Holland and Willoughby models overestimate  $R_{max}$  by more than 50% compared to its observed value (Fig [1.32a](#)), which results in a significant shift in the wind profile. This is one of the main limitations of these seminal models, as they are  $R_{max}$ -based parametric solutions and are therefore dependent on its accuracy. While they have long dominated, a new generation of parametric models has emerged in recent years ( $R_{fit}$ -based model), going in line with the growing interest in size parameters, that incorporate information from an outer core radius to guide the description of the wind gradient mainly [[Rappin et al. 2013](#)]. This is particularly the case for the modified Rankine model, which, despite a slightly steep gradient, is closer to the observed wind profile for this example. This new kind of process can be a complementary or alternative solution in the absence of high resolution observation.

The representation of the cyclonic forcing is not straightforward, and the extraction of the TC vitals demand to address specific issues:

- **Can the new capabilities of MW measurements document the inner core and provide reliable  $V_{max}/R_{max}$  estimates for any TC category ?**

- Does the current collection of MW instruments allow a sufficient sampling of the surface wind field to bypass Best-track wind radii information ?
- Can the use of these wind radii information combined with Rfit-based model allows a better description of the complete wind profile? Does it take into account inter- and intra-storm variability of inner and outer core parameters? And can this approach be generalised to a large sample of cases and what is its range of validity?

#### 1.4.4 Drag coefficient issues:

The interface is the place where transfers of momentum, enthalpy exchange via latent and sensible heat fluxes occur, where the turbulent atmospheric and marine boundary layers couple and shape a wavy turbulent layer. These interactions can be expressed in terms of the drag coefficient, that translates the rates of vertical exchange of horizontal momentum between the two media, and links the wind speed to the wind stress through the bulk formula:  $\tau = \rho_a C_D U_{10}$ . As a result, the dependence of this synthetic variable to wind intensity has been the subject of many studies, as shown in the Figure 1.33. It is indeed a paramount parameter to evaluate for cyclone forecasting, since it reflects the momentum exchange at the ocean surface and impacts the TC intensification [Donelan 2018]. Important changes in the marine environment occur at hurricane force wind, all suggesting a turning point in the  $C_d$  behavior. For instance air/sea exchanges, largely mediated by surface ocean waves in normal circumstances, are mainly conveyed by sea bubbles, spray and spume generated from breaking waves under extreme wind regime [Andreas and Emanuel 2001; Holthuijsen et al. 2012], the resulting foam coverage creating a thin layer that caps and suppresses the short capillary waves [Kudryavtsev 2006]. These mechanisms transform the reflective and emissive properties of the ocean surface, and may also be responsible for the saturation of the bragg backscattered signal of copolarized scatterometer measurements (section 1.2.3).

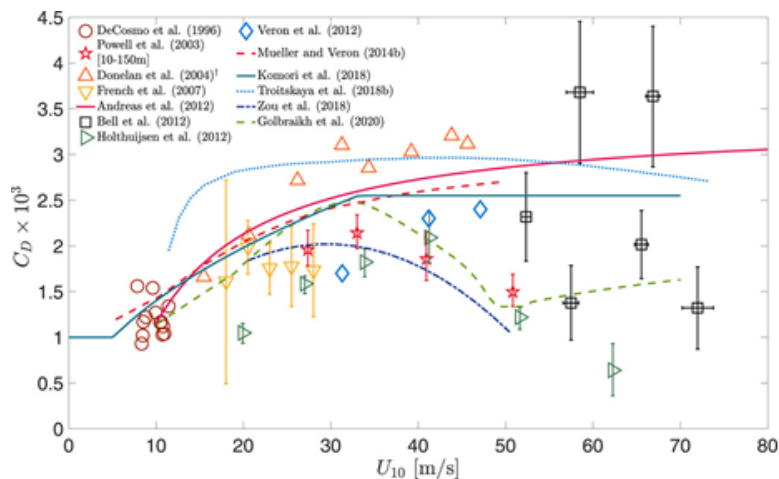


Figure 1.33: The sea surface drag coefficient  $C_d$  as a function of  $U_{10}$  from many experiments carried out over the last decades. Most studies find that  $C_d$  increases with  $U_{10}$  until about  $U_{10} = 30 \text{ m.s}^{-1}$ . Extracted from [Sroka and Emanuel 2021].

Several theories have been put forward on the shape of the  $C_d$  profile, but these seem to be gradually converging towards two preferred scenarios passing through a maximum value: an asymptotic or a parabolic profile. Although first studies initially suggested an increase of  $C_d$  for stronger winds, this assumption was made on the basis of in-situ



observations limited in intensity to 25 m.s<sup>-1</sup> [Curcic and Haus 2020]; the curve was then projected for higher values following this monotonic behaviour. Powell et al. [2003] showed that the linear increase in  $C_d$  from low to high winds can not be extrapolated for values above the hurricane force winds (33 m.s<sup>-1</sup>). Indeed, subsequent studies have agreed on  $C_d$  to level off around 30-40 m.s<sup>-1</sup> at values between  $2 - 3 \times 10^{-3}$  as displayed in Figure 1.33. If the vast majority have conceded the existence of an asymptotic value in the vicinity of the cyclonic intensity threshold, some doubts remain on the upper fraction of the  $C_d$  profile, for which two trends are essentially observed. On one hand, in-tank experiences predict a saturation of  $C_d$  for all winds above 33 m.s<sup>-1</sup> [Donelan et al. 2004]. On the other hand, indirect measurements methods based on GPS dropsondes [Powell et al. 2003; Holthuijsen et al. 2012], angular momentum budget [Bell et al. 2013], and upper-ocean momentum budget [Jarosz et al. 2007; Zou et al. 2018] advocate a decrease of the drag coefficient for cyclonic intensities. Theoretical analyses of the air/sea interface are generally in line with this last behaviour [Kudryavtsev 2006; Soloviev et al. 2014; Donelan 2018]. Even before the first studies on  $C_d$  decay, Andreas and Emanuel [2001] drew attention to the major role of sea spray in the enhancement of air/sea flux and the resulting intensification of TCs, for which wave drag alone could not explained the major intensity cases. Kudryavtsev [2006]; Soloviev et al. [2014] went further and postulated that sea spray and spume torn from breaking waves, generate a thin foam layer that damps the turbulence and suppresses short capillary waves. This slippery effect along with the separation of the air-flow from breaking wave crest [Donelan 2018] induce reduction of the drag. Soloviev et al. [2014] indicated that this decay implies a local maximum in the ratio of the exchange coefficients of enthalpy and momentum  $C_k/C_d$  (vital parameter for intensification [Emanuel 1995]), and would even be necessary to explain rapid intensification situations [Donelan 2018]. Those different mechanisms underline the limiting aerodynamic roughness of the surface waves [Donelan et al. 2004], and possibly explain the saturation of the backscattered signal in high-wind conditions, due to the short waves sheltering [Curcic and Haus 2020] and suppression. The analysis of the dependence of this parameter is thus not only interesting for the parametrization of the momentum fluxes at the interface, it is also revealing of the surface roughness properties, essential to MW measurements.

Finally, different slope gradients are also observed among the parabolic  $C_d$  profiles (Fig 1.33), with smooth decline for Powell et al. [2003] and steep gradient for Holthuijsen et al. [2012]. While these differences are partly due to the different methods used, the steepness of the profile would also depend on the properties of the wave field [Holthuijsen et al. 2012; Magnusson et al. 2019], which fluctuates according to the quadrants (section 1.3). Usually, determination of the air-sea momentum transfer are made from atmospheric measurements of the wind profiles, few studies have passed by a bottom-up approach looking at the ocean response like Jarosz et al. [2007]; Zou et al. [2018] with the induced ocean currents. The advantage of this process is to avoid the quite precarious conditions of in-situ surface measurements with sea droplets and breaking wave contamination [Jarosz et al. 2007], but it relies on sparse data (like mooring) that restrict the sampling of cases. Thereby, the observation of TC-induced anomalies with our multiplatform satellite database and the use of an analytical framework based on remote-sensed forcing and environmental parameters can help to infer the drag coefficient, for a wide range of intense cyclones. Once again, this innovative solution will highlight capability of MW sensors, but before these results can be interpreted, some clarification is required:

- Which analytical scheme to link the  $C_d$  to the wake anomalies ? Does our bottom-up approach coincide with previous surveys ?

As we will have different level of information about forcing, it is relevant to explore the robustness of our methodology and look at its impact on the  $C_d$  dependence.

- **Does the degradation of the information maintain the trend achieved with the full observation capability?**

#### 1.4.5 Synthesis:

Through these various problematics that border the wake of cyclones, and the shortcomings that have limited certain aspects of its analysis so far, our thesis aims at filling some gaps left by previous surveys on the scaling of anomalies (notably SSHA), in a context now favourable to a better exploitation of EO data thanks to a novel combination of instruments. To synthesize the stakes of our study, we can formulate the main questioning that led our entire approach as follows:

**How to integrate the different capabilities of the current constellation of satellites and in-situ systems into a coherent semi-empirical framework, to describe the complete TC wake anomalies and investigate, with a bottom-up approach, the air/sea interactions under extreme wind conditions ?**

## 1.5 Strengths and Outline of the Thesis.

To address our main motive and the different points raised in previous section, the thesis will build on five major assets, five essential pieces to complete our puzzle, and whose strengths have been demonstrated or illustrated individually by pioneering works. The bulk of these seminal surveys was carried out within the LOPS laboratory, under the direction or with the participation of my three thesis supervisors (Bertrand Chapron, Yves Quilfen, Alexis Mouche), and it underlines the expertise of the SIAM team in the use of satellite data. It is this corpus that has gradually shaped the thesis and led to the different results that we will present after this last section of Chapter 1.

### 1.5.1 MW radiometers and in-situ

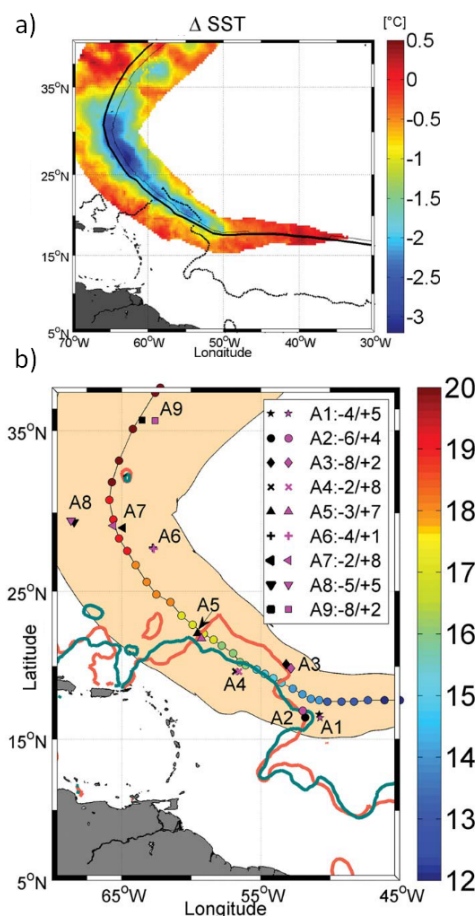


Figure 1.34: *Example of co-analysis of Igor wake in 2010, a) using a merged remote-sensed SST product and b) argo floats. The location of the available float pairs are indicated with the corresponding numbers of day between the profile measurement and the passage of Igor. The orange and gray-blue contours are indicating the pre-hurricane and post-hurricane horizontal extent of the Amazon-Orinoco plume, respectively.* (extracted from Reul et al. [2014]).

As a partial answer of 1.4.1 first question, Reul et al. [2014] illustrated the tremendous collusion between MW instrument (SMOS) and Argo floats to investigate the effects of ocean stratification on the ocean response, first for the Igor case (2010 Hurricane, Figure 1.34), and then at the basin scale. Some previous studies have also benefited from the massive use of Argo data in the vicinity of a cyclone, mainly for the northwest Pacific basin [Liu et al. 2007; Lin et al. 2009; Park et al. 2011; Wu and Chen 2012], but very few have used them to look after salinity features [Balaguru et al. 2012; Reul et al. 2014] and their effects on the oceanic answer. In that respect, these articles are pioneering, illustrating for the first time the remote-sensed salty wake of a storm. It highlights the ability of this multi-looking approach, to take advantage of both the spatial and temporal coverage of surface cooling by satellite MW instruments (Fig 1.34a), and of the fine description and short cycle of Argo floats (10 days), to document the pre-storm oceanic structure and analyse the post-storm deep anomalies (Fig 1.34b). In the scope of our multi-sensor analysis, we want to elaborate upon this combined spaceborne and in-situ measurements approach to go further in the analysis of the wake anomalies. The measurement of high winds by the new generation of L-band radiometers was also discussed in section 1.2.2, several studies have proven their skills, notably Reul et al. [2016] made a comprehensive review of TC structure performed by the two L-band instruments SMOS and SMAP and the multi-frequency sensor AMSR-2, for the three brothers with a high acquisition rate. Thanks to these two seminal works [Reul et al. 2014, 2016],

we will take full advantage of the radiometric data through two separate chapters, which will be dedicated to the exploitation methods of these measurements: one in order to document wind radii and define TC size (Chapter 2), and the other to examine both the ocean pre-storm conditions and response in combination with in-situ data (Chapter 5).

## 1.5.2 Altimetry

Altimetry observation can be seen as a paradox. Despite the sea level status as a primary geophysical variable and the strong interest carried by the climate science community for many applications, the use of along-track measurements to determine TC trenches is really overlooked (see 1.4.1 & 1.4.2), L4 product being preferred. Apart from an illustrative Figure in Ginis review [Ginis 2002], the study carried out by Y.Quilfen in the scope of OSTST 2012 (Figure 1.35) offers a unique view of the sea surface height anomalies left in the aftermath of Igor. The measurements are selected cross-track to the heading of the storm, along the characteristic dimension of the wake (Fig 1.35a), a privileged altimeter passage crossing Igor was then chosen to depict the temporal evolution of the anomaly (Fig 1.35b).

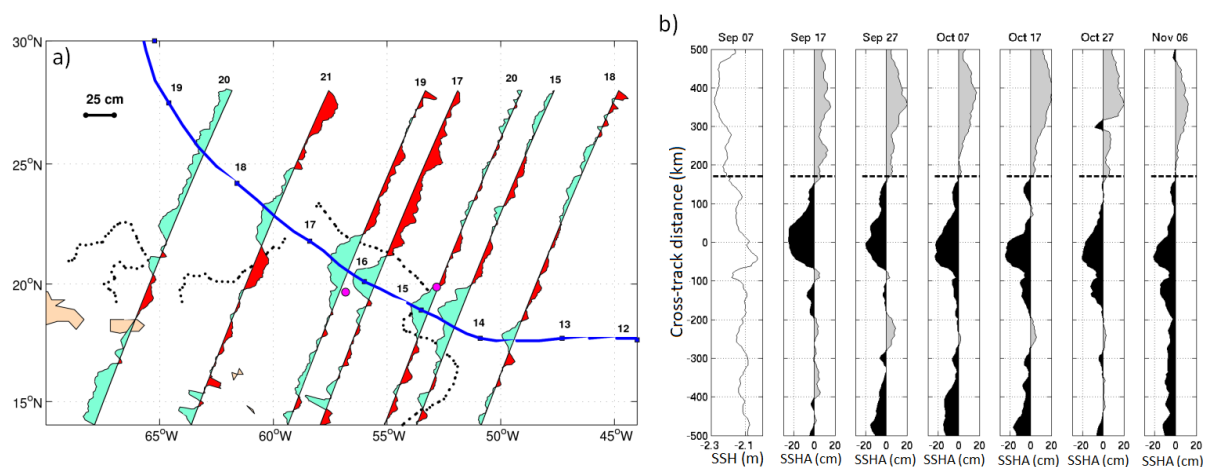


Figure 1.35: a) Spatial and b) temporal monitoring of Igor sea level anomalies from Jason series. In a), the differences in sea level (elevation in red, depression in cyan) are derived from successive Jason-1/2 altimeter tracks before and after Igor passage. Numbers above the tracks give the calendar day for one altimeter track after Igor passage. The track is pictured with the blue line for each location of Igor at noon. b) represents the evolution of the sea surface height anomalies for the Jason-2 orbit before (left, Sep 07) the Igor passage (Sep 16) and differences, for the same orbit number, with values for the subsequent cycles. (extracted from Y.Quilfen in OSTST 2012 meeting).

Through these two panels, several points stand out:

- The use of only two instruments seems to provide already a good sampling of Igor track. The current large number of instruments should therefore allow the sampling assumption made in section 1.4.1 to be met. (Panel a)
- Along track measurements display the classical salient features of the TC wake without smoothing, with near the TC center, a negative signature from the diverging flow, and around the trench, positive anomalies conveyed by the converging flow from the wind stress curl. (Panel a)

- The persistence of the SSH signature will allow to increase the temporal window for collecting altimeter tracks (7-10 days), therefore sampling will also be strengthened. (Panel b)

The results performed through this analysis justify our desire to bypass the L4 products and try an alternative approach. However, not all altimeters have a short repetition cycle, so we cannot use, as shown here, the subsequent cycles of an altimeter orbit to subtract the pre-storm measurement from the post-cyclonic tracks to retrieve SSHA. A compromise can still be considered, using the L4 products only for the initial conditions. Chapter 5, in addition to handling Argo and SST data, will explore different methodologies to extract the information from post-storm along-track measurements.

### 1.5.3 SAR

To complete our collection of MW instruments, we have to tackle our latest observation asset, the Synthetic Aperture Radar. This instrument was not foreseen in the initial scheme of the thesis, but it has become an essential part of the reading of wake anomalies in view of the precarious remote-sensed surface information of the inner core (mostly Rmax as evidenced by the section 1.2.2). This instrument has therefore been gradually incorporated in our story through personal initiative to explore its ability to retrieve extreme winds, even for major hurricanes (Figure 1.36).

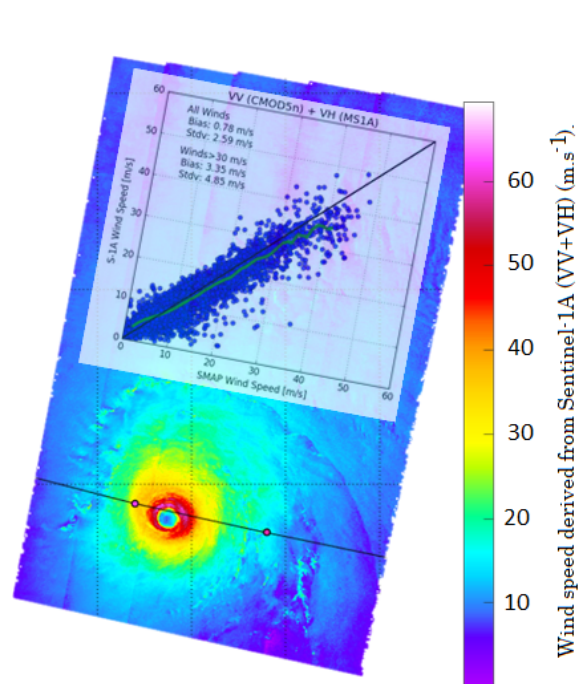


Figure 1.36: *Sentinel-1A derived-wind speed of Hector from combined co- & cross-polarized  $\sigma_0$ . The comparison between SMAP winds and those from the new SAR algorithm for 20 cases is superimposed on the SAR scene (extracted from Mouche et al. [2017]).*

Although this type of instrument is not new (first was on board Seasat), it has been enriched with new technologies, such as cross-polarized signal measurements used in dual- or quad-polarization mode on the satellites of the sentinel-1 constellation (S1A/B: VV+VH or HH+HV) or on Radarsat-2 (RS2: HH+VV+HV+HV). SAR instruments compensate for the classical issues encountered by other MW sensors, like the lack of resolution of radiometers [Reul et al. 2017] and the saturation of scatterometers [Brennan et al. 2009], thanks to the better wind dependence of its cross-polarized signal combined with the high resolution [Dagestad et al. 2013]. While the Bragg backscattering mechanisms, responsible from the signatures measured by co-polarised active sensors, tend to saturate due to change in sea surface roughness properties at hurricane-wind force (section 1.4.3), measurements dependent to breaking waves mechanisms (e.g. volume scattering) remain sensitive to extreme winds sea surface imprints [Zhang and Perrie 2012], like radiometers with foam emissivity [Reul et al. 2012].

It is the case for present SAR instruments; indeed breaking waves events induce volume scattering that depolarizes the radar signal [Phillips 1988], VH/HV signal is a measure of this contribution, which becomes dominant at high winds [Powell et al. 2003; Holthuijsen et al. 2012; Reul et al. 2012].



Hwang et al. [2010] and Vachon and Wolfe [2011] have shown the cross-polarized signal is strongly related to wind intensity, particularly for high winds, observing a nearly cubic relation with no apparent sign of saturation, nor dependence on wind direction or angle of incidence (unlike HH or VV).

Some studies have thus been undertaken to develop a crosspolarized based wind retrieval [Zhang and Perrie 2012; Hwang et al. 2015]. [Mouche et al. 2017] went further and chose to combine both polarizations to retrieve TC wind speed, thus taking advantage of the strong signal-to-noise ratio of co-polarized signal for low wind speed estimates and the strong sensitivity of cross-polarized signal for high winds (as shown in Figure 1.36). Mouche et al. [2017, 2019] have demonstrated this capability by comparing respectively this new wind retrieval with concurrent observations of SMAP, for 20 cases (up to 40 m.s<sup>-1</sup>, graphic on Fig 1.36), and with SFMR data for major Hurricane Irma, respectively. No quantitative comparison has been made to date for a wide range of intensity categories with any SAR wind retrieval. An extensive exploitation of SAR images will be conducted in Chapter 3, to assess the efficiency of this new SAR algorithm in retrieving reliable hurricane force wind, and in probing inner core parameters. The benefits of ingesting SAR information into the ocean response analysis will be highlighted in Chapters 5 and 6.

### 1.5.4 Scaling Laws

The wake of tropical cyclones requires an analytical framework to probe its properties, to scale the thermal and sea level anomalies thanks to the collection of remote-sensed and in-situ parameters that mediate those fluctuations. The two-part study: Kudryavtsev et al. [2019a,b] is a first attempt to bring together several satellite observations around the near-inertial wake of three case studies (Jimena, Ignacio and Kilo), and to reproduce their patterns through near-inertial wave theory, which has benefited from little development since the last works of Ginis [2002]. These two studies are the main pieces that provide guidance to our thesis, with Kudryavtsev et al. [2019a] describing the complete analytical model and simulating some oceanic answers, while Kudryavtsev et al. [2019b] derives from it, the semi-empirical scaling that will be used in Chapter 5 & 6. In that respect, we will provide here some short insights of the model, its building equations, its main assumptions and the simplification that lead to the semi-empirical laws, in order to show how the other seminal studies are structured around the analysis. Reference articles can be found respectively in Annex D and Annex E for both the full model and the scaling laws.

This analytical model is mainly based on the mixing layer momentum and heat budget as described by Geisler [1970] and Price [1981] respectively:

$$\frac{\delta M_\alpha}{\delta t} + \epsilon_{\alpha\beta} f M_\beta = -gh \frac{\delta \xi_s}{\delta x_\alpha} + \tau_\alpha^s \quad (1.14)$$

$$\frac{\delta T_m}{\delta t} = -1/2 \bar{\Gamma} w_e - \frac{q_s}{h} \quad (1.15)$$

where  $\alpha, \beta = 1, 2$  are the axis indexes,  $\xi_s$  is the ocean surface displacement,  $f$  is the Coriolis parameter,  $h$  the mixing layer depth,  $T_m$  and  $M_{\alpha,\beta}$ <sup>12</sup> the sea surface temperature and the total wind driven transport,  $\tau_\alpha^s$  and  $q_s$  are the surface wind stress (scaled by water density) and kinematic heat fluxes,  $\epsilon_{\alpha\beta}$  is the unit asymmetric tensor ( $\epsilon_{\alpha\beta} = 0$  if  $\alpha = \beta$

<sup>12</sup> $M_{\alpha,\beta} = h u_{\alpha,\beta}$ , with  $u_{\alpha,\beta}$  the current velocity components.

and  $\epsilon_{12} = -1$ ,  $\epsilon_{21} = 1$ ),  $\bar{\Gamma}$  and  $w_e$ <sup>13</sup> are the vertically-averaged pre-storm temperature gradient and the entrainment velocity. Both equations are linearized, advection terms have been removed to focus on the turbulent mixing response. As seen in 1.3.1, the cold wake is mainly driven by mixing from the wind-induced shear current at the base of the ML, which resonates at near-inertial frequency, and excites NI waves that propagate through the shape of a 3D baroclinic wake, dependant on the stratification. ML budgets are thus resulting from the wind action in the ocean interior (i.e baroclinic response). To treat the coupling between the different layers, Orlanski and Polinsky [1983] provided the third governing equation that describes interior layer dynamic, phrased in terms of vertical velocities ( $w$ ):

$$(\Omega^2 - f^2)\hat{w}'' - k^2(\Omega^2 - N^2)\hat{w} = 0 \quad (1.16)$$

More precisely, equation 1.16 pictures the linear wave motion within the ocean interior, where hat over any quantity denotes Fourier transform,  $\Omega = k_1 V_{fm}$  is the frequency,  $k_\alpha$  is wave number component<sup>14</sup>, double prime indicates second derivative over the depth. Once it is coupled with the ML momentum it finally gives<sup>15</sup>:

$$(\Omega^2 - f^2)(\hat{w}_s - \hat{w}_h) - ghk^2\hat{w}_s = \hat{F} \quad (1.17)$$

1.15 and 1.17 are the two governing equations that depict the cold wake dynamic. Solving this system requires to determine the complete TC-baroclinic wake in the Fourier space, as a function of the wind stress transform ( $\hat{F}$ ). It is itself defined as the combination of the vorticity and divergence of the surface wind stress:

$$F = \underbrace{f \text{Rot}(\tau)}_{\substack{\text{induced} \\ \text{inertial} \\ \text{pumping}}} - V_{fm} \underbrace{\frac{\delta}{\delta x_1} \text{Div}(\tau)}_{\substack{\text{induced ekman} \\ \text{pumping}}} \quad (1.18)$$

The analytical model solves 1.17 for an ocean stratification approximated as a three-layer system with three associated constant values of  $N$ , to describe the vertical propagation of near-inertial internal waves of frequency  $\Omega \sim f$ . The baroclinic motions of each layer is thus related to specific vertical velocity of n-modes<sup>16</sup>:

$$w_n(x, z) = \begin{cases} a_{1n} W_n(x) \sin(N_1 z / C_n), & \text{at } -d < z < 0. \\ a_{2n} W_n(x) \sin(N_2 (D + z) / C_n + \phi_n), & \text{at } -D < z < -d. \\ w_n(x, D)(H + z) / (H - D), & \text{at } -H < z < -D \end{cases} \quad (1.19)$$

where  $a_{1,2n}$  are dimensionless vertical velocity amplitudes,  $W_n(x)$  defines the 2D field of vertical velocity for n mode and relies on the integral of the Fourier transform of the wind stress (1.17). Analytical solutions for 1.19 are provided and modified from Geisler [1970] (see Annex E).

From the computation of the different vertical velocities, the displacement of water masses at given depth can be then derived as an integration over the horizontal space:

$$\xi_n = V_{fm}^{-1} \int w_n(x, z) dx_1 \quad (1.20)$$

The total displacement is accordingly defined as the superposition of the different modes. Once these vertical motions elucidate for the ocean interior, the resulting surface baroclinic motion is described from the boundary condition as:

---

<sup>13</sup> $w_e = \delta h / \delta t + \delta M_\beta / \delta x_\beta$

<sup>14</sup> $k = (k_1^2 + k_2^2)^{1/2}$

<sup>15</sup>subscripts "s" and "h" means at the surface or at the ML base

<sup>16</sup>Seasonal Thermocline (depth d), Permanent Thermocline (depth D) and Abyss (depth H).

$$w_{sn}^{bc}(x) = (C_n^2/g)w'_{n|z=0} = a_{1n}(N_1 C_n/g)W_n(x) \quad (1.21)$$

The surface anomalies associated to baroclinic activity ( $\xi_s^{bc}$ ) can be estimated using the relation 1.20. Yet, sea surface displacements are also caused by the depth-averaged current related to the barotropic answer (section 1.3.1). To recall, SSHA is made from both the oscillating wake of near-inertial gravity waves (i.e baroclinic response) and the geostrophically-adjusted surface circulation (i.e barotropic) .Ginis and Sutyris [1995] demonstrated that these two responses are interaction-free and can be determined independently, so the total SSHA reads as a simple sum:  $\xi_s = \xi_s^{bc} + \xi_s^{bt}$ . The calculation of  $\xi_s^{bt}$  follows the same scheme as  $\xi_s^{bc}$ , and is inferred from 1.20 with the information of the surface barotropic vertical velocity  $w_s^{bt}$ , expressed as a non-oscillating motion acting on an unstratified ocean of depth H (see Annex E). These different stages help to solve the ML momentum budget and determine the full TC wave-wake. Subsequently, the knowledge of the baroclinic vertical velocity also allows to complete the ML description, by documenting the entrainment velocity, necessary to solve 1.15 which provides the SSTA field associated to the mixing process, and to close the model.

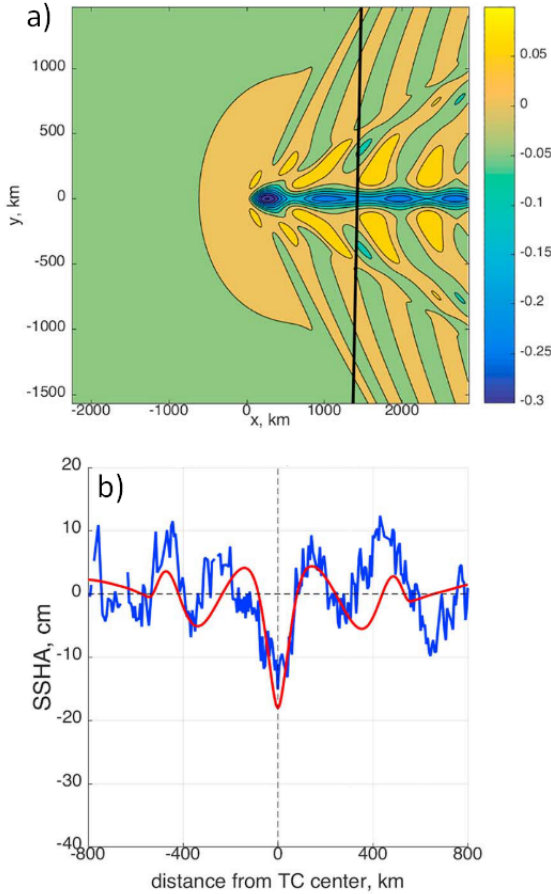


Figure 1.37: a) *Simulation of the surface wake of TC Jimena, with the location of the altimeter track (solid black line).* b) *The corresponding observed (blue) and simulated (red) SSHA anomalies along the altimeter track. Extracted from Kudryavtsev et al. [2019a]*

The Figure 1.37 outlines some of the possible explorations of the wake properties by the model. Of course, these analytical solutions of the cold wake are reached through several assumptions, essential to the simplification of the problem, such as :

- 1) The vertical stratification is solely driven by the temperature gradient (Eq. 1.15). Generally valid outside river plumes and intense pre-existing monsoon episodes (section 1.1.1).
- 2) The vertical stratification can be split into three constant gradient layers (Eq. 1.19). Especially true for tropical regions where a seasonal thermocline remains even in the late fall, in contrast to subtropical regions (section 1.1.3).
- 3) Mixing layer is a small perturbation in the ocean stratification, i.e in the wave-wake spreading process ( $h \ll d, D, H$ ). The vertical propagation of baroclinic motions is therefore first described for the three interior ocean layers (Eq. 1.19) with no explicit dependence to ML parameters. Ocean interior dynamic is then connected to ML parameters to finally solve heat budget (erosion of the ML).

These first three axioms set out the necessary considerations for the structure of the ocean, the following postulates are more concerned with wake properties and mixing mechanisms.

- 4) The frequency  $\Omega$  of NIW  $\sim f \ll N_{1,2}$ . The sharper gradient generally observed at tropical latitudes strengthens this hypothesis (section 1.1.3).
- 5) During the mixing process, Richardson number remains at a critical value as defined in Price [1981] and refined in Price et al. [1986]:  $Ri_{cr} = \frac{\Delta\rho}{\rho} \frac{gh}{u^2} = 0.6$ , with  $u$  the current velocity at ML base. It brings simplification to document the MLD progress:  $([2Ri_{cr}]^{1/4} \sim 1)$ .
- 6) The advection and heat flux terms ( $q_s/h$  in 1.15) are neglected from the ML budget (Eq. 1.17 & 1.15), which allows to linearize and drastically simplify the equations. The SST response is solely based on the baroclinic contribution. As we have seen in 1.3, this assumption is mainly fulfilled for TC cases with intensity higher than category 1, for which the wind-induced turbulent mixing is strongly dominant Price [1981]. It is even more true in the vicinity of the strongest surface anomalies and TC inner core [Vincent et al. 2012a].

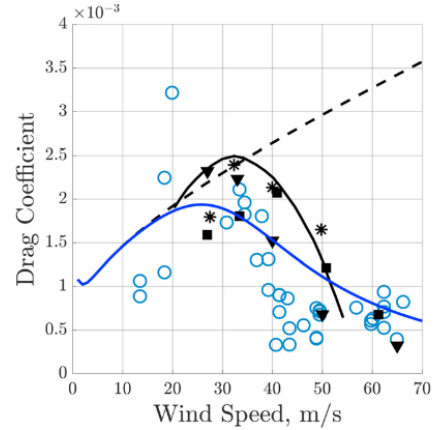


Figure 1.38: Drag coefficient as a function of wind speed at 10 m height. Dashed line is  $C_d$  calculated for the roughness scale predicted by the Charnock relation; black squares and stars are data from experiments conducted by Powell [Powell et al. 2003; Powell and Reinhold 2007]; black solid line, fitted quadratic curve to the empirical data by Jarosz et al. [2007]; triangles are estimates by Kudryavtsev [2006]; open circles show the drag coefficient derived from the altimeter SSH anomalies of Kudryavtsev et al. [2019a] with the associated trend in blue solid line.

- 7)  $F$  in 1.18 is reduced only to the vorticity term ( $Rot(\tau)$ ), at the origin of the resonance answer of the inertial current to the wind stress curl (simplification for Eq. 1.17). As illustrated by its hyperbolic shape in the panel a) of Figure 1.37, the TC wake is mainly a baroclinic phenomenon, which is also reflected by the alternating pattern of upwelling and downwelling cells ([Geisler 1970], see section 1.3.1). The resulting isopycnal displacement can be considered as an inertial pumping, different from the Ekman pumping caused by the divergence of the wind transport (second term 1.18), and most often associated with slow TCs and their barotropic response ([Greatbatch 1984; Jansen et al. 2010; Jullien et al. 2012]). Here, only the inertial pumping is considered.

Ultimately, based on this analytical model and its assumptions, some scaling laws can be extracted to directly diagnose the maximum amplitude of the wake anomalies. For instance, the heat budget defined in 1.15 can be written in the shape of the 1D heat conservation of Pollard et al. [1973]:  $\delta T = 1/2\Gamma h$ . As the stratification is only dictated by the temperature profile in our approach, the brunt-Väisälä definition (Eq. 1.2 in section 1.1.1

c) leads to the straightforward relation:  $\Gamma = -N^2(g\alpha)^{-1}$ , which only leaves us to relate the deepening to some summary parameters. From the scaling of the momentum conservation:  $M = \tau/f\Phi(R_{max}f/V_{fm})$  and the assumptions of  $R_{i_{cr}}$  from Price works on the deepening (point 5 above), we can read  $h$  as a function of:  $\frac{\tau^{1/2}}{f^{1/2}N^{1/2}}\Phi(R_{max}f/V_{fm})$ . Using  $\tau \propto V_{max}^2$ , we obtain a semi-empirical relation, whose coefficients are to be determined via a linear regression [Kudryavtsev et al. 2019b]:

$$\frac{SSTA}{\langle SSTA \rangle} \propto \Phi(R_{max}f/V_{fm}) \quad (1.22a)$$

$$\langle SSTA \rangle = \frac{V_{max}N^{3/2}}{g\alpha f^{1/2}} \quad (1.22b)$$

where  $\langle SSTA \rangle$  is a scaling term. Similarly, a simplified rule can be extracted for SSHA, based mainly on the simplifications from the model and Geisler [1970] observations. Considering the pressure continuity condition at the surface (eq. 1.21) with the shape of the vertical velocity ( $w(z) \propto \sin(Nz/c)$ ) in the seasonal thermocline for a three-layer stratification ( $h$ : small perturbation), we can relate the TC-induced baroclinic displacement of the thermocline with its associated signature in surface:  $\delta\xi_s^{bc} \approx (NC/g)\delta D$ . Geisler [1970] showed that this displacement is proportional to:  $\delta D \propto \tau R_{max}/(cV_{fm})$  and thus to  $V_{max}^2$ . Substituting  $\delta D$ , we can extrapolate a synthetic law:

$$\xi_s^{bc} g/V_{max}^2 \propto R_{max}N_1/V_{fm} \quad (1.23)$$

1.22 & 1.23 govern the baroclinic response at the ocean surface. They integrate the vertical column processes occurring in the TC wake through consistent scaling, involving a limited number of parameters to retrieve from observations, in addition to monitoring the geophysical variables of the ocean response (SSH,SST):

- TC forcing parameters:  $V_{max}$  and  $R_{max}$ .
- Coupling parameter:  $V_{fm}$ .
- Environmental parameters:  $f$  and  $N_1$ .

This analytical structure, with its assumptions and "philosophy", is the cornerstone on which the other founding studies (from 1.5.1 to 1.5.3) gravitate and which together build the thesis. This two-part study is also a first concrete exploitation of the along-track altimeter data capability (panel a/b Fig 1.37) illustrated in 1.5.2, it does not include however the capability of wind-derived satellite information (1.5.1, 1.5.3) neither in-situ information. Chapter 5 and 6 will pursue this effort by merging the different sensors abilities, presented in this section, into the analytical framework. They will also keep on exploring the scaling laws for a basin-case study (Chapter 5) and more extensively in a thorough analysis at global scale, in Chapter 6. Furthermore, since the wind stress  $\tau$  is expressed as:  $\tau = (\rho_a/\rho_w) C_d V_{max}^2$ , the scaling law can be inverted to probe the air/sea interaction through the estimation of the drag coefficient  $C_d$  (Chapter 7), as shown in the panel c of the Figure 1.37. At last, the forcing parameters in the scaling laws can also be derived from a parametric wind solution, like it is already the case in the full analytical model of Kudryavtsev et al. [2019a] with the Holland profile as wind stress input. Yet, in the light of the shortcomings of Rmax-based model (section 1.3 and 1.4.3), we will bring a last piece to our approach for the thesis, based on a recent parametric solution.



### 1.5.5 Chavas Parametric model

Finally, in order to generalise our approach to a wider range of cases, we will use an innovative parametric solution: the Chavas model, whose performance have been outlined in the two-parts study [Chavas et al. \[2015\]](#) and [Chavas and Lin \[2016\]](#). This model merges two solutions based on the partial conservation of angular momentum within the TC circulation, between the inner core and the outer core. The associated equilibria describe the thermodynamic regimes that drive the vorticity of the two distinct regions. They translate, in short, the primary and the secondary TC circulation into two synthetic equations:

For the inner core:

$$\frac{M(r)^{2-(C_k/C_d)}}{M_{R_{max}}} = \frac{2(r/R_{max})^2}{2 - (C_k/C_d) + (C_k/C_d)(r/R_{max})^2} \quad (1.24)$$

For the outer core:

$$\frac{\delta M(r)}{\delta r} = \frac{2C_d}{W_{cool}} \frac{(rV)^2}{r_o^2 - r^2} \quad (1.25)$$

where  $M_{R_{max}} = R_{max} V_{max} + \frac{1}{2} f(R_{max}^2)$ , and  $W_{cool}$  is an environmental parameter describing the radiative-subsidence rate in the free troposphere. Equation 1.24 provides a parabolic solution describing the deep convective mechanisms in the inner core, while equation 1.25 depicts an hyperbolic curve that represents the influence of the quiescent free-troposphere thermodynamic on the relative vorticity of the outer descending region. The combination of these two thermodynamically-independant regions into a singular TC-wind radial profile defines the actual Chavas model [[Chavas et al. 2015](#)].

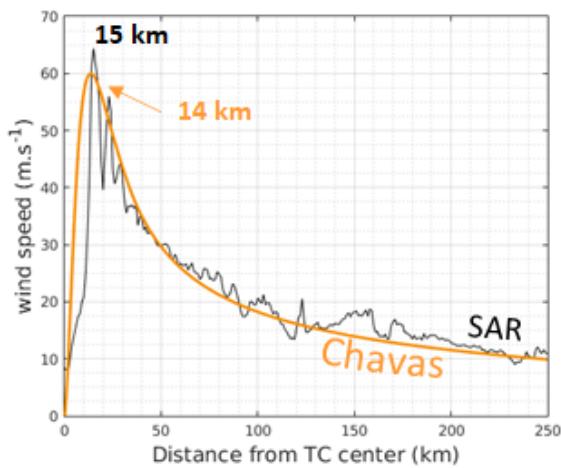


Figure 1.39: *Similar than Figure 1.32.a. but with Chavas wind profile.*

One of the most appealing features of this merged solution is to offer different modes of variability, i.e. it can be forced and fully specified by two different pairs of storm parameters :  $(V_{max}/R_{max})$  or  $(V_{max}/R_{outer})$ . These two pairs fix either the inner  $(V_{max}/R_{max})$  or the outer solution  $(V_{max}/R_{outer})$ , the other being then adjusted to provide an unique complete profile. For the second method, the  $R_{outer}$  is chosen in an interval between R10 and R17, as these wind radii are generally seen as indicative of the storm size. The advantage of using a radius from the outer profile is twofold. As already discussed (section 1.2.3), the external profile is easily accessible by most observations and with good accuracy [[Knaff et al. 2021](#)], unlike  $R_{max}$ , low-to-medium resolution MW instru-

ments can document the outer core at a high rate sampling (section 1.2.3). Secondly, as long been discussed by [[Merrill 1984](#); [Holland and Merrill 1984](#)], the intensity, strength and size of the cyclone are weakly correlated. If certain behaviours could be highlighted when the TC phase is taken into account, [Chavas and Lin \[2016\]](#) also showed that the fluctuations of TC parameters obey different variability patterns: an intra-storm one, corresponding to changes during the lifetime of a specific cyclone (linked to the phase), and an inter-storm pattern, corresponding to fluctuations observed between the different

TCs. While the TC size is relatively stable over its lifetime (in comparison to others parameters), it varies drastically from case to case, and cannot be inferred from  $V_{max}$  or Coriolis parameter, as is often the case with Rmax-based models. Thereby, the parametric model of Chavas with the pair  $(V_{max}/R_{outer})$  raises interesting perspectives to handle the decreasing gradient of the wind profile by forcing with an outer radius. This is briefly illustrated in the Figure 1.39 with the case of Hector shown earlier (Fig 1.32). Unlike previous wind models (section 1.4.3), both the eyewall and the outer regions are well captured as compared to the SAR profile, with very close  $R_{max}$  values and a well-depicted wind gradient.

The information of an external radius is provided by a concomitant MW radiometer or scatterometer observation. Despite their limitations in determining the parameters of the inner core, the combined use of this parametric solution with these low-to-medium resolution instruments can be an alternative method to reconstruct the complete wind profile, under certain conditions which must be specified. In that regard, Chapter 4 will examine the reliability of this method to infer both the inner and outer core from the information of an outer radii and Vmax, while Chapter 6 will try to demonstrate its capability to document the forcing parameters of the scaling laws. Finally, Chapter 6 will also put into perspective the use of this parametric solution as a consistent wind stress input for the full analytical model of the ocean response.

In this introductory chapter, we have drawn a relatively exhaustive picture of the various components that revolve around cyclone events. From the description of the ocean dynamics to the mixing processes occurring during TC forcing, this succession of concept have helped to establish the context for the thesis and to introduce the spaceborn instruments that are innately linked to the analysis of cold wakes. The numerous studies quoted were also intended to expose the grey areas and the limitations related to our topic, for which the thesis will attempt to provide leads through the solutions proposed by the pioneering work we have just highlighted. Throughout the six chapters that gather our results, we will try to syncretize the different observational approaches advocated by this section, within a semi-empirical framework able to solve most of the issues addressed around the predictive analysis of the cold wake anomalies. To this purpose, the second part will be devoted to the analysis of the cyclonic vortex, whose three constituent chapters bring together efforts to extract the essential parameters of the outer (Chapter 2) and inner core (Chapter 3), and to reconstruct the complete wind profile using a parametric model. The third part will take advantage of these information and incorporate them into the analysis of the oceanic response via the scaling laws. The chapter 5 will first detail the progressive addition of ocean observations and cyclonic forcing into a consistent methodological framework, and in a second time, Chapter 6 will further analyse the variability of the surface wake signatures across the entire database and in relation to the deeper anomalies observed by Argo. The concluding chapter (Chapter 7) will investigate the drag coefficient behavior according to the different forcing input.

# **Part II**

## **Tropical Cyclone Wind-Field.**

The first step of our research work was to get inside the cyclonic vortex, in order to extract the essential parameters for the description of the forcing (inner+outer core). Figure 1.40 shows the sections of the wind profile seen through the eyes of the different sensors and analyses that this thesis will exploit. From the discrete and smoothed information provided by the BTK ( $V_{max}/R_{max}$ , wind radii), the outer and near-inner core profile captured by the low to medium resolution MW instruments, to the high resolution SAR measurements that allow a complete portrayal of the TC, within its swath.

While radiometers and scatterometers allow rapid coverage of the ocean with high sampling rates, SAR instruments suffer from greater temporal intermittency due to their acquisition procedure. However, in the case of real measurements, they are an ideal and unique source of information. The strengths and limitations of each of these observations must therefore be considered. The first two chapters of this forcing section will focus on these observational capabilities, and will follow a clear strategy.

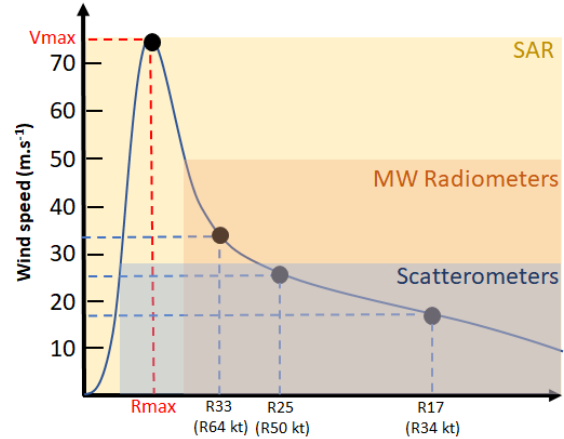


Figure 1.40: *Sketch of a radial hurricane wind profile and the structures generally visible from the MW satellite instruments. Black dots represents the wind radii available in the BTK.*

**Chapter 2** focuses on the direct exploitation (L2/L3 product) of "classical" satellite methods, commonly used by operational community. It has a double objective. 1) To build the backbone of our multi-platform database, the 6-hourly synoptic positions of the BTK provide a spatio-temporal skeleton on which the different observations will be aggregated. They also provide an estimate of the intensity and radius of the maximum winds at each time step. 2) The corpus of radiometric and scatterometric instruments should document very broadly the size and structure parameters of a large part of the tropical systems from 2010 to 2018. The representativeness of our sampling over the entire cyclonic activity of this period will be a convincing proof of the current strength of MW satellites.

**Chapter 3** aims to synthesise the efforts of the last two decades to validate the potential of SAR to investigate cyclonic force winds. It builds on one of our papers published in Monthly Weather Review in 2020, using one of the largest samples of SAR imagery collected during cyclonic events between 2015 and 2018. This unprecedented base has enabled us to carry out several missions.

- 1) To extensively demonstrate the capability of SAR instruments to infer both the structure of the extreme winds from the inner-core and the outer profile, and to compare them with a reference measurement (SFMR).
- 2) To show that SAR measurements can capture the variability associated to limiting situations.
- 3) To quantify precisely the uncertainties associated with the BTK estimates, including those of  $R_{max}$  that have not previously been calculated.
- 4) To illustrate extensively the contribution of cross-polarised signal technologies to the estimation of extreme winds.

The combination of these two chapters will vouch for the robustness of the future instrument constellation (cross-polarised scatterometers).

In order to propose a generalisation method to the high resolution but intermittent SAR approach, our work on forcing will lead to the use of a parametric wind profile. This solution is based on the Chavas model which, from the information previously extracted from BTK and low/medium resolution MW satellites, deduce the structure of the inner core ( $R_{max}$ ) and reconstruct the radial wind profile. In order to obtain the complete TC wind field, it is however necessary to integrate in this model, initially in 1D, information on the asymmetry. The results and findings of these experiments are decanted in the last chapter of this first part.

**Chapter 4** inherits from a long series of studies on the size and structure of TCs. The duality of the Chavas model (merge of two solutions) may lead us to question the real limits between inner and outer core with respect to these performances as regards to satellite observations. This chapter must therefore, in addition to proposing a generalisation method, enfold the different MW instruments of the two other chapters, which will be used as a source of comparison. The final challenges of this part are indeed multiple.

- 1) Validate the parametric model over the entire radial wind profile.
- 2) Provide an auxiliary source of information on  $R_{max}$ , based on easily observable parameters.
- 3) Capture asymmetries and intra/inter-peak variabilities.
- 4) Establish a validity framework for the parametric solution and generalise our approach.

At last, a graph and tables summarising the organisation and data of the following three chapters are provided.

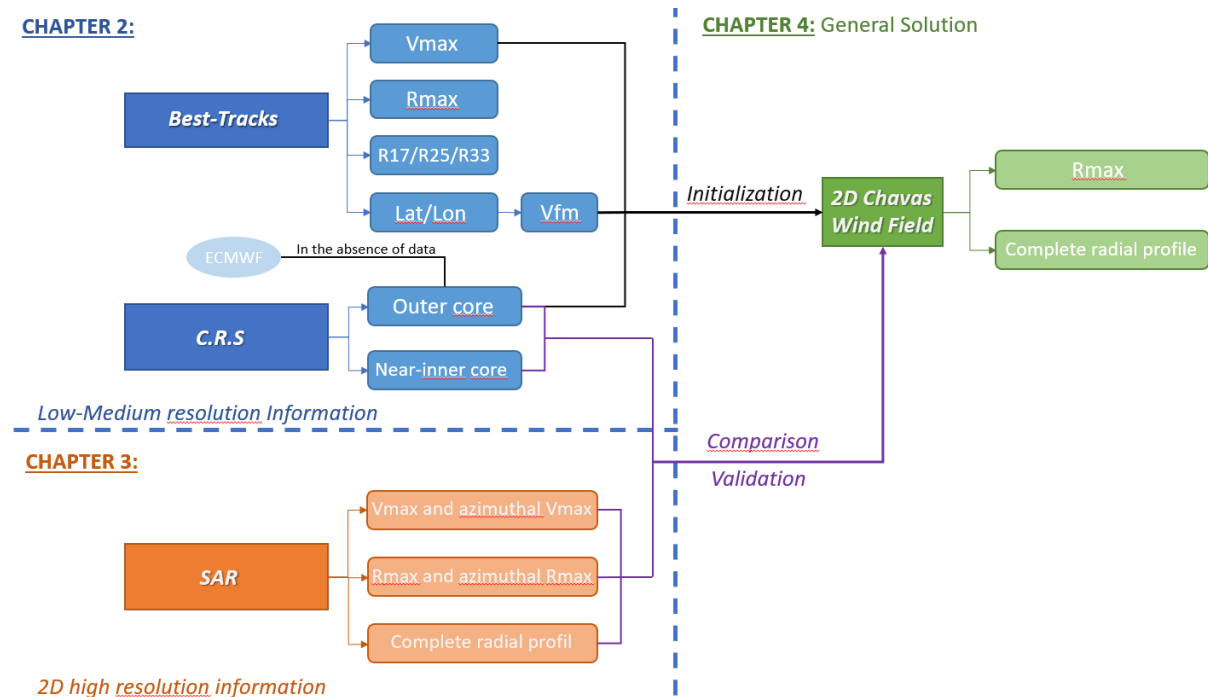


Figure 1.41: Schematic overview of the chapters in the Forcing part of the manuscript.



Table 1.2: List of low/medium resolution wind-derived sensors.

| Instrument   | Metop-A/B<br>ASCAT-A/B | SMOS<br>MIRAS | SMAP<br>SMAP | GCOM-W<br>AMSR2 |
|--------------|------------------------|---------------|--------------|-----------------|
| Type         | Scatterometer          | Radiometer    | Radiometer   | Radiometer      |
| Chanel       | C-band                 | L-band        | L-band       | multi           |
| Polarization | VV                     | H,V           | H,V          | H,V             |
| Orbit        | SSO                    | SSO           | SSO          | SSO             |
| Resolution   | 12.5 km                | 40 km         | 40 km        | 50 km (LF)      |
| Center/GMF   | ESA/KNMI               | ESA/LOPS      | ESA/REMSS    | JAXA/SOLab      |
| Level        | L2                     | L2            | L3           | L2              |
| Years        | 2006 (B/2012)-now      | 2010-now      | 2015-now     | 2012-now        |
| Database     | 2010 (B/2012)-2018     | 2010-2018     | 2015-2018    | 2012-2018       |
| Bassin       | All                    | All(no NI)    | All          | All             |
| Samples      | 1234 (B/957)           | 678           | 725          | 1900            |
| TC parameter | Router                 | Router        | Router       | Router          |

Table 1.3: List of high resolution wind-derived sensors.

| Instrument      | S1A/B<br>SAR-C        | RS2<br>SAR              | WD-3P<br>SFMR |
|-----------------|-----------------------|-------------------------|---------------|
| Type            | SAR                   | SAR                     | Radiometer    |
| Chanel          | C-band                | C-band                  | C-band        |
| Polarization    | Dual<br>(VV+VH/HH+HV) | Quadri<br>(HH,VV,VH,HV) | Linear        |
| Orbit           | SSO                   | SSO                     | N/A           |
| Resolution(GMF) | 3 km                  | 3 km                    | 1 km          |
| Center/GMF      | ESA/LOPS              | CSA/LOPS                | AOML+HRD      |
| Level           | L2                    | L2                      | L2            |
| Years           | 2014/2016-now         | 2007-now                | 1970-now      |
| Database        | 2016-2018             | 2015-2018               | 2016-2018     |
| Bassin          | All(no NI)            | All(no NI)              | AT+EP         |
| Samples         | 110                   | 59                      | 28            |
| TC parameter    | 2D-full               | 2D-full                 | 1D-full       |

Table 1.4: List of products used.

|                | ECMWF                | CFSv2     | Blended                 | BTK                                                        |
|----------------|----------------------|-----------|-------------------------|------------------------------------------------------------|
| Type           | Forecast             | Forecast  | Analysis                | Reanalysis                                                 |
| Resolution     | 0.125/3h             | 0.20/1h   | 0.25/1h                 | 6h                                                         |
| Center/Version | ECMWF                | NCAR      | Oceandatalab            | 1)NCDC/IBTrACs<br>2)JTWc/ATCF                              |
| Database       | 2010-2018            | 2010-2018 | 2015-2017               | 2010-2018                                                  |
| Input          | Scatterometers       | sat       | SMOS,<br>SMAP,<br>AMSR2 | SFMR, buoys, plat-<br>form, dropsondes,<br>IR, MW scat+rad |
| TC Parameters  | Router/<br>Asymmetry | Router    | Router                  | R17/R25/R33,<br>Rmax/Vmax, Lat,<br>Vfm                     |

# Monitoring of TC from low-to-medium resolution microwave remote sensors: investigation of the outer core profile.

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Our first chapter of Part II, although very short, represents the work done on the current constellation of passive and active microwave instruments, which derive the surface wind from brightness temperatures or surface roughness at resolutions ranging from 12.5 to 40 km. The results presented in this chapter are not meant to be innovative, but they are an essential step in the creation of our multi-platform satellite database, and an important source of information on cyclone structure and size that will be useful to provide a spatial framework for our analyses or to guide a parametric model.

## 2.1 Data description.

In order to create the skeleton of our multi-platform database, we will rely on three different types of data.

Firstly, **the Best-track (BTK) data**, which is a complete archive of the 6-hour positions of each tropical system, from the first signs of deep precursor disturbances to their development into a tropical cyclone and dissipation. In most cases, only systems that have exceeded the gale force wind threshold are given a name and considered in the total seasonal cyclone activity count (see Fig 1.13). While intensity and center positions (Lat/Lon) are the historical parameters that have been available since the genesis of the BTK analysis, it has been enriched with other estimations due to the needs of the scientific community, including the maximum observed extents of the gale- (r17), damaging-(r25) and hurricane- (r33) force wind radii distributed over the four geographical quadrants

[Cangialosi and Landsea 2016], but also information on the Rmax and ROCI/POCI (radius/pressure of the outermost closed isobar); they generally all benefit from a postseason reanalysis (except Rmax) [Sampson et al. 2018]. The six-hourly positions of Hector and some of the parameters estimated by the BTK are displayed in the Figure 2.1.

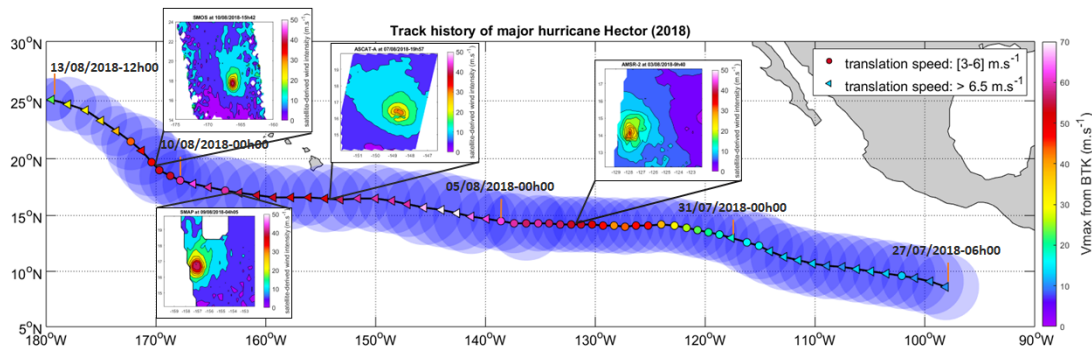


Figure 2.1: 6-hourly Hector BTK positions as a function of vortex intensity (colorbar) and motion speed (average in circles and fast in triangles). The blue shaded area indicates the extent of the radius of the outermost closed isobar (ROCI). The four satellite images representing the MW instruments in the figure are also projected temporally on the map.

The techniques on which the estimates are based have already been presented in Chapter 1, Section 1.2, and are mainly derived from the interpretation of IR/Visible observations and MW sounders, or from in-situ/platform data and SFMR measurements when available [Landsea and Franklin 2013]. The limitations are thereby mostly driven by these indirect methods, especially in reading the properties of the inner core (elusive Rmax, rapid intensification/weakening issues). It should be remembered that Best-track analysis also remains by definition a subjectively-smoothed representation of the cyclone track history, and that abrupt changes in structure, intensity and trajectory, over periods of less than 24 hours are not necessarily well taken into account in order to be also representative of the situation around the six hour positions [Landsea and Franklin 2013], its performance is therefore constrained both by the instrumental capabilities of the time and by its smoothing nature.

Significant progress has already been made in the quality of the estimates (Reduction of 20% and 75% of uncertainties related to cyclone intensity and position respectively, [Landsea and Franklin 2013]), with the data we use over the 2010-2018 period benefiting from the contribution of scatterometer data [Brennan et al. 2009], the development of the Advanced Dvorak Technic (ADT) [Olander and Velden 2007], better surface adjustments of the flight-level winds [Landsea et al. 2004] and a more routine use of SFMR data [Uhlhorn and Black 2003]. The contributions of the L-band radiometers are not noticeable here, as their incorporation is subsequent to our study period [Bushnell et al. 2018]. Of course, the methods used to assess and forecast cyclone activity are specific to each operational meteorological centre, and the data and conventions used, as well as the weight attributed to each observation in the analyses, fluctuate greatly from one region to another [Knapp et al. 2010]. The various tropical basins are in fact divided into several sectors and are themselves supervised by an official entity (6 RMSCs, 4 TCWCs) to which are added the complementary analyses of other centres operating with their own methodology (e.g. JTWC) [Knapp et al. 2010]. It can therefore be as challenging to gather TC tracks as it can be contradictory between the different analyses [Kruk et al. 2010]. No less than four different advisories are provided for the WP basin [Kruk et al. 2010], where the

greatest discrepancies are also found between centres [Schreck et al. 2014], with sometimes deviations largest than 15 kt ( $\sim 7.7 \text{ m.s}^{-1}$ ), possibly due to different constraint rules in the Dvorak technique [Knaff and Harper 2010].

To cope with this heterogeneity, the International Best track Archive for Climate Stewardship, the so-called IBTrACS has been built [Knapp et al. 2010]. Once the post-season reanalyses have been carried out by all the centres in the different basins, this archive collates and attempts to harmonise this information into a coherent and consistent database, producing averaged TC positions and intensities (normalized to 10-min wind) from the values of all available reports, although individual centre estimates are also kept with their conventions [Kruk et al. 2010]. For our study, we chose to refer to the merged positions from IBTrACS, which will also be used later to calculate the Coriolis parameter and the associated translation speeds. For all other parameters (wind radii,  $R_{\max}$ ,  $V_{\max}$ ,  $ROCI$ ), however, we chose to focus on the individual estimates generated by the combined US centres NHC and JTWC, which together provide a global coverage of hurricane activity [Schreck et al. 2014]. We have chosen this option merely for consistency, as both centres rely on broadly the same routines with the deployment of the forecasting and analysis system known as ATCF (Automated Tropical Cyclone Forecasting system). They also both provide intensity estimates based on the same convention, the maximum 1-min sustained winds, which is considered closer to the high resolution observations used in Chapter 3.2. The conversion of wind averaging periods remains quite hazardous, the techniques used simply provide an approximation as it is normally impossible to convert from a gust (e.g the maximum 1-min sustained winds) back to a specific time-averaged mean wind speed like the 10-min winds [Harper et al. 2010]. It is therefore more convenient to work with the uniform NHC and JTWC measurements. In addition, Song and Klotzbach [2016] suggested that the JTWC provides better estimates for the WP area.

Secondly, we rely on **micro-wave satellite** acquisitions of three passive and two active instruments, namely the two L-band radiometers **SMOS** and **SMAP** provided by IFREMER and REMSS respectively [Reul et al. 2016; Meissner et al. 2017], as well as the multiband radiometer **AMSR-2** with the SOLab wind retrieval method [Zabolotskikh et al. 2016], and finally the two active sensors **Ascatter-A and B** of the Metop satellite series, provided by KNMI [Verspeek et al. 2010]. They all display very heterogeneous characteristics, acquisitions made by each sensor for the Hector case are illustrated in Figure 2.2. Firstly, the two L-band radiometers can cover a wide intensity spectrum ranging from 12 to 70  $\text{m.s}^{-1}$  over a large swath of 1000  $\text{km}^1$ , even though they are affected by a coarse resolution of 40 km, they are relatively insensitive to rain which make them all-weather satellite [Meissner et al. 2017; Reul et al. 2017]. Unlike these two peers, AMSR-2 is a multi-frequency radiometer that measures winds in the range of 0-70  $\text{m.s}^{-1}$  with slightly better resolution due to its higher frequency bands ( $\sim 25 \text{ km}$ ) and provides the largest swath among these different sensors (1450 km), but is much more sensitive to rain issues [Zabolotskikh et al. 2014]. At last, Scatter A and B, offer the finest resolution of 12.5 km but are constrained by the inherent properties of their measurement (saturation for winds above 25  $\text{m.s}^{-1}$ ) which limit their field of application to winds ranging from 0 to 25  $\text{m.s}^{-1}$  with high accuracy around 1.5  $\text{m.s}^{-1}$ .

Finally, we use the **ECMWF** global products (forecast product) and the **CFSR** analysis from NCEP as a potential backup method, in case of incomplete data. The other hidden purpose of these comparisons is to check the consistency of these results in the

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<sup>1</sup>L-band radiometers have particular issues with low to moderate wind speed, due to a lack of sensitivity [Reul et al. 2012]



light of previous studies [Reul et al. 2017; Schenkel et al. 2017]. In addition, the Ascet data are assimilated into ECMWF, it is therefore also a kind of calibration of our data, to ensure the quality of our methodology.

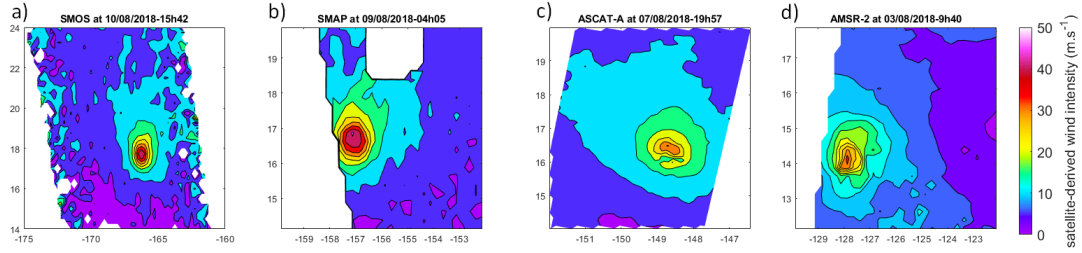


Figure 2.2: *Examples of satellite acquisition of TC Hector by: a) SMOS, b) SMAP, c) Ascet-A and d) AMSR-2.*

## 2.2 Extraction of wind radii from MW data.

In order to exploit the satellite data, our methodology consists of three main steps:

- 1) Find the cyclone passages in the satellite swaths.
- 2) Locate the TC center.
- 3) Extract a panel of wind radii.

1) The first step requires the positions of the Best-Tracks to intercept adjacent satellite measurements in time and space. For each of the cyclones in our database, we select the start and end dates of the trajectory, in order to have the temporal reference for image collection. We only keep the satellite swaths intercepting the BTK positions in this time interval. From all the images collected, we extract the acquisition times around the points that intercept the cyclone path. If these times meet the time interval described by the bordering BTK points, it is indeed a concordant acquisition. This method allows us to discriminate between possible concurrent cyclonic events (necessary for the eastern Pacific basin for example). We then keep all measurements within a 15 degree zone around the intersection points to eliminate superfluous data. Once these images have been collected, we apply the same filters as in some previous studies based on QuikSCAT observations [Chavas and Emanuel 2010; Chan and Chan 2012]. Namely, we keep only systems larger than  $20 \text{ m.s}^{-1}$  to avoid too weak structure with unorganized circulation. More than half of the cyclonic circulation must be located in the swath, with a center necessarily located inside it. At last we eliminate data with too much land contamination or noise.

This first process allowed us to obtain a coherent base of more than 5000 passages, proof of the incredible tracking capacity of the current constellation. However, in the end we only used a small fraction of this data ( $\sim 600$ ). Indeed, the second step consists in fixing the TC centers. While this step is the most crucial, as it determines the quality of the ray measurements, it often requires some manual checking in addition to the automated process, which limits the size of our analysis sample.

2) We then search for the center in each of the collected scenes. This step is very dependent on the properties of the instruments, especially in terms of resolution. We

interpolate the 6-hourly BTK positions every half hour, assuming a linear displacement between each adjacent points. The interpolated position closest in time to the image is used as a first estimate. We then check whether the first estimate lies within a specified isocontour (depending on the storm intensity, usually r17/r25). If not, we look for the maximum intensity as the first estimate. Once this step is verified, we look for the area of weakest winds within the isocontour, which is then defined as the final TC center. However, we set an intensity criterion that must be satisfied (must  $<15 \text{ m.s}^{-1}$ ), if this condition is not met, we keep the first estimate. For most Ascet data, this process does not require any manual checking, as the center is correctly fixed into the visible eye. For radiometric data, such as SMAP and SMOS, however, a manual control is required due to their coarser resolutions.

3) Once the centers are fixed, we transform the initial grid into polar coordinates, in order to obtain the different wind radii. All swaths are interpolated to a grid with a resolution of 10 km in radius, and  $1^\circ$  in azimuth. For each angle, we then look for the maximum intensity of the transect, in order to eliminate the wind radii located upstream of the eyewall. We then look for the radius values for each intensity thresholds (r10 to r25). This last step grant us an azimuthal distribution of several wind radii.

At last, from the centers of the satellite images, we collocate our satellite data with the CFSR and ECMWF wind fields at the nearest time points. We then apply the same processes to find the centers and extract the wind radii. These analyses are much easier on this type of product because of the smoothness of their wind fields.

## 2.3 Analysis and Climatology of TC wind radii estimates.

### 2.3.1 Consistency of MW observations.

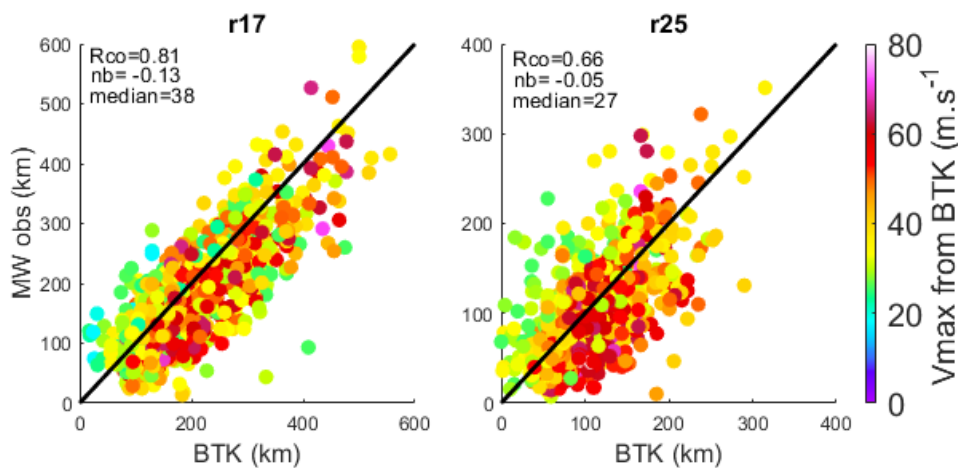


Figure 2.3: *Comparison of the r17 and r25 from BTK and MW observations.*

In order to assess the consistency between the different measurements, we first compare them with the BTK data (Figure 2.3). Since the scatterometer saturate at hurricane-force winds, we focused on the first two characteristic wind radii, r17 and r25, which also have

the advantage of suffering from lower uncertainties than the r33 ( $> 50\%$ , [Landsea and Franklin 2013]). In addition, the r17 from BTK is mainly informed by the scatterometer information, whose optimal measurement range (both in wind direction and intensity) is in this range of values [Chou et al. 2013], which will allow us to evaluate the quality of our methodology. Although SMOS and SMAP acquisitions are now also integrated in the BTK analyses, their incorporation is subsequent to our time range. For the r17, all three instruments are found very consistent with each other in their behaviour and in the quality of their estimates. They are in very good agreement with BTK data, with a slight negative bias and a median scattering of 38 km, which are in lines with other studies for radiometers [Reul et al. 2017; Zhang et al. 2021]. The R25 also shows good agreement between the two sets of estimates, but with a moderate decrease in correlation. In fact, two attitudes can be distinguished, those of the radiometers and the ASCAT radars. Indeed, if we separate the statistical properties individually per instrument, the radiometers show correlations roughly similar to those of R17, while the correlation of the Ascet sensors drop to about 0.5 and shows a negative bias. Both results are expected in the literature [Chou et al. 2013; Reul et al. 2017]. Overall, both wind radii are very well interpreted by the three radiometers and show similar performance to that observed in previous analyses.

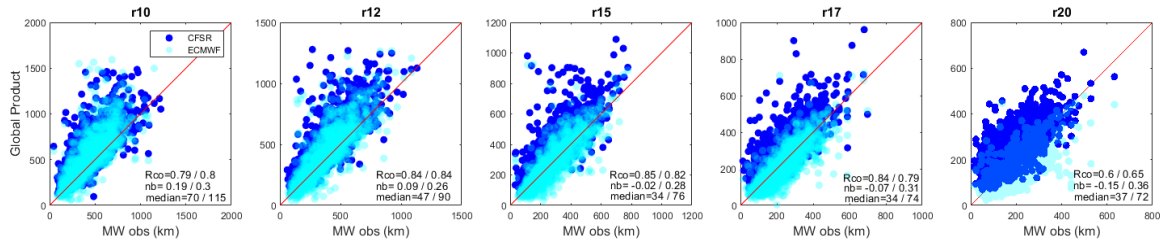


Figure 2.4: Comparison of wind radii from ECMWF and CFSR with MW observations. statistics values on the left stand for ECMWF comparison, and CFSR on the right.

Now that we have ensured the quality of our extraction methodology and the congruence of our observations, we would like to observe whether the global product estimates coincide with those of the MW measurements. We do not seek here to evaluate the quality of these products, nor of their data assimilation methods, but simply to find an alternative solution to complete our wind radii in case of incomplete data, either because of swath limitations or because the wind field is located too close to the edges. Our goal is to force a parametric model with a wind radius to inform the shape of the outer core profile, so this radius must be a good metric for the absolute length scale of the cyclone. Most metrics used in climatologies of TC size are in the range [r10 r20] [Chan and Chan 2014; Chavas et al. 2016], the Figure 2.4 shows the comparison of these radii between the MW data and the ECMWF and CFSR estimates.

Two main findings emerge from these results. The first is the systematic overestimation and large deviation of the CFSR data, despite an excellent correlation with the MW observations. The second is the excellent congruence of the ECMWF with the MW observations. Stopa and Cheung [2014] had already noted this problem of global widening of the cyclonic structure by CFSR, which explains the good correlation observed but the biases and discrepancies that exist. Although CFSR has other advantages such as a better description of the cyclone intensity [Murakami 2014; Stopa and Cheung 2014; Schenkel et al. 2017], in the scope of our study we have to eliminate this data source as a possible solution. As for ECMWF, its estimates coincide very strongly with the measurements of the different radii and display excellent correlations, although there is a slight decrease

in r10 with a much larger median deviation and a non-negligible bias, as well as a loss of coherence in r20 with a lower correlation. The discrepancies in the two marginal wind radii (r10 & r20) of the analysis are due to instrumental limitations and heterogeneity. The figure C.1 in Annex C discretizes the individual contributions of each instrument for the different radii. It shows essentially the agreement of all MW sensors for the radii from r12 to r17. Outside this range, two groups of trends can be observed, similar to the one in the previous figure: the radiometers on one side and the scatterometers on the other. The drop-out observed for r10 is due to the limitations of the radiometers for intensities below 12 m.s-1 [Reul et al. 2012]. On the other hand, while the loss of correlation observed for r20 also stems from the the radiometers that are starting to deviate slightly from the behaviour of the scatterometers (Figure C.1), it mainly reflects the limitations of ECMWF in providing information for more intense wind radii. Indeed, the forecast product is in perfect agreement with the Ascet data over the whole range of radii, which is rather reassuring given the strong assimilation of the ASCAT data, but it implies a loss of performance of ECMWF for wind radii beyond r20, given the saturation of the scatterometers and the divergence already observed with the BTK comparison for r25. However, despite this loss of quality for more intense wind radii, ECMWF fulfils the criteria to correctly inform on the size of the cyclone, specifically on radii from r10 to r17.

From these two figures, we can extrapolate that our observations are very much in line with the existing bibliography, the measurements of the different instruments are very consistent with each other for the range [r12 r20], above, the scatterometers saturate, below, the L-band radiometers suffer from a lack of sensitivity of the emissive properties of the ocean at this frequency and for low to moderate wind intensities [Reul et al. 2012; Meissner et al. 2017]. The preferred radii are therefore r12, r15 and r17, due to the instrumental properties and ECMWF performance. Taking into account the bias and the smallest median deviations, we have chosen to focus on r17. This wind radius has the advantage of being often more complete and allows us to deeply rely on the observations, it also corresponds to the radius where we observe the maximum convergence between the different sensors, which attests to the general quality of the r17 estimates.

### 2.3.2 Distribution of TC size.

Finally, the extraction of some radii indicative of the cyclone length scale (r10, 12, r15 and r17) allows us, like Chavas and Emanuel [2010]; Knaff et al. [2014]; Chan and Chan [2015a], to evaluate the relationship between size and latitude. This finding is important, as most wind models determine the size from the Coriolis parameter. The results from Figure 2.5 converge with the conclusions of the studies of Chavas et al. [2016]; Chan and Chan [2014, 2015b], namely that the size does not follow a monotonic relationship with latitude, even if some trend can be found. This behaviour is observed for both r12 and r17. It follows a first rise with a peak around the 20-25°N/S band, as previously demonstrated by observations and numerically [Chan and Chan 2014, 2015b], and then decreases at the edge of the tropical basins before rising again at the mid latitudes. If we look at the 25% of the strongest values (small graph inside), the vast majority also comes from these particular bands (> 40% occurrence) of 20-25° N/S band. The second rise observed for systems above 45°, corresponds to extratropical transitions that do not respond to the same mechanisms and that out of scope of the analysis. This non-monotonic trend is an important outcome, as it will be used to justify our choice of parametric wind model.

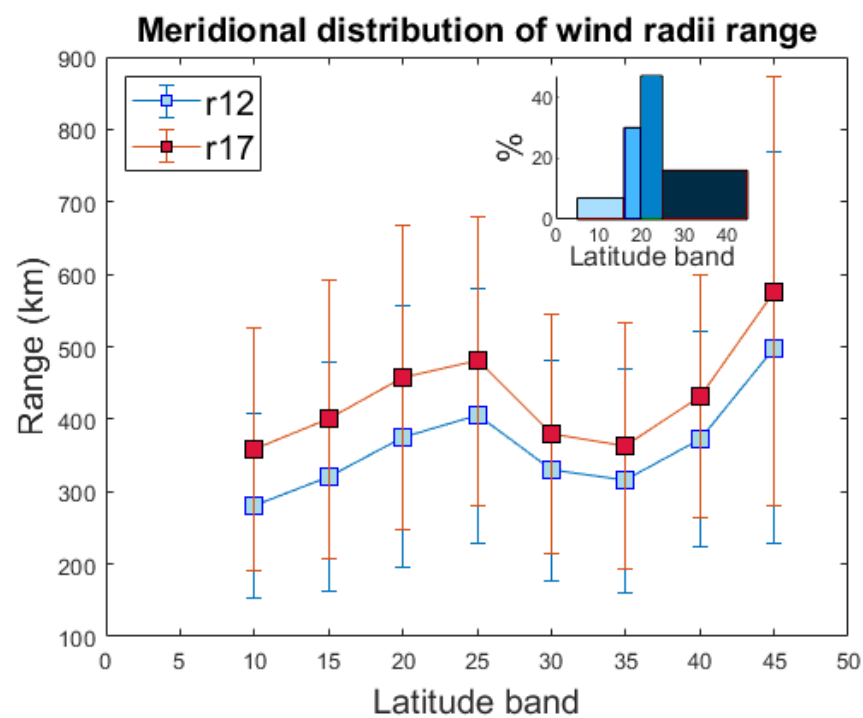


Figure 2.5: *Distribution of the TC size according to latitude.*



# Synthetic Aperture Radar: a magnifier for the inner-core.

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## 3.1 Preamble

Low- and medium-resolution MW satellite observations provides a comprehensive view of wind profiles at the periphery of the deep convection area, for most cyclones that occurred during the 2010 to 2018 seasons. The amalgamation of multi-frequency and L-band radiometric instruments, with C-band scatterometers, has allowed us a significant temporal follow-up of the characteristic wind radii, which compensates for the individual temporal intermittency, specific to each sensor. However, a paradox emerges from this analysis. If radiometers and scatterometers allow to approach the internal structure of the most intense winds, none of them allows a real description of its architecture. The finer resolution of active sensors generally offers a glimpse of the structure around the eye, yet the saturation of the signal prevents the evaluation of extreme winds (see Fig C.1 in Annex C), and thus an accurate estimate of  $R_{max}$  (as illustrated in the sketch 1.40). As for radiometers, despite an increasing emissivity with the pronounced foam activity under an intense forcing, they are handicapped by a resolution too coarse to effectively investigate the inner core, except for a large cyclone. The loss of coherence between these two types of measurements (beyond  $25\text{-}30 \text{ m.s}^{-1}$ ) makes any attempt at a combined approach difficult, or even obsolete for smaller systems. Irretrievably, the knowledge of the surface wind structure of the inner core is devolved to the emissivity measurements of the upper and middle atmosphere ingested in the BTK, thanks to MW sounders or IR-based techniques. In addition to the very vicarious nature of these methods (as discussed in 1.2.3), convective clouds on the inner edge of the eyewall follow a slight vertical slope [Shea and Gray 1973; Jorgensen 1984b], but steep enough under crescent shear to deflect estimates of the inner core metric between the cloud top and base. When it comes to determining  $R_{max}$ , the associated mean errors of individual techniques can be as severe as a dozen to more than 20 kilometers [Mueller et al. 2006; Kossin et al. 2007], whereas those finally provided in the 6-hourly interpolated Best-Track data remain unknown. Whether these measurement errors are the result of poor resolution [Klotz and Jiang 2017], drastic cyclone evolution, or abrupt vertical displacement, it is therefore hazardous to estimate  $R_{max}$  in the absence of valuable, albeit spatially localized, information from the SFMR data.

The emerging capabilities of SAR measurements seem crucial in a satellite landscape

lacking in high-resolution MW information of the inner core. Despite a decades-old instrumental tradition, tropical cyclone wind field retrieval using SAR imagery remains a new paradigm. If first techniques looked at combination of co-polarization signal with parametric model, to handle saturation issue ( $\sim 25\text{m.s}^{-1}$ ) [Reppucci et al. 2010], or the SAR-derived dominant swell spectrum [Hwang 2016], second efforts focused on cross-polarization due to its straightforward relation with wind speed and absence of saturation for high winds [Hwang et al. 2010; Vachon and Wolfe 2011]. Recent studies have even pointed out the need of using combined polarization, as cross-polarization face some signal-to-noise ratio issues under weak wind regime and some dependencies to wind direction and incidence angle [Zhang and Perrie 2012; Horstmann et al. 2015] for low to medium wind speed conditions. Thus, new generation of wind retrieval arise from either cross-polarized NRCS only or from merged capability of both polarization. Yet, most of the dedicated TC studies are rather on short number of cases, or barely up to  $40\text{ m.s}^{-1}$ . To answer the latter point, Mouche et al. [2019] provides a new GMF from combined co & cross polarization, to fully solve TC wind field. Looking qualitatively at the specific case of Irma, his wind retrieval gives consistent wind estimates with SFMR flights, intensity of  $75\text{ m.s}^{-1}$  was even observed. This outcome introduces the possibility of quantitatively probe TC parameters from SAR measurements, with this time a large dataset.

## 3.2 Analysis of TC wind field from SAR (Article)

NOVEMBER 2020

COMBOT ET AL.

4545

## Extensive High-Resolution Synthetic Aperture Radar (SAR) Data Analysis of Tropical Cyclones: Comparisons with SFMR Flights and Best Track

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**ABSTRACT:** To produce more precise descriptions of air–sea exchanges under tropical cyclones (TCs), spaceborne synthetic aperture radar (SAR) instruments provide unique capabilities to probe the ocean surface conditions, at very high spatial resolution, and on synoptic scales. Using highly resolved (3 km) wind fields, an extensive database is constructed from *RadarSat-2* and *Sentinel-1* SAR acquisitions. Spanning 161 tropical cyclones, the database covers all TC intensity categories that have occurred in 5 different TC basins, and include 29 cases coincident with SFMR measurements. After locating the TC center, a specific methodology is applied to filter out areas contaminated by heavy precipitation to help extract, for each acquisition, the maximum wind speed ( $V_{\max}$ ), its associated radius ( $R_{\max}$ ), and corresponding outer wind radii ( $R_{34/50/64}$  kt). These parameters are then systematically compared with best track (BTK), and when available, SFMR airborne measurements. For collocated SFMR and SAR observations, comparisons yield root-mean-squares of  $3.86 \text{ m s}^{-1}$  and 3 km for ocean surface wind speeds and TC  $R_{\max}$ , respectively. High correlations remain for category-5 cases, with  $V_{\max}$  exceeding  $60 \text{ m s}^{-1}$ . The largest discrepancies are found between BTK and SAR  $R_{\max}$  estimates, with  $R_{\max}$  fluctuations poorly captured by BTK, especially for rapidly evolving category-3, -4, and -5 TCs. In heavy precipitation ( $>35 \text{ mm h}^{-1}$ ), the SAR C-band measurements may be impacted, with local ambiguities associated with rain features, as revealed by external rain measurements. Still, this large dataset demonstrates that SAR measurements have unique high-resolution capabilities, capturing the inner- and outer-core radial structure of the TC vortex, and provide independent and complementary measurements than those used for BTK estimates.

**KEYWORDS:** Hurricanes/typhoons; Tropical cyclones; Wind; Algorithms; Remote sensing; Satellite observations

### 1. Introduction

Tropical cyclones (TCs) are better observed than ever before with improvements in coverage and capabilities from both remotely sensed and in situ observations. However, because of the heavy reliance of satellite-based observations in the oceanic regions where TCs form and track, creating precise descriptions of the TC surface wind field remains challenging as the necessary observations are often lacking (Knaff et al. 2016). In fact, estimates of most TC parameters including maximum sustained winds ( $V_{\max}$ ) (Velden et al. 2006; Demuth et al. 2004) and wind radii—the radial extent of 34-, 50-, and 64-kt ( $1 \text{ kt} \approx 0.51 \text{ m s}^{-1}$ ) winds (Demuth et al. 2006; Kossin et al. 2007; Knaff et al. 2011, 2016) are heavily weighted toward indirect satellite-based methods. Nonetheless, historical records or best tracks (BTK) containing TC location and intensity, and in some cases wind radii, have been compiled using such methods (Knapp et al. 2010, 2018; Landsea and Franklin 2013).

The radius of maximum wind speed ( $R_{\max}$ ) is a critical parameter for a number of applications. It helps to locate the maximum momentum injected in the ocean when strong mixing occur (Vincent et al. 2012; Price 1981).  $R_{\max}$  is also an essential scaling factor for the barotropic and baroclinic ocean responses (Geisler 1970; Ginis 2002; Kudryavtsev et al. 2019a), wind model parameterization (Holland 1980; Willoughby and Rahn 2004; Wood et al. 2013; Chavas et al. 2015) and operational

applications/models (Knaff et al. 2007, 2016), as well as to anticipate trapped-wave effects (Young et al. 2013; Kudryavtsev et al. 2015). Integrated kinetic energy (IKE) and wind power index (WPI) also rely on TC size information. IKE and WPI estimates were reported to outperform the sole use of  $V_{\max}$  estimates, to assess damages (Powell and Reinhold 2007), sea surface temperature TC wake signatures (Vincent et al. 2012), and poststorm ocean heat content capacity response (Knaff et al. 2013).

However,  $R_{\max}$  is often very difficult to estimate in absence of a well-defined eye feature in satellite imagery (Kossin et al. 2007; Lajoie and Walsh 2008) or aircraft reconnaissance due to the small spatial scales and strong wind gradients associated with  $R_{\max}$ , and the shortcomings associated with the various satellite methods (Lajoie and Walsh 2008; Knaff et al. 2011). The difficulty in estimating  $R_{\max}$ , likely leads to the reason why  $R_{\max}$  is not best tracked. Wind radii, on the other hand, are routinely used by operational centers for wave and storm surge forecasts (Sampson et al. 2010; NHC 2016), as well as the wind speed probability forecasts (DeMaria et al. 2013). In fact, since 2004, wind radii have been systematically reanalyzed for the best track at the National Hurricane Center (NHC) (Landsea and Franklin 2013). Similarly, the Joint Typhoon Warning Center (JTWC) has best track wind radii available since 2013 and in the final best tracks since 2016 (Sampson et al. 2018).

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To help in analyzing wind radii, operational scatterometers, radiometers, microwave sounders, and IR-based techniques are utilized (Sampson et al. 2017). A new generation of spaceborne instruments has recently emerged, the wide-swath L-band passive microwave sensors SMAP and SMOS (Reul et al. 2016; Meissner et al. 2017), able to estimate ocean surface wind speeds exceeding  $40\text{--}50\text{ m s}^{-1}$ , and also the CYGNSS constellation (Ruf et al. 2016; Morris and Ruf 2017) now assisting in that effort. Yet, while SMAP and SMOS have large swaths coverage that is ideal for estimating TC size and wind radii (Reul et al. 2017), these sensors have low spatial resolutions ( $40\text{--}50\text{ km}$ ). This precludes precise inner-core TC descriptions. To date, only limited airborne Stepped Frequency Microwave Radiometer (SFMR) measurements provide means to probe the high wind area of hurricanes (Uhlhorn et al. 2007; Klotz and Uhlhorn 2014). In this context, numerous studies already demonstrated the potential to use satellite synthetic aperture radar (SAR) measurements to provide TC surface wind field at very high resolution (Katsaros et al. 2000; Li et al. 2013). This unique mapping capability, further supported by highly sensitive cross-polarization measurements to increasing winds, has often been highlighted with respect to SFMR (Zhang and Perrie 2012; Horstmann et al. 2015; Mouche et al. 2019), buoys (Vachon and Wolfe 2011; Zhang et al. 2012), L-band passive remote sensing measurements (Mouche et al. 2017; Zhao et al. 2018), and global models (Fang et al. 2018). Still, most of these TC studies were not based on a large number of cases, especially those cases reaching winds higher than  $40\text{ m s}^{-1}$ . Using collocated SFMR measurements, Mouche et al. (2019) demonstrated how combined co- and cross-polarization C-band SAR measurements, efficiently resolve TC inner-core wind field characteristics. In that study, SAR measurements sampling Hurricane Irma (2017) when it had category (CAT)-5 intensity were compared to SFMR measurements and yielded comparable ocean surface wind speeds, with bias and root-mean-square of about  $1.5$  and  $5.0\text{ m s}^{-1}$ , respectively. Retrieved wind structure parameters outside the high wind inner core were also reported to be in agreement with NHC's best track and combined satellite- and aircraft-based analyses. The Irma study showed that within the TC inner core, SAR measurements alone can provide instantaneous and independent measurements of  $V_{\max}$  and  $R_{\max}$ , even in high wind speed gradients ( $6.5\text{ m s}^{-1}\text{ km}^{-1}$ ).

It is worth noting that contrary to most of the low-orbit Earth observation satellite missions, SAR instruments cannot continuously acquire wide swath data in high-bit rate modes. Anticipating and tasking SAR acquisitions with respect to the hurricane tracks forecast are thus required. Such tasking has been demonstrated over the past several years. In fact, since 2016, Sentinel-1 acquisition campaigns have thus been specifically designed, to test the instrument capabilities for mapping TCs (see Mouche et al. (2019) for details). A similar TC data acquisition program using *RadarSat-2* was also conducted by the Canadian Space Agency (CSA) (Banal et al. 2007). Today, the combined efforts of these campaigns have provided many TC cases—strongly maximizing the utility of SAR acquisitions from both Copernicus/ESA Sentinel-1 and MDA/CSA *RadarSat-2* missions.

In the present study, our main motivation is to describe how the SAR-derived wind field can be used to extract important TC parameters and evaluate their consistency with respect to best track and SFMR airborne measurements. Specifically, the potential of SAR high-resolution observations for estimating  $R_{\max}$  is discussed. 161 SAR acquisitions have thus been collected resulting in the first extensive SAR-TC database. They sample all TC intensity categories occurring in five different TC basins, and include 29 collocations with SFMR. After precisely locating TC center and areas contaminated by heavy precipitations,  $V_{\max}$ ,  $R_{\max}$ , and outer wind radii ( $R_{34/50/64}\text{ kt}$ ) are extracted from each acquisition. The data and methodology used are in section 2. TC structure parameters are then compared with best track from JTWC and NHC and SFMR estimates in section 3. Section 4, provides insights and in-depth analysis about limitations, e.g., rain impacts on C-band SAR measurements. Finally, section 5 summarizes the main outcomes of the present analysis, and provides prospects for future investigations.

## 2. Data and method for SAR analysis

### a. Synthetic aperture radar

This study benefits from three different C-band synthetic aperture radar (SAR) missions including *Sentinel-1A* (*S1A*), *Sentinel-1B* (*S1B*), and *RadarSat-2*. *S1A* and *S1B* are polar-orbiting satellites operated by Copernicus/European Space Agency and were launched in 2014 and 2016, respectively. *RadarSat-2* is a polar-orbiting satellite operated by the Canadian Space Agency (CSA) and was launched in 2007. All current SAR missions have several exclusive acquisition modes and cannot continuously acquire data, but rather acquisitions are scheduled. A dedicated acquisition strategy is thus mandatory to maximize tropical cyclone (TC) observations. The hurricane watch program (Banal et al. 2007) for *RadarSat*, and more recently, the Satellite Hurricane Observation Campaign (SHOC) (Mouche et al. 2019) for *Sentinel-1* and *RadarSat-2* were dedicated efforts that have maximized the collection of SAR measurements over TCs.

*S1A*, *S1B*, and *RadarSat-2* can be operated in wide swath modes and can acquire C-band backscatter measurements with different polarization states. Polarized radar images of the same scene can thus be combined for geophysical parameters retrieval. For ocean surface wind measurements over TCs, one generally uses the polarization configuration  $VV + VH$ , the antenna emitting electromagnetic waves in  $V$  polarization and receiving in both  $V$  and  $H$  polarization states. Here, the database solely builds on SAR observations in this polarization configuration acquired in wide swath modes. The swath widths range from 250 to 500 km depending on the sensor and/or modes. In comparison to other polar orbiting sensors such as radiometers or scatterometers, SAR swaths are smaller. However, SAR product resolution is much higher than other active or passive polar orbiting systems. Native product (i.e., Level 1 product from the Space Agency before applying the wind retrieval algorithm) resolutions used here to estimate the ocean surface wind speed are 20, 50 or 100 m depending on sensor and/or modes.

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The ocean surface wind retrieval algorithm follows the initial approach proposed by [Mouche et al. \(2017\)](#) and further refined for major hurricanes ([Mouche et al. 2019](#)). The method relies on the joint use of both co- and cross-polarized signals, befitting from both the copolarized signal's sensitivity and high signal-to-noise ratio for low to moderate wind speeds (i.e., below  $25 \text{ m s}^{-1}$ ), and from the cross-polarized signal sensitivity to very high wind speeds ([Zhang and Perrie 2012](#)). Compared to actual scatterometer measurements, the use of cross-polarized signals is key to mitigate the wind speed sensitivity issues often reported for copolarization backscatter signals ([Quilfen et al. 1998](#)).

A representative example of this co- and cross-polarization capability was provided in [Mouche et al. \(2019\)](#) where *SIA/B* measurements led to wind speed estimates ranging from 10 to  $75 \text{ m s}^{-1}$  in Hurricane Irma (2017) on 7 September when the storm was estimated to have category-5 intensities. Irma's wind speed estimates were compared to collocated SFMR estimates and the overall bias, root mean squared error (RMSE), and correlation were  $1.5 \text{ m s}^{-1}$ ,  $5.0 \text{ m s}^{-1}$  and more than 90%, respectively. It is, however, important to note that the C-band backscatter measurements can have contributions from ocean surface winds and rainfall ([Katsaros et al. 2000](#); [Li et al. 2013](#)), possibly leading to significant localized contamination in the wind speed estimates. In Hurricane Irma, the respective contribution of both wind and rain in areas with rain rates exceeding  $45 \text{ mm h}^{-1}$  and close to the TC center (30–35 km) was unclear. Rainfall was found to lead to about a 10% uncertainty in wind estimates ([Mouche et al. 2019](#)). Local gradient analysis of the radar backscattered signals ([Koch 2004](#)) is systematically performed in the present analysis to help identify localized regions where both rain and wind contribute to the C-band signals.

[Figure 1](#) shows an example of SAR wind speeds acquired in Hurricane Irma on 8 September and [Fig. 2a](#) shows the coincident cross-polarization backscatter. The retrieval resolution, in this case, is 3 km and the backscattered signals range from  $-35$  to  $-17 \text{ dB}$ . The backscatter has a clear minimum within the hurricane's eye. Backscatter then increases outward this minimum area with rapid signal increases, corresponding to the radius of maximum winds. Specific features related to rain events can be traced in the backscatter signal. In the north-western part, a bright pattern with a semicircular shape is detected and corresponds to an area of significant rainfall. [Mouche et al. \(2019\)](#) showed that rain impacts can also be associated with a small darker circular ring encircling an area just outside the largest backscatter signals and winds near the eye region (noticeable in [section 4b](#) example). Such a sudden signal decrease corresponds to very localized and heavy precipitation—fitting the general eyewall structure found in Hurricane Allen (1980) discussed in [Jorgensen \(1984\)](#). SAR thus provides instantaneous measurements of radar backscatter signals from very localized ocean surface areas. Here, a 3-km-resolution grid is adopted, which corresponds to the area that would be affected by a 1 min sustained  $50 \text{ m s}^{-1}$  wind speed.

### b. Best track and SFMR

TC best tracks that are prepared by different Regional Specialized Meteorological Center (RSMCs) and Tropical

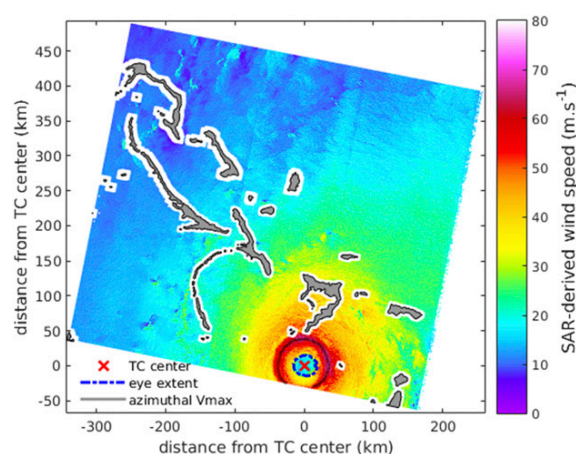


FIG. 1. Illustration of our methodology approach with the example of the SAR-derived wind field of category-5 Irma on 8 Sep 2017. The grid displayed is a Cartesian projection of 1 km and  $0.5^\circ$ -resolution polar grid, with TC eye center as origin. Dashed blue and black solid contours delineate, respectively, the eye extent and the azimuthal Vmax ring.

Cyclone Warning Center (TCWCs) depending on their areas of responsibility, provide 6-hourly estimates of location, intensity and other parameters covering each TC's life cycle. Initiated from near-real-time observations, these analyses are revisited after the TC seasons to take benefit of all available measurements, from surface information such as buoys, weather radars, platforms, up to aircraft, dropsondes and satellite remote sensors ([Knapp et al. 2010](#)), and are a subjectively smoothed representation of a tropical cyclone's history ([Landsea and Franklin 2013](#)). The analyses are performed by each RSMC, according to the data availability, not the same for each area, e.g., geostationary observations, their own strategy and conventions, e.g., time averaging periods. The gathering of all these information into a global and homogeneous database is thus not straightforward ([Kruk et al. 2010](#)), and is part of the efforts carried out to build the IBTrACS database ([Knapp et al. 2010](#)). A TC analysis starts by determining the TC location and the maximum sustained wind speed (Vmax), before including other parameters, such as wind radii or radius of maximum wind speed (Rmax), to refine the wind structure characterization. Specifically, three different wind radii are defined for each of the four geographical quadrant (NE, SE, SW and NW). These parameters provide the estimates of the maximum extent with wind speed greater than 34 (R34), 50 (R50), and 64 (R64) knots. Note, all RSMCs and TCWCs do not distribute these parameters.

In the following, we only focus on analyses from the NHC and the JTWC centers. At present, NHC and the JTWC both reanalyze R34, R50, and R64 following the season, but do not reanalyze Rmax ([Knaff et al. 2016](#)). Indeed, both centers rely on the Automated Tropical Cyclone Forecasting (ATCF) system, designed to ease and harmonize forecast processing and track record ([Miller et al. 1990](#); [Sampson and Schrader 2000](#)), with estimates of Rmax and wind radii, and the same convention for time averaging. In addition, both centers create



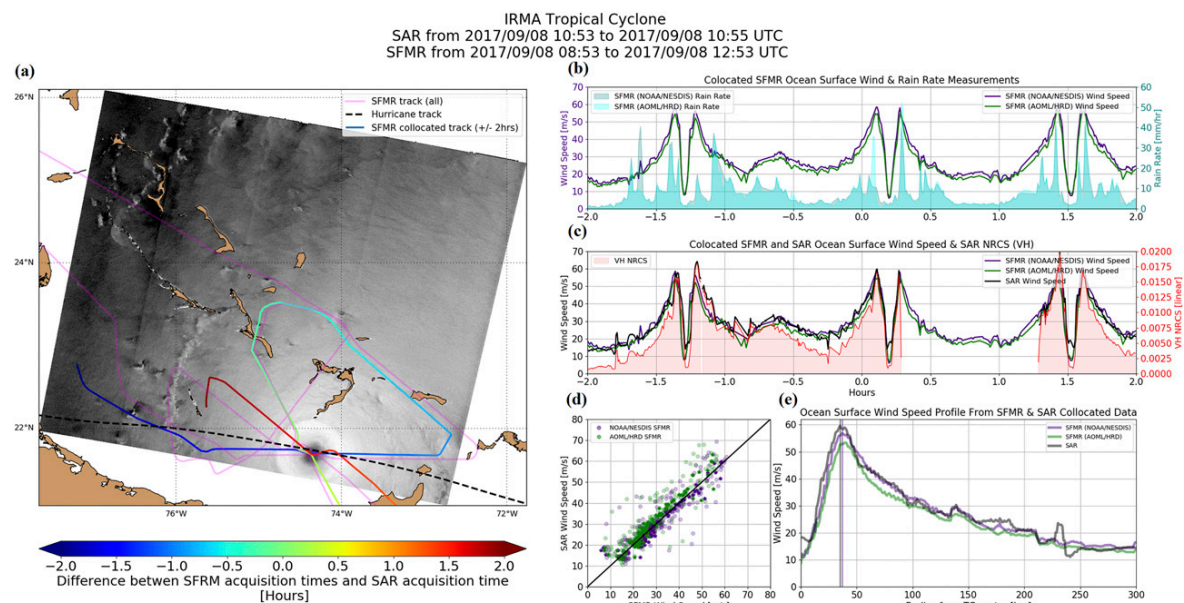


FIG. 2. Example of collocation and coanalysis between SAR and SFMR measurements in the case of Irma on 8 Sep 2017 (a) SAR backscattered signal in cross-polarization from *RadarSat-2* SAR from 1053 to 1055 UTC. (solid pink line) SFMR measurements location during hurricane hunters flight from 0853 to 1253 UTC. (colored line) Collocated SFMR measurements location within a time window of  $\pm 2$  h centered on SAR acquisition start time. Changes in color indicate the time difference between SAR and SFMR measurements. (b) Collocated SFMR ocean surface wind speed (purple and green) and rain rate (blue and cyan) measurements with respect to time difference between SAR and SFMR measurements. (c) Ocean surface wind speed measurements from SAR (black) and SFMR (purple and green) and SAR backscattered signal (red) in cross-polarization evolution with respect to time difference between the two sensors. (d) Direct comparison between SAR and SFMR wind speed (green and purple). (e) Ocean surface wind speed profile with respect to distance from TC center as measured from collocated SFMR (green and purple) and SAR (black) measurements within a  $\pm 2$  h time window. Vertical bars indicate the maximum of wind as given by the two sensors.

best track analyses for all storms, offering a homogeneous dataset for comparisons with SAR derived parameters over all five different ocean basins. In absence of IBTrACS data, ATCF archive (only 2018 cases) is solely used. Overall, the different strategies and methods to estimate TC parameters remain similar. Vmax is mainly inferred by the Dvorak analysis in combination to cloud pattern recognition from visible and infrared (IR) satellites to TC intensity (Velden et al. 2006, 2017), but at times can also consider other methods and aircraft reconnaissance, when available. R34, R50, and R64 wind radii are derived from scatterometers, cloud/feature-tracked winds, new L-band passive radiometer measurements (Reul et al. 2016; Meissner et al. 2017) and other operational techniques (Knaff et al. 2011, 2015), including IR (Kossin et al. 2007) or microwave sounder (Demuth et al. 2004, 2006). At last and despite its aforementioned importance, Rmax is generally subjectively estimated, except when airborne data from SFMR or flight level are available.

If other methods exist (Lajoie and Walsh 2008), they are all indirect methods as they do not retrieve any ocean surface wind field before providing the wind radii estimates. To note, neither JTWC nor NHC have been using SAR data for their analyses. Therefore, in addition of being the most integrated and quality-controlled data source, best track analysis is an independent source of comparison.

Rain rates and ocean surface wind speeds from the SFMR are also used as independent measurements. Since its first experimental flight in 1980 through Hurricane Allen, SFMR is now installed on all U.S. hurricane reconnaissance aircraft to routinely and operationally provide wind and rain estimates during TC events. The concept relies on the use of a C-band radiometer operating at six different frequencies ranging from 4.5 to 7.2 GHz with different sensitivities to foam coverage at the sea surface (related to ocean surface wind speed) and to rain (Uhlhorn and Black 2003).

Over the course of time, retrieval algorithms have been further improved, as the possibilities to refine the filtering of the data and better describe the brightness temperature dependency to wind and rain increase with the number of available flights (Uhlhorn et al. 2007; Klotz and Uhlhorn 2014). Recently, Sapp et al. (2019) proposed new improvements including a bias correction to calibrate the whole dataset, a new model for flat-surface emissivity (Meissner and Wentz 2012), a new frequency dependence for the atmospheric transmissivities, leading to a new method and formulation to derive the relationship between the wind excess emissivity and the ocean surface wind speed (so-called wind GMF, for Geophysical Model Function). In particular, arguing on possible non-geophysical contamination of the lower-frequency channel, the highest-frequency channel has been considered to derive a new

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wind GMF (and its frequency dependency). This also imposed modification of the rain absorption model coefficient to maintain the initial rain rates performances (Klotz and Uhlhorn 2014). Overall, SFMR wind speeds are expected to be higher than operational products for wind speeds larger than  $15 \text{ m s}^{-1}$  and are judged as nonreliable for rain rates measurements larger than  $45 \text{ mm h}^{-1}$  (Sapp et al. 2019). This study does not aim to discuss the two existing SFMR products, and mainly relies on products processed and provided by NOAA/NESDIS (Sapp et al. 2019). For sake of completeness, results obtained with AOML/HRD products are also considered in the methodology section (see section 2c), and in the discussion section (see section 4).

Because the SFMR design involves a single nadir-viewing antenna, only transects are available. Despite these limitations, the combined estimates of rain rate and ocean surface wind speed at very high resolution (temporal resolution is 1 s) makes this instrument unique for validating SAR-derived wind speeds and to discuss the possible rain impacts, especially in the TC inner core.

### c. Methodology

While the subjectively smoothed best track content already results from a multisource reanalysis, SFMR and SAR provide more localized measurements. Although SAR and SFMR can both be used to derive local ocean surface wind speeds, their differences in coverage and temporal resolution need to be considered in order to build meaningful comparisons. A specific methodology to coanalyze SAR measurements with each of these sources is needed.

To prepare for the evaluation of SAR's ability to estimate  $V_{\max}$ ,  $R_{\max}$ , and wind radii parameters, a three-step approach is proposed to extract those same parameters from the SAR wind products: 1) find the TC center, 2) identify the azimuthal  $V_{\max}$  ring and  $R_{\max}$ , 3) compute mean wind radii values. Irma's SAR-derived wind field on Fig. 1 illustrates the different processing steps:

- **TC center:** The TC position is first linearly interpolated using the best track at satellite-acquisition time. In the vicinity of this position (100 km), the signal contrast is computed for the two polarized images. Selecting the one with the strongest gradient, we search for the location of the signal intensity minimum. These locations are then averaged to get a second TC center guess, used to remap the SAR wind on a polar grid. The retained polarization channel and the wind speed map are further jointly coanalyzed with the heterogeneity mask to estimate the maximum gradient in all azimuth directions and to derive the eye extent (see dashed blue line on Fig. 1), from which the mean center is computed to obtain the TC center (see red cross on Fig. 1). A new polar grid centered on this TC center is then defined, resolution 1 km in distance and  $0.5^\circ$  in direction.
- **$V_{\max}$  and  $R_{\max}$ :** The first  $R_{\max}$  estimate is defined as the closest peak to the TC center obtained from the azimuthally averaged 1D radial wind speed profile. Based on this first guess, we further derive  $R_{\max}$  for each available azimuth angle of the polar grid (black line on Fig. 1). SAR-derived

$V_{\max}$  is defined as the 99th percentile of the wind speed values associated to the azimuthal  $R_{\max}$  values. SAR-derived  $R_{\max}$  is then simply the radius corresponding to this percentile.

- **Wind radii:** The 34-, 50- and 64-kt wind radii are estimated for the four geographical (NE, NW, SW, SE) quadrants. For each quadrant we search for the closest radial values above the wind speed threshold (34, 50, or 64 kt) in all of the quadrant's azimuth directions. We then keep the 10% of the largest values associated to the corresponding wind speed, to provide a SAR-derived wind radii. Estimates are considered optimal when quadrants are at least 50% complete. Under this threshold, they are not calculated. The use of the 90th percentile for deriving the maximum extent of the wind radii from high-resolution satellite wind measurements allows to filter unrealistic outliers.

Finally, for SAR-BTK comparisons, we linearly interpolate best track at satellite acquisition time for each TC parameters as instantaneous SAR measurements rarely match synoptic times.

Here we examine SAR's capabilities for measuring ocean surface wind speed near the inner core of TCs, including  $V_{\max}$  and  $R_{\max}$ , by using nearly coincident SFMR wind speed measurements. The two main differences between SAR and SFMR observations are (i) the duration required to sample a given TC and (ii) the coverage of the TC structure. For the Irma TC case presented in Fig. 2, SAR data are acquired in less than 3 min, while the SFMR data collection lasts more than 9 h. While the "multi-alpha" reconnaissance pattern (the solid purple line) is designed to sample the 2D aspects of the TC, it can only do so in a low-spatial-resolution and temporally averaged manner. A two-step procedure has been adopted (see Mouche et al. 2019 for details) to best compare SFMR and SAR wind speeds. SFMR measurements are resampled at 3 km resolution, and the time differences are taken into account between each SFMR measurement and the SAR acquisition time following storm motion. As shown in Fig. 2a, this procedure shifts the initial SFMR measurements locations (purple solid line) with respect to SAR acquisition time into a collocated track (colored solid line). In addition, to further mitigate the time difference impact, only collocations within  $\pm 2 \text{ h}$  are considered. Figure 2b illustrates SFMR wind and rain measurements along the transect used for collocating in the Hurricane Irma, while Fig. 2c shows the corresponding radar backscattered signal in cross-polarization (red) and the wind speed measured by the SAR (black). As expected from previous studies (Zhang and Perrie 2012; Mouche et al. 2019), the correlation between radar signal and SFMR wind speed is very high, and the two sensors are able to capture the TC characteristics within the inner core. Finally, wind speeds from SAR and SFMR can be directly compared (see Fig. 2c). In particular, when the collocation time is less than 30 min and during the transect across the hurricane eye, we note the remarkable agreement for wind speeds ranging from  $15$  to  $60 \text{ m s}^{-1}$ . This example also illustrates the difference between wind speed measurements obtained when using NOAA/NESDIS or AOML/HRD products. Although the shape of the two wind

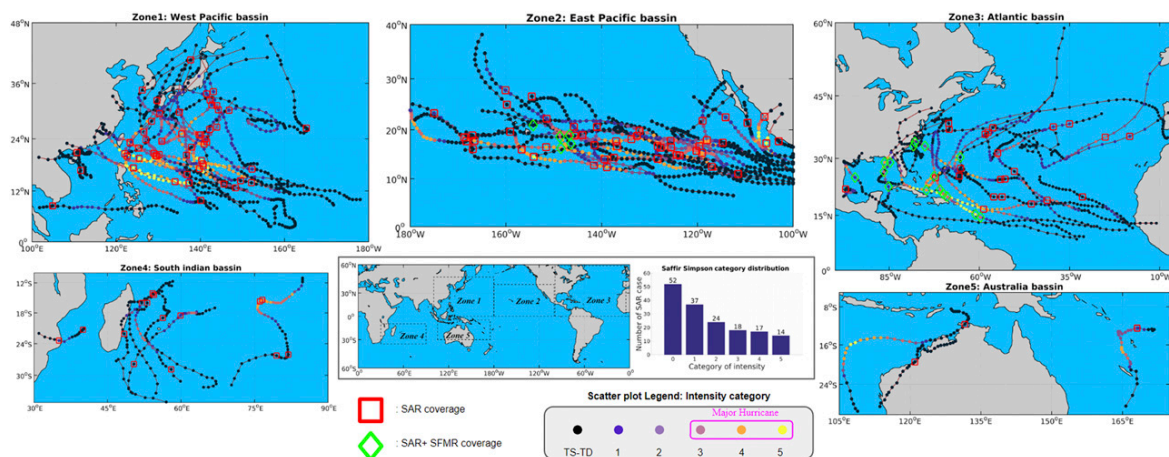


FIG. 3. Composite view of TC cases data constellation for each geographical zone; basin locations are indicated in the global map. Lifetime positions from best track are displayed for each TC; colors depict intensities with respect to Saffir–Simpson scale. Markers are stated for TC positions with measurements. Red squares indicate SAR measurements only and green diamonds indicate sequential measurements of SAR and SFMR.

speed transects from SFMR remain very similar, we can notice differences for the highest values of wind speeds. The impact on SAR comparison is further illustrated on Fig. 2d. However, SFMR measurements may not necessarily cross the area corresponding to the maximum wind speed. Consequently,  $R_{\max}$  parameter cannot always be derived from SFMR measurements. To overcome this limitation, we introduce the “effective  $R_{\max}$ ” parameter. It is defined as the radius of maximum wind speed obtained from all the SFMR transects collocated with SAR measurements. These transects are combined to derive a single averaged wind speed profile for each of the two sensors from which the maximum of wind speed and corresponding radius can be estimated. This is illustrated on Fig. 2e. The averaged wind profile derived from SFMR collocated transects is shown (solid purple line) with respect to the distance from TC center. Here the maximum wind speed and the corresponding effective  $R_{\max}$  are about  $60 \text{ m s}^{-1}$  and 37 km, respectively (see solid vertical purple bar). When applying the same method to the SAR collocated wind measurements, we obtain a very similar wind profile (see solid gray line), yielding an effective  $R_{\max}$  of 35 km that can be directly compared to SFMR estimate. Figure 2e also presents the two different wind speed profiles obtained with the two SFMR products available. As anticipated with the transect illustration, these products yield to some differences for the highest wind speed values, but not for the effective  $R_{\max}$ . This analysis has been applied to all available collocated SAR/SFMR data (see section 3).

#### d. Dataset overview

Thanks to the SHOC campaign, a total of 194 acquisitions have been obtained, enabling an unprecedented SAR TCs collection over all five distinct TC basins. So far, only the North Indian Ocean is missing, but acquisitions have been pursued in 2019. Best track analysis is available for all cases, with a total of

29 collocated SFMR flights. Yet, as discussed in section 2c, specific requirements such as the mandatory presence of a complete eye structure are imposed. Two situations prevent our analysis procedure: 1) when the swath border intersects the  $R_{\max}$  and 2) when land contaminates the retrievals ( $>75\%$  of the full scene). Both situations effectively prevent a full eye or eyewall description. In addition, cases without clear eye structures (~seven cases discussed in section 4) were removed. Overall, about 85% of initial acquisitions are preserved.

After this quality control step, 161 snapshots corresponding to 72 different tropical systems in the period 2015–18 can then be analyzed. Figure 3 synthesizes the dataset. For each storm, the 6-h best track locations with corresponding storm intensity (colors) is indicated. Specific markers highlight the collocation opportunities: A red square when only-SAR is available and a green diamond when simultaneous SAR+SFMR measurements coexist. Because aircraft measurements are restricted to North Atlantic and eastern North Pacific basins, with a majority occurring in the Atlantic, collocations with SFMR amount to only 13% of the dataset, with a total of 23 Atlantic and 6 Pacific flights. 70% of Atlantic hurricanes cases are actually retained. The intensity histogram shown in the bottom panel of Fig. 3 illustrates the spectrum of TC intensities. Unlike most of previous SAR-based studies, all Saffir–Simpson categories are sampled. Therefore, this dataset captures the general distribution and basins properties observed in climatology studies (Chan and Chan 2012; Knaff et al. 2014; Chavas et al. 2016), in terms of size and activity.

To complement this SAR dataset, we also collocate rainfall information from two different sources: the half-hourly Integrated Multisatellite Retrievals for GPM (IMERG) product, with  $0.1^\circ$  resolution and global coverage (Huffman et al. 2019), and specific NOAA Next Generation Weather Radar (NEXRAD) network samples, with an 1 km range and  $1^\circ$ -azimuth high resolution for a 450 km coverage. IMERG is systematically used to identify cases



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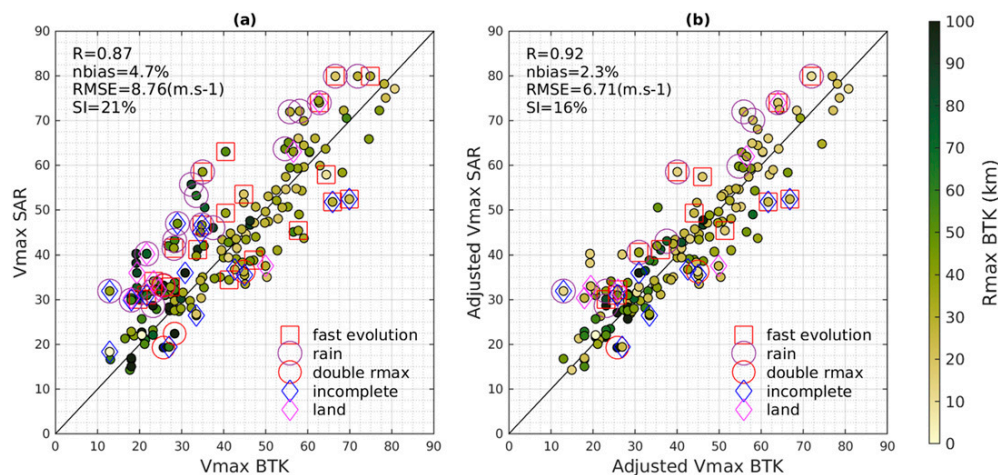


FIG. 4. Comparison of best track and SAR-derived Vmax. Disturbance sources are indicated with specific markers to guide analysis. Correlation ( $R$ ), normalized bias (nbias) calculated as  $(V_{\max_{\text{SAR}}} - V_{\max_{\text{BTK}}})/V_{\max_{\text{BTK}}}$ , root-mean-square error (RMSE), and scattering index (SI) are added as statistical tools. Comparison (a) before and (b) after treatments of flagged values.

with heavy rain situations ( $>35 \text{ mm h}^{-1}$ ), around the inner-core region. Regarding NEXRAD, specific cases have been selected to help discussing precipitation impacts on C-band sensors (Katsaros et al. 2000; Mouche et al. 2019).

### 3. Comparison of TC parameters

#### a. Ocean surface wind speed

We follow the three-step approach detailed in section 2c for the 161 quality-controlled SAR cases. Results obtained for Vmax, are shown in left panel of Fig. 4. Overall, there is a strong correlation ( $R = 0.87$ ) and low bias (4.7%) between best track and SAR Vmax estimates. A larger scatter is noticeable, mainly in the vicinity of  $10\text{--}40 \text{ m s}^{-1}$ . It leads to a RMSE around  $9 \text{ m s}^{-1}$ . We further define specific flags to isolate potential contamination sources in these comparisons, possibly resulting from uncompleted TC wind field due to land (pink diamond) or swath issues (blue diamond), TC with double eye Rmax (red circle) and rain signature (purple circle). It is also possible that Vmax comparisons are biased by the 6-hourly and smoothed nature of best track analysis, especially for cases with rapidly evolving intensities [ $dv/dt > 10 \text{ kt (6 h)}^{-1}$ ]. Specific data processing methods are thus considered, corresponding to three possible situations: 1) Fast evolving cases: as a TC intensity may not vary linearly, we simply choose the closest time instead of the interpolated one; 2) heavy rain ( $>30 \text{ mm h}^{-1}$ ): in presence of ambiguous peak, we select the second maximum wind speed in area where IMERG rain rate is lower than  $30 \text{ mm h}^{-1}$ ; 3) double Rmax: in case of an eyewall replacement cycle (ERC), we select the SAR Vmax associated with the closest ring to the best track Rmax estimate. These corrections define what we called the “adjusted” SAR or best track Vmax, to yield new comparisons presented in the right panel of Fig. 4. Statistics slightly improve, with higher correlation ( $R = 0.92$ ) and lower normalized bias

(2.3%). Some of the scatter is also related to uncertainties in the best track intensity estimates (Torn and Snyder 2012; Landsea and Franklin 2013).

The methodology discussed in section 2c, is applied to the collocated SFMR/SAR dataset presented in section 2d. Results and summary statistics are presented in Fig. 5. Overall (blue and red dots together), a high correlation ( $R > 0.90$ ), small bias ( $<0.5 \text{ m s}^{-1}$ ) and RMSE ( $<5 \text{ m s}^{-1}$ ) are obtained. An additional filtering to solely select collocated measurements obtained within  $\pm 30 \text{ min}$  can be applied (red dots). This improves the statistics—particularly for RMSE now at about  $4 \text{ m s}^{-1}$ . As discussed by Mouche et al. (2019), most outliers are associated with subtle errors in collocating points, which can adversely affect verification statistics performed at such high resolution. This sensitivity is particularly large in the eyewall region where a few kilometers of error in the collocation can lead to significant differences in wind speeds in a steady intensity state. To note, these results are obtained when SFMR measurements are considered as valid, i.e., excluding measurements with rain rates larger than  $45 \text{ mm h}^{-1}$  (Sapp et al. 2019). Using only the 29 SFMR cases clearly shows that SAR provides high quality ocean surface wind speed measurements up to  $70\text{--}75 \text{ m s}^{-1}$ .

#### b. Wind structure

Here we compare best track estimates of Rmax and wind radii to those calculated from SAR (see section 2c). Figure 6 presents all values obtained for the three wind radii (R34: green; R50: blue; R64: yellow), and includes all geographical quadrants. Correlation is high, larger than 0.85 for the three wind radii. The normalized bias is negligible (about  $-3\%$ ) for R34 and R50, but about 10% for R64. R64 from SAR are generally larger than values given in the best track. The scatter index increases with wind speed values associated to the radii, yielding to 29% for R64. As mentioned, best track parameters

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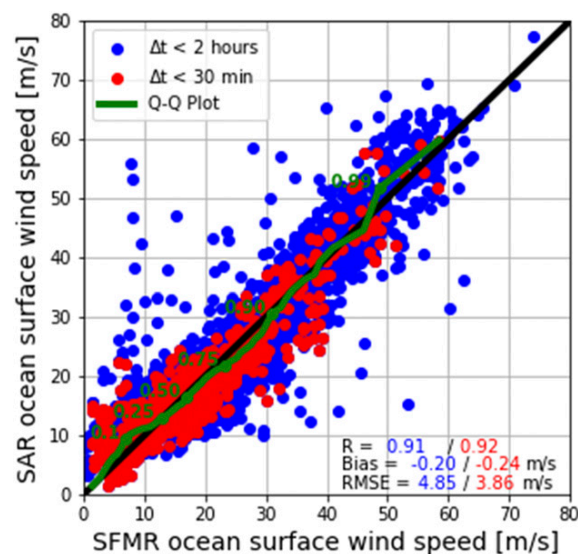


FIG. 5. SFMR-derived ocean surface wind speed vs SAR-derived ocean surface wind speed. Resolution is 3 km. Blue dots indicate the collocated data within  $\pm 2$  h and red dots within  $\pm 30$  min. Green line is the quantile–quantile (Q–Q) plot applied to collocated data within  $\pm 30$  min, some of the main quantiles are indicated for convenience (green).

are mostly derived from indirect methods (section 2b) or low-orbit satellite missions using medium to low-resolution (12.5–50 km) observations (Brennan et al. 2009; Meissner et al. 2017). This could explain the observed spread, especially, the increasing scatter index for R64 that typically has values less than 100 km. In addition, radii associated to higher winds may have higher temporal variability (Chavas and Lin 2016), possibly not captured by the 6-hourly best track analysis—especially if the estimate relies on sparse low-orbit satellite measurements. The number of wind radii available in the best track ( $N_{\text{total}}$  in legend of Fig. 6), from SAR data ( $N_{\text{SAR}}$  in the legend of Fig. 6), and that are used for the comparison ( $N_{\text{coloc}}$  in legend of Fig. 6) are also indicated for each wind radii on Fig. 6. Due to swath issues (e.g., the southern part in Fig. 1) preventing a complete TC structure description, SAR measurements are sometimes unable to estimate wind radii in all geographical quadrants and/or at all wind thresholds. Consequently, for R34 generally associated with the largest radius, only about 50% of the best track values can be compared, whereas this number rises to about 70% and 90% for R50 and R64, respectively. Note that the existence of the best track wind radii is dictated by the best track intensity (for instance R50 and R64 do not exist for tropical storms with  $V_{\text{max}} < 50$  kt). So the best track does not always have wind radii values to compare to SAR. This explains the difference between  $N_{\text{SAR}}$  and  $N_{\text{coloc}}$  parameters. In particular,  $N_{\text{SAR}} = 354$  and  $N_{\text{coloc}} = 286$  parameters for R64 indicates that SAR observations provide more R64 estimates than best track. This difference decreases for R50 and R34, suggesting resolution issues in R64 estimates.

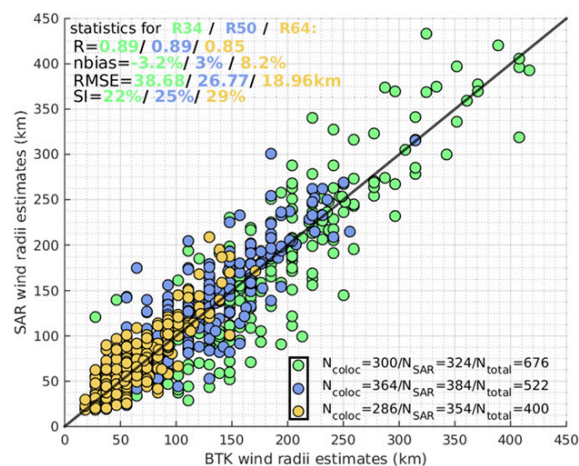


FIG. 6. Comparison of best track and SAR-derived maximum extent quadrant wind radii. Extents from SAR are inferred from the mean quadrant of the 10% largest values. All quadrants' wind radii estimates are included; colors of both statistics and chart are set accordingly: green indicates R34, blue indicates R50, and yellow indicates R64. Statistical tools used are as in Fig. 4.

Results of comparisons of  $R_{\text{max}}$  are shown in Fig. 7a. As for  $V_{\text{max}}$ , additional corrections can be applied (see Fig. 7b) to account for best track or SAR parameter uncertainties. Contrary to  $V_{\text{max}}$ , large scatter index ( $\sim 50\%$ ) and RMSE ( $\sim 25$  km) are found and persist despite additional corrections. The bias remains almost constant. Correlation increases (from 0.47 to 0.73), but it is mostly governed by double eyewall cases, which strongly impact the estimates. Below category-1 TC, tropical depressions and storms rarely exhibit a complete eye (Vigh et al. 2012). When discarding these cases, for which  $R_{\text{max}}$  and TC center definition can be quite subjective, a clear improvement is obtained, with a reduced RMSE (11.6 km) and scatter index (32%). Yet, the spread is still high and bias remains unchanged (SAR-derived  $R_{\text{max}}$  are globally smaller). A binning of the  $R_{\text{max}}$  values from best track is also noticeable in the vicinity of 20–40 km. Overall, this comparison reveals an overestimation of  $R_{\text{max}}$  parameters from the best track when compared to SAR estimates, for all TC intensities available in our study.

To compare the effective  $R_{\text{max}}$  from SAR and SFMR, as defined in section 2c, we only examine hurricane strength cases in the SAR–SFMR collocated dataset. Seventeen cases of the 29 available are ultimately used. Figure 8a presents comparisons and associated statistics of effective  $R_{\text{max}}$ . The color code indicates SFMR maximum wind speed for each case. Again, the agreement is very good with correlation coefficient larger than 0.70, RMSE of 12 km and bias lower than 5 km. SAR measurements provide  $R_{\text{max}}$  values from 10 to 70 km, with no significant dependency with respect to current intensity. The present analysis, however, yields two clear outliers, i.e., the two dots within the purple ellipse on Fig. 8a. They are both corresponding to data acquired over Hurricane Florence around 1100 UTC (SAR acquisition



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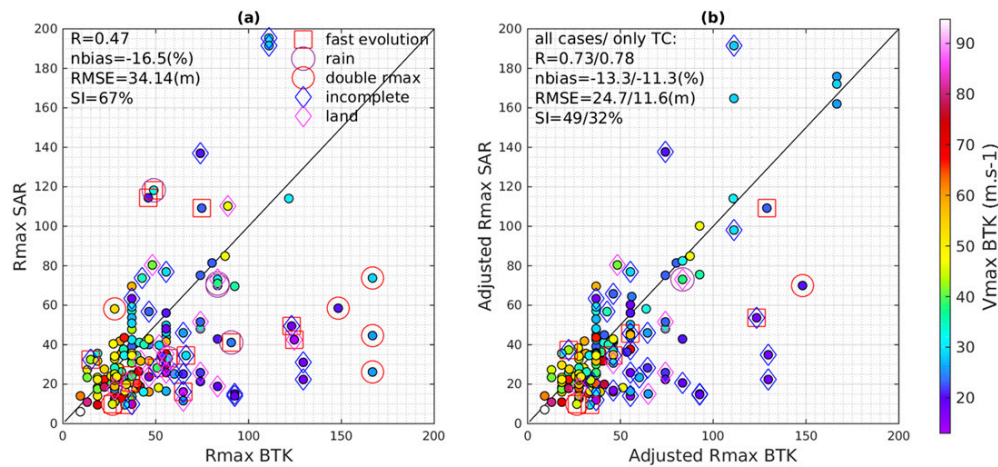


FIG. 7. Comparison of best track and SAR-derived Rmax. Markers and statistical tools are similar to the Fig. 4 convention. Comparison (a) before and (b) after treatments of flagged values. Statistics for solely cases  $>33 \text{ m s}^{-1}$  are also included.

time), 13 and 14 September 2018. Figures 8b and 8c shows SAR-SFMR collocated measurements and the corresponding SAR- and SFMR-derived wind profiles for the first outlier. Collocated transects exhibit a significant difference in the wind speed variation close to the eyewall (see within red circle in Fig. 8b), where SFMR measured a rain rate larger than  $30 \text{ mm h}^{-1}$  and wind speed of  $40 \text{ m s}^{-1}$ . This inconsistency between SAR and SFMR leads to a slightly different shape for the wind speed profile, as displayed Fig. 8c, and finally to a difference in the effective Rmax estimate. When the two outliers are filtered out, correlation coefficient increases, to become larger than 0.95, whereas RMSE and bias decrease to about 3.5 km and 500 m, respectively. Although a simple filtering of local maxima with respect to the TC center distance would have removed these outliers, we considered

them to further illustrate possible rain impact on C-band derived wind speed.

### c. Best track error estimates

Assuming SAR-derived parameters as reference values, we further analyze TC inner (Vmax and Rmax) and outer (wind radii) core parameters as given by best track with respect to storm category. Figure 9 displays the mean values from SAR and best track for each of these TC parameters as a function of storm category. The related uncertainty (top panels) and mean absolute error (MAE) (bottom panels) are indicated for both inner (left panels) and outer core parameters (right panels).

Most SAR- and best track-derived TC parameters (i.e., Vmax and wind radii) have similar trends in mean values and MAE with respect to intensity (with exception of TS/TD).

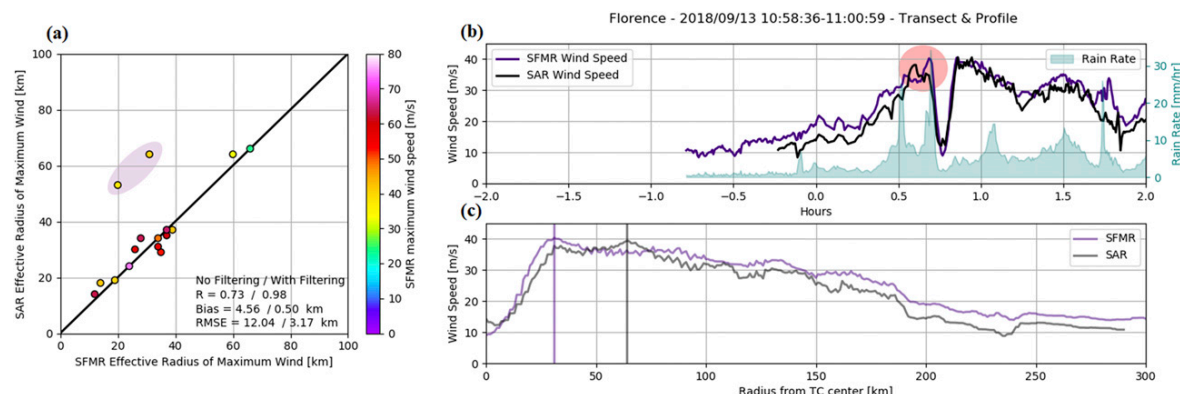


FIG. 8. SFMR- and SAR-derived effective Rmax. (a) SFMR-derived effective Rmax vs SAR-derived effective Rmax. Color code indicates SFMR maximum wind speed. Outliers are within the purple ellipse. (b),(c) Case study of Florence TC on 13 Sep 2018 corresponding to one of the two outliers. (b) Collocated SAR-SFMR measurements along the transect. (c) SFMR- and SAR-derived wind speed profiles and corresponding Rmax.

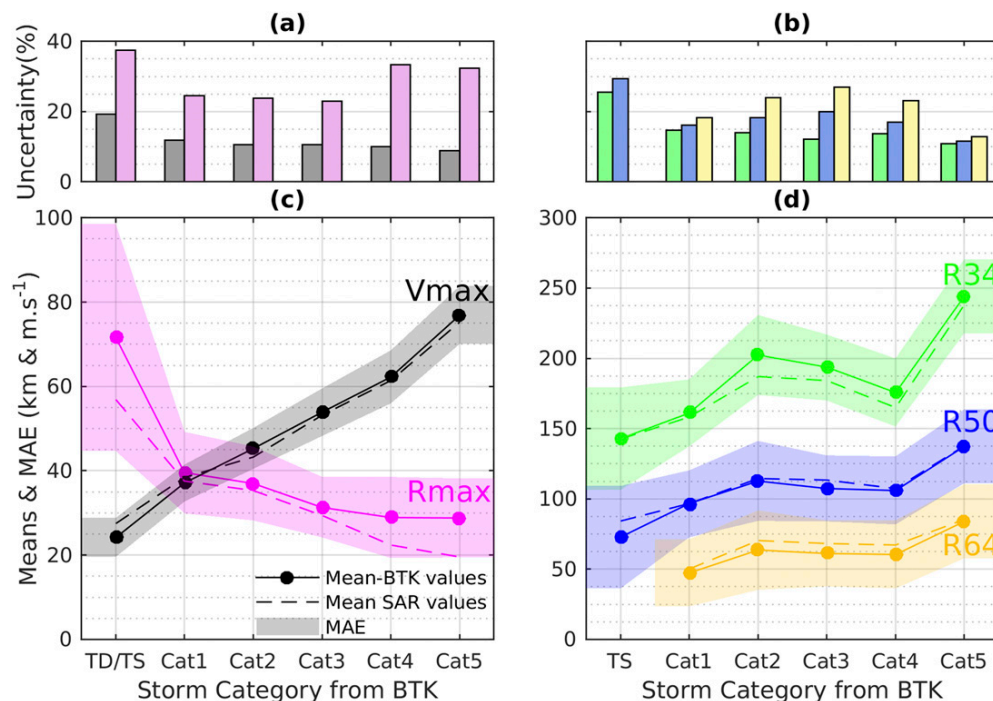


FIG. 9. Best track errors and uncertainties estimated from SAR-derived values for each TC parameter. (left) Inner core ( $V_{\max}$ : black;  $R_{\max}$ : magenta) and (right) wind radii ( $R_{34}$ : green;  $R_{50}$ : blue;  $R_{64}$ : yellow). (a),(b) Histograms of TC parameters uncertainties for each intensity category. Uncertainties are computed as  $\text{MAE}(\text{category})/\text{mean}(\text{category})$ . (c),(d) Mean category values are displayed for SAR (dashed colored lines) and best track (solid colored lines); shaded areas outline best track errors computed as mean absolute error (MAE).

Significant differences are found for  $R_{\max}$  mean values and MAE for TD/TS cases and CAT-4/5 TCs, with  $R_{\max}$  value uncertainties larger than 30%. This confirms the disagreement between SAR and best track  $R_{\max}$ , previously discussed.

For wind structure parameters describing the TC outer core, uncertainties are found to increase with the prescribed wind speed.  $R_{64}$  uncertainties are the largest (peak about 25%), followed by  $R_{50}$  and  $R_{34}$  uncertainties, regardless the TC intensity.  $R_{50}$  and  $R_{64}$  wind radii uncertainties seem to be storm category dependent, with larger discrepancies obtained for intermediate categories. In particular,  $R_{50}$  and  $R_{64}$  uncertainties peak for CAT-3 TC, with values reaching about 20% and 25%, respectively. The analysis of best track used in this study further reveals more variability for  $R_{50}$  and  $R_{64}$  from one time step to another for CAT-2 or CAT-3 TCs. This could explain the larger uncertainties when compared to SAR estimates.

Remarkably, all TC parameters suffer from large MAE and uncertainties for TD/TS cases, whereas only  $R_{\max}$  parameter uncertainties rise for major TCs. This deficiency linked to the complexity of weak systems is addressed in section 4. A significant difference in  $R_{\max}$  trends is also found, with a clear plateau for the highest category of best track analysis. It may be indicative of a specific issue regarding  $R_{\max}$  estimates for major TCs. As already mentioned, this apparent lack of

sensitivity for  $R_{\max}$  may be due to the low- to medium-resolution observations used to estimate these parameters as well as operational constraints and/or procedures. Our analysis shows differences up to 30 km when SAR-derived  $R_{\max}$  are about 20 km (see Fig. 6c). Such cases will be further discussed in section 4.

Our SAR-based results are consistent with previous studies: a range of 10%–40% were found by Knaff and Sampson (2015) and Sampson et al. (2017) for wind structure, and 10%–20% by Torn and Snyder (2012) and Landsea and Franklin (2013) for intensity. After averaging estimates over all categories, we end up with the following MAE and uncertainties values for TC parameters:  $V_{\max}$ : [4.8 m s<sup>-1</sup>, 9.5%];  $R_{\max}$ : [15 km, 32%];  $R_{64}$ : [14 km, 22%];  $R_{50}$ : [20 km, 19%]; and  $R_{34}$ : [29 km, 17%].

#### 4. Discussion

##### a. Best track–SFM–SAR comparison limitations

Although comparisons between best track- and SAR-derived parameters reveal an overall high consistency, both  $V_{\max}$  and wind radii parameters can display noticeable scatter, of about 20%. High-resolution SAR acquisitions using both co- and cross-polarizations are still recent. Current uncertainties governing the relationship between radar parameters and wind speed or rain, as well as possible calibration issues, will certainly improve with increasing systematic

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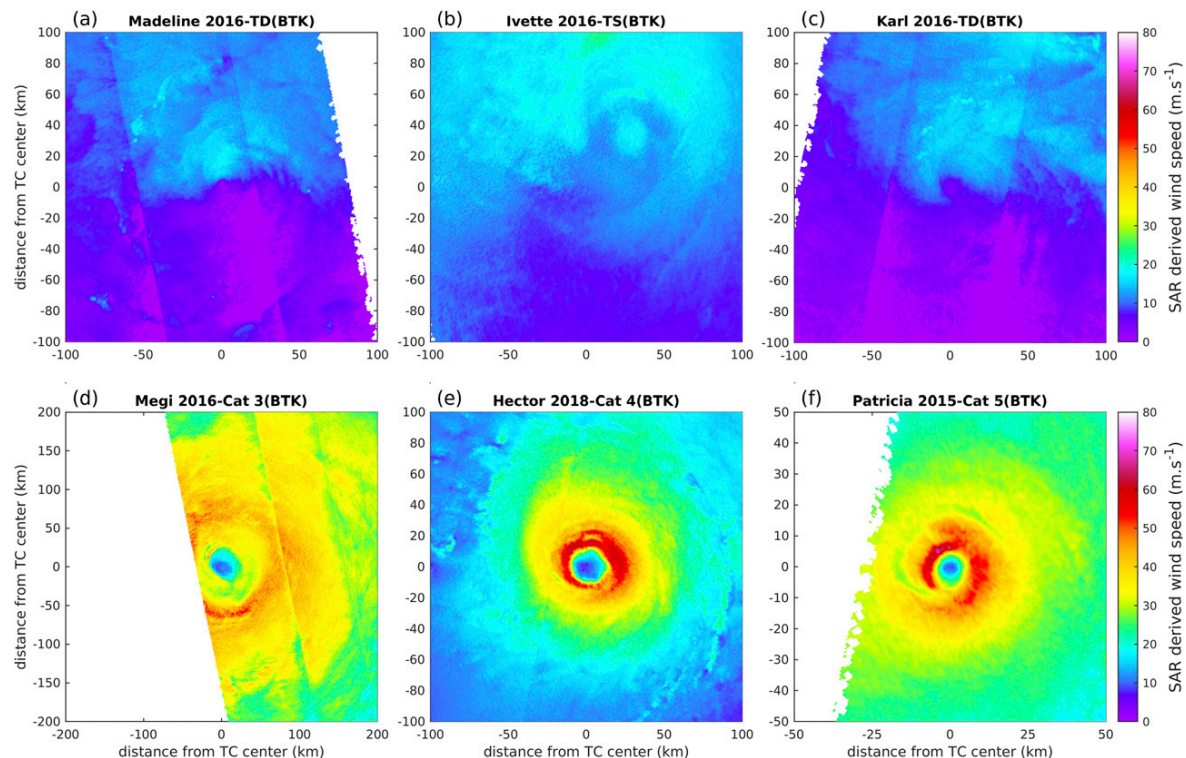


FIG. 10. Mosaic of six specific tropical systems from SAR acquisitions, represented on a polar grid centered at each TC origin. (a),(b) Madeline and Ivette, two weak systems eliminated from analysis due to eye location issues. (c),(d) Two complex structures: (c) Karl, a disorganized TD and (d) Megi, a double-eyewall case. (e),(f) Two intense cases with discrepancies between SAR and BTK: (e) Hector, an open-ocean case without any SFMR measurements, best track-limiting case, and (f) Patricia, a coastline case with SFMR measurements (not for this snapshot), SAR-limiting case. The beam seam effect (signal jump) observed in some panels is induced by the noise floor variation according to incidence angle.

acquisitions. The Vmax difference could be the result of (i) the wind field variability, a SAR acquisition being instantaneous, (ii) the convention used to define Vmax, best track estimates relying on 1-min average wind whereas instantaneous SAR measurements are analyzed at 3 km resolution, and (iii) the use of indirect methods (e.g., Dvorak) to estimate Vmax values in best track analysis. For wind radii, best track estimates are probably affected by the use of low- to medium-resolution sensors and sensors that signals saturate at higher wind speeds. High-resolution SAR estimates are certainly adequate to assess the sensitivity of TC parameters to resolution, as different spatial resolutions can be considered. As for Vmax, the impact of the wind field variability within 6 h on wind radii estimates is also certainly a limitation for direct comparisons. The large scatter and associated uncertainties obtained for Rmax, together with the remarkable consistency observed between SAR and SFMR, indicate the need for establishing more robust and homogeneous methodologies for this parameter. Specific comparisons are shown in Fig. 10, further illustrating complex cases and sources of disagreements between SAR and best track analysis.

As shown Figs. 7 and 9, most of the differences regarding the TC wind structure are found for tropical storms or tropical

depressions, which exhibit the largest uncertainties compared to other storm categories (see Fig. 9). When only considering wind radii corresponding to TD or TS categories ( $V_{\max} < 33 \text{ m s}^{-1}$ ), the scatter index and the RMSE obtained for R34 increase from 22% to 40% and from 39 to 50 km, respectively, and those obtained for R50 increase from 25% to 40% and from 27 to 32 km, respectively. Weaker systems tend to be more asymmetric in their complete radial profile, even in low shear environment (Klotz and Jiang 2017), with a less organized circulation (Fig. 10). TD/TS wind radii (Knaff and Sampson 2015) estimates are thus more difficult to estimate than for hurricanes/typhoons. In addition, at this intensity stage, TD/TS likely do not have an eye structure (Vigh et al. 2012), a constraining situation for our SAR-based methodology. When existing, the eye is often asymmetric (Li et al. 2013), a limiting factor for IR-based techniques that require symmetrical eyes (Mueller et al. 2006; Kossin et al. 2007). Such techniques are also sensitive to the cirrus darkening effect, especially in weak eyes or developing eyewall situations (Velden et al. 2006). Regarding the present dataset, seven weak cases (6 TD and 1 TS) were discarded due to unclear eye structure (see Madeline on Fig. 10a). Another noteworthy case corresponds to an unexpected situation, with a high wind area



measured within the eye region (see Ivette on Fig. 10b). Although likely correct, this case is also removed for the present analysis. After screening, 52 TDs/TSs are included in this study (Fig. 3). The SAR capabilities to examine this range of intensity was already demonstrated in previous studies (Zhang and Perrie 2012). Wind structure can then be used to initialize models to help improving forecast in terms of intensity (Bender et al. 2017), tracks (Kunii 2015) and TC structure (Wu et al. 2010). Low wind speed systems are thus important to precisely describe.

Complex structures also occur during ERC events, for which two distinct wind speed maximum regions can coexist, leading to the existence of double Rmax. During the ERC process, TC goes through significant structural and intensity changes of the inner core, with a widening and phases of weakening and reintensification (Maclay et al. 2008; Kossin and Sitkowski 2012). For instance, Typhoon Megi (2016, WP) encountered this situation, with a first 25 km Rmax ring and a second one at 60 km (Fig. 10d), as observed by SAR measurements. In such a case, significant differences on Rmax estimates can be expected if the two applied methods do not pick the same Rmax. Beyond the comparison, an ERC is a critical situation for forecasters, as it changes the TC size with a potential large increase in integrated kinetic energy (Sitkowski et al. 2011). Wind structure is also of paramount importance to assess storm surge (Irish et al. 2008) and many other applications. Rapid intensification may also follow after eyewall replacement [Andrew in 1992; Landsea et al. (2004)], such rapid changes remain challenging for forecasters and those preparing the best track (Leroux et al. 2018). In total, our dataset includes eight observations of ERC, and SAR observations are particularly suitable to infer the double Rmax at finescale (see Fig. 10d) and to possibly guide on the onset or the reached phase of the ERC. This may help forecasters to reduce errors associated with such event (Kossin and DeMaria 2016).

It is also noteworthy that the inner-core wind structures of major TCs are challenging to estimate. When evolving over open ocean, and lacking aircraft data, best track analyses heavily rely on satellite information (Landsea and Franklin 2013). During intensification, Rmax generally tends to decrease, a limiting situation for most of low- to medium-resolution sensors. Hector was one of these small (Rmax = 16 km) intense systems, developing within the eastern Pacific in 2018 (see Fig. 10e). During its second intensification phase (on 9/10 August 2018), Hector's inner core shrank, and this was not captured by the best track. Rmax was evaluated at 46 km, a large overestimation of about 30 km (200% error). Similar examples can be found—notably in east Pacific basin, such as hurricane Ignacio (3 September 2015), for which a 20 km underestimation was observed. Those significant structure discrepancies can lead to dramatic change in IKE (Powell and Reinhold 2007), and the upper ocean response (Ginis 2002; Kudryavtsev et al. 2019b). Vmax discrepancies also occur during Rapid Intensification (RI) or Rapid Decay (RD) events. For instance, Typhoon Jebi (2018) experienced an explosive increase of  $25 \text{ m s}^{-1}$  in 24 h. In this particular case, up to  $10 \text{ m s}^{-1}$  intensity difference is reached between the two sources. Best track analysis may fail due to strong intensity

variations ( $2.5 T$  maximum; Velden et al. 2006), for which smaller TC cases tend to be more affected (Xu and Wang 2015; Carrasco et al. 2014; Leroux et al. 2018), possibly combined with the best track binning effect. To note, one limiting situation was found using SAR observations with Hurricane Patricia in 2015 (Fig. 10). Analysis of this powerful TC benefited from temporal and spatial sampling from aircraft, surface and satellites (Rogers et al. 2017). Ocean surface wind speeds, with values as high as  $90 \text{ m s}^{-1}$  were measured. SAR Vmax estimates only reached  $72 \text{ m s}^{-1}$ , but Rmax corresponded well with observations with the smallest radius of the present dataset (7 km). Since wind retrievals were performed on a 3 km resolution grid, the full peak intensity may not have been well resolved. Regardless of this specific case, SAR seems to address most situations to complement the quality of TC parameter estimates. This is especially true for highly variable situations, where 50% of the values exceeding MAE of both Vmax and Rmax in Figs. 4b and 7b are defined as fast-evolving cases (including ERC situations). The nature of the highly variable cases is the most limiting factor for SAR–best track comparisons. It is also noted that half of the total adjustments for Vmax, section 3a, were performed for best track estimates identified as highly variable situations for which comparison can be particularly tricky (interpolation failure). For these situations, objective analyses merging microwave and infrared satellites could be used to further investigate these discrepancies between SAR and best track. Those methods include the Satellite Consensus (SATCON; Herndon et al. 2012) product developed by the Cooperative Institute for Meteorological Satellite Studies (CIMSS), or the Multiplatform Tropical Cyclone Surface Wind Analysis (MTCSSA; Knaff et al. 2011) system by the Cooperative Institute for Research in the Atmosphere (CIARA). In particular, because they combine different satellite data sources, they have higher temporal resolution than best track allowing for a more accurate TC intensity variability description, they may provide an interesting complement for further comparisons.

Finally, we also evaluate the impact of the SFMR processing method. As expected from Sapp et al. (2019) and the example presented in section 2c, no matter the product considered, a great consistency is found between SAR and SFMR data. Regarding wind speed parameter, the RMSE and correlation remain very similar, whereas the largest differences are obtained for the bias. Correlation coefficients are 0.93 and 0.92 for AOML/HRD and NOAA/NESDIS processing methods, respectively. RMSE are  $4.32$  and  $3.86 \text{ m s}^{-1}$  for AOML/HRD and NOAA/NESDIS, respectively. Values for bias are different:  $1.49$  and  $-0.24 \text{ m s}^{-1}$  for AOML/HRD and NOAA/NESDIS, respectively. Regarding Rmax parameters, differences are negligible. When comparing the two processing versions, correlation is higher than 0.90, RMSE lower than 4 km and a bias around 0 km. Further comparing the two SFMR processing versions is clearly out of the scope of this study and the choice of SFMR product does not impact our conclusions on SAR–best track comparisons. In particular the main result concerns Rmax parameter for which the consistency between SAR and best track is dramatically improved when SFMR measurements are

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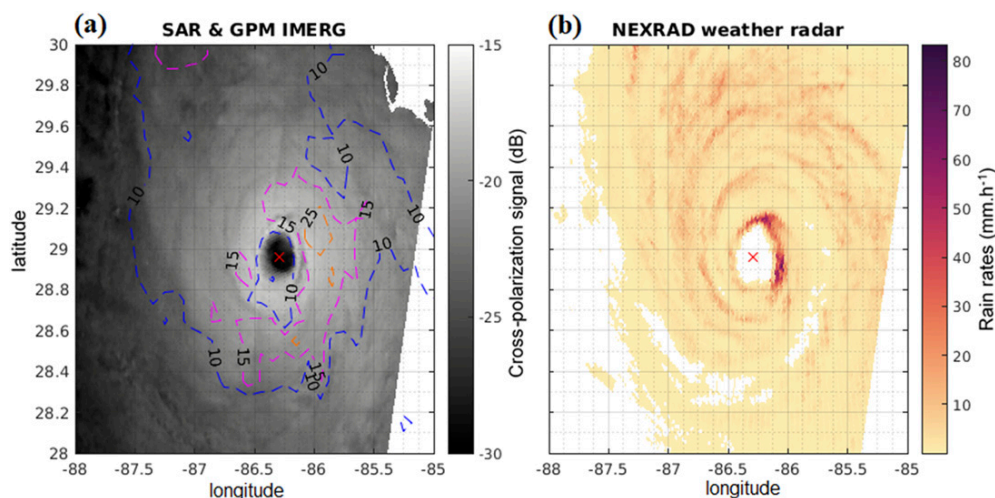


FIG. 11. Comparison of two rain products for the Hurricane Michael case on 10 Oct 2018. (a) Cross-polarization signal of *Sentinel-1A* with rain-rate contours of GPM IMERG ( $\text{mm h}^{-1}$ ). (b) NEXRAD base reflectivity converted to  $\text{mm h}^{-1}$ .

available. And the processing version of SFMR does not impact the location of the strongest wind speeds along the transects. In contrast, the SFMR impact on SAR–best track comparison for  $V_{\text{max}}$  parameter is found to be negligible.

#### b. Rain impact on SAR

Although SAR measurements appear relevant to provide guidance in many limiting situations, specific processing steps are still required to provide qualified estimates (see section 3). Specifically, a special method was developed to remove rain effects for wind estimates. This is important because rain can cause both increase and/or decrease of C-band backscatter signals (Katsaros et al. 2000; Alpers et al. 2016). This results from either modification of the ocean surface waves, damping and enhancement of the roughness by rain droplets impinging the surface, or from direct interactions with hydrometeors (scattering or attenuation) in the atmosphere column along the path of the electromagnetic waves.

For the present database, rain rate estimates are systematically inferred from IMERG product, collocated in time and space with SAR acquisitions. The rain intensity for a given TC is defined by the mean rain rate ( $\text{mm h}^{-1}$ ), estimated on a  $\pm 10$  km ring around SAR-derived  $R_{\text{max}}$ . In the case of CAT-1 and higher TC, heavy rain occurrence is defined by mean rain rates larger than  $30 \text{ mm h}^{-1}$ . This threshold is lowered for TD/TS to  $10 \text{ mm h}^{-1}$ . Applying these criteria, 70 (42%) cases correspond to heavy rainfall, and 23 of them (1/3) were found to be significantly affected by rain. As briefly mentioned in section 3, a case is considered significantly affected by rain, if an anomalous peak is found in the SAR-derived  $V_{\text{max}}$  azimuthal distribution, and can be associated to heavy rainfall as indicated by IMERG. To help distinguish several maxima, we use the Jelesniansky description [Jelesniansky (1966), detailed in

Pan et al. (2016)], taking into account the TC translation induced  $V_{\text{max}}$  asymmetry.

When applied, this procedure improves consistency between SAR and best track (Fig. 4a). However, four remaining cases exhibit anomalous peaks in the  $V_{\text{max}}$  azimuthal distribution, for which the heavy rainfall criterion derived from IMERG is not met. Figure 11 illustrates one of these cases, Hurricane Michael (2018). In Michael, IMERG measurements barely reach  $25 \text{ mm h}^{-1}$ . In contrast, the high reflectivity measurements by the KEVX radar (from NEXRAD network) around the eyewall tend to indicate heavy rainfall [ $>50 \text{ dBZ}$ , corresponding to about  $80 \text{ mm h}^{-1}$  using  $Z(R)$  relation from Fulton et al. (1998)] that was not captured by IMERG. This case demonstrates the limitation of using medium-resolution rain products such as IMERG, for both time and space sampling, 30 min and  $0.1^\circ$ , respectively. In fact, rain rate is often related with maximum wind intensity (Lin et al. 2015), and its variability (Rodgers and Adler 1981) with maximum activity generally located in the inner TC core (Lin et al. 2015; Jiang et al. 2008). In this deep convective region, the diameter of rain cores can be smaller than IMERG resolution ( $\sim 5\text{--}10 \text{ km}$ ), with rain rate exceeding  $20 \text{ mm h}^{-1}$  (Begum and Otung 2009). High-resolution weather radar (1 km) can capture this activity. Among the 17 cases collocated with NEXRAD, four additional heavy rain cases like the Michael case were found. 70% of the adjusted SAR-derived  $V_{\text{max}}$  estimates correspond to cases with impacting heavy rainfall, yielding mean adjustments of  $6.2 \text{ m s}^{-1}$  ( $4.5 \text{ m s}^{-1}$  for hurricanes and  $8.1 \text{ m s}^{-1}$  for TDs and TSs). This confirms that intense rain is the most limiting factor to retrieve  $V_{\text{max}}$  from C-band SAR measurements. Interestingly, SAR-derived  $R_{\text{max}}$  are generally quite insensitive to rain issues for hurricane-force situations where rain-induced ambiguities are close to the region of maximum winds and rather symmetric. However, for five TDs, SAR backscattered signal enhancement associated with outer rainband activity, can compete with the



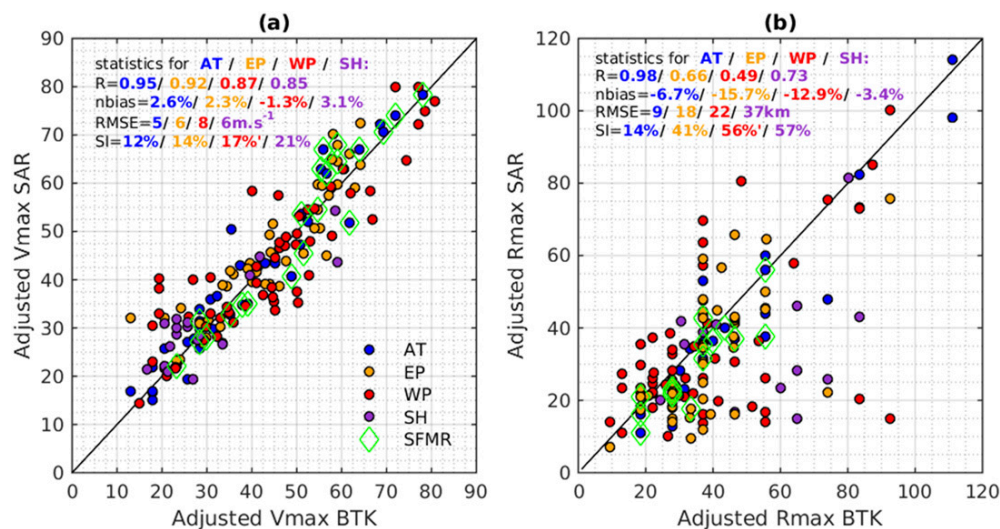


FIG. 12. Interbasin comparison of inner-core parameters. As in Figs. 4b and 7b, but with respect to basin location. Cases with simultaneous SFMR measurements are framed with green diamond. Statistics are added for each basin.

weak TD wind signature (e.g., Karl in Fig. 10). Still, the global impact on Rmax determination is generally minimal.

### c. Discrepancies among TC basins

Due to its global coverage and unique high resolution, SAR can also assess TC parameter variability and document the best track quality homogeneity for both inner and outer core, among different ocean basins. TC characteristics regarding size, intensity and trajectory (Chavas et al. 2016), as well as best track quality (Landsea and Franklin 2013), also depend on ocean basin. In particular, improved performances are achieved when aircraft reconnaissance and SFMR measurements are available (Landsea and Franklin 2013; Sampson et al. 2018). Figure 12 shows Vmax and Rmax comparisons as function of basin location and SFMR coverage. TD cases are not considered for this interbasin comparison, as none of them benefited from SFMR observations. To note, unlike in section 3, we do not compare direct SFMR information but rather best track analyses that are influenced by SFMR measurements. Inherent smoothing in the best track analysis may slightly reduce the correlation presented in Fig. 8. Still the benefit of using SFMR is obvious for both Vmax and Rmax parameters (see Fig. 12) with improved agreement between SAR and best track when SFMR observations are available. As listed in Table 1, this especially applies for R50, R64 and Rmax wind structure parameters. Only R34 parameters are found to be very similar. This certainly reflects the use of scatterometer measurements to improve R34 estimates for best track (Brennan et al. 2009).

Regarding performances for each individual basin (Fig. 12), the North Atlantic basin logically experiences higher consistency with SAR ( $R > 0.90$ ), thanks to SFMR observations. Similar performances for TC intensity are obtained in the east Pacific ( $R = 0.91$ ). Overall, and despite a slight decay for non-U.S. basins, Vmax performances are found very robust for all

basins. Significant discrepancies between basins primarily exist for Rmax. More specifically, east Pacific undergoes the strongest normalized bias ( $-16\%$ ). This is likely associated to the prominence of smaller storms (Chavas et al. 2016; Chan and Chan 2012) combined with a general lack of aircraft data in this basin. In addition, those small systems appear to experience higher RI/RD rate (Carrasco et al. 2014; Xu and Wang 2015), making TCs developing in the east Pacific more temporally variable. For the west Pacific basin, a larger distribution of TC sizes (Chavas et al. 2016), together with the largest systems observed (Knaff et al. 2014) is expected. This, combined with an absence of SFMR measurements, may explain the large RMSE and scatter in estimates. Likewise, the Southern Hemisphere (SH) shows large scatter. However, the small number of available cases in the present dataset prevents any strong and definitive conclusions. To note, specific adjustments are applied to the initial Dvorak analysis scheme by each RSMC, with respect to their measurements capabilities (Velden et al. 2017) and performances could be different than those obtained here as we are only using the JTWC analysis. Furthermore, Southern Hemisphere never benefits from SFMR measurements. This can explain the low consistency achieved for Vmax.

## 5. Conclusions and prospects

Thanks to an unprecedented large dataset, consisting of 161 acquisitions from three different SAR instruments (RS2, SIA/B), SAR observations are shown to not only explore the full spectrum of TC intensities, but also to precisely detail TC structure parameters from wind radii well removed from the eyewall region to the location of the maximum wind and Rmax in small TCs. Rmax is a parameter of paramount importance for the assessment and forecasting of ocean-atmosphere interactions, damages (Powell and Reinhold 2007), storm surge

TABLE 1. Mean absolute errors and uncertainties in parentheses of each TC parameter according to presence of SFMR information.

|                       | Vmax (m s <sup>-1</sup> ) | Rmax (km)  | R34 (km)   | R50 (km)     | R64 (km)   |
|-----------------------|---------------------------|------------|------------|--------------|------------|
| BTK <sub>SFMR</sub>   | 4.0 (8%)                  | 6.0 (18%)  | 26.5 (15%) | 12.3 (12.8%) | 9.5 (15%)  |
| BTK <sub>NoSFMR</sub> | 5.1 (13%)                 | 16.5 (33%) | 30.0 (17%) | 21.6 (20%)   | 16.0 (26%) |

(Irish et al. 2008), and also to initialize parametric models (Holland 1980; Wood et al. 2013; Willoughby and Rahn 2004). In addition to the potential to resolve the complete wind field (Fig. 1), SAR measurements are found fully consistent with SFMR ( $R > 0.9$ ). SAR acquisitions can be performed everywhere, independent of TC basin and best track. Although an acquisition scheme methodology is required (see section 2), comparisons (see section 3) highlight the ability of SAR to routinely guide analysis, especially in limiting situations, where significant variability is noticed and jeopardize best track estimates (50% of discrepancies).

As previously reported (Horstmann et al. 2013, 2015), SAR signals still require special processing steps and screening before being used to infer TC parameters, especially Vmax estimates made in heavy rain conditions. Here we have shown that heavy rain possibly account for 20% of observed scatter of Vmax (Fig. 4a). More dedicated efforts are certainly required, with use of IMERG products, GPM's satellite measurements, weather ground-based radar, or other satellite-based product such as MIMIC (Wimmers and Velden 2007) to provide best-qualified Vmax estimates. Following the steps detailed in this paper, high-quality Vmax estimates can be obtained with respect to both SFMR ( $R = 0.91$ ) and best track ( $R = 0.92$ ). Unlike Vmax, Rmax estimates appear only weakly affected by heavy precipitation, leading to almost perfect agreement with SFMR ( $R = 0.98$ ). More significant discrepancies in Rmax and Vmax comparisons with best track occur when storms are evolving quickly or when there are double Rmax structures (Fig. 7). We also observe that wind radii uncertainties increase with TC intensity (Fig. 9). Since temporal variability has an impact on best track wind structure estimates (50%), the lack of more direct methods seems to be the most impactful to the results of our study. These impacts are highlighted by comparing performances over different TC basins: highest scores are obtained for the Atlantic basin, which benefits from good SFMR coverage (70% of our TC cases). Likewise, R34 was found very consistent regardless of TC basins or SFMR available observations (Table 1), which highlights the real benefit of using scatterometer measurements in best track analysis.

This apparent reliance on scatterometry implies that TC wind structure analyses could immediately benefit from the new capabilities of the recent generation of rain-free L-band passive radiometer sensors: SMAP/SMOS that depict TC, still at the low resolution of 50 km, but with a wide swath and twice-daily temporal coverage (Reul et al. 2016, 2017). In such a context, SMAP measurements have been recently added to ATCF (JTWC 2017). It has been shown that the coarse resolution of these observations generally precludes direct estimation of the inner-core parameters for TCs with Rmax less than 40 km (most of cases  $\geq$  CAT-2; see Fig. 9). But these satellites can provide very accurate wind radii that have been

found to be in good agreement with both SAR (Mouche et al. 2017; Zhao et al. 2018) and SFMR (Reul et al. 2016; Meissner et al. 2017) estimates. Today, the SMOS/SMAP constellation, possibly augmented by AMSR-2 measurements, combined with the unique capability of SAR to inform the eyewall region, offer quite unique opportunities. In addition, the CYGNSS constellation with its high temporal resolution and insensitivity to high rain rates (Ruf et al. 2016; Morris and Ruf 2017) can also be an asset to complement SAR measurements, especially in situation of heavy rain conditions.

As shown here, improved TC parameter estimates can be robustly derived using SAR winds corrected for heavy rainfall to lower errors associated to more indirect and subjective analysis methods. However, to date, no MDA-operating *RadarSat-2* or Copernicus/ESA-operating Sentinel-1 ensures any operational service to systematically acquire data over TC and then process and disseminate them into wind field in near-real time. The new version of 3-hourly IBTraCs (Knapp et al. 2018) and annual best track preparation at NHC and JTWC may benefit from this growing combined capability using these different sensors in the future. It will bring finer description of TC evolution and wind field variability that should directly benefit operational, applied and/or research communities. Presently, best track uncertainties appear to largely depend on the availability of SFMR information (Table 1), leading to large discrepancies between basins (Fig. 12). Our results have also shown that best track errors are in agreement with previous studies (Torn and Snyder 2012; Landsea and Franklin 2013; Knaff and Sampson 2015): 10% uncertainties for intensity and 20% for wind radii (section 3). More importantly, uncertainties for Rmax, which we found to be 32%, were not previously assessed. Clearly, SAR observations can thus guide poststorm analysis of this important parameter, which is not yet reanalyzed as part of the best track process. In a future study, SATCON (Velden and Herndon 2020) and MTCSWA analysis products may be considered to extend this work. Their higher temporal resolution and the use of recent methods, like specific microwave algorithms (Wimmers and Velden 2016) can be an alternate source of comparison, notably for Rmax and complex situation such as ERC event. Furthermore, these methods are generally based on 89–91 GHz microwave radiometers that provide elevated eye features and midlevel winds (MTCSWA). A joint use of C-Band SAR and 89–91 GHz microwave radiometers could allow estimates of the vertical eyewall slope and possibly lead to additional guidance in the surface adjustment to be performed. Finally, SAR measurements can also document the TC eye dynamics (Li et al. 2013), and provide high-resolution details of the nature of the TC's wind field. Thanks to the growing database, future investigations will be conducted to more carefully analyze the TC's finer-scale wind structures, to not only provide details of the

radial and azimuthal wind variations, but details of the organization of large eddies that occur at various spatial scales (Foster 2005) within the TC.

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# Parametric models: Generalisation of the approach.

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Parametric solutions provide a synthetic, low-cost (computational time) and easy-to-implement approach to depict the cyclonic vortex, which they attempt to interpret through a simple semi-empirical relationships based on a few key parameters (section 1.2.2). They are convenient for forcing an oceanic response model [Young 1988; Vincent et al. 2012a; Kudryavtsev et al. 2019a], statistically analysing wind field properties [Knaff et al. 2011b, 2016], or for completing the wind profile and providing estimates of outer/intermediate radii from partial surface observations [Reppucci et al. 2010] or parameters extrapolated from upper/middle atmosphere measurements [Demuth et al. 2006; Mueller et al. 2006; Knaff et al. 2016]. Their relevance is intrinsically bound to their founding assumptions and the free parameters required as input, which together define their validity framework. In section 1.2 we presented some of the most common models and in section 1.4 we highlighted the classical limitations. Most of them are based on the Holland model scheme, and rely on the information of Vmax, the latitude and Rmax to guide the behaviour and extent of the tangential winds. Two major issues arise from this initial configuration though, the dependence on Rmax and the description of the outer region. The discrepancies revealed with the Hector profile in the figure 1.32 of section 1.4.3 and between SAR and BTK information in Chapter 3, reflect both the difficulty of measuring Rmax and the misreading of the wind field variability by the Rmax-based models, mainly for the outer core. So far, we have mainly addressed the first point, but we will now add some additional elements to delve into these two issues and shed light on our choice of parametric solution. This, together with the following results, will fully justify the use of the Chavas model to capture both inner and outer core structure.

## 4.1 Model paradigm, description and evolution.

### 4.1.1 Problematic of the two modes of variability.

At this stage of the part II, we seek to provide an auxiliary method with infrequent SAR information, to correctly infer  $R_{\max}$  and the overall cyclone structure. This solution must answer the two main issues:

#### 1) $R_{\max}$ variability

As we saw in the introduction,  $R_{\max}$  remains an elusive parameter, accessible only by a small handful of highly accurate, but spatially (SFMR) or temporally (SAR) intermittent, high resolution instruments. So far, only SAR sensors have been able to document the entire 2D wind field and accurately capture  $R_{\max}$  over all basins (Chapter 3). Conversely, the methods deployed in the BTK suffer from uncertainties that are too large to allow a reliable diagnosis of the area of strongest winds, especially for major hurricanes (>30%, see Chapter 3). As this parameter is also not reanalysed [Knaff et al. 2021], it further exposes BTK to the vagaries of the fine-scale (hours-days) temporal variations of the inner core [Weatherford and Gray 1988b; Sitkowski et al. 2011; Kossin and Sitkowski 2012; Musgrave et al. 2012], including rapid intensification events [Carrasco et al. 2014; Sangster and Landsea 2020]. Indeed, of all the estimated metrics,  $R_{\max}$  is the one that undergoes both the largest intra- and inter-storm evolution [Chavas and Lin 2016], i.e. whether they are dominated by changes in intensity, internal structure or cyclone displacement (intra-storm), or whether they are related to variations in the absolute length scale (i.e "size") of the storms within and between basins (inter-storm). It is therefore not straightforward to bypass the  $R_{\max}$  information and try to derive it from the other parameters of the cyclonic wind field, given the differences in the existing dynamics.

If  $R_{\max}$  is deeply attached to the strong convective dynamics of the eyewall, it remains sensitive to variations in size between tropical systems. When the wind field is averaged over the TC lifetime, the wind radii from  $R_{\max}$ ,  $r_{33}$  to  $R_{17}$  are found strongly correlated (0.5-0.7) to each other [Chavas and Lin 2016], meaning the  $R_{\max}$  also follows a strong inter-storm variability, dictated by the absolute TC length scale [Chavas et al. 2016]. Larger initial vortex thus tend to have larger overall structure [Chan and Chan 2014]. Environmental factors may play a part, higher relative humidity will enhance the development of eyewalls and primary and secondary rainbands, resulting in higher  $R_{\max}$  and in the meantime, larger outer radii ( $r_{17}$ , Xu and Wang [2010]). Similarly, if the life cycle variations of external radii are more stable in time, their intra-storm variability can in some cases lead to changes in  $R_{\max}$  through accumulation of angular momentum [Chan and Chan 2015b]. It is therefore difficult to examine  $R_{\max}$  solely on the basis of  $V_{\max}$  and/or latitude like in Quiring et al. [2011]; Knaff et al. [2015], especially as the two parameters of the inner core also follow different fluctuations [Shea and Gray 1973], even if an inversely proportional relationship can be found between the two terms [Quiring et al. 2011; Vigh et al. 2012] (see Figure 9 of Combet et al. [2020a]), not to mention regression laws using this method are often very basin dependent and cannot be extrapolated [Quiring et al. 2011]. Their covariation remains tenuous, as the standard deviation of the observed  $R_{\max}$  values for each wind intensity category is of the same order of magnitude as the  $R_{\max}$  itself. To deduce  $R_{\max}$ , it is necessary to reconcile the different time scales associated with the inner and outer core and, above all, the two modes of variability that profoundly affect its evolution.

## 2) Outer core behavior

The second main problem then arises from the interpretation of the outer radii, which are often underestimated in most parametric models, as illustrated in Section 1.4.3. Finding key parameters to anticipate the cyclone external profile requires to have a comprehensive understanding of the impressive collection of mechanism that modulate its dynamic, as listed by [Chan and Chan \[2015a\]](#), and space-time parameters such as region, season and latitude that shape the synoptic environment [[Knaff et al. 2014](#)]. While their significance in the observed size distribution is not necessarily fairly understood, their influence depends as well on the chosen metric <sup>1</sup>, whose sensitivity is not the same to external forcing (shear, planetary vorticity...) [[Klotz and Jiang \[2017\]](#); [Chavas et al. \[2016\]](#)], nor its behaviour when the intensity drops [[Kimball and Mulekar 2004](#)]. To recall, whenever we mention size, we are referring to the outermost wind radii of the tail of the radial profile (r17 to ROCI), which are characteristic of each storm because of their relative temporal stability [[Chavas et al. 2016](#)] (as compared to the inner core). At this distance from the eye, the outer circulation is mainly in equilibrium with the free large-scale troposphere, which controls both its geometry (through shear, [Klotz and Jiang \[2017\]](#)) and partly its evolution [Chan and Chan \[2013\]](#).

Studies carried by [Holland and Merrill \[1984\]](#); [Weatherford and Gray \[1988b\]](#); [Chan and Chan \[2013, 2014\]](#) have shown, both observationally and numerically, that if the factors favouring the vertical flux (convection) and the upper-tropospheric outflow are mainly conducive to intensification, those impacting the lower tropospheric import of angular momentum mainly leads to size changes. Coexistence of anticyclonic flows bordering the TC, which occurs during the late seasonal shift of the subtropical ridge activity (October), provides an increased source of convergent flow to the TC circulation [[Chan and Chan 2013](#)]. These distinct changes of the wind field properties are also acting on different time scales. In contrast to the highly turbulent inner core [[Gray and Shea 1973](#)], the size of the TC is more stable and fluctuates over longer time periods ( $\vartheta(10 \text{ days})$ ), as it is mainly influenced by the large-scale [[Weatherford and Gray 1988b](#); [Chavas et al. 2016](#)]. The inner and outer cores thus appear to be relatively independent [[Merrill 1984](#); [Holland and Merrill 1984](#)], although the deep convection area is sustained by the inflow of angular momentum starting at the edge of the TC circulation. [Chavas and Lin \[2016\]](#) nicely resumed the cyclonic wind field as a vortex within a vortex. Several observations have indeed illustrated the weak relationship between intensity and size [[Weatherford and Gray 1988a](#); [Chan and Chan 2012](#)], or at least its non-linear nature [[Wu et al. 2015](#)]. Some interdependencies may still emerge when the TC life cycle stages are decoupled [[Musgrave et al. 2012](#); [Chan and Chan 2013](#)], or when the eye extent is taken into account, as it also reflects changes in structure [[Weatherford and Gray 1988b](#); [Kimball and Mulekar 2004](#)]. For instance, [Knaff et al. \[2014\]](#); [Chan and Chan \[2015b\]](#) showed that during the initial and intensification stages, intensity affects the cyclone growth rate, as well as eyewall replacement cycles induce significant size changes [[Maclay et al. 2008](#)]. Inversely, [Carrasco et al. \[2014\]](#); [Xu and Wang \[2015\]](#) observed a stronger occurrence of rapid intensification for small tropical cyclones, potentially due to higher sensitivity to large-scale interaction. Despite these relationships, the dynamics of the outer core still prevails on the evolution of TC size [[Chan and Chan 2015b](#)], and the covariation of the intensity with wind radii extent of r17, r15 or r12 never exceeds 0.3 correlation [[Merrill 1984](#); [Chavas and Lin 2016](#)].

However most parametric models ingest inner core variables ( $V_{\max}/R_{\max}$ ), and rel-

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<sup>1</sup>Among these are r33, r25 but more generally r17, r15, 12 and ROCI which correspond more to the classical definition of cyclone "size".

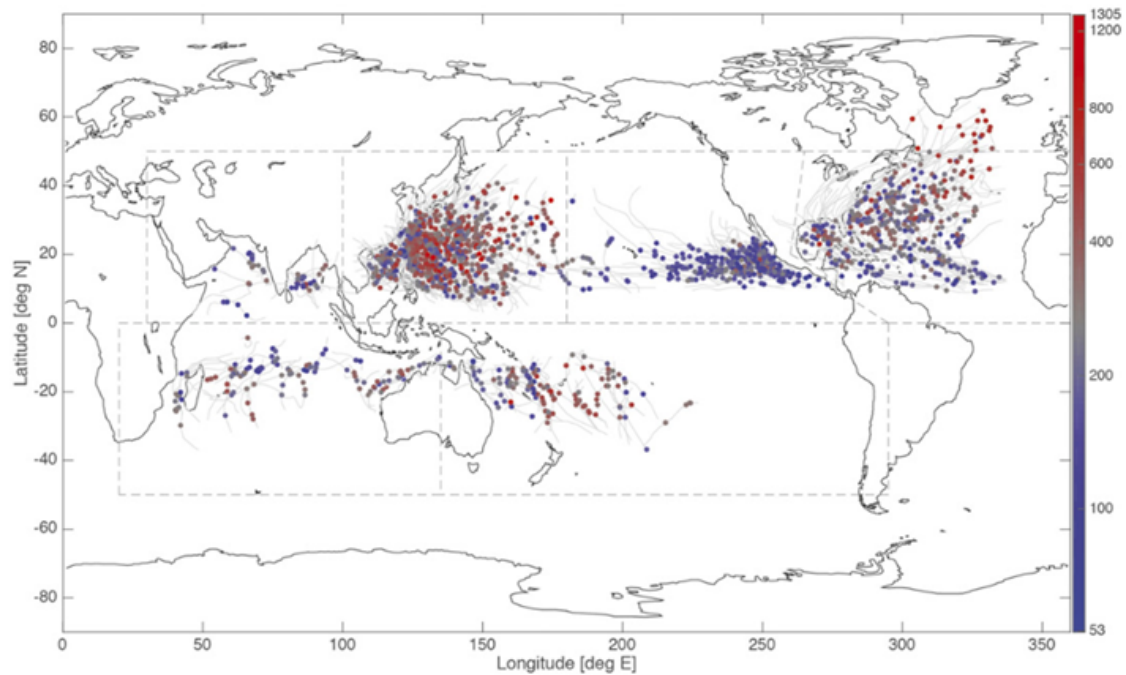


Figure 4.1: Map of spatial distribution of storm size:  $r_{12}$  (km) estimated from the QuikSCAT database (1999–2009) from the article [Chavas et al. \[2016\]](#)

egate the control of the outer core to the Coriolis parameter. The impact of latitude has since been mitigated, with planetary vorticity obviously playing a crucial role in the primary circulation and its extent [[Knaff et al. 2014](#)], particularly during the TC lifetime as it moves poleward and gains EAM, but its contribution does not explain the size distribution within and between the basins (Fig 4.1). Several recent studies have shown that there is no simple relationship between latitude and cyclone size, an optimal band even exists around 20–25° N/S and reflects a trade-off between the gain in planetary vorticity and the resistance to influx due to the increase in inertial stability when moving to higher latitudes [[Chan and Chan 2015b](#)]. This non-monotonic relationship was demonstrated mainly on the basis of MW observations from QuikSCAT [Chan and Chan \[2012\]](#); [Chavas et al. \[2016\]](#) and was later confirmed by numerical analyses [[Chan and Chan 2015b](#); [Wang and Toumi 2022](#)]. The results we conducted in Chapter 2, with  $r_{12}/r_{17}$  sampled by L-band radiometers and the support of AMSR-2 and the ASCAT constellation (Fig 2.5), converge in that direction. The Figure 4.1 taken from [Chavas et al. \[2016\]](#), clearly displays the wide dispersion of sizes observed across and within the basin for a same zonal band, particularly in the WP, SH, and NA regions. While latitude remains a critical parameter, particularly in understanding the marginal sizes of EP systems and extratropical storms (obvious in NA basin), the initial vortex has been proven to be dominant to interpret the inter-storm variability [[Chan and Chan 2014](#); [Wang and Toumi 2022](#)]. A more direct connection has also been found between relative SST <sup>2</sup> and the global distribution of absolute TC length scale [[Lin et al. 2015](#)], since it would be a good indicator of the higher concentration of relative humidity in the troposphere, favouring the development of larger structures as mentioned earlier.

To this end, [[Holland et al. 2010](#)] has added an oceanic control parameter to its original model, which transforms the relationship of its shape parameter (B-parameter) and the definition of the rest of the wind profile. As the TC size variability is mainly related to

<sup>2</sup>difference between the local temperature and the mean tropical SST band [30°S–30°N].



the initial vortex properties and the environment in which it is embedded, one can even assimilate a cyclonic vortex into a global forcing model [Vincent et al. 2013], in order to resolve both the structure of the inner core and the large scale flow. Both solutions, however, remain highly dependent on Rmax information and add to the complexity of the external parameters required. Since size can be considered as an inherent property of the cyclone itself, it would be more obvious to guide the profile with an outer core metric in combination with easily observable inner core information. It is in this context that the Chavas model has been designed and may address the different variability issues, as we have very briefly illustrated with Hector in section 1.5.5. The different points raised in this subsection allow us to partly justify our choice of parametric model.

### 4.1.2 Description of the radial profile.

We have already introduced the two fundamental formulas of the model, which describe the dynamics of the inner and outer region of the cyclone respectively, to recall:

$$\frac{M(r)^{2-(C_k/C_d)}}{M_m} = \frac{2(r/R_{max})^2}{2 - (C_k/C_d) + (C_k/C_d)(r/R_{max})^2} \quad (4.1)$$

$$\frac{\delta M(r)}{\delta r} = \frac{2C_d}{W_{cool}} \frac{(rV)^2}{r_o^2 - r^2} \quad (4.2)$$

and with  $M_{R_{max}} = R_{max} V_{max} + \frac{1}{2} f(R_{max}^2)$ . Three main strengths emerge from this framework:

- 1) It dissociates the two distinct and independent thermodynamic regions of the cyclone, in order to better account for their own dynamics, and it builds on previous works by [Emanuel 2004; Emanuel and Rotunno 2011]. On the one hand, it allows the description of the deep convection region, characterised by the strongest winds and the ascending flux at centre, which link the angular momentum distribution to the stratification of the outflow aloft (Eq 4.1) through the exchange term of the enthalpy and momentum flux ( $C_k/C_d$ ) [Emanuel and Rotunno 2011]. On the other hand, equation 4.2 reflects the radial distribution of angular momentum in the convection-free region defined by its extension  $r_0$  (or any radii associated to the absolute TC length scale, like  $r_{12}$ ,  $r_{17}$ ) in radiative-subsidence balance with the free troposphere [Emanuel 2004]. Under the radiative cooling, air subsides at constant rate ( $W_{cool}$ ), while in return the frictional surface inflow ( $C_d$ ) acts on the relative vorticity ( $rV$  term) and induces Ekman suction at the top of boundary layer, that equates  $W_{cool}$ . These two equations mirror the turbulent convective regime of the inner core and the dominant large-scale regime of the free troposphere over the outer core circulation, merging them provide the complete profile [Chavas et al. 2015].
- 2) The intersection method which ensures the versatility of the model. Much more than a simple fusion method, it allows the two regimes to be combined, but above all to interconnect and play with the variabilities associated with each region [Chavas and Lin 2016]. Indeed, for a set of initial variables, there is a unique solution which passes through the fusion radius  $R_a$  and which respects the conditions of continuity between the two equations, so that the angular momentum ratio ( $M_m/M_o = \frac{M_m}{M_a} \frac{M_a}{M_o}$ )<sup>3</sup>

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<sup>3</sup>with  $M_m$ ,  $M_a$  and  $M_o$  respectively the angular momentum at Rmax, the merge radius and outer radii ( $r_o$  or any outer radii)

is written as a function of the two initial free parameters and the environmental parameters  $C_k/C_d$ ,  $W_{cool}/C_d$  and  $f$ .

- 3) The model allows two possible routes to solve the complete profile and determine the merge radius. Either describe the inner core completely with the  $V_{max}$  /  $R_{max}$  (classical method), in which case the model fixes the inner core and attaches the outer solution to its tail, or use the information of the  $V_{max}$  with that of an outer radius to constrain the size of the system, in which case this approach fixes the outer region and finds the optimal and singular solution that passes through the  $V_{max}$  and matches to the outer profile.

We of course choose this second option because it encompasses all the different modes of TC variability, unlike the ( $V_{max}/R_{max}$ ) pair, as explained earlier. In his second paper, [Chavas and Lin 2016] argues that the mechanisms underlying the radial distribution of angular momentum are much more explicit with this mode of variation, since any local increase in angular momentum (intensification, TC growth) requires an inward inflow from the largest radii, notably from  $r_0$ , which corresponds for the rest of the system to the initial source of angular momentum and to the radius where it is maximal ( $fr_0^2$ ). The fixing of this metric ought to better depict the  $M/M_o$  fluctuations within the cyclone. However, as this radius is difficult to observe, the model allows the use of another near radius from the outer profile and indicative of the TC size, from which  $r_0$  is then deduced. The radii that have so far proved most interesting for measuring the width of the cyclone are  $r_{12}$ ,  $r_{15}$  and  $r_{17}$ , according to the literature we aforementioned. Although all three provide robust definition of the TC radial extension, we find  $r_{17}$  more suitable as it combines several advantages, notably the fact that it is more often complete when sampled by our MW surface winds observations. We also found that while  $r_{12}$  provides a better representation of the quiescent region below its intensity threshold,  $r_{15}$  and  $r_{17}$  help to better capture the rest of the profile, including the high wind region which is more critical for us. This has been established by several preliminary analyses, which would require more effort to be truly conclusive, as they were performed on a rather small sample of cases (not shown here).

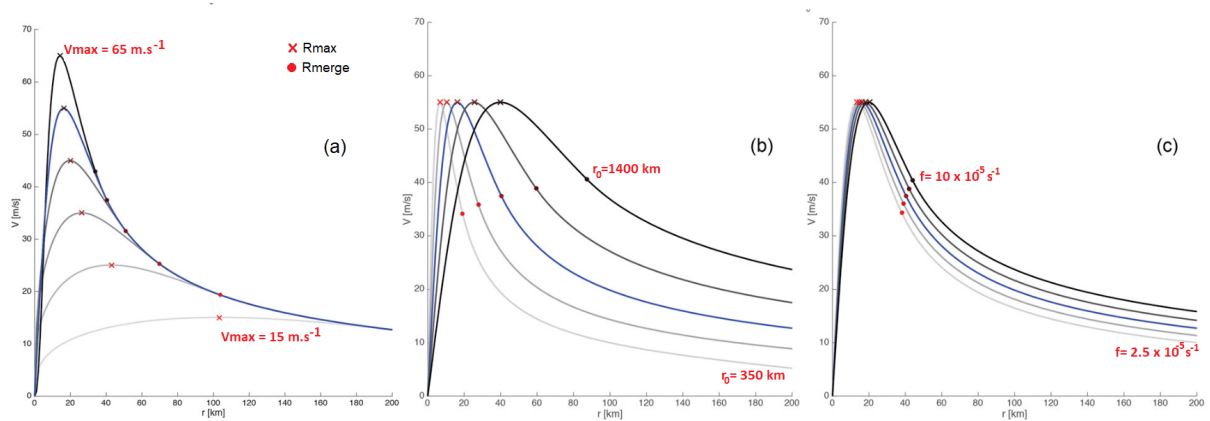


Figure 4.2: *Model variability associated with a) varying  $V_{max}$  at constant  $fr_0$ , for  $V_{max} \in [15:65] \text{ m.s}^{-1}$ , and with varying  $fr_0$  b) for constant  $V_{max}$  and  $f$  but varying  $r_0 \in [350:1400] \text{ km}$  and c) constant  $r_0$  with fluctuating  $f \in [2.5:10] \times 10^{-5} \text{ s}^{-1}$ .*

Once the  $V_{max}$ ,  $r_{17}$  and latitude information has been ingested, the model is then able to reproduce the changes dictated by their modulations. This is what the Figure 4.2

extracted from [Chavas and Lin 2016] shows theoretically, with the three major structural changes associated with the a) intensification, b) the widening of the absolute length of the cyclone or c), the poleward shift. For example, we note the contraction/expansion of the inner core relative to a stable outer circulation, following the intensification/decrease of the storm with an inward/outward shift of the merge radius, as predicted by the conservation of angular momentum [Emanuel and Rotunno 2011]. This is accompanied by the growth of some intermediate radii that can be also observed in reality [Knaff et al. 2014]. The model thus reflects the intra-storm variability related to intensity changes that occur independently of the outer core. Furthermore, panels b and c describe respectively the subtle latitude-induced variations that smoothly modify the short and intermediate radii, and the drastic changes on the global radii induced by the fluctuation of the  $r_0$  extent. This last variation accounts for the very large inter-storm variability observed between and within basins, which evolves independently of the intensity. In the model, the information on the outer radius acts on the whole profile and monotonously but non-uniformly rescales each radius with respect to its dimension. Latitude and  $r_0$  (or  $r_{17}$  for us) contribute to the outward propagation of the merge radius. These different experiments conducted by [Chavas and Lin 2016] on the model are very consistent with the observations and numerical analyses we have cited above. Moreover, they have been compared with QuikSCAT observations for winds below 35 m.s<sup>-1</sup>, and with the Hwind analysis for the inner core, which reconstructs a 2D field from several platforms and SFMR measurements. These two datasets were also used to calibrate the environmental parameters  $C_k/C_d$  and  $W_{cool}$  by optimally fitting the parametric profiles to the observations. The former increases progressively with intensity over a short interval and the  $W_{cool}$  is set as a constant. For information purposes, we have kept their respective values. Their influences on the shape of the radial profile are marginal, due to the strong constraint exerted by the two free parameters, at least for the mode (Vmax,r17).

Ultimately, the Chavas model predicts Rmax and its variations as a function of competing variations in Vmax,  $f$  and an outer radius. Each of these parameters induces a different impact that respects the observed variability modes and covariations between the different radii. The most significant changes in the structure are generated by fluctuations in the size parameter, in agreement with the observations [Chan and Chan 2014, 2015b; Wang and Toumi 2022]. The information from the outer radius helps to guide the model and indirectly incorporates the environmental conditions that have shaped its extension. If, like the Holland model, the model equations originally describe a gradient wind (typically at flight level), the use of a surface wind radius ought to rescale to surface level [Holland et al. 2010]. At last the combined use of Vmax and  $r_{17}$  information has a double advantage. Both parameters are readily available, with Vmax documented every six hours by the BTK, which provides robust estimates in excellent agreement with high-resolution SAR and SFMR data ( $\sim 5\text{m.s}^{-1}$  accuracy, Combot et al. [2020a]), and  $r_{17}$  easily observable by all radiometers and scatterometers, which are in the upper performance range of all these instruments [Chou et al. 2013; Reul et al. 2017; Knaff et al. 2021] and benefit from high-ratio spatio-temporal sampling.

While there are of course limitations to the model capabilities, notably the discrepancies found for winds around the merge point, as well as the difficulties in predicting large Rmax for intense cyclones (intra-storm mode limitation) [Chavas and Lin 2016], one of the main shortcomings we need to address is the 1D nature of the model. Indeed, the model we have described so far is a 1D profile, integrated over all azimuths. We therefore need to apply an asymmetry factor.

### 4.1.3 Transformation to a 2D wind field.

From the integrated solution, we want to transform the parametric profile into a 2D wind field. We can use to our advantage the alternative route proposed by the model (Vmax, r17) because it relies on two parameters whose azimuthal distribution we can easily estimate.

- 1) The work of [Shea and Gray 1973; Klotz and Jiang 2017] showed that at first order, the asymmetry of the inner core is mainly due to the advection of the cyclone into the synoptic atmospheric field. Like Holland model, an azimuthal distribution can be applied to Vmax, which varies with the alignment of the vortex winds with the heading direction of the storm. This sinusoidal variation is based on the [Jelesnianski 1966] formula and only requires information on the TC displacement vector, which is available in the BTK. In contrast to Holland parametric model, we do not wish to apply the asymmetry of the motion speed to the rest of the profile, since we have other information for the outer core.
- 2) The skewness of tropical cyclones outside the eyewall is mainly governed by the large-scale atmosphere via shear processes and has an impact on all intensity categories and especially on the weaker ones. This asymmetry is already contained in the r17 satellite swath measurements. We can therefore extract the isotach of the gale-force wind from the different scenes.

Thus, for each of the acquisitions of our combined radiometric and scatterometric database, we have the azimuthal estimate of r17 as well as the BTK information interpolated at the time of sampling. We then derive for each azimuth a wind profile given by the parametric Chavas solution, with information on the TC motion speed, the heading direction, the azimuthal Vmax value and the associated r17 size. This process allows us to resolve the inner and outer core variabilities over the whole cyclonic vortex using the model interpolation method with the help of the 2D surface measurements of r17. In addition, we respect the dichotomy of the model, which dissociates the two regions of the cyclone, by giving it two distinct asymmetry regimes.

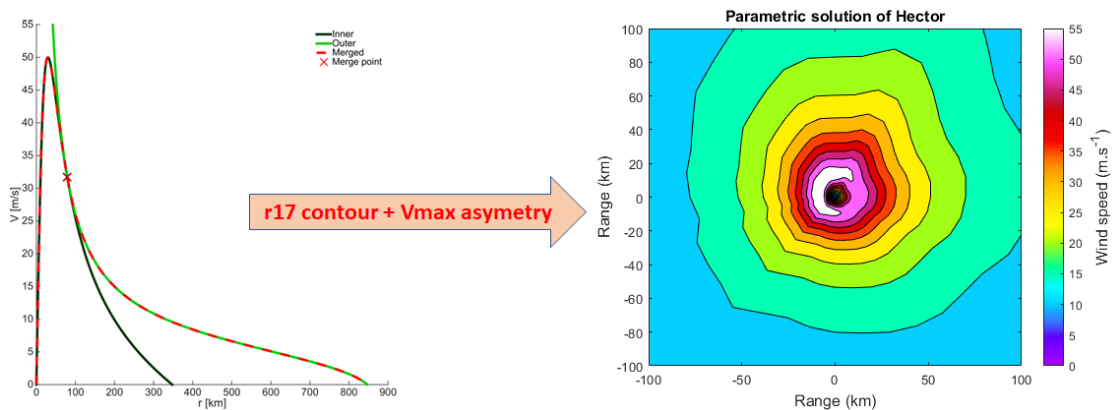


Figure 4.3: Transformation of the parametric solution from a 1D wind profile to a 2d wind field, with the example of Hector.

The figure 4.3 shows the model transformation with the example of the Hector wind field, which corresponds to the same run as presented in the introductory section. The

specific crescent shape of the eyewall reflects the typical asymmetric structure of the inner core of a relatively fast moving cyclone.

Now that we have a 2D wind field, we will estimate its performance by first confronting it with low and medium resolution microwave observations and BTK data, mainly on the outer core and for some inner core radii visible by radiometric instruments. In a second step, in order to test the capabilities of inferring the inner core from an outer radius, we will qualitatively compare the performances of the model to SAR observations and examine the reliability of the  $R_{\max}$  predicted by the model, first using the  $r_{17}$  measured by the SAR (to eliminate any resolution effect), then reusing the radiometer and scatterometer information, which will allow us to quantify the possible degradation of the model performances. Finally, we will quantitatively compare the full model wind profiles with the SAR observation, over all categories, in order to propose a validity frame for the model.

Although [Chavas and Lin 2016] has already made comparisons with observations, these were made on the integrated profile and with two very heterogeneous surface wind sources (a satellite product and a cross-platform product). Here, we want to evaluate for the first time the capabilities of a 2D version of the Chavas profile based on  $V_{\max}$  and  $r_{17}$ , compared to a homogeneous, highly resolved and complete surface wind source (SAR). In addition, we want to highlight the performance of the model over another historical parametric solution, the Holland vortex. To date, only a modified 2D Chavas model based on  $R_{\max}$  has benefited from such a complementary analysis with other parametric solutions [Wang et al. 2022].

## 4.2 Outer-core comparison.

### 4.2.1 Comparison with radiometers and scatterometers

We first compare different metrics of the surface wind field extracted from our combined database of radiometers and scatterometers, with those estimated by the Chavas model. Like we did with the MW observations in Chapter 2, we extract from the 2D parametric wind field the mean quadrant values for several wind radii, by averaging azimuthally the wind transects of each geographical sector (NE, NW, SW, SE). This results in a set of 4 values for  $r_{10}$  to  $r_{60}$ , which should allow us to check whether the model correctly reads the cyclone geometry. The two figures (Fig 4.4 and 4.5) gather these comparisons in a first instance for the outermost radii ( $r_{10}$ ,  $r_{12}$ ,  $r_{15}$ ) generally used to characterize the size, and the intermediate ones ( $r_{20}$ ,  $r_{25}$ , Fig 4.4); and then for the radii at the edge (depending on the intensity of the TC) or within the inner core ( $r_{30}$ ,  $r_{35}$ ,  $r_{40}$ ,  $r_{45}$ ,  $r_{50}$ ,  $r_{60}$ , Fig 4.5), which determine the strength of the cyclone. These results are notable for the prolific number of different observations and metrics used, they advocate once again the quality and complicity of the radiometer/scatterometer constellation to provide information on the cyclonic wind field. For both graphs, we have also normalised the mean values by the  $R_{\max}$  guessed from the parametric model, in order to provide an order of magnitude to the observed gaps.

Overall, the Chavas parametric solution faithfully reproduces the different radii, with a global correlation higher than 80% and a normalized bias lower than 10% until  $r_{40}$ . ASCAT data were removed from the study for winds higher than  $25 \text{ m.s}^{-1}$  for which we start to see a dropout, as expected, due to saturation issues. On the other hand, the parametric model is very much in phase with the different radiometers, in particular for those operating at 20 cm wavelength (SMAP and SMOS). In order to synthesize this



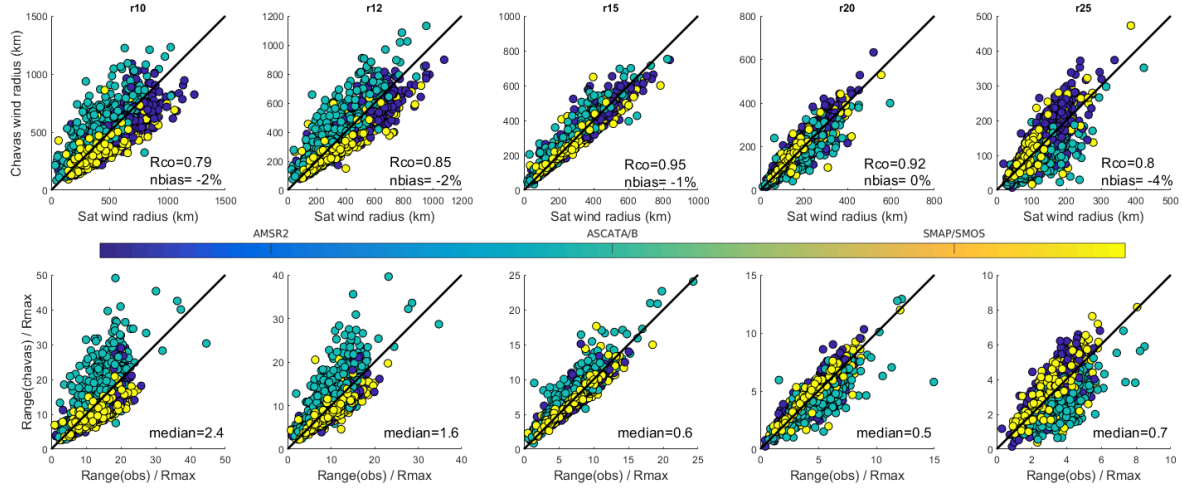


Figure 4.4: Comparison of outer wind radii derived from low-to-medium resolution MW instruments and the 2D parametric solution. Upper panels displays the range, while lower panels are normalized with the  $R_{max}$  given by Chavas model.

behavior, the two L-band passive sensors were grouped into a single label, given the similar performance reported in previous studies [Reul et al. 2016; Zhao et al. 2018] and that displayed against the parametric model.

Regarding the outermost section (Fig 4.4), the generated 2D field perfectly reads the variability of the different wind radii observed with the MW measurements, with almost zero normalized biases ( $<4\%$ ) and median deviations of the order of 1-2  $R_{max}$  ( $R_{max} \sim 30\text{km}$ ) that decline for increasing values of wind intensity up to  $r_{20}$ , visible on the lower panels. As the gale-force wind radius is ingested in the model to anticipate the behavior of the outer core, this trend is expected. These first comparisons demonstrate the model ability to capture the wind fringe where the scatterometers are extremely reliable in terms of wind inversion ( $12\text{-}18 \text{ m.s}^{-1}$ ) [Chou et al. 2013]. It is therefore comforting to see that the best results are observed with ASCAT instruments. The values of the four geographical quadrants are represented, so the parametric solution correctly interprets the asymmetry of the external field. This is quite conclusive on the interest of using a size parameter to guide the model in this region of the TC, whose geometry is more subject to the influences of the large-scale atmosphere (shear) [Chavas and Lin 2016; Klotz and Jiang 2017], and thus difficult to interpret in a  $R_{max}$ -based parametric model that often misread the outer wind profile, like Holland one [Willoughby and Rahn 2004]. Finally, the lower panels also show that the model captures the full range of values, including the largest systems.

As the inner core is approached (Fig 4.5), the performance of the model is maintained with a correlation of around 80% or even 90% for  $r_{30}$ ,  $r_{35}$  and  $r_{40}$ . The exacerbated sensitivity of the radiometers at these intensity levels makes these comparisons particularly relevant [Reul et al. 2016]. If the median deviation progressively increase up to 1  $R_{max}$ , the model performs well up to  $r_{60}$ , with a bias that never exceeds 30% overestimation, and remains relatively low up to  $r_{40}$ . These six additional metrics allow us to state that the model accurately interprets the more edge-of-the-wall portion. Although MW radiometers provide an excellent reading of the foam emissivity under extreme winds condition [Reul et al. 2012], we have stopped at the radii  $r_{60}$ , since the above intensities are too difficult to interpret by the more coarsely resolved measurements of these instruments. However, the Figure C.3 from Annex C shows the ability of radiometric instruments to measure coherent winds up to  $80 \text{ m.s}^{-1}$  compared to BTK and SAR measurements, as

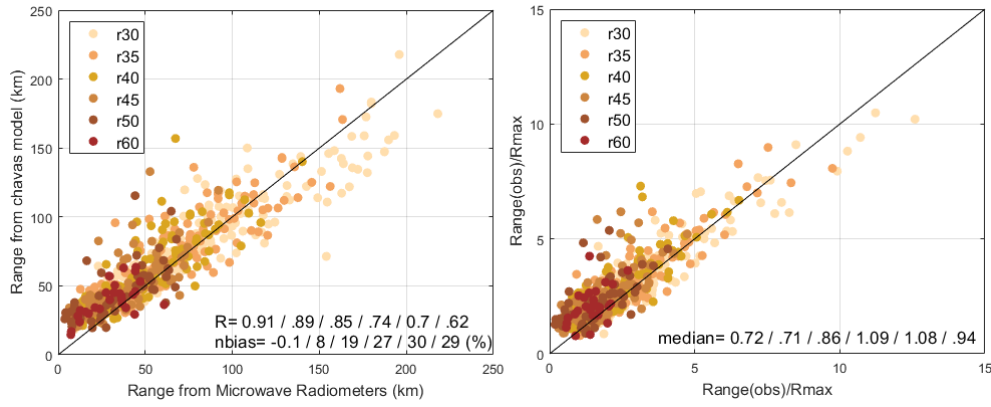


Figure 4.5: *Scatter plot of cyclonic wind radii from r30 to r60 m.s-1. Only data from the SMOS, SMAP and AMSR-2 radiometers were used. Correlations and normalized biases follow the increasing order of the wind radii intensity.*

previously demonstrated [Reul et al. 2017], notably with SFMR measurements [Meissner et al. 2017]. Some of the observed scatter of R30/35 may be attributed to the resolution of the measurement, however the appearance of a positive bias ( $>5\%$ ) for r35, suggests instead a slight overestimation of the width of the high wind area. This tendency is probably inherent to the nature of the model, where this type of bias has been observed in studies devoted to the model [Chavas et al. 2015], especially in the eyewall before the merging area between the two solutions ( $\sim 2R_{\max}$ ) [Chavas et al. 2015], where an overestimation of winds was observed ( $2 \text{ m.s}^{-1}$ ). The overestimated wind gradients must be interdependent on the cyclone intensity, since the intersection depends on both the  $V_{\max}$  information and the wind radii. In this sense, we observe a disappearance of this positive bias for R35 for more intense cyclones, when we compute for Cat-1 to Cat-5 only (not shown here). In order to establish a reliability framework for the model, high resolution SAR analyses will confirm these results in section 4.3. Finally, the correlations displayed between the model and the observations are very consistent with the observed values of the comparisons made between the radiometers and the BTK and/or SFMR data of other studies ( $\sim 80$  and  $70\%$ ) for a similar intensity threshold [Zhang et al. 2021], which indirectly proves the good orientation of the outer radius information on the profile behaviour.

### 4.2.2 Comparison of the three wind radii with BTK

Finally, to finish with the analyses of the outer and near-inner core profile, we also wished to confront these estimates with the BTK data (Fig 4.6), which integrates this type of instruments in its procedure, particularly for R17 and R25 [Bushnell et al. 2018]. These comparisons are to be put in the light of those made previously between high and medium/low resolution MW measurements with BTK data (Chap 2/ 3).

Indeed, the same observations can be made for the first two metrics (r17/r25 in Fig 4.6), with good similarities between the two sets of values and scattering of the order of 30-40% (as observed with SAR in Chapters 3 for r17), and of the same order of magnitude as with the satellite data (Chapter 2). The cloudiness of the data is slightly higher than that observed with the SAR, as we did not do conduct the fine filtering as those done for Combot et al. [2020a]. Nevertheless, this again indicates that the model correctly reconstructs the profile with very little degradation of the wind radius estimates.

Regarding r33, we note larger deviations and a significant positive bias, which also ex-

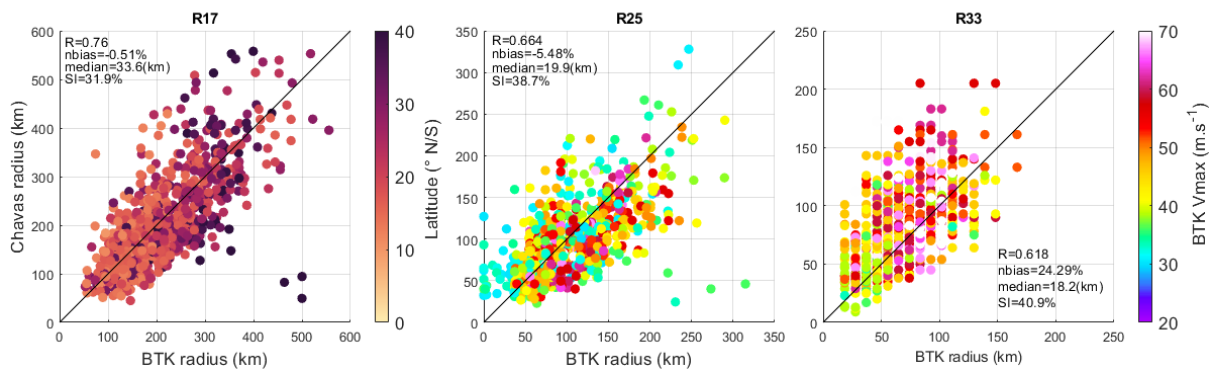


Figure 4.6: Comparison of the three BTK wind radii ( $R17, R25, R33$ ) with the predicted values from the model. The colorbar of  $r17$  is a function of latitude, while that of  $r25$  and  $r33$  is set according to the  $V_{max}$  information from BTK.

isted in the comparison with the SAR data but in a lesser extent ( $\sim 10\%$ ). This result is related to three factors. 1) The high uncertainty associated with this parameter in the BTK [Landsea and Franklin 2013]. As discussed in Section 1.4, the hurricane-force winds quality is the most erratic of the three wind radii, with larger errors observed. 2) The reduced dependence of MW instruments in the BTK analysis of  $r33$  [Herndon et al. 2012; Knaff et al. 2021], which may induce larger uncertainties due to deeper reliance on indirect methods and their limitations (resolution...). This point can be verified by the systematic observation of a positive bias in  $r33$  found by all MW observations, all resolutions included (see also Fig C.2 annex C). 3) The overestimation linked to the model, which we have already observed in Figure 4.5 and which must be added to the scattering already present in the data, which could explain the larger differences obtained with the estimated values.

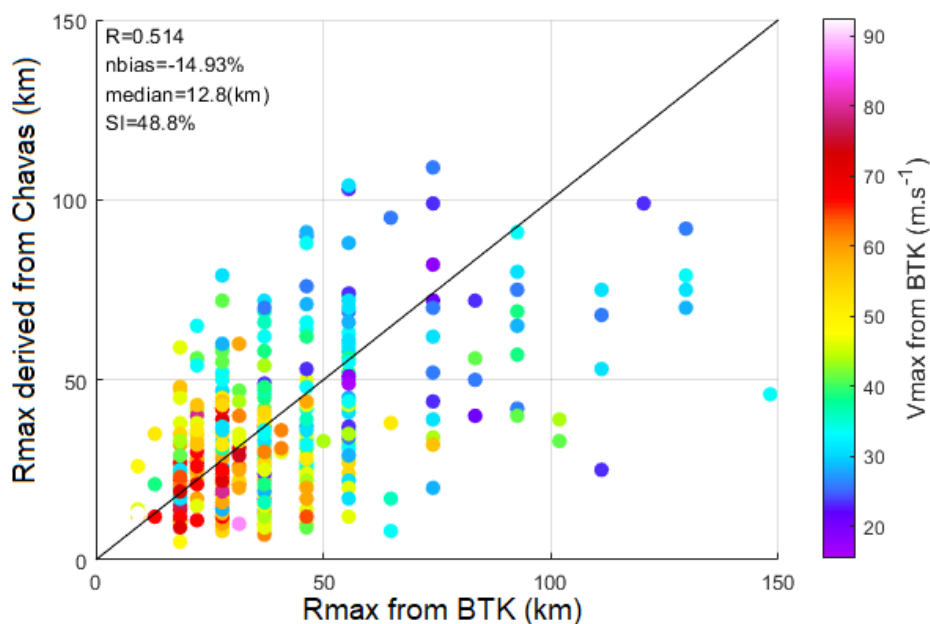


Figure 4.7: Distribution of the  $R_{max}$  given by BTK and Chavas model.

We also compared the values of  $R_{max}$  (Fig 4.7), even though it is the most conflicting

parameter of the operational products, with the largest errors [Combot et al. 2020a]. However, this comparison raises an interesting point. We can note a great agreement between the differences in this figure, and those already observed in Chapter 3 between the  $R_{\max}$  derived from the SAR and the BTK. We observe the same bias around 15%, which underlines the overestimation of the values asserted by the BTK. This similarity may suggest a better fit between the predicted values and those obtained by the high resolution measurements. In order to verify this hypothesis, we will now compare the Chavas wind profiles with those of the SAR.

## 4.3 Inner core comparison

The SAR analysis performed in Chapter 3 provided us with a substantial database of approximately 100 highly resolved cyclonic wind field, representative of a wide range of geometric and geophysical properties. The knowledge of the inner core structure allows us to access the complete wind profile for each azimuth within the spatial sampling limit of the SAR instrument. We will now exploit this advantage to ensure the quality of the model over the entire profile, and more specifically for winds above the cyclonic threshold. To highlight the possibilities of inferring the inner core, in this section we first use the  $r_{17}$  extracted from the SAR data to force the model, in order to get rid of the potential errors associated with the scattering/resolution of the low/medium resolution MW data. In a second step, we again use the radiometric and diffusometric measurements to inform the  $r_{17}$ , and observe changes in the model predictions.

### 4.3.1 Case study

We will start to illustrate the qualities of the model and present its limitations through a qualitative analysis of four case studies, selected according to specific criteria. 1) Intensity. The previous analysis quickly showed us an asymmetry in the model performance depending on the wind category considered. We therefore selected two major cyclones (cat 4 and 5), a category 2 and a tropical storm. We focused of course on the intense cyclones, for which salient errors were found when estimating  $R_{\max}$  with BTK (Chap 3). 2) Inner core geometry. Since the outer core is guided by a wind radius, it is the variability of the inner core that we will try to capture here, especially since the geometry of the eye and the eye wall are indicators of potential changes in cyclone structure or intensity [Musgrave et al. 2012; Kossin and Sitkowski 2012; Li et al. 2013]. Of the four cases we have chosen, two experienced strong asymmetries in their azimuthal  $V_{\max}$  distributions (Fig 4.8 & 4.9) and are associated with fast translation speeds ( $V_{\text{fm}} > 6 \text{ m.s}^{-1}$ ), while the other two have milder variation (Fig 4.10 & 4.11) and slow advection ( $V_{\text{fm}} < 2 \text{ m.s}^{-1}$ ). 3)  $R_{\max}$  size. Chapter 3 revealed the vagaries of BTK in estimating  $R_{\max}$  from narrow inner core, or from structures that deviate from the standard values.

- **Hector 2018:**

Of course, we could not ignore the case of Hector, whose one wind transect has already been introduced in section 1.4 and 1.5. This midget cyclone perfectly embodies the current shortcomings of the different analysis and forecasting products to correctly asses the  $R_{\max}$ , due to a lack of resolution from the retrieval methods[Klotz and Jiang 2017]. We want to see if the parametric solution of Chavas can infer the inner core parameter from the  $V_{\max}$  and  $r_{17}$  information. The Figure 4.8 depicts the wind profiles of Hector at azimuths  $45^\circ$ ,  $135^\circ$ ,  $225^\circ$ , and  $315^\circ$ , to convey the geometry of the four geographical quadrants. In

addition to the radial wind distribution derived from SAR and Chavas, we incorporated wind radius information from BTK, although they represent the maximum extent values of wind radii for each quadrant. The associated 2D wind fields are presented in the Figure 4.9 via the top panels a b and c. They display respectively the high resolution TC portrayal from SAR, the one predicted by the model, and the residual errors coming from the difference between these two wind fields (following convention: SAR-Chavas). The other three cases are illustrated in the same way. This passage of Hector was misinterpreted by the BTK analysis, with an Rmax greater than 40 km, while it does not even exceed 20 km (see Chap 3), we also notice here, a much broader behavior for the different BTK wind radii. The model on the other hand captures remarkably well the different profiles for each of the quadrants and predicts identical Rmax with ( $\sim 1/2$ km accuracy). The geometry of the cyclone is well respected, as can be seen on the Figure 4.9, where the same extents are observed for the different contours of similar magnitude. For each of the 2D fields, we have drawn the same intensity contours between the panels a) and b). The model thus provides a smoothed version of the observed wind field and respects its symmetry degree. The residual errors are very small as compared to the storm intensity, with a median deviation around  $2 \text{ m.s}^{-1}$ , and a dispersion mainly contained in a range between  $\pm 4 \text{ m.s}^{-1}$  (mean+ $\sigma$ ), for the regions outside the eye. On the contrary, salient ambiguities stand out within the eye, due to the very sharp gradient from the parametric solution of the ascending wind profile located between the TC center and the Rmax (Fig 4.8). Existing winds around  $10 \text{ m.s}^{-1}$  (see Chapter 3) are almost systematically observed in the SAR images Li et al. [2013], which contrast with the sharp drop to zero values described from the parametric solutions. There is a recurring tendency from the Chavas solution to provide narrow eyes [Chavas et al. 2015; Chavas and Lin 2016] and thus logically widen the eye wall extent. This limitation has generally little impact here for our study, since this region is generally very confined especially for intense cases, the influence from these local errors on the total energy displayed by the cyclone forcing is reduced [Powell and Reinhold 2007]. Furthermore, since we are interested in the oceanic response, the variabilities of the wake anomalies are mainly sensitive and driven by the actual range of maximum forcing and the size of the wind profile [Ginis 2002; Kudryavtsev et al. 2019b]. We are therefore primarily interested in validating the outer core up to Rmax. The errors associated with the eye, although drastic, have a small impact on the model performance, as shown by the evolution of the statistical parameters when the whole structure is included in the calculation of the median and mean errors (Fig 4.9 c).

- **Jose 2017:**

Jose was a major cyclone of 2017, evolving in conjunction with Hurricanes Irma and Maria during a very active season for the Atlantic. It has a particularly strong inner core asymmetry (Fig 4.8) between the mean intensity of the four quadrants ( $\sim 10 \text{ m.s}^{-1}$  oscillations), perfectly captured once again by the model. However, as with Hector, there is a slight global overestimation at the top of the curve, related both to the faulty structure of the eye that widen the strong winds area, but also probably to the definition of the Vmax asymmetry, as we see a wider maximum wind contour given by the model. For this situation, there is a good match between the BTK wind radii and those given by the SAR and model. As for Hector, the 2D field geometry is globally well captured, with the exception of the rainband visible in the NW quadrant wind profile (Fig 4.8), where we observe an underestimation of the model for winds between 50-100 km, which corresponds to the wider extension of the  $25 \text{ m.s}^{-1}$  contour (Fig 4.9 d), visible through the positive anomalies in panel f of Figure 4.9. The 2D wind profiles of the two TCs remain extremely well anticipated, with a median deviation always close to  $2 \text{ m.s}^{-1}$ .



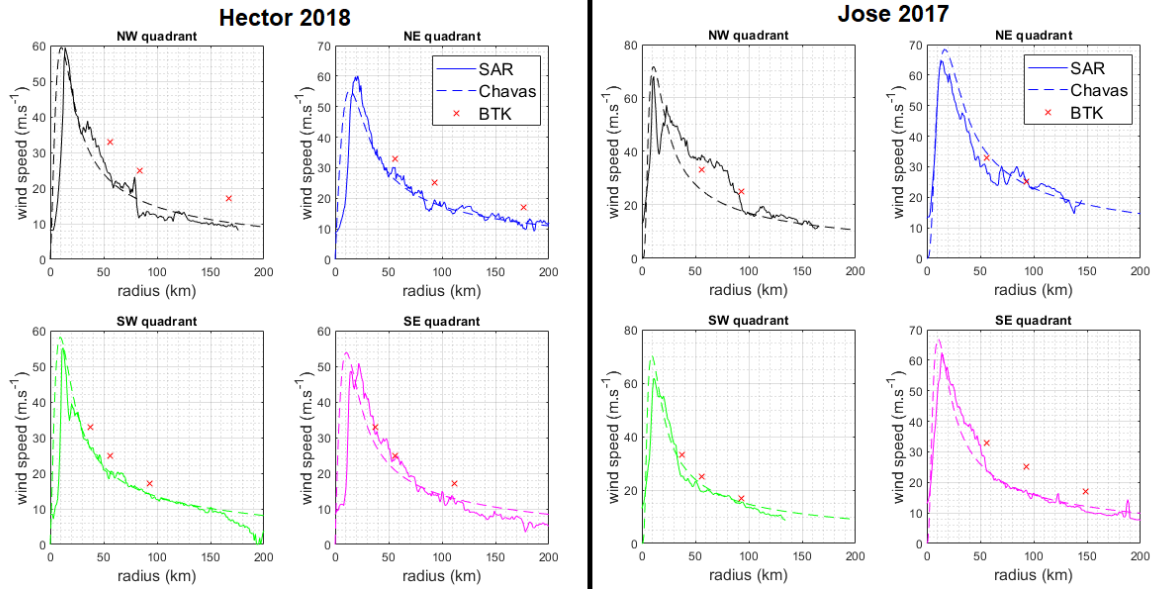


Figure 4.8: Analysis of the radial wind profile of Hector (left side) and Jose (right side) derived from SAR measurements (solid line) and from the model (dashed line). The four geographic quadrants NW, NE, SW and SE are represented according to transect from  $135^\circ, 45^\circ, 225^\circ$  and  $315^\circ$  azimuth. Complementary information of the BTK wind radii are added for each quadrant (red crosses).

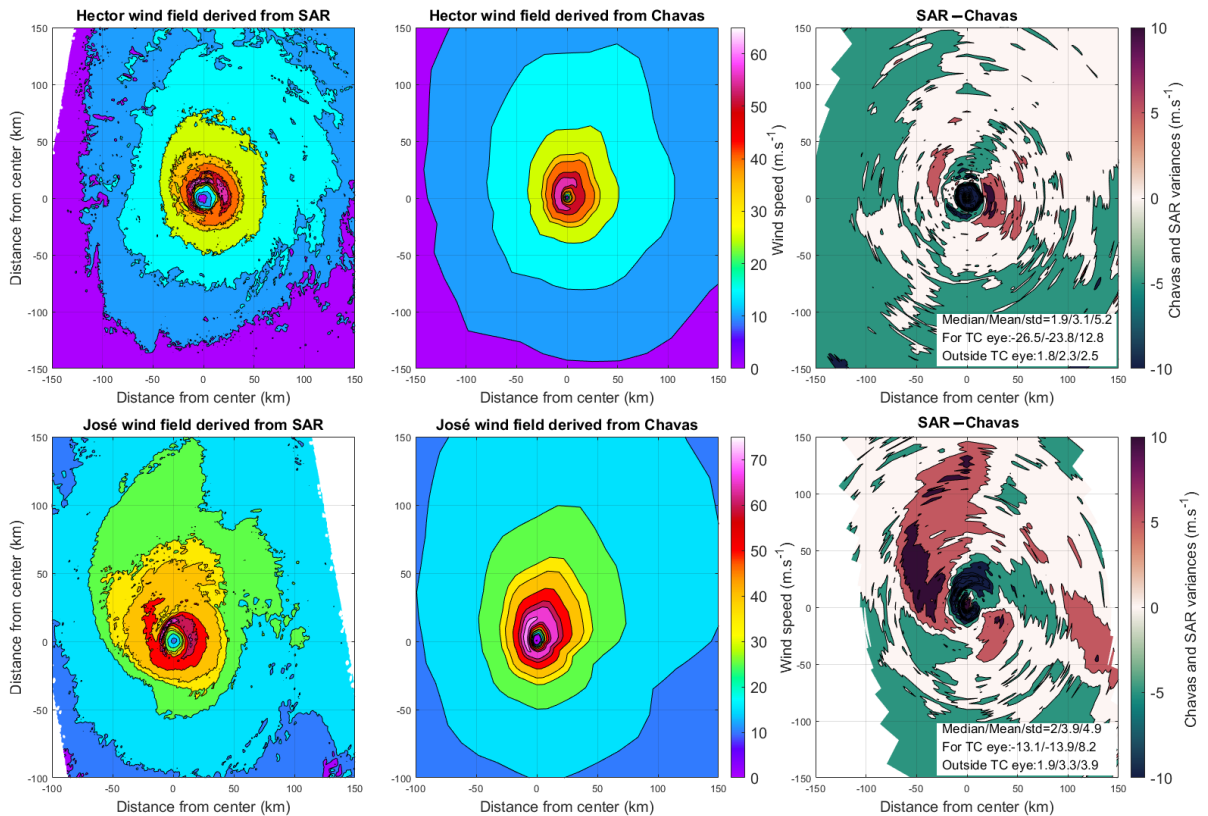


Figure 4.9: View of the 2D wind field of Hector (upper panels) and Jose (lower panels) got from SAR measurements (left) and from the model (middle). Residual errors are derived from the difference between SAR and Chavas wind field (right).

- **Trami 2018:**

Trami was a major typhoon of 2018, which roamed the very warm waters of the China Sea and featured an asymmetric triangular-type eye [Li et al. 2013] with an excessively large Rmax (80km). This type of cyclone is likely to generate very deep oceanic responses [Ginis 2002] and dithyrambic sea states [Kudryavtsev et al. 2015], due to its large intense wind extent. So far, the model has captured the subtleties of the cyclonic wind field perfectly, but here we see some of the apparent limitations of the model. While the hyperbolic decay is very well anticipated in all quadrants, and in agreement with the BTK values, the position of Rmax is clearly underestimated ( $\sim 20$  km). Such a large extension is probably not expected by the parametric solution and represents a physical limitation. Indeed, the value of the Rmax of Trami is located within 3 times the standard deviation of the Rmax normally measured for this intensity ( $37 \pm 12$  km for  $v_{\max} = 40 \text{ m.s}^{-1}$ <sup>4</sup>). We see through panel c of Figure 4.11, significant errors extended in a wide area going from TC center to Rmax. However, we note the excellent fit of the rest of the cyclonic wind field with deviation close to zero. Despite the large discrepancies around the central region, we observe a total median deviation that remains relatively small and at  $3.2 \text{ m.s}^{-1}$ . The model still manages to represent a significant part of the wind field, but with errors around the eye that are no longer negligible.

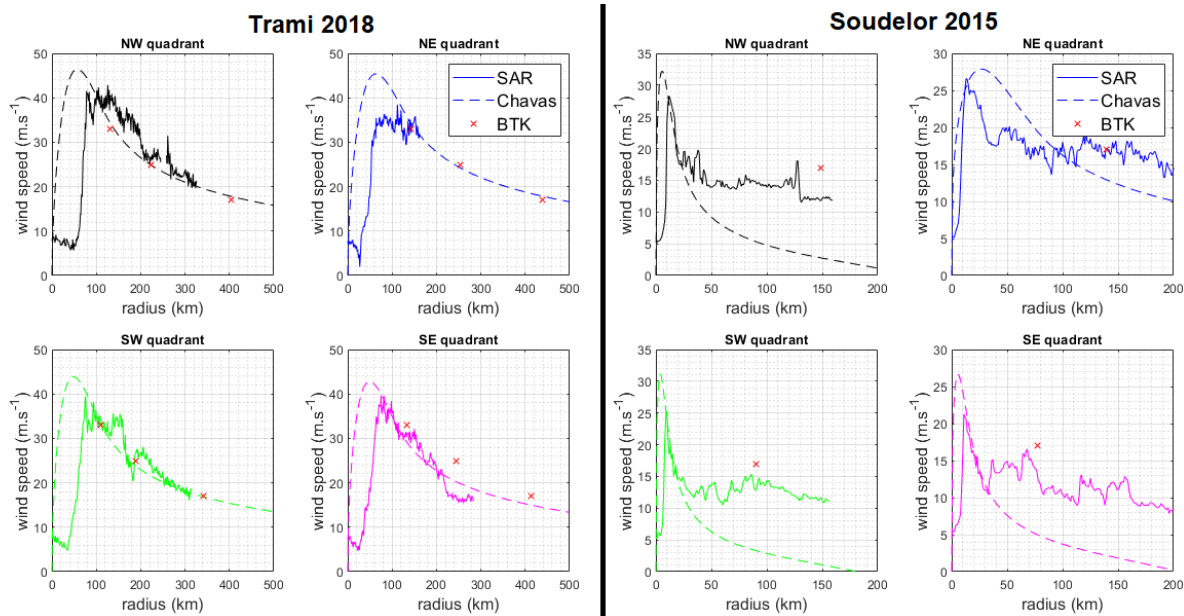


Figure 4.10: Same as Fig 4.8 but for Trami and Soudelor cases.

Finally, Soudelor is a major typhoon that crossed the western Pacific zonally until reaching the Asian shores. At the time of the satellite acquisition, the tropical storm is just beginning to intensify, but its circulation is not yet fully organized, nor fully-developed into a deep convective system that surrounds the low-pressure center in the shape of an eyewall and concentric outer bands. An evident strong asymmetry emanates from its cloud cluster (see Fig 4.11 d) and the whole cyclonic field seems to be still greatly under the influence of the large-scale atmosphere. These two statements are outside the physical paradigms of the model, and can explain the global rout of the parameterized wind field. This is the only one of the four examples to show stronger errors outside the eye area, and

<sup>4</sup>extracted from our SAR database

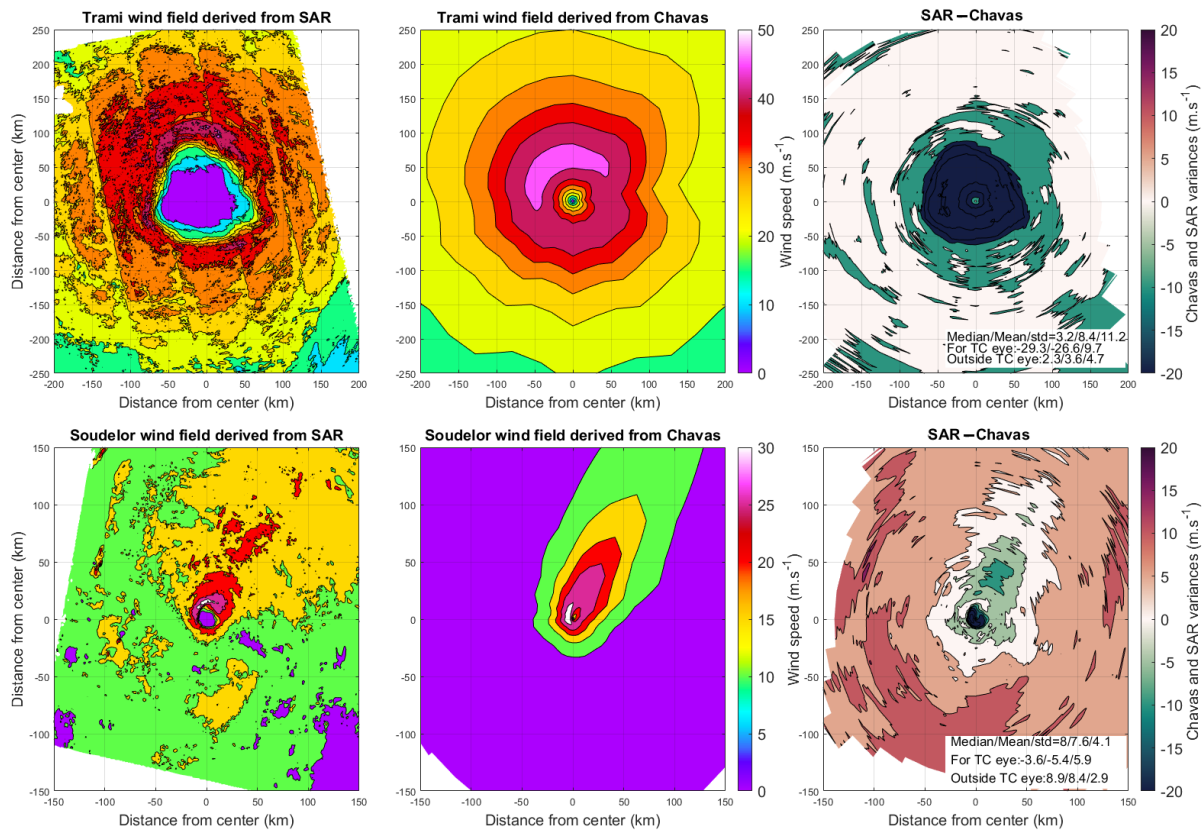


Figure 4.11: Same as Fig 4.9 but for Trami and Soudelor cases.

thus over a much wider area. The strong asymmetry existing in the northwest quadrant with a full-width r15 contour is not reproduced by the model (Fig 4.11). The hyperbolic wind decline predicted by the parametric solution is indeed countered by a plateau around 10-15 m.s<sup>-1</sup> for each quadrant Fig 4.10. This divergent case may raise the question of what external radius to incorporate into the parametric solution for weak systems. The r12/r15 have already been used as a metric to determine the size of the forcing, and may limit some of the observed discrepancies, although this broad, disorganized circulation and the outer wind plateau can not be anticipated by the mechanisms governing both the inner core and outer core (see section 1.5).

These four studies qualitatively demonstrate the model capability to incorporate the different asymmetries and variability of the cyclonic wind field, at least for systems that are organized or have exceeded the cyclonic threshold and do not have an inner core structure several times larger than the standard values. This last point must be balanced, however, as the predicted wind field nevertheless manages to capture the rest of the profile.

### 4.3.2 Rmax and asymmetry

To assess the global performance of the model to retrieve the Rmax, we then performed a more quantitative experiment by comparing the Rmax values advocated by the SAR and those estimated by the model. This is shown in the two-panels Figure 4.12, which respectively shows the comparison of the effective Rmax, associated with the maximum wind speed, and the value averaged over all azimuths. This double analysis allows us to verify if the model captures both the location of the most extreme winds, but also integrates its azimuthal distribution. The Figure 4.12a summarizes the results already

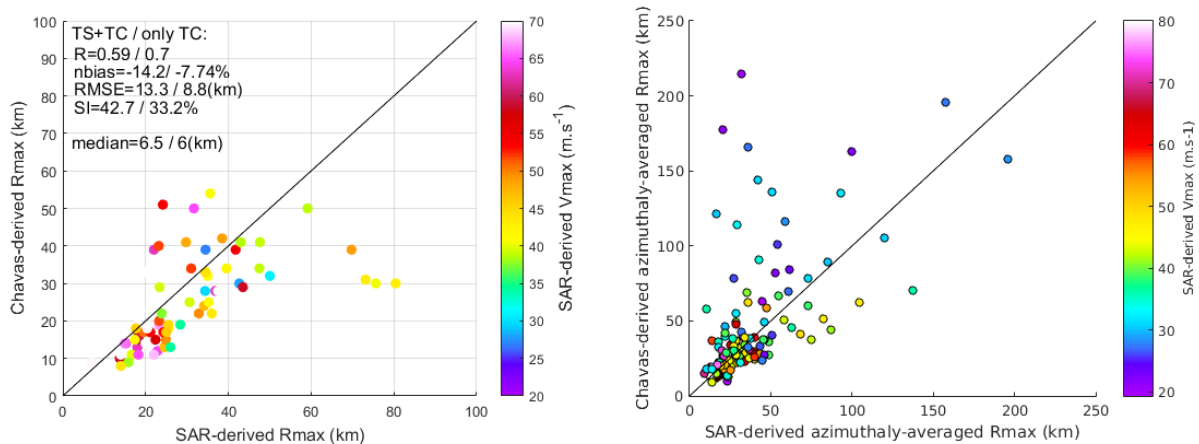


Figure 4.12: a) Comparison between SAR-derived and Chavas-derived  $R_{max}$ . b) The right panel represents the mean  $R_{max}$  from the azimuthally-averaged wind profile.

observed by our qualitative analysis, with a better agreement of the model for systems that have crossed the cyclonic threshold (according to statistic indicators). The performances for hurricane-force cases are much better than those achieved with BTK (Figure 7 of [Combot et al. 2020b]), with a median deviation of 6km and an RMSE of 8km, but with a slight negative normalized bias of 8 %. The  $R_{max}$  asymmetry is perfectly integrated by the 2D parametric solution for hurricane cases (Fig 4.12 b), with values very close to those averaged azimuthally from SAR data. The gap is even more striking in the Figure 4.12b between tropical storms and higher categories, where we see very marginal mean- $R_{max}$  values for TS, unlike hurricane-force cases. This means that the effects conveyed by the information of the translation speed, and by the  $r_{17}$  (as it modulates the merging point [Chavas and Lin 2016]) in the model allow a good reading of the asymmetries, as foreshadowed by the old studies on the domination, at 1st order, of  $V_{fm}$  on the inner core symmetry properties [Jelesnianski 1966; Shea and Gray 1973; Klotz and Uhlhorn 2014; Klotz and Jiang 2017]. On the other hand, tropical storm asymmetries are induced by other large-scale atmospheric mechanisms, such as vertical shear, which has been shown to have a significant influence on the overall wind radii by Klotz and Jiang [2017] and are not taken into account for the inner solution of the parametric model.

Beyond a distance of 60 km, model estimates of  $R_{max}$  are generally erroneous and most often underestimated (Fig 4.12a). This physical limitation has already been mentioned for similar range by Chavas and Lin [2016], via comparisons with high resolution HWIND analyses. Our study come to confirm this result thanks to SAR observations. The reasons for this systematic underestimation stem from the variability mode chosen with the solution forced by an external radius. Since the wind field dynamics in the model can be essentially phrased in terms of the angular momentum ratio between the inner and outer core ( $M_m/M_0$ ), the  $R_{max}$  parameter is originally inferred from a set of three variables:  $f$ ,  $V_{max}$ , and  $r_{fit}$  ( $r_{fit}$  being here our outer radius information). Because of this ratio, the sensitivity of  $R_{max}$  appears to be much more sensitive to variations in  $R_{fit}$  than  $V_{max}$  ( $R_{max} \propto f r_{fit}^2 / V_{max}$ , Chavas and Lin [2016]), which is particularly relevant for capturing the inter-storm variability of cyclones, but has some limitations when assessing intra-storm variability sensitive to the intensity fluctuations and the internal structural changes [Kossin and Sitkowski 2012]. This imposed trend in  $R_{max}$  remains physically consistent, given the inversely proportional relationship between the two inner core parameters [Quiring et al. 2011], with intensifying systems generally followed by eye-



wall contraction <sup>5</sup>, as we also saw with the SAR data in the Figure 9 in Chapter 3 . This also remains statistically consistent given the weak representativeness of these large and intense cases in our TC sampling.

Still with a view to generalisation, we now want to check how these performances on the prediction of Rmax are maintained through the use of r17 from radiometric and scatterometric observations. We are therefore looking for possible matches with SAR data, within a window of more than  $\pm 12$ h. This new step demonstrates once again the strength of the current constellation of instruments, with 80% of our SAR acquisitions having a successful colocalisation. For 60% of them, the time difference is even less than 2-3 h. They are almost exclusively obtained from radiometers, with SMAP and SMOS leading these observations. We have retained only those situations where the intensity was greater than  $20 \text{ m.s}^{-1}$  and where the tropical systems are far from the coast, to ease the extraction of r17. The coarser resolution of the instruments makes them more sensitive to the noise generated by the approach to a coastal area. In our SAR sample, more than half of the cases are retained in total ( $\sim 90$  cases), including most of our major cyclones.

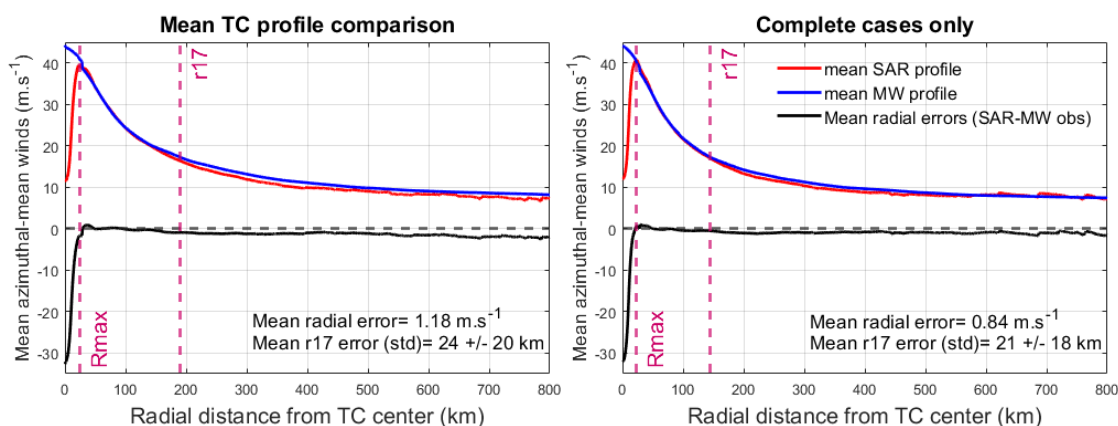


Figure 4.13: a) Mean azimuthally-averaged wind profile derived from 70 cyclonic passages ( $> 33 \text{ m.s}^{-1}$ ) captured by SAR instruments (red) and from colocated MW radiometer measurements (blue). b) Right panel displays only cases for which the r17 is almost complete in the SAR scene ( $> 75\%$  of azimuth).

Before examining the prediction of Rmax induced by the lower resolution observations, we compare the full radiometer <sup>6</sup> wind fields with those from the SAR to investigate the possible discrepancies. The 2D surface vortex from the satellites are azimuthally-averaged to give an integrated profile. The figure 4.13 shows the mean 1D profile averaged over all cases for the SAR in red, and for the radiometers in blue, the mean radial differences between each of the SAR and radiometer 1D profiles is shown in the black curve. Overall it shows the excellent correspondence between these two types of measurements, with on average a very similar behavior from Rmax to the cyclone tail (median radial error  $\sim 1 \text{ m.s}^{-1}$ ). The median offset of r17 is of the order of 20 km, which represents an uncertainty of the order of 10%, i.e a lower value than that assigned to the BTK. Quite remarkably, the radiometers manage to capture the gradient of the strongest winds up to Rmax, even though the eye structure is not captured and induces an increase of the mean winds in the area before the eyewall. These results are in even better agreement when the observed

<sup>5</sup>as predict by partial angular conservation theory

<sup>6</sup>the seldom colocated-ASCAT data are removed from this comparison due to saturation issues



wind fields are complete for both measurements (panel b).

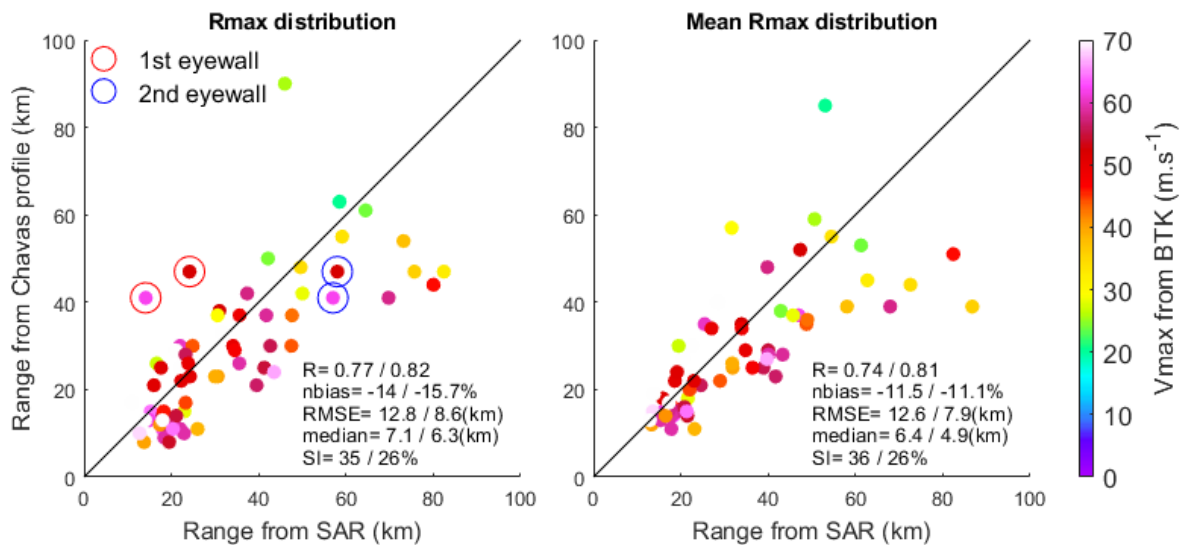


Figure 4.14: Same as Figure 4.12. Distribution of the a)  $R_{max}$  and the b) mean  $R_{max}$  values from SAR and derived from Chavas model, but with  $r17$  information from low and medium resolution satellites. .

Now that we have discerned the degree of dispersion between the two measures, we finally compare the  $R_{max}$  predicted by the model with the  $r17$  derived from classical MW observations and those estimated by the SAR in figure 4.14. Not surprisingly, the results are very similar to those observed in the Figure 4.12, given the very small differences in the radius estimates. The better statistical results obtained for all categories (with a correlation of 0.77) must be balanced by the fact that some of our weaker cases do not have concurrent acquisitions with the SAR. When we isolate the cases above the cyclonic intensity threshold, we find exactly the same performances with a median value around 6/5 km, depending on whether we consider the real  $R_{max}$  or the mean  $R_{max}$  of the 1D profile. These estimates are therefore of better accuracy than those given in the global BTKs, but are found similar to the estimates benefiting from the SFMR data (Table 1, [Combot et al. 2020a]). Two particular and limiting situations of double  $R_{max}$  associated with the eyewall replacement cycle can also be seen. The values circled in red represent the primary eyewall observed by SAR, which is closest to the eye center, while the blue circles indicate the distance to the secondary eyewall. Our predictions are found closer to the second maxima, as the  $r17$  extracted our observations were associated with the furthest eyewall. This kind of internal forcing cannot be anticipated by the model and are further removed from the analysis. As in the Figure 4.12, there is also slight underestimation of the  $R_{max}$  guessed by the model especially for small system (10% bias), and a loss of sensitivity for radii greater than 60 km.

Looking at our entire database of tropical cyclone cases, the period 2010-2018 displays about 619 tropical systems and include 400 tropical cyclones (Cat-1 to Cat-5), for only 12% of them experiencing at least once in their course a maximum wind radius larger than 60 km. If we look at all the six-hour positions given by the BTK, we can draw several remarks:

- 40% have a  $R_{max} > 60\text{km}$ , when we consider every wind categories ( $TD + TS + TC$ ).

- This percentage falls to 10%, when we only look at systems stronger than  $28 \text{ m.s}^{-1}$  (average intensity threshold of a well-defined eye appearance [Vigh et al. 2012]).
- Less than 5% of the tropical systems higher than  $33 \text{ m.s}^{-1}$  experience such large  $R_{\text{max}}$ .

The distribution of  $R_{\text{max}}$  indicates that the Chavas model would therefore be able to reproduce the vast majority of category 1 to 5 cyclones, and that this physical shortcoming is mainly imputable to tropical storms and depressions. For the weakest of them, the definition of an  $R_{\text{max}}$  can be ambiguous, as the circulation is not properly established, and does not display a properly defined concentric eyewall or eye [Vigh et al. 2012; Li et al. 2013]. These cases are inherently outside the validity of the model assumptions. For tropical storms evolving above  $25\text{-}28 \text{ m.s}^{-1}$ , the results seem to be more contrasted with situations much more in agreement with the SAR measurements (see Fig 4.12).

## 4.4 Complete wind field model: comparison and scope.

### 4.4.1 Global performances of the model.

We will now take advantage of the capability of the SAR to derive the full radial profile, in order to quantitatively assess the quality and reliability of the 2D model on the different sections of the TC, for the different wind categories. The Figures 4.15, 4.16, are showing respectively the final analyses between the mean 1D profiles of the SAR (black), the Chavas model (blue) and the Holland model (red), for the whole database and split by intensity; as well as the associated residual values for the global mean profile and for the six categories. The figures 4.17 and 4.18 are similar but normalise the radial distribution by the maximum wind radius. We have chosen to add the comparison of the Holland model to the study, in order to highlight the advantages of this new type of model based on an outer radii. For information, the residual curves are the average differences between each 1D SAR profile and those of the two parametric solutions.

From these four figures, several assertions can be made. First, from the comparison of the direct profiles, we can state that:

- The Chavas model reproduces very well the behaviour of the cyclonic vortex from the  $R_{\text{max}}$  region to the model boundary for all intensity categories from 1 to 5, with residuals always below  $5 \text{ m.s}^{-1}$  and most often contained in an interval around  $\pm 2 \text{ m.s}^{-1}$ . Hurricane-force system at the margin of the cyclonic spectrum (i.e. Cat 1 and 4-5) show the best agreement along the radial distribution.
- The Chavas model overlaps with the observations globally around 100 km from the centre and matches almost perfectly the gradient described by the SAR for all categories, except tropical storms.
- The eye definition is always underestimated, resulting in narrow eyes followed by a too sharp gradient of the ascending winds up to  $R_{\text{max}}$ . This shortcoming may stem from the definition of  $C_k/C_d$  responsible for the sharpening of the inner model solution at increasing  $V_{\text{max}}$  [Chavas and Lin 2016].
- A slight overestimation of the wind gradients after the  $R_{\text{max}}$  also exists and impacts every intensity category, in a range of 50-100km centred around the mean value of the merge radius ( $R_a$ ).

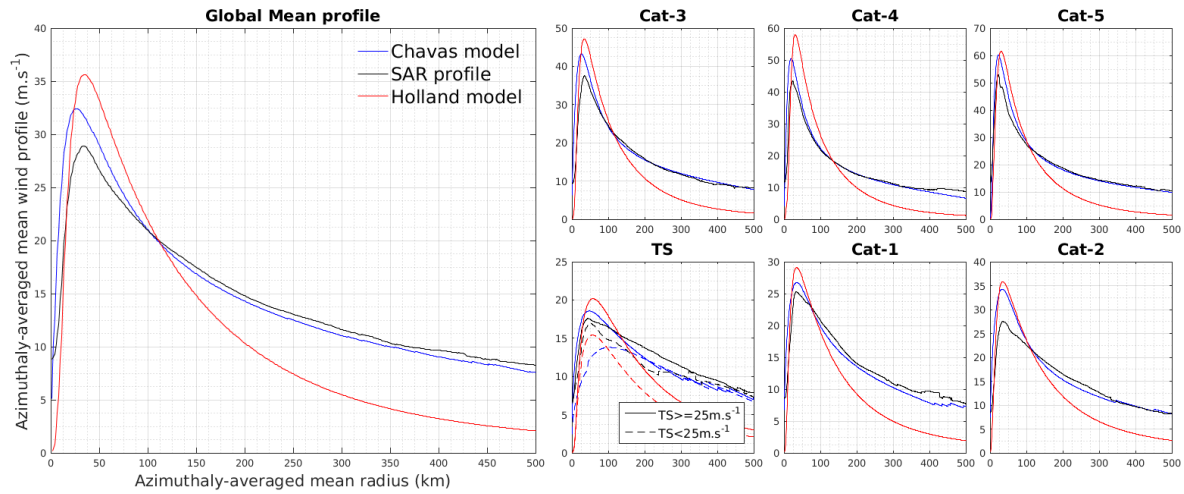


Figure 4.15: Comparison between the global azimuthally-averaged mean wind profile derived from SAR database with the ones predicted by Chavas and Holland model. The azimuthally-averaged mean wind profiles of each TC wind category are also displayed from Tropical storm to Cat-5 cyclones.

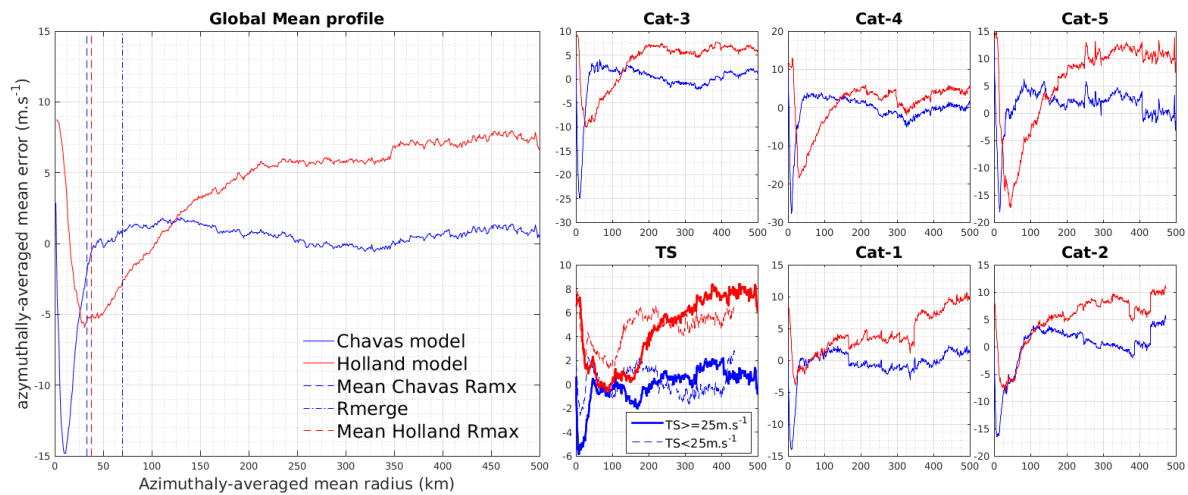


Figure 4.16: Radial distribution of the azimuthally-averaged mean errors (SAR-models) computed from the difference between SAR profile and the two models. Mean errors profiles of each TC wind category are also shown.

- Tropical storms show, as expected, more contrasted results, with drastic differences in the inner core for the weakest systems, especially in the position of the mean Rmax, and more moderate differences for systems at the limit of the cyclonic intensity threshold, as their circulation almost corresponds to the description of a vortex organised around a deep convection centre.
- The agreement of the integrated profiles (SAR & Chavas) suggests that the description of the geometry given by the r17 perfectly encompasses the model asymmetries along the outer core. However, the asymmetry dictated by the translation speed does not capture all azimuthal variations in intensity, resulting in a slight to moderate overestimation of the strong wind areas for both models. The largest differences

appear for intermediate categories. As a reminder, we mentioned in Chapter 3 that it is for these categories that the fastest evolution have been observed in our database.

- While the Vmax asymmetry is dictated for both models by the translation speed, the exacerbated asymmetries of the Holland model for the eyewall may stem from the uniform location of the azimuthal maximum winds. In the case of the Chavas 2D model, although the asymmetry advected by the cyclone displacement is a simple modulation of Vmax, the non-uniform rescaling by the azimuthal distribution of r17, in combination with the Vmax variation, also creates an asymmetry in the location of the maximum winds by the intersection method of the model, in contrast to the Rmax-based models.
- The largest differences in mean inner core amplitude observed for categories 2 and 3 are due to the highly transient nature of their wind fields and the existence of a highly asymmetric eyewall and eye structure (see examples from Trami, Fig 4.11 upper panels). As a reminder, we indicated in Chapter 3 that we observed the fastest phase changes for cyclones belonging to these categories, which in the case of decaying cyclones can be associated with a stronger asymmetric structure such as a triangle-like shaped eye and eyewall [Li et al. 2013], as in the previous Trami example. By averaging the different transects, this creates a greater smoothing of the profile due to the skewed position of the maximum tangential winds contrary to the more symmetric inner core of the model. Besides multiple short time-scale variations, are not fully taken into account in the intra-storm variability mode of the model (like ERC) and can also explain part of the discrepancies.
- As regards to Holland model, the profile systematically underestimates the outer core beyond a distance of 100 km, and overestimates the winds near the eyewall, as demonstrated by previous study in comparison to SFMR data [Willoughby and Rahn 2004].
- Overall, Holland model is much more dispersed throughout the TC sections, with the exception of the eye area. Errors are found greater than  $5 \text{ m.s}^{-1}$  or even  $10 \text{ m.s}^{-1}$  for category 5 cyclones.

When combined with the analysis of the standardised profiles, several conclusions can be drawn about the validity frame of the model:

- The profiles normalised by their respective Rmax show that the performance of the Chavas model is superior to that of a classical Rmax-based model for the whole profile from Rmax to at least 20Rmax and for all categories.
- The normalised profiles also reveal that the r17-based model respects much better the different wind gradients observed on all sections, for all intensity categories, with the exception of the very narrow area of increasing winds at the eye.
- The joint analysis of the evolution of the merge radius (Fig 4.19 as a function of the Vmax, shows that on average the Rmerge is around 2.5 Rmax ( $\sim 75\text{-}80\text{km}$ ) and oscillates essentially in an interval ranging from 2-3Rmax (on average  $60\text{-}95\text{km}$ ). This interval is not insignificant since it is the zone around which a slight overestimation of the model is observed, as already pointed out by [Chavas et al. 2015], and which was observed on the 2D comparisons (section 4.3.1). This is due to the existence of a transition zone between the inner and outer core dynamics, with the presence of intermittent convection (primary rainbands [Didlake and Houze 2013a]) which is not taken into account by the model [Chavas et al. 2015].

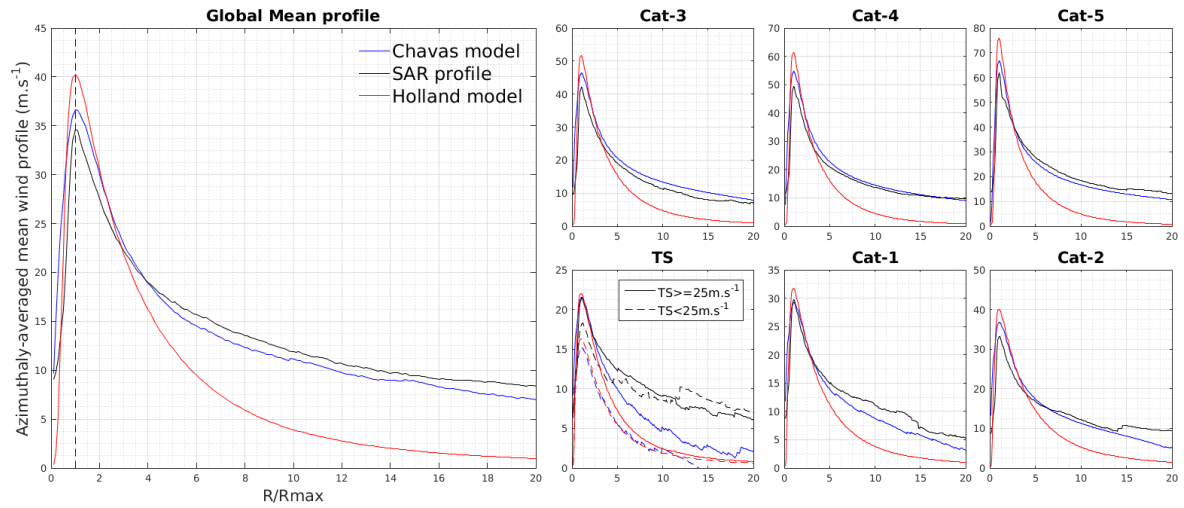


Figure 4.17: Same as Fig 4.15 but normalized respectively by the  $R_{max}$  derived from SAR (black), Chavas(blue) and Holland(red) model.

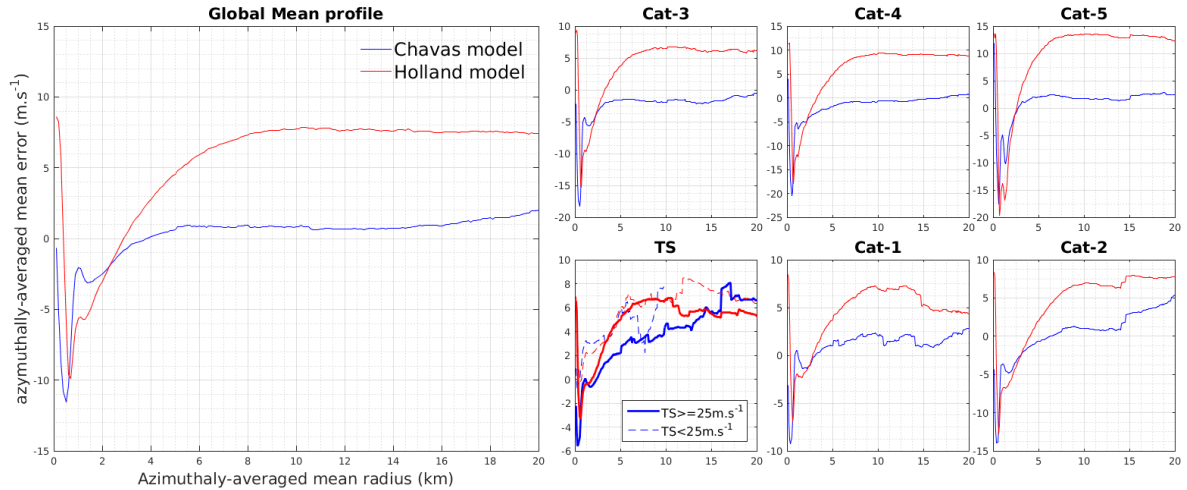


Figure 4.18: Same as Fig 4.16 but normalized by the  $R_{max}$  values as in Fig 4.17.

- A band of optimal performance exists after the  $R_{merge}$ , around  $3-4R_{max}$  and extending over the rest of the profile, with an observed asymptotic minimum error that fluctuates essentially around  $\pm 1/2 \text{ m.s}^{-1}$ , whereas this value is on average  $7-10 \text{ m.s}^{-1}$  for the Holland model. The too rapid decay of the outer core winds by the Holland model is even more marked for category 5, with a significant difference of  $12 \text{ m.s}^{-1}$ .
- The contraction theory also seems to be respected on average. The model predicts a decrease in  $R_{max}$  and convective surface for a stable outer circulation as the system intensifies, which should result in a decrease in  $V_{merge}/V_{max}$  and  $R_{merge}/R_{max}$ , which is well observed in the Figure 4.19. The standard deviation of this trend is of course much more pronounced for  $R_{merge}/R_{max}$ , as there may be simultaneous and independent growth in cyclone size, which counteracts the contraction effect. The r17 model represents both modes of variability of the  $R_{max}$  dynamics well, mainly for the span (0-50 km). Above 60 km, however, there is a loss of sensitivity,



as the model does not predict as large  $R_{max}$  for intense cyclones. This is therefore a limitation in describing the mode of intra-storm variability, as the model omits some turbulent-scale effects. The plateau for  $R_{max} > 100$  km for tropical storms is rather related to a metric problem (r17).

- The 17 represents an excellent size parameter given the performance of the model, especially in the outer part of the profile with an excellent fit for categories 2 to 5. In view of the mean ratio of  $V_{merge}/V_{max}$  (Fig 4.19) and the comparison of normalized profiles (Fig 4.17), r17 is not appropriate for tropical storms, especially for those below  $25 \text{ m.s}^{-1}$ , where the radius is located in the inner core. The r12 or r10 metric should be preferred. Category 1 storms also experience greater lift-off for radii greater than  $10R_{max}$ , which may suggest that another metric should also be used, even if its impact is less. On the other hand, the large rise in the mean SAR profile around  $15R_{max}$  for category 2, is consistent with satellite swath problems (incomplete cyclone).

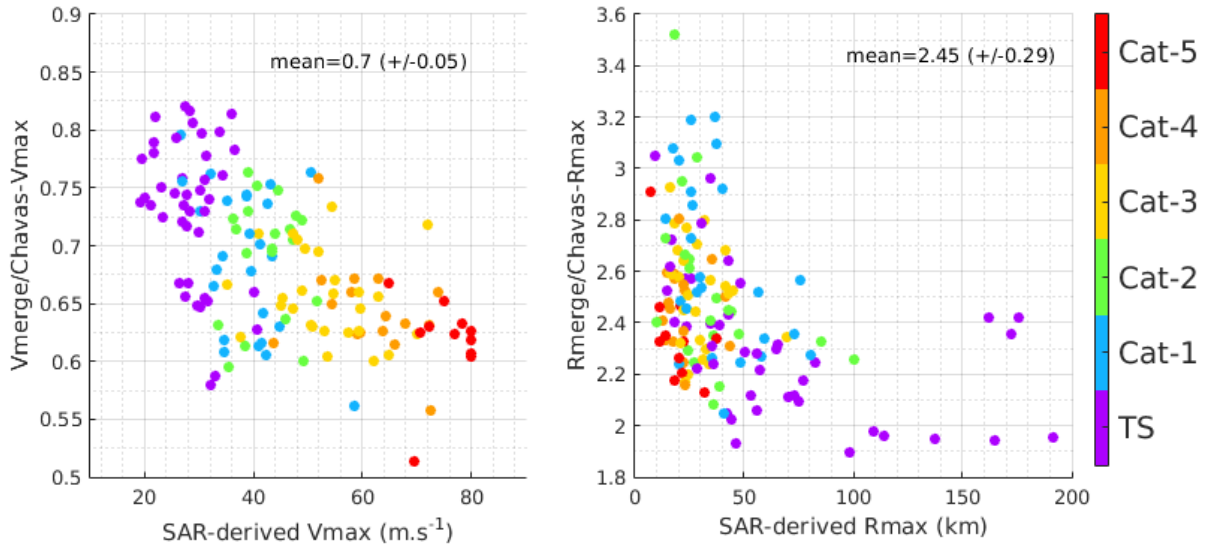


Figure 4.19: Analysis of the merging point between the outer core and inner core solutions according to: a) its intensity ratio and b) its range from  $R_{max}$ , as a function of TC wind category. There are more cases, because we also examined these values for situations without simultaneous radiometric acquisitions, quickly incorporating the BTK r17s, in order to get an overview of the evolution of  $R_{merge}$ .

#### 4.4.2 General Outcomes.

In this chapter, we compared all our MW observations of different resolutions with an innovative parametric model which, using the  $V_{max}$  information from the BTK and an r17 from the constellation of radiometric and scatterometric instruments, allows us to capture the complete profile of the TCs over all categories and to predict  $R_{max}$  with an accuracy comparable to the BTK when informed by the SFMR data, respecting the modes of variability specific to each wind radius. The combination of the translation speed with the geometry of the r17 contour also makes it possible, through the model intersection

method, to represent globally the asymmetries of the wind field, and to adjust  $R_{\max}$  to the azimuthal variations of both  $V_{\max}$  and  $r_{17}$ , in order to obtain a 2D field consistent with the high-resolution SAR observations. However, some limitations were noted, such as the slight overestimation of the mean intensity of the inner core, the difficulty in describing the wind field of tropical storms below  $25 \text{ m.s}^{-1}$ , and the underestimation of  $R_{\max}$  above 60 km. Despite these shortcomings, the model is capable of interpreting the vast majority of tropical cyclones forcing.

More generally, Part II will have highlighted the formidable capabilities of the low (radiometers), medium (scatterometers) and high resolution (SAR) MW instruments. On the one hand, the impressive combined sampling of radiometers and scatterometers allows a global follow-up of most of the cyclonic passages that took place during our observation period, 2010-2018. Although we have only analysed a handful of our database (10-20%), more than 5000 passages could be identified over this period, and 80% of our SAR observations can benefit from concurrent information of less than 12 hours with one of these sensors (70% less than 5 hours). On the other hand, the incredible collusion between radiometric and SAR instruments allows us to describe the most intense winds but with different resolutions. At last, the paper presented in Chapter 3 summarises recent efforts to use high-resolution, cross-polarised SAR measurements to assess the full spectrum of cyclonic intensities and, more importantly, to allow an accurate examination of the full structure of the wind field, including the extraction of  $R_{\max}$ . Furthermore, the paper shows that the reliability of the SAR estimates is comparable to SFMR, but is available over all basins.

We have now reached the end of our efforts to represent the cyclonic wind field. Thanks to these three chapters, we have three different sources of information to inform the forcing parameters in the scaling laws. An ideal solution in case of acquisition (SAR), a source of interpolated hourly data (BTK) and a possible solution for generalisation to SAR data (modified Chavas model in 2D). These three different solutions allow us to explore the oceanic response and the part of the variability explained by the chosen source of forcing information.

## **Part III**

# **Tropical Cyclone Wake Analysis: Mixing Process.**

Now that we get the depiction of the forcing and the history of cyclone positions, we need to turn our attention to the oceanic response along this track. Contrary to the previous observations, centred around the specific measurement of the wind field, we will deal here with instruments that are heterogeneous in their measures. Based on these different sensors, the first chapter will describe the methodology used to collect the different physical quantities useful for the analysis of the cold wake, such as surface temperature and sea level (SST/SSH), and those related to the description of the pre-storm oceanic structure. We will also see how these different elements fit into a coherent framework, and how they are ingested in scaling laws that link the oceanic response to the wind field and stratification parameters.

Due to the scale of the circulations that occur during the forcing and the ubiquity in the ocean of mesoscale structures of similar size, the SSH signature can be particularly difficult to interpret, especially because of its dynamics. Several methodologies were thus developed before the one finally used. We will evoke them through their impact on the results, which will provide us with elements for discussion at the conclusion of this thesis. An article will close this chapter with a case study on the East Pacific basin. This focus will highlight the outcome of our methodology for the exploitation of these different satellites and in-situ data. These results will then be taken to a more global scale with the other chapter that comes afterwards.

It will echo Chapters 3 and 4, as they will in turn discuss the wake study through the exploitation of information induced by the SAR, then through information derived from the parametric model. Strong of more than a hundred cases sampled by SAR measurements, we will observe the dynamics of the oceanic response via temperature and sea level anomalies. We will go further in this analysis of the wake, notably through its spatial dimension: whether in the horizontal plane with the width of the signature, or the vertical aspect thanks to the use of argo data. Emphasis will also be put on the importance of the different measurements in the quality of the results, which will further corroborate the "philosophy" of this thesis. Finally, we will take advantage of this base and the scaling laws to explore the behaviour of the drag coefficient, this key term linked to the parameterisation of forcing but also more widely to air/sea interactions.

Once again to address the study stakes, several instruments and products will be used in this part along with several physical variables. In order to ease the reading, and to offer some first key elements, we provide again tables summarising the data used and the quantities observed by each of them.

Table 4.1: List of L2 altimeters from CMEMS database.

| Instrument   | Jason 1<br>Poseidon-2 | Jason 2/3<br>Poseidon-3/3B | HY-2A<br>ALT | SARAL<br>Altika | Cryosat-2<br>SIRAL | S3A<br>SRAL |
|--------------|-----------------------|----------------------------|--------------|-----------------|--------------------|-------------|
| Chanel       | C/Ku                  | C/Ku                       | C/Ku         | Ka              | Ku                 | C/Ku        |
| Orbit        | 66                    | 66                         | SSO          | SSO             | near polar         | SSO         |
| Cycle (days) | 10                    | 10                         | 14           | 35              | 29                 | 27          |
| Center       | CNES                  | CNES                       | NSOAS        | CNES            | ESA                | ESA         |
| Version      | j1n/j1g               | j2/j2n/j3                  | h2/h2g       | al/alg          | c2                 | s3a         |
| Years        | 2002-2013             | 2008(/2016)-now            | 2011-now     | 2013-now        | 2011-now           | 2016-now    |
| Database     | 2010-2013             | 2010(/2016)-2018           | 2011-2018    | 2013-2018       | 2011-2018          | 2016-2018   |

Table 4.2: List of products and in-situ data.

|            | L4-SSH       | OSTIA     | REMSS           | ISAS-15     | Argo                        |
|------------|--------------|-----------|-----------------|-------------|-----------------------------|
| Type/Level | L4-daily     | L4-daily  | L4-daily        | Climatology | Individual Profile          |
| Resolution | 0.25         | 0.25      | 0.25            | 0.01        | fluctuating                 |
| Version    | Delayed mode | N/A       | 1)MW<br>2)MW+IR | N/A         | Delayed only                |
| Center     | CMEMS        | CMEMS     | REMSS           | IFREMER     | CMEMS                       |
| Database   | 2010-2018    | 2010-2018 | 2010-2018       | N/A         | 2010-2018                   |
| Variables  | SSH          | SST       | SST             | N-frequency | N-frequency,<br>MLD/SST/SSS |



# Extraction of sea surface anomalies

## - Methodologies and synergy between observations.

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|       |                                                                                                    |     |
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As we have seen in Chapter 1, the structure of the oceanic surface parameters is driven by multiple interactions mostly ruled by turbulent processes [Cronin and Sprintall 2001; Rudnick 2001; Ferrari and Wunsch 2009; Josey and Taylor 2019]. While mesoscale eddies are ubiquitous and dominate the kinetic energy spectrum of the ocean circulation [Wunsch and Ferrari 2004], prominent sea surface height anomalies are linked to the large scale surface current systems, with variations of the mean circulation that can exceed one meter [Le Traon and Morrow 2001]. Such signatures can easily blur the TC wake SSH signatures, whose values are of the order of tens of centimeters [Ginis 2002]. Thus, achieving a database capable to detect and analyse the TC cold wake is not straightforward, particularly at the western edge of subtropical regions where fronts [Fu et al. 2010] and large-scale eddies [Chelton et al. 2007] from current meanders are raging. As a result, a robust methodology need to be developed to fully interpret the processes beneath the storm. In this chapter, we will look at the different methods used to probe the ocean response and at the limits encountered, before introducing the one finally used. In this way, we will highlight the efforts carried out to build this database, and the capability of current observations to provide high-quality TC wake analysis. In the scope of our thesis, this will bring a strong argument about the necessity of synergy between space-borne instruments.

## 5.1 Methodologies to derive sea surface anomalies in the TC wake from daily-interpolated data.

### 5.1.1 Data and space-time frame of the cold wake analysis.

To ease the synergy between the different observations, Best-Track is used to give us a temporal and spatial skeleton of storm positions, on which we can aggregate the different observations of the ocean response and get the TC parameters ( $V_{max}/R_{max}$ ) every six hours. In the meantime, this position history can help to put the TC in the context of its basin and locate the errors sources. But before an inventory can be made, the time frame of the analysis must be set up.

As a TC is a transient source of wind forcing with variations of the order of the inertial period, its oceanic response is of similar time scale, with a mixing process going from one to three days [Price et al. 1994]. This perturbed state is followed by a rapid decay (e-folding time  $\sim 15$  days) of the surface anomalies [Hart et al. 2007; Dare and McBride 2011], although anomalies related to the vertical structure can persist for several weeks, as shown by the SSH trench that remains deep several days after the storm. Considering both the short time scale of this forced regime and the persistence of sea level anomalies, we will focus on a ten-day window [d+1 to d+10] to collect the maximum amplitude anomalies from the different post-storm observations.

Similarly, a pre-storm condition is usually computed to retrieve the anomalies from the post-storm variables field [Vincent et al. 2013; Mei and Pasquero 2013]. This estimate is often built on a mean field, whose precise averaging window needs to be a good compromise portraying the ocean conditions: not too long to capture the variability of the mesoscale activity, and not too short to remove the small scale turbulence that are not representative of the ocean state. To meet these conditions, we will lay on a two weeks mean [d-21:d-5] before the storm passage. This interval will help us to capture the fluctuations and suits to the temporal mesoscale variability. The few days before the TC passage are typically removed from the analysis, due either to interactions with its outer core or its remotely-generated swell, or some pre-existing artefacts in the ocean signature coming from the temporal smoothing of daily products.

These pre-conditions will be derived from L4 observations for both SST and SSH, using the microwave-combined product from REMSS and the altimetry gridded product from CMEMS respectively. These two data sources are composite of different observations interpolated in time and space to create a daily global field at 0.25 degree resolution in longitude and latitude.

For SST, four multi-channel microwave radiometers are merged together to generate the global field: GMI, AMSR-2, AMSR-E and WINDSAT. Taking benefit of the 7- and 11-GHz bands properties [Gentemann et al. 2010], these sensors can measure SST from sea surface properties even under cloud coverage [Wentz et al. 2000]. In the same way, the daily composite of absolute dynamic topography is made from the gathering of all available altimeters, at the present time. Unlike for SST, the number of sea level observations has evolved over the decade, passing from four sensors in 2010-2016 to six since 2016, which implies higher resolution in space and time [Ballarotta et al. 2019; Ciani et al. 2020]. To mention, only the reprocessed version is selected to benefit from optimal data coverage and reanalysis of along track measurements. As the data ingested in both products (SST and SSH) are irregularly sampled, an optimal interpolation method is used to generate a consistent global field in space and time [Pujol et al. 2016; Reynolds and Smith 1994]. To do so, this mapping method requests a statistical knowledge of the sensors errors and

the ocean field (variance, noise, correlation scale) to merge the observations on a regular grid [Pujol et al. 2016], which leads to a compromise in the scale of the physical signal represented. At the end, if these two global fields are delivered on daily basis, the effective temporal resolution is actually lower. It is constrained both by the number of sensors and by the time correlation of the interpolation method. In any case, these L4 data will help us to define pre-storm means based on a regular sampling, in the course of two weeks prior to the storm.

In contrast with the rather short period of the forcing stage [Price et al. 1994], the analysis of the mixing is one month scale. We must take into account the interactions that may occur in this time frame. In addition to this aspect, it is also paramount to monitor the spatial features of the basin in case of interference with the TC track. Therefore, several issues must be addressed:

- Shallow waters and coastlines.
- Western boundary currents and fronts.
- Pre-existing storms signature.
- Interaction with concurrent post-storm events.
- Complex trajectory.

This list includes the various factors that can be interpreted as sources of errors, they hinder either (i) the understanding of the signature itself, via pre-existing or subsequent events, (ii) or its visibility with a stronger activity (currents), (iii) or either the validity of our model itself, like in the coastal areas. Indeed, their shallower depth do not offer the same dispersion relationship or the same dynamics for the development of baroclinic waves [Ginis and Sutyrin 1995; Jansen et al. 2010]. Similarly, loop trajectories can be closer to a steady state that emphasizes the response of the barotropic mode and thus worsen our analysis. The global eddies activity is not included in this list, as we already addressed this issue with the pre-storm mean condition. For each of the disturbance above, we have developed specific solutions to bring out the TC cold wake from this turbulent background.

### 5.1.2 Filtering of the ocean background activity

We start by filtering out the coastal regions thanks to the land mask already provided in the L4 products, and we extend it to 200 km around mainlands to obey the deep ocean assumption required by the model [Geisler 1970]. Finer filtering will be performed later when we analyze the information on the vertical profiles (depth criteria), which will be more relevant around small islands. As mentioned in supplementary material of Jansen et al. [2010], depth can be an important criteria with regard to the ratio between the barotropic and baroclinic components. Indeed, while the barotropic response can be scaled as an inverted function of the depth, the ratio between the two responses ( $\frac{\eta_b}{\eta_T}$ , upper term: baroclinic response) is in contrast proportional to the vertical dimension of the ocean structure:

$$\eta_T \sim \frac{2\tau f R^2}{\rho g H U_H} \quad (5.1)$$

$$\frac{\eta_b}{\eta_T} \sim \frac{\Delta\rho}{\rho} \frac{H}{R^2 f^2} \quad (5.2)$$

with  $\tau$  the wind stress,  $R$  the size of the TC and  $U_H$  its translation speed,  $f$  the Coriolis parameter and  $g$  the gravitational acceleration,  $H$  the ocean depth and  $\rho$  its density. In view of the effect of the  $H$  parameter in the two equations, it is obvious that the barotropic response can get significantly stronger or even dominant for shallow waters [Ginis and Sutyin 1995]. The coastal regions are therefore limiting areas for the scaling laws.

Speaking about the specificity of ocean borders, we then try to isolate Western boundary surface currents that trigger intense SSH anomalies, by making the difference between one day of L4-daily SSH product and the yearly-averaged mean. Taking a threshold of 0.5 meter amplitude in absolute value, this give us roughly the areas dominated by the current regime (Fig 5.1). In addition to this overwhelming signature, eddies from those regions display some of the largest spatial scale ( $> 200$  km) and circulation speed ( $> 20\text{cm.s}^{-1}$ ) observed [Chelton et al. 2007, 2011], going along with an high mean amplitude ( $> 0.25$  m per 1 deg square) [Chelton et al. 2007; Fu et al. 2010] and a great temporal variability as a lot of short-time eddies occur there [Chen and Han 2019]. These salient features coupled with fast fluctuations are a real obstacle to analyse the wake and to define a pre-storm mean state, that we need to compute the anomalies. Around this troublesome concentration of activity, we create polygons to delineate the current systems and keep them as limiting areas (see Fig 5.1). Thanks to these two masks (the land mask and this one), every TC position crossing those particular regions will be automatically removed from the analysis.

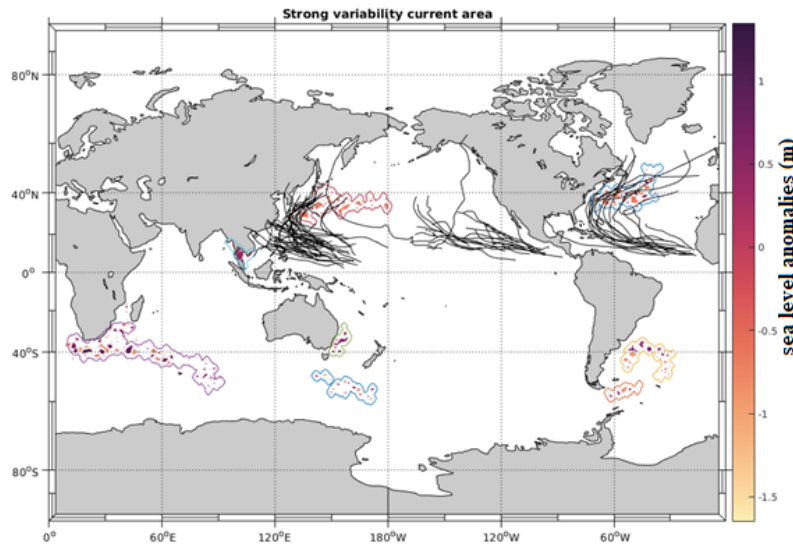


Figure 5.1: *Global map of the first methodology samples. High-variability areas are delimited by contours of singular colours to depict the different surface current systems. Sea level anomalies are computed from L4-daily gridded ADT product of CMEMS with the difference between one day grid and 2018 yearly-mean value (shaded colors), only the strongest anomalies were kept to isolate strong variable areas ( $> |0.5|m$ ).*

Now that we have dealt with the environmental context, we have to focus on the surrounding cyclonic events. As TCs can leave quite persistent signatures (see 1.3) that change the oceanic structure and influence future activity [Balaguru et al. 2020], it is important to contextualize each case in its season and location. As already mentioned in 1.1, each basin has its own features in terms of TC numbers, size, category distribution [Knapp et al. 2010; Chavas et al. 2016],... . It means that the TC density is fluctuating from one basin to another according to its extent, shoreline and its activity. On this level, the North-West (WP) and East Pacific (EP) come to head with the largest annual range

of cases with respectively 24 and 19 cases in average per year [Bell et al. 2013]. While WP offers a broad crossing place between concomitant cases, EP has the highest monthly density (maximum at 5 per month per  $10^6 \text{ km}^2$ , Bell et al. [2013]) due to the narrowness of its extent and its reduced season. Generally speaking, they both experience higher density than other basins standards with peak values up to 3 storms per month per  $10^6 \text{ km}^2$ , [Bell et al. 2013]. This parameter is important to consider, as the analysis of the mixing anomalies is done at the scale of the month, considering both the timespan of the pre-mean condition (up to d-21) and the post-storm sampling (up to d+10). If other basins can experience such gathering of cases, particular care must be paid to these two very busy regions. For each TC, we thus generate a grid of  $\pm$  five degrees at the edge of its trajectory to determine the possible intertwining TC tracks, and we first keep every singular event that crosses spatially this box. In a second time, we refine our selection by finding the crossing areas between the different TCs, thanks to r17 information from Best-Track. Once we have monitored the potential intersections at each TC position, we look if they also meet one of the two time spans (pre or post-storm), in which case we reduce the interval to the closest crossing time. When another event appears to intercept the track in a time window as near as -5 to +3 days, the TC points are removed from analysis, because neither the pre-mean condition nor the post-storm anomalies can be properly assessed.

To conclude on interaction between TCs, there is also the particular scenario where the storm is interacting with itself. This situation occurs when its trajectory forks and describes a loop. According to its diameter, the effective time of the forcing acting on the same sea surface can be dramatically increased. Similarly to a steady state, this allows time to set a deep geostrophic adjusted circulation on surface, that transports waters away from the center and compete the internal isopycnal displacement [Geisler 1970; Ginis 2002]. In the meantime, the diverging Ekman transport and its associated upwelling take a significant part in the cooling process at the expense of the wind-induced turbulent mixing [Vincent et al. 2012a; Jullien et al. 2012]. This paradigm shift represents a pitfall for the scaling laws, as the barotropic response can become significant (Uh very small in the equation 5.1). In the same way as concurrent events, we look at the intersections inside the track itself to isolate the points describing a loop. If the different radii of  $17\text{m.s}^{-1}$  are overlapping, the associated TC positions are removed from the analysis.

To finish on these filtering processes, we have also considered some additional criteria related to the dataset quality, such as the basin topography, weak and high-latitude cases filtering. Due to the confinement of the North Indian basin and the amount of island or land interactions in the southern hemisphere ocean, which complicate the SSHA analysis with an automatic approach, we first restrain to four basins: North Atlantic, West-, East- and Central Pacific. For the remaining cases, we then put an intensity filter that removes cases which never reached hurricane intensity threshold, which are often associated with a too weak ocean response. At last, high-latitude storms that experience Extra-tropical transition are also removed (typically higher than 40 degrees of latitude). They do not obey at the same mechanisms than tropical cyclone [Beucher 2010], and the ocean interior description also differs from tropical and subtropical regions, with a higher seasonal stratification variability and an important meridional gradient of SST(fronts). Besides, during fall and winter, the strong deepening induced by the more frequent winds in mid-latitude regions and the decrease of radiative fluxes, usually prevents subsurface anomalies from persisting, which is not relevant for long-term analysis. Thanks to all these filtering operations we end up with a dataset of 134 cases, for which we are going now to use different set of observations to study the wake.



### 5.1.3 Extraction of ocean variables from daily interpolated products

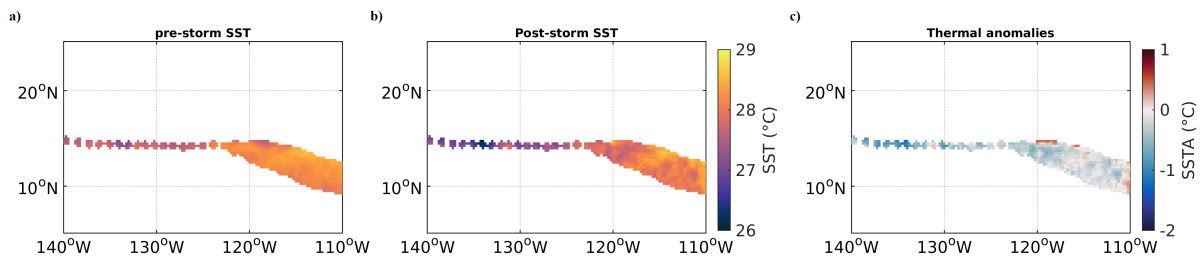


Figure 5.2: *Global map of the first methodology samples. High-variability areas are delimited by contours of singular colours to depict the different surface current systems. Sea level anomalies are computed from L4-daily gridded ADT product of CMEMS with the difference between one day grid and 2018 yearly-mean value (shaded colors), only the strongest anomalies were kept to isolate strong variable areas ( $> |0.5|m$ ).*

For each TC, we have also followed a very specific methodology to extract all the required ocean parameters: SST, SSH and the Brunt-Väisälä frequency of the seasonal thermocline. A same scheme is applied to sample the two surface variables, before and after the TC passage, in order to infer the respective anomalies. We start by looking at the singular post-storm observations, at each time step from  $d+1$  to  $d+10$  along the TC trajectory (Best-Track), for both the L4-daily REMSS product (SST) and the along-track altimetry measurements. Around each successive position, a circle of 2.5 times the maximum wind radius ( $R_{max}$ ) is drawn to keep the ten daily post-storm temperature values, with a minimum authorized threshold of 50 kilometers (similarly as in Fig 1.34). The choice of this diameter is not arbitrary, as the maximum cooling radius is usually located between the center and a distance of  $2R_{max}$  [Vincent et al. 2012a]. We have chosen to take an high estimate (see Hector example in Fig 5.2).

Regarding SSH, the 6-hour Best-track is interpolated every minute to know the exact time difference between the TC passage and the whole set of altimeters that crossed its trajectory during its lifespan, only the post-storm tracks included in the ten-day time window are kept (section 5.1.1). We also take advantage of this step to adjust at the different crossing times, each TC parameter:  $V_{max}$ ,  $R_{max}$  ... . The translation speed  $V_{fm}$ , derived from the 6-hour positions of BTK, is interpolated at the crossing time by a weight function of the two surrounding values. Thus, for each altimeter measurement, a set of interpolated parameters is associated.

At this stage of the study, only the SSH retrieved from the Jason constellation was considered in this first dataset, to benefit from their short-time cycle repetitiveness ( $\sim 10$ days). The Figure 5.3a illustrates one of the along track measurement operated by Jason-3 on Hector case (black line), with a noticeable trench at the location of the storm center. Indeed for all the remaining tracks, we have computed the distance from the respective center for each point sampled by the altimeters, following the classical convention dictated by the direction of cyclone motion, namely: the positive side stands for the right of the trajectory, while the left side is set as negative (Northern Hemisphere). The SST analysis follows this same rule. A last convention on the inclination of the altimeter tracks is applied. As cyclones leave an along-track wake, the characteristic direction to analyse the cold trench at a given time is perpendicular to its trajectory. Taking into account the geometry of the ocean signature, a tilt criterion of 30 degrees minimum is added. Because

of the mainly zonal displacement of the low latitude systems, most of the measurements can be kept [Knapp et al. 2010]. The black curve observed in the Figure 5.3a is thus a cross-track point of view of the Hector wake. To mention, the large signature on the left of the track (around -500 km) represents the intense equatorial upwelling.

Now that we have the post-storm observations, we need to portray the pre-storm condition to derive the TC-induced anomalies. From the two L4-daily products we derive a mean global field on the time window [d-21:d-5], centered around each 6-hour BTK position for the temperature (see Fig 5.2a), and at each crossing time for the SSH. To be comparable with the post-storm estimates, the pre-SST are then arranged on the same circular areas than used before, while the 2D pre-SSH are interpolated along the respective 1D altimeter tracks and set according to their distances to center, like in the example of Hector in Figure 5.3a (magenta curve). Finally, both anomalies are computed from the difference between singular post-storm observations and the averaged pre-storm conditions (Fig 5.2c & 5.3b). This process is repeated for thermal anomalies at each time step from d+1 to d+10, in order to find the the maximum cooling occurrence and referenced it as the final TC-induced SSTA. On the other hand, a reversed peak analysis is performed to capture the SSHA (black crosses in Fig 5.3b), the strongest negative extremum located in the vicinity of storm center is defined as the TC-induced SSHA (red circle in Fig 5.3b). The illustrated case of Hector (Fig 5.3b) introduces the typical cross-track patterns left by the TC, with nearby the center, a negative anomaly in the shape of a geostrophically adjusted trench induced by the penetration of near-inertial baroclinic internal waves [Geisler 1970; Price et al. 1994; Shay et al. 1992], and away from the storm, the positive anomalies from the outward converging flow [Price 1981; Ginis 2002; Mei et al. 2013]. From the retrieval of these two ocean variables, the picture is almost complete to analyze the TC wake response. Yet according to equation 1.22 in Section 1.5, the pre-storm ocean interior description is now the last missing piece required.

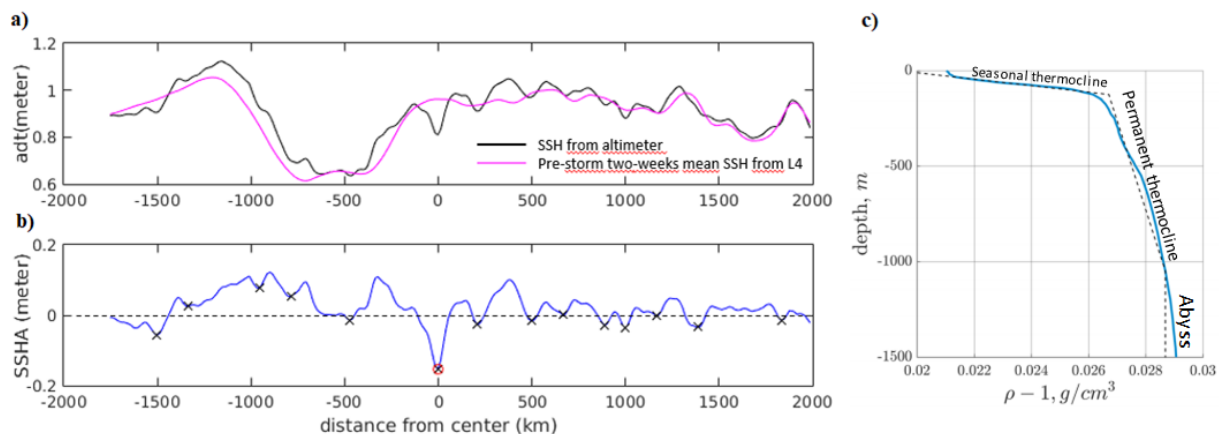


Figure 5.3: *Illustration of the analysis method for sea level anomalies and ocean stratification. a) Example of post-storm SSH from direct L2-altimeter measurements (black) along Hector track and pre-storm mean SSH computed from L4-daily product (magenta) over a period ranging from d-20 to d-5 before the storm. b) Corresponding SSHA computed from the difference between the post-storm values and the pre-storm mean condition, the identification method of hollows (black crosses) allow to dissociate the TC-induced trough (red circle) from the other mesoscale signatures. c) Decomposition of an ocean vertical profile into three layers: seasonal thermocline (upper ocean), main thermocline and abyss, this sketch was extracted and modified from [Kudryavtsev et al. 2019a].*

As indicated in section 1.1.1 and 1.5, the modulation of the oceanic response to any forcing lays on its stratification described by the Brunt-Väisälä frequency  $N$  [Shay 2009; Reul et al. 2014]. A sharper density gradient induces a stronger stability, a higher energy is in return required to disrupt this structure and trigger an efficient mixing. On the other hand, a greater value of  $N$  implies the proximity of colder water near the base of the ML, an enhanced cooling may occur in case of perturbed state. This condition is assumed to be fulfilled during the TC forcing [Kudryavtsev et al. 2019b], for which the  $N$  frequency represents a favorable parameter to SSTA. As regards to the SSH signal, a higher stratification also emphasises the development of the baroclinic near-inertial waves (gradient of density in equation 5.2). It is thus an essential variable to monitor, as it is a vector of the ocean response. To document the vertical structure of the ocean, we look at profiles given by ISAS-15 climatology at each Best-track points. This climatology is built on the Argo constellation over the period 2002-2015, and provides monthly-averaged vertical profiles [Gaillard et al. 2016]. The colocated information are then split into four distinct parts: mixing layer, seasonal thermocline, main thermocline and the abyss (Fig 5.3c).

We start to assess the ML depth using three different approaches: the curvature method, described by [Chu and Fan 2010]; the maximum of the second derivative (temperature), similarly as in [Kudryavtsev et al. 2019a], and finally the temperature criteria of 0.2 degree [de Boyer Montégut et al. 2004]. The former is a linear fit technique based on the geometric properties of the vertical structure, we choose to apply it on the density profile unlike the two other methods. Through the computation of two parameters, this method allows to identify the steep boundary that separates the homogeneous layer from the slope of the thermocline. Due to the narrowness of the mixing layer during the cyclonic period, we limited the analysis to the first 100 meters. Several steps are then necessary to retrieve the MLD:

- 1) We Interpolate the climatological profile at a finer scale (1 meter).
- 2) We derive several linear fits from the surface to a depth  $Z_k$ , with  $Z$  going from 1 to 100 meter.
- 3) For each increment, we compute the root-mean square error  $Rs(k)$  between the linear ( $\hat{\rho}_i$ ) and the climatological ( $\rho_i$ ) profiles of  $Z(k)$  depth. Small values of  $Rs$  indicate that the profile from ISAS climatology is closed to a linear behavior for the associated range  $[1:k]$ :

$$Rs(k) = \sqrt{\frac{1}{k} \sum_{i=1}^k (\rho_i - \hat{\rho}_i)^2} \quad (5.3)$$

However, in order to know the maximum depth for which the ISAS profile looks like a homogeneous layer, a second criterion is necessary.

- 4) For each profile of  $Z(k)$  depth, we extrapolate the linear fits to  $k+5$ . We then compute the bias between the fitting and the observation for the extrapolated points  $[k:k+5]$ :

$$Bias(k) = \frac{1}{n} \sum_{j=1}^n (\rho_{k+j} - \hat{\rho}_{k+j})^2 \quad (5.4)$$

- 5) We make for each increment the ratio between the bias (in absolute value) and  $Rs$ . The value of  $k$  which gives the maximum ratio is defined as the MLD.

Indeed, at the base of the mixing layer the ratio is maximum, with a very low root-mean square error due to the homogeneous behavior of the density, followed by a significant bias for the extrapolated points of the linear fit due to the thermocline jump. This multiple-step approach will provide our reference MLD. However, in case of a good agreement between the two others methods ( $<10$  m) and strong discrepancies with this curvature technique ( $>15$  m), we choose the value given by the 0.2 degree method as the final MLD. Actually, the 4-layer description of the density profiles can be inaccurate in presence of particular salinity features (barrier and compensated layers). Disagreement between the temperature and density based methods can help to identify this type of situation. For those cases, the temperature profile is more suitable to infer the homogeneous layer depth. Once the base of the ML is found, we look at the structure below it.

We use this time the maximum angle method from [Chu and Fan 2010], to distinguish the sharp gradient of the seasonal thermocline from the softer permanent one. We search for the maximum angle between subsequent sections of 10 meters, in order to find the approximate depth of the seasonal thermocline ( $ST_h$ ). Around this depth, we derive two sets of linear fit: one joining the mixing layer base to a range of  $\pm 20$  m around the approximate value, and the other joining this area to the lower boundary of the main thermocline, which was simply defined as the depth that contains 95% of the total observed density drop [Kudryavtsev et al. 2019a]. At last, the point from the [Ast-20: Ast+20] interval that gives the best linear approximations of both thermocline trends, is set as the seasonal thermocline depth.

Thanks to this analysis of the vertical structure, we are then able to infer the N-frequency of the seasonal thermocline:

$$N_1 = \sqrt{\frac{-g}{\rho_0} \frac{\rho_1 - \rho_0}{h_1 - h_0}} \quad (5.5)$$

with  $\rho_0$  the density at the ML depth ( $h_0$ ),  $\rho_1$  the density at the basis of the seasonal thermocline ( $h_1$ ). From the extraction of this different parameters, we can also extract the speed of the first baroclinic mode:

$$(c_1)^2 = g \frac{\Delta\rho}{\rho_1} \frac{h_0 h_1}{h_0 + h_1} \quad (5.6)$$

with  $\Delta\rho = \rho_1 - \rho_0$ . To mention, this parameter will be only analysed in the next chapter, to see its influence on the scattering of results. Thanks to the exploitation of the in-situ and satellite data we have explained through this section 5.1, we have all the parameters required to now analyse the wake.

#### 5.1.4 Interpretation of the scattering in the scaling laws.

Thanks to this set of observations from different levels, we can derive the scaling laws and link the ocean response variables to the forcing and stratification parameters (Fig 5.4). To recall, they are expressed as:

$$SSTA \cdot \left( \frac{U_m^2 N_1^{3/2}}{g \alpha f^{1/2}} \right)^{-1} \propto \frac{V_{fm}}{R_{max} f} \quad (5.7)$$

$$\frac{g \cdot SSHA}{U_m^2} \propto \frac{R_{max} \cdot N_1}{V_{fm}} \quad (5.8)$$

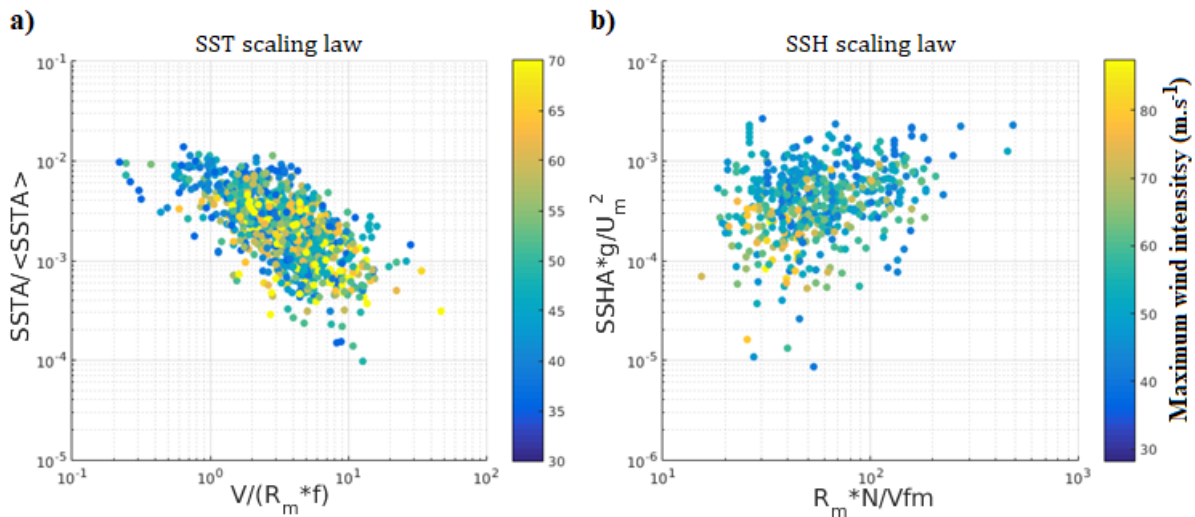


Figure 5.4: *First scaling law results for a) SSTA and b) SSHA.*

Two things are very striking from these results: (i) If we consider the linear regression inferred from the equation 5.7 and 5.8, a great disparity exists between the two correlations. On one hand, the SSTA shows a good value around 0.7, when the SSH is struggling to reach 0.5 (0.49). (ii) A wide dispersion is observed for the two answers, and it is even more marked for the SSHA.

In view of the methodology applied and the outcomes, it is legitimate to wonder about the noisy answer of the sea surface height signature. If several elements have already been addressed during the methodological display, there are other points more intrinsic to the nature of the signal that deserve attention as well. This multi-factorial issue raises its share of caveats on the use of L4 daily products for this kind of study. Among the various problems, we can mention five particularly important points:

- The statistical properties of the dataset. Among the previous studies that have looked at sea level anomalies, most of them used a much larger database :[Jansen et al. \[2010\]](#); [Mei and Pasquero \[2013\]](#); [Vincent et al. \[2013\]](#). The analysis of a smaller cases sample may require a more refined analysis. [Jansen et al. \[2010\]](#) also mentioned that because of the dynamics of the SSH signature, a large number of cases is often needed to observe a behavior and to be statistically relevant. This aspect of the signal brings us to the next point.
- The dynamic of SSHA. At the rear of the storm, a complex mesh of upwelling and downwelling cells is set up in combination with a geostrophic along track current, over a distance proportional to the product of the translation speed by the local inertial period [[Geisler 1970](#); [Shay and Elsberry 1987](#)]. As already discussed, this circulation takes the shape of a hyperbolic wedge composed of a depression along the trajectory and positive anomalies that radiates away from the center. The wedge opening (its angle), the distance between the different cells and the decay of their magnitudes rely on the translation speed of the storm, the baroclinic wave speed (mainly first and second mode) and the inertial wavelength [[Geisler 1970](#); [Shay and Elsberry 1987](#); [Price et al. 1994](#); [Ginis 2002](#)]. Due to wake geometry, the altimeter observation can give different trench estimates depending on its orbit orientation and crossing. Still the amplitude of this depression is in most cases below the ocean background variability [[Jansen et al. 2010](#)], and ambiguities can alter our method



of peak selection and lead to misinterpretation and shifting of the TC footprint. After the forcing regime, the circulation is geostrophically adjusted and starts also to propagate with a westward motion of 0.1 degree per day [Jansen et al. 2010; Mei et al. 2013; Vincent et al. 2013]. It follows an eddy-like process by trapping ocean materials whose SST is also advected at the same rate [Jansen et al. 2010; Mei and Pasquero 2013], suggesting a non linear process. Within our ten-day post-storm time window, this may also dephase our along-track observations with the pre-mean condition for the latest measurements. Owing to the smaller spatial sampling of altimeters at low latitudes, some smoothing issues from the daily product may superimpose on this effect and shift the structures of the unperturbed ocean between the pre-condition and the direct observations. Depending on the situation, it can lead to an overestimation or underestimation of the wake depression.

- Smoothing issue of L4-daily product. As mentioned in 5.1.1, these products use an interpolation method that first allows to go from irregularly sampled data to a regular gridded product, and in a second step, to fill the gaps. Because of this interpolation and the uneven zonal coverage of the altimeters, part of the mesoscale activity cannot be resolved ( $< 200$  km) [Ballarotta et al. 2019; Archer et al. 2020]. This effect is even more crude for low latitudes [Pujol et al. 2016]. Moreover, the effective temporal resolution of such a product is evaluated in weeks, this is why most studies [Chelton et al. 2007, 2011; Fu et al. 2010; Faghmous et al. 2015] related to eddies have focused on cases with a long lifetime ( $> 28$  days). This latter aspect raises two major issues. Because of the necessary and broad temporal correlation of the gridded product, it is possible that some of the cyclone-induced anomalies are already incorporated in the pre-storm mean, especially for cases with a large signature. Secondly, due to the temporal characteristics of both the daily product and the SSH signature, it is possible that a strong pre-existing cyclonic activity may be poorly resolved temporally and spatially and that some artifacts persist longer in the mean. These two problems raise the question of our filtering method and the definition of our time window. However, it is important to balance our statements on the effective resolution of gridded products, which have also evolved with the improvement of techniques and the growth of the altimeter constellation, which has increased from 4 altimeters to 6 in 2016. Thus the average temporal resolution has increased from 34 days to 10-15 days, with an estimated maximum of 7 days [Ciani et al. 2020]. The same goes for the spatial resolution.
- Filtering and Time window. It is now obvious that in the case of our database, a more drastic filtering has to be performed on the pre-cyclonic activity. It is also likely, in view of the temporal resolution of the gridded product, that a more distant definition from the cyclone date is required for the time window. The two weeks pre-mean should start at least 10-15 days before the storm (like in Mei et al. [2013]), as the time range previously used may underestimate the trench left by the TCs, especially those with the strongest forcing. Another solution may be to abstain from the pre-cyclonic condition [Kudryavtsev et al. 2019a] and the use of L4-daily SSH product. This last clue is the one that will be exploited in the following section 5.2. Finally our spatial sampling and the variability of the SSH signature are also a source of underestimation.
- The variability of the signature. As the wake is generated through the excitation of near-inertial gravity waves, it is an oscillating answer by nature. We have already evoked through the dynamics of SSHA the alternation of the spatial features which constitute the wake, it is question here of its temporal aspect. Several studies have

highlighted the time dependency of the three-dimensional baroclinic wake, with the spreading of internal waves of different phases according to depth levels [Price 1983]. These vertical oscillations induce variations of several centimeters in the observed sea surface trough [Price 1983; Ginis and Sutyrin 1995; Shay and Chang 1997]. Therefore it raises the question of the post-storm sampling. Because of these oscillations, it is possible that we do not capture the maximum amplitude at the altimeter transit time. Besides we only used measurements from the Jason constellation, this choice was motivated by the desire to use their short repetition cycles to analyze the temporal evolution of the anomalies. However, this brevity implies a more reduced spatial sampling, and may cause an underrated assessment of the cyclonic trenches and an under-sampling of the TC wake. It is therefore more suitable to use the whole altimeter capacity afterwards. Figure 5.5 illustrates this last aspect through the example of Hector, where 6 different altimetry tracks from 3 altimeters could be co-located around the same best-track position. They show an oscillation of the response, ranging from 0.06-0.08 m for the low trend (Fig 5.5b), to a response of 0.11-0.12 m for the high trend (Fig 5.5a). It highlights the variability of the signature which cannot be attributed either to the measurement qualities of the instruments, since the three altimeters show the same variation (Altika, Jason 2, Sentinel 3A are in both panels), or to the decline of the signature alone, since we have in both cases tracks ranging from 0-6 days (here 1 IP= 44 hours). On the other hand, the 8-day track (4 IP) of Figure 5.5b (Jason-2) can be impacted by the recovery of the signature. Finally, due to the orientation of the altimeter tracks with the cyclone wake, some measurements are not totally cross-track and may underestimate the altimeter trough. This is an additional justification for the use of the whole altimeter constellation to compensate those effects.

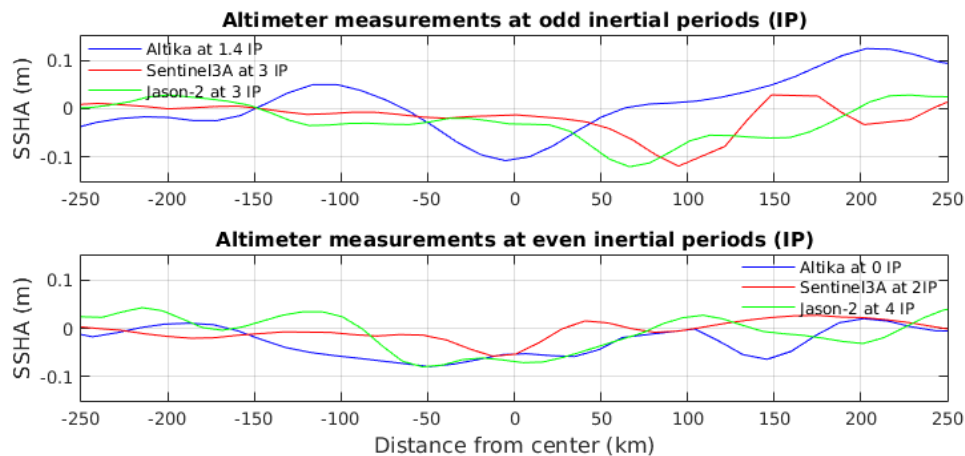


Figure 5.5: *SSHA from multiple altimeter measurements around same Hector BTK position. Upper panel: maximum values given by altimeters at odd inertial periods (0.11-0.125 m). Lower panel: low range of SSHA values given by altimeters at even inertial periods (0.06-0.07 m). Here IP=2 days.*

In the light of these explanations, several methodological adjustments are therefore necessary in addition to the massive use of different altimeters. However, before developing a new satellite data exploitation strategy, it seems important to make a few comments on the better correlation observed in the SST signature.

While we used the same level of product for the precondition, the scaling law in SST offers a good correlation of 0.7. This difference with SSH can be explained by two factors:

(i) the different nature of the observations ingested in the gridded fields. Indeed, the SST-daily product is built on 2D-swath observations, whereas the product in SSH uses nadir-looking instruments that provide only 1D-along track measurements. This implies a better spatial and temporal coverage by the different radiometers and results in a finer effective resolution (100 km in space and 2-3 days in time [Reynolds et al. 2007; Ciani et al. 2020]). The temporal smoothing effects of the interpolation method are thus less for SST, because the method is highly dependent on the data coverage capability of the instruments. Moreover, due to the 2D swaths of the radiometers, the SST product is much less latitude-dependent than SSH one [Reynolds et al. 2007; Pujol et al. 2016], which contributes to a more stable resolution. Since the wake size is of the order of the forcing size [Zhang and Lin 2019], the cooling is well resolved regardless of latitude, meaning that the pre-existing signatures of previous cyclones in the mean condition are better depicted in space and time. Of course it is not only because of the smoothing we observe such a difference, as SST still suffer from this effect to a lesser extent, but it is the addition with all the mentioned parameters that affect the correlation.

(ii) Secondly the scale of the two responses. As mentionned by Ginis [2002], TCs generate trenches going from a few centimeters to 0.5-0.6 meter, the main part of this activity is actually of the same order that the ocean background [Le Traon and Morrow 2001]. In contrast, the cooling corresponds to much larger and wider scales than the average ocean activity, with a greater variability ranging from 0 to more than 10 degrees for the slowest systems [Chiang et al. 2011]. With the exception of the costal or equatorial upwelling, the TC-induced SSTA stand out from the ocean background. In practice, it results in thermal anomalies that are more trackable by a daily product of medium resolution and that require less effort to monitor.

Finally we will conclude this section on the important scattering observed for the two signatures. If the one impacting SSH has been already explained through the different points, we have to clarify the one noticed for the SST. It is obvious that, despite a higher spatio-temporal resolution, the daily L4-data triggers some smoothing for the ocean response, especially since we are also using this product for the post-storm observations as contrary to the SSH. Thus, for a good sampling of the pre-storm condition, the post-storm trench observed by a direct altimeter observation can be of higher quality. At last, our spatial sampling is not optimal to follow the thermal anomalies at each Best-Track position. Using a circle-averaged domain can create interference with neighboring positions, so a new spatial area will be used in the next section to focus on the cross-track direction. This change in spatial wake sampling will be briefly presented, as it represents a minor change in the methodology. The next section will mainly focus on the evolution of the SSHA behavior.

## 5.2 Towards a complete tracking of TC inertial wakes from L2 MW measurements of SSH and winds.

### 5.2.1 Input from direct post-storm altimetry measurements.

Previous results have highlighted the shortcomings of our methodology for the analysis of SSH anomalies. From this experience we can draw three fundamental lessons: 1) the sampling with only 10-day cycle altimeter can hardly capture the signature variability (only 2 instruments at same time <sup>1</sup>), 2) that the inherent smoothing of the daily fields-based precondition can alter the signature dynamic, and misread prior activity or integrate a

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<sup>1</sup>L4 products number

pre-existing anomaly from the storm, at last, 3) that the accumulation of these effects lead to a significant scattering and decorrelates the signature. To address these issues, we have to change our initial approach to be able to see how the results are sensitive to observation levels.

The new proceed is based on a box method which uses only the post-storm along-track measurements of the altimeter constellation. In this way our method is only focused on direct observations of sea surface height. Two questions then arise: how to interpret the variability of the response from multiple successive measurements, and how to extract the anomalies from the sole use of post-storm tracks?

From the Best-track, we decompose the TC trajectory into a regular pattern. The distance between two adjacent positions is divided into three equal boxes: two linked to the 6-hourly positions and one intermediate (Fig 5.6a). This operation is repeated along the track in order to obtain a series of numbered boxes. The intermediate parts are self-sufficient, while for the 6-hourly BTK positions, the contributions of the two sections framing them are merged. The visual description of this decomposition is illustrated by the sketch 5.6a. Each box will then be associated to a certain number of measurements, as shown in the Figure. For each intersection, the crossing point helps to determine which box the measure is affiliated with. The Hector example (Fig 5.6b) displays the box-method on a concrete case and emphasizes at the same time the current capability of the satellite coverage. For that purpose the different instruments are also labelled. Thanks to this new approach, we expect to bring a finer sampling of the wake and in the meantime a better accounting of the SSHA variability. Once these tracks are collected and gathered in a box, we have to assess the different cyclonic trenches.

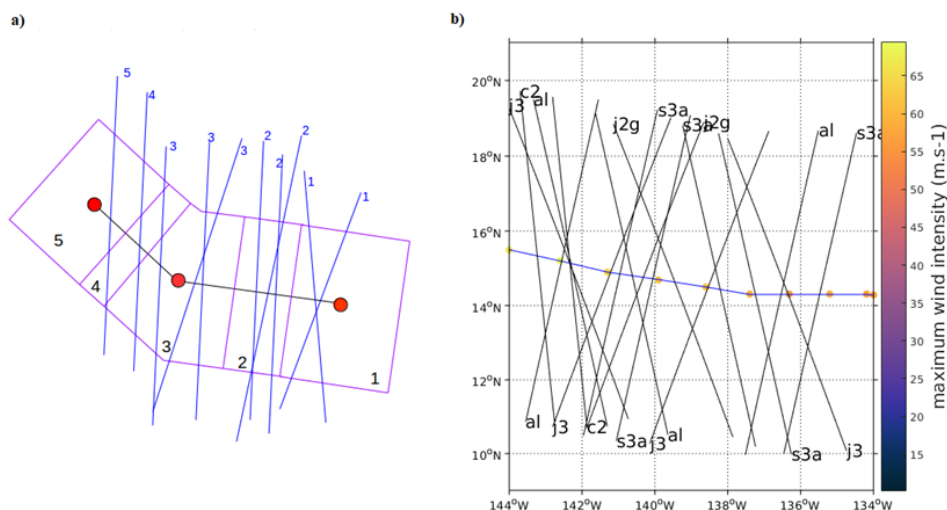


Figure 5.6: *Description of the box method for SSHA analysis. a) Sketch of the box decomposition, TCs tracks are split into 3 equal sections between each adjacent Best-Track points, the intermediate section makes its own box (box 2-4) while the two boundary sections are one side of the box centered around the BTK points (box 1-3-5), each altimeter is then associated to the box of the crossing point. b) Zoom on Hector path crossed by several altimeter passing.*

To complement this box-method, we then infer the wake depression directly from the singular post-storm observations. Thus one abstains from all the issues encountered pre-

viously with the daily product. At the location of the storm center the trench is generally noticeable in the along-track measurement, as a disruption in the general behavior of the ocean field, like it was already the case in the previous example of Hector (black curve, Fig 5.3a). From the shape of the sea surface height profile, we estimate manually the TC-induced SSHA directly from the difference between the top of the general profile and the bottom of the trough. The Figure 5.7a depicts the new retrieval of anomalies for two altimeter passing on Hector (top curve) and Lane (bottom curve) examples. For a large panel of cases, this process is quite straightforward (like in the example), as the orbits are oriented in the meridional direction where SSH follows at first order a global distribution similar to SST [Calman 1987]. Yet, at finer order the mesoscale activity bring some divergence to this state and in some situations can make it difficult to distinguish the general trend, i.e the top of the trench left by the TC. To investigate those signatures, we derive four different pre-storm conditions (d-5:d-10,d-10:d-30,d-20:d-40,d-5:d-40) to provide guidance on the general behavior of the SSH profiles and to evaluate the time-dependency of the surrounding disturbances (if they are prior to the storm or not). In case of a too ambiguous signal or too erratic ocean background activity (i.e the four pre-conditions strongly diverge), the altimeter track is removed from the analysis. For each box, we repeat this operation to all the altimeter measurements and we compare the different TC trench estimates. At last, only the maximum amplitude of the wake depression is kept, we end up with a single estimate for each box. This extraction process along with the box-method define our new methodology to try to meet the requirements previously seen in 5.1.4.

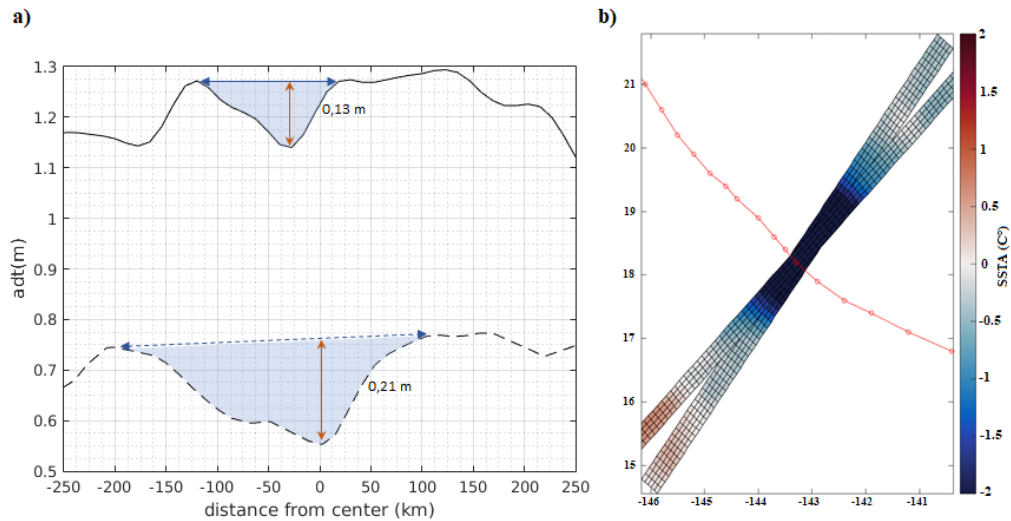


Figure 5.7: *New methodology set to extract TC-induced ocean anomalies from satellite observations. a) estimation of the sea level trench directly on post-storm altimetry track, illustrated with the examples of Hector(top curve) and Lane (below curve), b) new grid method to estimate SSTA from L4 daily product.*

To note, a small change occurred also on SSTA strategy. Instead of using a circle-shape that includes the along-track component, we focus on double cross-tracked bands (90 degrees swath apart from TC center location) of 250 km long and 25 km wide (Fig 5.7b), centered around each 6-hourly BTK position and oriented according to the trajectory before and after the TC passage (time interval remains unchanged). Both the pre-storm mean and the singular post-storm observations are interpolated on those areas, before computing the anomalies. The two rectangular areas are then averaged along-track in



two 1D SST profiles that we also merge together. From this final cross-track observation we search the lowest SSTA values to determine the TC-induced cooling at the specific storm location. From the new domain definition, a reduction of the scattering was observed (not shown here), especially for slower or larger storms, for which previous method encompassed anomalies from adjacent time periods.

Even if some different strategies have been applied in the extraction process, it is important to mention that we kept several aspects of the previous methodology, namely: the masks of limiting boundaries (coastline and currents), the range of the post-storm time window, the loop processing and the calculation of distances from the center for each track (with same convention). Regarding the seasonal activity, we have gone further in the filtering by removing this time TC points that cross several prior TC events (more than one). This more drastic selection help us to limit the impact of prior TC interactions on the wake analysis, as we no longer retrieve SSHA from pre-storm mean (just a guide), but also impact on the ocean stratification, as changes made in the density profile by previous storms are of course not taken into account in a climatology product. At last our dataset is focused on a smaller sample of 40 cases. It is obvious that this reduction is also due to a human bias, the estimation of the altimeter trough requires a semi-manual control and is therefore a time-consuming step. Among the storms that met all the conditions, we favored the simplest systems (trajectory, temporal and spatial isolation...) while taking care to keep a varied panel in terms of intensity, size and translation speed. This compromise will allow us to quickly validate our effort on an ideal case basis.

From these two main changes in the reading of anomalies come an appreciable improvement in the response behaviour. Indeed, this time we observe a strong correlation of 0.8 with the scaling law (Fig 5.8). If, however, the dispersion remains relatively large (it is a logarithmic scale), this encouraging result highlights the efforts made and is indicative of two important postulates:

- 1) The current altimeter coverage can capture the variability of the signature with a high-frequency sampling.
- 2) From the sole use of direct post-storm measurements comes a clear enhanced assessment of the wake trench.

It is also true that this simplified dataset takes better account of the basin context and is a factor of improvement. By retrieving these anomalies with the precondition method for this database, we also found a beginning of correlation at 0.6, which also proves we have a more refined methodology and that pre-storm activity can be a real limiting situation for the daily field resolution. Still an important bias was also present, with an underestimation for cases up to 0.3 m and an overrated trench for cases higher than 0.3 m. Despite the progress, the pre-storm method remains complex to depict the dynamic of the SSHA signature, while our new methodology is more straightforward. This result thus strengthens the idea of the current capability of remote sensors to directly infer wake parameters instead of smoothed daily fields. From this prospect, we have to discuss now about a possible source of scattering in Figure 5.8: the wind field input.

### 5.2.2 Influence of high-resolution surface wind measurements on SSHA prediction.

The use of direct along-track observations was beneficial to the wake analysis, it is thus legitimate to ask the same question about the contribution of different sources for TC pa-

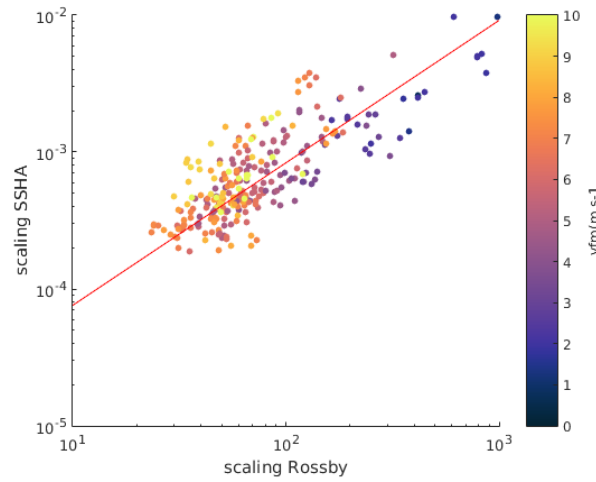


Figure 5.8: *SSHA scaling obtained with the new methodology (Fig 5.6) for the first samples shown in Figure 5.1. TC parameters are still documented from Best-Track.*

rameters. Until now, the  $R_{max}/V_{max}$  couple used for the scaling laws came exclusively from the BTK. However, as we have seen in Chapter 3, this data source can undergo large errors and important biases for several limiting situations such as rapid intensification/weakening, and more generally for the estimation of  $R_{max}$ . In order to bring some elements of discussion, we will review three cases with different distribution between the model estimates and the measurements.

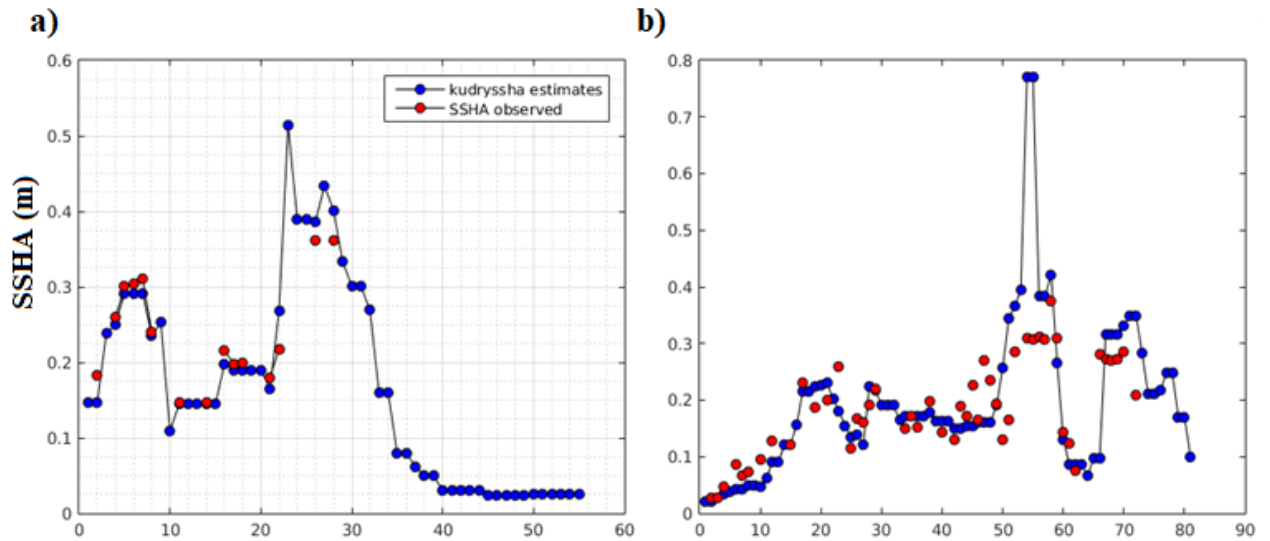


Figure 5.9: *temporal evolution of the SSHA along a) Atsani and b) Jimena tracks observed by altimeters (red points) and the SSHA retrieved from model at each Best-Track point (blue).*

The Figure 5.9 shows results for the typhoon Atsani which occurred in 2015, an ideal case with an almost homogeneous translation speed. The evolution and amplitude of its anomalies are well represented by the inverse scaling law, with observed values ranging from 0.12 to 0.38 m for the altimeter (red points), and from 0.10 to 0.42 m for the model (blue points) at similar times. The last part of the trajectory was not sampled due to the

bifurcation of the cyclone towards the North of the basin and the Kurushio region. Panel b introduces the well-studied case of cyclone Jimena, a long-lasting storm that underwent different variations in its parameters. While globally similar evolution is observed in the estimates, there is a slightly stronger dispersion with notable bias before and around the peaks. In particular, there is an overestimation of the values capping the activity, with an ambiguity at 0.7-0.8m. The latter corresponds to a borderline situation for the system, the cyclone being almost stationary, its oceanic response therefore experienced a barotropic regime. Around this ambiguity the values of the model remain higher by at least 0.06 m, and this behavior is also repeated for the next peak. We can also note the presence of a long plateau before the first extremum, with an underestimation of the activity. These biases may be due to quick fluctuations of the cyclone forcing that was not anticipated by the BTK [Combote et al. 2020b]. To investigate this last point, we are going to use the now well-documented and known case of Hector.

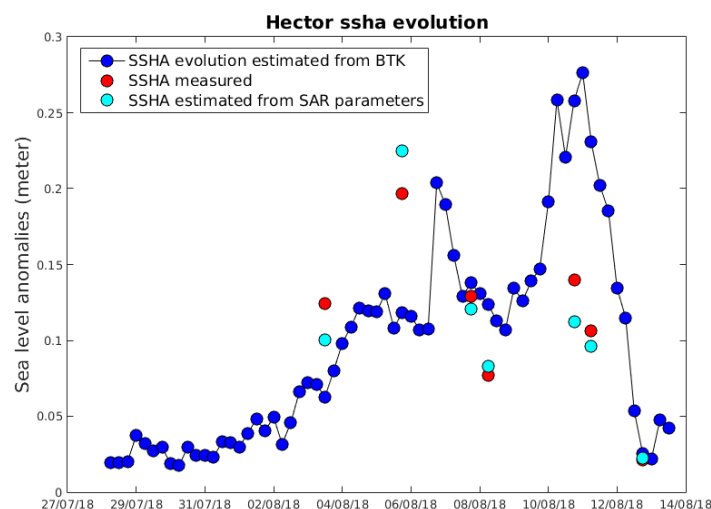


Figure 5.10: Same than Fig 5.9 but for Hector track and with SSHA estimates derived from SAR measurements for the cyan points.

This typical eastern Pacific cyclone is a midget storm that has followed a rapid intensification and has been monitored several times by SAR observations, it thus offers a good playground for comparison. Since this type of situation is difficult for BTK, in the absence of high-resolution instruments like SFMR data, it strongly relies on indirect IR-based methods (or flight level-based like AMSU) which are not surface measurements and also do not capture fast fluctuations, or on low to medium resolution microwave sensors with a limited spatial sampling of the inner-core (radiometers) or limited sampling of the intensity range (scatterometers). This case can bring some insights on the influence of the level of information of TC parameters in the modelling response. The Figure 5.10 compares the estimates given by the scaling laws inferred from BTK parameters (blue) and from SAR ones (cyan) with the measurements (red). In addition to a large dispersion, the BTK-based scaling no longer follows the fluctuations indicated by the altimeter measurements, especially for the second peak. As discussed in [Combote et al. 2020b], this case suffers from a large misinterpretation of its Rmax, with an estimation around 38 km in BTK instead of the 18 km measured by SAR for the period between the 8th and 12th august 2018. Hector and some passages of Jimena show the limits of using a 6-hourly analysis in favor of a more direct source of observations. Besides, as we have seen in

Chapter 3,  $R_{\max}$  is the parameter with the largest uncertainties among those estimated in the BTK, mainly for intense cases where the behavior of  $R_{\max}$  were not captured. For the operational products or analysis that use this parameter, it can trigger large discrepancies in the variables estimated [Young 1988; Irish et al. 2008]. We finally illustrate here one of the consequences of this effect for the SSHA, with some short examples.

Taking in consideration this last point, we decide to conduct a last experiment to verify this assertion with a larger panel of cases and by taking advantage of the work carried out in the previous part with the SAR database. The next section will present a zoom-study of the Eastern Pacific during the 2018 season and will conclude on our progressive efforts to establish a robust and consistent methodology for the wake analysis. The choice of this basin is not innocuous, as it includes many difficult cases due to their smaller size [Chavas and Lin 2016; Combot et al. 2020a], and due to their underlying trend to undergo fast fluctuations and different cycles of intensification [Knaff and DeMaria 2017]. Combot et al. [2020a] will definitely promote the capability of a full direct satellite observations constellation to study the TC wake.

### 5.3 Benefit of high-resolution wind forcing. (Article)

### Section 4.5. Space-based observations of surface signatures in the wake of the 2018 Eastern Pacific tropical cyclones

Authors: *Clément Combot, Yves Quilfen, Alexis Mouche, Jérôme Gourrion, Clément de Boyer Montégut, Bertrand Chapron, Jean Tournadre.*

Statement of main outcome: In this section, a new focus is given on the ocean response induced by Tropical Cyclones. Tropical Cyclones are among the most devastating and destructive natural hazards. Unfortunately, predicting the intensity and evolution of such individual event is still extremely difficult, owing to various internal and environmental factors, including interactions with the ocean interior. In that context, multiple satellite remote sensing observations are essential, and today, combined with denser ARGO interior measurements, the upper ocean responses to moving tropical cyclones can be more efficiently captured and monitored.

| Ref No. | Product name & type                                                                                                                                                                                                 | Documentation                                                                                                                                                                                                                                                                                                                                                   |
|---------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| 4.5.1   | SST_GLO_SST_L4_NRT_OBSERVATIONS_010_001                                                                                                                                                                             | PUM: <a href="https://marine.copernicus.eu/documents/PUM/CMEMS-SST-PUM-010-001.pdf">https://marine.copernicus.eu/documents/PUM/CMEMS-SST-PUM-010-001.pdf</a><br>QUID: <a href="https://resources.marine.copernicus.eu/documents/QUID/CMEMSOSI-QUID-010-001.pdf">https://resources.marine.copernicus.eu/documents/QUID/CMEMSOSI-QUID-010-001.pdf</a>             |
| 4.5.2   | SEALEVEL_GLO_PHY_L3_NRT_OBSERVATIONS_008_044                                                                                                                                                                        | PUM: <a href="https://marine.copernicus.eu/documents/PUM/CMEMS-SL-PUM-008-032-062.pdf">https://marine.copernicus.eu/documents/PUM/CMEMS-SL-PUM-008-032-062.pdf</a><br>QUID: <a href="https://resources.marine.copernicus.eu/documents/QUID/CMEMS-SL-QUID-008032-062.pdf">https://resources.marine.copernicus.eu/documents/QUID/CMEMS-SL-QUID-008032-062.pdf</a> |
| 4.5.3   | INSITU_GLO_TS_NRT_OBSERVATIONS_013_001_B                                                                                                                                                                            | PUM: <a href="https://marine.copernicus.eu/documents/PUM/CMEMS-INS-PUM-013.pdf">https://marine.copernicus.eu/documents/PUM/CMEMS-INS-PUM-013.pdf</a><br>QUID: <a href="https://resources.marine.copernicus.eu/documents/QUID/CMEMSINS-QUID-013.pdf">https://resources.marine.copernicus.eu/documents/QUID/CMEMSINS-QUID-013.pdf</a>                             |
| 4.5.4   | Non-CMEMS product: ISAS13-clim Monthly climatology of temperature and salinity profile used as a background stratification information, in case of argo absence. Developed by F.Gaillard & al 2016 (LOPS/ IFREMER). | Downloaded from: <a href="https://www.seanoe.org/data/00348/45945/">https://www.seanoe.org/data/00348/45945/</a><br>Reference paper: <a href="https://journals.ametsoc.org/doi/pdf/10.1175/JCLI-D-15-0028.1">https://journals.ametsoc.org/doi/pdf/10.1175/JCLI-D-15-0028.1</a>                                                                                  |

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|-------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| 4.5.5 | Non-CMEMS product: IBTrACS<br>Best-tracks are post-storm analyses at every 6-hours of Tropical and subtropical cyclones, giving several characteristic parameters (position, intensity, size) . They are produced by several dedicated centres and gathered in a homogeneous database named IBTrACS: International Best Track Archive for Climate Stewardship.                                 | Downloaded from: <a href="https://www.ncdc.noaa.gov/ibtracs/index.php?name=ibtracsdata-access">https://www.ncdc.noaa.gov/ibtracs/index.php?name=ibtracsdata-access</a><br>User Manual: <a href="ftp://eclipse.ncdc.noaa.gov/pub/ibtracs/v04r00/doc/IBTrACS_v04_column_documentation.pdf">ftp://eclipse.ncdc.noaa.gov/pub/ibtracs/v04r00/doc/IBTrACS_v04_column_documentation.pdf</a><br>Quality information: <a href="https://journals.ametsoc.org/doi/pdf/10.1175/2009BAMS2755.1">https://journals.ametsoc.org/doi/pdf/10.1175/2009BAMS2755.1</a> |
| 4.5.6 | Non-CMEMS product: Sarsentinel-1A/B wind field:<br>C-band radar with high resolution and dual polarisation (Co- & Cross-pol) signal. The wind product used was retrieved by the inversion scheme published by Mouche & al 2017 (LOPS/ IFREMER) from L1 sigma0 of ESA Copernicus. Data were collected in the context of ESA's SHOC campaign. (SHOC: Satellite Hurricane Observations Campaign). | - ESA Copernicus L1 sigma0: Copernicus open access hub ( <a href="https://scihub.copernicus.eu/">https://scihub.copernicus.eu/</a> ) and Sentinel-1 Mission Performance Center (MPC S-1)<br>- SAR's Tropical Cyclone Wind Product:<br>(Distribution site to come soon)<br>Contact: alexis.mouche@ifremer.fr                                                                                                                                                                                                                                        |

#### 4.5.1. Introduction

Over the Eastern Pacific region, 2018 has been a very active Tropical Cyclone season. The accumulated cyclone energy reaches a record value of  $316.10^4 \text{ kt}^2$  (NOAA, Annual 2018 report; Kruk and Schreck 2019). As shown in the background of Figure 4.5.1a, a persistent positive Sea Surface Temperature Anomaly (SSTA) up to  $1.5^\circ\text{C}$  extended over the entire region, to explain this very active 2018 season in combination with an enhanced convection (Kruk and Schreck 2019).

Associated to the extreme wind forcing conditions, distinctive localised impacts have long been attracting considerable attention. Indeed, quite systematically, a tropical cyclone will leave persistent signatures in its wake (Leipper 1967; Price 1981). The vigorous induced mixing and resulting intense upwelling generally cool the upper ocean mixed layer, stirring warm surface waters with colder waters below (Ginis 2002). Consequently, a tropical cyclone wake is generally characterised by a surface cold anomaly, possibly accompanied with nutrient blooms. Moreover, governed by intense isopycnal displacements (Geisler 1970), a tropical cyclone can also leave prominent sea-surface height anomalies in its wake. Resulting surface





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depressions can reach 0.3–0.5 m, depending upon the forcing intensity, size, translation speed, and ocean stratification conditions (Kudryavtsev et al. 2019a). Building on the actual satellite altimeter constellation (presently up to 6 satellites are available), satellite sea surface height estimates may more likely cross such trenches. Using both sea surface height and temperature observation, a more consistent view of the tropical cyclone characteristics can thus be obtained with additional support of a semi-empirical 2D model to interpret the oceanic answer. Here, the wakes of seven Eastern Pacific tropical systems are presented. To illustrate and further discuss the influence of the forcing parameters, the analysis mainly focuses on three particular major tropical cyclones: Hector (August), Lane (August) and Sergio (October). Hector experienced two intensification periods with maximum winds over 110 kt that lasted up to 7 days (NHC archive). Lane was the second wettest storm in USA territory with 1341 mm accumulated rainfall (NHC archive). Finally, Sergio left a particularly deep ocean signature (Figure 4.5.1). Above all, these three tropical cyclones are relevant cases as they benefit from longer time monitoring by satellites.

#### 4.5.2. Data analysis

A database has been set up that merges satellite observations, from altimeters, radiometers, and high-resolution Synthetic Aperture Radars (SAR), with Argo float data. The 6-hour Best-Track data set further provides several parameters over the tropical cyclone life such as position, intensity and specific wind radii (Product ref. 4.5.5). For each track, all available altimeter observations from the L3 CMEMS products were assembled for the different storm positions (Figure 4.5.1, product ref. 4.5.2), keeping only altimeter passes crossing the tropical cyclone wake afterwards. The well-defined trough left behind the tropical cyclones (blue shaded areas on the altimetry tracks in Figure 4.5.1) was used to directly estimate the sea surface height anomaly, i.e. the maximum difference between the bottom and the top of the altimeter sea surface height profile. As the tropical cyclone generates near-inertial oscillations (Geisler 1970), the altimeter may not sample the maximum trough value. For each Best-Track synoptic time, we thus kept the largest sea surface height anomalies among all the closest time/space colocated altimeter passes. When only one altimeter pass is available, the unique anomaly estimate was kept. This

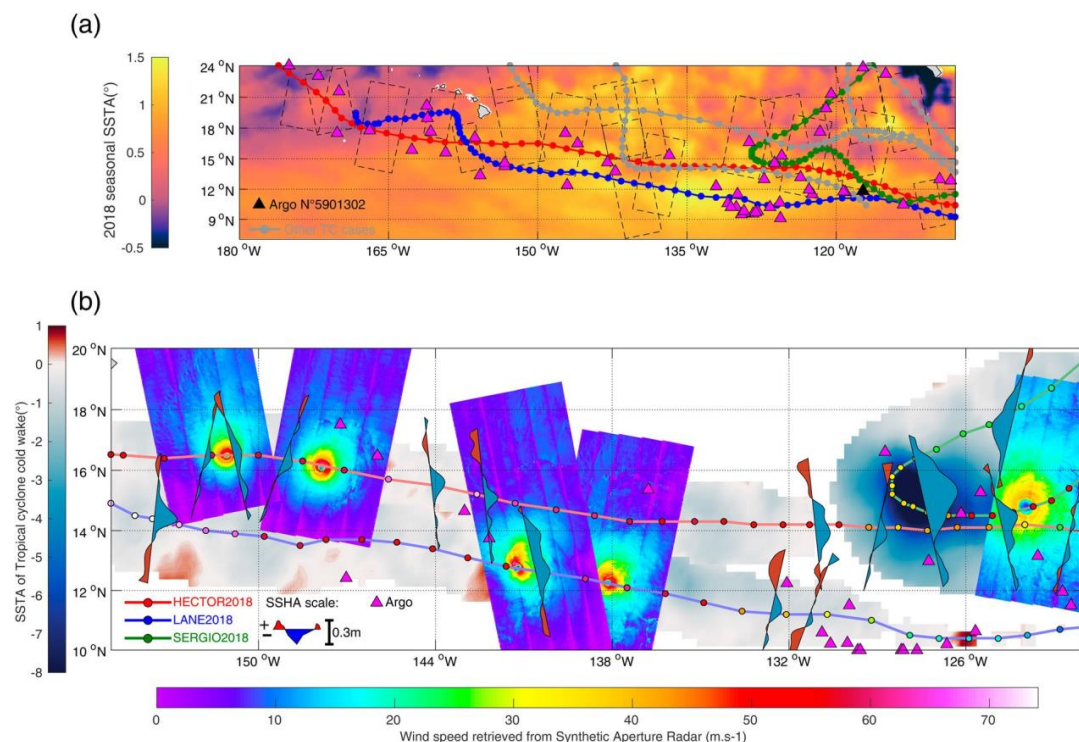


Figure 4.5.1. Constellation of satellite (product ref 4.5.1 & 4.5.2) and in-situ observations (product ref 4.5.3 & 4.5.4) for wake analysis over Eastern Pacific in 2018. (a) Overview of the database with all the tracks (coloured lines), SAR swaths (dashed contours) and Argo floats (magenta triangles) represented. The background field depicts the 2018 seasonal thermal anomalies (globally positive). (b) A zoom on Hector, Lane and Sergio. The 3 tropical cyclones wakes are represented by shaded areas for SSTA (see colorbar) as well as for SSHA anomalies (blue = negative, red = positive). Specific Sentinel 1A/B swaths were selected to map the tropical cyclone wind fields (product ref. 4.5.6). To clarify, not all the altimetry and SAR crossings available for the 3 tropical cyclones are displayed. Best-track information were added as the synoptic positions and intensities (coloured circles, product ref 4.5.5), with Hector and Lane heading westward and Sergio drawing a loop that ends northeast.

can thus introduce a slight underestimation of the altimeter trough.

The daily CMEMS L4 sea surface temperature product is used to estimate the thermal anomalies at each BestTrack synoptic position (product ref. 4.5.1). The cold wake intensity was estimated as the difference between a 2-week average of pre-storm sea surface temperature conditions and the daily post-storm sea surface temperature field. In case of pre-existing storms or loop trajectory, this pre-storm time window was reduced to withdraw the cooling of previous forcing. Shaded areas in Figure 4.5.1b trace the maximum cooling, from 1 to 3 days after the Best-Track synoptic time (Dare and McBride 2011; Vincent et al. 2012). Finally, using wind radii information from Best-Track, all Argo data inside the 34 kt wind radius were selected to infer the ocean stratification parameters (magenta triangles in Figure 4.5.1, product ref 4.5.3). From the nearest profile, we extracted the Brunt-Väisälä frequency  $N$  in the thermocline (Figure 4.5.2c), known to strongly influence the oceanic response to tropical cyclone forcing (Reul et al. 2014). In absence of Argo floats, the ISAS-13 climatology was used (product ref 4.5.4). As obtained in Figure 4.5.1b, rather moderate signatures are found for Hector and Lane, with maximum thermal and height anomalies amplitude of  $-1.7^\circ$  and  $-0.17$  m, respectively. A stronger cooling and a deeper trough are found for Sergio, with respective signatures larger than  $-7^\circ\text{C}$  and  $-0.3$  m. The first two cyclone cases were small intense storms with fast translation speed ( $V_{\text{fm}} = 7 \text{ m s}^{-1}$ ), while Sergio had a weaker wind intensity but is a large and slowly moving cyclone ( $V_{\text{fm}} = 4 \text{ m s}^{-1}$ ). As anticipated, the overall tropical cyclone induced-stirring process depends on its intensity, but also on its size and translation speed (Reul et al. 2014; Kudryavtsev et al. 2019a). The tropical cyclone wind structure is therefore of paramount importance, and the Best-Track data may not be precise enough to depict it, especially the determination of the maximum wind radius ( $R_{\text{max}}$ ) can be subject to large errors (Kossin et al. 2007; Knaff et al. 2015).

To overcome such a limitation, SAR measurements have therefore been used (product ref. 4.5.6), which can provide unique observations of high resolution wind patterns in tropical cyclones (Mouche et al. 2019). As collected, 1-km resolution data from Copernicus/ESA's Sentinel 1A/B operating in C-band and VH cross-polarisation provide reliable surface winds up to  $70 \text{ m s}^{-1}$  (Mouche et al. 2017; Mouche et al. 2019). A total of 23 SAR scenes were collected with maximum winds ranging from 25 to  $70 \text{ m s}^{-1}$  (Figure 4.5.2a), representing

small to medium size systems (Figure 4.5.2b).

For each SAR scene, an objective analysis was used to determine the centre of the storm and to extract the maximum wind speed ( $V_{\text{max}}$ ) and radius of maximum wind information (Combot et al. 2020). These estimates are compared to the Best-Track ones in Figures 4.5.2a, b. An excellent consistency is found for the maximum wind speed parameter, with a correlation coefficient of 0.95, a very weak normalised bias, and a root mean square difference of about  $4 \text{ m s}^{-1}$ , that is partly associated with the discretization of the Best-Track estimates. For the radius of maximum winds, the correlation is significantly lower but still high (0.80). Yet, compared to Best-Track values, SAR estimates are not discrete and therefore more likely to provide accurate values for small size TCs ( $R_{\text{max}} < 30 \text{ km}$ , Combot et al. 2020).

Pre-storm ocean stratification and heat content are key variables that determine ocean/atmosphere feedback mechanisms and the associated evolution of tropical cyclones. The lower panels in Figure 4.5.2 show the temperature, salinity and density profiles measured by an Argo float a few days before ( $\sim 7$  days, in blue) and after ( $\sim 3$  days, in magenta) Sergio's passage in category 4 at that time. The corresponding ISAS profiles are also displayed for the two climatological months close to the Argo profile time (black lines). As obtained from the prestorm Argo profile, the ocean stratification was much stronger than the climatological values (values indicated in the lower right panel), mainly due to a 30-m thick surface layer much fresher than usual, which is confirmed by the surrounding floats. As a result, mixing and associated salinity and temperature anomalies are reduced, and the stratification remains very strong during and after Sergio's passage which is an important limiting factor for oceanic feedback on tropical cyclones evolution.

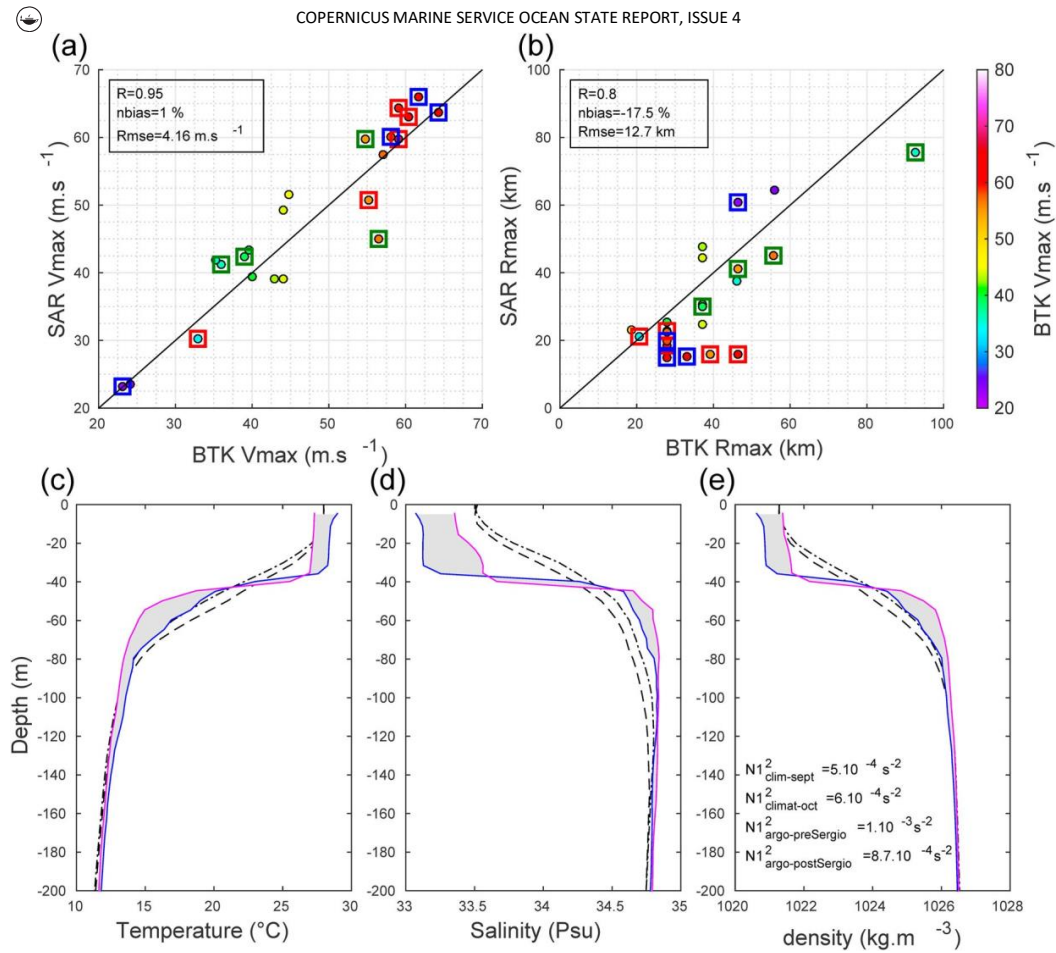


Figure 4.5.2. Upper panel: Comparison between Best-Track (product ref 4.5.5) and SAR (product ref 4.5.6). Maximum wind speed (a) and radius of maximum wind speed (b) with dots coloured as a function of Vmax. Following the Figure 4.5.1 convention, the three particular TCs dots are notified with coloured squares, red = Hector, blue = Lane, green = Sergio. Lower panel: Profiles of temperature (c), salinity (d) and density (e) of Argo and ISAS climatology (product ref 4.5.3 & 4.5.4) at the Sergio's location on 3rd October 2018 at 13h40, when it was particularly strong ( $\sim 60 \text{ m s}^{-1}$ ). The two surrounding months, September (dashed line) and October (dashed dotted line), from the ISAS climatology profiles are shown. Two Argo profiles were selected, one before (26th September 2018 in blue) and one after (6th October 2018 in magenta) the Sergio path. Shaded areas illustrate the anomalies left by Sergio, a slight deepening ( $\sim 10 \text{ m}$ ) and cooling ( $\sim 2^\circ \text{C}$ ) of the mixed layer are observed. These profiles were acquired near  $117^\circ \text{W}$  and  $12^\circ \text{N}$  (black triangle in Figure 4.5.1.a, profiler number: 5901302).

#### 4.5.3. Analysis from a semi-empirical model

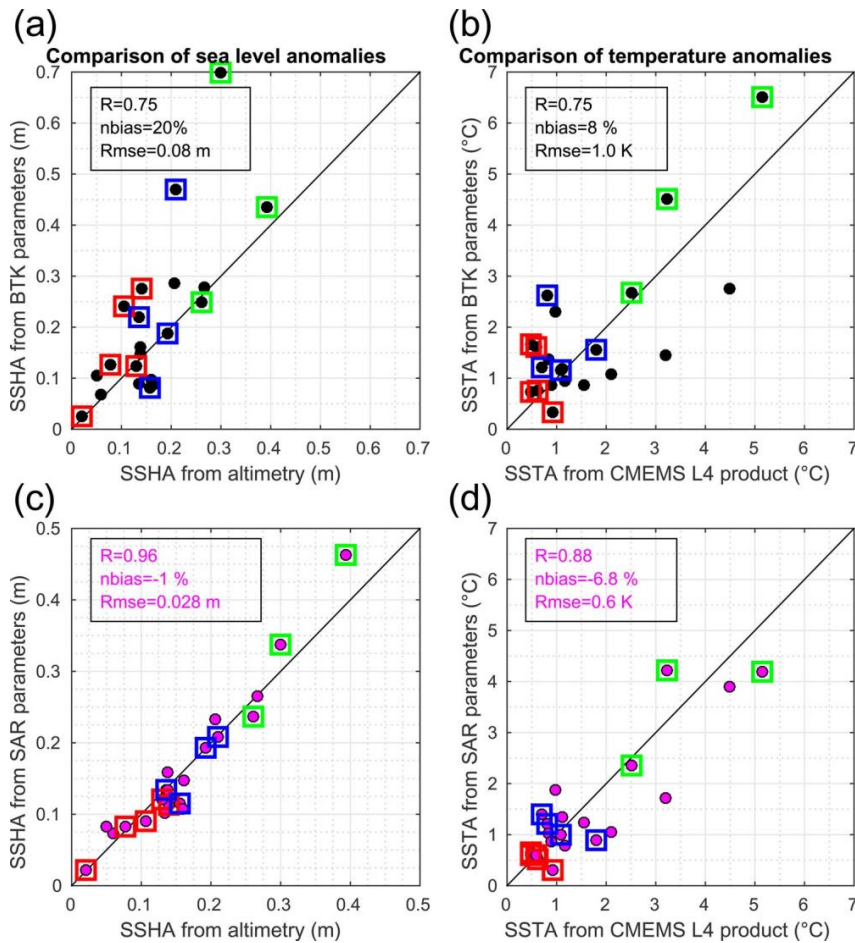
Following Geisler (1970) and Price (1981), Kudryavtsev et al. (2019a) proposed a semi-empirical framework to jointly analyse and interpret tropical cyclones sea surface temperature and height anomalies. It is assumed that the ocean response to a moving TC is largely dominated by baroclinic effects. Considering this assumption, scaling laws issued from the semi-empirical modelling approach developed by Kudryavtsev et al. (2019) can be expressed as:

$$\frac{SSTA}{V_{max} \cdot N^{(3/2)} / (g \cdot \alpha \cdot f^{(1/2)})} \propto \left( \frac{V_{fm}}{f \cdot R_{max}} \right)^{-1} \quad (1)$$

$$\frac{g \cdot SSHA}{V_{max}^2} \propto \frac{R_{max} \cdot N}{V_{fm}} \quad (2)$$

with  $f$ , the Coriolis parameter,  $V_{fm}$  the translation speed,  $\alpha$  the thermal expansion coefficient,  $g$  the gravitational

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within the maximum wind region, SAR estimates give a significantly closer agreement with the



**Figure 4.5.3.** Comparison of Sea Surface Height (left column, product ref 4.5.2) and Sea Surface Temperature (right column, product ref 4.5.3) anomalies between satellite measurements (x-axis) and estimation from the semi-empirical model (y-axis): (a) and (b) represent the estimates inferred from Best-Track information (product ref 4.5.5); (c) and (d) from SAR-derived wind field (product ref 4.5.6). The 3 reference cases are reported according to the same convention than in Figure 4.5.2. The required spatial collocation between SAR and altimetry has slightly reduced the numbers of points (21) as used for Figure 4.5.2(a,b).

acceleration on Earth, and  $N$  the Brunt-Väisälä frequency. Best-Track and SAR offer two different sets of  $V_{max}/R_{max}$  estimates, giving different semi-empirical predictions that are compared in Figure 4.5.3. Overall, the results agree well with large correlation between observations and predictions ( $R \geq 0.75$ ). These correlation coefficients are significant at the 95% confidence level but, due to the rather small sample size, this cannot be accurately assessed. However, ongoing research confirms these findings with a much larger sample size including  $N > 100$  SAR scenes collocated with TCs for the recent years and over all the ocean basins. Bringing improved high-resolution description

proposed scaling laws ( $R \geq 0.88$ ). The Hector case (red squares) remarkably illustrates the improvement, as large maximum wind radius discrepancies were indeed observed between the two data sources (Figure 4.5.2b). As also shown in Figure 4.5.3, the Sergio case highlights a larger dynamical range for its wake signature, going from  $2.5^\circ\text{C}$  to  $5.2^\circ\text{C}$  cooling and from 0.26 to 0.39 m sea level drop. This overall stronger induced oceanic answer arises from its larger size and lower translation speed and is well accounted for by the semi-empirical model when fed by the SAR observations. This is particularly true with regard to the two points departing the most from the 1:1 line, Sergio and Lane (Figure





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4.5.3a), for which overestimation of Best-Tracks for maximum wind speed and its radius (Figure 4.5.2) led to unrealistic predictions of sea surface anomalies.

The proposed interpretation framework can thus guide the combined use of sea surface temperature and height amplitude changes measured in the wake of tropical cyclones. It can help to inform about the resulting strength of hurricane-induced mixing and upwelling. This opens a bottom-up approach to guide future investigations to help document the resulting wind forcing and practical drag coefficient under extreme tropical cyclone conditions (Kudryavtsev et al. 2019a, 2019b).

#### 4.5.4. Summary

This study highlights the use of multi-platform observations for the analysis of tropical cyclones and their induced oceanic answer. As a main outcome, it is shown that the combination of low and high resolution satellite sensors are of paramount importance to better depict and monitor the tropical cyclones wind patterns, and to interpret the air/sea coupling that imprints the cyclone wake. In particular the study shows that altimeter and SAR measurements can provide unique information to help analysis and monitoring of extreme events. The wide-swath altimetry SWOT mission foreseen in 2021 will be a unique opportunity for further research and applications, notably since it will provide a 2D mapping of sea level anomalies left by tropical cyclones.

#### Acknowledgements

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# TC wake analysis: interpretation & cross-comparison of the variability of surface anomalies.

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## 6.1 Global scaling of the cold wake anomalies.

Chapter 5 demonstrated the ability of scaling laws to infer the oceanic response to the passage of a tropical cyclone, based on parameters that synthesise both the description of the forcing and the pre-existing ocean structure.

The performance of these temperature and sea level analyses has shown a significant dependence on the level of information in the observations used, i.e. whether they are under track or interpolated data, climatologies and analysis products or highly resolved in-situ measurements. Equally crucial is the methodology required for the exploitation of this information, which requires special efforts to obtain readable and relevant signatures. After several successful experiments, our latest approach takes maximum advantage of direct observations <sup>1</sup>, and recommends their use. Of course, these various findings were the result of work carried out on a limited number of cases from a single study-basin, the Eastern Pacific in 2018 [Combot et al. 2020b].

In order to bring further our statements and to confirm our multi-platform approach, we will in this last chapter explore deeper the oceanic response by extending our wake

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<sup>1</sup>in the sense of under track or in-situ measurements

analysis to all existing basins. In this way, we will be able to demonstrate the incredible complicity between the different space and in-situ observations, a true three-dimensional experimental laboratory that will allow us to characterise a wide range of cyclonic signatures. Finally, if the previous results have fully demonstrated the great altimetry coverage, this chapter will also highlight the important collection of Argo data, which gives us direct information on the pre-cyclonic ocean structure.

### 6.1.1 Description of the complete database.

In this respect, Figure 6.1 shows the extent of our SAR database (Sentinel 1A/B or Radarsat-2) benefiting simultaneously from an Argo measurement (triangles) or, failing that, from information from ISAS climatology (circles). Of the 169 SAR images, no less than 132 cases can benefit from in-situ information (i.e. 80%), which demonstrate the importance of the current fleet of instruments. The R17 extent provided by the satellite observations determine the collection area. The closest float from TC center is used as a reference for the geophysical variables present in the scaling laws, with however an initial priority attributed to those located in the inertial resonance region (to the right/left of the track in the northern/southern hemisphere), where the maximum response is generally observed. In the absence of floats, ISAS climatology is used.

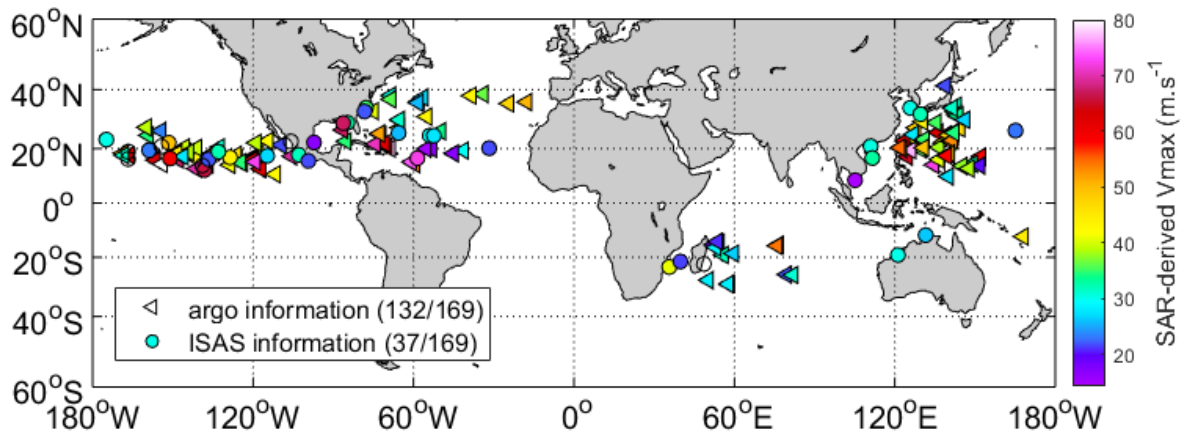


Figure 6.1: *Geographical distribution of cases benefiting from Argo float data (colored triangle) or from ISAS-15 climatology (colored circles) for our complete SAR database, with maximum surface wind intensity in colorbar.*

The entire database could not be exploited and our sampling was reduced to 135 TC footprints for 111 collocations with argo measurements (% coverage unchanged), due to too coastal cases that are difficult to exploit, cyclone tracks that are too northerly to allow cross-track altimetry measurements, or because of too scattered low-pressure systems for which the associated responses are indistinguishable from the mean ocean activity (1-2 cm only of trough). As these different constraints can be asymmetric in their effects on the measurements, some cases present only one of the two observables (SSTA/SSHA), with respectively 8 situations of uncorrelated sea level and temperature anomalies observed. Thus, out of our 135 altimetric and radiometric measurements, 127 situations are common to both observations (94%), thus ensuring that most of our sample has a complete and coherent view of the wake. As indicated in Chapter 4, no SAR observations had been made in the North Indian Basin at the time of the data processing, and this gap was filled

by more recent acquisitions outside the time frame of our study.

To complete the description of our database, it is also worthwhile to make a small clarification on the global temperature products. In the course of our various experiments to study surface cooling, several data were tested, in particular in the framework of the CMEMS study where the use of European products was strongly requested. On this occasion we changed our initial data source (REMSS) to use the aggregated measurements of the OSTIA product, which incorporates in-situ data in addition to remote sensing data. This constraint gave us the opportunity to check the consistency between these different sources of information with respect to the scaling law, as presented in Figure 6.2. In this perspective, the two versions of the daily REMSS product were also confronted to the analysis, namely the one using only MW satellite measurements and the one blending them with IR observations. Figure 6.2 compares the measurements of the three products on panels a, b and c, and shows their overall good performance. A slight heterogeneity in the correlations is observed, with a lower value attributed to the OSTIA product, which also suffers from a higher normalised bias, with lower measured SSTA values than those reported by the two versions of the REMSS product.

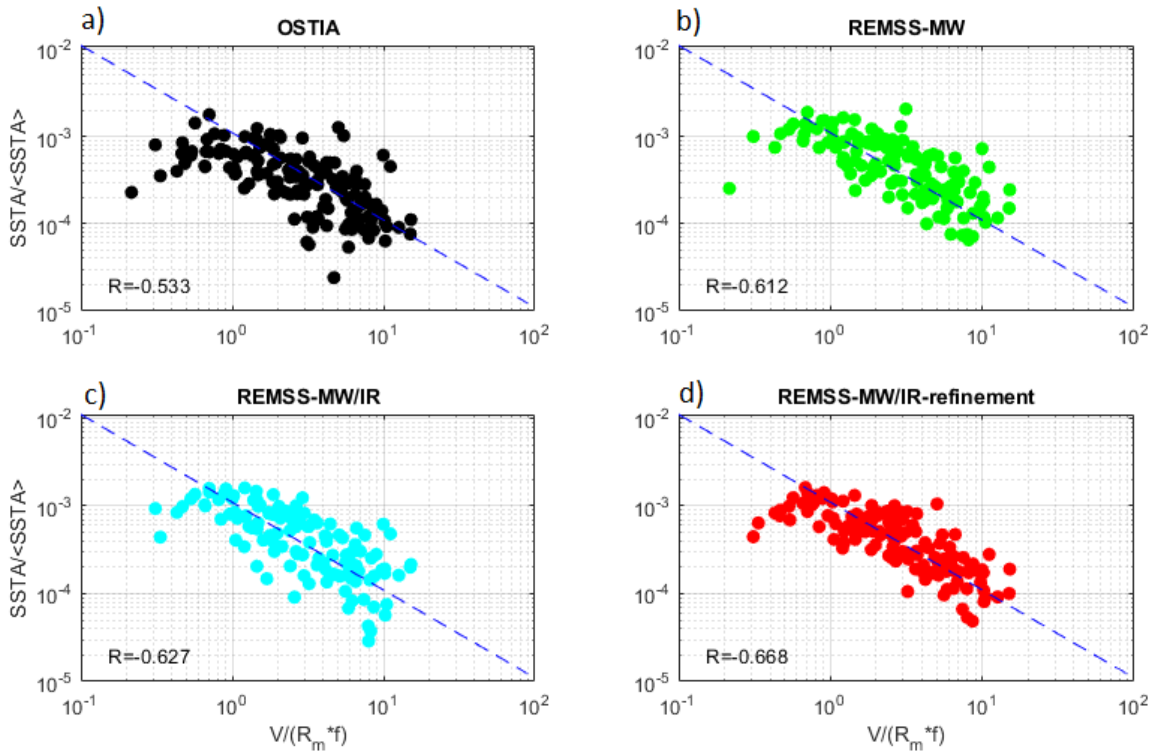


Figure 6.2: Visualization of the surface temperature anomalies scaled by as a function of the cyclonic rossby number, for the different daily products used: a) OSTIA, b) REMSS-MW, c) REMSS-MW/IR, and d) REMSS-MW/IR but with a filtering control

These results may stem from a consistency problem between the spatial-temporal scales involved in the ocean response and those involved in the OSTIA temperature description, where argo float information prevails. Due to the interpolation methods used (temporal correlation matrix) and the properties of the in-situ data (repetition cycle, horizontal coverage), it is possible that the maximum cooling values are underestimated. These observations are of course only hypotheses and do not allow to determine any hierarchy in the quality of the products, nor in their performances for the complete horizontal

analysis of the cold wake. However, in the analysis of the temperature response maxima, the use of purely satellite products seems more appropriate and our final choice was therefore naturally the REMSS daily products, more particularly the one mixing both IR and MW strengths. When a double control of the filtering of the peaks is carried out (elimination of the spurious peaks), this blended daily field provides a finer capture of the maximum values of the cooling and observes a better behaviour with the temperature scaling law (panel d). This quick comparison is just a parenthesis, which allows to close the one opened by the use of different products in Chapter 5, and to justify our final choice of data.

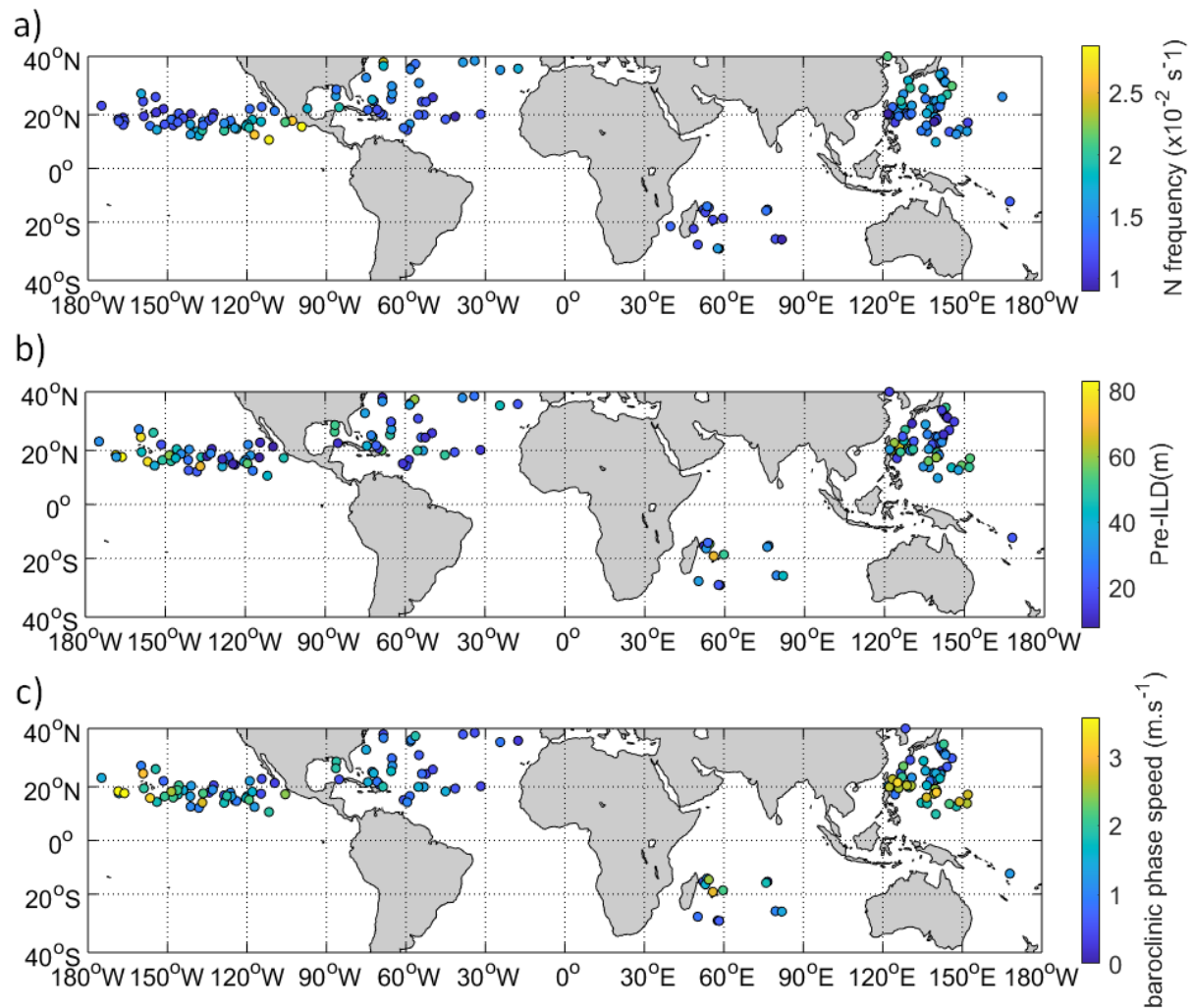


Figure 6.3: *Global distribution of the brunt-Väisälä frequency of the seasonal thermocline  $N1$  (a), the pre-cyclonic mixing layer (b) and the phase velocity of the 1st baroclinic mode (c) for the whole database.*

Finally, to conclude the description of our complete database, Figure 6.3 presents the distribution of some key ocean variables, necessary for understanding the wake dynamics. The distribution of the forcing parameters was carried out in Chapter 3 ( $V_{\text{max}}/R_{\text{max}}$ ) and 4 ( $V_{\text{fm}}$ ). From this in-situ information, we can accurately describe the stratification parameters needed for the scaling laws (panel 1), such as the brunt-Vaisala frequency of the seasonal thermocline, which we have already used in the previous chapter and which turns out to be an essential parameter to describe the turbulent mixing and the propagation of near-inertial waves. Ancillary parameters can be extracted in order to better

understand the oceanic response (panel 2 and 3), such as the initial mixing layer which also informs us about particular situations such as barrier layer zones, or deep ML zones (warm eddies...) which influence the wake dynamics. In addition, the depth is also essential in the calculation of the 1st baroclinic mode velocity (panel c), which, when compared to the translation speed of the cyclone, allows us to characterise the nature of the oceanic response, i.e. whether the energy of the forcing is rather transferred to the inertial circulation (baroclinic regime) or to the geostrophic surface circulation (barotropic regime). Thus for Froude number values below 1, there will be no inertial wake observed but a quasi-stationary non-oscillating circulation with a simple trough elongated along the track.

The North-West and East Pacific basins also contain several deep mixed layer situations, with associated baroclinic velocity values that can be quite high, which in the case of relatively slow TC movement can lead to barotropic regime situations or a non-negligible barotropic component. In addition, there are particularly strong stratifications in the low latitudes of the East Pacific, thereby increasing the variability of a basin that is often considered complex to study. It should be noted that several cases have been added in this basin compared to the previous analysis, due to restrictions linked to the CMEMS study (European product, only TCs with several acquisitions). The incorporation of Radarsat-2 data and some Sentinel-1A/B acquisitions complete the picture.

### 6.1.2 Scaling laws

In order to corroborate the results put forward in CMEMS, Figure 6.4 presents the SSH (panels a, b, e, f) and SST (panels c, d, g, h) scaling laws, similar to those used in both [Kudryavtsev et al. 2019b] and Combot et al. [2020b], and derived from the information of the Best-tracks (panels a, c, e, g) or the SAR instruments (panels b, d, f, h). This Figure shows again, but for all basins, the benefit of the SAR measurements in the reading of the wake anomalies and is underpinned by the observed correlations. Although the difference is slightly smaller for the SST between the two forcing sources, it remains higher with the variables extracted from the high-resolution wind field, especially for the SSH, which also benefits from a much smaller dispersion.

This database is very different from the one introduced in Chapter 5.2 (see Figure 5.8) with our final methodology, where the scaling laws fed with BTK information seemed to provide good statistical performances, better than those presented here with the same data archive; innately, the reason is deeply rooted in the nature of the cases selected. Previously, we carried out a major selection process in order to keep only the "perfect" cases, i.e. those with simple characteristics: like a low variability in terms of forcing parameters, average and relatively stable translation speeds, very clean and mainly zonal trajectories, as well as cases that are spatially and temporally spaced from all other cyclonic interactions. To this end, a large proportion of the wakes studied in Chapter 5.2 came from the Atlantic region, which also has a better observation network [Black et al. 2007; Foltz et al. 2019], as shown by the quality of the BTKs for this region [Landsea and Franklin 2013]. Here, we follow a somewhat opposite approach, constrained both by the SAR observation and by our desire to broaden our field of study to the plurality of existing cases, in order to highlight the need for more direct observations. Several complex situations are therefore included in our sample, in particular several systems offering very strong evolution in terms of intensity, size and motion speed, and possibly evolving in turbulent pre-cyclonic conditions. The b/d or f/h panels illustrate the contribution of these finer measurements for a better representation of the spectrum of oceanic responses.



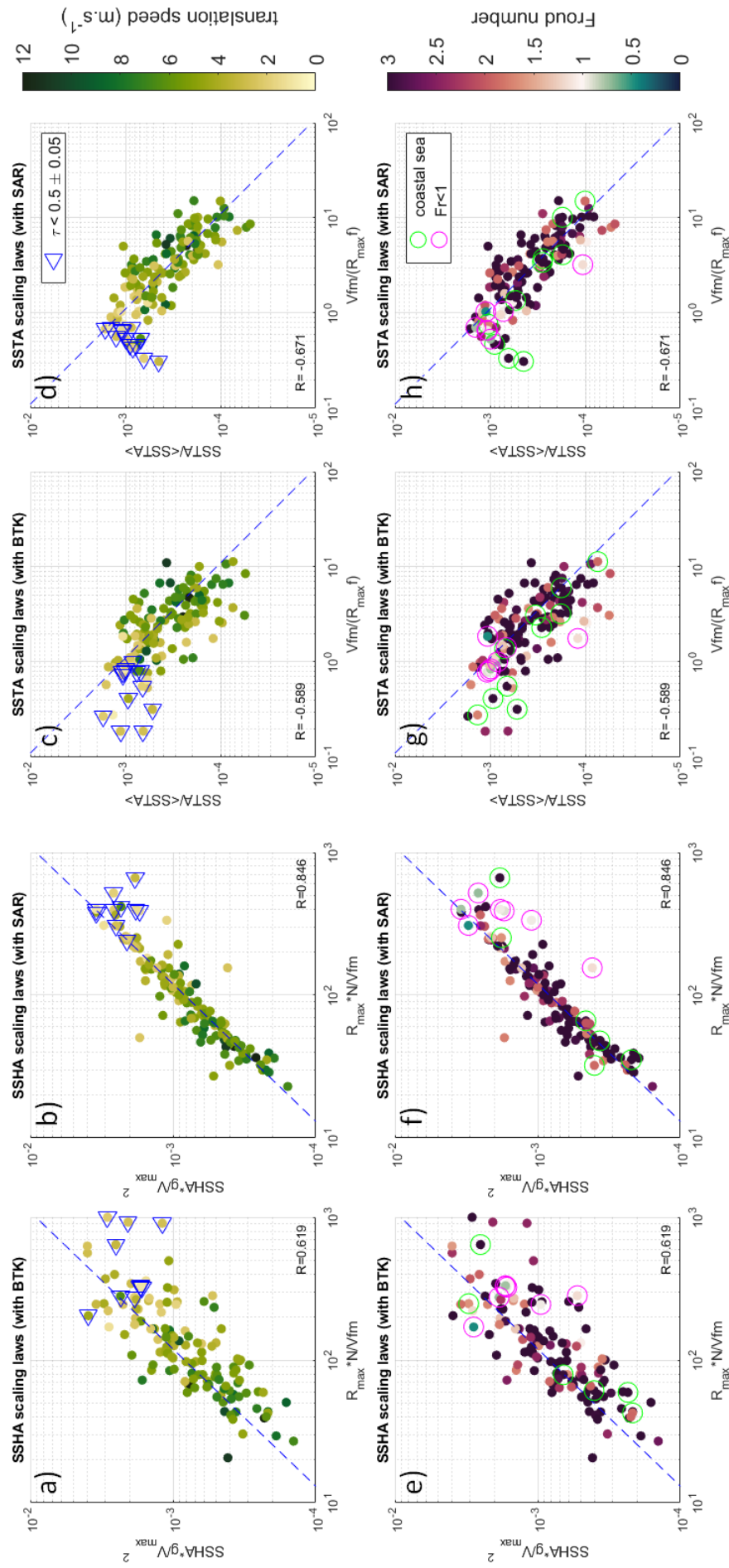


Figure 6.4: Comparison of SSHA (a,b,e,f) and SSTA (c,d,g,h) scaling laws documented by SAR measurements (b,d,f,h) or BTK information (a,c,e,g), with indications of translation speed (upper panels), or Froude number (lower panels). The boundary situations corresponding to the development of a barotropic-like response are indicated by pink circles for sub-critical translation speeds, by blue triangles for long residence times, and by green circles for cases in coastal environments.

As it is an important parameter for classifying the ocean response regime, the translation velocity serves as the colorimetry for the upper panel. It shows a relatively monotonic relationship with the two scaling laws, where progressively slower translation velocities are observed for increasing amplitudes, in agreement with many previous studies. This trend can be observed simultaneously for the SSH and the SST. We can also observe a stalling of some of the slowest systems, where notable divergences occur around the regression laws. These disparities can be explained in several ways, notably through the analysis of the Froude number. When the translation speed is scaled by the 1st mode velocity, we observe mostly situations where this ratio is more than double, which induces a mainly baroclinic response and the observation of an inertial wake. On the other hand, the divergent situations show Froude number below the unit value, especially for the SSH, which implies a rather barotropic response and the absence or the decline of the inertial response. These situations therefore fall outside the validity of the scaling laws, which are a simplification of the ocean response model and only take into account the baroclinic component [Kudryavtsev et al. 2019a].

For the SST, these deviations are better explained by the presence of a shallow vertical oceanic structure near from the coastal environment. Such a situation is conducive to the amplification of a depth-averaged current, as demonstrated by the Jansen parameterisation [Jansen et al. 2010], where the cyclone no longer sees the ocean as stratified but as a single thin layer, for which the barotropic response, that already propagates quickly over the whole column, is therefore more effective and strongly manifests itself at the surface. These two effects have the consequence of inhibiting or diminishing the contribution of the inertial response. They are limiting cases and can be easily identified with the two markers <sup>2</sup> used in Figures 6.4 e/f/g/h. If they induce a different dynamic, one should not forget the contribution of advection in this surface circulation, a term neglected due to the assumptions of the model and de facto absent from the scaling laws. Its influence can however be critical for the most stationary systems. These cases are generally associated with Froude numbers very close to or less than 1 and are therefore already identified as barotropic. However, size is a factor that can also significantly exacerbate Ekman transport within the upper ocean, thus bringing a lot of non-linearity to the behaviour of geophysical variables. The Ekman pumping subsequent to the divergence of a very large (and/or slow) cyclonic transport has time to set up as a dominant stationary wind stress curl signature [Jullien et al. 2012] rather than an oscillatory response [Lin et al. 2003]. This general upwelling geostrophically adjusts with the surface barotropic circulation and amplifies its contribution. This is therefore a potentially very limiting situation, as it is not taken into account in the model assumptions.

In order to identify this type of situation, the ratio between the inertial period and the residence time of the cyclone can be used, like in Reul et al. [2014], to show the influence and competition of upwelling in the mixing process. This dimensionless number, close to cyclonic Rossby number, puts more weight on the size parameter. Most of the long residence times ( $\tau < 0.5$ ) are of course correlated with subcritical displacement velocities ( $Fr < 1$ ), although a few more cases can be identified due to a much larger than normal  $R_{max}$ . All cases identified by any of these three markers represent situations outside the context of the scaling laws, and are therefore removed from the analysis of the ocean baroclinic response.

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<sup>2</sup>Some cases are both in areas where  $H < 2000$  m and  $Fr < 1$ .

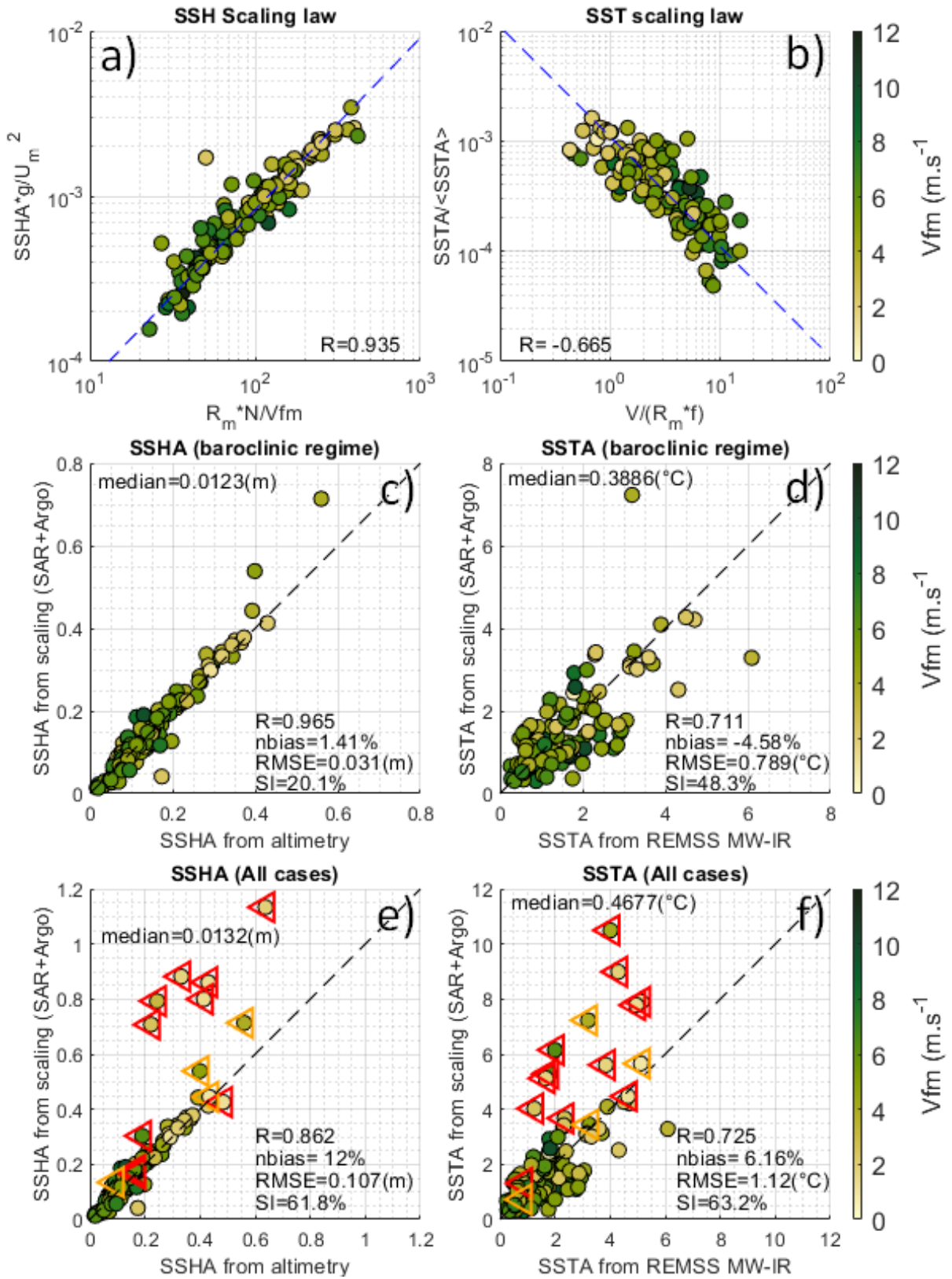


Figure 6.5: Analysis of scaling laws (a,b) and anomaly inversions (c,d) for the SSHa (a,c) and SSTa (b,d) in the baroclinic regime (a,b,c,d). Anomaly inversion for all cases (baroclinic+barotropic regime) for SSHa and SSTa are also shown in panel (e) and (f) respectively. The red triangles indicate the limit cases (barotropic response) and the orange triangles the transient cases ( $Fr$  and  $c1 \sim 1$ ).

A filtering of these 3 situations is performed to only keep the situations with a dominant inertial response (Figs. 6.5 a) and b), which mainly increases the relationship in SSH ( $R=0.935$ ) but implies little change for that of SST, except in terms of dispersion. Similar to what is presented in the paper [Combot et al. \[2020b\]](#), an inversion of the scaling laws is then performed to estimate the two geophysical variables of the ocean response (Fig. 6.5 c and d). It can be seen that the two laws established by [Kudryavtsev et al. \[2019b\]](#) prove to be very consistent and robust in estimating the SSH and SST anomalies, as revealed previously in Chapter ?? and now with the entire SAR database, where we find strong correlations ( $>0.7$ ) and almost negligible biases ( $<|5|$  %), especially for SSH. A large range of oceanic responses is thus correctly captured by the model, for diverse pre-cyclonic conditions (Figure 6.3) and forcing situations. Several observations can be made about the inversion of the scaling laws:

- 1) The estimation of the cyclone-induced sea level anomalies leads to an excellent correlation ( $R=0.965$ ) with the values measured by the altimeter constellation, under the assumptions of the baroclinic response of the model. 93% ( $R^2$ ) of the variability is thus explained by the the inertial wake spreading mechanisms and its resonance with the shear currents during turbulent mixing. This statement is reinforced by the negligible bias (1.41%) and the very low RMSE values ( $\sim 3$  cm) observed, even if they are slightly increased by the values of the two extremums above 0.5 m. They correspond to interesting transient situations ( $Fr \sim 1$  and  $T \sim 0.5$ ) that we will discuss in section 6.3. This slight inflection that they cause in the statistical properties can be felt especially when compared with the median of the differences between the estimated and measured values, where we observe a lower value (1.2 cm) than that of the RMSE, a sign of a slight dispersion of the data around the observations (20%), mainly driven by these two ambiguities. These two indicators (RMSE and median) are also equivalent to the known error of the altimeter measurement ( $\sim 2$  cm), which is a further argument for the excellent agreement between the observed and estimated inertial wake troughs. The variability of the SSH is therefore mostly understood by the scaling law. This first observation allows us to validate the hypotheses that lead to the regression laws.
- 2) As for the estimation of temperature anomalies in the wake, it leads to a weaker correlation with the observations, even if it remains very robust ( $R=0.71$ ). The erosion of the thermocline by the entrainment process during turbulent mixing seems to explain 50% of the variability of the wake cooling. This estimate is slightly lower than the values suggested by the studies of [Price \[1981\]](#) and [Vincent et al. \[2012a\]](#), when considering the contribution of this term over all categories (60%-70%). The representativeness of our sample cannot be questioned, given the distribution of the different parameters framing the response, and the filtering of the barotropic cases. However, this low value may be the result of certain hypotheses leading to the scaling variables, especially since the results are marred by a significant scattering ( $\sim 50$  %), proof of a potential lack of dynamics in the SSTA law. This is undoubtedly due to the lack of description of the pre-cyclonic MLD, which reflects the efficiency of the mixing as demonstrated by [Vincent et al. \[2012a\]](#) and the control it exerts on the ocean thermal response. The cooling inhibition by deep ML or BL phenomena are now well documented. This aspect of the thermal response was already pointed out in [Kudryavtsev et al. \[2019b\]](#) and tends to be confirmed here.
- 3) In addition, these results highlight the variability of the baroclinic response, whose dynamics range from 0 to 0.6 metres of trough and  $0^\circ$  to  $6^\circ\text{C}$  of surface cooling. This latter range of values is in agreement with the pioneering studies of the kind,



which rarely report higher values for standard forcing conditions (i.e. non-stationary and no bifurcations), as cyclone-induced cooling has long been confined to this measurement range [Bender et al. 1993]. As for the SSH, few studies have focused on the amplitude of the response itself. The few existing analyses have been clustered around interpolated products and climatology, and rarely report observations under the track, except for a few specific cases. The study by Mei et al. [2013], however, shows values up to 0.5 m of trough amplitude, in agreement with the limits evoked by Geisler [1970] and the review by Ginis [2002] around 0.5 to 0.6 m. Apart from our study and the related one by Kudryavtsev et al. [2019a,b], the estimation of the TC trench amplitude and its evolution as a function of the coupling parameters is thus relatively new, and suffers from few comparisons to be able to feed these findings further. Nevertheless, it can be stated that these values are very consistent with those in the already existing literature, and that these observed scales correspond essentially to the baroclinic response, via inertial wake propagation.

The observation of cooling above 6°C requires, on the other hand, the concurrence of a generalised upwelling event, whether directly induced by a quasi-stationary translation speed ( $\sim 1 \text{ m.s}^{-1}$ ) or artificially reproduced by bifurcation zones in the TC trajectory, provided that the parameters of the forcing (intense or broad) and of the oceanic structure are also favourable, as highlighted by several case studies mentioned in sections 1.3 and 1.4. Zhang et al. [2020] noted that the most impressive responses are systematically observed for sub-critical translation speeds ( $\text{Fr} < 1$ ), as evidenced by the 14°C cooling episode in Blanca in 2015. Such amplitudes cannot therefore take place without the development of an appropriate and adjusted barotropic response at the surface with an intense Ekman circulation.

In that respect, Figure 6.5 e and f show the SSHA and SSTA inversions for the full range of responses, baroclinic and barotropic, without prior filtering of limiting events. From a statistical point of view, they show much more contrasted results than for the purely baroclinic responses, with significant differences in RMSE and dispersions ( $\sim 60\%$ ). The SSTA measurements, on the other hand, do not exceed 6°C for the limiting cases either. Unfortunately for our SAR database, the context was not conducive to the appearance of temperature anomalies of this calibre. Few of our cases approach quasi-stationary translation speeds, or are found in a tight bifurcation zone (loop). The rare systems corresponding to one of these two criteria are either located too far upstream or downstream of the yaw paths, or are too weak in intensity (below the cyclonic threshold) to lead to an exacerbated response. The measurements related to these limiting situations (in abscissa) therefore remain confined to the same intervals as those observed in Figures 6.5 c and d. Conversely, the model estimates for these cases diverge greatly from the observations and provide significantly overestimated values for both geophysical variables, above the 0.6 m and 6°C thresholds. As a result, we note the appearance of a notable bias for SSHA around 12% (compared to 1.41 %), as well as its change of sign for SSTA, from -4.58% to 6.16%, proof of a generalised overestimation of the responses for these cases at the margin of the model assumptions. We also note the notable divergence between the RMSE values and those of the medians, which remained substantially close to those previously calculated (see Fig. 6.5 c and d). This significant difference underlines that most of the estimates still remain confined to minimal errors, close to those of the instrumentation, as also shown by the still very high, if not unchanged, correlations of the two cold wake anomalies.

In the case of the scattering indexes, they are this time quasi-similar between the SSHA and the SSTA, both dominated by the dynamics of the extreme points. Several outliers can be observed for the SSHA and mark a real break with the situations operating in the



validity framework of our analysis. This observation is more nuanced for the SSTA, due to the more dispersed nature of the results, even if the borderline cases are gradually moving away from the observed trend. This behaviour reflects the non-linearity of the oceanic response for events under the increasing influence of the barotropic regime or favourable to its development.

Apart from our collection of SAR images, several anomalies higher than 6°C can be observed for our time frame (2010-2018), notably through the examples of Goni in 2015, Sergio in 2018 [Combote et al. 2020b] or Noru in 2017, with anomalies higher than 7°, 8° and 9°C respectively. They all correspond to stationary events, with a Froude value below 1 and an intensity above the cyclonic threshold. Some of these storms could have been captured at other times by SAR observations, but were not in the immediate vicinity of the areas concerned. This collection of predominantly barotropic responses thus covers a temperature scale ranging from 0°C to about ten degrees (14°C with Blanca). On the other hand, if the sea level depressions easily exceed 40 centimetres, they seem to be capped around troughs of 0.7 to 0.8 metres, and do not exceed the variability displayed around the currents ( $O(1\text{m})$ ). The non-linear property of the response is therefore also based on a diverging behaviour between SSTA and SSHA. It may be rooted in an antagonistic effect of the advection on the two variables, or in our method of estimating troughs based solely on post-storm measurements, which does not take into account the intense transport that takes place in such cases and potentially sinks the surrounding mean sea level.

More generally, Figure 6.5 highlights the predominance of baroclinic signatures, with the TC wind stress energy being mainly dissipated within a wake of quasi-inertial waves. The relationship of its amplitude with the translation speed follows a rather monotonic law until the transfer of the response to a stationary circulation (or to another circulation regime), as underlined by the variations between the c/d and e/f panels, and by Figure 6.4 with the Froude number information. In the vein of the studies presented in 1.3.1, the advection of the cyclonic vortex also appears here as an essential criterion for determining the nature of the cyclonic footprint, with which it has a more direct relationship than other coupling parameters, such as intensity.

### 6.1.3 Evolution of anomalies and Impact of the level of information.

In order to summarise the scaling issues, the following figures will present additional arguments on the contribution of our observational data to the study of the wake.

Following the example of the studies of Reul et al. [2014]; Mei et al. [2015]; Lin et al. [2009], Figure 6.6 illustrates the evolution of the averaged anomalies as a function of the cyclonic intensity categories (according to BTK), both for the observations and for the estimates, whether they are derived from SAR measurements (purple curve) or from BTK analyses (blue curve). These averages are of course performed within the validity of the model, with only baroclinic situations (i.e. points in panels c/d Figure 6.5). We have also specified the sampling for each of the categories in the lower panels c and d, as well as the total number of cases for the two variables in the legend. The low intensity of tropical depressions, and the particularly fast translation speeds for systems at the end of their life, make their signatures difficult to interpret from the ocean background, which explains their smaller number in the analysis ( $<10$ ). The other categories are rather well supplied in number of cases, with samples fluctuating between 10 and 30 events.

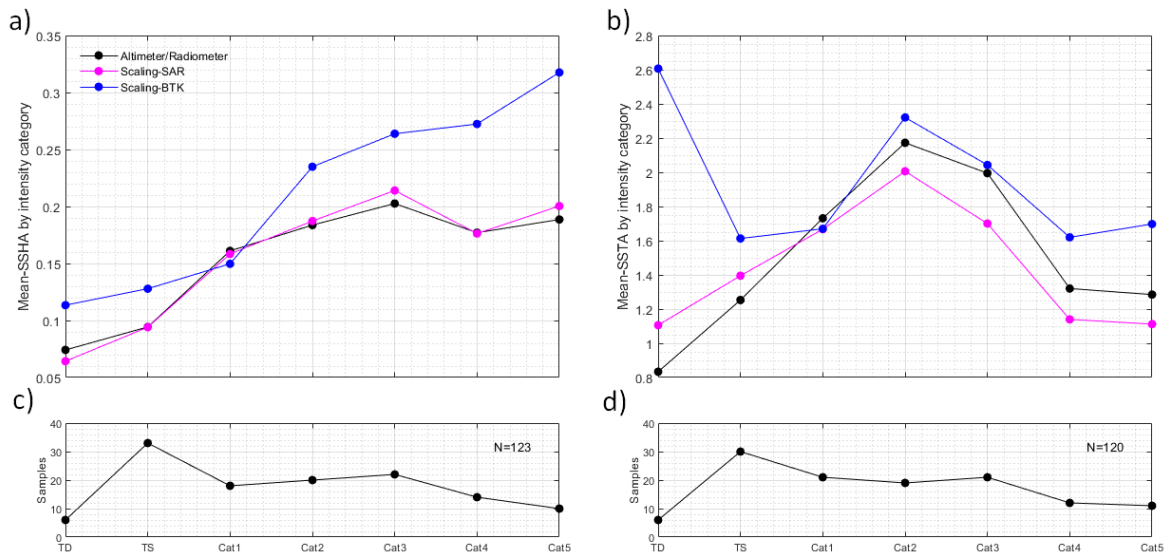


Figure 6.6: *Evolution of the mean (a) sea level and (b) surface temperature anomalies amplitudes averaged by intensity wind category, for SAR-derived (magenta) and BTK-derived scaling retrieval (blue) and for observations (black). The lower panels show the sampling of cases by intensity category for SSHA (c) and SSTA (d).*

As far as the results are concerned, we can only note the great consistency of our observations (black curve) with the trends shown in the literature. The capping generally observed at categories 2 or 3 is indeed present for the SST, and similar to the analyses of [Lloyd and Vecchi \[2011\]](#); [Reul et al. \[2014\]](#); [Mei and Pasquero \[2013\]](#), with a decrease for the major TCs. A plateau is also observed for the SSH, which again corroborates the non-monotonic behaviour of the oceanic response to cyclone intensity for both variables. However, this plateau is asymptotic, since the mean amplitudes remain relatively constant for categories 3 and above. We thus find a divergence in the dynamics of the two variables, this time with regard to the behaviour of the major TCs. For information, this analysis of the SSTA and SSHA as a function of intensity is quite innovative since it has never been carried out simultaneously for the two variables. The figure 6.6 also shows the excellent correspondence between the observed means and those estimated by the scaling deduced from the SAR data, both for SST and SSH, the two curves have indeed a similar shape contrary to those derived from the BTK information. This is especially true for sea level anomalies, for which both curves (SAR and altimetry) first describe an almost monotonous rise up to category 3, then saturate by oscillating slightly around 0.2 metres, unlike the BTK curve which follows a continuous progression up to the most intense cyclones and globally overestimates the observed amplitudes. These results echo to some extent the comparison of the maximum wind radius in Chapter 3 ( see Fig.9 of [Combot et al. \[2020a\]](#)), where overestimated values were observed for all categories. There was also a loss of sensitivity of Rmax information for BTK, with a notable plateau for categories 3 and above, instead of the decrease perceived by the SAR measurements. This constant threshold of Rmax for major TCs is thus translated here by a continuous increase of the SSHA deduced from the BTK scaling, for increasing values of intensity. Looking now at the SSTA trend, the shape of the curve described by the radiometric observations is well reproduced by the averaged estimations extrapolated from SAR measurements, with, however, positive and negative biases for the systems respectively below and above the cyclonic threshold. This result is somewhat expected due to the larger discrepancies observed previously between the model and the measurements for the temperature

anomalies. More surprisingly, the averaged values prescribed by the BTK information seem to perfectly capture the fluctuations around the intermediate categories (Cat 1-2-3), with performances comparable to those of the SAR, and even better for category-3. However, these results must be qualified in view of the more severe median deviations and the very wide scattering displayed around the observations in Figure 6.8 c, which statistically reduces the biases for these intensities (variable in parenthesis) and thus brings them closer to the mean values. Significant discrepancies can be noted for systems located at the two extremes of the cyclonic activity (weak and intense), where there are significant gaps between the actual behaviour of the mean cooling and the one estimated. The trend associated with tropical depressions is particularly poorly anticipated, the hazardous geometry of these systems can be a fairly large vector of error in the BTK (see chapter 3), which is reflected in the estimated response.

In order to complete our speech, we then carry out an interesting experiment of information degradation on all our cases in the baroclinic regime, through Figures 6.7 and 6.8. The objective is to summarize the impact of the type of data we ingest in the parameterization of the oceanic response, showing the progressive evolution of the performance of the scaling laws as we gradually move away from more direct observations (under-track and/or in-situ measurements). The multiple panels of Figure 6.7 represent the evolution of the SSHA estimates, while those of Figure 6.8 focus symmetrically on the SSTA estimates.

There is a clear and common loss of quality in the estimates of the two variables as the level of information in the ingested data increases. It is even more striking when we compare the two diametrically opposed situations, i.e., the one using observations directly (SAR+Argo), or the one using only interpolated products (BTK+ISAS), for which the deterioration of the results is significant and generalized to all the statistical indicators. However, this spoilage occurs in successive stages, more or less visible, depending on the variable considered. The deviations are, for example, clearly identifiable and brought about gradually for SSHA if we look at the values of R and SI in Figure 6.7 (panels a b and c), whereas for SSTA these variations, although real, are less transparent because of the already pre-existing dispersion in the results (see panels a and b Figure 6.8). In this respect, the median deviations are a particularly relevant indicator of the quality of our estimates, and representative of our sample, by weighting the contribution of the most dispersed data. This statistical quantity thus reveals a monotonic progression of the gaps for the two variables, as the more direct data in our analysis are stripped.

The personal impact of the different sources is also seen quite uneven. Indeed, the divergences with the observations are more or less abrupt depending on whether we change the information on the forcing or on the oceanic structure, the most important variations being observed when we document the inner core by BTK data, as shown by the differences between panels a and c of the two Figures 6.7 & 6.8. The differences are just as pronounced between inversions combining SAR+ISAS and BTK+Argo, despite the use of in-situ data, the latter combination cannot compensate for the lack of high-resolution information on the cyclonic vortex. Accurate description of the vertical structure is therefore only beneficial if the properties of the forcing are fully translated, otherwise a poor understanding of the effects of the pre-cyclonic structure on the response ensues, as demonstrated by the comparison between panels c and d.

In the case of a good reading of the inner core, the transition from in-situ data to ISAS climatology (panels a to b) also results in a decrease in the performance of the scaling law, even if it is more moderate and localized at a number of points, proof that a non-negligible

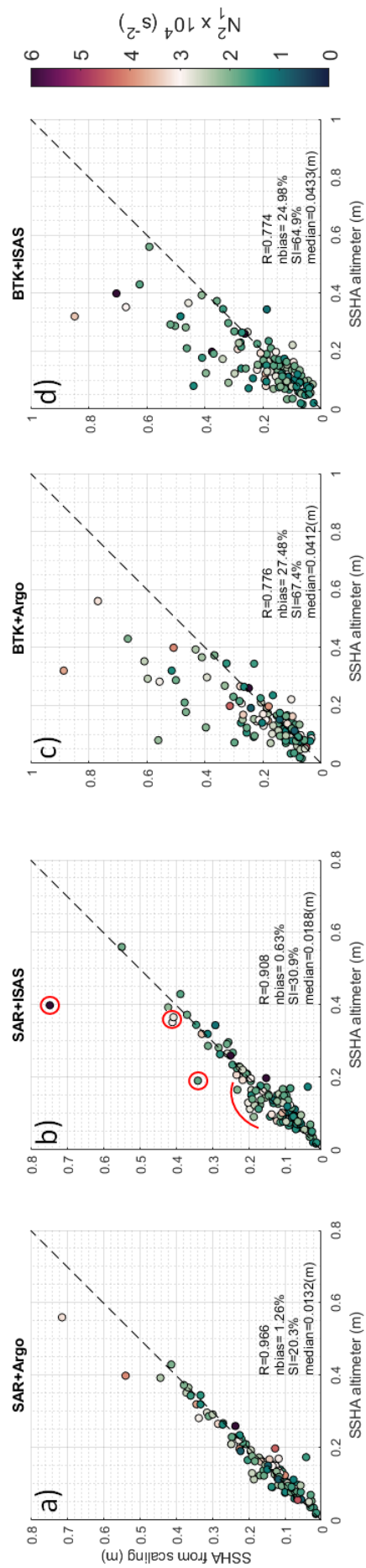


Figure 6.7: Evolution of the SSHA prediction, by progressively downgrading the forcing information and the initial ocean structure indicated by the SAR and Argo measurements (a), with interpolated data: SAR + ISAS (b), BTK+argo (c), BTK+ISAS (d). The colours follow the  $N_1$  values deduced from Argo floats (a, c) or ISAS data (b, d).

fraction of our TCs takes place in oceanic conditions close to the one described by the climatology, or in areas of low temporal variability. However, we observe an increase of more than 10% of the dispersion, including several situations that split away from the observed trend, especially in the vicinity of hollows of 0.1-0.2 m (below the red curve). Other particular cases stand out more roughly for anomalies of higher values (red circles). The impact of the N1 variability thus appears to be proportional to the intensity of the response. For a weak forcing, deviations from climatological conditions will thus have little effect on the amplitude of the measured troughs, as indicated by the N1 variations (see colorbar Fig 6.7) between 6.7a and 6.7b for SSHAs below 10 cm. This result makes physical sense, since very weak or/and fast cyclones inject less momentum into the mixed ocean layer and therefore weakly perturb stronger stratifications. On the other hand, for more intense cyclonic conditions, oscillations around climatological standards can have increasingly dramatic consequences on the scaling law and lead to biased estimates of the response (Fig 6.7 b).

Overall, performance increases with the simultaneous use of direct vertical profile information and a highly resolved inner core source. If the use of climatology allows to observe a behavior already well visible, the use of Argo floats proves to be essential to describe precisely the response and to widen the panel of wake represented. Oceanic areas with high variability, or witnessing unusual or strong seasonal activity, are therefore sources of significant discrepancies that can induce drastic errors in the estimates, even if they are conditioned by the intensity of the forcing. Like the studies of [Lin et al. \[2009\]](#) and [Pan and Sun \[2013\]](#), the differences in performance between inversions with Argo and ISAS show the importance of oceanic control over its own response to TC forcing, even if of course the description of the inner core remains paramount and the main source of discrepancies from observations. If the combined use of Argo floats and SAR observations ensures the best possible anticipation of response anomalies, this reading of the cyclonic wake cannot be complete without estimating the global error of our inversion laws and of the variables measured by altimetry and radiometry.



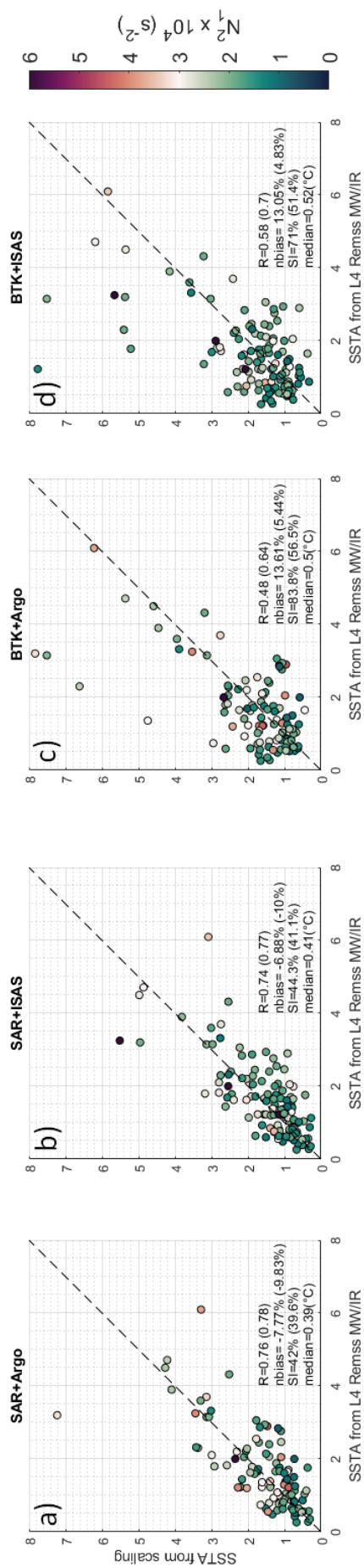


Figure 6.8: Same as 6.1.7 but for SSTA

### 6.1.4 Error estimation.

The benefits of coupled observations, SAR and Argo, are now clear for our semi-empirical basis. Their combination greatly eases the understanding of the cold wake and signature variability, by a better portrayal of the dynamics of the cyclonic vortex and pre-cyclonic ocean conditions. For this optimal version of the scaling laws, we now want to give the error margins relative to both the inversion of the wake geophysical variables, and their measurements, in order to estimate the potential role of these uncertainties in the observed dispersions, especially for the SSTAs. This analysis will also provide orders of magnitude for the quality of our projections. With this in mind, Figure 6.9 revisits the comparison between observations and projected values of anomalies, specifying the associated errors for each point. The horizontal bars refer to the uncertainties related to the extraction of anomalies by altimetry (panel a) or radiometry (panel b), while the vertical bars are attributed to those of the scaling laws. These two types of indicators are derived from distinct methods, responding to the different nature of the x- and y-axis values.

For instance, the errors associated with the observations come from methodological biases, whether by taking into account the oscillations measured around the troughs by all the altimeters in the area, for the SSHA, or by averaging the cooling obtained by the different preconditions, for the SSTA. On the other hand, the uncertainties of the simulated values are to be attributed to the different parameters involved in the scaling laws, the vertical errors, as shown in Figure 6.9, therefore need to take into account each of their influences. To do this, we used the method of propagation of uncertainties, it allows to estimate the individual contributions of each variable at work, by calculating the partial derivatives of the regression law, to first order. Each of these derivatives is assigned an uncertainty ( $Dx_i$ ) of the corresponding variable, known in advance, and which together with the derivative represents the propagation of the error for the variable  $X_i$ .

In the scope of our scaling laws, five parameters are needed to determine the inertial footprint of the cyclone, the quality of their extraction is evaluated as follows. For the couple  $V_{max}/R_{max}$ , their accuracy is known thanks to the comparisons made previously with the SFMR information, for which we use the MAE as a reference value (2.5 m.s<sup>-1</sup>, [Mouche et al. 2019; Knaff et al. 2021] and 3 km [Combot et al. 2020a]). For the translation speed, we choose to take the minimum deviation from the two neighboring values informed by the BTKs for each TC. For the seasonal thermocline brunt-Vaisala frequency, we take into account the error margins of the argo floats in terms of density measurement (2.4dbar), as well as their depth resolution ( $\sim 10$  m). Finally, the contribution of the Coriolis parameter reflects the ambiguities regarding the position of the center of the cyclone, which could be calculated either from the deviations between the position extracted from the SAR images and that interpolated from the BTK, or dictated by the size of the eye of the cyclone. In either case, these deviations are far too small to have a real impact on the quality of the estimates (generally  $< 50$  km), they are withdrawn from the calculation. The sum of these partial derivatives thus assess the quality associated to the scaling laws, the exact details of the different formulas are given in the Appendix for the two markers of the response.

After computation, the global uncertainties are respectively evaluated at 0.05 m and 0.46 °C for the SSHA and SSTA law, they increase individually in a quasi-linear way with the amplitude of the anomalies, due to the increasing weight of the parameters in the scaling laws. The errors associated with the smaller responses (for  $SSHA \leq 0.2$  m and  $SSTA \leq 2^\circ$ ) are thus roughly at the same level as the instrumental biases ( $\pm 2$  cm and  $\pm 0.3^\circ$ ). If we take a closer look at the interval around the SSHA predictions, we see that the margin of error completely encompasses the deviations previously observed

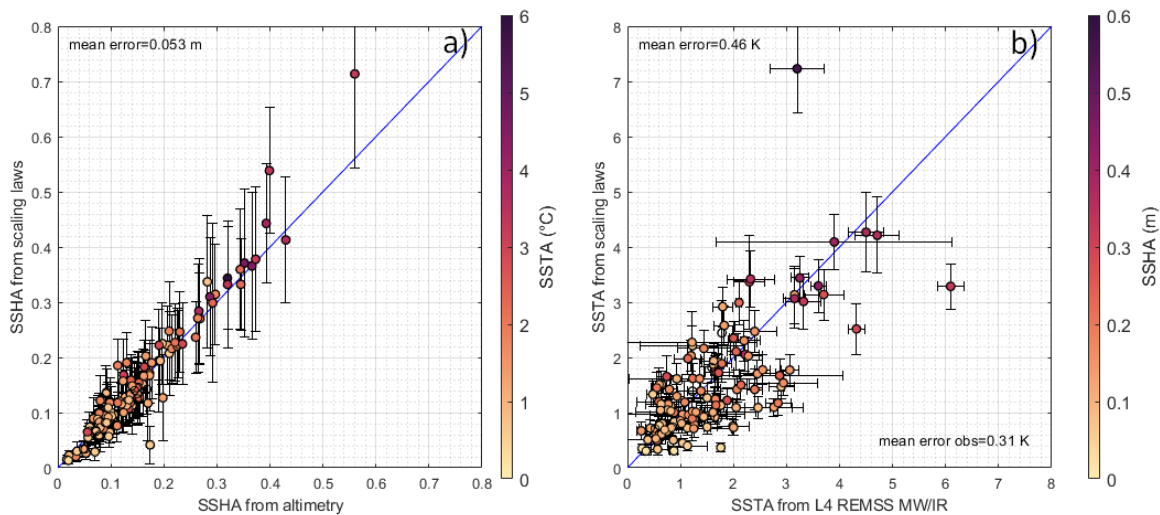


Figure 6.9: *Similar to Figure 6.5 c & d but with the information of the uncertainties associated with the scaling laws (vertical bar) and satellite measurements (horizontal bar) for SSHA (a) and SSTA (b).*

in Figure 6.5, with respect to the RMSE (3 cm) and the median deviations (1.2 cm). If only the points corresponding to the transient situations still remain on the margin of the observed values, (see error bar in Figure 6.9 a), this interval allows to explain most of the dispersion of the projected anomalies (SI=20%). The two source terms of the inner core,  $V_{max}$  and  $R_{max}$ , are responsible for the largest variations, they both contribute on average to 34% of the total uncertainty of the scaling law (5.3 cm), even if in most of our cases (54%) it is indeed those of associated to  $V_{max}$  that dominate, because of its power law (exact details in appendix). They are then followed by the contributions of  $N1$  (21%) and  $V_{fm}$  (11%). To this analysis we must add the inaccuracies associated with the observation of the troughs, which include both the oscillating nature of the wake and the instrumental biases (1-2cm), they prove to be extremely consistent with the uncertainties characterizing the SSHA law, meaning the quality of our predicted values is of the same order of magnitude as our observations. Their combined fluctuations explain the scattering around the trend.

On the other hand, neither the errors related to the measurement of cooling, nor those associated with the prediction of temperature anomalies can fully explain the greater dispersion of the SSTA, in view of the gap between several points and the line, and the larger value of the RMSE. The observations cannot be blamed here for the discrepancies found. While it is true that unlike the other observations in our semi-empirical database, we use an interpolated source of information for the SST, its spatial and temporal qualities are better than those of the equivalent SSHA products, due to the instrumental properties of the radiometers (Chapter 5). This can be seen with the globally low error of about 0.3 °C, which reflects the stability of our methodology and from our various pre-mean conditions, except for a few situations where the time window crosses the presence of concurrent cyclonic events, or the bifurcation of the cyclone, as is the case of Sergio with a large delta of  $\pm 2^\circ\text{C}$ . Moreover, the errors of the instruments incorporated in the REMSS daily solution are generally of comparable accuracy (0.3-0.5 °C) and remain lower than the error of the scaling law. In this regard, the contributions of the stratification parameter  $N1$  (35%) and  $R_{max}$  (33%) largely dominate the total estimated uncertainty (0.46°), followed by the contributions of  $V_{max}$  (21%) and  $V_{fm}$  (11%).

The analysis of the propagation of uncertainties shows us once again how important it is to benefit from under-track and in-situ measurements, particularly in view of the considerable weight of the size and stratification variables in the error margins of the estimates, despite the high precision of our instruments ( $Dx=3\text{km}$  and  $0.001\text{ s}^{-1}$ ). If the  $V_{\text{max}}$  is generally well captured by the BTK, several situations are also tricky (ERC, fast intensification, midget storm), the use of a high resolution measurement of surface wind becomes thereby critical in view of the strong influence of the  $V_{\text{max}}$  power law on the quality of estimates. On the other hand, the errors related to the translation speed are systematically analyzed as the least impacting for the evaluation of the wake anomalies, which is fortuitous, since it is the only parameter related to the forcing for which we cannot absolve ourselves from the BTK data.

As a comparison, we have calculated the global uncertainties of SSHA and SSTA deduced from the BTK+ISAS degraded information. The error associated with each of the parameters is determined from comparisons with L2 and in-situ measurements. The estimation intervals for the vortex TC parameters are thus based on the MAE values calculated previously with respect to the SAR measurements, i.e.  $4.8\text{ m.s}^{-1}$  for  $V_{\text{max}}$  and  $15\text{ km}$  for  $R_{\text{max}}$ , while the  $N1$  values are based on the difference between the climatological values and those described by the Argo data. We have removed from the calculation the errors related to the translation speed, since we take the BTK positions as reference and it is mainly a question here of giving orders of magnitude on the impact of the observations in the accuracy of the predicted values. Irreversibly, the uncertainties explode dramatically up to  $14\text{ cm}$  and  $1.2^\circ\text{C}$ , respectively for the SSHA and SSTA law. The error margins are thus multiplied by a factor of 3, when the use of interpolated data is preferred.

### 6.1.5 Scaling conclusion

In this first section, we analyzed the behavior of the scaling laws on all our cases collocated with the SAR observations. These results have allowed us to question the variability of the signatures, the influence of the different parameters and the impact of the information sources. To summarize the aspects presented here in this section, we can already formulate 5 main conclusions:

- (i) The excellent results displayed by Figures 6.4 and 6.5, demonstrate that the scaling laws allow a partial (SSTA) or quasi-total (SSHA) synthetic understanding of the baroclinic signatures of the oceanic response, for a large panel of situations and for all cyclonic categories (Fig. 6.6), especially when the ingested information is extracted from SAR (Fig 6.6) and Argo (Fig. 6.7 and 6.8) measurements, for which the laws provide fine global estimation uncertainties, evaluated at  $5\text{cm}$  and  $0.46^\circ\text{C}$  respectively for SSHA and SSTA (Fig. 6.9), and which lead to even smaller median differences of  $1.2\text{ cm}$  and  $0.39^\circ\text{C}$  with the observations.
- (ii) This statement is particularly true for the SSHA law, whose inversion allows to deduce the amplitudes with an error close to the instrumentation, and explains 90% of the observed fluctuations. When we add the limiting cases in barotropic regime, the part of the total variability included is lowered to 70%. This number is not meaningless, since Jullien et al. [2012] had estimated that the same proportion of tropical cyclones own a predominantly inertial signature, with the remaining 30% having a stationary barotropic trough composed of an Ekman-type pattern. This

conclusion is also shared by the more recent study of [Zhang et al. \[2020\]](#), for which our analysis offers an alternative method to confirm this postulate.

- (iii) Both the sea level and temperature anomalies follow a non-monotonic behavior with respect to the intensity of the TCs (Fig 6.6), and run respectively along a spectrum of values from 0 to 0.7/0.8 meters and from 0 to about 10 degrees. If the barotropic circulation generates anomalies covering the entire range of values observed, these intervals of measurements can be dissociated into three different dynamics. The purely baroclinic signatures are focused and dominate the first region ( $[0^\circ - 4^\circ]$ ,  $[0 - 0.4m]$ ), which is the richest in cases. The upper region ( $>6^\circ$ ,  $>0.5/0.6m$ ) is occupied almost exclusively by barotropic signatures, for which mixing alone cannot explain the intensity of the observed cooling without the competition, this time dominant, of an Ekman upwelling. A hinge zone ( $[4^\circ - 6^\circ]$ ,  $[0.4 - 0.5m]$ ) lies between these two trends, consisting of responses belonging to both categories and from transient cases with both strong baroclinic and barotropic components. More details will be given on the estimation of the contributions of the two circulations on the total response in section 6.3.
- (iv) The spoilage of signatures (Fig 6.7 and 6.8) occurs abruptly when switching from high-resolution SAR observations to 6-hourly Best-track analyses, and in a more localized manner but with drastic differences, when switching from Argo data to ISAS climatology. This change in observation paradigm also leads to a threefold increase in prediction uncertainties.
- (v) There is a double asymmetry between the two geophysical variables, with the first associated with the non-linearity of the oceanic response for subcritical velocities, which results in a capping of the SSHA around 0.6-0.7m and a rise of the cooling to more than double of the baroclinic signatures. This divergence in dynamics is also reflected in the more limited statistical performance of the SSTA inversion (Fig. 6.5), with notably a twofold scattering, making the interpretation of stratification information ambiguous (Fig. 6.8). These performances are also countered by an incomplete reading of the cooling variability, indicated by the invariance of the correlation and the lighter rise of the RMSE for the addition of limiting cases.

This gap in the cooling monitoring, may come from an asymmetry in the relationship of SSHA and SSTA to the dynamics of the MLD. In the scope of the scaling laws, the uniform layer depth is seen as a small and negligible perturbation to the benefit of the seasonal thermocline stratification. If the results tend to confirm this hypothesis for the inertial wake (SSHA), [Kudryavtsev et al. \[2019b\]](#) pointed out, on the other hand, the need to incorporate this information to widen the range of SSTA predicted by the model. Yet, this section will have allowed us to estimate the part of the variability included by our SSTA scaling ( $R^2=50\%$ ), without taking into account the MLD dynamics. The following section will remedy this, by further exploring the asymmetry between the two response variables with respect to the Argo data.

## 6.2 Investigation of signature asymmetry and vertical wake analysis by Argo.

### 6.2.1 Scattering of the SST and SSH response across basins.

So far, we have dealt with the two cyclone footprint markers individually. We now want to explore their relationship further. Figure 6.9 suggests a roughly synchronous evolu-



tion of SSHA and SSTA, given the simultaneous growth of the colour gradient with the anomaly values on the ordinate. In the previous section, however, the examination of their magnitudes as a function of a number of parameters revealed asymmetries in their respective dynamics, which leads us to question the degree of real affinity between the two signatures and the intertwining of the mechanisms of their responses. This confrontation is all the more interesting as it has never been carried out in the analysis of the TC cold wakes.

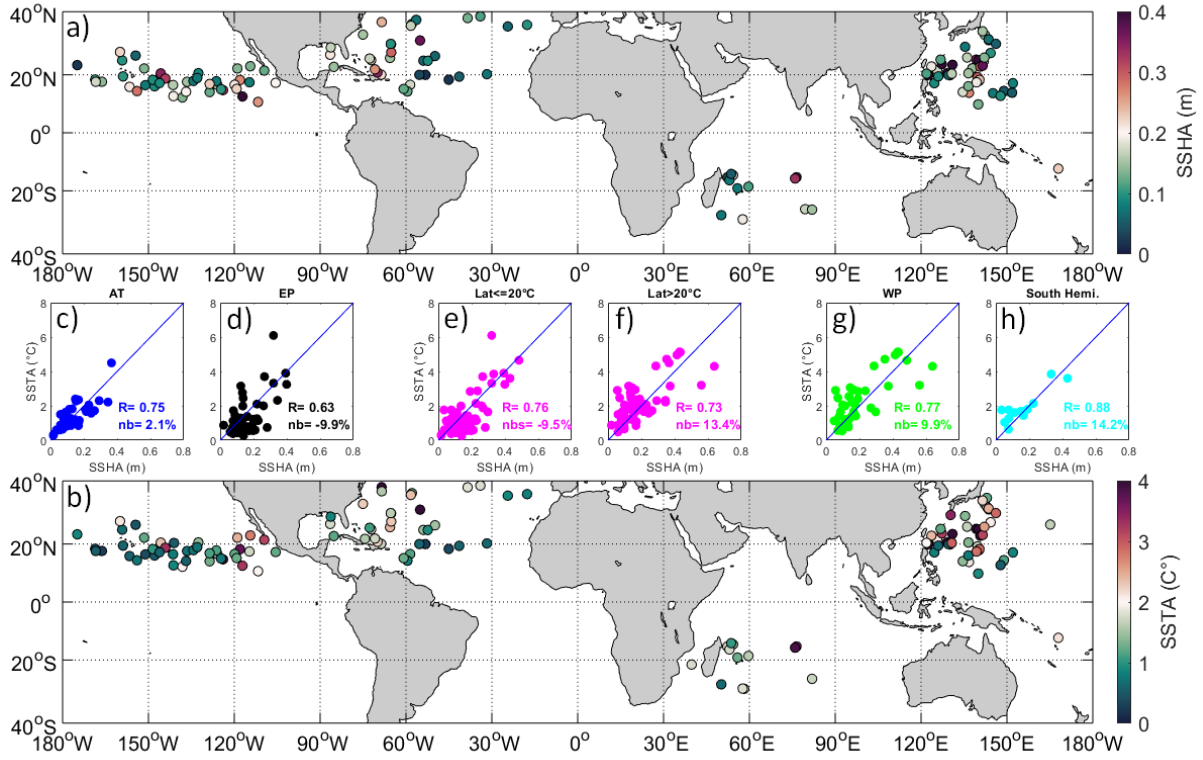


Figure 6.10: *Summary map of observed amplitudes of surface troughs (a) and surface cooling (b), with their comparison discretised by area, for the Atlantic (c), Eastern Pacific (d), Western Pacific (g) and Southern Hemisphere (h) basins, and by geographical area as a function of latitude, for the tropical (e) and subtropical (f) regions.*

On this occasion, Figure 6.10 provides a comprehensive overview of the global scattering of the wake anomalies, with their comparison discretised zonally and by basin. This choice of decomposition will allow us to highlight the areas of divergence between the two markers, and is motivated by the outcomes of previous published work on the influence of environmental parameters. The local planetary vorticity, expressed by the Coriolis parameter in the SST law (which depends on latitude), is as essential for the development of the cyclonic circulation (section 1.2) as it is for the wake (section 1.3), with the resulting inertial period information reflecting, among other things, the efficiency of turbulent mixing [Mei and Pasquero 2013]. Shay [2009], Lloyd and Vecchi [2011] and Wu and Chen [2012] have demonstrated in this sense the influence of latitude on the cooling amplitude, as well as the parameterisation of Knaff et al. [2013]. While the 20°-30° N/S band generally has the strongest SSTA [Wu and Chen 2012], which is also consistent with our results (see Tables, Annex F.3), this observation is often strongly interdependent on oceanic and forcing conditions, which are more favourable in this area. Many studies, mainly based on different sectors of the Atlantic [Black et al. 2007; Shay 2009] and West-

ern Pacific [Lin et al. 2009; D’Asaro et al. 2014; Mei et al. 2015] have focused more on the influence of oceanic structures present in the interior of the basins, some properties of which may coincide with a zonal distribution, like the thinning of the MLD (see Fig 6.3). As the cyclonic forcing parameters are common to both descriptions (SSTA and SSHA), our analysis needs to be supported by such local oceanic information. Thereby, the addition of the basin description attempts to syncretize these different aspects to better understand the dissonance between the two signature models. However, the paralleling of the two geophysical variables requires a framework for comparison. In the previous Figure 6.9, an average trend of  $1^{\circ}\text{C}$  cooling per 10 centimetres of trough could be detected, this order of magnitude is used to ease the reading and scaling of the SSHA and SSTA in panels 6.10 a) and b). The regression lines in panels c) to h) also follow this same rule and impose a factor of 10 between the two variables. The question we can now ask is how this axiom is verified at the global scale and across basins?

At first sight, this proportionality relationship seems to be consistent when looking at panels a and b, with globally congruent activities between surface temperature and sea level anomalies. This is particularly the case for the Southern Hemisphere (6.10h) and Atlantic (6.10c) basins, which offer remarkable correlations (0.75 & 0.88) and slight biases between the two markers of the wake dynamics. A closer look at the variations in ocean signatures nevertheless reveals areas of strong divergence, mainly driven by the North Pacific basins (6.10d & g). This is visible in the very strong sea level anomalies around 20-40 cm, and the associated weak SSTA around  $0-2^{\circ}\text{C}$  in the eastern Pacific (nbias  $\sim -10\%$ , 6.10d). In contrast, exacerbated temperature anomalies ( $2-4^{\circ}\text{C}$ ) proliferate in the North-West Pacific (nbias  $\sim 10\%$ , 6.10g), in the vicinity of the central region, but also near the Philippines and the China Sea (6.10b). These disparities are zonally-distributed and reflect the reversed patterns of the two basins, with the more tenuous anomalies from EP concentrated at lower latitudes ( $< 20^{\circ}\text{N}$ ) and the sharper WP anomalies at higher latitudes (6.10b), which also overlaps with the effects of the Coriolis parameter on cooling (Shay [2009], Fig. 6.10e & f). Inevitably, these biases are echoed in regional comparisons, where the points corresponding to the two dissident basins can be detected, with reduced SSTA values for the tropics (nbias  $< 0$  6.10e) and amplified for the subtropics (nbias  $> 0$  6.10f). While some of the dispersion is attributable to the effect of the Coriolis parameter on the inertial properties of the mixing, it seems to be mainly driven by the dynamics of the basins, and in this case more specifically by that of the North Pacific.

Unsurprisingly, this is also the region of the world with the greatest spatial variability in ocean features (see Figure 6.3). Indeed, the eastern Pacific accumulates the largest standard deviations on all stratification parameters (Appendix F.1 & F.2), generally followed by the western Pacific. While the former monopolises most of the deep ML situations ( $\sim 40\%$  of those  $> 40$  m), its western neighbour hogs the thinnest ML ( $\sim 50\%$  of the 0-20 m). As the cyclonic activity in the  $110^{\circ}$  to  $180^{\circ}$  W bands is mainly confined to low latitudes, the surface oceanic layer deepens due to the strong advection from the equator; conversely, typhoons further north, cross territories marked by a gradual zonal thinning, particularly severe in the open ocean of the WP [Mei et al. 2015]. These two basins can also experience opposite situations, such as hurricanes travelling through the upwelling areas near the US west coast, where a more significant cooling takes place (fig 6.10 b). Of course, the aim here is not to make a precise inventory of the basins, but to discern the dominant trends that explain the existing divergences. The main thing to note from these findings is that EP and WP together account for nearly 75% of the situations bordering the observed mixing layer depth interval (thin and broad), suggesting its crucial role in the asymmetry of the signatures. WP also benefits from the greatest temporal variability

in the ocean, with local stratification of the thermocline (i.e. as seen by Argo) well outside climatological values (Appendix F.1). For information, the differences between the SAR+Argo and SAR+ISAS inversions in Figure 6.7 are mainly from this basin, where mesoscale variations (eddy) are 50% to 100% larger than in the open Western Atlantic [D’Asaro et al. 2014], so it is possible that some of the ocean patterns in the area are eddy-induced. In parallel to these discrepancies, the excellent match of the SSTA and SSHA in the North Atlantic and Southern Hemisphere goes along with more moderate spatio-temporal fluctuations (Appendix F.1 & F.2) and the presence of more standard structures. For instance, hurricanes circulate mainly (60%) over a medium mixed layer (20m-40m). These conclusions are of course in the context of our database, many ocean features (WCR, CCR) evolve in the Gulf of Mexico and the Caribbean sea; our study being constrained by the capabilities of satellite instruments, most cases in this area are obscured in favour of the open ocean.

The Atlantic is also accustomed to a limiting situation that we have so far ignored: barrier layer scenarios. Of the ten or so cases recorded, half come from the eastern Pacific, with the other Northern Hemisphere basins sharing the rest equally. Although the Atlantic is generally a breeding ground for BL [Reul et al. 2014; Balaguru et al. 2020], due to the persistence and extent of the Amazon-Onorico plume, very few of our SAR acquisitions here have ventured into this area. As for the Pacific, as mentioned in section 1.1.1, they mainly arise from monsoon episodes for the western part, and from a mixture between equatorial advective flows and ITCZ precipitation for the eastern part. In addition to their prevalence in tropical basins, one of the most striking features of BL is their strong potential to inhibit mixing. Although we do not know their actual effects on SSHA, this property suggests a tangible role for them in the observed divergences, especially in the EP where they are more numerous. To satisfy our curiosity, two cyclonic passages crossing such saline structures will be analysed in section 6.2.2 (Lane and Hector again!). At last, the ubiquity of salient oceanic features in the EP basin, and its high variability, explain the lower correlation between SSHA and SSTA and the larger disparities obtained (Fig 6.10 d).

The inter-comparison of the two cold wake anomalies has therefore allowed us to explore their dissymmetries as a function of two environmental factors. On the one hand, the latitude, which intervenes unilaterally in the scaling laws (only the SSTA), and on the other hand the MLD, whose impact on the cooling has already been documented. While the dynamics of the Coriolis parameter is already taken into account in the parameterisation of the thermal response, that of the MLD is absent. When we substitute the ocean response measurements (SST, SSH) by the expected values from the scaling laws in Figure 6.10, asymmetries are not faithfully reproduced in the North Pacific (not shown here). Yet, the relationships described by the other two basins remain unchanged (AT+SH). In a way, AT and EP represent diametrically opposed situations and isolate the dynamics of the Coriolis parameter and the MLD. While the first one provides a broad basin with homogeneous surface layers (for our sampling), the second provides a narrow basin but with a wide range of mixed layer depths. Given the differences in bias and the different impacts mentioned, the influence of the ML appears to be dominant overall. In order to provide more direct arguments for this statement, we will now exploit further the Argo data, by analysing the relationship between surface deepening and dynamic wake markers.

### 6.2.2 Mixed layer deepening and surface anomalies.

We look for structures in our collection of ocean responses that have already been examined by an Argo profile and that can be diagnosed by a float after the cyclone's arrival, in a short time window between  $T+1$  and  $T+5$  days ( $T$ : day of arrival) due to the rapid temperature recovery period (e-folding 7-15 days). Of the 111 collocations previously made between SAR and Argo, about 30 allow a complete temporal analysis of the wake, both before and after the cyclone passage. Although a few more cases meet this temporal condition, our sample was reduced to meet another requirement, this time spatial; we kept only those situations for which the post-cyclone in-situ measurements are in the wake resonance zone (right/left of the track in the northern/southern hemisphere). As the asymmetries are relatively large as a function of translation speed, the cyclone footprint can be very weak and thin on the left side of the track (northern hemisphere), which can lead to large underestimates of the actual forcing. All selected floats are therefore located close to the area of maximum cooling. This final sample of cases benefits from both complete satellite monitoring at the surface and at depth via Argo profiles. Thanks to this temporal follow-up, we can subtract the two vertical profiles and access the cooling and deepening of the mixing layer.

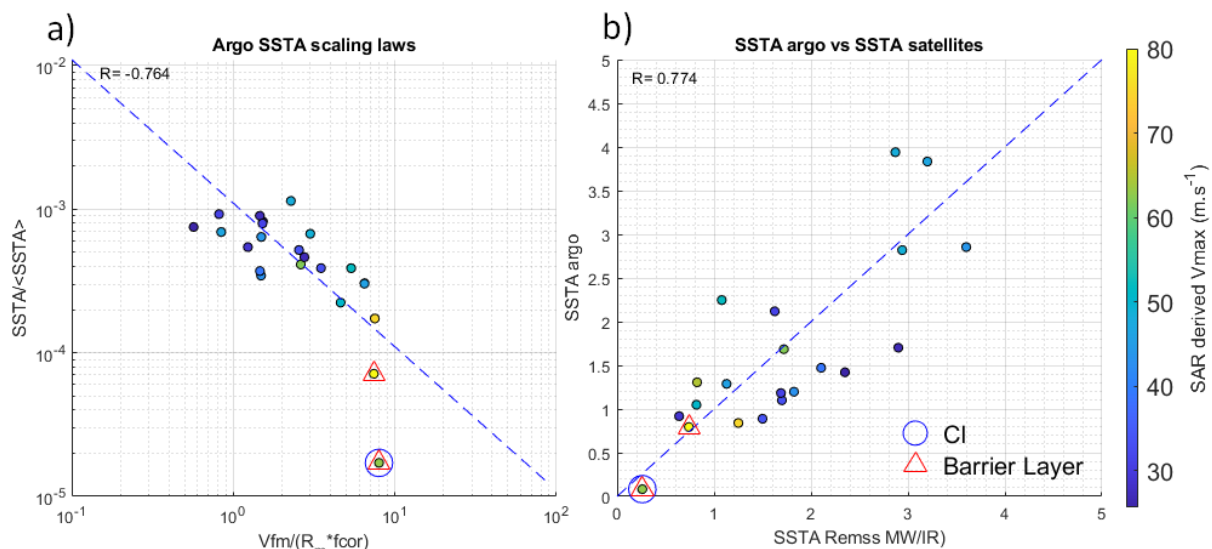


Figure 6.11: Verification of the scaling law with temperature anomalies measured by a pair of argo floats before and after the passage of the cyclone (a), and comparison with SSTA deduced from satellite observations (b). Barrier layer and cooling inhibition situations are indicated by red triangles and blue circles respectively.

Before comparing the behaviour of the anomalies as a function of layer depth variations, we wish to ensure the consistency of the SSTA measurements between in-situ and satellite observations. Figure 6.11 shows the robustness of the model (panel a) and the observations (panel b) through two analyses. The scaling law developed from the radiometers is very convincing when using the Argo derived information and is very close to the behaviour observed with the satellite measurements. As for the amplitudes of the anomalies, they are very consistent between the two sources of observations. The slight scattering observed probably implies the spatio-temporal shift between the two measurements, as the Argo floats are not necessarily located exactly in the extremum zone, just as the anomalies given by REMSS may be reduced due to the temporal smoothing associated with the product interpolation method. Two limiting barrier layer situations were

also captured by pairs of Argo floats, corresponding to the passage of two EP hurricanes, Hector (circle + triangle) passing through a very thick mixed layer (70 m) accompanied by a very thin BL (10m), and Lane (red triangle) encountering the opposite situation with a thin ML (10m) accompanied by a broad BL (30-40 m). The points corresponding to these two cases progressively deviate from the trend dictated by the scaling law, but are extremely consistent between radiometric and in-situ observations, with very little associated cooling. This preliminary comparison thus attests in some way to the quality of our measurements and their good correlations, which invites us to continue our analyses.

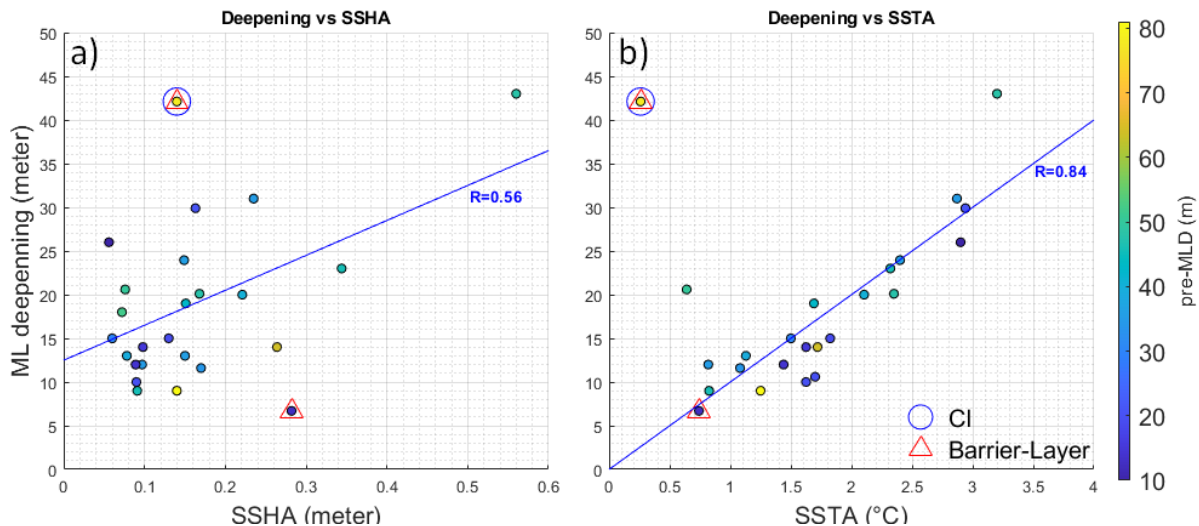


Figure 6.12: *Analysis of the behaviour of the surface anomalies, as a function of the depth of the mixed layer measured by a pair of argo floats before and after the passage of the cyclone, for the SSHA (a) and the SSTA (b). Special cases such as cooling inhibition (blue circle) and barrier layer (red triangle) situations are also specified. The colours are indicated according to the initial mixing layer.*

In that regard, we assess the deepening from the differences between the initial and post-cyclonic MLD of the Argo pairs. These results are then compared to the magnitudes of the surface anomalies, those of sea level (6.12 a) and those of temperature (6.12 b). Figure 6.12 summarises the existing asymmetries perfectly, and is direct evidence of the different dynamics of the SSHA and SSTA with respect to the surface ocean layer, with on the one hand surface trough amplitudes that remain unaffected to its deepening, and on the other hand temperature anomalies that are totally sensitive to it. These results validate the hypothesis of the Kudryavtsev model on the transparency of the mixed layer to the development of the near-inertial wake, as shown by the scattering of the point cloud and the low correlation obtained (0.56). Still, it is reinforced by the isolated situation of the transient case of Trami, without which the scatterplot falls to  $R = 0.34$ , explaining only 10% of the observed variability. The ML thus appears to be a small perturbation, which can be neglected for the SSHA estimates.

This reasoning is not valid for temperature, as the experiments conducted by Mei et al. [2015] between the open ocean of the WP and the China sea have clearly shown. In particular, Price [1981] mentioned that SST remains mainly a mixing layer problem. Under the assumptions of the baroclinic model, its cooling is mainly induced by the entrainment flows at its base, which are the result of the turbulent motions injected by the cyclone. The resulting erosion of the thermocline leads to the deepening of the ML.



The greater the deepening, the greater the erosion and entrainment of cold water from the thermocline towards the surface. It is therefore normal to observe in Figure 6.12(b) a congruent evolution between the intensity of mixing and the widening of the surface layer. This close link between temperature and depth variations (observed in Figure 6.12b) is modulated both by the parameters of the forcing and the thermocline, as transcribed in the models of Pollard et al. [1973] and Price [1983], but also by the description of the mixing layer which reflects the accessibility of cold water at its base [Neetu et al. 2012; Vincent et al. 2012b]. Furthermore, in Figure 6.12b, we observe a stronger relationship between the two observables for changes in ML than that described by the scaling law. Moreover, we observe in Figure 6.12b a stronger relationship ( $R=0.84$ ) between these two variables of the mixing layer than the one described by the scaling law.

The ambivalent nature of the wake with respect to surface dynamics is also illustrated by the salient divergences caused by the two barrier layer cases aforementioned. While the SSTAs are strongly attenuated by the presence of this haline stratification ( $< 1^\circ\text{C}$ ), the SSHAs remain impermeable ( $> 10\text{cm}$ ) and in agreement with the values predicted by the scaling laws. Lane dug a steep trench of nearly 30 cm, well below the induced thermal signature ( $\sim 0.7^\circ\text{C}$ ), while Hector has a more moderate wake of 14 cm and an almost negligible cooling of  $\sim 0.2^\circ\text{C}$ . The widening of its MLD should be taken with caution, as it is actually a thinning of 42 m due to the very deep initial layer in its path, which in turn inhibits mixing (CI: cooling inhibition) with the help of the BL at its base. Hector and Lane represent an important failure, a flaw in the suggested proportional relationship between the two wake variables ( $1^\circ\text{C}=10\text{cm}$ ). The features altering the properties of the ML therefore also contribute to the asymmetry of the observed responses. In general, this cross-comparison shows us that the inertial response is highly insensitive to the different dynamics impacting the surface layer, contrary to the cooling which is totally dependent on it. The use of the complete solution of the oceanic response allows us to prove this qualitatively on these two examples.

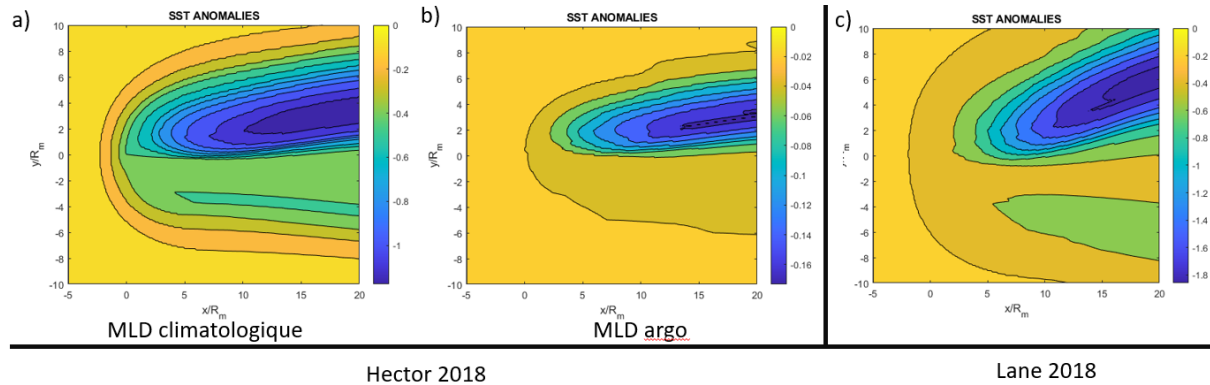


Figure 6.13: *Simulation of the surface cooling of Hector (a,b) and Lane (c) with the complete model of [Kudryavtsev et al. 2019a], using the argo profile (b) or the climatology information from ISAS15 (a) for Hector cases.*

Contrary to the scaling laws, which serve a need for simplification, the mixing layer is involved in the complete model [Kudryavtsev et al. 2019a] to characterise the heat budget. If the surface layer is initially neglected to determine the vertical displacements associated with the propagation of near-inertial waves, the depth parameter is necessary to close the model and evaluate the entrainment flux at its base. Ultimately, its description has an impact on the prediction of temperature anomalies, as is the case for the

Hector simulation in Figure 6.13. Injecting the climatological MLD (30 m) or the local Argo value (70 m) brings a significant difference of more than  $1^{\circ}\text{C}$  in amplitude, as well as changes in the geometry of the wake. If the climatological value is chosen, then the predicted cooling ( $1.2^{\circ}\text{C}$ ) seems very consistent with the amplitude of the inertial response (14 cm), as it follows the average trend of  $1^{\circ}\text{C}$  per 10 cm but remains far from reality. Conversely, the local value (i.e real value) leads to an estimate very close to that observed with satellite observations and the Argo float pair. The inhibition of mixing as Hector passes is therefore well reproduced, as is its near extinction to the left of the trajectory. Throughout the experiment, the SSHA estimates remained stable regardless of changes in the initial thickness. The assumptions and mechanisms of the wake at the origin of the model, therefore verify this asymmetric behaviour between the two variables with respect to the dynamics of the ML.

Lane's case is more conflicting <sup>3</sup> due to the absence of a description of the salinity gradient, temperature being the only driver of stratification. The cooling is therefore logically overestimated. However, the occlusive effects of BL can be partially simulated by considering the depth of the isothermal layer instead. These two examples show in a synthetic way the interest of integrating the pre-cyclonic surface structure. A better anticipation of cooling thus requires taking into account the initial ML pattern, such as the cooling inhibition parameter of Vincent et al. [2012b], or the reduction parameter presented in Kudryavtsev et al. [2019a]. Without this information, the predictions are misleading, like with the temperature scaling law which indicates the same values as the model with the climatological MLD, for the Hector case.

### 6.2.3 quantification and incorporation of the effects of the MLD.

Now that we have demonstrated that the asymmetries between SSHA and SSTA are built around the properties of the ML, we want to quantify these effects on the amplitudes of the observed signatures and incorporate its dynamics into the scaling law. In this respect, Figure 6.14a shows the comparative evolution of the two anomalies according to depth category. The thin mixing layers from 0 to 20 metres (magenta), the intermediate layers from 20 to 40 metres (black) which correspond to the climatological averages generally observed for the tropical open ocean areas, and finally the deep mixing layers (cyan) which extend beyond 40 metres. For each of these groups, the temperature anomalies follow distinct trends, which diverge from the initial hypothesis when departing from standard conditions. Only the events occurring on middle layers (black circle) actually meet the initial standard of  $1^{\circ}\text{C}$  cooling per 10 cm of trough (black line), which explains the symmetry of the signatures in the Atlantic, where this type of structure predominates.

For the other cases at the edge of the MLD spectrum, they cluster at the periphery of the central curve and crystallise around two opposing patterns. The thin surfaces gather around a line corresponding to a 70% amplification of the SSTA, while the thick surfaces coalesce around a 30% reduction of the mixing. Incidentally, this duality is also reflected in the previously observed biases in the EP/WP. It should be noted that the group of deep ML enfolds a wider range of measurements (40-90m), and for the thickest the reduction can be as much as 70%. Remarkably, these two groups of points never overlap or intermingle, they both remain confined to their portion of the graph and rarely outreach the climatological mean curve. This abrupt separation is indicative of the tight control that the surface structure exerts on the temperature trend. At the same time, it allows us to assume the physical limits that this strict behaviour implies.

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<sup>3</sup>hector also passes through a barrier layer, but this is too thin and negligible compared to the depth of the ML (70 m).

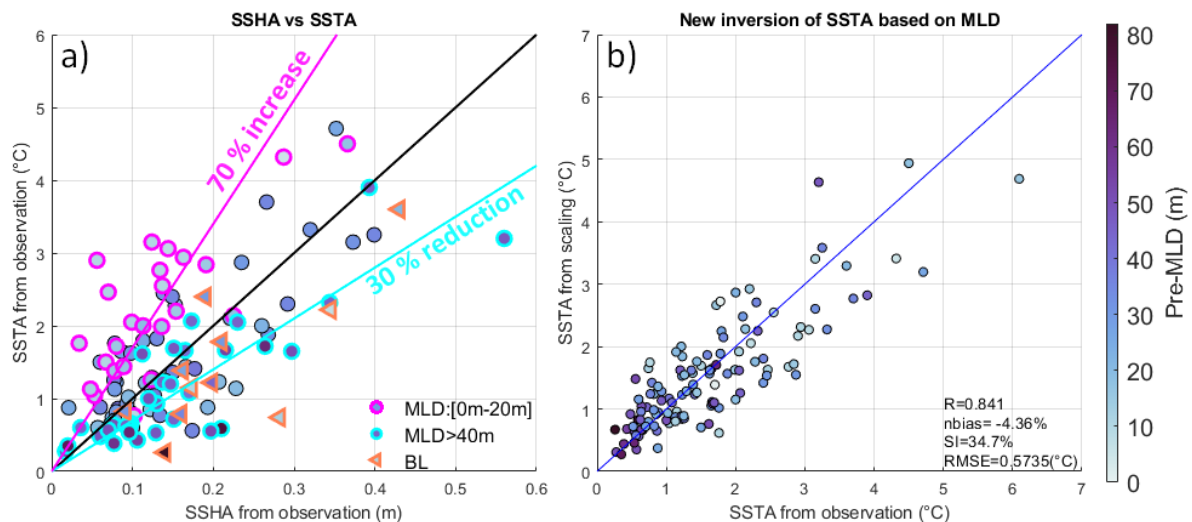


Figure 6.14: (a) Relationship between SSHA and SSTA anomalies for thin (magenta circle), medium (thin black circle), and deep (cyan circle) mixed layer situations, with corresponding regression lines. Barrier layer cases are indicated by an orange triangle. (b) New temperature anomaly inversion law, based on scaling discretised by mixed layer category. The pre-storm ML is used for colorimetry.

From the previous section, we know that purely -or strongly - baroclinic responses provide cooling centred around  $0^{\circ}\text{--}5^{\circ}\text{C}$  (or  $5^{\circ}\text{--}6^{\circ}\text{C}$  for transient cases), and SSHAs up to 40-50 cm trough (or 50-60 cm for transient cases), as exhibited by systems evolving on intermediate ML. This means that TCs can hardly trigger a cooling of more than  $3^{\circ}\text{C}$  when passing a deep ML, according to the regression line (cyan curve). For episodes occurring on structures close to or beyond 70-80 m depth, this limit even falls to the  $1^{\circ}\text{C}$  threshold, due to much less efficient mixing at the base of the ML (70% reduction). Although temperature anomalies of thin surface outperform those of the other categories for the same SSHA variation, few of them goes beyond  $4\text{--}5^{\circ}\text{C}$ . Most of our strong thermal signatures are indeed associated with medium-sized layers. The apparent scarcity of strong responses may stem from another physical limitation concerning enthalpy fluxes. While Lin et al. [2008] and Vincent et al. [2012a] have shown the ocean control on the cold wake, Lin et al. [2009], Lloyd and Vecchi [2011] and Walker et al. [2014] have highlighted its influence on the cyclone intensity. Indeed, if the mixing is more efficient when the interface is thin, the energy reservoir coming from the warm waters is also more limited due to the proximity of the thermocline. The rapid intrusion of cold water towards the surface implies a significant reduction in air/sea fluxes, and even a negative feedback from the ocean, more or less severe depending on the translation speed of the cyclone [Lin et al. 2009]. Cione and Uhlhorn [2003] calculated that a difference of only  $1^{\circ}\text{C}$  induced changes of 40% in the enthalpy fluxes, which can lead very quickly to the suppression of the cyclone intensification for small surface layers.

The direct consequence of these coercive effects is the depletion of the number of major cyclones in these areas, or the acceleration of their motion speeds. Mechanically, this also leads to cap the response amplitudes, which may explain the moderate cooling. A quick analysis of the Vmax, averaged for different depth categories, confirms this assertion for our database. Hence, Table F.4 of Appendix F shows a monotonic growth of the average intensity with thickness, with notably a difference of more than  $10\text{ m.s}^{-1}$  between the

values at the lower and upper limits. This is obviously a coarse analysis, we try to set the approximate cooling limits dictated by the mixing layer thickness, under the baroclinic assumption. These limitations are of course obsolete with the help of a generalized upwelling event, whose depth response is able to disrupt the thickest layers [Lin et al. 2017], or even remove subsurface warm anomalies [Jullien et al. 2012] due to a more extensive vertical cooling [Price 1981]. Among the most disruptive points we also find the stigmas of the barrier layers (orange triangle). All the situations identified are below the central curve or even the line of 30% mixing reduction. Despite the overall attenuation of the thermal signatures, their heterogeneous behaviour prevents them from being gathered into a precise trend. The inhibition rates relies on both the thickness and the depth of the BL, and this information is not taken into account by the graph. As demonstrated with the Hector and Lane examples in the previous subsection, the presence of such salinity stratification does not prevent the development of very strong inertial wakes. The Figure 6.14 a) generalises this insensitivity to all our cases.

The effects of oceanic control on mixing are expected. The cross-comparison of wake amplitudes in the light of the surface layer dynamics is however new. It has allowed us to address in an original way the amplification or reduction of the cooling, to quantify them and to highlight the asymmetric nature of the SSHA/SSTA relationship conveyed by the mixed layer. The reverse reasoning can be equally relevant. A salient response of one of the two variables is also indicative of a singular surface stratification, which deserves the use of in-situ information. While Price et al. [1994] stated that the thermal answer was completely intertwined with the current field, our results qualify this statement and show that the intertwining of the mechanisms may be partial and conditioned by the ML. Its depth is a measure of the degree of resonance between the two responses, leading to a gain or loss of efficiency of turbulent mixing. The dissonance also stems from the heterogeneous nature of the two variables. The SST, as understood by the IR and MW satellite observations, measures skin (10  $\mu\text{m}$ ) and sub-skin (1 mm) temperatures respectively, which, when adjusted and stripped of the influence of the diurnal cycle, represent the temperature of the uniform surface layer of the ocean, and the mechanisms that contribute to its evolution. This is, of course, a very shallow measure of the ocean, rarely exceeding 100 metres. While the SSH is a description of the dynamic ocean height, set in motion by atmospheric and mesoscale forcing, it incorporates all the mechanisms leading to steric and mass transport variations in the upper ocean, i.e. the internal displacement of the isopycnets. The cyclonic inertial imprint generated as a three-dimensional wake consists of a superposition of n-mode baroclinic currents, with vertical wavelengths ranging from about 100 m to 1000 m, which innately leads to intense vertical motions with alternating upwelling and downwelling cells. Vertical velocities are paroxysmal within the mixing layer and the seasonal thermocline (100-400 m) where maximum energy and shear are accumulated. The result of these internal displacements is the so-called surface trough to which the geostrophic circulation adjusts. Cyclone-induced sea level anomalies can be seen as the direct manifestation of these mainly subsurface phenomena, while cooling is the result. The MLD thus mediates between the deeper subsurface dynamics (100-1000m) of the near-inertial wake and the surface dynamics of the cooling (0-100m).

Unlike sea level anomalies which are written as a function of forcing and stratification, thermal anomalies must be written as follows:  $\text{SSTA} = \Phi(\text{forcing}, N1, \text{Lat}, \text{MLD})$ . In order to incorporate this last missing component, we have redefined the SSTA scaling law, decomposing it according to the depth bands already presented. The inversion of this optimised law provides new estimates of the thermal response, which we have compared with the observations in Figure 6.14b.

This addition leads to a significant improvement in all statistical indicators of variance and regression metrics, with the exception of the bias which remains constant. The root-mean square error is generally affected by the outliers in contrast to the median deviation. Here, their convergence and the decrease in the RMSE reflect a better reading of the oceanic responses from a pre-cyclonic borderline structure, which is also shown by the decrease in the scattering index. The most conclusive improvement remains the correlation, which is close to the observed relation between the deepening and the thermal anomalies, meaning that our scaling parameters perfectly synthesise the driving mechanisms of ML erosion and cooling. In this respect, the fraction of variability explained by the wind-induced turbulent mixing is now similar to the orders of magnitude decreed by previous studies, around 70%-80%. The remaining fraction is due to air/sea flux for the weakest systems, and to advective flows including upwelling for the slowest and/or largest cyclones, both of which are absent from the wake modeling. Let us recall that the scaling laws remain above all a linearisation of the three-dimensional cold wake problem, reducing it to its two surface variables and to a restricted number of tropical cyclone and ocean vitals. Our various efforts have resulted in two scaling laws that now enfold most of the baroclinic response. Although the depth of the ocean surface layer is not really integrated in the parameterisation, its main trends have allowed us to derive a set of regressions that allow a quick diagnosis of the cooling. This in-depth examination of the wake with the dynamics of the vertical structure will have allowed us to confront for the first time the more or less resonant evolution of the two anomalies, on which we can draw some major conclusions to sum up this section.

#### 6.2.4 Conclusion about SSHA and SSTA dynamics and asymmetries

- 1) The prediction of the oceanic response combines the stratification of the seasonal thermocline with the description of the forcing, to which is added for the SSTA the influence of the mixing layer and the Coriolis.
- 2) These auxiliary parameters are the vectors of the dissonance between the two geophysical variables of the wake, even if the contribution of the ML largely dominates and imprints its dynamics on the cooling amplitude (section 6.2.1).
- 3) The origin of the asymmetries stems from the ambivalent nature of the wake as regards to the deepening. The MLD appears to be a transparent layer for the propagation of NIWs that govern the SSHA, while it is an essential metric for assessing mixing efficiency. It determines the propensity of exchanges at the interface between the thermocline and the surface layer and the degree of resonance between inertial oscillations and entrainment fluxes (section 6.2.2).
- 4) The new SSTA law reproduces the asymmetries observed between the two variables (reduction or amplification of mixing) and reduces those associated with the previously observed statistical performance (section 6.2.3). In addition, it removes the noise and scatter shown in Figure 6.8, in order to observe the real contribution of the observations (see Figure 6.15).

The conclusions we have drawn so far have been made on the basis of the analysis of the main wake component, the baroclinic response. To what extent does this contribution remain dominant in our measurement range? And what fraction does it represent in the amplitude of the transient and boundary case troughs? These questions require the use of the full model to describe the entire wake current system.



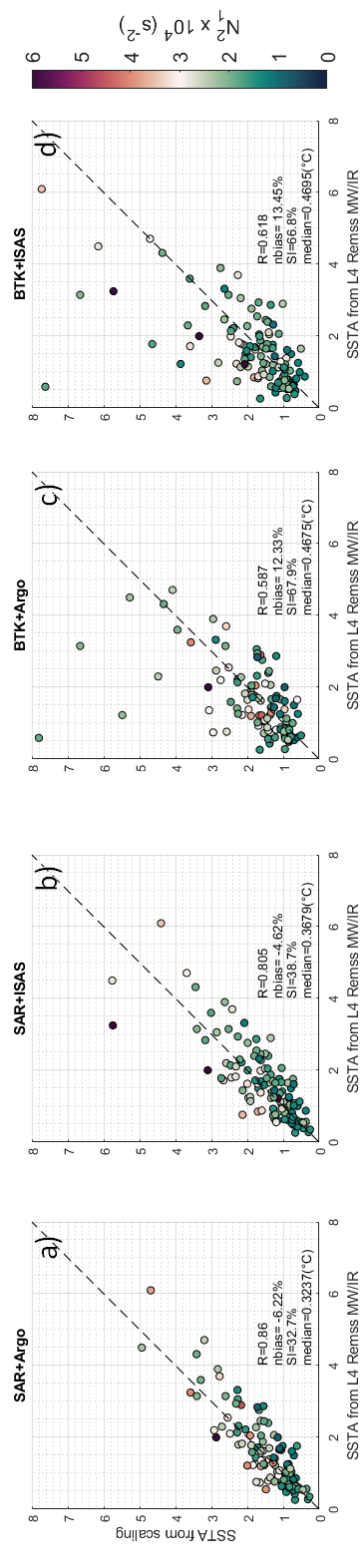


Figure 6.15: Same as 6.1.8 but using the updated SSTA inversion law.

### 6.3 Analysis of the complete solution and generalisation method of the wake anomalies estimation.

In section 1.5, we explained the complete model from which the scaling laws are derived. It is based on analytical solutions of the baroclinic and barotropic response to the cyclone passage. In order to assess the ability of the full model to interpret the signatures of the limiting cases, we have chosen to study first the particular case of Trami, which corresponds to one of the four SAR observations that we compared to the Chavas 2D model in Chapter 4, before generalising to all our cases.

As a reminder, this is a cyclone at the edge of categories 2 and 3, which was evolving in the China Sea, nearby the south of Japan. Trami was moving at an average speed of about  $3.9 \text{ m.s}^{-1}$  over an oceanic structure defined by a moderately deep mixed layer of 45-50 m, slightly deeper than the tropical average, and a seasonal stratification of the thermocline of about  $3 \cdot 10^{-4} \text{ s}^{-2}$ . The eye structure was strongly asymmetric with an Rmax of about 80 km. Finally, the velocity of the 1st baroclinic mode is estimated to be about  $3.5 \text{ m.s}^{-1}$ , which makes the Froude number close to the unitary value. Trami, due to its spatial and dynamical properties, and the oceanic conditions, presents an extremely pronounced response around 60 cm of trough, but was not considered to be totally limiting because a good fraction of its response was correctly estimated by the scaling laws (slightly above 40 cm). Trami thus corresponds to a transient case where both regimes coexist. Although the baroclinic response seems to dominate, a non negligible barotropic response exists. We wish to see if the model correctly represents the wake mechanisms for this type of transient situation.

The altimeter measurements are from a Jason-2 pass, 56h after the cyclone passed. The cyclone remained in place for about 35 hours, of the order of magnitude of the inertial period ( $\sim 30\text{h}$ ). A strong baroclinic inertial response is therefore expected. The altimeter track is therefore located at 2IP, which should allow us to observe the maximum response. In the figure 6.16, we observe a baroclinic component around 40 cm, which corresponds to our estimated value from the scaling law, and a strong barotropic response around 16 cm. The total response is in excellent agreement with the altimeter measurements. The same is true for the temperature anomaly predictions and the radiometric measurements. We also note in passing the very good agreement of the cooling observed by a pair of Argo floats collocated with the model grid. This example shows the ability of the analytical model to decompose the wake signature, the baroclinic and total components are in agreement with the values estimated by the scaling law and the altimeter respectively. Another boundary situation could be correctly anticipated by the model, that of Michael, which corresponds to a coastal situation where we have altimetry observations. The model shows an essentially barotropic response (14 cm) which is very close to the observed value (16 cm). On the other hand, the signatures of stationary cyclones such as Noru do not seem to be resolved by the model, due to the absence of the Ekman pumping contribution. We will check these results with our dataset.

The figure 6.17 shows the comparison between the SSHA/SSTA measured by the altimeters/radiometers and the estimates of the full model forced by the Holland and Chavas vortices respectively. The values provided by the predictions correspond to the total component, which includes both the displacement induced by the baroclinic and barotropic response. These results echo the disparities in performance already observed between the two forcings, with overall better response anticipation for the Chavas forc-

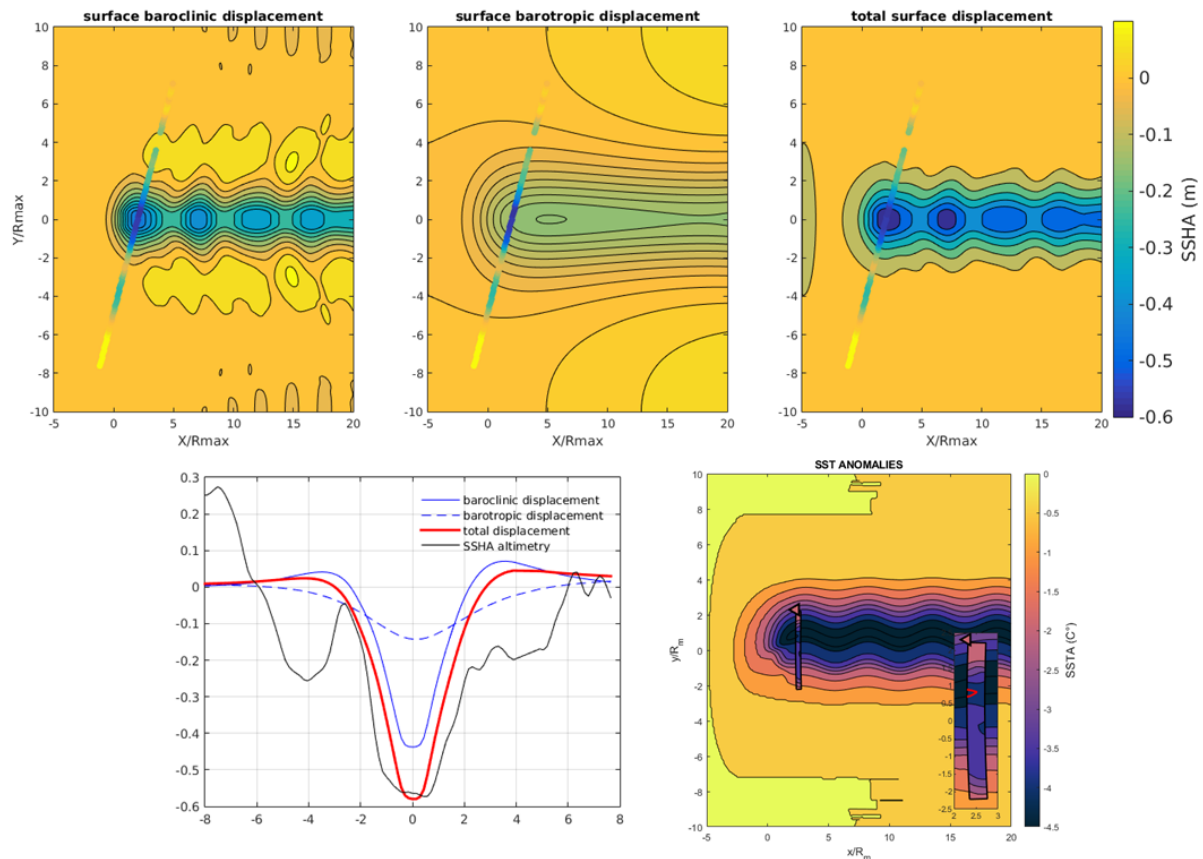


Figure 6.16: Analysis of the complete near-inertial wake response of Trami inferred from the complete model [Kudryavtsev et al. 2019a], with the horizontal sea surface anomalies a) from the baroclinic and b) the barotropic component, and c) the total response. The altimeter track of Jason-2 is also colocated with the Trami wake with the associated SSHA estimation. d) SSHA values along the altimeter transect e) SSTA derived from the model and compared with the argo measurements.

ing. The cases circled in blue represent situations where the Froude number is less than 1, and thus correspond to a predominantly barotropic regime. Contrary to the scaling laws, a part of these situations can be solved with the full model. As far as the SSHA is concerned, several situations remain limiting. These include the slow cases ( $\leq 2$  m.s<sup>-1</sup>), where a significant part of the energy is transferred to the geostrophically-adjusted surface circulation associated with the Ekman pumping, rather than to the inertial response [Geisler 1970]. As Ekman pumping has been neglected, these cases remain limiting for the full model. It can be seen that no situation exceeds 50% contribution of the barotropic component in the total surface displacement, with the exception of the Michael case, visible in panel a. Most of the variability in the observed anomalies is explained by the baroclinic component, across our cases and for the range of values observed. However, this statement must be qualified for the borderline cases, due to the absence of Ekman pumping. This outcome still converge with previous findings [Ginis 2002; Zhang et al. 2020]

The Chavas model therefore provides a better reading of the temperature and sea level anomalies than the Holland model. When coupled with the full analytical model, it provides more accurate estimates of the anomalies than the scaling laws. To conclude on these performances, and to complete our approach, we now wish to verify the quality

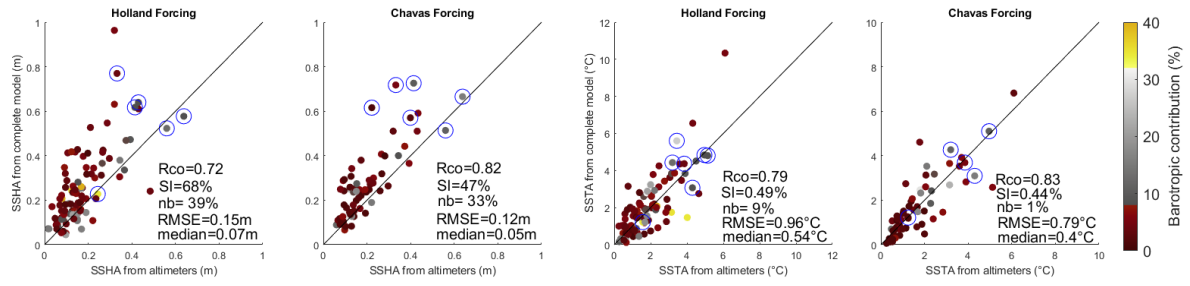


Figure 6.17: Comparison of the SSHA /SSTA anomalies derived from the complete analytical model forced by a)/c) Holland and b)/d) Chavas model

of its predictions of  $R_{max}$  in order to generalise the scaling laws in the absence of SAR observations. The Figure 6.18 shows the generalisation of our scaling laws with an  $R_{max}$  deduced from the Chavas model. In these analyses, only the baroclinic responses are of course retained. We observe an excellent correlation between the predicted SSHA and the Chavas  $R_{max}$ , which explains 75% of the variability (compared to 90% with the SAR). Despite a slight decrease in performance, this result shows that the analysis of SSHA in the absence of a high resolution forcing source can be made with the competition of an  $r17$  provided by a radiometer or a scatterometer and the Chavas model. The colorbar shows the percentage of degradation with respect to the SSHA value estimated with the SAR. In most cases, this degradation is less than 10%. The same observation can be made for the SSTA, but with a greater scattering that was similar to that observed with the SAR, without the information on the depth of the mixing layer. We also note for both figures, a slight negative bias which reflects the one observed between the  $R_{max}$  estimates of SAR and the parametric model.

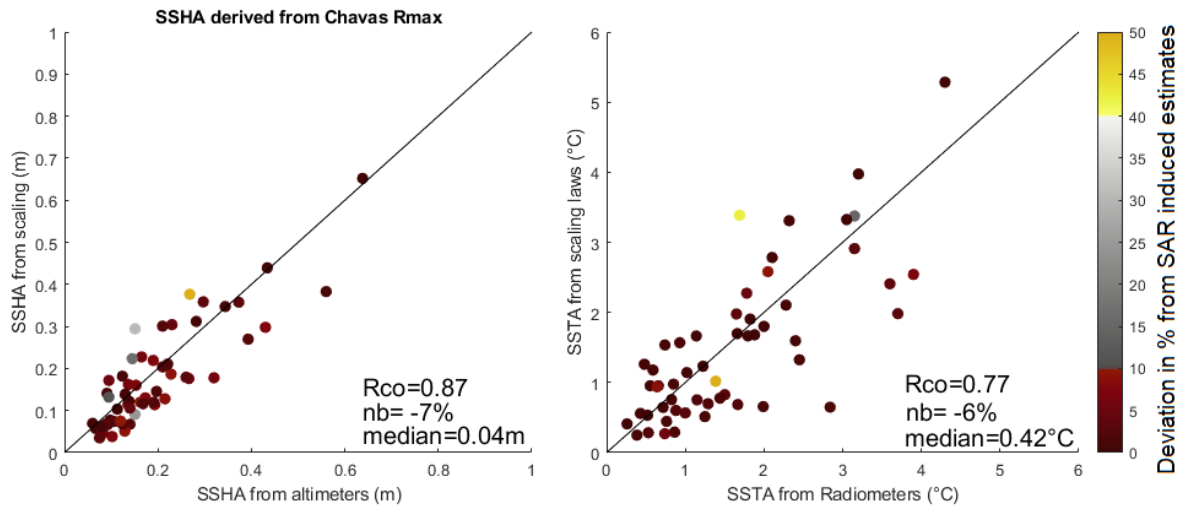


Figure 6.18: SSHA and SSTA derived from the scaling laws documented with the  $R_{max}$  from Chavas model.

# Discussion and Prospects

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## 7.1 Main outcomes.

The thorough analysis of the wake of tropical cyclones is not straightforward, as evidenced by the limited amount of work that studies the inertial response of ocean mixing in a holistic manner (see Chapter 1). Indeed, the collection of atmospheric forcing and ocean structure data, as well as the monitoring of geophysical variables before and after the cyclone, require considerable effort. To take into account the variability associated with each of the wake parameters, we had to converge different instrumental approaches and develop a robust methodology to read properly the wake anomalies (see Chapter 5). As the depiction of turbulent mixing induced by tropical cyclones encompasses several issues, four main axes have fuelled and oriented most of our speech.

### 7.1.1 Multiplatform observation.

The first and main axis concerns the creation of a multi-platform satellite base that combines the recent capabilities of the MW instruments of the last decade, without which our thesis could not exist and around which the different issues have been articulated. It is therefore the real thread of our manuscript. The article in Chapter 3 of the Monthly Weather Review summarised the progress made in wind inversion from the cross-polarised signal of high-resolution C-band SAR measurements, confronting them to 169 different cyclonic situations. Our paper demonstrated the unique ability of these instruments to interpret the full cyclonic wind field over all categories and with similar accuracy to the SFMR instruments, which have always served as a reference. These results led us to evaluate the quality of the different BTK parameters, of which the SAR offers for the first time a fully independent and homogeneous source of comparison and allows a complete inventory of the different wind radii, from the outer profile to the inner eyewall. While most BTK parameters follow the trends observed by the SAR measurements, significant uncertainties appear for increasing wind radius intensities. The largest uncertainties are



largely associated with the  $R_{\max}$  estimates, whose decay for the main cyclone categories is not observed by BTK and leads to a large overestimation. SAR observations are therefore a valuable source of information, as extremely accurate for the evaluation of  $R_{\max}$  as the SFMR, but globalised over all basins unlike the latter.

In parallel, the massive exploitation of other MW instruments demonstrates the extensive spatio-temporal monitoring of forcing and wake parameters from the current armada of radiometers, scatterometers and altimeters. While a few hundred cyclonic structures have been extracted from satellite images, more than 5,000 acquisitions have been intercepted by the coalition of active and passive low- and medium-resolution sensors over the 2010-2018 period, meeting specific quality criteria (Chapter 2). Several size parameters could be extracted from the different scenes collected, and compared to BTK data and global products, thus revealing the synergy of these observations. Equally prolific, the altimeter constellation, with six instruments, has enabled a new strategy for interpreting sea level signatures and moving away from the smoothed and more conflicting information of the daily gridded products that lack of resolution and variability. Thanks to this approach, the maximum amplitudes of the troughs can be captured and monitored in much greater detail. Several chapters have also highlighted the versatility of radiometers, both for the observation of intense cyclonic winds by L-band sensors, and for the observation of ocean signatures by multi-frequency sensors such as AMSR-2, which are incorporated into the REMSS product.

When these remote sensing measurements are linked to the large Argo float array, a 3D wake analysis is obtained, which fully characterises the pre-cyclonic conditions for over 80% of our SAR database. Chapter 6 provided the final critical arguments for the necessary use of the full set of observations for an optimal reading of the mixing processes in the cold wake. In particular, the combination of SAR with Altimeters and Argo provides an unprecedented description of the evolution of the SSHA as a function of ocean and forcing parameters. The degradation of this information (transition to higher level products) leads to a significant loss of the variability accounted for by the scaling laws, and results in an at least threefold increase in the uncertainties associated with the estimates.

Finally, to conclude on this aspect, Chapters 4 and 6 have highlighted the excellent coherence that exists between the different observations of the same nature, such as between the SAR-derived surface wind measurements and the radiometers over the majority of the wind profile, but also between the radiometers and the Argo float pairs for temperature anomalies, which indirectly attests to the quality of our different methodologies and our analyses.

### 7.1.2 Cold Wake analysis and scaling laws.

Throughout Part II, we have capitalised on the potential of the scaling laws developed in the two parent studies [Kudryavtsev et al. 2019b] and [Kudryavtsev et al. 2019a] (Annex D and E). The ingestion of high-resolution, in-situ and along-track data into this semi-empirical database allowed us to jointly depict the maximum cooling and sea level trough amplitudes with greater accuracy, but also for the first time to analyse the co-variability of the anomalies in the wake. The existing asymmetry between the two anomalies confirmed some of the model assumptions about the differences in the dynamics of the two variables, and the existing divergences. While the mixing layer is particularly transparent

in the modulation of the SSHA, which are essentially sensitive to deeper variations of the seasonal thermocline, the temperature anomalies are dependent on its dynamics (initial depth and deepening).

A new scaling law for SSTA could thus be derived for different categories of pre-cyclonic mixed layer depth, depending on the climatological ones observed in the tropics. The addition of the MLD provides a better description of the cooling in the wake, and has increased the explained variability by 20-30%.

Chapters 5 and 6 also validated the performance of the scaling laws for a very wide range of cyclonic forcing over all basins and in the case of the particularly difficult eastern Pacific region. As these laws are based solely on the baroclinic response, most cases in our database verify the dominance of this response in the observed anomalies (>80% of our cases). When the limiting cases are included in the analysis, the SSH scaling law still explains 70-80% of the observed variability, which is in line with the standards suggested by previous studies [Jullien et al. 2012; Zhang et al. 2020].

In addition, the full analytical model of the oceanic response enriches the scope of our analysis, by including cases in the shallow ocean with a predominant barotropic response, and those under transient regimes that have strong responses in both modes (like Trami). Finally, the analysis of the discrepancies between the estimates and the observations demonstrated the influence of each of the parameters in the quality of the estimates, and mainly the critical role of  $R_{max}$ .

### 7.1.3 Size parameter.

Whether for atmospheric forcing or oceanic response, the size parameter has received particular attention, whether associated in the scaling laws with the maximum wind region ( $R_{max}$ ), or more broadly with  $r_{17}$  to describe the absolute TC length scale. While in the early 2000s, wind radii are of increasing interest and are reanalysed over all basins by the NHC and JTWC centres, especially  $r_{17}$  which benefits from the valuable support of scatterometers [Sampson and Knaff 2015], there is still a dearth of information regarding  $R_{max}$ . SAR observations are therefore an ideal solution in case of acquisition, as they are the only ones able to accurately extract the region of strongest winds. The most significant decrease in scaling performance occurs overall when BTK information is preferred to SAR information (SST and SSH). The strong dependence of the ocean response estimates on  $R_{max}$  is also clearly visible when comparing the evolution of SSHA as a function of cyclone intensity and information source (BTK/SAR), where the strongest discrepancies observed for BTK-guided predictions is a direct consequence of the loss of sensitivity of the  $R_{max}$  estimates for categories 3 and above within the BTK.

Similarly, the TC size is also a crucial issue in the representation of the forcing, in particular to inform the gradient of the outer core wind decay and its extent. While the amplitude of the response is of course dependent on the region of maximum winds, the structure and wavelengths of the near-inertial waves are strongly dependent on the size of the forcing [Ginis 2002; Shay 2009]. Indeed, the spatial scale of the cold wake is correlated with that of the cyclone [Zhang and Lin 2019]. The correct representation of the inertial wake by the full analytical model therefore requires both an accurate description of the inner core and the absolute size of the wind field. To this end, we have used the plethora of information from the constellation of radiometric and scatterometric instruments to lead

a new parametric wind model, whose performance to derive the wind profile from  $R_{\max}$  to the outermost wind radii is superior to that of classical  $R_{\max}$ -based models (Chapter 4).

### 7.1.4 Generalization of our approach.

As a final major focus of the thesis, the addition of a general wind field solution allows both to compensate for the higher temporal intermittency of the SAR instruments (due to their acquisition method) and to force the full ocean response model. The main advantage of the Chavas parametric solution is that it relies on the information of  $V_{\max}$ , which is strongly estimated by the Dvorak analyses and the available measurements incorporated in the BTK analyses, and an outer profile radius such as  $r_{17}$ , which is easily measured by the set of satellite measurements of sea surface winds. Unlike traditional models, it reconstructs the complete profile from the  $r_{17}$  and  $V_{\max}$  information and derives the  $R_{\max}$  value from the intersection of the two solutions (inner and outer core). When the azimuthal distribution of  $r_{17}$  and the translation speed are appended, the complete 2D field can be derived.

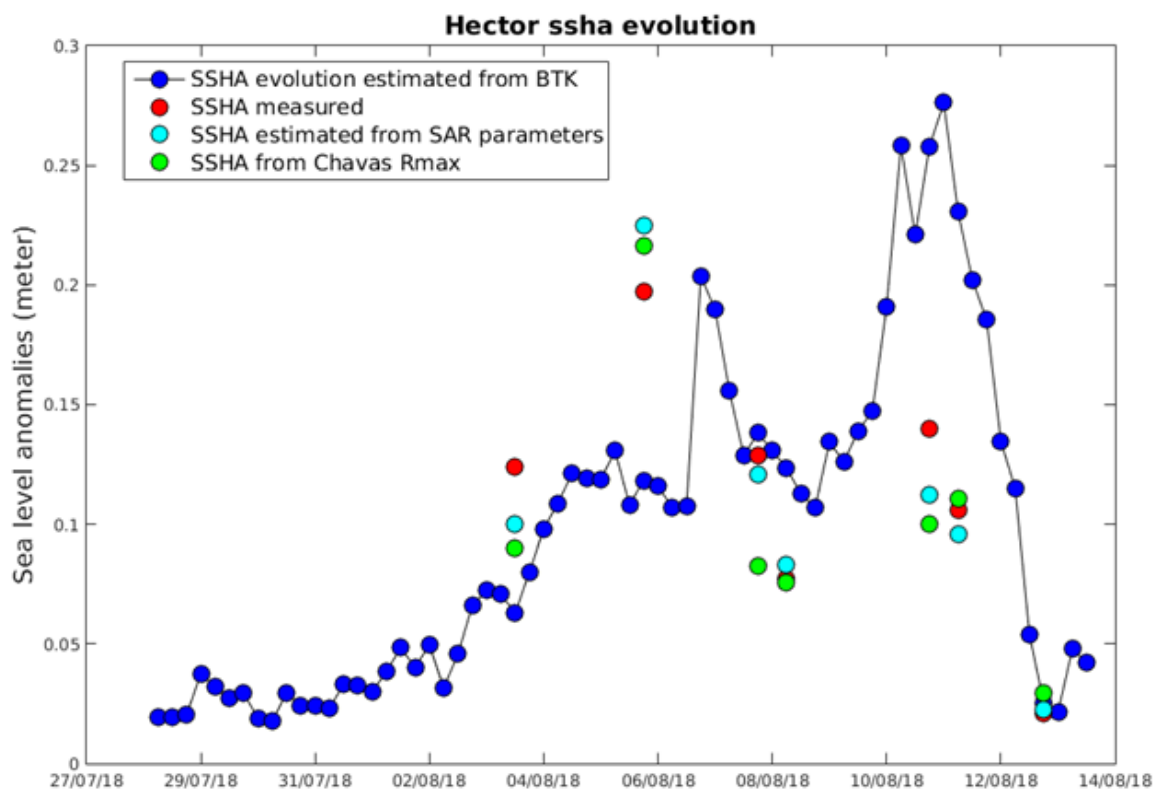


Figure 7.1: Same as Figure 5.10 but with the additional SSHA estimates derived from the  $R_{\max}$  calculated from the parametric model.

The ability of this parametric vortex to reproduce the wind profile and to guess the  $R_{\max}$  was examined using the previously exploited set of MW observations. These different comparisons have shown qualitatively on some particular situations and quantitatively on the whole SAR database, the excellent performances of the model to interpret the great majority of tropical systems, above the cyclonic intensity threshold, along the whole profile. Comparisons with azimuthally averaged SAR wind profiles have also shown that

the model is able to capture most of the inner core geometry described at first-order by translation speed, and that guided by the r17 distribution for the rest of the structure. Furthermore, the model provided reliable estimates of  $R_{max}$  over a relatively wide range of 0-60 km, with a small negative bias. Ultimately, Chapter 4 demonstrated that the model skills are optimal from 2.5  $R_{max}$  for all cyclonic categories, and from 2.5 to 5  $R_{max}$  for tropical storms with an organised structure ( $> 25 m.s^{-1}$ ).

The Chavas model thus represents an alternative path that has superior qualities to the classical  $R_{max}$ -based models, as illustrated by the comparison with the Holland vortex. These statements are even more true when ingesting and comparing the two types of parametric solution into the full analytical ocean response model, for which the best estimates of sea surface anomalies were obtained with the Chavas model (Chapter 6). The  $R_{max}$  deduced from the model can also directly document the scaling laws and be used as a general solution in case of no SAR acquisition, it results in a better anticipation of surface anomalies than with the BTK  $R_{max}$ , and gives predictions almost similar to those displayed with SAR values (with slight decrease of performance), which is perfectly illustrated once again with our well-known case of Hector (Figure 7.1).

Finally, the different wind radius comparisons carried out with the BTK have shown the same large discrepancies with both MW instruments and the Chavas model, which is an additional proof of the uncertainties impacting some BTK parameters (e.g R33).

## 7.2 Limits and prospects.

### 7.2.1 Methodology and new instrumental strengths.

In order to complete our conclusions, we must now mention the limitations of our study and the perspectives it raises. The great strength and the great weakness of our analyses stem simultaneously from our methodology for extracting anomalies. While it allowed us to assess the stigma of cyclonic forcing with great precision, it also raises questions about the repeatability of our results. Indeed, our method relies on a subjective analysis, somewhat similar to Dvorak one, with a recognition of cyclonic signatures and patterns related to pre-cyclonic mesoscale activity, which can be particularly time consuming and difficult to adapt to a very large sample. The analysis of high-resolution surface wind data also requires a lot of effort to correctly interpret the cyclonic wind field, but also to gather all the observations needed to interpret the cold wake of cyclones.

While the objective of this thesis was clearly to provide the most accurate description of the wake using current observational methods, it also demonstrated the need to use a multi-platform approach to fully capture the variability of signatures and parameters influencing the three-dimensional response of the cyclone. This thesis should therefore be seen as a blueprint whose very promising potential is closely linked to the capabilities of future instruments over the next two decades (2020s and 2030s). Indeed, a new generation of instruments is emerging, with the addition of cross-polarised signal technology within the second generation of scatterometers onboard MetOp-SG, directly inspired by the progress made with SAR measurements [Dagestad et al. 2013]. This new sensor series should allow a better reading of intense cyclonic winds by scatterometers, without the saturation issues associated with Bragg scattering mechanisms. The development of this type of measurement should improve the monitoring and ease the extraction of inner-core parameters, by complementing the more intermittent high-resolution SAR observations,

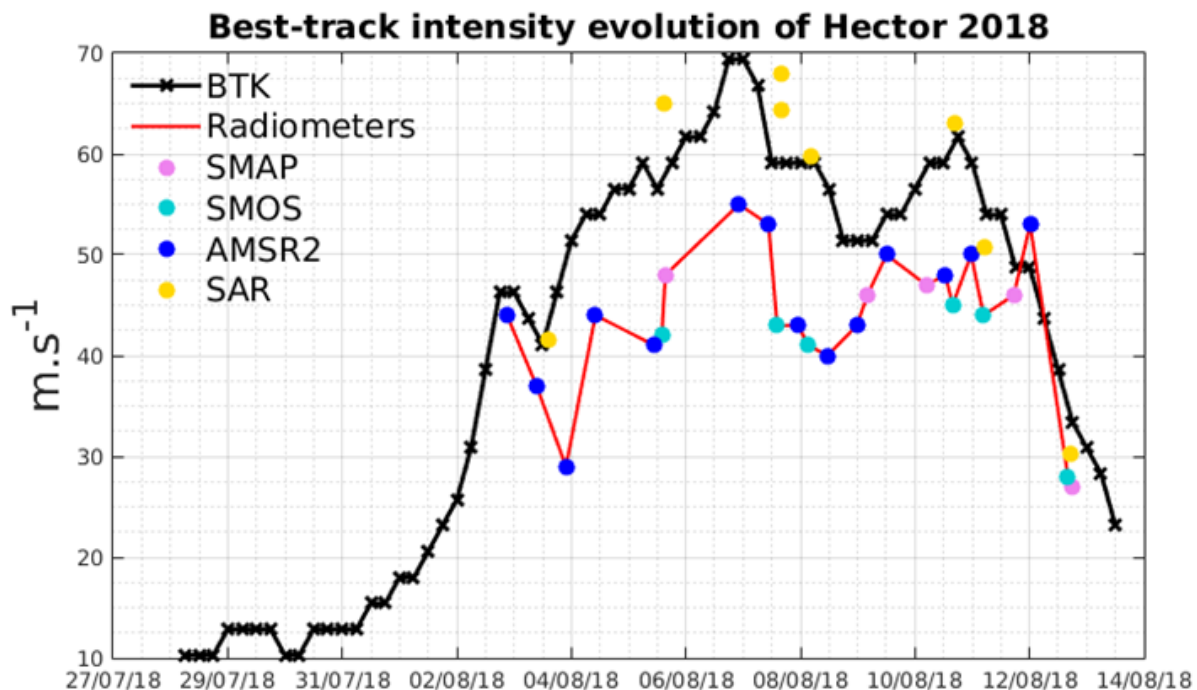


Figure 7.2: *Temporal evolution of Hector Vmax (black line) compared to the intensities extracted from the interpolated observations of radiometers and SAR.*

and are also expected to be in good agreement with the radiometers, due to the excellent correlation existing between the brightness temperatures of the excess of foam emissivity and the cross-polarised normalized radar cross-section under strong wind regime [Zhao et al. 2018]. Although the coarse resolution of passive sensors does not always allow maximum winds to be captured, they remain a valuable tool for assessing intensity and especially its trend [Reul et al. 2017], as illustrated in the Figure 7.2, which shows the excellent alchemy between the SAR observations that capture intensity and the numerous radiometric acquisitions that allow its temporal variations to be followed in the case of Hector. The addition of the new generation of scatterometers could bridge the gap between the more prolific and coarse radiometric measurements and the more precise and sporadic SAR measurements. The addition of these observations should also benefit BTK analyses, and extend the influence of scatterometers on the estimation of inner core parameters, including Rmax.

An even more important revolution is also underway with the emergence of a new type of altimeter that will provide for the first time a 2D mapping of the ocean surface topography, using SAR interferometric techniques, the Surface Water and Ocean Topography (SWOT) Mission [Morrow et al. 2019]. Combined with other remote sensors and in-situ observations, SWOT can resolve most of the mesoscale and sub-mesoscale activity down to a wavelength of 15-30 km depending on sea state, in contrast to previous along-track altimeter data alone (60-70 km) or the global product ( $\sim 150$  km) [Morrow et al. 2019]. Because of the strong signatures left by the cyclone along their tracks, the induced SSHA should be easily observable by SWOT, which will provide for the first time a 2D view of the horizontal structure of the inertial wake, similar to the surface cooling observed by radiometers. The swath is narrower<sup>1</sup>, but should, if well collocated, capture the structure of the maximum troughs. This new observation will therefore have several advantages in

<sup>1</sup>120 km with 20 km gap at nadir, [Morrow et al. 2019]



the analysis of the ocean response:

- 1) A portrayal of the surface wake structure that can be directly compared to the complete analytical model, and which will provide accurate description of the TC trench length scale.
- 2) A better spatial and temporal monitoring of the sea level anomalies that will help to track the persistence of the signatures. We have seen in chapter 1 that disturbances generated through turbulent mixing and spread of near-inertial waves leave long-term anomalies, particularly in the subsurface, where the surface flows cannot participate in their recovery and they can persist on a scale of several months. At the surface, the cooling signature remains on average up to 30 to 40 days, with a maximum generally observed 2 to 3 days after the passage of the cyclone and a e-folding period around 10 to 15 days [Dare and McBride 2011; Vincent 2011], exactly as we have observed with our database (see Figure 7.3, e-folding time  $\sim 12$ ). Performing the same analysis on SSHA proved to be more challenging. Part of our work, which has not been presented in this thesis, has been to examine the temporal evolution of the troughs along the track, using the repeatability cycle of the Jason series altimeters. As SSH is an integrated response and very dependent on the deeper dynamics of the thermocline, these signatures can persist for much longer time, (SSHA on the Igor example lasted over 90 days, Figure 1.35). However, assessing the persistence of these signatures with our extraction method makes the task laborious and complicated for the more evanescent signatures. Interpolating Jason measures with SWOT data offers a possible solution for automation without using a daily gridded product.
- 3) a better reading of the pre-cyclonic mesoscale activity (eddy), which can improve our methodology.

The combined use of these new instruments (SWOT and MetOp-SG) with existing technologies should allow simplified wake reading and better sampling of cyclonic forcing, which could lead to further automation of our methodology. In addition, recent modifications of IBTrACs [Knapp et al. 2018] now include a three-hourly analysis. If this is an interpolation that takes into account the multiplication of geostationary instruments, it should not necessarily impact the quality of parameters such as  $V_{max}$ , which remains constrained by rules dictated in the Dvorak proceed [Song et al. 2018], or  $R_{max}$  and  $R_{33}$ , which remain limited by the nature of the observations. On the other hand, it should provide a better estimation of the successive positions of the cyclone, and thus of the translation speed, the only parameter that we cannot estimate outside of the BTK.

### 7.2.2 Limits of the parametric wind solution.

The parametric model developed by Chavas et al. [2015] has major advantages over the classical  $R_{max}$ -based model, although its application has some limitations. The model underestimates the size of the TC eye, the too sharp ascent of its winds leads to a (generally slight) widening of the eyewall structure. Although in most cases this is not really a problem, as the different oceanic responses are generally more sensitive to the maximum forcing and/or the area of sustained winds [Geisler 1970; Price 1981; Ginis and Sutyrin 1995; Kudryavtsev et al. 2015], for large cyclones with a wide eye this can lead to an over-estimation of the area of strong winds and a misinterpretation of the oceanic response by the full analytical solution. In contrast, there is also a threshold beyond which the predictions of  $R_{max}$  are erroneous and underestimated by the parametric vortex, located

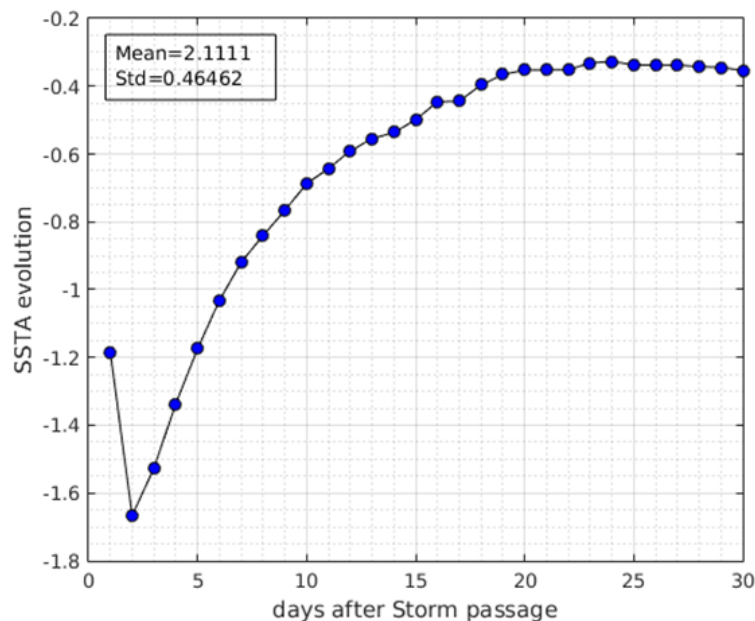


Figure 7.3: *Temporal evolution of sea surface cooling induced by tropical cyclone turbulent mixing.*

around 60 km. This limit had already been reported by [Chavas and Lin 2016], and may be due to internal processes that are not taken into account by the model, as the variations in  $R_{max}$  are dictated by the intensity information and especially the influence of the size parameter, the model therefore does not predict such a large radius for intense cyclones with respect to partial angular momentum conservation. While this threshold (60 km) only concerns a small fraction of the cases above the cyclonic intensity ( $\sim 5\%$ ), it is not so uncommon for tropical storms, which also suffer from larger uncertainties along the radial profile. Their more ambiguous results are mainly due, for less intense situations, to disorganised structures that do not fit the assumptions of the model, or, for systems that already have a clear circulation (generally above  $25\text{--}28\text{ m.s}^{-1}$ , Vigh et al. [2012]), to an irrelevant choice of the wind radius. Indeed, for these cases we should probably choose an  $r_{12}$  or even an  $r_{10}$ , as the  $r_{17}$  is no longer describing an outer radius for tropical storms, and may be too close to the inner core, but this would imply a greater reliance on information from global products such as ECMWF to fill in incomplete wind radii due to swath boundaries (Chapter 2).

Finally, we also noticed a slight overestimation of the maximum winds when we azimuthally averaged all the radial profiles of the cyclone, the description of the  $V_{max}$  asymmetry is thus slightly biased, the model predicts on average a stronger intensity distribution. Several parallel works have been carried out on this issue to improve the description of the asymmetry, and have not been presented in this manuscript. One approach has been to extract the azimuthal distribution of  $V_{max}$  from the ECMWF surface wind fields, if the predictions systematically underestimate the intense winds, one could instead focus on the asymmetry factor, i.e. the ratio of each azimuthal value to the true maximum value. An alternative to the ECMWF data could be the advent of cross-polarised scatterometers, which would allow regular access to this asymmetry factor. Finally, another possibility would be to take advantage of the work conducted by [Wang et al. 2022] on the Chavas model (but  $R_{max}$ -based version), which includes a wind shear term in the definition of asymmetry.

### 7.2.3 Limitations and evolution of wake analysis.

Finally, we can make several criticisms of our wake analysis and discuss some interesting perspectives. Firstly, the main shortcomings concern:

- 1) The absence of the SSS. As we have seen, salinity plays a key role in modulating the ocean response, but also in explaining the asymmetries between temperature and sea level anomalies. While the surface cooling is a more superficial response [Vincent 2011], SSSA are mainly related to the thermocline dynamics. It would therefore be interesting to see how the SSH and SSS anomalies are related to each other. While at the time of the development of our methodology, the temporal resolution of the gridded products and the accuracy of the individual observations did not guarantee a meaningful examination of the salinity anomalies, Reul et al. [2021]; Sun et al. [2021] showed that this was possible with the composite products (SMOS and SMAP), due to the pronounced cyclone-induced signature ( $\sigma(1psu)$ ) compared to the variability generally observed (0.1 psu, [Wunsch 2015]).
- 2) The absence of the Ekman upwelling contribution limits the scope of the full model, and underestimates an unknown fraction of the barotropic response for slow and/or very large systems, which prevents an accurate assessment of the dominance of this response mode. This is one of the possible uses of SWOT mission. The future combination of SAR cases along with SWOT observations, should allow an accurate spatial description of the TC wake. A new scaling law similar to the one developed for baroclinic cases could be established for the barotropic regime. Due to the properties of its orbit and resolution, SWOT is also expected to provide a better view of coastal areas [Morrow et al. 2019]; this would allow for broader analyses of TC barotropic signatures in shallow water.
- 3) The use of daily SST products. Given our efforts to use under-track observations, it may make sense in the future to perform the same procedure to characterise cyclone-induced cooling, especially since the large collection of radiometers allows for the individual use of different sensors. Part of the higher scattering of the SSTA may come from the smoothing that still exists in the SST L4 products.

Finally, several more or less immediate perspectives can be foreseen in the light of our results and are the next goals:

- If the work carried out by Reul et al. [2021]; Sun et al. [2021] allows the monitoring of salinity anomalies, Pivaev et al. [2022] has attempted to parameterise this response, similarly as [Kudryavtsev et al. 2019b]. Previous work conducted by Kudryavtsev et al. [2015] has also derived an analytical model to describe the significant wave height distribution along the main transect of tropical cyclone, using observation of the constellation of altimeters. By extrapolating our methodology and instrumental means to all these different semi-empirical bases, we should obtain an even more complete picture of all the geophysical markers of cyclone-induced ocean response.
- The combined use of the conventional altimeter constellation and SWOT mission should make it easier for us to track sea level anomalies over time, and if we take advantage of its observations of the horizontal surface wake field, we can then estimate changes in the OHC in the same way as Mei et al. [2013], with unprecedented accuracy.

- The most impending prospect remains the estimation of the drag coefficient, using the same bottom-up approach as described in Kudryavtsev et al. [2019a]. As the scaling laws are also derived from the simplification of the stress to the intensity alone ( $\tau_s \sim V_{max}^2$ ), reusing the wind stress definition and inverting the SSHA scaling law gives the following relationship:  $C_d \propto (\rho_w/\rho_a)(V_{fm}/N_1 R_m)(gSSH A/V_{max}^2)$ . In contrast to the analysis performed in Kudryavtsev et al. [2019b], we can take advantage of the SAR and Argo information to refine the  $C_d$  trend as a function of wind intensity. The Figure 7.4 compares the different  $C_d$  trends according to the SAR and BTK information. The dispersion shown in panel b is reminiscent of the one displayed in the parent study Kudryavtsev et al. [2019b]. This scattering is again the result of the lack of variability of BTK, mainly in the estimation of  $R_{max}$ . While both panels confirm the decay of  $C_d$  for increasing intensities Powell et al. [2003]; Jarosz et al. [2007]; Holthuijsen et al. [2012]; Zou et al. [2018], the gradient and the dispersion around the mean values are not at all similar. The  $C_d$  behaviour deduced from the SAR parameters describes a less sharp decay. The use of high-resolution forcing data and in-situ data within the semi-empirical database allows the estimation of  $C_d$ , which is an essential parameter in the exchanges, through the analysis of wake anomalies. However, this estimation is only partial for the moment. The  $C_d$  values calculated represent a trend, as they are deduced from a proportionality relationship, whose coefficient we still have to evaluate in order to compare them with the observed values. The calculation of this coefficient must be done using the complete model and the calculation of the constraint, which will be one of the immediate follow-ups of our thesis work, with the necessary improvements to the parametric model of Chavas et al. [2015].

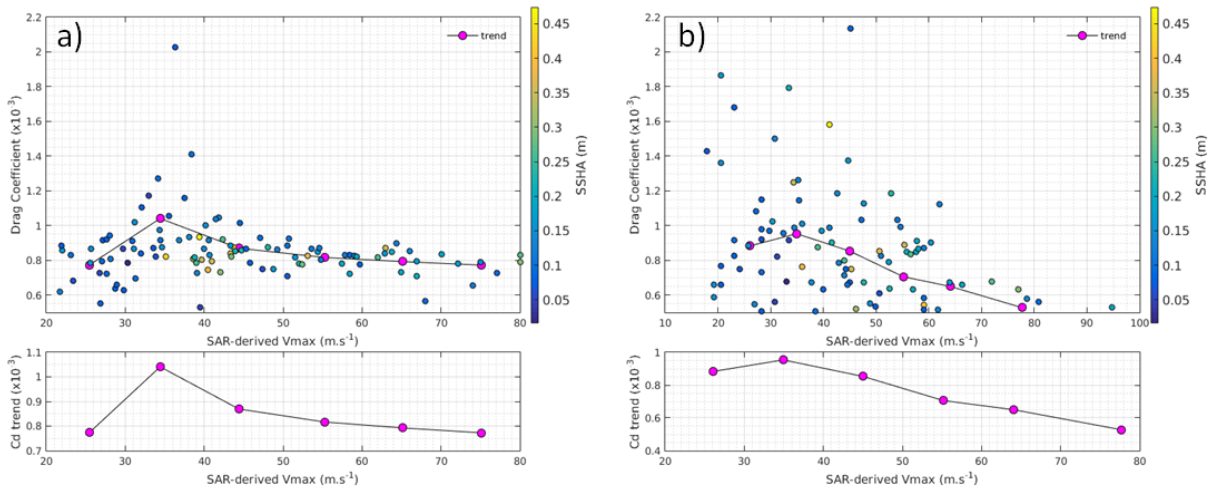


Figure 7.4:  $C_d$  estimate as a function of TC intensity, derived from forcing information of a) SAR and b) BTK. The black line with the magenta dots represents the running average over  $-5/+5$  m.s<sup>-1</sup>.

Globally, the use of the recent instrumental capabilities associated with the scaling laws of the oceanic response has allowed us to establish a very broad panorama of the oceanic response induced by tropical cyclones, and to describe precisely the variabilities associated with each of the parameters involved in the dynamics of the cold inertial wake. With the proliferation of new MW instruments, it is becoming more than relevant to use this type of multiplatform database, and even necessary in the context of cold wake

analysis due to the extreme wind conditions, the spatial scale and temporal variations of the ocean features, and the countless interactions that occur. Although the evolution of cyclone activity in the face of climate change is still undetermined, several lines of evidence suggest that there will be an increase in the number of TCs that undergo fast intensification episodes [Leroux et al. 2018]. This increase would be very problematic, as these events remain extremely limiting for BTK analyses and difficult to predict. Although SAR observations are currently the only means of accurately analysing the surface wind field of cyclones in all basins [Mouche et al. 2019; Combet et al. 2020a], they are not yet routinely examined or included in an operational procedure [Knaff et al. 2021]. The work of this thesis therefore also advocates for greater versatility in operational analyses to integrate current and future MW instrumental strengths, which suffer from some latency, to meet the needs of the tropical cyclone community [Knaff et al. 2021].



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# Supplement material for Part I.

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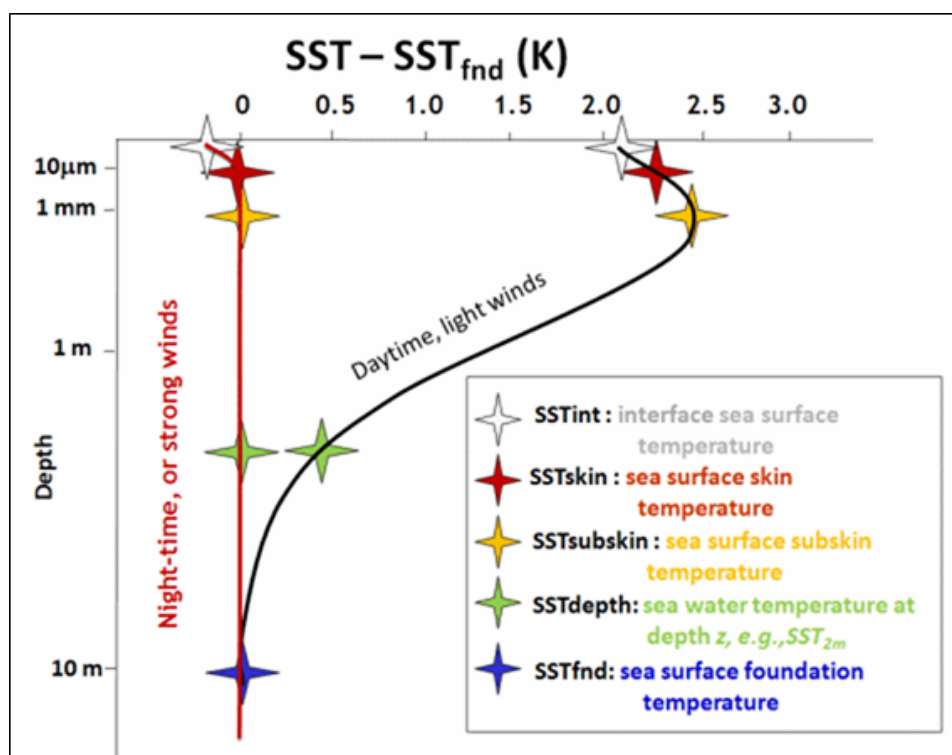


Figure A.1: the different definitions of SST depending on the depth of the measurement. The amplitude variations are shown as a function of the wind regime and whether it is daytime or nighttime. The SST<sub>fnd</sub> is the surface temperature stripped of diurnal effects. SSTs obtained from radiometers are usually adjusted to create an SST<sub>fnd</sub> using a diurnal model.

## Tropical cyclone Track history

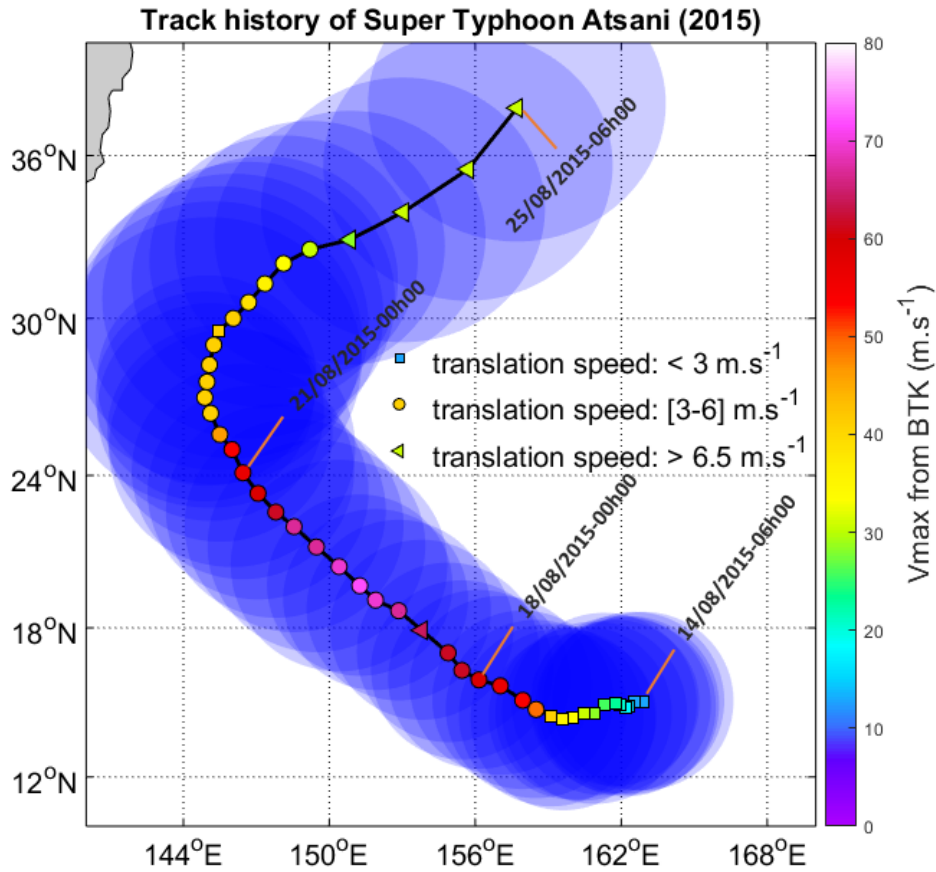


Figure A.2: *BTK positions of super Typhoon Atsani (2015) as a function of vortex intensity (colorbar) and motion speed (average in circles and fast in triangles). The blue shaded area indicates the extent of the radius of the outermost closed isobar (ROCI).*

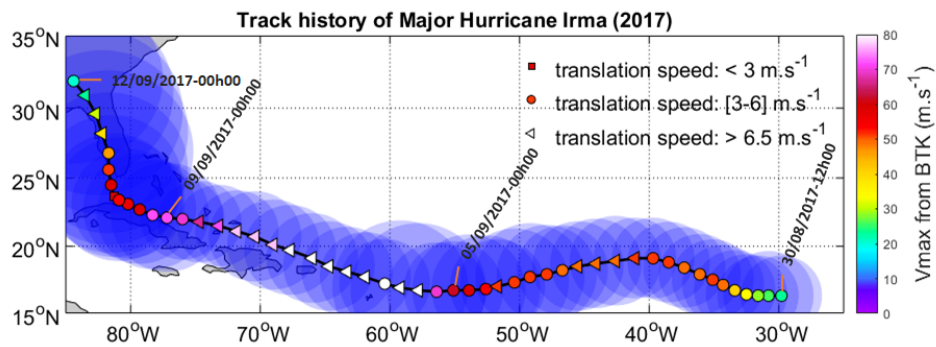


Figure A.3: *Same as A.2 but for Major Hurricane Irma (2017).*



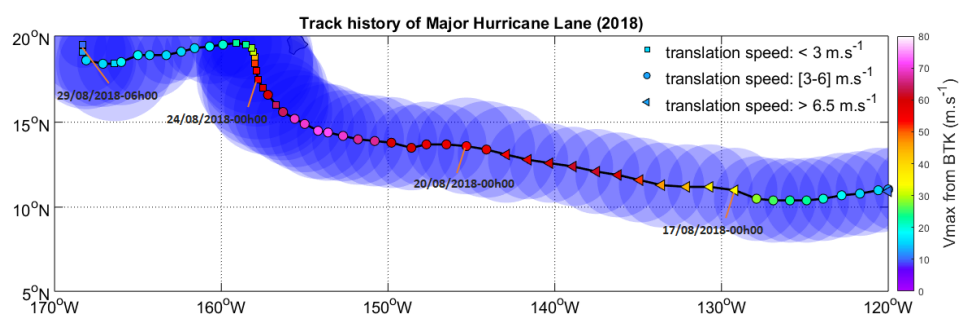


Figure A.4: Same as A.2 but for Major Hurricane Lane (2018).

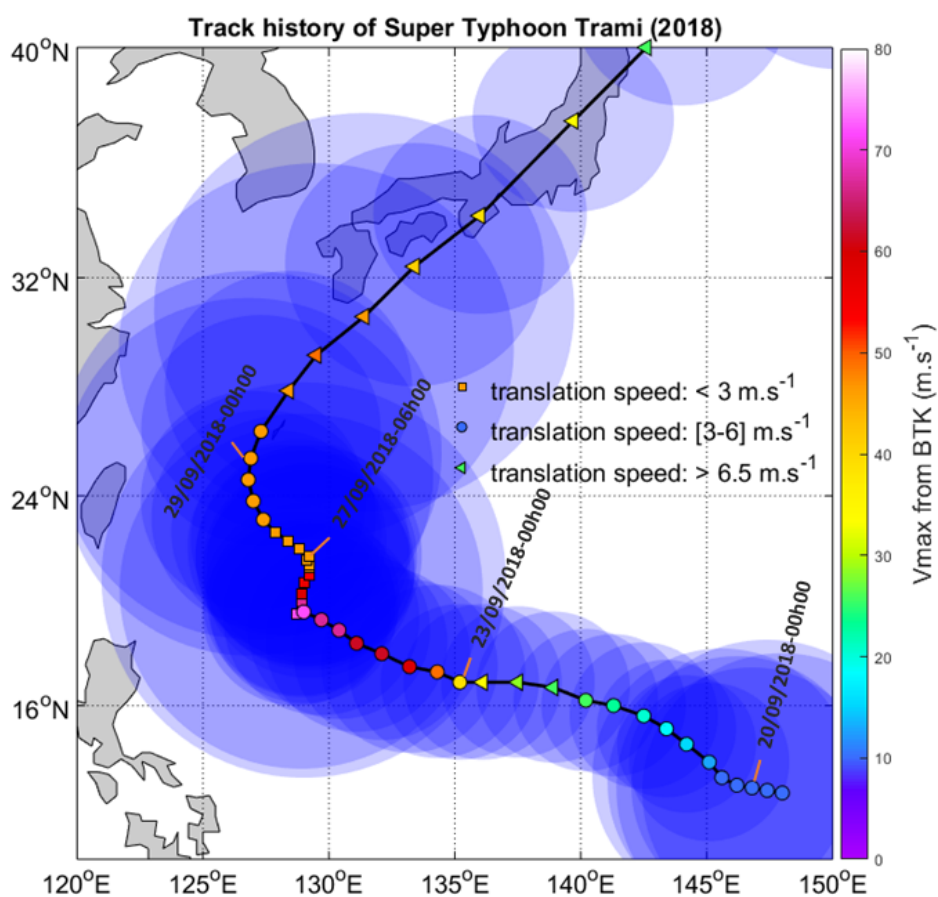


Figure A.5: Same as A.2 but for Super Typhoon Trami (2018).

# High resolution observation of Irma wind field with co- & cross-polarization SAR signal

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## RESEARCH ARTICLE

10.1029/2019JC015056

## Special Section:

The Three Major Hurricanes of  
2017: Harvey, Irma and Maria

## Key Points:

- No saturation of the cross-polarized C-band ocean backscattered signal for surface wind speeds up to 75 m/s is observed
- Combined copolarization and cross-polarization are used for ocean surface wind retrieval from Sentinel-1 and Radarsat-2 SAR over major hurricanes
- Irma category 5 hurricane wind structure is described at high resolution O(1 km) and compared with other existing and independent analysis

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## Copolarized and Cross-Polarized SAR Measurements for High-Resolution Description of Major Hurricane Wind Structures: Application to Irma Category 5 Hurricane

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**Abstract** C-band high-resolution radar (synthetic aperture radar [SAR]) is the only spaceborne instrument able to probe at very high resolution and over all ocean basins the sea surface under extreme weather conditions. When coanalyzed with Stepped Frequency Microwave Radiometer wind estimates, the radar backscatter signals acquired in major hurricanes from Sentinel-1 and Radarsat-2 SAR reveal high sensitivity in the cross-polarized channel for wind speeds up to 75 m/s. The combination of the two copolarized and cross-polarized channels can then be used to derive high-resolution surface wind estimates. The retrieval methods and impacts of intense rainfall are discussed in the context of a Hurricane Irma (2017) case study. On 7 September 2017, Sentinel-1 measurements intercepted Hurricane Irma when it was at category 5 intensity. When compared to Stepped Frequency Microwave Radiometer, SAR-derived wind speeds yield bias and root-mean-square of about 1.5 and 5.0 m/s, respectively. The retrieved wind structure parameters for the outer core are found to be in agreement with the Best-Track and combined satellite- and aircraft-based analyses. SAR measurements uniquely describe the inner core and provide independent measurements of the maximum wind speed and the radius of maximum wind. Near the radius of maximum wind a 65-m/s increase in wind speed in less than 10 km is detected, corresponding to an instantaneous absolute vorticity of order 210 times the Coriolis parameter. Using a parametric Holland model and the environmental surface pressure (1,011 hPa), SAR-derived wind speeds correspond to a central surface pressure of 918 hPa (921 hPa from the Best-Track) in Irma's eye.

### 1. Introduction

While a variety of sources for tropical cyclone (TC) wind data is now available (e.g., (Reul et al., 2017)), the only routine observations in the high-wind inner core region comes from aircraft reconnaissance, limited to TCs occurring in the North Atlantic and East Pacific. In this context, C-Band high-resolution radar (or SAR for synthetic aperture radar) is the only spaceborne instrument able to probe and uniquely quantify, at very high spatial resolution O(1 km), ocean sea surface information under extreme conditions (Fu & Holt, 1982; Horstmann et al., 2005, 2013; Katsaros et al., 2000; Zhang & Perrie, 2012). Indeed, SAR measurements can be acquired day and night, regardless of the cloud coverage, with pixel resolution of few meters in swaths of several hundred kilometers. In addition, recent SAR missions, with enhanced polarization diversity, have led to new prospects to use radar measurements (cross section, Doppler) to retrieve geophysical parameters (Mouche et al., 2012). In particular, Radarsat-2 capabilities to measure the backscattered signals in copolarization and cross-polarization (antenna emits in Vertical polarization and receives in Horizontal polarization; or vice versa) opened for new analysis strategies (Kudryavtsev et al., 2013, 2014), to help reveal and interpret surface roughness changes related to upper ocean dynamics. For extreme weather events, such as TCs, the expected high sensitivity of cross-polarized signals to ocean wave breaking further translated into a new potential: the use of these new cross-polarized signals to map, at very high resolution, variations in ocean surface winds in and around the TC eyes (Mouche et al., 2017; Zhang & Perrie, 2012). Early demonstrations rapidly triggered vast interests, leading to refinements of future concept missions (Fois et al., 2015) to best exploit this high sensitivity to high wind conditions. In that context, EUMETSAT (European Organization for the Exploitation of Meteorological Satellites) together with ESA (European Space Agency) already



planned to add a cross-polarized channel for the next generation of operational scatterometer mission (i.e., the next Polar System Second Generation) dedicated to the ocean surface wind vector measurements at medium resolution (Stoffelen et al., 2017).

The new European SAR mission, namely, Sentinel-1, also benefits from these polarization diversity. As part of the European Copernicus program, the continuity in data acquisitions will be ensured for the next decade. In the present study, the motivation is thus to further document the benefit of having a cross-polarized channel to characterize and map ocean surface winds in extreme TC weather conditions. As reported below, investigations are performed at high resolution, as available with SAR missions, but results also directly contribute to the next generation of European medium-resolution scatterometers. Moreover, the focus is to further assess the potential of using both Sentinel-1 and Radarsat-2 SAR measurements and more specifically to report on cases corresponding to very intense events, that is, category 3, category 4, and category 5 TCs. Based on Sentinel-1 acquisitions over the major Hurricane Irma (2017), we illustrate how combined C-band dual-polarized SAR measurements can uniquely provide quantitative information on the wind structure in both the inner and outer cores.

For our purpose, the analysis concentrates on specific acquisitions over North Atlantic TCs, to maximize the chance to compare satellite measurements with coincident airborne ones. The goal is first to more precisely document the relationship between the radar cross-polarized cross-section measurements and ocean surface wind speeds during very extreme events, with maximum wind speeds higher than 50 m/s. Cases have thus been selected with respect to their intensity and, more importantly, for the collocation opportunities with airborne measurements from the Stepped Frequency Microwave Radiometer (SFMR) instrument. The SFMR provides independent measures of rain rate and ocean surface wind speed (Klotz & Uhlhorn, 2014; Uhlhorn et al., 2007). Then, we consider particular acquisitions during Irma, one of the strongest North Atlantic TCs recorded during the 2017 season. To recall, most previous studies hardly sampled wind speed conditions higher than 40 m/s (Hwang et al., 2015) or 50 m/s (Mouche et al., 2017). Section 2 presents the data sets we used and the collocation method we applied. Section 3 focuses on the relationship between the normalized radar cross section (NRCS) and the wind speed as given by SFMR measurements. Section 4 is the Irma case study, detailing the wind field obtained from SAR and discussing the results with respect to independent analysis.

## 2. Data Set

### 2.1. Sentinel-1 and Radarsat-2 C-Band SAR

Sentinel-1 mission is part of the European and operational Copernicus program space component. This is a constellation of two satellites (S-1A and S-1B units). Both Sentinel-1A and Sentinel-1B carry a C-band SAR and provide continuity with previous European (ERS-1 and ERS-2) and ENVISAT missions. Sentinel-1A and Sentinel-1B were launched in April 2014 and 2016, respectively. They have four exclusive imaging modes: Interferometric Wide swath (IW), Extra Wide (EW) swath, Strip Map (SM), and Wave (WV) modes. This study solely concentrates on IW mode. The IW swath is 250 km wide and covers incidence angles from about 30° to 46°. When processed into Level-1 (L1) GRDH (Ground Range Detected High resolution), IW Sentinel-1 products have a resolution of about 20 m in range (across track) and 22 m in azimuth (along track). C-band Radarsat-2 SAR was launched in 2007 and is the first SAR able to provide dual-polarized (and quad-polarized) images (VV+VH or HH+HV) with different acquisition modes including different resolutions and swath dimensions. In this study, we rely on Radarsat-2 SCANSAR Wide (SCW) imaging mode, with incidence angles ranging from 20° to 49° and a resolution of about 100 m in both range and azimuth directions. These two RS-2 SCW and S1 IW wide swath modes allow acquisitions in dual polarization, two images being acquired over the same area at the same time. One is in copolarization (VV or HH), and the other is in cross polarization (VH or HV). In this study we only use L1 data acquired in dual polarization (VV+VH) to take benefit of the two polarization channels for the SAR wind retrieval (Mouche et al., 2017). To note, the EW Sentinel-1 acquisition mode has also the dual-polarization capability with a larger swath (400 km wide) more adapted to hurricane monitoring. Moreover, CSA (Canadian Space Agency) will launch in 2019 the Radarsat-2 mission follow-on (Radar Constellation Mission), and ESA already planned Sentinel-1C and Sentinel-1D to ensure continuity up to 2030.

At C-band over the ocean, the NRCS acquired in cross polarization is typically 10 to 100 times lower than in copolarization, depending on radar incidence angle and geophysical conditions. The NRCS accuracy and the signal-to-noise ratio are thus the main limitations for applications based on cross-polarized signals. The



potential of Sentinel-1 dual-polarized acquisitions for ocean applications (Mouche & Chapron, 2015) and a first algorithm for ocean surface wind measurements over extremes (Mouche et al., 2017) have already been presented. Sentinel-1 mission requirements for radiometric accuracy is 1 dB. According to measurements acquired over DLR (Deutsches Zentrum für Luft- und Raumfahrt) transponders, the relative radar cross section for IW modes has been evaluated in 2017 at  $-0.06 \pm 0.16$  dB for Sentinel-1B and at  $-0.1 \pm 0.15$  dB for Sentinel-1A. Sentinel-1 IW and Radarsat-2 SCW Level-1 data analysis give noise equivalent sigma zero (NESZ) values between  $-30$  and  $-23$  dB depending on elevation angle. The NESZ is different for each subswath and further range dependent (i.e., modulated across the subswaths) inside each subswath (see Figure 2 in Mouche & Chapron, 2015). For Sentinel-1, since this preliminary assessment, ESA has already improved the NESZ estimates with respect to range direction but also azimuth direction in order to mitigate subswath jumps and azimuth scalloping (Miranda et al., 2017). The Level-1 processor has been recently revised accordingly (IPF V2.90) to annotate more accurate NESZ in the product. A technical note on the annotated noise in Sentinel-1 Level-1 products has been published and details all stages to properly correct the NRCS from noise (Piantanida et al., 2017). In this study, we strictly rely on Sentinel-1 products reprocessed with IPF V2.90 to take into account the latest and most accurate annotated noise available, and we apply the technical note methodology.

## 2.2. Hurricane Tracks

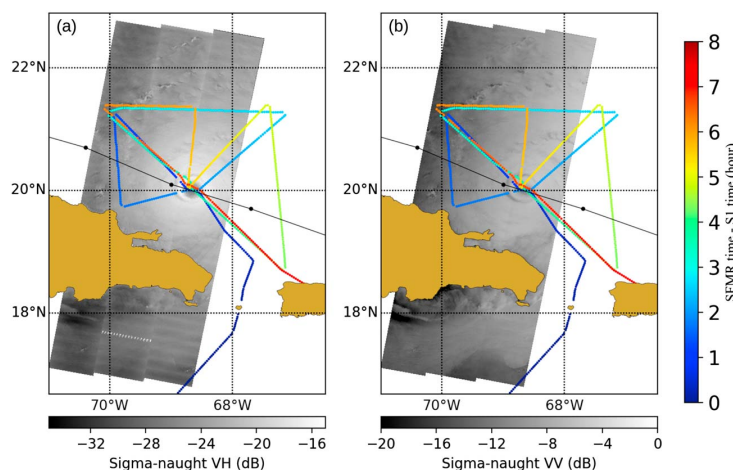
Best-Tracks are post-storm analyses of the intensity, central pressure, position, and size of tropical and subtropical cyclones. They are produced by Regional Specialized Meteorology Center or Tropical Cyclone Warning Center (RSMC and TCWC) after the hurricane season and include all available data from buoys, platforms, and surface and satellite (no SAR) observations to provide TC parameters at each synoptic hour (00, 06, 12, and 18 hr). Additional information such as wind radii or time and intensity of landfall can also be provided by RSMC and TCWC. Given the different ocean basins, the variety of involved agencies, and the nonhomogeneity of the different sources, building a homogeneous (format and quality) and global database is not straightforward (Kruk et al., 2010). This is the main purpose of the IBTrACS database (Knapp et al., 2010). This data set gathers all available Best-Tracks for all storms and all basins and follows the World Meteorological Organization (WMO) recommendations. In this study, we use the IBTrACS database because it provides storm intensities and the wind structure parameters such as wind radii (64, 50, and 34 knots), or maximum sustained wind radius. To note, in case of unavailability of these Best-Tracks (as they are produced several months after the hurricane season), one can rely on the Automated Tropical Cyclone Forecasting System proposed by Sampson and Schrader (2000). Indeed, this system, fed by National Hurricane Center (NHC; for east Pacific and north Atlantic basins) and Joint Typhoon Warning Center (JTWC; for the others), was designed to optimize forecasting processes and warning messages issues and to maintain a tracks archive for past and active storms (Miller et al., 1990). These “operational Best-Tracks” are produced in near-real time and are reanalyzed after the season to feed the HURDAT database (Rappaport et al., 2009). Although some of the parameters can suffer from large errors (e.g.,  $>25\%$  for radius of maximum wind speed (Landsea & Franklin, 2013; Schreck et al., 2014) or discrepancies between different agencies (mostly in Western Pacific basin as shown by Knapp et al., 2010; Sampson et al., 2018), Best-Tracks remain a valuable source of information regarding the evolution of a TC during its whole lifetime (Knapp et al., 2010). In this study, because SAR observations are not used to produce hurricane tracks (including reanalysis of Best-Tracks), they are used as independent measurements for comparison.

## 2.3. Combined Satellite Aircraft Analysis

In this study, a combined satellite aircraft analysis is used for comparisons with SAR-based wind retrievals. This analysis combines the multisatellite platform TC surface wind analysis (Knaff et al., 2011) with aircraft reconnaissance data collected from the real-time high-density observations distributed via the global telecommunication system. Reconnaissance data contain flight-level wind, temperature dew point, and altitude information and surface wind speed estimates from SFMR. The aircraft collect data over several hours, and for most of the sorties sample in an  $\alpha$ -shaped pattern (see Figure 1); making multiple passes through the center of circulation.

The temporal composite of data occurs over a 6-hr time window, and the final analyses are centered at the synoptic times. To perform the data composite, cubic splines are used to estimate the position of the storm at the time of the observation from which a radius and azimuth is calculated following the storm. Storm positions come from a combination of the operational best track (before and after the aircraft sortie) and the routine aircraft-based center fixes (during the time the aircraft is sampling the storm).





**Figure 1.** Maps of normalized radar cross section (NRCS) as measured by Sentinel-1A over Irma category 5 hurricane on 7 September 2017. (a) NRCS in VH polarization. (b) NRCS in VV polarization. The locations of SFMR measurements after collocation are overlaid with color code indicating SFMR-derived ocean surface wind speed. SFMR = Stepped Frequency Microwave Radiometer.

The variational data-fitting methodology of the analysis follows that described in Knaff et al. (2015) where flight-level (700 hPa) wind vectors and SFMR wind speeds (corrected to flight level) are analyzed on a polar grid with radial resolution of 2 km and azimuthal resolutions of 10°. The enhancements of wind speeds to flight level and the reductions of the final analysis to a 10-m marine exposure follow recommendations in Franklin et al. (2003). The two ways these analyses differ from those used in Knaff et al. (2015) are that the satellite-only analyses are used as initial wind estimate or first guess and that a residual analysis is performed. The residuals are analyzed using the same variational methodology with smaller azimuthal and radial data filters. When the residuals are added back into the analysis, the resulting winds are closer to the observations as the fine-scale residual analysis acts much like multiple passes through a Barnes (1964) analysis as demonstrated in Koch et al. (1983). Following the final analysis and residual correction, winds are reduced to a 10-m marine exposure and inflow angles applied as suggested in Zhang and Uhlhorn (2012). If a point is actually determined to be over land, winds are reduced an additional 20% and the inflow angle is increased an additional 20°. These winds were then interpolated to a  $0.05^\circ \times 0.05^\circ$  Cartesian grid.

The resulting analysis produces a spatially continuous data set that closely fits the input data, objectively replicating what a detailed subjective analysis of the data would produce. It is noteworthy that the variational analysis tends to more heavily weight flight-level vector wind components compared to the scalar SFMR wind speeds because (1) wind speed is the scalar magnitude of the wind vector and (2) the cost function minimizes the sum of errors in tangential vectors, radial vectors, and wind speeds. Thus, the resulting analysis is fundamentally different than one produced using SFMR wind speeds alone. We will refer to this as the Multi-observation Tropical Cyclone Surface Wind analysis (MTCSSWA), hereafter.

#### 2.4. Next-Generation Radar

NEXRAD (for Next-Generation Radar) is a network of 160 high-resolution Doppler weather radars operated by the National Oceanic and Atmospheric Administration (NOAA) National Weather Service, the Federal Aviation Administration, and the U.S. Air Force (USAF). They are located in the contiguous United States, Alaska, Hawaii, U.S. territories, and at military base sites. Thirty-four stations out of the 160 stations are located in coastal areas. NEXRAD Level-II and Level-III products are routinely available from the NOAA National Centers for Environmental Information. Level-II products provide the three meteorological base data quantities: reflectivity, mean radial velocity, and spectrum width, while there are over 75 different Level-III products such as storm relative velocity, vertical integrated liquid, or 1-hr precipitation (Federal Meteorological Handbook No. 11—Doppler Radar meteorological observations (WSR-88D). Part C: WSR-88D products and algorithms, 2017). In this study, we only consider Level-III products of base reflectivity. The radar scan time is 4.5 min, and the resolution is 1 km in range direction and 1° in azimuth with a



radius coverage of about 450 km. Note that the 1-hr precipitation product has been discarded as it is much less correlated with SAR signal due to the large (60 min) integration time.

### 2.5. Stepped Frequency Microwave Radiometer

Each Atlantic and East Pacific hurricane season, the Hurricane Research Division conducts a field program in which they collect data with the NOAA and, since 2007, the USAF aircrafts. Airplanes fly directly into the eye of the hurricane, usually adopting an “alpha” or “multi-alpha” pattern (see Figure 1), to best sample the four geographical quadrants. They carry SFMR, a stepped frequency microwave radiometer providing wind speed and rain rate estimates. Initially operated in 1980, SFMR measurements are now routinely acquired by NOAA and USAF aircrafts and operationally processed to be transmitted to the NHC (in Miami and Honolulu). The wind estimate principle relies on the use of six different C-band frequencies between 4.5 and 7.2 GHz to correct for rain-induced effects. SFMR measurements' temporal resolution is 1 s. The validation of retrieved wind speeds against Global Positioning System (GPS) dropwindsondes using 3 years of data by Uhlhorn and Black (2003) for a range of wind speeds from 10 to 60 m/s yields to a root-mean-square error of 3.3 m/s and indicates a possible bias due to ocean surface waves. SFMR processing algorithm has been further revised by Uhlhorn et al. (2007) to extend the range of wind speed up to 70 m/s and remove a remaining bias for moderate SFMR-measured wind speeds (10–50 m/s). Because the SFMR design involved a single nadir-viewing antenna, the data only provide transects of the wind speeds and not the full two-dimensional wind-speed pattern of the hurricane system. However, the combined estimates of rain rate and ocean surface wind speed makes this instrument unique for SAR wind speed validation and to discuss the possible rain impact on SAR measurements.

### 2.6. SAR Data Collection

To date, SAR missions cannot continuously acquire wide swath data in high-bit rate modes. As a consequence, tasking SAR with respect to the hurricane tracks forecast is required to jointly maximize acquisitions over TCs and mitigate the impact on the whole acquisition plan. This implies solving potential conflicts between users regarding the duty cycle along the orbit and the acquisition modes to be used over a given area of interest. Since 2016, ESA organizes specific S-1 acquisition campaigns to test the instrument capabilities for mapping, at very high resolution, extreme (TC) ocean wind conditions. These campaigns of dedicated acquisitions are named as SHOC, for Satellite Hurricane Observations Campaign (Mouche et al., 2017). Fully coordinated as for the hurricane watch program approach (Banal et al., 2007), SHOC campaigns help maximize the number of SAR acquisitions for both Copernicus/ESA Sentinel-1 and also MDA/CSA Radarsat-2 missions. For this study, the strategy to collect the data includes two different approaches. (i) As part of SHOC, we collect the data through acquisitions requests for both Radarsat-2 and Sentinel-1 missions, respectively, to MDA and ESA. These requests were based on 5-day forecasts of the hurricane track and satellite orbit. This approach requires great flexibility for the data provider. (2) We also analyze the SAR data archives and the maximum wind speed with respect to the hurricane Best-Tracks database. The period considered for the archive analysis is from 2015 and 2017. We only applied the second method to the Radarsat-2 data. Indeed, SHOC provided all dual-polarized acquisitions over TC since the beginning of Sentinel-1 mission. Then, as a final and common step, we apply two additional criteria to consider only SAR data acquired (1) over hurricane eyes when maximum wind speeds are larger than (or equal to) 50 m/s and (2) with airborne data available.

In total, the SHOC campaign provided six SAR observations corresponding to four different TCs in 2015–2017. The list of the SAR observations, associated TC name, and intensity category during the observations are given in table 1. Three have been acquired during category 5 intensity, one during category 4 TC intensity, and three during category 3 intensity. An example of SAR acquisition and the collocated SFMR measurements locations is shown for Sentinel-1 data acquired in VH (Figure 1a) and VV (Figure 1b) polarizations over Irma TC, on 7 September 2017. In Figure 1, SFMR measurements locations are indicated by colored dots, and the color code indicates the time difference between SAR acquisition and SFMR measurements. In this case, up to five transects across the eye have been performed for a flight lasting about 8 hr.

## 3. Data Analysis

### 3.1. Collocation

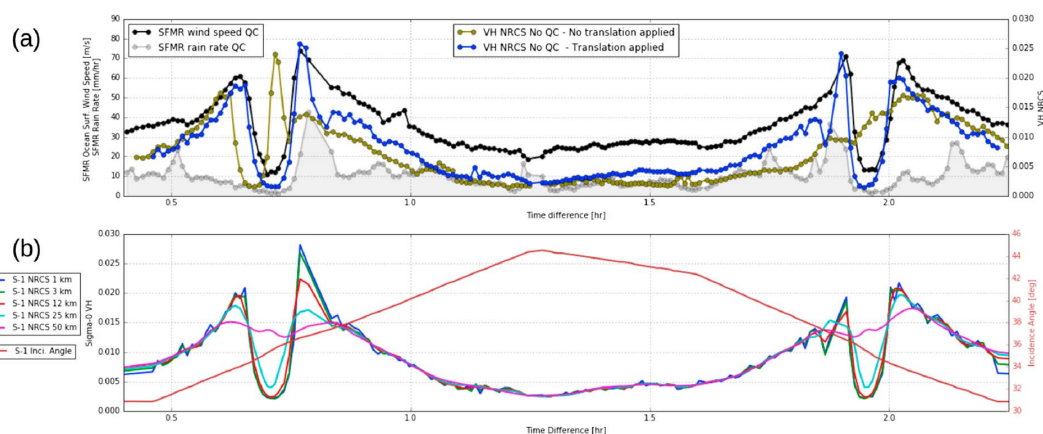
The collocation of rain and ocean surface wind speed measurements involves several steps. First, SFMR measurements are smoothed using an averaging moving window of 10 s and a spatial resampling at 3 km. During

**Table 1**  
SAR Acquisitions and Corresponding Hurricane Information

| Sensor      | Acquisition mode     | Polarization | SAR observation   |                 | TC Name | TC category before/after | Collocated SFMR |
|-------------|----------------------|--------------|-------------------|-----------------|---------|--------------------------|-----------------|
|             |                      |              | starting date UTC | ending date UTC |         |                          |                 |
| Radarsat-2  | ScanSAR              | VV, VH       | 2015/10/03        | 2015/10/03      | Joaquin | 4/4                      | 1               |
|             | Wide                 |              | 10:44:58          | 10:45:12        |         |                          |                 |
| Radarsat-2  | ScanSAR              | VV, VH       | 2016/09/02        | 2016/09/02      | Lester  | 3/3                      | 1               |
|             | Wide                 |              | 03:53:04          | 03:53:17        |         |                          |                 |
| Sentinel-1A | Interferometric Wide | VV, VH       | 2017/09/07        | 2017/09/07      | Irma    | 5/5                      | 1               |
|             |                      |              | 10:29:51          | 10:31:10        |         |                          |                 |
| Radarsat-2  | ScanSAR              | VV, VH       | 2017/09/08        | 2017/09/08      | Irma    | 5/4                      | 2               |
|             | Wide                 |              | 10:53:52          | 10:54:04        |         |                          |                 |
| Radarsat-2  | ScanSAR              | VV, VH       | 2017/09/18        | 2017/09/18      | Maria   | 3/5                      | 1               |
|             | Wide                 |              | 22:07:22          | 22:07:36        |         |                          |                 |
| Radarsat-2  | ScanSAR              | VV, VH       | 2017/09/22        | 2017/09/22      | Maria   | 3/3                      | 1               |
|             | Wide                 |              | 10:45:38          | 10:45:51        |         |                          |                 |

Note. Dates are formatted as year/month/day.

this preprocessing step, the quality flag included in SFMR data can be used to possibly filter out low-quality data. Second, the hurricane translation speed is computed from the hurricane track. The locations of SFMR measurements are then shifted with respect to the time difference between each SFMR measurements and the SAR acquisition time using the TC motion vector. The duration of a SAR acquisition is typically few seconds (see the fourth and fifth columns in Table 1), whereas a flight with SFMR can last up to 6 hr and more. SAR time is thus considered constant with respect to SFMR times. Starting acquisition time is used for SAR acquisition time. If several acquisitions performed in a row are necessary to describe the whole hurricane, the duration of SAR acquisitions increases. For instance, in the case of Irma presented in Figure 1, a total of four successive acquisitions are concatenated to produce the image, leading to an acquisition duration of 1 min and 29 s. Finally, for each SFMR storm-motion-relative location, oceanic SAR measurements are averaged within a radius  $R$ . Note that this step results in a collocated SAR data set with a pixel spacing of 3 km but a spatial resolution depending on  $R$ . In this study we used  $R = 1.5, 2.5, 6, 12.5$ , and 25 km to mimic different spatial resolutions.



**Figure 2.** Transect of collocated SFMR and SAR measurements at 3-km resolution. (a) Quality-controlled SFMR ocean surface wind speed (black) and rain rate (gray). NRCS in VH polarization with (blue) and without (olive) taking into account Irma translation speed during the collocation. (b) Same Sentinel-1A NRCS transect as in (a) but for different resolutions (1, 3, 12.5, 25, and 50 km). Variation of SAR incidence angle along the transect is indicated in brown (right y axis). SFMR = Stepped Frequency Microwave Radiometer; SAR = synthetic aperture radar; NRCS = normalized radar cross section; QC = quality controlled



An example of collocation obtained for Irma between Sentinel-1A and SFMR is given in Figure 2. Figure 2a shows the SFMR ocean surface wind speed and rain rate estimates as a function of time. Here the quality flag included in the SFMR data has been taken into account. *X* axis indicates the time difference with respect to SAR acquisition time. In the present study, we only focused on collocations with an absolute time differences less than 2.5 hr. In this case, measurements up to 75 m/s have been measured by SFMR. These large values of wind speed strongly correlated with the highest values of rain rate, corresponding the western part of the TC eye, very close to the radius of maximum wind (RMW). The nonflagged observations appear to contain wind estimates where rain rates, which are discussed later, are up to 40 mm/hr. To note, flagged measurements reach rain rates up to 80 mm/hr (see Figure 8). This suggests that the quality procedure removes measurements for the most intense rain rates. Figure 2a also presents the collocated cross-polarized NRCS (blue) with respect to time corresponding to the same SFMR measurements. As observed and already reported with SAR data from Radarsat-2 mission, there is a very strong correlation between SFMR ocean surface wind speeds and cross-polarized NRCS at C-Band (Zhang & Perrie, 2012). Figure 2a also shows the match-ups that result if the collocation procedure is not performed. The olive line shows NRCS that has not been aligned with TC motion. In this case the strong correlation between SAR and SFMR measurements is completely lost.

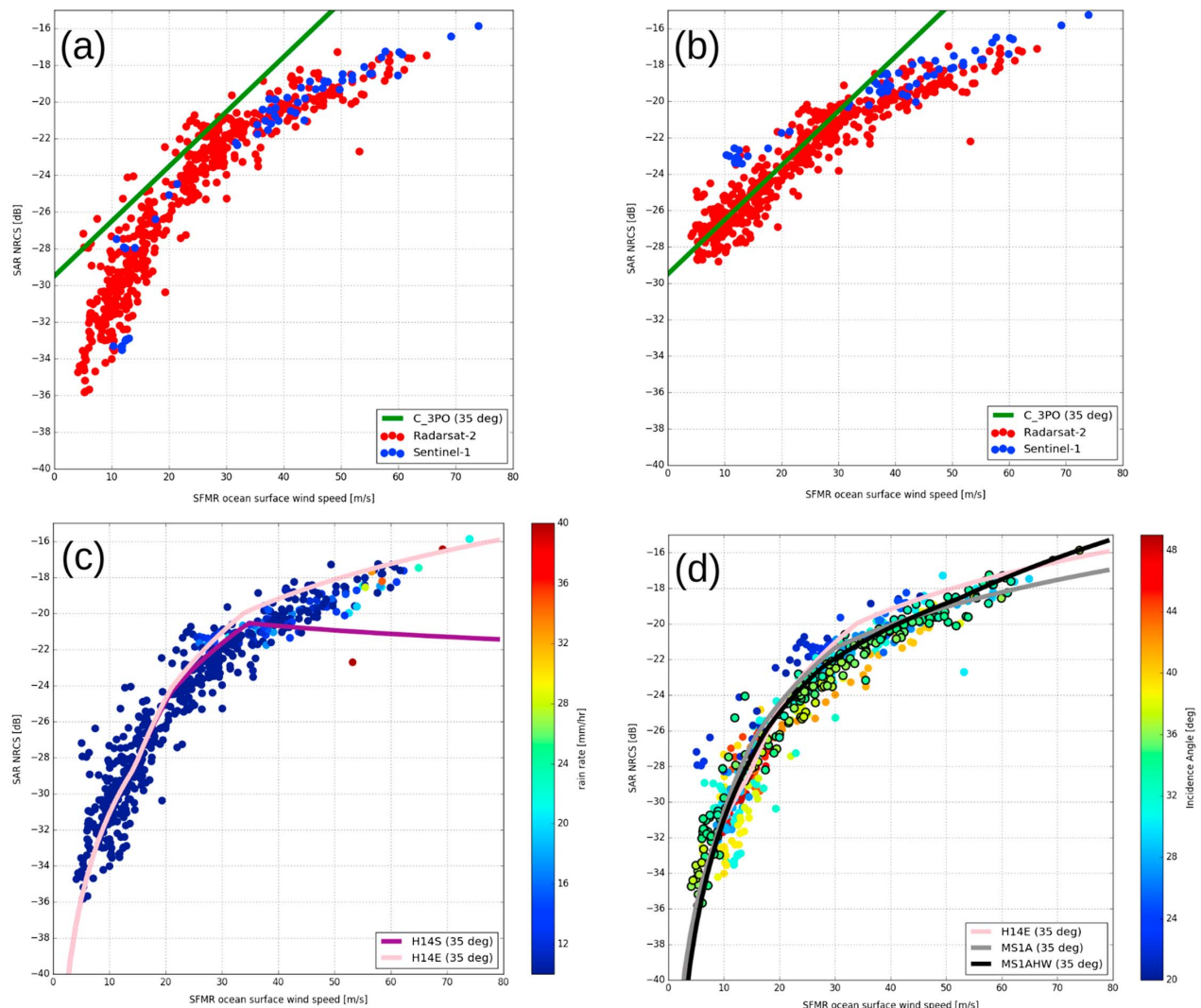
The impact of the different spatial resolutions is presented in Figure 2b. This comparison shows the benefit of high-resolution compared to medium-resolution missions such as L-band radiometers (about 40 km for Soil Moisture Active Passive (SMAP); Meissner et al., 2017; Yueh et al., 2016; or Soil Moisture and Ocean Salinity (SMOS); Reul et al., 2012) or scatterometers (about 25 km for nominal resolution with the next European SCA on MeTop SG; Stoffelen et al., 2017). In this case, larger averaging areas blur the interpretation of the data, clearly showing that both the central wind minimum and peakedness of the wind maxima are severely impacted at spatial resolutions exceeding 12 km. In fact, at the time of acquisition, Irma eye diameter was about 30 km (about 16 nautical miles) as given by the Best-Track analysis, corresponding to less than three measurements at 25-km resolution. When the pixel sampling decreases, results are even worse (not shown). More generally, the use of SAR acquisitions with NRCS computed at different resolutions to systematically mimic lower resolution sensors can certainly help to characterize the resolution impact on the measurements depending on TC characteristics, particularly in the high wind-speed gradient areas near the RMW.

### 3.2. Radar Backscattering Versus SFMR Measurements

Using a 3-km resolution, all collocated measurements between SAR and quality-controlled SFMR measurements can now be used to more precisely document the relationship between cross-polarized NRCS and ocean surface wind speeds, including local extremes up to 80 m/s. For this purpose, Sentinel-1 and Radarsat-2 data are combined.

As explained in section 2, and especially for weak cross-polarized backscattered signals, the signal-to-noise ratio is a key measurement, highly dependent upon the accuracy of the noise annotated in the SAR products. To note, noise corrections were often neglected (Zhang & Perrie, 2012) or not always properly taken into account for the entire data set (Hwang et al., 2015). Here, all SAR data are noise-corrected before collocation.

The impact of the noise correction on the NRCS relationship with ocean surface wind speed is illustrated in Figure 3a without noise correction and in Figure 3b with noise correction. Both cases show that the NRCS clearly increases with ocean surface wind speed, without any apparent NRCS saturation in wind speeds up to 75 m/s. With or without noise correction, these comparisons confirm the substantial potential of C-Band cross-polarized NRCS for retrieving oceanic surface wind speeds in TC environments over TCs, including category 4 and category 5 hurricanes. Nonetheless, NRCS to wind relationships, obtained with and without noise correction, are impacted by noise over the whole range of wind speeds. As anticipated, this impact can become particularly significant for the lowest branch of the high wind speed range (wind speed  $\approx 20$  m/s). Without noise correction, our present analysis consistently recovers previously reported relationships (or GMF, for geophysical model function) between NRCS and wind speeds, especially those developed for wind speeds lower than 30 m/s (Zhang & Perrie, 2012; Zhang et al., 2017). The application of noise correction fully explains the difference observed between our analysis and reported GMFs in this range of wind speeds. For higher than 30 m/s wind speeds, these GMFs are not adapted and simply fail to explain the collected data. The present analysis demonstrates that noise-corrected NRCS values increase from about  $-36$  dB at 5 m/s up to  $-16$  dB at 70 m/s. This analysis also reveals, for the first time, the great consistency of the two



**Figure 3.** Three-kilometer Sentinel-1A (blue) and Radarsat-2 (red) NRCS in VH polarization (a) with noise correction and (b) without noise correction, and as a function of the ocean surface wind speed measured by SFMR. Green solid line stands for the GMF proposed by Zhang et al. (2017). (c) Same as (a) but with color code indicating SFMR rain rate. Pink and purple solid lines respectively indicate H14E and H14S GMF from Hwang et al. (2015). (d) Same as (a) but with color code indicating SAR incidence angle. Pink and gray solid lines respectively indicate H14E GMF from Hwang et al. (2015) and MS1A GMF from Mouche et al. (2017). NRCS = normalized radar cross section; SFMR = Stepped Frequency Microwave Radiometer; GMF = geophysical model function; SAR = synthetic aperture radar.

different SAR missions (RS2 and S-1), regarding their calibration. This clearly demonstrates the potential to systematically combine the different instrument measurements in order to increase the temporal sampling of a given storm. In the following, we only consider NRCS corrected for noise annotated in the product.

Comparisons with other existing GMFs are presented in Figure 3c. The two GMFs proposed by (Hwang et al., 2015) display significantly different behaviors for wind speeds higher than 35 m/s—note the relative saturation in the H14S GMF. The present analysis reveals that, between 35 and 70 m/s, H14E GMF is clearly more adapted than H14S GMF. For this figure, the rain rate measured by SFMR is also reported. In our analysis, data show no evidence of significant degradation due to the presence of precipitation. For wind speeds between 5 and 30 m/s, the rain rate is lower than 20 mm/hr, and its effect on NRCS computed at 3 km





is hardly visible. Few measurements with rain rate between 20 and 40 mm/hr and wind speeds larger than 30 m/s have been collected, and in those cases, precipitation impacts are unclear. It is however noteworthy that around 52.5 m/s wind speed, an outlier in NRCS is obtained for 40 mm/hr rain rate. This latter result tends to indicate possible significant decreases of NRCS for high rain rates. The impact of rain rates on NRCS and retrievals is further discussed in section 3.3 based on a detailed case study.

In Figure 3d colors provide the SAR incidence angles. Overall, NRCS measurements decrease when the incidence angle increases. Here, this decrease can be observed up to 50 m/s. This result agrees with previous studies (Hwang et al., 2015; Mouche et al., 2017; Zhang et al., 2017). The lack of cases at higher wind speeds prevents any conclusion of the impact of incident angle on NRCS at those speeds. Future studies that make use of larger numbers of cases will be used to address this issue. Results of H14 E GMF from Hwang et al. (2015) and MS1A GMF from Mouche et al. (2017) are reported at 35° incidence angle and are also shown in Figure 3. For comparison, circles surrounded with black lines indicate measurements with incidence angle between 32.5° and 37.5°. Both GMFs agree quite well with the data. At wind speeds above 30 m/s, differences appear, with an overestimation of the NRCS for H14E GMF between 30 and 60 m/s, and an underestimation of the NRCS for MS1A GMF for wind speeds larger than 50 m/s. This is not surprising. The H14E GMF has been derived with wind speed measurements up to 40 m/s, while MS1A GMF definition relies on medium resolution wind speed from SMAP L-Band radiometer up to 50 m/s. Based on these new collected data, we have slightly modified MS1A, to fit the observations up to 70 m/s. We will refer to this new formulation as MS1AHW throughout.

### 3.3. Applications and Discussions

Hereafter, a case study is detailed. The case corresponds to Sentinel-1 measurements acquired over Hurricane Irma, as already presented in the previous sections. Following the approach from Mouche et al. (2017), an algorithm, jointly using copolarized and cross-polarized SAR signals, is applied to estimate ocean surface wind speeds using the MS1AHW GMF. Because concurrent high-resolution surface information does not exist, it is not straightforward to validate the resulting 2-D structure of SAR ocean surface wind speeds. As a first step, we directly compare SAR and SFMR wind speeds along aircraft translation segments. In addition, the overall TC structure derived from the SAR analysis can be compared to the MTCSWA.

Comparisons are also performed with standard operational TC parameters used to describe the hurricane wind structure at Regional Specialized Meteorological Centers. Particular attention is given to the wind radii in the four geographical quadrants NE, NW, SW, and SE provided by the Best-Track analysis. To recall, operational techniques subjectively combine a variety of sensors, including radiometer- and scatterometer-based measurements. In comparison with these other observations of chance, SAR systems provide very high resolution ocean surface information, largely interpreted in terms of wind intensity and structure. SAR wind speed estimates can provide similar structural estimates as aircraft reconnaissance, but observations nearly instantaneously made over the whole inner core area of the TC.

#### 3.3.1. Ocean Surface Wind Structure

When hurricane Irma was observed by Sentinel-1 SAR, on 7 September 2017, it had category 5 intensity according to National Hurricane Center report (Cangialosi et al., 2018) and Best-Track analysis. The wind speed map, as derived from SAR along the SFMR transects, is presented in Figure 4a, where the quality-controlled wind speed measurements from SFMR are also superimposed using the same color code. An overall agreement is clearly obtained between Sentinel-1 and SFMR wind speeds. Two rings of maximum wind speeds are noticeable, around the eye at radii of about 20 km ( $\approx 10$  n.mi) and 30–35 km ( $\approx 15$ –20 n.mi). For the highest values of wind speed, the agreement is even more remarkable (see Figure 4c). This is essentially the result of using the cross-polarization information in the retrieval scheme (Mouche et al., 2017). Note that SAR's high resolution captures the full dynamic range of the wind speed gradient in the eyewall, where wind speeds vary from 10 to 80 m/s in less than 20 km ( $\approx 10$  n.mi). A scatter plot of SAR wind speed estimates versus SFMR ones is presented in Figure 4b. Blue dots indicate data collocated within 3 hr, whereas red dots are collocated data without any criteria time criterion. The overall bias is 1.41 m/s, with a standard deviation of 5.14 m/s, a correlation coefficient of 94%. As expected, such a performance largely degrades as the collocation time difference increases, especially for this intense TC. The time difference becomes more crucial than in other typical studies where wind gradients are smoother or satellite-derived winds are retrieved at coarser resolution. As an illustration of the collocation limitations, a slight misalignment between Sentinel-1 NRCS and SMFR ocean surface wind speeds can be observed in Figure 4c. It is even more obvious during the second transect across the hurricane eye, for which the time difference is about 2

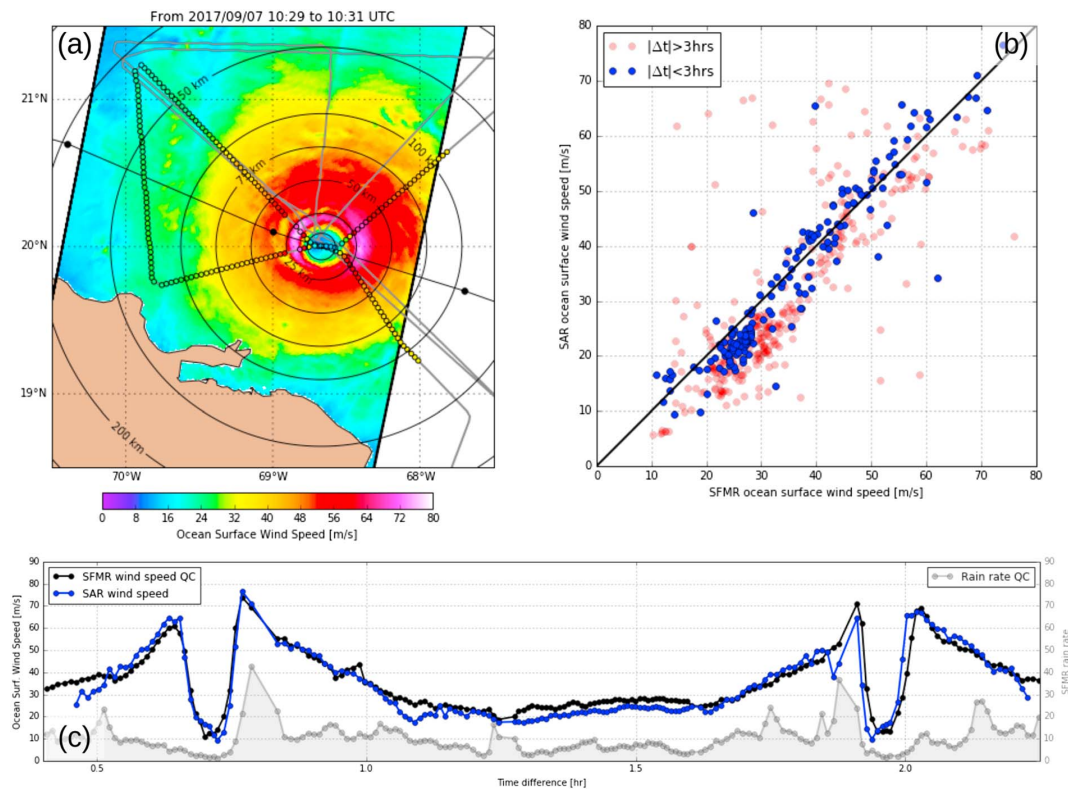
hr. The misalignment is, to first order, likely due to the accuracy of the translation speed, which is based on the 6-hourly and purposely smooth best track. In addition, simply applying a translation, the rotation of the system is also neglected in our collocation method. Nevertheless, these comparisons already provide very encouraging results.

Figure 5 presents the 2-D ocean surface wind speed (a) derived from Sentinel-1 using the dual-polarization approach, (b) MTCSWA, and (c) wind radii-based wind structure provided in the Best-Track. The same color code is used and is defined to indicate the extent of areas corresponding to specific values of 34, 50, and 64 knots that are used by operational services. To ease the visual comparison, the whole Sentinel-1 image is relocated to the storm center location (derived from the MTCSWA product). The original SAR swath limit is also indicated by the black rectangle. The analysis date is 7 August 2017 at 12 UTC for the surface wind analysis and Best-Track products. From Sentinel-1 wind and MTCSWA, we derive the mean ocean surface wind speed profile versus the distance to the TC Eye center, combining data from all geographical quadrants (see Figure 6b), and for each of them (NW, SW, SE, and NE, respectively, in Figures 6c–6f). Wind profiles are indicated in knots to match the units used in operational hurricane centers. We also compute the standard deviation from SAR measurements to highlight the spatial variability at a given radius. The number of SAR measurements used to evaluate this variability as a function of the distance from the TC eye center, is indicated (see y axis on right hand side). As expected, the number of SAR data increases with the distance to the TC eye center. Note that over land or for SAR measurements considered of poor quality, data are removed. Finally, Figures 7b–7f also indicate the 34-, 50-, and 64-knot wind radii and the maximum wind speed from Best-Track analysis before (>) and after (<) SAR acquisition time. In spite of the evident differences between the two approaches used to derive the ocean surface wind fields, the two TC structures are in very good agreement. More precisely, the shape of the mean wind profile, the increase of the wind speed in the TC inner core, the RMW, the maximum wind speed value, but also the decay of the wind speed in the outer part of the TC, all these parameters aligned very consistently for both approaches (see Figure 1b). To further note, averaged wind radii are also very close to Best-Track analysis.

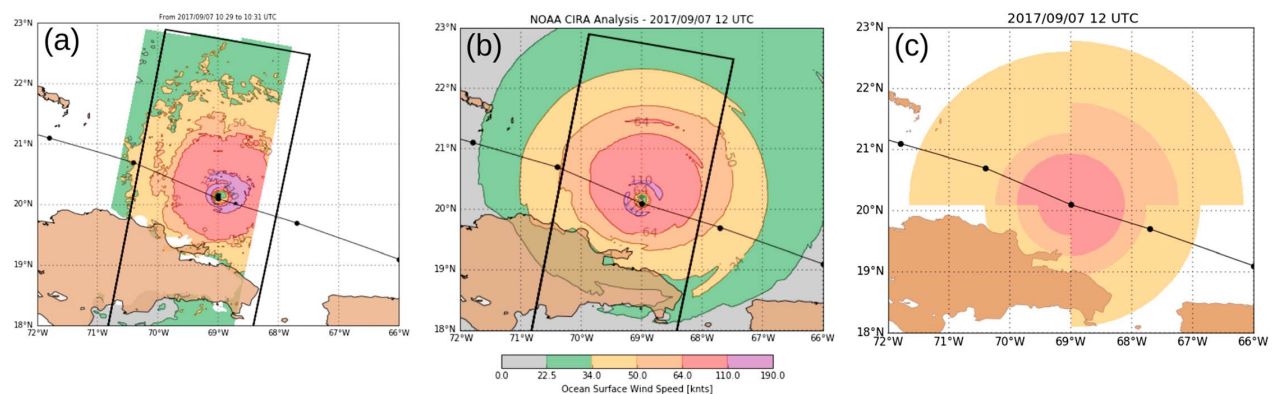
Yet, such an overall excellent agreement is not obtained for all quadrants. As a matter of fact, because of the swath coverage and/or land, SAR cannot always equally cover the four geographical quadrants. In this particular case, we observe that the SE, NE, and SW quadrants are less favorable for reliable estimate of the wind radii. The SE and NE quadrants mainly suffer from swath coverage limitations, whereas the SW quadrant is affected by land for a radius between 100 and 300 km. In addition, the aircraft does not cover the whole region where 34- and 50-knot winds are occurring. Moreover, the combined satellite aircraft analysis is likely weighted to the satellite-based first guess. These may explain small discrepancies for 50-knot wind radii. In spite of this limitation, 34-knot wind radii from MTCSWA and SAR still match very well, as more points are usable far from the TC. As computed, derived estimates provide lower values in comparison to Best-Track analysis. Regarding the 64-knot wind radii, the agreement between the three products is clearly satisfying. For the eyewall, a nice match between the two products is obtained, except in the NW quadrant.

The shape of the wind speed profiles in the inner core derived from MTCSWA and SAR are strongly dependent on the precise hurricane center location. As indicated in Figure 6a, they are not the same for the two products. In MTCSWA product, this location appears to be less certain, to possibly explain this difference. Finally, the most significant discrepancies are obtained for radius values between the maximum of backscattered signal and the 64 knots. This is noteworthy for the maximum wind speed and associated radius derived from the wind profile in the NW and SW quadrants. Sentinel-1 wind profiles presented in Figures 6c and 6d indicate sudden changes, with decrease and increase signals between 20 (about 10 n.mi.) and 30 km (about 15 n.mi.). This is not captured in MTCSWA and is directly related to NRCS variations. The combined satellite aircraft analysis is weighted toward flight-level wind vectors, and flight level measurements do not get these variations. Also, the sampling of the aircraft data is every 30 s or  $\approx 2$  km and the analysis has radial smoothing filters with a half-wavelength of 24 km that can prevent fully capturing such kind of short-scale variations in the wind field. As observed on SAR images around the TC eye, NRCS forms a quite distinct darker ring. It is hypothesized that this feature corresponds to localized heavy rain conditions.

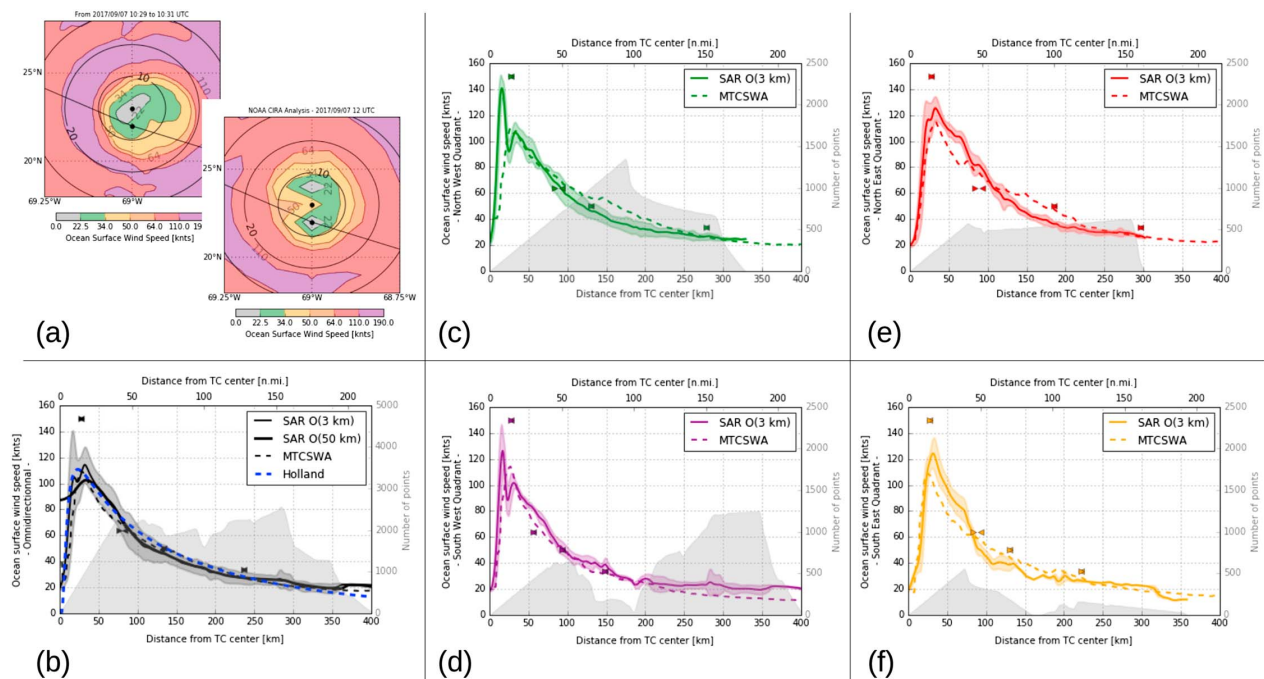
For intense TCs, the region of maximum convective rainfall is typically several kilometers outside the RMWs (Jorgensen, 1984). The high NRCS values, adjacent to the dark ring in the SW and NW quadrants, then explain the retrieved very high wind speeds at 20 km (about 10 n.mi.) from the TC center, observed in Figure 4a and in the NW and SW quadrants in Figures 6c and 6d. This is further discussed in the next section.



**Figure 4.** Ocean surface wind speed for Irma category 5 hurricane on 7 September 2017. (a) Map of SAR-derived wind speed. Gray solid line stands for collocated SFMR track. Colored circles along this track indicate SFMR measurements with time difference lower than 2 hr and 30 min. Color code is the same for SAR and SFMR wind speeds. Concentric circles indicate radius from the cyclone center, and dotted black line stands for the Best-Track trajectory, black circle being hurricane position every 6 hr. (b) Scatter plot between SAR-derived and SFMR-derived ocean surface wind speed. (c) Same as Figure 1 but for 3-km resolution ocean surface wind speed from SFMR and Sentinel-1. SAR = synthetic aperture radar; SFMR = Stepped Frequency Microwave Radiometer.



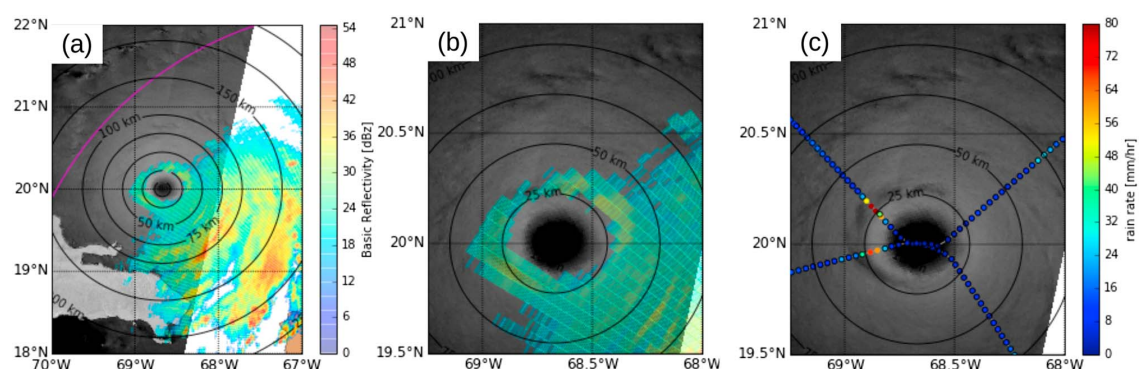
**Figure 5.** Irma two-dimensional wind structure on 7 September 2017 obtained with (a) Sentinel-1A SAR measurements, (b) the combine satellite and aircraft analysis, and (c) given in the Best-Track. Sentinel-1A-based wind structure is from quasi-instantaneous analysis measurements acquired between 10:29:51 UTC and 10:31:27 UTC, while the two others are based on measurements and data collected during 6 hr. The color code has been selected to specifically highlight the 34-, 50-, and 64-knot wind radii. Note that in (a), Sentinel-1A wind structure has been relocated to the exact storm location used for the two other analyses. Black box indicates the original SAR coverage. SAR = synthetic aperture radar.



**Figure 6.** Irma ocean surface wind profiles with SAR observation and multiple satellite platform analysis. (a) Zoom of the wind structure obtained by the two approaches. (b) Omnidirectional ocean surface wind profile as a function of distance from the hurricane center up to 400 km. “>” and “<” indicate 34-, 50- and 64-knot wind radii given by the Best-Track analysis before and after Sentinel-1A acquisition time. Solid black line stands for SAR analysis (see legend for resolution) and dashed black line for multiple satellite platform analysis. Dashed line is the result given by Holland model. (c, d, e, and f) Same as for (b) but for the four geographical quadrants (NW, SW, SE, and NW). SAR = synthetic aperture radar; MTCSWA = Multi-observation Tropical Cyclone Surface Wind analysis.

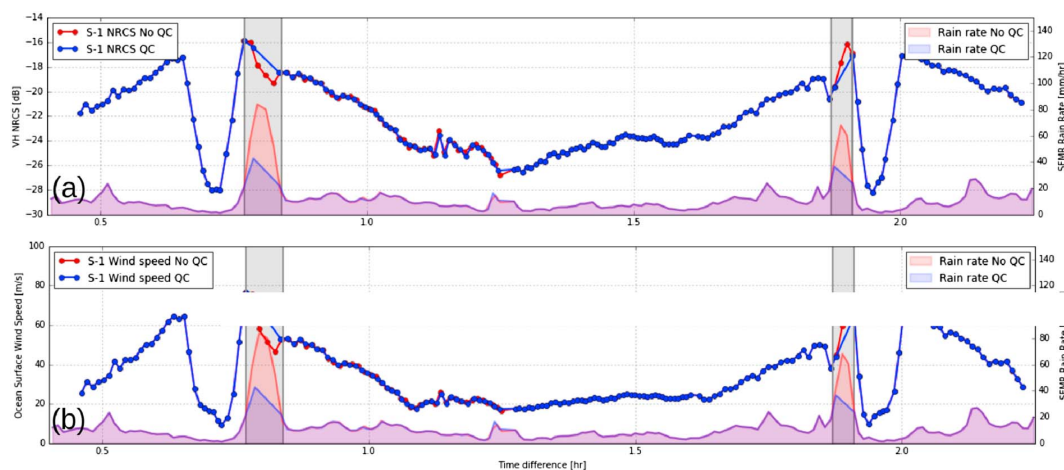
Comparisons to quality controlled SFMR wind speeds tend to endorse these high wind speed estimates, which are not included in the MTCSWA.

Thus, within the inner core region, SAR measurements quite uniquely provide essential means to derive information in the vicinity of the eyewall. As anticipated for a category 5 TC, the SAR-derived maximum wind speed and the one-dimensional wind speed profile (solid black in Figure 1b) are very well approximated



**Figure 7.** Rain during Irma category 5 hurricane on 7 September 2017. (a) SAR cross-polarized normalized radar cross section and base reflectivity measured NEXRAD at the same time. Purple line indicates the limit of NEXRAD coverage. (b) Zoomed on Irma eye. (c) Same as (b) but the color code indicates rain rate measured by SFMR. Circles with (respectively, without) black contours are considered as good (respectively, bad) SFMR data as indicated by the product quality flag. SAR = synthetic aperture radar; NEXRAD = Next-Generation Radar; SFMR = Stepped Frequency Microwave Radiometer.





**Figure 8.** Impact of rain on SAR measurement near the maximum wind speed area. (a) Transect of SAR cross-polarized normalized radar cross section (NRCS) and (b) SAR-derived ocean surface wind speed collocated with SFMR at 3-km resolution. Solid blue (respectively, red) lines are measurements obtained after collocation when a (respectively, no) filtering of the data with respect to the quality flag in SFMR products is applied. SAR = synthetic aperture; SFMR = Stepped Frequency Microwave Radiometer. radar

using a parametric Holland model (dashed blue line in Figure 1b). Combined with the pressure of the outermost closed isobar given by the Best-Track, it yields a central surface pressure of 918 hPa. Remarkably, this estimate of the central surface pressure almost perfectly coincides with the 921-hPa value, as given by the Best-Track and based on aircraft reconnaissance. The analysis of the two-dimensional wind structure further provides some asymmetrical descriptors (e.g., Zhang et al., 2017), including for each quadrant or azimuth direction, the corresponding maximum wind speed and its associated radius, but also the full wind gradient ( $dV/dr$ ) profile. For the present Irma case study, in the direction of the maximum wind speed, a 64-m/s increase in less than 10 km is obtained, to correspond to an absolute vorticity of  $210f$ , where  $f$  is the Coriolis parameter, within the inner core.

### 3.3.2. Rain Impact

As a short-microwave active radar instrument, C-Band SAR signals can be affected by rain. Typically, it has been reported (Tournadre & Quilfen, 2003, 2005) that intense rain reduces the Ku-band QuikSCAT retrieved wind speed by about 5 m/s for rain rate 10 mm/hr and wind speed 30 m/s. Thanks to their dual-frequency capability (Ku- and C-band), altimeters signals have also been jointly processed using adapted algorithms (Quilfen et al., 2006, 2010) to retrieve surface wind/wave information, along with the rain rate in extreme weather events. Compared to Ku-band, C-band data were reported to be far less affected by rain. For a maximum attenuation of 10 dB for the Ku-band instrument, the attenuation at C-band is only 1 dB, in line with Marshall-Palmer relationships (Marshall & Palmer, 1948). For off-nadir measurements, this rain impact on C-band measurements has already been addressed in the literature, especially under moderate wind speeds. In particular, first studies documented rain cell signatures associated with downdraft (Atlas, 1994a, 1994b). Although not necessarily strongly attenuated, C-band SAR images often clearly imprint rain events, including diverse forms such as stratified rain, rain bands, and squall lines (Alpers et al., 2016). But, as backscatter signals are usually only slightly hampered, and not fully attenuated to be easily flagged, these different analyses report very well the challenges to face for an accurate interpretation of rain effects at C-band.

It is likely that raindrops impinging onto the sea surface can affect the backscattered radar power. In the particular case of very high rain rates (above 40 mm/hr), both scattering from hydrometeors at the melting layer and attenuation of the radar signal by rain drops in the atmosphere coexist (see Figure 3(a) from Alpers et al., 2016, with C-Band data acquired over the rain forest). In addition, the importance of these different effects is likely dependent on the background sea state and surface roughness. To note, rain signatures at high resolution often consist as adjacent areas of both enhanced and reduced NRCS values (relative to the ambient rain-free and downdraft-free area). For rain signatures detected from C-Band SAR images acquired over TCs, RADARSAT-1 images in copolarization, together with coastal rain radar and radiometer measurements, have already been discussed (Katsaros et al., 2000) and then further documented (Li et al.,





2013). As concluded, attenuated and enhanced backscattered signals coexist in TC SAR images. But, to date, lack of dedicated investigations preclude definite conclusions. Most studies remain insufficiently documented with few reference concurrent data (including rain, wind, and sea state), and no systematic and robust methods exist to possibly correct effects in SAR images.

In the present study, the NEXRAD rain radar (IJUA) located in Puerto Rico was still operating when Irma approached the Dominican Republic. We can thus use NEXRAD base reflectivity measurements and take advantage of its very high temporal resolution. A base reflectivity map collocated with SAR acquisition is presented in Figure 7a. A zoom in on the TC eye is shown in Figure 7b. In Figure 7a, the purple line is the limit of the NEXRAD radar field of view. NEXRAD radar measurements indicate that more rain occurred on the east side of the storm. In contrast, NEXRAD reflectivity is absent in the western part of the storm, corresponding apparently to a less rainy area. A clear 5-km ring rain is observed around the TC eye at about 30–35 km (about 15–20 n.mi.) from the TC center. For the western part of the TC, the accuracy of the NEXRAD is questionable as the TC is close to its field of view limit. The obtained reflectivity map differs significantly from measurements when the TC was very close to Puerto Rico on 6 September (see NHC report). Collocated SAR measurements and rain rates measured from SFMR along the flight are also available and superimposed to the SAR image intensity in Figure 7c. Measurements that are considered as good (respectively, bad) quality in the SFMR product are indicated by circles with (respectively, without) black contours. SFMR measurements indicate high-quality rain rates up to 40 mm/hr around the TC eye in agreement with NEXRAD high reflectivity. In addition, we also note rain rates around 30 mm/hr in the NE quadrant ( $R \approx 75$  km; about 40 n.mi) and 20 mm/hr in the NW quadrant ( $R \approx 70$  km; about 38 n.mi) corresponding to visible and narrow rain bands in SAR image.

Around the TC eye, where significant rain rates are detected by both NEXRAD and SFMR, Sentinel-1 NRCS signals display a general attenuation. To get a quantitative assessment of this intensity loss, Sentinel-1 NRCS are collocated with SFMR, without and with the SFMR product quality control criteria applied. As observed in Figures 8a and 8b, during the first transect across Irma TC eye ( $\Delta T \approx 0.8$  hr, i.e., 50 min), a signal drop of about 2 dB is found for the NRCS in the area associated with high rain rate (NW area). The comparison obtained during the second transect ( $\Delta T \approx 1.9$  hr, i.e., 1 hr and 55 min) shows an increase. It is noteworthy that, and as already pointed out in the previous section, this apparent inconsistency between these two comparisons is, to a large extent, attributed to the relative inaccuracy of our collocation method. This certainly confirms the inherent limitation of collocating airborne and spaceborne measurements, especially considering very high resolution data. If more carefully and manually corrected, to best align NRCS profile and SFMR wind profile, a 1.5-dB decrease is found for the NRCS in the area associated with high rain rate. It is noteworthy that in these two particular cases the nonqualified rain rates are about 80 and 60 mm/hr, respectively. This suggests very extreme rain rate conditions. Accordingly, close to the TC eyes, where very intense rain occurs, a significant decrease of the NRCS is observed. Note that it applies to both copolarized and cross-polarized signals (not shown). These results agree with the typical location of eyewall rainfall as discussed in Jorgensen (1984).

At this stage, there is not enough evidence to conclude if these overall signal changes are solely due to rain impacts, to wind changes, or to their combined effects. As computed with the proposed wind retrieval algorithm, the resulting localized decrease of the estimated ocean surface wind speed corresponds to about 5–10 m/s near the hurricane maximum wind speed ( $\sim 75$  m/s) area. As shown and detected as the dark circular shape in Figure 7, visible impacts cover almost all the western part of the TC eye, at a distance of about 30–35 km (15–20 n.mi.), and over a relatively limited width of about 5 km. As such, this naturally induces apparent double rings on the wind speed map (see Figure 4a). This also explains the trend of the wind field profiles obtained in NW and SW quadrants at this radius (see Figure 6).

#### 4. Conclusions and Perspectives

Following previously reported studies, the present analysis fully confirms the potential to quantitatively interpret, at very high resolution, the sea surface imprints captured by spaceborne SAR measurements under extreme weather conditions. Specifically, the combined copolarized and cross-polarized signals provide unique high-resolution and nearly instantaneous views of the expected intense and rapidly evolving interactions, occurring near the ocean surface. In these extreme hurricane conditions, heavy precipitation, clouds, and high values of water vapor content combine to significantly mask the ocean contribution to



high-frequency microwave brightness temperatures and Vis-IR measurements. Least affected under extreme environmental conditions, lower-frequency satellite-borne passive microwave instruments (e.g., Reul et al., 2012, 2017; Zabolotskikh et al., 2013), do not operate at sufficient spatial resolution. Quantitatively interpreted, imaging radar systems can thus offer new independent means to complement existing observing systems, with very high resolution O(1 km) description of the TC ocean boundary layer structures. SAR backscatter signals have generally sufficient sensitivity to precisely map surface wind structures in and around the TC eyes, as demonstrated in the present investigation. More specifically for cross-polarization measurements, the radar intensity emerge well above noise and remains significantly sensitive to infer ocean surface information for category 5 hurricanes, with wind speeds reaching up to 75 m/s, as also estimated by SFMR local measurements.

Our conclusions apply to both Radarsat-2 and Sentinel-1 SAR missions. The remarkable consistency between the two missions is obtained after considering proper noise correction. Findings indicate noise correction is an absolute necessity and an essential aspect of processing SAR data. It primarily helps the geophysical interpretation to further pave the way for a combined use of these missions to improve the spatio-temporal coverage for hurricane monitoring. Currently, with three SAR systems operating in copolarization and cross-polarization, and possibly four with Gaofeng-3, a dedicated acquisition strategy could be defined, such as to ensure that a particular TC would be followed throughout its life cycle. Combined with medium resolution observations (e.g., Reul et al., 2017; Zhao et al., 2018), this would certainly provide unprecedented observations to both quantify high-resolution air-sea interactions and describe the organized wind structures during intensification mature and weakening phases.

In this study, the two polarized (copolarized and cross-polarized) signals are combined. This helps to retrieve ocean surface wind speeds at 3-km resolution. In the Irma case study presented here, the resulting wind estimates are found to be in very close agreement with collocated SFMR measurements. The overall bias is about 1.5 m/s, standard deviation around 5 m/s, and correlation coefficient higher than 90%. The large SAR swath also provides the 2-D structure of the wind over hundreds of kilometers. Accordingly, this synoptic capacity is directly used to infer typical TC structure parameters, routinely derived by operational hurricane centers. Again, the SAR-derived parameters are found to be in line with those provided by operational hurricane centers. This certainly suggests the potential for SAR winds to aid existing operational methodologies, already relying on various available data, to infer practical and essential TC parameters: precise wind radii and the degree of storm symmetry. SAR-derived TC information can also be used to help constrain theoretically based parametric wind models and possibly to inform forecasters about inner core processes that could possibly lead to improved intensity forecasts.

The precise analysis of high-resolution data has also limitations. At first, the collocation method, classically used to analyze radar parameters or radar-derived geophysical parameters, certainly needs to be revisited. Indeed, in the case of major hurricanes, such as the Irma case study analyzed in section 3, extreme gradients occur within the eyewall. Using high-resolution SAR images, a small shift in the collocation can then lead to strong misalignment between measurements and huge impacts on direct comparisons. Not only the proposed relationship could certainly benefit for more measurements, but this shall be further constrained using more severe criteria for collocations ( $\Delta T \leq 30\text{min}$ ). As foreseen, necessary future studies shall be devoted to improve the quality control of SAR data, to again augment comparisons with reference measurements, especially including ocean surface wind speed and also surface waves and subsurface currents (Mitarai & McWilliams, 2016).

Moreover, heavy and very localized rainfalls around the eye are also important to consider and possibly monitor using high-resolution SAR measurements. An automated flag for the rain in these situations, to trace and possibly compare radar signatures with high-frequency microwave brightness temperature measurements, would be beneficial, especially to help characterize the rain structures in the inner core and near maximum wind areas. Local gradient analysis developed by Koch (2004) have already been applied to hurricanes (Horstmann et al., 2013; Mouche et al., 2017). Very localized rain events in Irma, as detected using collocated ground-based rain-radar reflectivity measurements, lead to strong variations in the radar signals, translating to 5–10 m/s changes in less than tens of kilometers. Alternating radar intensity variations can then produce double-ring effects on the retrieved wind speed field. Yet, to date, it is not possible to unambiguously estimate and separate the contributions of precipitation from those associated to winds.



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Further investigations, possibly building on available rain radar measurements, and SFMR rain estimates, are beyond the scope of this present paper but will be considered in near future studies. In particular, efforts will be dedicated to evaluate and document the possible differing polarization sensitivity to heavy precipitations. Finally, the combination of SAR measurements with parametric TC wind model has already proven to be useful to overcome SAR limitations such as saturation of the signal in copolarization (Reppucci et al., 2010) but also rain attenuation (Zhang et al., 2017). Although applied to a limited data set with wind speeds lower than or equal to 40 m/s, this method is certainly one way to flag of heavy precipitation.

To conclude, the continuation of routine acquisition of SAR data over TCs and creation of wind speed estimates database would fulfill both research and operational needs. While operational centers are required to estimate the extent of 34-, 50- and 64-knot (17, 25, and 33 m/s, respectively) winds, and the RMW, in reality there are very few observational platforms that can provide accurate assessments of metrics associated with higher wind speeds (e.g., above 50 knots or 25 m/s). And in the absence of aircraft reconnaissance and/or a clearly defined satellite eye, the RMW is typically no more than an educated guess. These platforms include scatterometers, aircraft reconnaissance, microwave sounder-based techniques, and multiplatform techniques, but each has their shortcomings. Scatterometer-based winds begin to saturate above 25 m/s, aircraft reconnaissance is rarely available, microwave sounders suffer from low spatial resolutions, and attenuation and scattering in precipitating scenes and multi-platform techniques rely heavily on statistical proxies for the winds based on infrared imagery. The existing SHOC data set is also a gold mine of information to be further explored for research, providing unique nearly instantaneous views of TC of all intensities and in all the global basins. Research and applied research communities can use these data for developing and testing models and multisensor techniques to better estimate structural features, particularly 64- and 100-knot wind radii, and the RMW.

The location of the RMW, particularly for systems without an eye feature, is particularly difficult to estimate from the current space-based observing platforms. Its location is however very important. RMW is a key parameter in many parametric vortex models (Deppermann, 1947; Holland, 1980; Holland et al., 2010; Willoughby & Rahn, 2004; Wood & White, 2011), and these models are routinely used to derive storm surge and risk models (Lin & Chavas, 2012, and references therein). In addition, the location of the RMW with respect to convective heating has been shown to influence the response of the TC secondary circulation (Schubert & Hack, 1982) and thus is potentially important for the short-term forecasting of intensity change (Carrasco et al., 2014; Wang et al., 2015). Finally, SAR data could be used to study, in detail, the existence and formation of secondary eyewall features and eyewall replacement cycles (see Kossin & Sitkowski, 2009 and references within), which are often monitored using ice scattering signals from microwave imagers. Adopting a clear and sustainable strategy for SAR acquisitions over TC is thus a key point to further secure research and operational activities.

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# Supplement material for Part II.

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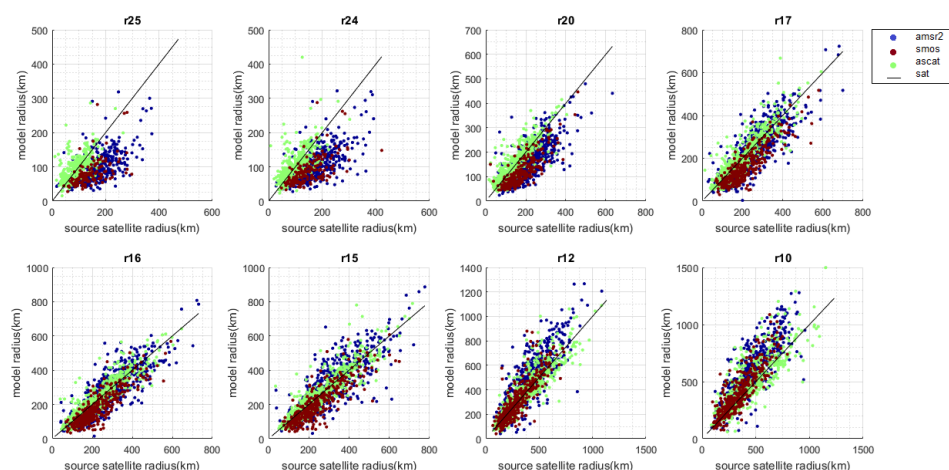


Figure C.1: Comparison between the wind radii from ECMWF and the MW instruments.

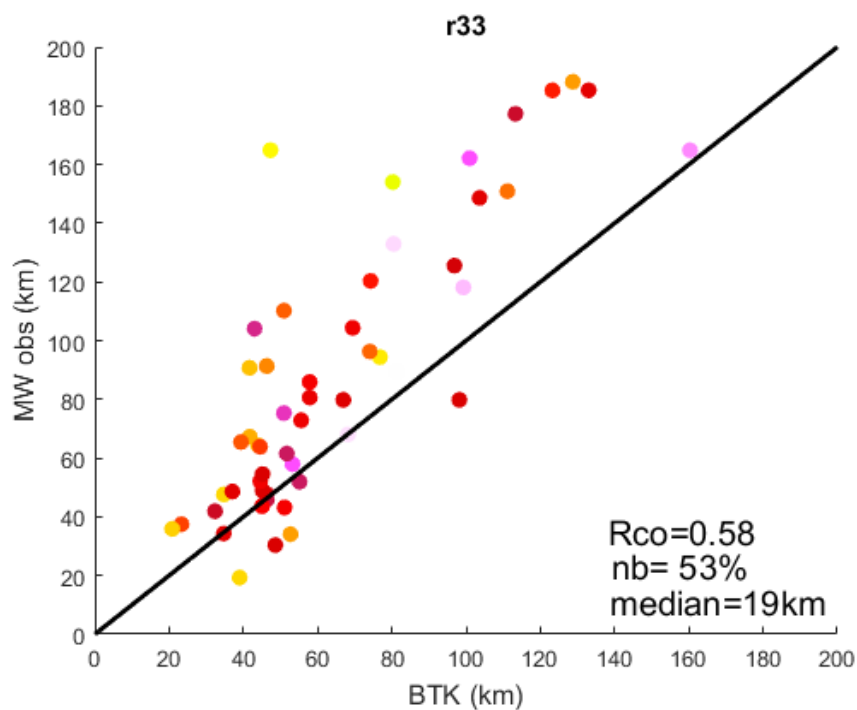


Figure C.2: Comparison between the  $r_{33}$  derived from MW observations and extracted from BTK.

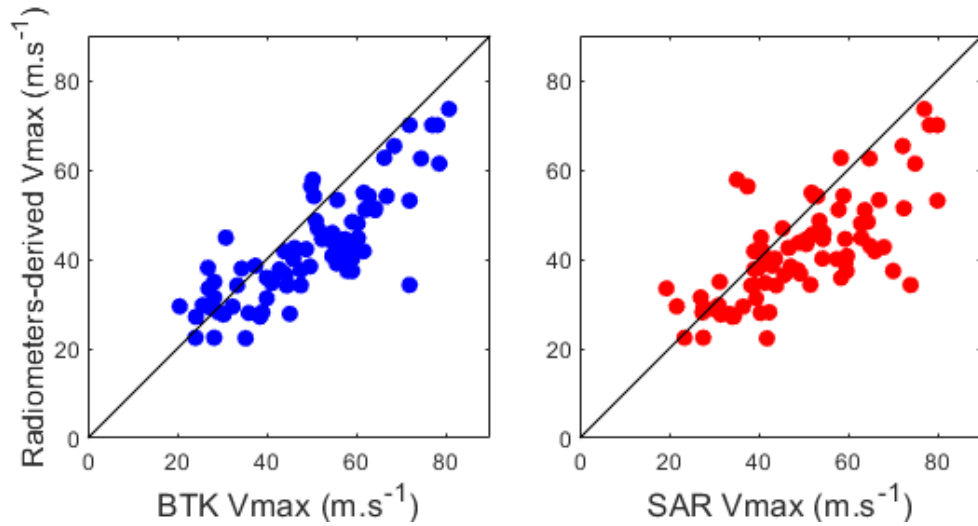


Figure C.3: Comparison between the  $V_{\max}$  derived from MW observations and a) BTK information, and b)  $V_{\max}$  derived from SAR scene.

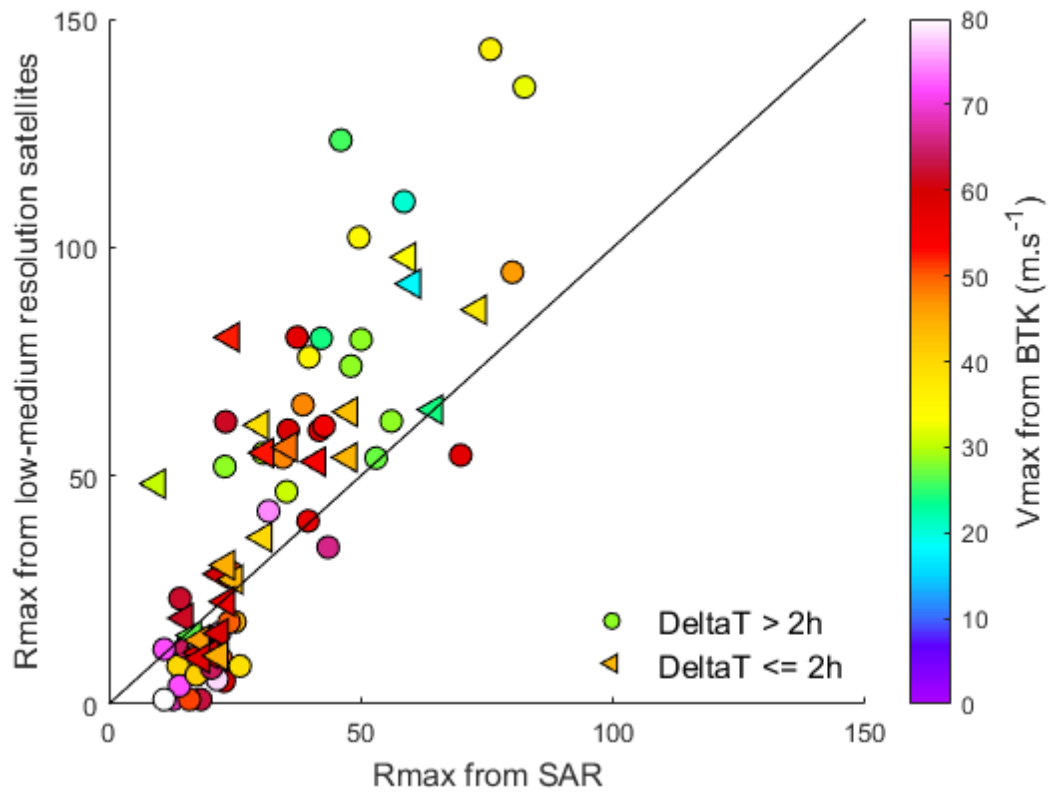


Figure C.4: Comparison between the  $R_{\max}$  derived from MW observations and derived from SAR measurements. Legend indicates the time difference between the acquisitions.

# Simplified scaling laws of the ocean response

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# JGR Oceans

## RESEARCH ARTICLE

10.1029/2018JC014746

This article is a companion to Kudryavtsev et al. (2019), <https://doi.org/10.1029/2018JC014747>.

### Key Points:

- Semiempirical relations for SST and SSH anomalies derived from 20-day satellite measurements of three TCs are suggested
- SSH anomalies provide direct estimates to evaluate impact of the upwelling on amplification of the SST wake
- A marked drag reduction for wind speed higher than 35 m/s is revealed from the measured SSH anomalies

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## A Simplified Model for the Baroclinic and Barotropic Ocean Response to Moving Tropical Cyclones: 1. Satellite Observations

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**Abstract** Changes of sea surface temperature and height, derived from 20-day passive microwave and altimeter measurements for three tropical cyclones (TCs), Jimena, Ignacio and Kilo, during the 2015 Pacific hurricane season, sampling different stages of intensification, wind speeds, radii, Coriolis parameter, translation velocities, and ocean stratification conditions, are reported and analyzed. As triggered along the path of moving TCs, very large interior ocean displacements can occur to leave prominent sea surface height (SSH) anomalies in the TC wake. Resulting surface depressions can reach 0.3–0.5 m, depending upon size, translation speed, and ocean stratification conditions. These signatures can be quite persistent, that is, more than few weeks, to possibly be intercepted with satellite altimeters. To interpret sea surface temperature (SST) and SSH anomalies, a semiempirical framework is adopted, based on the heat and momentum conservations laws for the upper wind driven mixed layer. As interpreted, SSH anomalies provide direct estimates to evaluate the upwelling impact, that is, the upwelling amplification on the SST wake. For the reported cases, the influence of the upwelling is found rather moderate, of order 10–40%. More promising, the proposed bottom-up approach can help document the resulting wind forcing and practical drag coefficient under extreme TC conditions. As found for these three TCs, a marked drag reduction for wind speed higher than 35 m/s is inferred to ensure consistency with the measured SSH and SST anomalies.

### 1. Introduction

Thanks to multiple satellite remote sensing observations and improved available in situ measurements, the upper ocean responses to moving tropical cyclones (TCs) are today often very well captured and monitored (e.g., Shay, 2010, and references therein). TCs generate a variety of responses: asymmetrical sea states (e.g., Hwang & Fan, 2017; Kudryavtsev et al., 2015; Wright et al., 2001), internal motions at superinertial and inertial frequencies (Geisler, 1970; Gill, 1984; Longuet-Higgins, 1965; Meroni et al., 2017; Price, 1983), geostrophically balanced motions, and turbulence, all contributing to irreversible vertical mixing through the combination of surface stirring, shear at the base of the mixed layer, and convective cooling. As results of all these interactions and adjustments, upper ocean responses to extreme wind forcing by moving TCs still remain difficult to fully elucidate.

From a satellite perspective, distinctive features of these upper ocean transient and localized impacts attract considerable attentions. Quite systematically, TCs passages exhibit measurable persistent signatures in TC wakes, for example, changes of the sea surface temperature (SST; e.g., Cornillon et al., 1987), ocean color (e.g., Babin et al., 2004; Huang & Oey, 2015), and/or salinity (Grodsky et al., 2012; Pudov & Petrichenko, 2000). Vigorous hurricane-induced mixing and intense upwelling act to entrain cool thermocline water into the upper ocean mixed layer, stirring warm surface waters with colder waters from below. Consequently, the wake produced by the passage of a TC is generally characterized by a surface cold anomaly, accompanied with nutrient blooms and a subsurface warm anomaly (Jansen et al., 2010). As also reported, passage over freshwater plumes can cause strengthening of hurricanes due to localized enhanced SST, and minimization of the cold-water intrusion from below, due to the presence of a barrier layer effect (e.g., Balaguru et al., 2012; Reul et al., 2014).

In the Northern Hemisphere, a more or less pronounced rightward bias also occurs, consequent to TC forward motion, resulting in resonant couplings between surface winds and clockwise inertial currents,

accelerated (respectively decelerated) on the right-side (respectively left side). Mixed layer stirring and entrainment from below the thermocline are thus amplified (Huang & Oey, 2015; Price, 1981; Skillingstad et al., 2000). Overall, TC-induced ocean cooling has then been reported to be more pronounced when the storm is intense, the mixed layer shallow with a sharp thermocline, and is slowly moving (e.g., D'Asaro et al., 2014; Lin et al., 2009; Mei et al., 2012). For the later conditions, the resulting sea state is more symmetrical, as trapped fetch effect mostly occurs for a fast-moving TC (Kudryavtsev et al., 2015). For slow-moving TCs, the circular wind pattern will further trigger Ekman pumping, driving surface water away from the storm center, with associated isopycnal uplifts that can typically reach 50–100 m (Babin et al., 2004; Walker et al., 2005).

Consequently, as mostly governed by these isopycnal displacements, TCs can also leave prominent sea surface height (SSH) anomalies in their wakes (Ginis, 2002, Figure 3). Resulting surface depressions near the storm centers, as trenches behind TCs, can reach 0.3–0.5 m, depending upon size, translation speed, and ocean stratification conditions. Following geostrophic adjustment, such a signature (encompassing barotropic and baroclinic effects) can be quite persistent, that is, more than few weeks. Accordingly, building on the actual satellite altimeter constellation (presently up to six satellites are available), satellite SSH measurements may likely cross such trenches, to provide quantitative sea surface trough amplitudes in the wakes of TCs.

In the present paper, our motivation is to essentially dwell on this overlooked capability and to report on the potential to combine SST and SSH satellite observations to analyze the ocean responses to TCs. Hereafter, this capability is demonstrated using 20-day satellite observations intercepting three major TC events, namely Jimena, Ignacio, and Kilo, developing and traveling over the central and eastern parts of the Pacific Ocean, simultaneously reaching category 4 on 29–30 August 2015 (Reul et al., 2017; Figure 7). As gathered, multisensor observations provide a comprehensive data set on local changes of SST (derived from passive microwave measurements) and SSH (altimeter measurements) along the TCs paths, sampling different stages of intensification, wind speeds, radii, Coriolis parameter, translation velocities, and ocean interior conditions. Description of the data is given in section 2. Analysis of the SST and SSH anomalies and their dependencies on TC and environment parameters, and parameterizations are discussed in section 3. In section 4, the potential to infer the hurricane-wind forcing is presented, and in section 5, the influence of the upwelling is discussed. A summary of the results is given in section 6.

## 2. Data

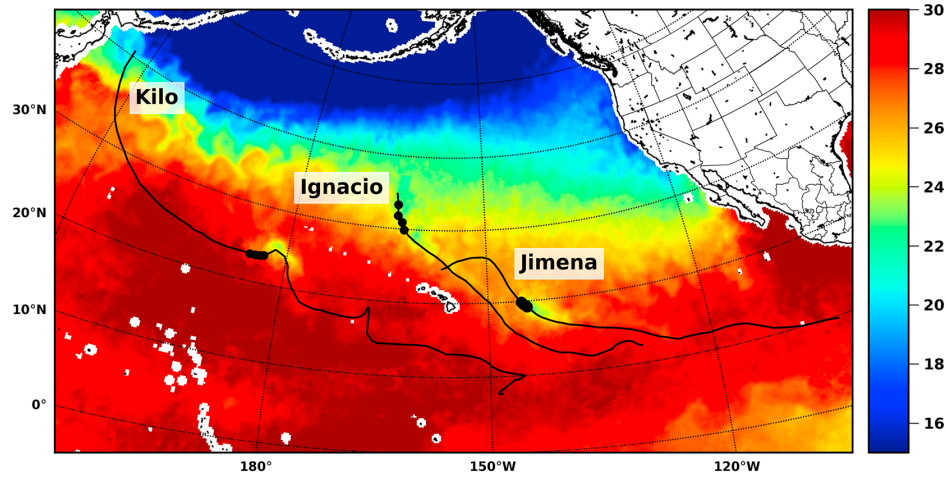
The 2015 Pacific hurricane season was the second most active one on record: 31 tropical depressions, of which 26 became named storms, and 11 major hurricanes. As already reported (Reul et al., 2017), three of these major hurricanes, namely Jimena, Ignacio, and Kilo, developed in between 20 August and 10 September. Hurricane/Typhoon Kilo became one of the longest-lived tropical cyclones on record, with a total lifespan of 22 days.

### 2.1. SST Wakes

Combining mult-sensor satellite measurements, daily averaged SST, at 25-km resolution ([http://data.remss.com/sst/daily\\_v04.0/mw/2015/](http://data.remss.com/sst/daily_v04.0/mw/2015/)), are produced using optimal interpolation. For 4 September 2015, Figure 1 illustrates such a Microwave Optimally Interpolated Sea Surface Temperatures (MW OISST) field, where black dots indicate TCs positions on this day.

While well expressed for this day, significant spatial and temporal variability of the background SST field can prevent the proper identification of these SST wakes. Commonly, differences between the SST field after and before TC passages are usually performed. The SST field averaged over 10–15 days before TC passage is then considered as the background SST field (e.g., Reul et al., 2014; Vincent, Lengaigne, Madec, et al., 2012). Yet for long time scales of order of weeks, as for the case of long-living TCs, it can be necessary to reduce as much as possible the contribution of the SST field variability. This helps focus on local features of the wake during the forced stage of its formation. Figure 2 shows fragments of the daily MW OISST data, corresponding to 3 days, 5 to 7 September 2015, around TC Kilo. For each day, its position is indicated by black dots. As revealed and anticipated, a rapid evolution of the SST wake is clearly taking place on daily time scales.





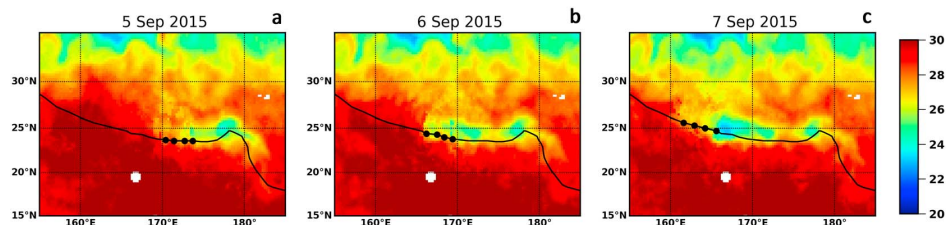
**Figure 1.** Daily averaged sea surface temperature field from Microwave Optimally Interpolated Sea Surface Temperature data for 4 September 2015. The black lines are the tropical cyclones TCs' tracks according to the Best-Track Data. Black dots indicate positions of the tropical cyclone on 4 September 2015.

To best evaluate the SST anomaly,  $\Delta\theta_s$ , generated by a TC on a given day,  $t$ , we thus consider the difference between the SST fields, corresponding to the day after cyclone passage,  $\theta_s(t+1)$ , and 1 day before,  $\theta_s(t-1)$ :

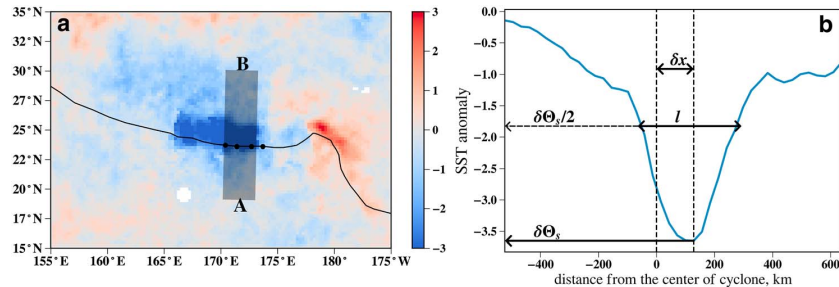
$$\Delta\theta_s(t) = \theta_s(t+1) - \theta_s(t-1). \quad (1)$$

As an example, Figure 3a illustrates a field of SST anomalies (on day  $t$  corresponding to 6 September) using equation (1) and the difference between the SST fields on 7 and 5 September, shown in Figure 2. We then consider transects of the SST anomalies in the direction perpendicular to the hurricane track. Transects are then averaged along the track direction. The averaging window approximately corresponds to the distance traveled by TC during a day, Figure 3a. The averaged transect of the SST anomaly is shown in Figure 3b. The following parameters are then defined: magnitude of the SST anomaly,  $\delta\theta_s$ , corresponding to the minimal value of the  $\Delta\theta_s(t)$  transect; right-biased offset of the anomaly,  $\delta x$ , defined as the distance between  $\delta\theta_s$  location and TC track positions; and the anomaly width,  $l$ , defined as the distance between the isotherms equal to half  $\delta\theta_s$ , see Figure 3b for more explanation. Since SST anomalies are sufficiently averaged, we introduced the minimum value,  $\delta\theta_s^{\min}$ , as an extra characteristic of the SST wake. It is defined as the minimal value of the SST anomaly  $\Delta\theta_s(t)$  found inside the averaged transect region and  $\delta\theta_s^{\min} \leq \delta\theta_s$ .

As illustrated Figure 4, these resulting SST wake parameters can significantly vary over the course of the TCs, corresponding to a wide range of changes in TC characteristics (i.e., maximum wind speed, translation velocity and size) and environmental conditions (i.e., ocean stratification and the Coriolis parameter). To note, the minimal value of the SST anomalies,  $\delta\theta_s^{\min}$ , well follows variations of the minimum of the (averaged) SST anomalies,  $\delta\theta_s$ , but displays some local offsets (up to about  $2^\circ$ ).



**Figure 2.** Daily averaged sea surface temperature fields from Microwave Optimally Interpolated Sea Surface Temperature data for (a) 5 September, (b) 6 September, and (c) 7 September 2015. Black lines is the tropical cyclone Kilo track and black dots indicate its positions within the given day.

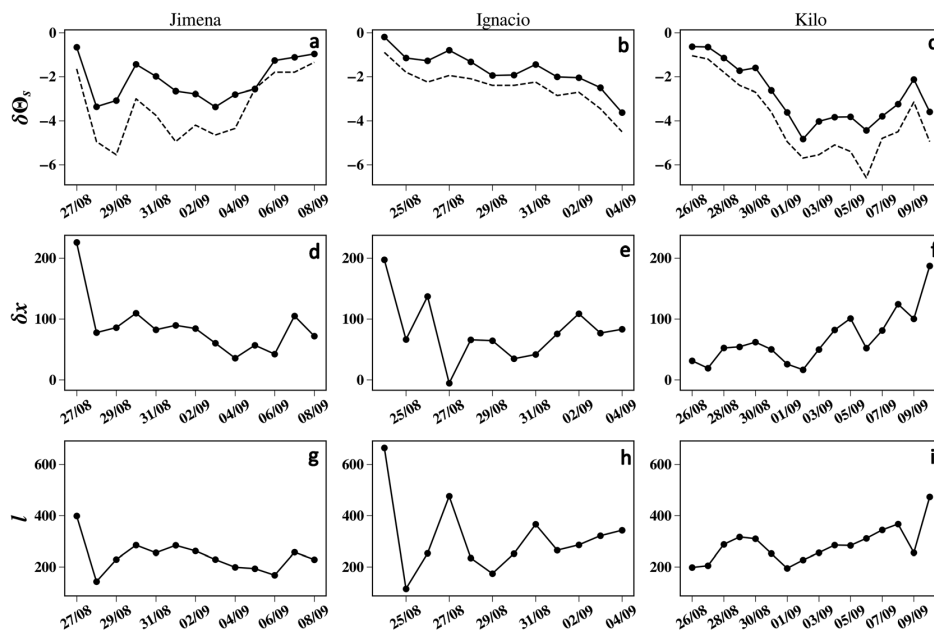


**Figure 3.** (a) Sea surface temperature (SST) anomaly field on 6 September and (b) profile of SST anomaly along cross section AB. The vertical dashed line crossing 0 on  $x$  axis corresponds to the position of the tropical cyclone track. Distance between two dashed lines indicates offset of the SST anomaly from tropical cyclone track;  $l$  is the width of the SST anomaly defined as distance between isotherms equal to half the SST anomaly.

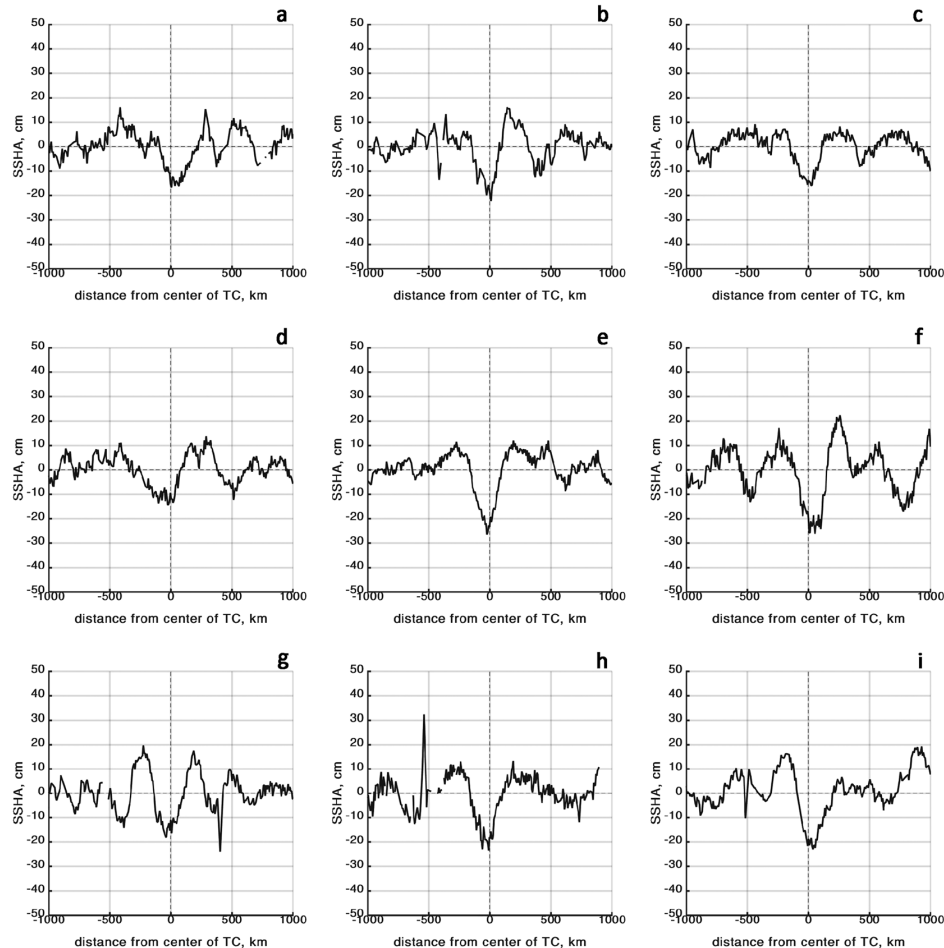
## 2.2. SSH Wakes

The SSH anomalies are investigated using multisatellite altimeter measurements. For the study, altimeter data from three missions are used: Jason-2, CryoSat-2, and SARAL/AltiKa. The Jason-2 Geophysical Data Records (GDRs) are provided by the National Oceanographic Data Center (NODC; <http://www.nodc.noaa.gov/>). The CryoSat-2 data are available at the European Space Agency (ESA) Earth Online website (<https://earth.esa.int/>). The SARAL/AltiKa GDRs are distributed through the Archiving, Validation and Interpretation of Satellite Oceanographic data (AVISO) portal (<http://www.aviso.altimetry.fr/>).

Operating at high frequencies (i.e., Ku- and Ka-band), radar altimeter measurements are sensitive to precipitations, and data affected by TC rain bands can often be lost. Yet as SSH anomalies are expected to be long-living surface features, possibly persisting for several weeks, the time difference between altimeter



**Figure 4.** Time-evolution using coordinate system moving with tropical cyclone (TC) of (a–c) the sea surface temperature (SST) anomaly magnitudes (solid line corresponds to averaged SST anomaly and dashed line to the minimal value inside the averaging window), (d–f) offset of the SST anomaly from TC's track, and (g–i) width of the anomaly for TCs Jimena (a, d, and g), Ignacio (b, e, and h), and Kilo (c, f, and i).

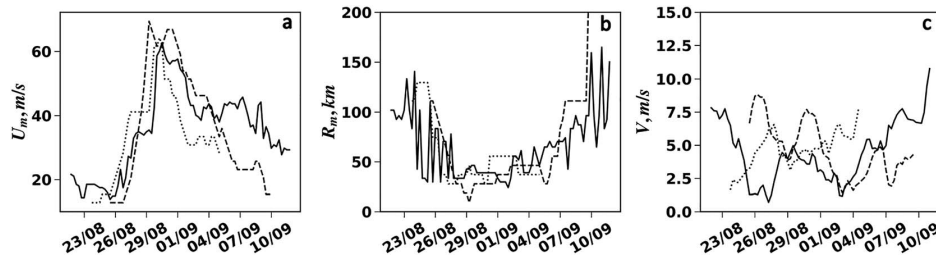


**Figure 5.** Examples of sea surface height (SSH) anomalies obtained from the altimeter passes crossing (a, d, and g) tropical cyclone (TC) Ignacio, (b, e, and h) TC Jimena, and (c, f, and i) TC Kilo. Vertical dashed lines crossing 0 on the x axis indicate position of TC track taken from Best Track data. Magnitude of SSH anomaly is further defined as minimum of the surface height around TC track.

measurements, also controlled by each altimeter cycle, can be relaxed. In total, 53 altimeter passes crossing TC tracks (after a TC passage) have been selected. As selected, SSH anomalies correspond to measurements for which the time differences between altimeter crossing a TC track and the time of its passage did not exceed 6 days. Figure 5 shows examples of SSH anomalies with well expressed depressions around the TC track. Apparent oscillations apart from the TC are also revealed, to possibly be interpreted as TC baroclinic wake signatures and/or residual SSH anomalies left by other preceding TCs, previously traveling in the same area before. The maximum magnitudes of these SSH anomalies are then collected to be compared with TCs characteristics and environmental conditions.

### 2.3. Best Track Data

TCs trajectories and main characteristics are taken from the Best Track (BT) data (<http://ftp.nhc.noaa.gov/atcf/>), that is, maximum wind speed,  $u_m$ , and its radius,  $R_m$ , with translation velocity,  $V$ , derived from the TCs coordinates, Figure 6. For instance, Kilo peaked in intensity on 30 August as a category 4 hurricane with 120-kt winds. After peaking in intensity, Kilo fluctuated between categories 3 and 4, as it slowly moved northwest, before weakening below major hurricane status as conditions became less favorable. More



**Figure 6.** Along-track evolution of (a) maximum wind speed, (b) radius of maximum wind, and (c) translation velocity derived from best track data for Hurricanes Kilo (solid line), Jimena (dashed line), and Ignacio (dotted line).

detailed comparisons between BT estimates and low-frequency passive radiometer satellite measurements can be found in Reul et al. (2017).

Among key parameters to determine impacts on the upper ocean layers, the dimensionless TC translation velocity is often considered

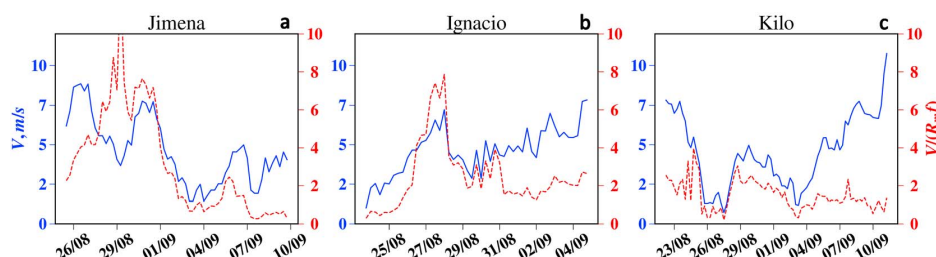
$$Ro = \frac{V}{R_m f} \quad (2)$$

also termed as the TC Rossby number,  $f$  is the Coriolis parameter. This parameter divides TCs in two groups: “slow” if  $Ro < 1$  and “fast” if  $Ro > 1$ . Reported observations revealed that SST anomalies generated by slow TCs are systematically larger than those generated by fast ones (Mei et al., 2012). As already mentioned above, contrarily to slow TCs, fast TCs generate right (left)-biased SST wakes in the Northern (Southern) Hemisphere (Cornillon et al., 1987).

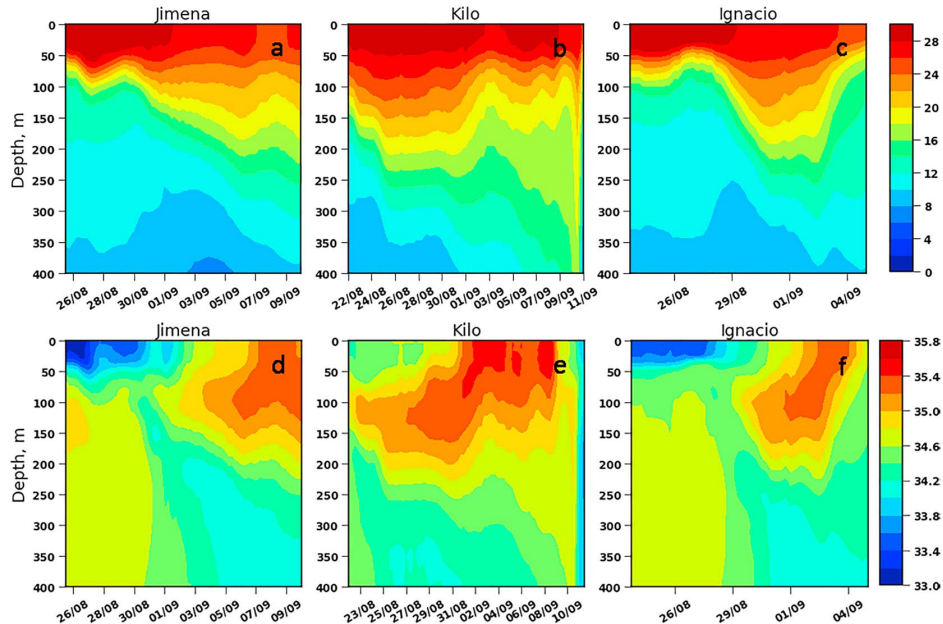
For each TC case, evolutions of the translation velocities and Rossby numbers are given in Figure 7. As obtained,  $Ro$  widely varies, from almost 0 to about 10. During most of their life span, the three TCs can be classified as fast ones, and only rarely as slow TCs. Changes of TC size can impact  $Ro$ , and in some case, for example, for TC Jimena around 29 August, the marked decrease of the TC radius leads to  $Ro$  behavior opposite to the translation velocity trend. Note, these TCs were all generated in the equatorial region to then travel northward. As such, the Coriolis parameter varies by a factor 3, also contributing to strongly modulate  $Ro$ .

#### 2.4. Ocean Interior

As anticipated and often reported (D’Asaro et al., 2014; Lin et al., 2009; Mei et al., 2012), resulting SST anomalies also significantly depend on the local upper ocean stratification under the moving TC. To specify a background stratification, that is, before a TC passage, ocean interior conditions are evaluated from the monthly averaged temperature and salinity data provided by the World Ocean Atlas 2013 version 2 (<https://www.nodc.noaa.gov/OC5/woa13/>). As a comparative ground-truth data set, ISAS-15 gridded ( $1/2^\circ$  resolution) monthly fields and climatology of temperature and salinity is also employed, which are



**Figure 7.** Along-track evolution of (red) the tropical cyclone-Rossby number,  $Ro = V/(R_m f)$ , and (blue) the translation velocity,  $V$ , calculated from the best track data for (a) Jimena, (b) Ignacio, and (c) Kilo.



**Figure 8.** Vertical cross sections of the (a–c) temperature and (d–f) salinity of the upper 400 m of the ocean along the tropical cyclones tracks: (a and d) Jimena, (b and e) Kilo, and (c and f) Ignacio. The data are taken from ISAS-15 gridded ( $1/2^\circ$  resolution) monthly fields.

constructed from ARGO profilers on 152 levels ranging from 0- to 2,000-m depth and entirely based on these in situ measurements (Kolodziejczyk et al., 2017). Resulting vertical cross sections of the ocean temperature and salinity along the TCs tracks are shown in Figure 8. As found, temperature stratification significantly varies along the TC tracks. For instance, at their initial stage of developments, Jimena and Ignacio traveled over an area with shallow thermocline, to then pursue over an ocean area exhibiting deep and smeared thermocline conditions. As also revealed, the sharp temperature drop at the end of Kilo's track, is likely related to inflow of cold freshened Arctic waters through the Bering Strait, reaching the eastern coast of Japan.

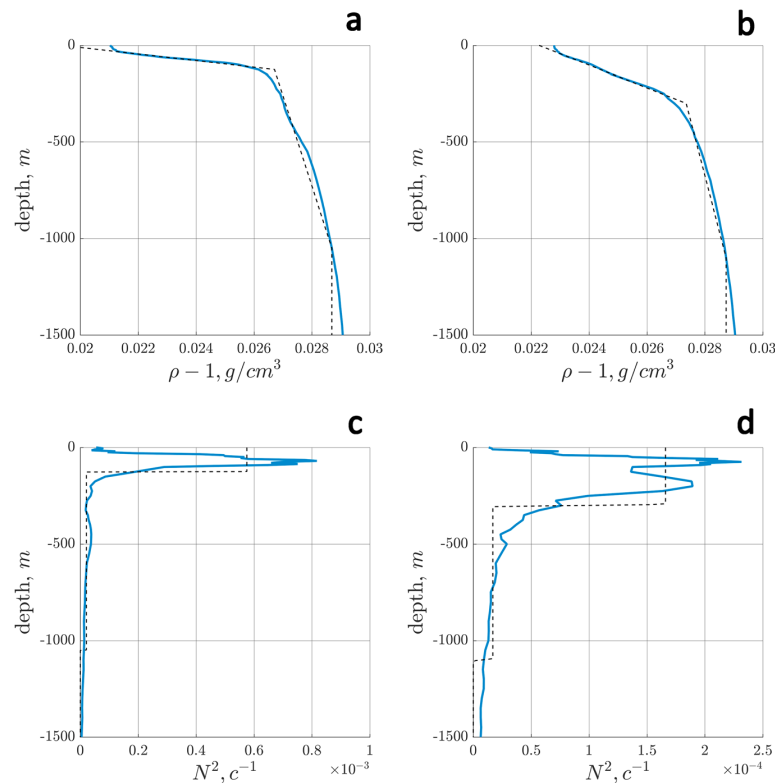
The density,  $\rho$ , is evaluated using a linear approximation for the UNESCO sea water state equation in the form

$$\rho = \rho_0[1 - \alpha(\theta - \theta_0) + \beta(s - s_0)] \quad (3)$$

where  $\alpha = 2.7 \times 10^{-4} \text{ } 1/^\circ\text{C}$  and  $\beta = 7.6 \times 10^{-4} \text{ } 1/\text{‰}$  are thermal and salinity expansion coefficients correspondingly,  $\theta$  is the water temperature in  $^\circ\text{C}$ ,  $s$  is salinity in per mille, and  $\rho_0 = 1,025 \text{ kg/m}^3$  is the reference density with  $\theta_0 = 22^\circ\text{C}$  and  $s_0 = 36\text{‰}$ . From density, two stratification parameters are determined: Brunt-Väisälä frequency,  $N$ , of the upper ocean below the mixed layer, and phase velocity of long gravity internal waves contributing to the baroclinic response.

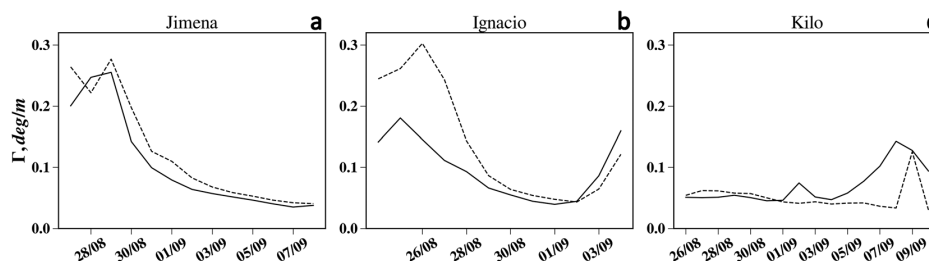
To infer these parameters, the local vertical ocean stratification (at each point of TC track) is approximated with a three-layer stratification model, adjusting the seasonal and the main pycnoclines with linear approximations of density over the depth, and the abyssal part with constant density. The fit parameters (Brunt-Väisälä frequency,  $N$ , in the seasonal pycnocline and its depth,  $d$ ) are derived using a least squares method. It ensures adjustment of the model density profile in the seasonal and main pycnocline to the observed profile in the layer  $h < z < D$ , where  $h$  is the mixed layer bottom (identified as the maximum of the second derivative of the density profile), and  $D$  is the depth of the lower boundary of the main pycnocline, defined as the depth of the layer containing 95% of the total observed density drop. Examples of adjustments for “shallow” and “deep” pycnoclines are given in Figure 9. As a measure of the ocean stratification, Figure 10, the vertical gradient of the water temperature in the seasonal pycnocline, that is, in the layer  $h < z < d$  prescribed by the three-layer model, is further estimated. It also demonstrates quite strong along-track variability.



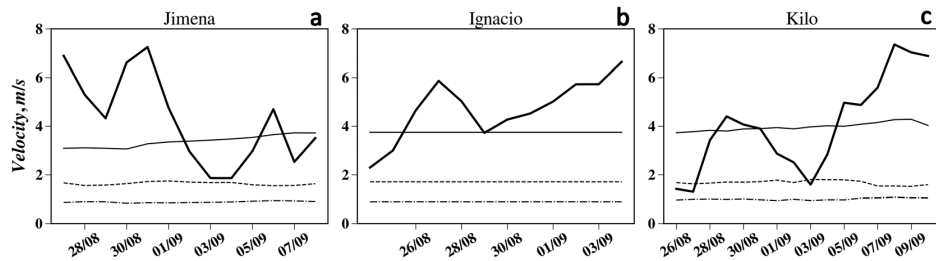


**Figure 9.** Solid lines are (a and b) density and (c and d) squared Brunt-Väisälä frequency profiles for the (a and c) shallow and (b and d) deep pycnoclines. Dashed lines in the upper row indicate fit of the density profiles by three-layer model, and corresponding three-layer models for  $N^2$  are shown in plots (c) and (d).

Finally, the phase velocities,  $C_i$ , of inertial-gravity internal waves (IW) corresponding to the first three modes defined within the frame of the three-layer stratification model along the TC tracks are given, Figure 11. IW phase velocities,  $C_i$ , for each of the modes do not significantly vary. As a transient and intense extreme event, a TC will trigger baroclinic response when its translation velocity exceeds the IW phase speed,  $U > C_i$ . From Figure 11, it may appear that TC Ignacio, during most of its life span, is relatively fast and capable to generate baroclinic responses formed by the composition of the first three IW modes. At variance, TCs Kilo and Jimena are slower during about half of their life spans, with translation velocities below the first IW mode, and sometimes even below the second mode.



**Figure 10.** Along-track evolution of the local vertical gradient of the ocean temperature in the seasonal thermocline calculated from (solid lines) WORLD OCEAN ATLAS and (dashed lines) ISAS-15 gridded ( $1/2^\circ$  resolution) monthly fields for (a) Jimena, (b) Ignacio, and (c) Kilo.



**Figure 11.** The along-track evolution of (thick solid lines) tropical cyclone translation velocity and internal waves phase velocity for three first modes (thin solid, dashed, and dash-dotted lines correspondingly) for tropical cyclones (a) Jimena, (b) Ignacio, and (c) Kilo.

### 3. Analysis and Scaling of SST and SSH Anomalies: Dependencies on TC and Environment Parameters

#### 3.1. SST Anomalies

At first, estimated SST anomalies,  $\delta\theta_s$ , can be compared with concomitant maximum wind speed, translation velocity, and/or temperature gradient in the seasonal thermocline. As found (not shown), no remarkable correlation is emerging from these comparisons. An overall wind trend, that is, the higher the winds, the larger the SST anomalies, could still be revealed. This is in qualitative agreement with previously reported observations (e.g., D'Asaro et al., 2014; Lin et al., 2017). Such a weak correlation with these individual variables clearly invites to consider more elaborated combinations of parameters, to best interpret SST anomalies.

Vincent, Lengaigne, Madec, et al., 2012, Vincent, Lengaigne, Vialard, et al., 2012) introduced two variables, the Wind Power index (WPI) and the Cooling Inhibition index (CI). The WPI characterizes the strength of the TC forcing on the upper ocean and is based on the power dissipated by friction at the air-sea interface (Emanuel, 2005). It combines in a single measure, the maximum wind, TC size, and translation speed:  $WPI \propto u_m(R_m/V)^{1/3}$ . WPI is a proxy to estimate the available amount of kinetic energy contributing to mixing in the upper ocean and thermocline erosion, leading to surface cooling. As cooling also depends on the ocean background stratification, that is, type of the thermocline—shallow/sharp or deep/broad, CI is introduced to characterize the conversion of kinetic energy to potential energy by vertical mixing. This helps identify the possible inhibition of mixed layer deepening in presence of a strongly stratified background. From their numerical experiments, Vincent, Lengaigne, Vialard, et al. (2012) demonstrated that TC-induced SST anomalies are largely controlled by WPI and CI, with CI affecting the cooling amplitude by up to an order of magnitude.

For the present analysis, we advocate a more straightforward scaling, based on classical upper mixed layer concept. Similar to Vincent, Lengaigne, Vialard, et al. (2012), maximum wind speed,  $u_m$ , its radius,  $R_m$ , and translation velocity,  $V$ , are considered as TC governing parameters, defining the upper ocean forcing. As also expected, governing environmental parameters are, prestorm values of the Brunt-Väisälä frequency,  $N$ , temperature gradient,  $\Gamma$ , in the seasonal pycnocline, and Coriolis parameter,  $f$ . Considering the ocean temperature and the density to linearly vary with the depth in the seasonal pycnocline, the SST,  $\theta_s$ , satisfying a 1-D heat conservation equation reads

$$\theta_s = \theta_s^0 - 1/2\Gamma(h_m + \delta h) \quad (4)$$

where  $\theta_s^0$  is the conventional “calm condition” SST of the ocean,  $h_m$  is the mixed layer depth, and  $\delta h$  is the displacement of the water masses at  $z = h_m$  due to the TC-induced upwelling effect. In equation (4), the solar heating and other components of the heat balance are ignored, and equation (4) is thus best valid during the initial stage (forced stage) of the transient TC impact on the upper ocean. Under enhanced wind forcing, equation (4) suggests that cooling results through mixed layer deepening, caused by an intensification of turbulent mixing (this corresponds to 1-D ocean model), and through the upwelling associated to the resulting vorticity of the surface stresses. Under a three-layer approximation of the ocean stratification (see Figure 9), the vertical velocity in the upper seasonal thermocline layer, with constant  $N$ , reads:

$w(z) \propto \sin(Nz/c)$  (see, e.g., Appendix A in Kudryavtsev et al., 2019). If the TC-induced displacement of the main pycnocline,  $\delta D$ , is chosen as a leading parameter, then  $\delta h$  is evaluated as  $\delta h = (h_m N_1/c) \delta D$ , where we simplify,  $\sin(h_m N_1/c) \approx h_m N_1/c$ . Then equation (4) can be rewritten as

$$\delta \theta_s = -\frac{N^2}{2\alpha g} h_m [1 + (N_1/c) \delta D] \quad (5)$$

where  $\delta \theta_s$  is the SST anomaly, and we assumed that the vertical gradient of the ocean temperature largely controls the amplitude of Brunt-Väisälä frequency, that is,  $\Gamma \approx N^2/(\alpha g)$ .

Following the concept of a critical regime for the mixed layer deepening (e.g., Price et al., 1986), the mixed layer may be postulated to evolve as to maintain the bulk Richardson number at a constant (critical) value:

$$Ri_{cr} = (\Delta \rho / \rho) g h_m / v_m^2 \quad (6)$$

where  $\Delta \rho / \rho$  is the relative density drop over the mixed layer base, and  $v_m$  is the magnitude of the wind drift current velocity in the mixed layer. The parameterization of  $h_m$  by equation (6) explains the possible offset of the SST anomalies from the TC track, since inertial drift currents are amplified to the right from the track where wind stress rotation coincides with the rotation of inertial currents (Price et al., 1986). Introducing the volume transport by the wind-driven current as  $M_w = v_m h_m$ , and estimating  $M_w$  from the momentum conservation as  $M_w = (\tau/f) \varphi(fR_m/V)$ , we further use  $\Delta \rho / \rho = (1/2) h_m N^2/g$ , to suggest

$$h_m = u_* / (fN)^{1/2} \varphi^{1/2}(fR_m/V) \quad (7)$$

where we took  $(2Ri_{cr})^{1/4} \approx 1$ . For  $fR_m/V \gg 1$ , the dimensionless function  $\varphi(fR_m/V)$  shall become constant, and equation (7) reduces to a classical relation for the mixed layer depth. For  $fR_m/V \ll 1$ , corresponding to a fast TC, function  $\varphi(fR_m/V)$  shall tend to  $\varphi \approx fR_m/V$ . Then TC induced SST anomalies from equation (4) with equation (6) shall follow

$$\delta \theta_s \approx \left[ u_* N^{3/2} / (g \alpha f^{1/2}) \right] \varphi^{1/2}(fR_m/V) [1 + (N_1/c) \delta D] \quad (8)$$

The term  $[1 + (N_1/c) \delta D]$  in equation (8) can be interpreted as an amplification factor of the SST anomalies due to the upwelling mechanism. Yablonsky and Ginis (2009) already pointed out that upwelling can play a significant role in the formation of SST anomalies: for a slow TC, with translation velocity less than 5 m/s, it becomes crucial. For very slow translation velocity, that is, 1–2 m/s, SST anomalies are dominated by the upwelling mechanism, and a classical 1-D ocean turbulence mixing model would not solely explain observations. As a first guess, we can thus assume the function  $\varphi^{1/2}(Ro) [1 + (N_1/c) \delta D]$  to mainly depend on the translation velocity, or on the TC Rossby number  $Ro = V/fR_m$  in our notation. Introducing  $\varphi_\theta(Ro) \equiv \varphi^{1/2}(Ro) [1 + (N_1/c) \delta D]$  and scaling  $u_*$  as  $u_* \propto u_m$ , TC-induced SST anomalies in equation (8) shall thus well be parameterized as

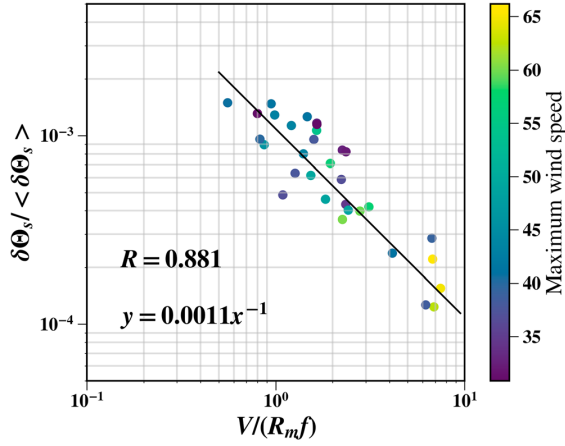
$$\delta \theta_s / \langle \delta \theta_s \rangle \propto \varphi_\theta(Ro) \quad (9)$$

where  $\langle \delta \theta_s \rangle$  represents an overall scaling of the SST anomalies:

$$\langle \delta \theta_s \rangle = u_m N^{3/2} / (g \alpha f^{1/2}) \quad (10)$$

In Figure 12, observed SST anomalies,  $\delta \theta_s$ , scaled by  $\langle \delta \theta_s \rangle$  are compared to TC Rossby numbers. As obtained, the suggested scaling seems to well apply. Consistent with results numerous reported in literature, the higher the wind speed and the shallower the thermocline, larger shall be the SST anomalies. As well, the slower and larger the TC, more pronounced will also be the SST anomalies. While satisfactorily, clear scatter is still noticeable. It suggests that the overall  $\langle \delta \theta_s \rangle$  is additionally impacted by other factors, in particular (as mentioned above) by the TC-induced upwelling impact.

Shown in Figure 13, SST wake offset,  $\delta x$ , and width,  $l$ , scaled by the radius of maximum wind speed are presented as a function of TC Rossby numbers. Again, conforming to previously reported observations (e.g.,



**Figure 12.** Observed sea surface temperature anomalies scaled by  $\langle \delta\Theta_s \rangle = u_m N^{3/2} / (g\alpha f^{1/2})$  versus tropical cyclone-Rossby number  $Ro = V / (fR_m)$ . Color indicates maximum wind speed. Solid line is the fit  $\delta\Theta_s / \langle \delta\Theta_s \rangle = 1.1 \times 10^{-3} Ro^{-1}$ .

D'Asaro et al., 2014), SST offsets (to the right from TC track) demonstrates a clear trend, increasing with increasing translation velocity.

### 3.2. SSH Anomalies

For the ocean surface height anomalies,  $\delta h_s$ , two contributions can be triggered: barotropic,  $\delta h_s^{bt}$  and baroclinic,  $\delta h_s^{bc}$ , as  $\delta h_s = \delta h_s^{bt} + \delta h_s^{bc}$ . For deep ocean conditions (depth about 5,000–7,000 m),  $\delta h_s^{bt}$  can be postulated much smaller than  $\delta h_s^{bc}$  (Geisler, 1970; Ginis & Sutyrin, 1995), and the barotropic contribution is further ignored in our analysis.

Invoking the pressure continuity condition at the surface, the following relationship is suggested between the surface vertical velocity,  $w_s$ , and the vertical velocity gradient,  $w'_z$ , beneath the surface:

$$w_s = (C^2/g)w'_z|_{z=0} \quad (11)$$

where  $C$  is the phase velocity of long internal waves (see, e.g., equation (3.9) from Kudryavtsev et al., 2019). For a two-layer approximation (Geisler, 1970), with a very shallow upper layer, and constant densities in each layer, relation (11) can be simply estimated, using  $C^2 = g(\Delta\rho/\rho)$  and  $w'_z|_{z=0} \approx w_D/D$  as follows:  $w_s = (\Delta\rho/\rho)w_D$ , where  $\Delta\rho/\rho$  is the density

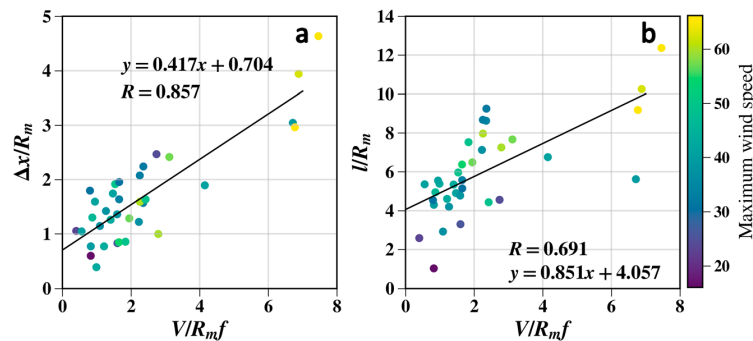
difference between layers scaled by the mean density value and  $w_D$  the vertical velocity of the pycnocline. Accordingly, compared to the pycnocline displacement, the surface displacement is attenuated by a factor  $\Delta\rho/\rho$ . As already discussed (see equation (5)), under a three-layer approximation, with constant density gradient in each layer, the vertical velocity in the upper seasonal thermocline layer becomes  $w(z) \propto \sin(Nz/C)$ , and equation (11) gives:

$$\delta h_s^{bc} \approx (NC/g)\delta D \quad (12)$$

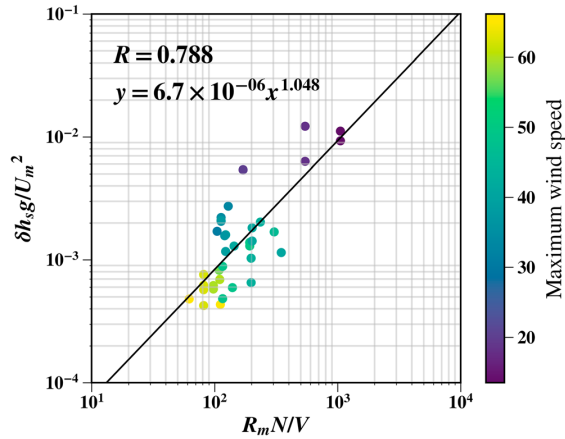
where  $\delta D$  is the displacement of the base of the main pycnocline. As an estimate of this displacement, we follow Geisler (1970). For a two-layer ocean response to moving hurricanes, the amplitude,  $\delta D$ , (Geisler, 1970; equation(37)) becomes:

$$\delta D \propto \tau_s R_m / (CV) \quad (13)$$

where  $\tau_s$  is the surface stress scaled by water density:  $\tau_s = (\rho_a/\rho_w)C_d u_m^2$ ,  $C_d$  the surface drag coefficient. Accordingly, equation (12) becomes



**Figure 13.** (a) Offset of sea surface temperature anomaly from tropical cyclone TC's track and (b) width of SST anomaly scaled by radius of maximum wind speed versus dimensionless translation velocity  $Ro = V/(fR_m)$  (the tropical cyclone-Rossby number). Color indicates wind speed. Solid lines are fits to the data by least squares method.



**Figure 14.** Dimensionless sea surface temperature anomalies,  $\delta h_s^{bc} g / u_m^2$ , versus parameter  $R_m N / V$  (symbols). Solid line is the fit to the data using least squares method. Color indicates tropical cyclone maximal wind speed.

$$\delta h_s^{bc} \propto (\tau_s / g) R_m N_1 / V \quad (14)$$

Assuming to first-order  $\tau_s \propto u_m^2$ , SSH anomalies shall scale as

$$\delta h_s^{bc} g / u_m^2 \propto R_m N / V \quad (15)$$

Dimensionless SSH anomalies as a function of parameter  $R_m N / V$  are shown in Figure 14. In spite of a rather large scatter, essentially caused by residual SSH anomalies left by mesoscale activities and other possible preceding TCs, previously traveling in the same area, the data demonstrate a clear trend. A fit to correlate dimensionless SSH anomalies versus parameter  $R_m N / V$ , gives

$$\delta h_s^{bc} g / u_m^2 = 6.9 \times 10^{-6} (R_m N / V)^{1.04} \quad (16)$$

The fit exponent is close to 1, and the proportionality constant scales very well with a surface drag coefficient of order  $C_d \propto 10^{-3}$  multiplied by  $\rho_a / \rho_w \propto 10^{-3}$ . This is further discussed in the next section.

#### 4. Hurricane-Force Wind Forcing

Proper definition of the drag coefficient at high wind speed condition, above 30 m/s, is still a fundamental issue. As generally recognized, the relation for  $C_d$  valid for moderate winds must not be extrapolated to hurricane-force wind conditions. Scanty amount of observations indeed demonstrates that  $C_d$  levels off and/or falls at wind speeds above 30 m/s (e.g., Powell et al., 2003).

From the present analysis, equation (14) may well provide unique opportunities to assess drag coefficients from observed TC-induced SSH anomalies,  $\delta h_s^{bc}$ . To further dwell on this anticipated property, it is thus tempting to infer a drag coefficient dependency as function of the TC maximum winds, as

$$C_d \propto (\rho_w / \rho_a) (V / N R_m) (g h_s^{bc} / u_m^2) \quad (17)$$

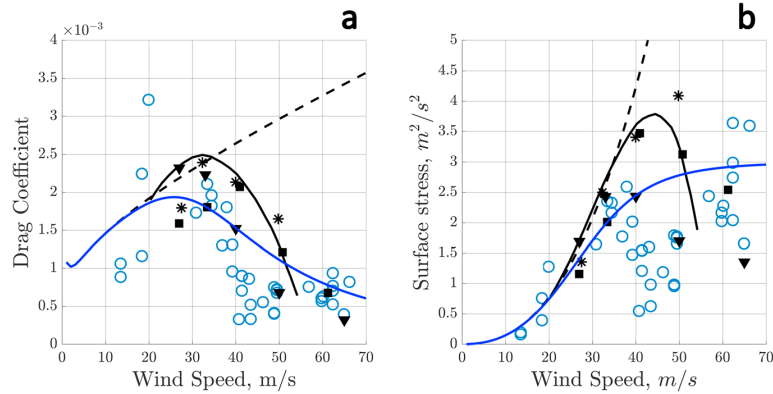
Adjusted with a proportionality constant equal to 1/6, the predicted drag coefficients are reported in Figure 15. As proposed, this analysis bears strong resemblance with the bottom-up approach applied by Jarosz et al. (2007). These authors infer estimates of drag coefficients using ocean current profile measurements. Though the collected SSH anomalies are rather scattered, due to presence of residual SSH anomalies associated to mesoscale eddy activities and/or left by other previous TCs traveling in the same area before, the “suggested direct dependency (17) remarkably recovers an apparent drag reduction for wind speed higher than 35 m/s. This is in line with estimates reported by Powell et al. (2003) and Jarosz et al. (2007). Estimates also favorably compare with a top-down approach using observations of the height of the planetary boundary layer (Powell et al., 2003, Figure 2; Kudryavtsev, 2006, Figure 9).

Moreover, calculations of the surface stress,  $\tau_s = C_d u^2$ , shown in Figure 15b, reveal remarkable feature—the surface stress has a clear trend to level off at hurricane-force wind conditions. Model simulations by Kudryavtsev (2006) and Kudryavtsev and Makin (2011) provide some theoretical grounds to interpret this behavior. From the present analysis, the surface stress can be parameterized as

$$\tau_s = (\tau_{s0}^{-m} + \tau_{st}^{-m})^{-1/m} \quad (18)$$

where  $\tau_{s0}$  is a reference stress calculated with the drag coefficient corresponding to the Charnock roughness length  $z_0 = 0.012 u^2 / g$ , and  $\tau_{st}$  is a threshold value of the stress,  $m$  is a tuning exponent. Surface stress parameterization in equation (18) and corresponding drag coefficient,  $C_d = \tau_s / u_m^2$  calculated for  $m = 2$  and  $\tau_{st} = 3$  m/s are reported in Figure 15; it fits the “cloud” of data and reproduce previously reported trends in the surface drag and the wind stress data.





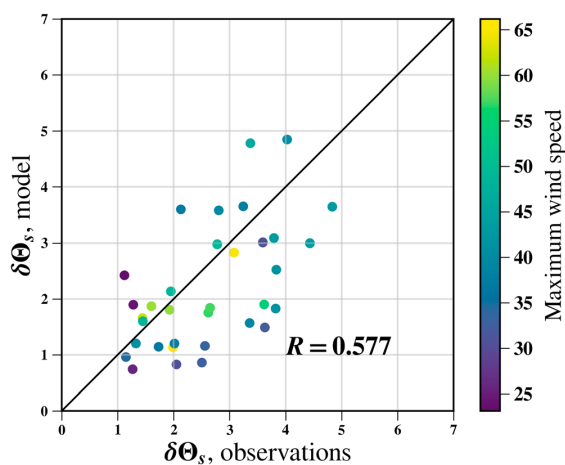
**Figure 15.** (a) Drag coefficient versus wind speed at 10-m height. Dashed line is  $C_d$  calculated for the roughness scale predicted by the Charnock relation:  $z_0 = 0.12u^2/g$ ; black squares are data by Powell (2006), compiled from his Figure 7, layer 20–160 m; stars are data by Powell et al. (2003) compiled from their Figure 3, layer 20–150 m; black solid line, fitted quadratic curve to the empirical data by Jarosz et al. (2007), their Figure 3; triangles are estimates by Kudryavtsev (2006); open circles show  $C_d$  derived from the altimeter SSH anomalies. (b) Corresponding surface stress versus wind speed calculated using  $C_d$  shown in Figure 12a.

### 5. Coupling Between SST and SSH Anomalies: Upwelling Mechanism Influence

As further interpreted, SSH anomalies provide direct estimates of the pycnocline upwelling associated to the passage of a TC. It can then be used to evaluate the upwelling impact, that is, the upwelling amplification, on the SST wake, equation (8). Using equation (12), this amplification factor can be expressed in terms of the SSH anomalies as

$$[1 + (N/C)\delta D] = 1 + g\delta h_s/C^2 \quad (19)$$

For the considered cases, SSH anomalies range from 0.1 to 0.5 m, and the IW phase velocities from 3 to 4 m/s, leading the amplification factor  $(1 + g\delta h_s/C^2)$  to vary between 1.1 and 1.5. The influence of the upwelling mechanism to control the strength of the SST wake is thus rather moderate. Nevertheless, the initial parameterization can now be extended, to explicitly introduce this upwelling factor as



**Figure 16.** Scatter plot demonstrating relation of observed SST anomalies versus anomalies predicted by (20).

$$\delta\Theta_s/\langle\delta\Theta_s\rangle = c_h \left[ 1 + c_u \frac{u_m^2 N}{C^2 f} Ro^{-1} \right] Ro^{-1} \quad (20)$$

with,  $\langle\delta\Theta_s\rangle = u_m N^{3/2}/(g\alpha f^{1/2})$  representing an overall scaling of the SST anomalies, and  $c_h$  and  $c_u$  the empirical constants:  $c_h = 0.8 \times 10^{-3}$   $c_u = 6.9 \times 10^{-6}$ .

Observed versus predicted SST by equation (20) anomalies are reported in Figure 16. An overall agreement is clearly achieved, suggesting that equation (20) correctly includes fundamental mechanisms governing the expected strength of the SST wake. Deviations are likely caused by numerous other factors, that is, the precise TC shape, estimated radius of maximum winds and associated wind stress spatial distribution, departure of the vertical gradient from a constant value, etc. Peculiar upper ocean salinity distribution can also contribute to build potential energy barrier to mixing, thus reducing the cooling magnitude (Balaguru et al., 2012). Precise prediction of SST anomalies should thus certainly be performed using a more elaborated approach. Yet, to first order, accounting for all mentioned factors may not seem fully necessary, as essential constraining parameters are included within the proposed parameterization.

## 6. Summary

In the present paper, a consistent-parameterization framework has thus been developed to motivate a more systematic combination of SST and SSH satellite observations to analyze the ocean responses to TCs. This framework is demonstrated using 20-day satellite observations intercepting three major TC events, namely Jimena, Ignacio, and Kilo, developing and traveling over the central and eastern parts of the Pacific Ocean. As gathered, multisensor observations provide a comprehensive data set on local changes of SST (derived from passive microwave measurements) and SSH (altimeter measurements) along the TCs paths, sampling different stages of intensification, wind speeds, radii, Coriolis parameter, translation velocities, and ocean interior stratification conditions.

To first order, measured SST and SSH anomalies do not exhibit clear dependences on any “individual” TC characteristic (e.g., maximal wind speed, size, translation velocity), or environment interior condition, for example, stratification. Yet, as suggested, scaled SST and SSH anomalies can still be robustly predicted and combined.

As derived, a semiempirical relation to interpret SST anomalies can essentially build on the heat and momentum conservations laws for the upper wind driven mixed layer (ML). Considering that the bulk Richardson number of the ML deepening (combining drift current, temperature change over the ML base, and ML depth) keeps a constant (critical) value, the following relationship to explain SST anomalies writes:

$$\delta\theta_s / \langle \delta\theta_s \rangle = \varphi_\theta (V/R_m f)$$

where  $\langle \delta\theta_s \rangle = u_m N^{3/2} / (g\alpha f^{1/2})$  is a scale of the SST anomalies, and  $\varphi_\theta$ , a function of the dimensionless TC translation velocity,  $V/R_m f$ , found empirically as  $\varphi_\theta = 1.1 \times 10^{-3} (V/R_m f)^{-1}$ . The offset of the SST anomalies from the TC track is then mostly governed by the TC translation velocity: offsets are larger for faster TCs.

As collected, TC passages have further been found to strongly imprint the ocean surface height. For the considered TC cases, expected amplitudes of the barotropic responses to the TC transient forcing were expected to be small, limited to about 1 to 5 cm. This is much smaller than the reported SSH anomalies. Therefore, for these cases, SSH anomalies must essentially be governed by the ocean baroclinic response. Following Geisler (1970), scaling arguments to estimate the thermocline displacement induced by a moving TC, lead to the following relationship of SSH anomalies with TC and ocean interior parameters:

$$gh_s^{bc} / u_m^2 \propto R_m N_1 / U$$

As demonstrated, this relationship may well provide unique opportunities to document the TC wind forcing and to assess drag coefficient from observed TC-induced SSH anomalies. From the collected SSH anomalies, a drag reduction is remarkably recovered for wind speed higher than 35 m/s, in line with estimates reported by Powell et al. (2003), Powell (2006) and Jarosz et al. (2007).

As interpreted, SSH anomalies thus provide direct estimates to evaluate the upwelling impact, that is, the upwelling amplification, on the SST wake. For the cases studied, the influence of the upwelling mechanism has been found to be rather moderate, of order 10–40%.

Building on the actual satellite altimeter constellation (presently up to six satellites are available), the proposed interpretation framework can thus guide the combined use of SST and SSH amplitude changes measured in the wakes of TCs. It can help to analyze the ocean response to TCs, and to first-order inform about the resulting strength of hurricane-induced mixing and upwelling. As mentioned above, the bottom-up approach can also guide future investigations to help document the resulting wind forcing and practical drag coefficient under extreme TC-conditions. In that context, it can also be anticipated that next NASA's Surface Water and Ocean Topography (Fu et al., 2012), with unprecedented 2-D altimeter mapping capabilities, certainly promises to greatly improve the analysis of TC-induced SSH wake: improved knowledge of the air-sea exchanges under TCs might thus be an unexpected outcome of this mission.

The proposed simplified framework is further extended in a 2019. It provides a more complete analytical description of ocean response to moving TC, especially detailing the wind-driven current field, ML



cooling and its evolution, and associated space–time variability of the pycnocline caused by TC-induced baroclinic motions.

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# Barotropic and Baroclinic model of the ocean response

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## RESEARCH ARTICLE

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## Key Points:

- Simplified analytical model for baroclinic and barotropic ocean response to TC and its impact on pycnocline erosions is suggested
- The model demonstrated its ability to reproduce SSH and SST observations in wide range of TC parameters and ocean stratification
- Given its low computational burden, the model can be introduced as a computational module into numerical coupled TC-ocean models

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## A Simplified Model for the Baroclinic and Barotropic Ocean Response to Moving Tropical Cyclones: 2. Model and Simulations

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**Abstract** A simplified analytical model is developed to describe the baroclinic and barotropic ocean response to moving tropical cyclones (TCs) and their associated pycnocline erosions. The model builds on classical mixed-layer (ML) models and linear models of ocean response to transient events. As suggested, disturbances of the upper ocean stratification caused by the ML development shall not strongly impact the dynamics of baroclinic modes. Accordingly, the baroclinic response can be estimated using the prestorm ocean stratification condition. To the contrary, the ML is strongly coupled with these interior motions, through the TC-induced upwelling response that affects the entrainment velocity. The ML temperature is then strongly dependent on the local temperature gradient in the upper layer. The model is represented by a set of analytical relationships providing rapid calculations for the ocean response to TC, given a prescribed wind velocity field traveling over an ocean with arbitrary stratification. Compared to satellite observations, simulations demonstrate the model ability to quantitatively reproduce the observed shape and magnitudes of the sea surface height and the sea surface temperature (SST) anomalies. Remarkably, the model is robust and efficient for a wide range of variability of TC characteristics (max wind speed, radius, shape of wind profile, and translation velocity), parameters of the ocean stratification, and Coriolis parameter. Simulations provide solid evidences about the key role of TC-induced upwelling in the ML cooling and formation of SST wake. Cross-track advection by wind-driven currents, though small compared with TC translation velocity, can significantly contribute to broaden the shape and offset of the SST wake. Given its effectiveness and low computational burden, the proposed model can be introduced as a computational module into atmospheric numerical models of TC-coupled evolution with the ocean, through the resulting local changes of surface enthalpy fluxes.

### 1. Introduction

A complete description of upper ocean responses to tropical cyclone (TC) transient and extreme wind forcing remains a difficult problem. Crucial to quantify both the momentum increase of surface currents and the efficiency of vertical mixing in cooling the ocean surface, the parameterization of the wind forcing under extreme conditions is indeed still poorly known. Moreover, besides the combining effects of the TC intensity and translation speed, upper ocean precyclone stratification can also strongly mitigate or exacerbate TC-induced cooling amplitude (Lloyd & Vecchi, 2011; Schade, 2000). The intensity-wake relationship can then be strongly modulated with evolving air-sea fluxes associated to sea surface temperature (SST) changes over the TC path (Cione & Uhlhorn, 2003). Passage over freshwater plumes has also been reported to cause strengthening of hurricanes, due to the presence of a barrier layer effect (e.g., Balaguru et al., 2012; Reul et al., 2014).

To assess the sensitivity of the parameterization of wind forcing at high winds and/or the impact of precyclone upper ocean stratification (Yablonsky & Ginis, 2013; Zedler et al., 2009; Zedler et al., 2012), advanced numerical ocean models are generally implemented to simulate the ocean response. These three-dimensional and time-dependent models (e.g., Price et al., 1994; Sanford et al., 2007) consistently solve the momentum, heat, and salt budget equations. Numerical outputs then provide detailed descriptions of ocean response to moving TC, including description of wind-driven current field, ML cooling, its evolution, and the space-time variability of the pycnocline caused by TC-induced baroclinic motions. However, these

advanced models require significant computer resources and simulations are time-consuming. This may appear as major hurdles to develop fully coupled ocean-atmosphere models and/or try to best sample the whole uncertainty space to apply inverse approaches, and more particularly to test and infer wind forcing parameterization.

In the latter context, different methods have been proposed. Sraj et al. (2013) proposed polynomial chaos expansions to construct a faithful surrogate of the response of upper ocean model simulations. For the same purpose, Zedler et al. (2013) adopt a numerical approach in an inverse problem setup, using the ocean model and its adjoint. Data can thus be assimilated and the drag coefficient adjusted, to help correspond to the minimum of a model minus data misfit or cost function.

To possibly serve these inverse approaches, as well as to derive a simplified framework to rapidly interpret satellite observations (e.g., Kudryavtsev et al., companion paper), it is tempting to circumvent the computational burden of advanced numerical simulations. This is the main purpose of the present paper. The goal is to develop a simplified analytical model to assess, on quantitatively correct levels, ocean responses to moving TCs.

As proposed, model derivations can essentially build on classical 1D mixed-layer models (e.g., Niiler, 1975; Zilitinkevich et al., 1979) and previously suggested analytical models for the barotropic and baroclinic responses (e.g., Geisler, 1970; Orlanski & Polinsky, 1983). To rapidly outline the proposed developments, it is recalled that wind-driven currents, generated by a TC wind stress field, are localized in the mixed layer (ML). In the open ocean, the ML depth,  $h$ , is typically much smaller than the depth,  $H$ , of the quiet interior layer,  $h \ll H$ . The ML layer is highly turbulent. The interior layer is not. Consequently, the TC-induced deepening of the ML shall lead to marked *discontinuities* in density and current velocity profile at the base of the ML, separating the highly turbulent upper ocean from the nonturbulent interior layer (Price, 1981). Caused by the vorticity of the TC surface wind stress, wind-driven currents in the ML are divergent, further leading to lift the interior layer, triggering inertia-gravity internal waves and associated baroclinic wake (e.g., Gill, 1982; Ginis, 2002). A barotropic response of the ocean to the wind stress vorticity action is also expected. Ginis and Sutyrin (1995) argued that barotropic and baroclinic modes do not interact and can thus independently be considered. Finally, feedbacks between ML and baroclinic motions can also encompass the impact of the thermocline upwelling on the ML cooling. Indeed, both numerical simulations and experimental observations (Yablonsky & Ginis, 2009) provide solid evidences about the key role of TC-induced upwelling to control the ML cooling, especially for slowly translating TCs.

The paper is organized as the following. In section 6 we introduce the governing equations for the upper ML (sections 6.1 and 2.2) and the interior layer (section 6.3). Simplification of the governing equations and analytical solutions describing baroclinic and barotropic responses to moving TC, and the SST anomalies is given in section 3. Section 4 presents some results of model simulations, demonstrating general properties of the model. Model simulations are applied to compare with satellite observations of the SST and the surface height anomalies, reported in the companion paper, and are presented in section 5. Conclusion section outlines suggested model and discusses its possible applications.

## 2. Governing Equations

Considering an integrated description, within the upper highly turbulent ML of depth  $h$ , the density,  $\rho_m$ , temperature,  $\theta_m$ , salinity,  $s_m$ , and current velocity,  $\mathbf{u}$ , are constant. Below the ML, the interior is nonturbulent and continuously stratified,  $\rho = \rho(z)$ . The ML being highly turbulent, sharp density/temperature jumps must develop, with associated current velocity changes at its base, to express a marked separation from the nonturbulent interior layer.

In coordinate system traveling with the moving TC, stationary solutions are further considered. The TC moves opposite to  $x_1$  direction with translation velocity  $-V$  ( $V$  is positive value), and the TC eye coincides with  $(x_1, x_2)$ -coordinate origin. As such, partial time derivatives in all equations below are equivalent to



$$\partial/\partial t = V\partial/\partial x_1$$

### 2.1. Upper Mixed Layer

Integrated over the ML depth,  $h$ , momentum and heat conservation equations read

$$h\left(\partial u_\alpha/\partial t + u_\beta\partial u_\alpha/\partial x_\beta + \varepsilon_{\alpha\beta}fu_\beta\right) = -gh\partial\zeta_s/\partial x_\alpha + \tau_\alpha^s - \tau_\alpha^{-h+0} \quad (1)$$

$$h(\partial\theta_m/\partial t + u_\beta\partial\theta_m/\partial x_\beta) = -q_s + q_{-h+0} \quad (2)$$

where  $\alpha, \beta = 1, 2$  are the indexes,  $\zeta_s$  is the ocean surface displacement,  $f$  is the Coriolis parameter,  $\tau_\alpha^s$  and  $q_s$  are the surface wind stress (scaled by water density) and kinematic heat fluxes, respectively,  $\tau_\alpha^{-h+0}$  and  $q_{-h+0}$  are the turbulent stress and heat flux at the ML base,  $\varepsilon_{\alpha\beta}$  is the unit antisymmetric tensor ( $\varepsilon_{\alpha\beta} = 0$  if  $\alpha = \beta$  and  $\varepsilon_{12} = -1, \varepsilon_{21} = 1$ ), and  $x_3$  axis is directed upward. The water density is defined by a known sea state law in the form  $\rho = \rho_0(1 + \alpha\theta + \beta s)$ , where  $\alpha$  and  $\beta$  are thermal and haline expansion coefficients. As heat and salt balance equations are very similar, we solely consider the heat conservation equation, a corresponding equation for salinity being derived by replacing  $\theta$  by  $s$  and surface heat flux by overall resultant flux of the fresh water mass (precipitation minus evaporation). Turbulent stress and heat flux at the ML bottom,  $x_3 = -h+0$ , are defined as

$$\begin{aligned} \tau_\alpha^{-h+0} &= u_\alpha w_e \\ q_{-h+0} &= -\Delta\theta w_e \end{aligned} \quad (3)$$

where  $\Delta\theta = \theta_m - \theta_{-h-0}$  is the sharp temperature change at the ML base,  $\theta_{-h-0}$  is the ocean temperature  $\theta(x_3)$  below the ML at  $x_3 = -h-0$ , and  $w_e$  is the entrainment velocity defined as

$$w_e = \partial h/\partial t + \partial M_\beta^w/\partial x_\beta \quad (4)$$

where  $\partial M_\beta^w/\partial x_\beta$  stands for the divergence of the total wind-driven transport,  $M_\beta^w = hu_\beta$ . Definition 4 applies as long as the ML is deepening/developing, that is, when  $\partial h/\partial t + \partial M_\beta^w/\partial x_\beta > 0$ . In all other cases,  $w_e = 0$ . Solely considering the linearized problem, use of equation (4) leads to the following momentum and heat balance equations:

$$\partial M_\alpha^w/\partial t + \varepsilon_{\alpha\beta}fM_\beta^w = -gh\partial\zeta_s/\partial x_\alpha + \tau_\alpha^s \quad (5)$$

$$\partial\theta_m/\partial t = -1/2\bar{\Gamma}w_e - q_s/h \quad (6)$$

where

$$\bar{\Gamma} = (2/h^2)\int_0^h z\Gamma dz \quad (7)$$

stands for the background (prestorm) temperature gradient averaged over the ML depth.

### 2.2. ML Depth

A review and developments of ML depth models can be found in Niiler (1975) and Zilitinkevich et al. (1979). Here we follow the assumption that ML deepens as to maintain a bulk Richardson number constant, that is, a critical  $Ri_c$  assumption. Considering the ML thickness,  $h$ , and sharp changes of buoyancy,  $\Delta\rho$ , and current velocity,  $|\Delta\mathbf{u}|$ , at the ML base, the critical,  $Ri_c$  is defined as

$$Ri_c = \frac{g(-\Delta\rho/\rho_0)h}{|\Delta\mathbf{u}|^2} \quad (8)$$

Parametrization 8 had been widely used as to define the closure scheme for subgrid processes in numerical models simulating ocean response to TC passage, starting from the pioneering work by Price (1981) and further refined in Price et al. (1986). A sharp buoyancy jump at the ML base,  $g\Delta\rho/\rho_0 = g(\alpha\Delta\theta + \beta\Delta s)$ , follows from the heat and salt balances within the ML. Ignoring effects of heat and salt surface fluxes, it comes



$$g(\Delta\rho/\rho_0) = -\frac{1}{2}h\overline{N^2} \quad (9)$$

where  $\overline{N^2}$  is the background (prestorm) Brunt-Väisälä frequency,  $N^2$ , averaged over the ML depth:

$$\overline{N^2} = (2/h^2) \int_{-h}^0 N^2 dz \quad (10)$$

Velocity sharp change,  $|\Delta \mathbf{u}|$  in 8, is associated to wind-driven velocities,  $\mathbf{u}$ , in the ML. Considering the magnitude of the volume wind-driven transport,  $|M| = h|\mathbf{u}|$ , the following relationship for the ML depth can be derived:

$$h^4 = 2Ri_c \frac{|M|^2}{N^2} \quad (11)$$

Scaling  $|M|$  as  $|M| \propto u_{10}^2/f$ , a classical relation for the ML depth in the stratified ocean is recovered,  $h \propto u_{10}/\sqrt{fN}$ , (Pollard et al., 1972). For TC applications, the volume transport in equation (11) can be considered as a *pure* wind-driven transport, defined as solution of momentum balance equation (5) written in the rigid lid approximation:

$$M(x_1, x_2) = f^{-1} \int_{x_1^0}^{x_1} \tau^s(x_1, x_2) \exp[-ik_0(x_1 - x_1')] d(k_0 x_1') \quad (12)$$

where  $M = M_1 + iM_2$  is the complex volume wind-driven transport,  $k_0 = f/V$  is the wave number of inertial oscillations, with boundary  $(x_1^0, x_2)$  far ahead of the TC eye, where wind-driven currents induced by TC vanish,  $M = 0$ .

### 2.3. Interior Layer

To treat the interior layer and its coupling with ML, we follow the model suggested by Orlanski and Polinsky (1983; see their section 2 for the details). The governing equation to describe the interior layer dynamics,  $x_3 < -h$ , is the equation for vertical velocity,  $w$ . In the Fourier space, it reads

$$(\Omega^2 - f^2) \hat{w}'' - k^2 (\Omega^2 - N^2) \hat{w} = 0 \quad (13)$$

where hat over any quantity denotes Fourier transform,  $\Omega = k_1 V$  is the frequency,  $k_\alpha$  is the wave number components, and,  $k = (k_1^2 + k_2^2)^{1/2}$ , double prime indicates second derivative over the depth. This equation must be coupled with vertical velocity in the ML. Transforming the momentum balance (5), the coupling reads

$$(\Omega^2 - f^2) (\hat{w}_s - \hat{w}_h) - g h k^2 \hat{w}_s = \hat{F} \quad (14)$$

where subscripts “s” and “h” for  $\hat{w}$  denote vertical velocity at the surface and at the ML base, respectively, and  $F$  is the wind stress forcing term

$$F = f \cdot \text{Rot}(\boldsymbol{\tau}) - V \partial / \partial x_1 [\text{Div}(\boldsymbol{\tau})] \quad (15)$$

combining vorticity,  $\text{Rot}(\boldsymbol{\tau})$ , and divergence,  $\text{Div}(\boldsymbol{\tau})$ , of surface wind stress field. Among these terms, only the vorticity term is capable to produce a steady state response in the ocean (Orlanski & Polinsky, 1983). Hereinafter, the second term in (15) is therefore ignored, and  $F$  solely governed by the wind stress vorticity. Note that  $w_h$  in equation (14) equals the divergence of the volume transport,  $w_h = \partial M_\beta^w / \partial x_\beta$ , and defines the entrainment velocity (4).

At  $x_3 = -h$  equations (13) and (14) are coupled through the kinematic,  $\hat{w}|_{x_3=-h} = \hat{w}_h$ , and the dynamic,  $(\Omega^2 - f^2) \hat{w}'|_{x_3=-h} = g k^2 [\hat{w}_s + (\Delta\rho/\rho) \hat{w}_h]$ , boundary conditions, where prime indicates derivative over the depth. At the bottom,  $x_3 = -H$ , solutions must satisfy the boundary condition  $\hat{w}|_{x_3=-H} = 0$ .



### 3. Simplification of Governing Equations

Considering open ocean conditions, the ML depth is  $O(10\text{--}100\text{ m})$ , small compared to both the ocean depth and the pycnocline depth,  $D$ ,  $O(1,000\text{ m})$ . Developing ML preserves mass, and the density jump at the ML base is also small,  $O(h/D)$ , compared to the total density difference over the pycnocline. Consequently, the ML somehow appears as a *small perturbation* of the vertical ocean stratification, localized in a *thin* layer beneath the ocean surface. This significantly simplifies the problem in the following way.

As a first iteration, we can indeed *decouple* the interior layer dynamic from the ML. A baroclinic response can be readily estimated using the prestorm stratification condition, that is, without accounting for the small perturbations caused by the ML development and associated pycnocline erosion. At variance, the ML, and first of all its temperature, is strongly coupled with the interior layer through the entrainment velocity (4) affected by the upwelling velocity below the ML base. Thus, once the vertical velocity in the interior layer is determined, the system of ML heat balance equations, (6), (7), (4), and (11), is closed.

#### 3.1. Baroclinic Response

##### 3.1.1. Three-Layer Approximation of Interior Layer

To derive a practical analytical solution describing the baroclinic response, the vertical stratification in the interior layer is approximated by a three-layer model with constant  $N$  in each of the layers: the seasonal,  $x_3 > -d$ , and the main,  $-D < x_3 < -d$ , pycnoclines with  $N_1$  and  $N_2$ , respectively, and abyssal,  $x_3 < -D$ , with  $N_3 = 0$  (see Figure 9 in Kudryavtsev et al., companion paper). As tested, a three-layer approximation well fits general properties of the ocean vertical stratification, at least with a sufficient accuracy to adequately estimate TC-baroclinic responses. In addition, we assume that seasonal and main thermocline Brunt-Väisälä frequencies are much larger than the Coriolis parameter,  $(N_1, N_2) \gg f$ . As TC-baroclinic responses are represented by near-inertial internal waves with  $\Omega \approx f$ , it also implies  $\Omega \ll (N_1, N_2)$ .

##### 3.1.2. Solutions

Solution of equation (13) for a three-layer approximation of the interior layer stratification, coupled with vertical velocity in the ML, equation (14), through boundary conditions, is detailed in equations (A.1) to (A.3). For typical ocean conditions, the ML depth is  $O(10\text{--}100\text{ m})$ , and Brunt-Väisälä frequency in the seasonal pycnocline,  $N_1$ , is about  $N_1 \propto 10^{-2}\text{ s}^{-1}$  or less. In this case, parameter  $N_1 h/C$ , where  $C$  is the phase velocity of long surface waves ( $C$  is about 2 to 3 m/s), is small,  $N_1 h/c = 3 \times (10^{-2} \text{ to } 10^{-1})$ . Solutions A.1 to A.3 with accuracy to small parameter  $1/2(N_1 h/c)^2 \ll 1$  can be simplified to relations (A.6)–(A.7), not explicitly dependent on the ML parameters. This fact suggests that while entirely driven by divergence of wind currents in the ML, the vertical structure of baroclinic motions is not strongly affected by *small* perturbations of the ocean stratification associated with the ML. Accordingly, the baroclinic response can effectively be calculated using prestorm stratification conditions.

The vertical velocity of baroclinic motions of the  $n$  mode in the seasonal and the main pycnocline can be written as (see section b in Appendix for details)

$$w_n(\mathbf{x}, z) = \begin{cases} a_{1n} W_n(\mathbf{x}) \sin[N_1 x_3 / C_n], & \text{at } -d < x_3 < 0 \\ a_{2n} W_n(\mathbf{x}) \sin[N_2(D + x_3) / C_n + \varphi_n], & \text{at } -D < x_3 < -d \end{cases} \quad (16)$$

and in the abyssal,  $-H < x_3 < -D$ , it linearly varies with depth, as

$$w_n(\mathbf{x}, z) = w_n(\mathbf{x}, D)(H + x_3)/(H - D) \quad (17)$$

In these relations, subscript “ $n$ ” for any quantity indicates the mode number,  $C_n$  is the phase velocity of long internal waves of the  $n$  mode defined from the dispersion relationship:

$$\sin \left[ \frac{N_1 d}{C_n} + \frac{N_2(D-d)}{C_n} + \varphi_n \right] - \frac{N_1 - N_2}{N_1 + N_2} \sin \left[ \frac{N_1 d}{C_n} - \frac{N_2(D-d)}{C_n} - \varphi_n \right] = 0 \quad (18)$$

with  $\sin \varphi_n = [N_2(H-D)/C_n] / \sqrt{1 + [N_2(H-D)/C_n]^2}$  (for the deep ocean  $\varphi = \pi/2$ ),  $a_{1n}$  and  $a_{2n}$  are dimensionless vertical velocity amplitudes defined by A.13, and  $W_n(\mathbf{x})$  is a function defining 2D field of vertical velocity for  $n$  mode:



$$W_n(x_1, x_2) = \frac{1}{4\pi^2 C_n^2} \iint \frac{\hat{F}(k_1, k_2)}{(V^2/C_n^2 - 1)k_1^2 - k_2^2 - \kappa^2} e^{ik_\beta x_\beta} dk_1 dk_2 \quad (19)$$

where  $\kappa = f/C_n$  is the inverse baroclinic radius of deformation. If the TC-translation velocity is large enough to satisfy  $V > C_n$ , relation (19) exhibits a singularity around a *resonant* curve in the wave number space:  $k_1^2 (V^2/C_n^2 - 1) - k_2^2 - \kappa^2 = 0$ .

An analytical solution for this case is given by Geisler (1970, his equation (30)). For our purpose, a slightly different form of the solution is suggested to be numerically estimated using fast Fourier transform. This solution reads (see section b in Appendix for details):

$$W_n(x_1, x_2) = \frac{1}{2\pi(V^2 - C_n^2)} \int_{-\infty}^x dx'_1 \int_{-\infty}^{\infty} \hat{F}(x'_1, k_2) \frac{\sin(k_{01}(x_1 - x'_1))}{k_{01}} e^{ik_2 x_2} dk_2 \quad (20)$$

where  $\hat{F}(x'_1, k_2)$  is the  $k_2$  Fourier transform of wind stress source for a given  $x'_1$ , and  $k_{01}$  is the resonant wave number defined by A.17. For slow TC,  $V < C_n$ , relationship (19) does not possess any singularity, and the vertical velocity is found directly as Fourier transform.

Solution 20 for fast TC,  $V > C_n$ , and solution 19 for slow TC,  $V < C_n$ , together with 16 to 18 provide a full description of the TC-baroclinic vertical motions. Given vertical velocities, displacements of the water masses at a given depth,  $\zeta(\mathbf{x}, x_3)$ , are then straightforwardly determined as the integral of vertical velocity over the horizontal space:

$$\zeta_n(\mathbf{x}, x_3) = V^{-1} \int_{-\infty}^{x_1} w_n(\mathbf{x}, x_3) dx_1 \quad (21)$$

Correspondingly, total displacement is to be found as superposition of the different modes.

### 3.2. Surface Height Anomalies Induced by TC

Ginis and Sutyrin (1995) argued that barotropic and baroclinic modes do not interact and can thus independently be considered. The TC-induced total sea surface height (SSH) anomalies,  $\zeta_s$ , thus writes as a sum of baroclinic,  $\zeta_s^{bc}$ , and barotropic,  $\zeta_s^{bt}$ , components:  $\zeta_s = \zeta_s^{bc} + \zeta_s^{bt}$ . To second order of the small parameter  $N_1 h/C_n$ , the surface vertical velocity associated to the  $n$  mode of the baroclinic motions follows from boundary conditions and reads as

$$w_{sn}^{bc}(\mathbf{x}) = (C_n^2/g) w_n' \big|_{x_3=0} = a_{1n} (N_1 C_n/g) W_n(\mathbf{x}) \quad (22)$$

The corresponding SSH anomalies,  $\zeta_s$ , follow from 21 with 22.

The barotropic component of the surface vertical velocity is taken into account in the governing equations (13) and (14). The barotropic mode is not sensitive to the ocean stratification, attenuates linearly with the depth, and the maximum of vertical velocity occurs at the surface. Evaluating  $w_h'$ , as  $\hat{w}_h' = \hat{w}_h/(H-h)$ , expressing then  $\hat{w}_h$  via  $\hat{w}_s$ , and substituting it to (14), we arrive at equation

$$\hat{w}_s^{bt} [(V^2/C_b^2 - 1)k_1^2 - k_2^2 - f^2/C_b^2] = \hat{F}/C_b^2 \quad (23)$$

which is similar to A.14, except that  $C_b$  corresponds to the phase velocity of long surface waves:  $C_b^2 = gH$ . Since  $C_b > V$ , barotropic vertical velocity can be directly found from 23 as inverse Fourier transform:

$$w_s^{bt}(x_1, x_2) = -1/(4\pi^2 gH) \iint \hat{F}/(k^2 + \kappa_{bt}^2) e^{ik_\beta x_\beta} dk_1 dk_2 \quad (24)$$

where we ignored  $V$ , very small compared to  $C_b$ , and  $\kappa_{bt} = f/C_b$  is the inverse barotropic radius of deformation. Correspondingly, the SSH anomalies caused by the barotropic mode can be calculated using 21 with 24.

### 3.3. Heat Balance in the Upper Mixed Layer

As considered, the three-layer model has been introduced to derive tractable analytical solutions for the baroclinic motions. Deviations of the real stratification from a three-layer approximation should not

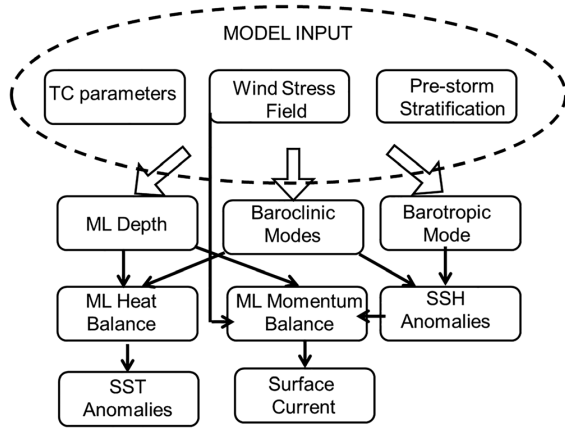


Figure 1. Flow diagram of the model.

quantitatively impact the main characteristics of the baroclinic modes. At variance, the ML heat balance, equation (6), and its depth, equation (11), drastically depend on the stratification in the upper ocean. Thus, to simulate the ML and the SST wake, realistic ocean stratification must be considered. As discussed in section 6.2, the ML evolves so that

$$Ri = (1/2)h^4 \overline{N^2} / M_m^2 \quad (25)$$

keeps a critical value,  $Ri = Ri_{cr}$ , see equation (11). This may only be valid for deepening and/or stationary conditions. The ML deepening is an irreversible process. Therefore, the ML depth must not decrease when the total wind-driven transport/stratification is locally reduced/enhanced. For such conditions, the ML depth either keeps a constant value or undulates following vertical movements:  $\partial h / \partial t = -w_h$ . Following such reasoning, the entrainment velocity at the base of the ML is defined as

$$w_e = \begin{cases} V \partial h / \partial x_1 + w_h, & \text{if } Ri = Ri_{cr} \\ 0, & \text{if } Ri > Ri_{cr} \end{cases} \quad (26)$$

During the forced stage cooling/warming of the ML due to the surface heat fluxes plays a secondary role (Price et al., 1994); therefore, term  $q_s/h$  in 6 can be ignored. Surface heating of the TC cold wake acts on time scale of order few days. Equation (6) thus simplifies to

$$\partial \theta_m(x_1, x_2) / \partial x_1 = -1/2 \overline{\Gamma} w_e / V \quad (27)$$

or in terms of the SST anomalies,  $\delta \theta_s$ :

$$\delta \theta_s(x_1, x_2) = -1/2 V^{-1} \int_{x_{10}}^{x_1} \overline{\Gamma} w_e dx'_1 \quad (28)$$

This relation together with 26, 7, 11 and vertical velocity 16 taken at the base of the ML completes the ML description.

Flow diagram, Figure 1, demonstrates links between different components of the model.

#### 4. Model Results

To specify the wind forcing, the radial wind speed profile can follow the form suggested by Holland (1980), as

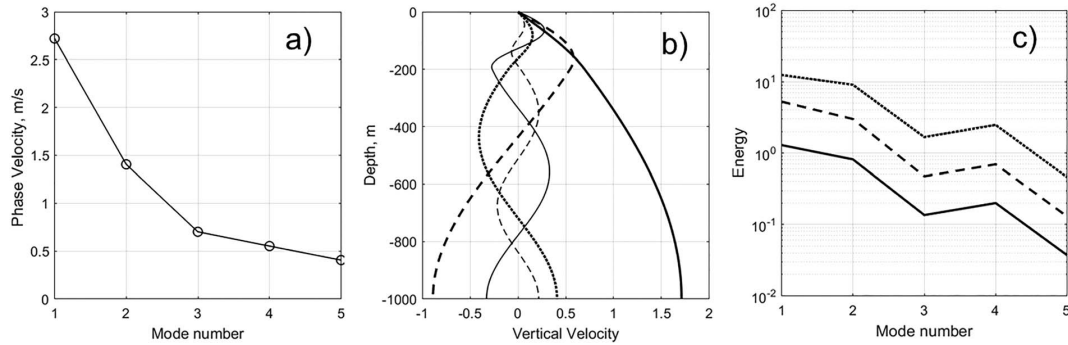
$$u(r) = \left[ (u_m^2 + u_m r f) \left( \frac{R_m}{r} \right)^B \exp \left( - \left( \frac{R_m}{r} \right)^B + 1 \right) + \left( \frac{r f}{2} \right)^2 \right]^{1/2} - \frac{r f}{2} \quad (29)$$

where  $u_m$  is the maximum wind speed at 10-m level,  $R_m$  is the radius of maximum wind speed, and  $B$  defines the shape of the wind field with increasing radial distance  $r$ .

Below we present some results, considering the following parameters, typical of ocean environmental conditions:  $N_1 = 1.36 \times 10^{-2} \text{ 1/s}$ ,  $N_2 = 3.9 \times 10^{-3} \text{ 1/s}$ ,  $d = 200 \text{ m}$ ,  $D = 1,000 \text{ m}$ , and  $H = 5,000 \text{ m}$ . Wind speed is specified by 29, and it is assumed that wind velocity spirals toward the TC eye, with a constant inflow angle of  $25^\circ$  (Shea & Gray, 1973), and  $R_m = 50 \text{ km}$ ,  $u_m = 50 \text{ m/s}$ , and  $B = 1.5$ . The wind stress acting to the ocean surface is

$$\tau_\alpha^s = (\rho_a / \rho_w) C_d u u_\alpha \quad (30)$$

where  $C_d$  is drag coefficient,  $u$  is wind speed at reference level (e.g., 10 m), and  $u_\alpha$  is wind velocity component. Parameterization of the drag coefficient at high wind speeds is still a matter of debate. As generally recognized, the relation for  $C_d$ , valid for moderate winds, must not be extrapolated to hurricane-force wind conditions. Scanty amount of observations demonstrates that  $C_d$  levels off and/or falls at wind speeds above



**Figure 2.** (a) Dispersion relation (18) for the five first modes; (b) profiles of vertical velocity for the five first modes: thick solid, dashed, and dotted lines, and thin solid and dashed lines, respectively; (c) redistribution of energy (conventional units) between baroclinic modes generated by TC traveling with translation velocities: (solid)  $V = 10$  m/s, (dashed)  $V = 5$  m/s, and (dotted)  $V = 2.5$  m/s. TC = tropical cyclone.

30 m/s (e.g., Powell et al., 2003; also see Figure 14 in the companion paper). For our demonstration purpose, the surface drag coefficient and wind stress parametrizations follow those suggested in Kudryavtsev et al. (companion paper).

#### 4.1. Baroclinic and Barotropic Responses

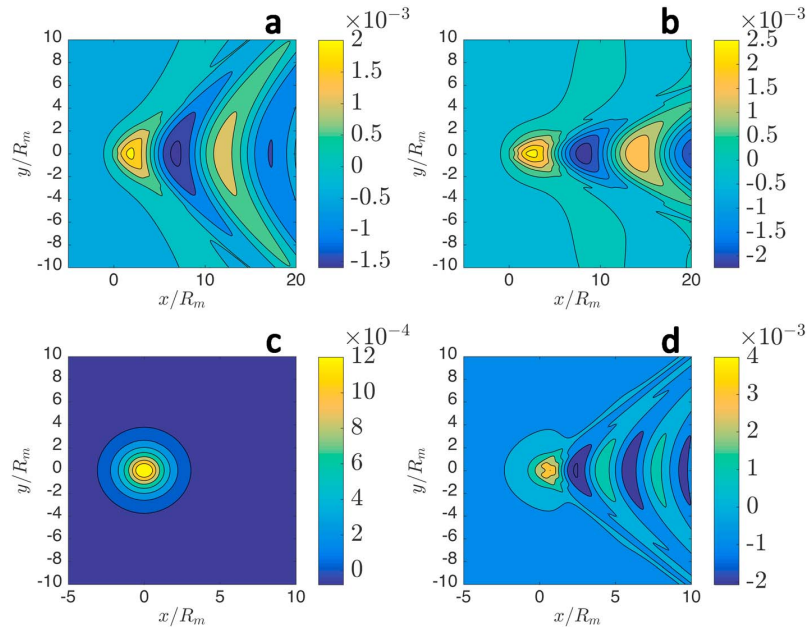
Phase velocities of long IW, using dispersion relation (18) and vertical structure for the first five modes, are shown in Figure 2. The low limit of the depth axis in Figure 2b corresponds to the depth of the main pycnocline base,  $x_3 = -D$ ; hence, vertical velocity of each of the modes at  $x_3 < -D$  attenuates linearly to meet boundary condition  $w = 0$  at  $x_3 = -H$ . Figure 2c illustrates the redistribution of the (conventional) energy

$$E(n) = \int_{-D}^0 N^2 \zeta_n^2 dx_3 \quad (31)$$

of the baroclinic motions, between the different modes triggered by a moving TC;  $\zeta_n$  in 31 is the pycnocline displacement for each mode defined by 21. From Figure 2c, most part of the energy of baroclinic motions is contained in the two first baroclinic modes. Similar results (not shown) are generally found for other TC parameters and ocean stratification conditions. Therefore, we only consider the baroclinic response as the composition of these two lowest baroclinic modes.

Figure 3 illustrates the 2D fields of vertical velocities, equation (19), generated by TC, either *fast* ( $V = 5$  m/s) relative to both IW modes,  $V/C_1 = 1.84$  and  $V/C_2 = 3.56$ , or *slow* ( $V = 2$  m/s) relative to the first mode,  $V/C_1 = 0.74$ , but fast relative to the second one,  $V/C_2 = 1.42$ . If the TC is fast, it generates wake of near-inertial IWs (Geisler, 1970). The wave number of these waves is about  $k_0 = (f/C_n)/\sqrt{V^2/C_n^2 - 1}$ , due to their dispersive nature, and the amplitude of the vertical velocity attenuates with the distance from the TC eye (along the central line,  $x_2 = 0$ ) as  $\propto (k_0 x_1)^{-1/2}$ . In the wake, IWs populate a wedge with half-angle,  $\beta$ , equal to  $\tan \beta = (V^2/C_n^2 - 1)^{-1/2}$ . Thus, the faster is the TC, the longer is the IW length and narrower will be the wedge filled by these IWs. If a TC is slow relative to the first mode,  $V/C_1 < 1$ , it does not generate IW wake behind, but it causes a localized upwelling, spatially confined to the area of the surface stress vorticity action, Figure 3c.

Baroclinic and barotropic components of the SSH anomalies are shown in Figure 4. For a fast TC, baroclinic SSH anomalies are formed by the IW wake, with a dominant contribution of the first mode. Barotropic SSH anomaly has a shape of the surface trough, with a depth much smaller than the baroclinic SSH anomaly contribution. This derives from considering *deep* ocean conditions. For shallow ocean conditions, relations between magnitudes of the barotropic and the baroclinic SSH anomalies could very well be opposite. For slow TC, with  $C_2 < V < C_1$ , baroclinic SSH anomaly is a composition of the surface trough caused by a non-resonant upwelling contribution from the first mode overlapped with *high-frequency* IW oscillations. Similar to fast TC, barotropic SSH anomalies for slow TC are significantly smaller than the baroclinic ones. The total



**Figure 3.** Two-dimensional field of vertical velocity described by 19 for (a, c) the first and (b, d) the second baroclinic modes generated by TC traveling with translation velocity (a, b)  $V = 5$  m/s and (c, d)  $V = 2$  m/s. Color bar in each of the plots indicates velocity in meter per second. TC = tropical cyclone.

SSH anomalies (sum of baroclinic and barotropic components) are shown in Figures 5a and 5b. Comparing magnitudes of SSH anomalies induced by fast and slow TC, one may find that the SSH anomalies are more marked for a slow TC.

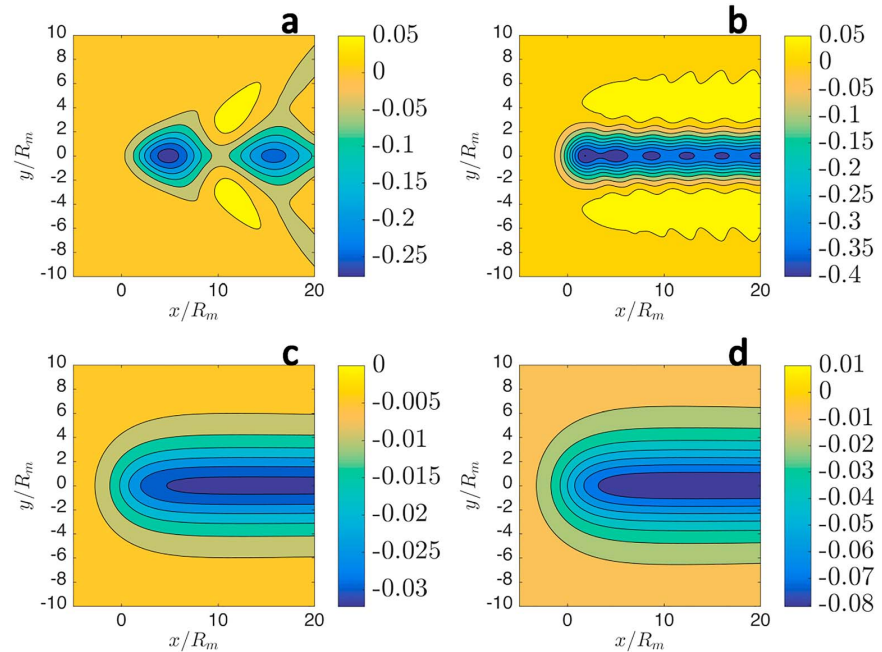
Once the SSH anomalies are found, the sea surface current field can be determined through solution of the integral momentum balance equation for the ML, equation (5), with prescribed *momentum source*—sum of the surface wind stress and the pressure gradient. Solution of this equation is straightforward and has the form similar to equation (12), where, however, the wind stress under the integral must be replaced by momentum source (equal to the right side of equation (5)). Then, with known ML depth, the surface currents,  $u_\beta$ , read  $u_\beta = M_\beta^w/h$ . Fields of the surface current for fast and slow TCs are shown in Figures 5c and 5d correspondingly. These fields are composed of superposition of the wind-driven currents and currents induced by the baroclinic and the barotropic motions.

The model SSH anomalies, scaled by  $u_m^2/g$ , as a function of dimensionless parameter  $R_m N_1/V$  characterizing the properties of the TC and the upper ocean stratification, are shown in Figure 6. An empirical relation  $gh_s/u_m^2 = 6.9 \times 10^{-6}(R_m N_1/V)$  suggested from satellite SSH measurements (see Figure 14 in Kudryavtsev et al., companion paper) is also shown. As found, the model is consistent with reported observations. Yet the model seems to indicate that dimensionless SSH anomalies are not fully self-similar on parameter  $R_m N_1/V$ .

#### 4.2. SST Wake

Figure 7 evidences the impact of the baroclinic wake on the 3D field of the ocean temperature, for a fast TC with  $V = 5$  m/s, Rossby number  $V/fR_m = 2$  and  $V > C_1 = 2.7$  m/s, and a slow TC with  $V = 2$  m/s, Rossby number less than 1,  $V/fR_m = 0.8$ , and  $C_1 > V > C_2 = 1.4$  m/s.

In both cases, the SST wake has a remarkable bias to the right side of the TC track, stronger for the faster TC condition. This can be attributed to resonant couplings between the rotation of surface winds and clockwise inertial currents, accelerated (respectively decelerated) on the right side (respectively left side). ML stirring and entrainment from below the thermocline are thus amplified (Huang & Oey, 2015; Price, 1981; Skillingstad et al., 2000).



**Figure 4.** (a, b) Baroclinic and (c, d) barotropic components of the sea surface height anomalies generated by TC traveling with translation velocity (a, c)  $V = 5$  m/s and (b, d)  $V = 2$  m/s. Color bar in each of the plots indicates SSH anomalies in meter. SSH = sea surface height.

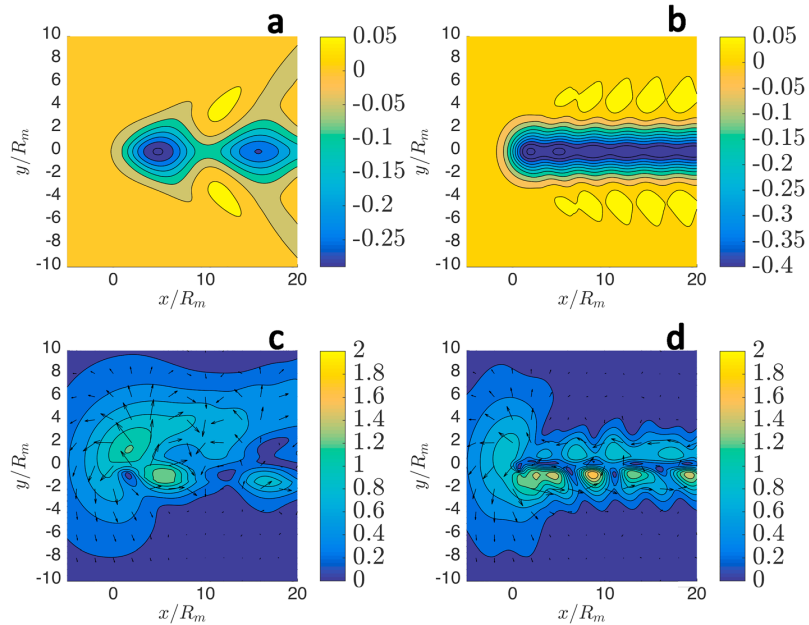
Vertical transects, along the TC track, reveal two stages of the SST wake development: a forced stage ( $x_1/R_m < 5$  for fast TC and  $x_1/R_m < 2$  for slow TC in Figures 7c and 7d correspondingly), characterizing deepening of the ML and its cooling due to heat fluxes caused by both the thermocline erosion (first term in equation (26), upper line) and upwelling (second term in equation (26), upper line), followed by a relaxation stage at larger  $x_1/R_m$  values. At this stage, turbulent mixing at the ML base ceases,  $Ri > Ri_{cr}$ . ML evolves keeping constant temperature, as surface heat fluxes are not taken into account, and its lower boundary undulates following quasi-inertial IW oscillations, equation (26)—lower line.

Vertical transects, across the TC track, Figures 7e and 7f, are related to the forced stage of the ocean response. The ML depth is controlled by the magnitude of the wind-driven transport, with a clear bias to the right of the track. The upwelling effect lifts up isotherms to impale the ML boundary, enhancing cooling of the upper layer.

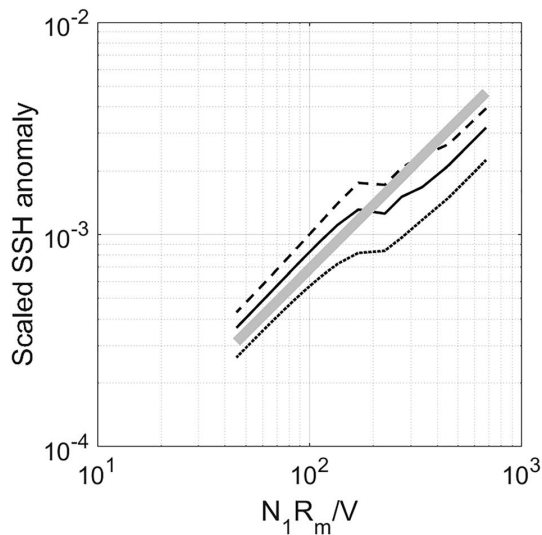
Model SST anomalies, with and without accounting for the upwelling impact, scaled by  $\langle \delta \theta_m \rangle = 2\Gamma u_m / (fN_1)^{1/2}$  are shown in Figure 8, as a function of the TC-Rossby number. As expected, taking into account upwelling remarkably impacts the SST anomalies, especially for low translation velocities. Effect of the upwelling on SST anomalies is also illustrated in Figure 18a below. The scaled SST anomalies are not self-similar on the TC Rossby number  $Ro \equiv V/(fR_m)$ , that is, model curves do not converge for different winds. Similar conclusions could be drawn from varying the radius of maximal wind speed, the shape parameter, and parameters related to the ocean stratification. Strictly, the model results do not fully support a self-similarity of the dimensionless anomalies on TC Rossby number, as initially suggested from the analysis of observations (Kudryavtsev et al., companion paper).

To first order, the empirical parametrization for the scaled SST anomalies (thick gray line in the right plot of Figure 8) does not perfectly compare with initial model simulations. Indeed, initial setups consider very idealized stratifications of the upper layer, with a constant temperature gradient up to the ocean surface. More likely, stratification of the upper top ocean shall be different, with rather uniform vertical temperature profiles, see, for example, Figure 9 from (Kudryavtsev et al., companion paper). As derived, equation (28) with 7,





**Figure 5.** (a, b) The total (sum of baroclinic and barotropic components) sea surface height anomalies and (c, d) surface currents generated by TC traveling with translation velocity (a, c)  $V = 5$  m/s and (b, d)  $V = 2$  m/s. Color bar in each of the plots indicates SSH anomalies in meter and surface current speed in meter per second correspondingly. Arrows indicate direction of the current velocity.



**Figure 6.** Dependence of dimensionless model SSH anomalies  $gh_s/u_m^2$  on parameter  $R_m N_1/V$  for different maximal wind speed: (dashed) 30, (solid) 50, and (dotted) 70 m/s. Thick gray line is empirical relation  $gh_s/u_m^2 = 6.9 \times 10^{-6} (R_m N_1/V)$  suggested by Kudryavtsev et al. (companion paper). TC parameters:  $R_m = 50$  km,  $1 \text{ m/s} < V < 15 \text{ m/s}$ . SSH = sea surface height; TC = tropical cyclone.

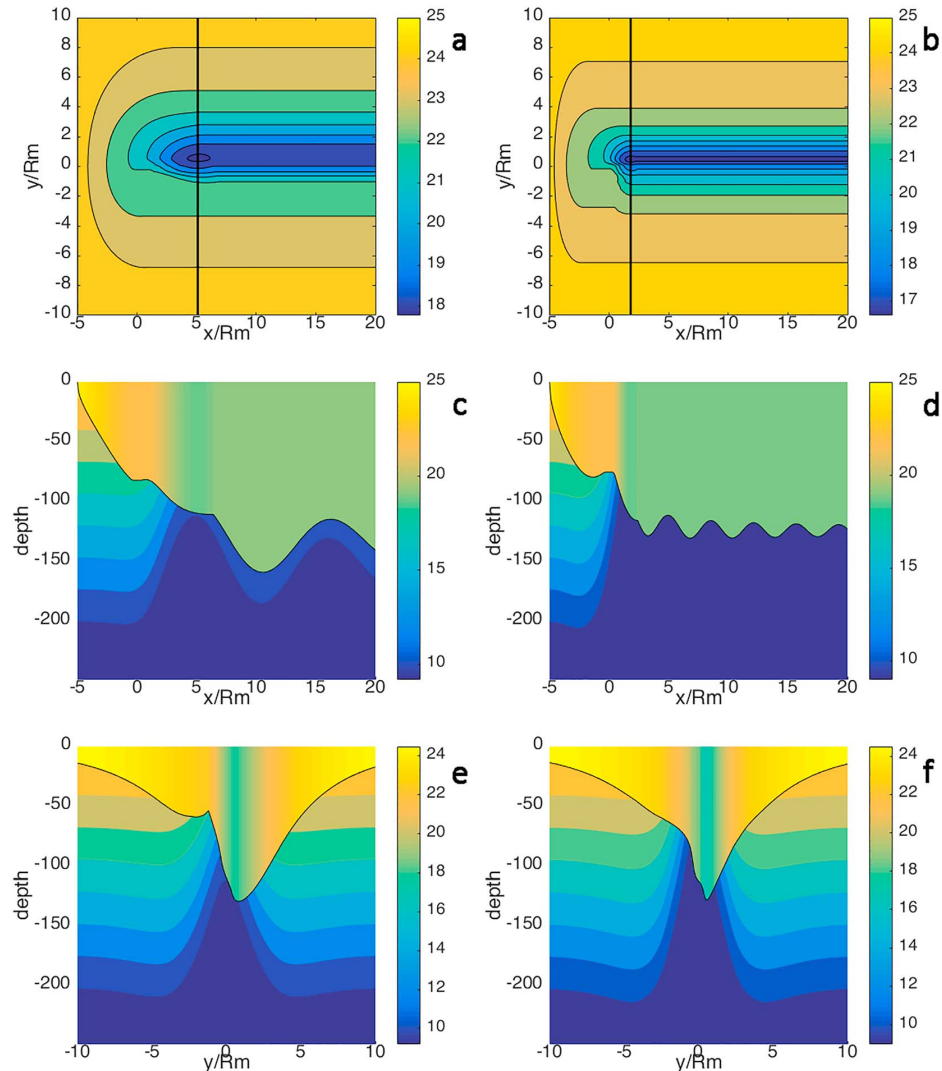
SST anomalies are very sensitive to this vertical distribution of the temperature. To illustrate this property, the prestorm stratification can be assumed as a composition of a constant temperature subsurface layer, with thickness,  $h_0$ , much smaller than the depth of seasonal thermocline,  $h_0 \ll d$ , superposed over a lower layer with constant temperature gradient. The averaged temperature gradient 7, defining an SST anomaly via 28, is then reduced by factor  $(1 - h_0^2/h^2)$ . As the sensitivity for the ML depth,  $h$ , to the temperature gradient is much weaker, power one fourth of the gradient, see equation (11), the SST anomalies in presence of an upper uniform layer can be simply estimated by multiplying the reference SST anomalies (for constant temperature gradient) by the factor  $(1 - h_0^2/h^2)$ . As obtained, Figure 8c, taking into account this uniform upper layer effect, largely widens the range of model estimates to encompass the empirical relation.

## 5. Simulations of Observations

In this section, simulations are performed to compare with satellite observations, SST, and SSH anomalies, reported in Kudryavtsev et al. (companion paper).

### 5.1. Input Parameters

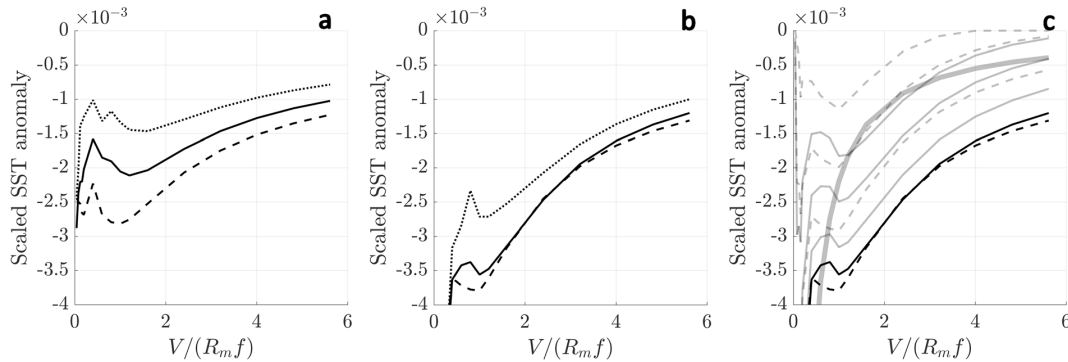
The best-track (BT) data describing key TC characteristics, that is, maximum wind speed and its radius, radii of given wind speed and translation velocity (derived from TC 6-hr position) for three TC were reported (Reul et al., 2017). To perform the model simulations, these input characteristics



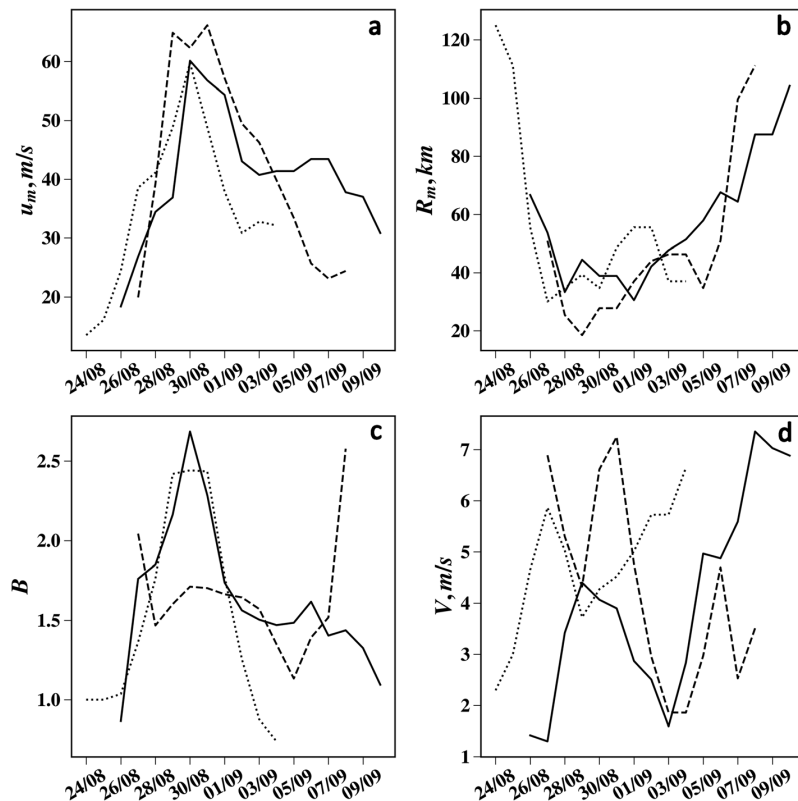
**Figure 7.** SST wake generated by (a, c, e) *fast* and (b, d, f) *slow* TC with translation velocity  $V = 5$  and  $V = 2$  m/s correspondingly. (a, b) SST fields (view from the top), (c, d)  $x_1 - x_3$  transects at  $x_2 = 0$ , and (e, f)  $x_2 - x_3$  transects at  $x_1$  indicated in the corresponding upper plots. SST = sea surface temperature.

are averaged over a 1-day sliding window, Figure 9. The BT radial wind speed distributions were then fitted to the model wind speed 29. The maximum wind speed,  $u_m$ , is fixed, and the two other model parameters are adjusted,  $R_m$  and  $B$  (see Figure 9), by minimizing the difference between the wind model and the radial velocity distribution averaged over TC sectors. Two examples of this procedure, demonstrating good agreement for the corresponding radial distributions of wind stress and its vorticity, are shown in Figure 10.

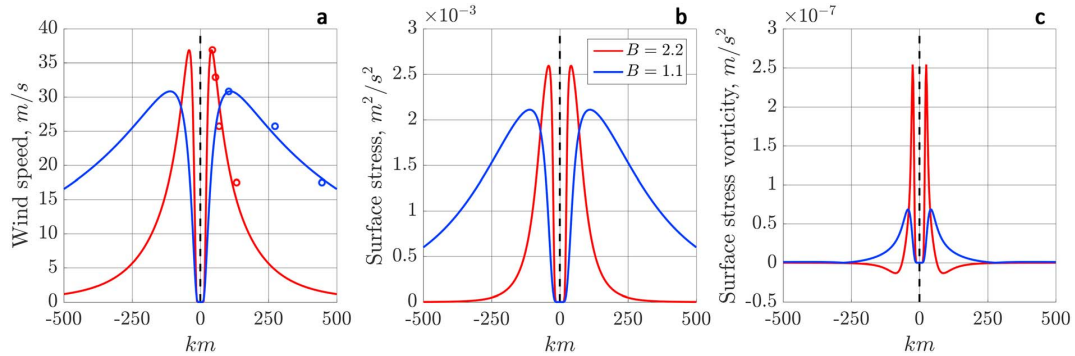
To specify the ocean stratification along the TC tracks, the WORLD OCEAN ATLAS 2013 version 2 (<https://www.nodc.noaa.gov/OC5/woa13/>) is used. As a comparative ground-truth data set, ISAS-15 gridded monthly fields of temperature and salinity based on ARGO profilers is also employed (Kołodziejczyk et al., 2017). For the evolution of the ML depth and SST anomalies, simulations are performed using *real*



**Figure 8.** Normalized SST anomalies,  $\hat{\delta\theta}_m = 2(\partial\theta_m/\Gamma)(fN_1)^{1/2}/u_m$ , as a function of TC Rossby number,  $V/(R_m f)$ , at different maximum wind speeds: (dashed) 30, (solid) 50, and (dotted) 70 m/s. Plots (a) and (b) show simulations when impact of upwelling on SST wake is either taken or not taken into account. Plot (c) illustrates the impact of a prestorm upper layer of uniform temperature with thickness 0 (thick lines), 20, 40, and 60 m (thin lines of the same style from down to top, respectively) on the SST anomalies at (dashed lines) 30 and (solid lines) 50 m/s. In these calculations  $R_m = 50$  km,  $f = 5 \times 10^{-5}$  1/s. Thick gray line in the right plot shows parametrization of the SST anomalies suggested from observations (Kudryavtsev et al., companion paper). SST = sea surface temperature; TC = tropical cyclone.



**Figure 9.** Parameters of TC wind field derived from BT data averaged over 1 day used in the model simulations: (a) maximum wind speed; (b) radius of max wind speed; (c) shape parameter  $B$  in model (3.16), and (d) translation velocity. Lines style: (dashed) Jimena, (dotted) Ignacio, and (solid) Kilo. BT = best-track; TC = tropical cyclone.

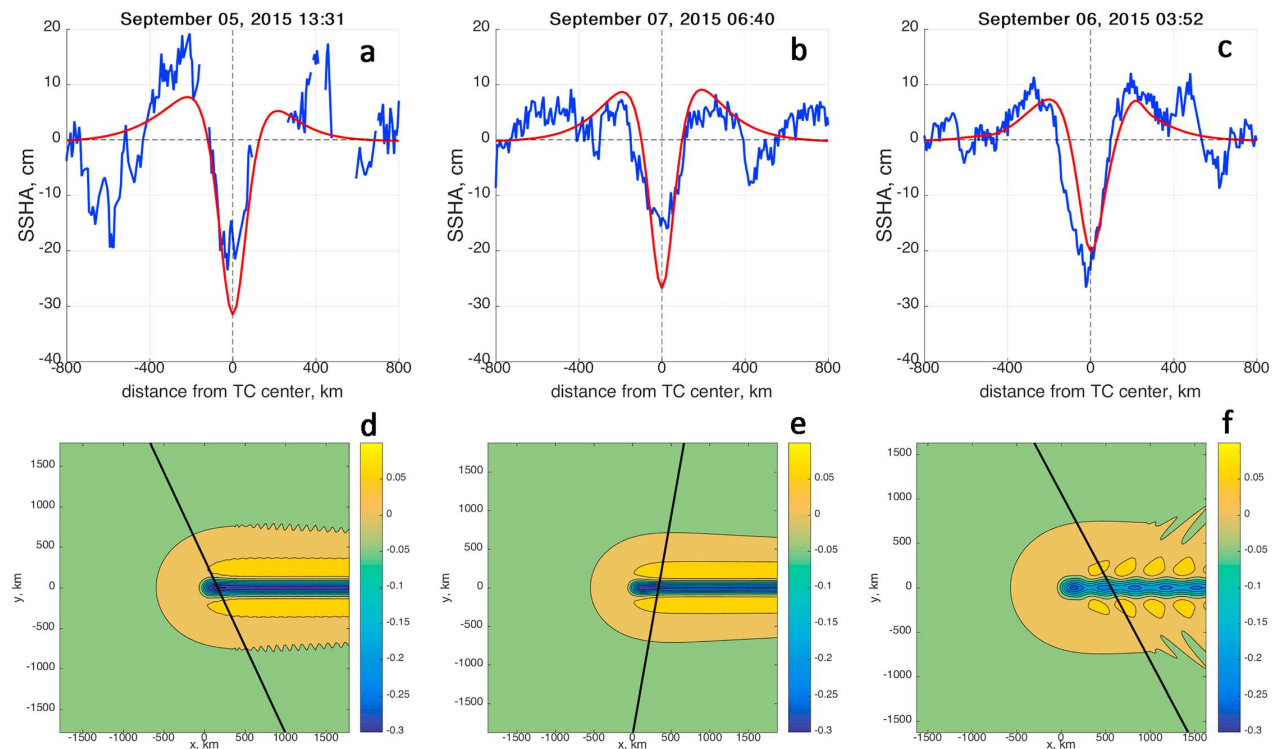


**Figure 10.** (a) Wind model (3.16) fitted to BT radial distribution of wind speed data shown by open circles. (b) Radial distribution of the surface wind stress corresponding to the wind speed profile shown in left plot. (c) Corresponding vorticity of the wind stress. BT = best-track.

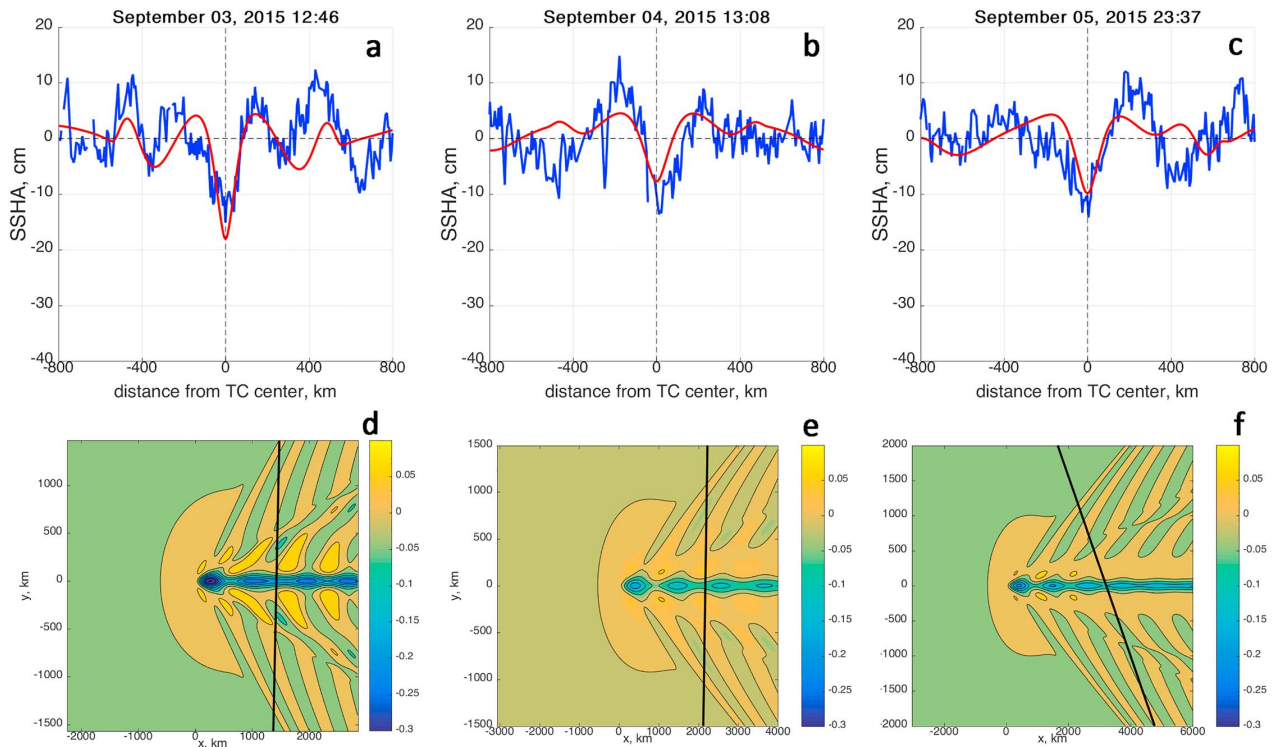
stratification. For the baroclinic responses, analytical solutions are considered, as described in section 3.1, with a three-layer approximation of the vertical stratification.

### 5.2. SSH Anomalies

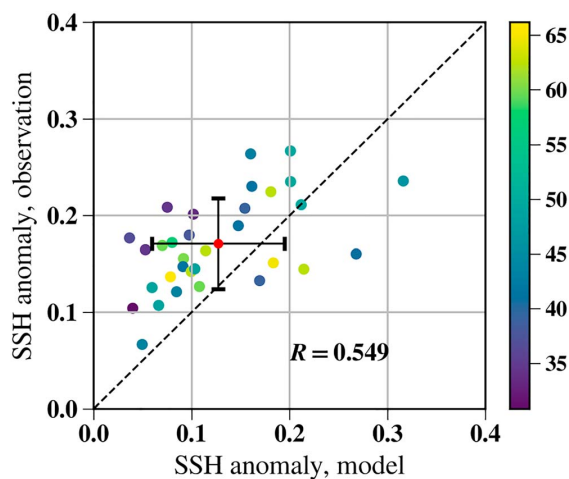
Some examples for the SSH anomalies, for slow,  $V < C_1$ , and fast,  $V > C_1$ , situations are shown in Figures 11 and 12. A slow TC does not generate wave-wake, except for the case shown in Figure 11f where the wave-wake results from the weak second mode. Accordingly, simulated SSH anomalies are shaped as a surface



**Figure 11.** Cases of *slow* TCs. (a, b, c) Observed (blue) and simulated (red) SSH anomalies along altimeter tracks for TC Jimena (a and c) and Kilo (b). (d, e, f) Corresponding model fields of the SSH anomalies with location of altimeter tracks (shown by black line), color bars indicate SSH anomalies in meter. Parameters of the TCs (from left to right):  $V = 1.9$ ,  $u_m = 46$  m/s,  $R_m = 46$  km,  $B = 1.6$ ;  $V = 1.6$  m/s,  $u_m = 41$  m/s,  $R_m = 48$  km,  $B = 1.5$ ;  $V = 3$  m/s,  $u_m = 49$  m/s,  $R_m = 44$  km,  $B = 1.6$ . SSH = sea surface height; TC = tropical cyclone.



**Figure 12.** Cases of fast TC. (a, b, c) Observed (blue) and simulated (red) SSH anomalies along altimeter tracks for TC Jimena. (d, e, f) Corresponding model fields of the SSH anomalies with location of altimeter tracks (shown by black line), color bars indicate SSH anomalies in meter. Parameters of the TC Jimena (from left to right):  $V = 4.3$  m/s,  $u_m = 65$  m/s,  $R_m = 19$  km,  $B = 1.6$ ;  $V = 7.3$  m/s,  $u_m = 66$  m/s,  $R_m = 28$  km,  $B = 1.7$ ;  $V = 6.6$  m/s,  $u_m = 6.2$  m/s,  $R_m = 28$  km,  $B = 1.7$ . SSH = sea surface height; TC = tropical cyclone.



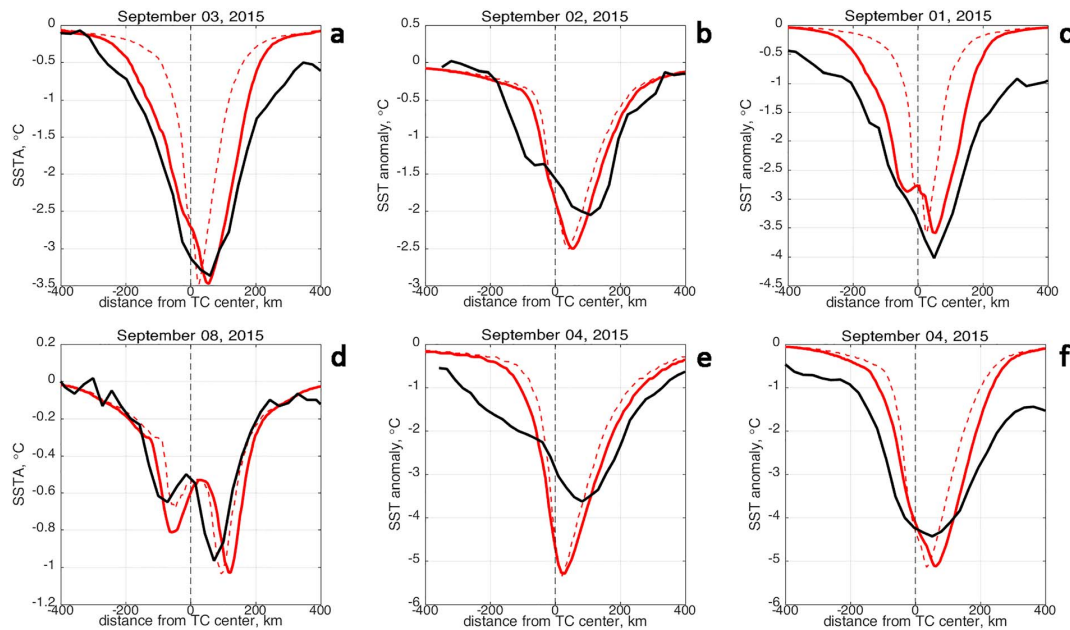
**Figure 13.** Scatterplot model versus observations. Color indicates wind speed, dashed line is one-to-one relation, red dot is mean value, and vertical bars are the std of the data in direction of each of the axis. SSH = sea surface height.

trench, mostly generated by baroclinic modes, with a weak contribution from the barotropic mode (see also Figure 4). At variance, a fast TC generates a spectrum of IW modes, and SSH anomalies combine ocean surface undulations with dominant contribution of the two first modes, Figures 12d to 12f. Simulated transects of the SSH anomalies display marked oscillations with main troughs around the TC track. For both slow and fast situations, model estimates are consistent with observations: the slower the TC, the deeper the SSH anomalies. Overall results are summarized in Figure 13. Given its simplicity, the model results remarkably reproduce observations, with a slight overall underestimation (about 5 cm). It should also be noticed that some remarkable differences between simulations and observations (as, e.g., in Figure 11a) result from the fact that SSH anomalies not only reflect immediate changes in the wind forcing but also anomalies associated with earlier forcing events, as well as mesoscale eddy activity. As such SSH measurements may appear *contaminated* by the SSH anomalies left by other TC traveled in this area before, as well as probably some other anomalies related to the mesoscale ocean dynamics.

### 5.3. SST Anomalies

SST wakes, developing behind TCs, are shown in Figure 14. Compared to observations, predicted SST wakes are generally narrower. Such a





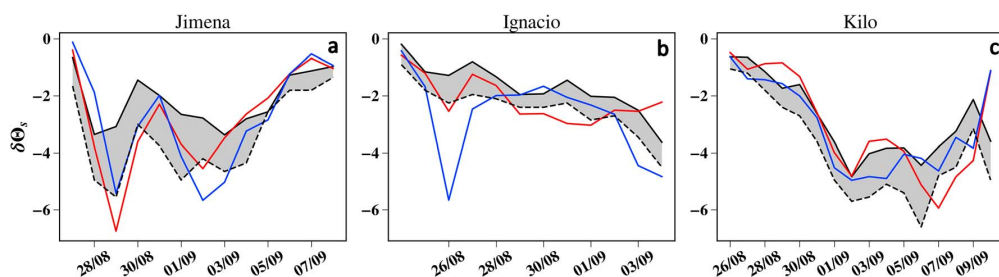
**Figure 14.** Comparisons of (black lines) observed and (red lines) modeled transects of the SST wakes just behind TCs (a, d) Jimena, (b, e) Ignacio, and (c, f) Kilo. Dash red lines correspond to model 28, solid red lines are model results accounting for the cross-track advection, equation (33) with 34. SST = sea surface temperature; TC = tropical cyclone.

discrepancy may result from a too idealized wind field description and from the broadening induced by the resolution of the satellite observations.

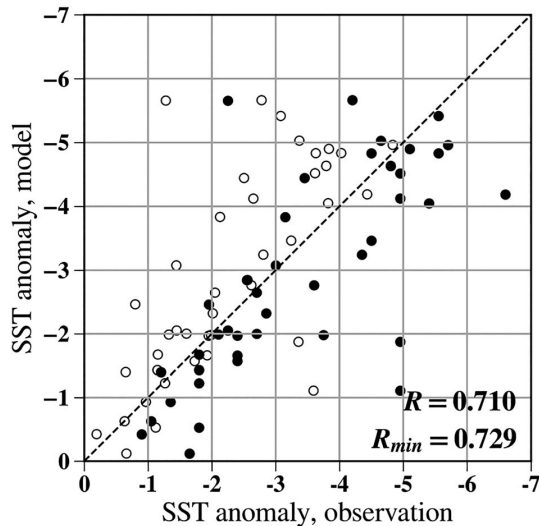
The role of advective transport of the SST anomalies, in the TC cross-track direction, can also be considered. The magnitude of TC-induced currents is typically much smaller than the TC translation velocity (linearized heat balance; equation (27)), but strong SST cross-track gradient may be large. To assess this contribution, equation (27) is thus modified as

$$\partial \theta_m / \partial x_1 + (u_2 / V) \partial \theta_m / \partial x_2 = -1 / 2 \bar{\Gamma} w_e / V \quad (32)$$

where  $u_2$  is the cross-track component of the surface current velocity in the ML, that is,  $u_2 = M_2^w / h$  with  $M_2^w$  defined by equation (5) (its solution is discussed in the end of section 4.1). Equation (32) rewrites as



**Figure 15.** Along TCs track evolution of the SST anomalies for TCs (a) Jimena, (b) Ignacio, and (c) Kilo. Line style: (black solid) observed SST anomalies, (dash solid) minimal values of observed SST anomalies, (red) model simulations for WORLD OCEAN ATLAS stratification, and (blue) for ISAS-15 stratification data. SST = sea surface temperature; TCs = tropical cyclones.



**Figure 16.** Observed SST anomalies versus model simulations for ISAS-15 stratification data. Dash line indicates one-to-one relation. Open circles indicate mean SST anomalies, and filled circles indicate minimal values of SST anomalies SST = sea surface temperature.

$$d\theta_m/dx_1 = -1/2\bar{\Gamma}w_e/V \quad (33)$$

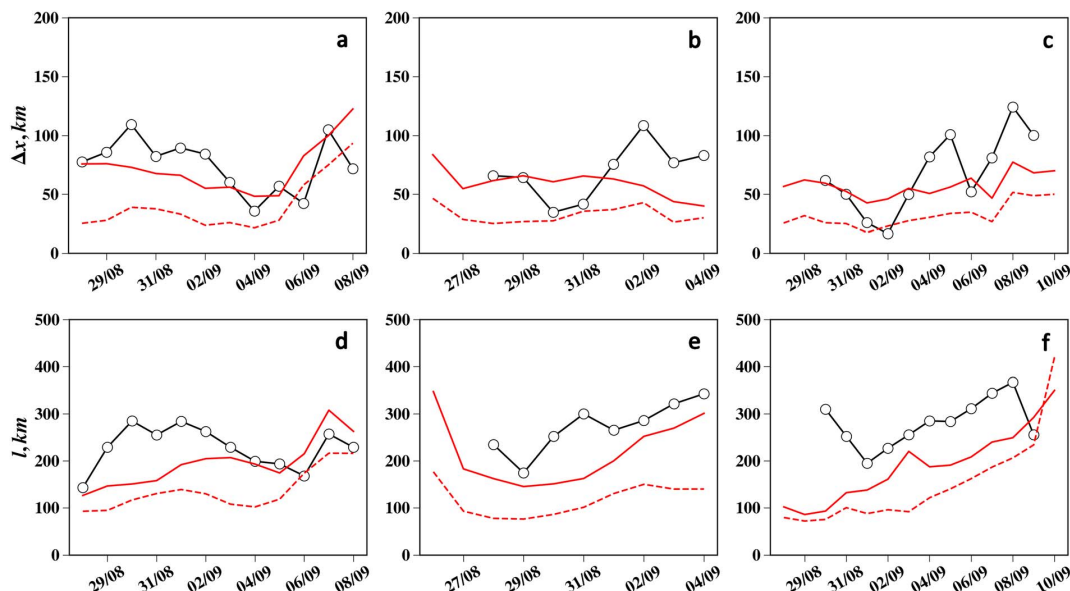
where  $d/dx_1$  stands for total derivative along characteristics

$$x_2(x_1) = x_2(x_{10}) + \int_{x_{10}}^{x_1} (u_2/V) dx_1 \quad (34)$$

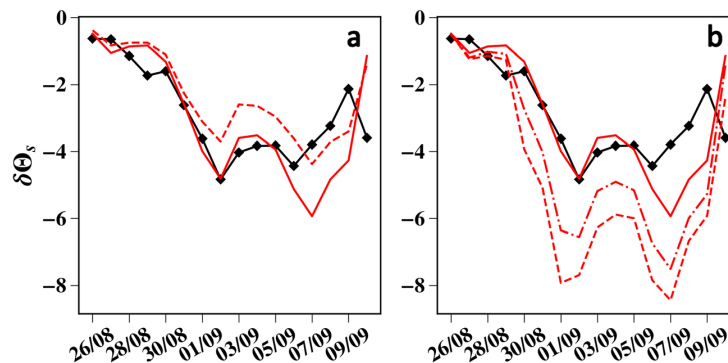
Since 33 is similar to 27, the solution 28 for the SST anomalies,  $\delta\theta_s$ , should now be treated as SST field along a family of characteristics 34. Solution of equation (33) with 34 is shown in Figure 14. The cross-track advection effectively widens the SST wake, with a stronger impact for slower TCs. This effect improves comparisons between model and observations, though some discrepancies still remain.

Figure 15 shows simulations of the along track evolution of the observed SST anomalies for the interior stratification data provided by the WORLD OCEAN ATLAS 2013 version 2, and, for comparative purposes, estimates provided by ISAS-15. Model simulations for both stratification data sets are in general very similar, justifying that shape of stratification transects provided by both data sets is similar (see Figure 8 in Kudryavtsev et al., companion paper, for ISAS-15 transects). In general, SST anomalies follows the main trend and oscillations of observed SST anomalies (both the mean and the minimal values) caused by changes of variety of the TCs parameters (wind speed, translation velocity, and radius) and environment parameters (ocean stratification and Coriolis parameter) in wide range. For TCs Kilo and Jimena, comparisons are quite remarkable, but

for TC Ignacio, some deviations between the model simulations using ISAS-15 and observations around 26 August can be noticed. Simulations with stratification evaluated from the World Ocean Atlas attenuate departures between model and observations. Following ISAS-15 data, in this particular case, the TC traveled over an ocean area with the largest values for both the temperature gradients and Brunt-Väisälä frequencies



**Figure 17.** Along TCs track evolution of (a–c) offset and (d–f) width of the SST anomalies for TCs (a, d) Jimena, (b, e) Ignacio, and (c, f) Kilo. Line style: (black line with open circles) observations and (red dashed and red solid) model simulations without (equation 28) and with (equation 33 with 34) cross-track advection by wind-driven current, respectively. SST = sea surface temperature; TCs = tropical cyclones.



**Figure 18.** Along-track evolution of SST anomalies for TC Kilo; observations are shown by black lines with dots. (a) Full model (red solid) and model without accounting for the upwelling (red dashed). (b) Model simulations with different drag coefficients: (red solid) suggested  $C_d$ , (red dashed)  $C_d$  corresponding to the Charnock relation extrapolated to high winds, (red dash-dotted), and  $C_d$  corresponding to the Charnock relation but saturated at  $C_d = 2 \times 10^{-3}$ . SST = sea surface temperature; TC = tropical cyclone.

in the seasonal pycnocline, see Figures 8 and 10 from Kudryavtsev et al., companion paper. According to the ISAS-15 data (see Figure 8 from the companion paper), these regions are also located in the vicinity of the ocean surface frontal zone. Actual bias of this frontal zone from an observed point may then lead to significant deviation of actual stratification from ISAS-15 data and thus to lower model estimates of the SST anomaly.

Scatterplot, Figure 16, exhibits quantitative correspondence of the model to observations with rather high level of correlation, especially for the model correlation with minimum of the SST in the wake.

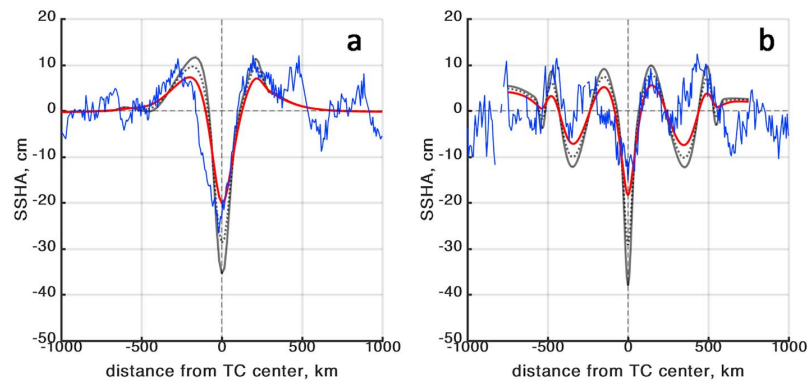
Model simulations of the TC-track offsets and widths of the SST anomalies underestimate the observation, Figure 17. Considering cross-track advection, solutions 33 with 34, improves the model performances. However, the cold wake width is still largely underestimated. It may be speculated that more realistic TC wind fields, with possible marked asymmetries between left and right sectors, could improve the comparisons.

#### 5.4. Influence of Drag Coefficient Parametrization

Given the proposed model simplicity, we can readily explore the idea that SST and SSH surface expressions, along with knowledge of the initial conditions, can be used to test and possibly infer the drag coefficient relationship with wind speed. Already, the suggested parametrization is consistent with reported observations (Powell et al., 2003) and altimeter-derived SSH anomalies, see equation (18) and Figure 15 from Kudryavtsev et al., companion paper. The present model can as well be used to test the sensitivity to drag coefficient,  $C_d$ , alternatively defined as (a)  $C_d$  prescribed by the Charnock relation extrapolated to high winds,  $C_{d1}$ ; and (b)  $C_d$  saturating at a given threshold value, for example,  $C_d^{\text{st}} = 2 \times 10^{-3}$ , that is,  $C_{d2} = \min(C_{d1}, C_d^{\text{st}})$ . Results are reported in Figure 18b. As expected, model simulations using different drag coefficients result in larger SST anomalies, largely overestimating the observations. This is consistent with simulations reported by Sanford et al. (2007).

Similar effects apply to SSH anomalies; two examples for slow and fast TCs are shown in Figure 19. While more investigations are certainly needed, observations and simulations highlight the potential to combine both SST and SSH wake signatures to provide necessary information to improve understandings and quantify momentum exchanges at the sea surface under extreme conditions. As presented, the present results suggest a relative saturation of the surface stress, that is, reduction of the drag coefficient, to bring model estimates in reasonable agreement with observations.

To note, the present model setup can also be used to test impacts of asymmetrical wave fields, that is, associated with possible trapping and enhancement of large surface gravity waves in the front-right storm quadrant (Kudryavtsev et al., 2015; Young, 2006). Indeed, wave-induced mixing, including breaking impacts and



**Figure 19.** Examples of (blue lines) observations and model simulations of the SSH anomalies for (a) slow and (b) fast TC, shown before in Figures 11 and 12. Red lines are model simulations with basic drag coefficient, solid and dotted black lines are model simulations with  $C_d$  defined by the Charnock relation extrapolated to high winds and  $C_d$  defined by the Charnock relation but saturated at  $C_d = 2 \times 10^{-3}$  correspondingly. SSH = sea surface height; TC = tropical cyclone.

the combined action of wave-induced drift and wind-induced shear (e.g., Reichl et al., 2016), certainly contributes to the turbulent kinetic energy (TKE) balance equation. Yet, within the TC core, the expected ML deepening, equation (11), may largely exceed the depth strongly influenced by wave breaking and nonbreaking waves that is of the order  $d_w \approx 1/(3k_p)$ , with  $k_p$  the spectral peak wave number (Babanin & Haus, 2009; Kudryavtsev et al., 2008). In that case, wave-induced mixing should not impact the SST anomalies induced by TC, as recently reported by Stoney et al. (2017). Moreover, the present results suggest a relative saturation of the surface stress. As such, while existing, local particular influences of surface waves and asymmetrical wave fields may not be strongly distinguishable.

On the other hand, nonlocal surface wave impacts may well be considered. Indeed, as discussed in paragraph 4.2, SST anomalies are very sensitive to the vertical distribution of the prestorm temperature. Fore-runner waves, traveling with group velocities largely exceeding the TC-translation velocity, will disturb (a priori) quite upper ocean areas. The prestorm stratification shall then be modified, before the TC-core arrival time. This enhanced nonbreaking wave-induced mixing process is thus nonlocal. The depth,  $d_w \approx 1/(3k_p)$ , influenced by surface waves (Babanin & Haus, 2009) shall thus be related to TC characteristics at previous time steps. Such a nonlocal impact can then be suggested to enter the correcting factor  $(1-h_0^2/h^2)$ , with a thickness,  $h_0 = d_w$ , now more directly related to wave parameters, for example, peak frequency/wave number (see, e.g., equations (6) and (15) for trapped waves in Kudryavtsev et al., 2015).

## 6. Conclusion

In this paper, the previously proposed framework, used to help interpret satellite observations (Kudryavtsev et al., companion paper), is extended to provide a more complete analytical description of ocean responses to moving TC. As developed, the proposed model helps detail the wind-driven current field, the upper ML cooling and its evolution, with the associated space-time variability of the pycnocline caused by TC-induced baroclinic motions. While more complete, the goal, and main purpose of the present paper, is to best bypass the computational burden of advanced numerical simulations. Simplified analytical solutions are thus elaborated. As tested, the derived solutions are very fast to compute, and help to assess, on quantitatively correct levels, the sensitivity of ocean responses to moving TCs, specified by various environmental (wind forcing and ocean stratification) conditions.

To go beyond classical 1D ML models, the suggested solutions build on previously developed linear models (Geisler, 1970; Orlanski & Polinsky, 1983) of ocean response to transient events. The final model is then represented by a set of analytical solutions, describing baroclinic and barotropic responses, and resulting SST anomalies. The baroclinic response is decoupled from the ML, in that sense that disturbances of the



upper ocean stratification caused by the ML development do not strongly impact dynamics of baroclinic modes. This assumption is accurate to the smallness of the parameter  $1/2(N_1 h/C)^2$  utilizing Brunt-Väisälä frequency in the seasonal pycnocline, the ML depth, and phase velocity of long internal waves. Accordingly, the baroclinic response can be estimated for prestorm ocean stratification. Simple analytical solutions are then derived by fitting the prestorm stratification with a three-layer description, to take into account seasonal and main thermocline, and abyssal characteristics. To the contrary, the ML is strongly coupled with these interior motions, through the TC-induced upwelling effect that affects the entrainment velocity 26. The resulting ML temperature is further strongly dependent on the integrated temperature gradient in the upper layer, equation (27) with 7.

Comparisons with observations proved the model capable to reproduce observed shape and magnitudes of the SSH and SST anomalies on quantitative levels. Supporting Yablonsky and Ginis (2009), the model simulations provide solid evidences about the key role of TC-induced upwelling, caused by vertical baroclinic motions, to control the ML cooling, see Figure 18a. This is especially effective for slowly translating TCs. Cross-track advection by wind-driven currents, though small compared with TC translation velocity, further remarkably contributes to broaden the shape of the SST wake, for example, Figure 17. These quantitative agreements are obtained for various TC characteristics, ocean stratification conditions, and Coriolis parameters. It is worthy to note that the proposed model does not possess special/dedicated tuning constants, besides the critical Richardson number, equation (8). For the simulations, it is here set as  $R_{cr} = 0.6$ , following previous studies (Price, 1981). Yet, the ML depth (11) solely depends on this critical  $R_{cr}$  in power one fourth, and changing this tuning parameter shall not impact the results significantly.

Given the proposed model simplicity, we further test the idea that SST and SSH surface expressions, along with knowledge of the initial conditions, can be used to test and possibly infer the drag coefficient relationship with wind speed. As presented, the obtained results suggest a relative saturation of the surface stress under extreme conditions, that is, reduction of the drag coefficient, bringing model estimates in reasonable agreement with observations. Since a nearly saturated surface stress may occur under high winds, it can be speculated that peculiarities of the resulting wave field may not particularly impact the upper ocean response within the TC-core area. More plausible, outrunning energetic surface waves, related to TC characteristics at previous time steps, may have important nonlocal contributions. As precursors to the forthcoming TC intense forcing, surface waves can destabilize the preexisting upper layer stratification. As such, surface waves may effectively, but through a so-called nonlocal effect, contribute to modulate the intensity and extent of TC-cooling wakes.

More investigations are certainly needed, but observations and simulations already highlight the potential to combine both SST and SSH wake signatures. In that context, analysis can certainly already build on the present-day altimeter constellation: up to six satellite altimeters are today available. It can further be anticipated that next National Aeronautics and Space Administration's Surface Water and Ocean Topography mission (Fu et al., 2012), with unprecedented 2D altimeter mapping capabilities, promises to greatly improve the analysis of TC-induced SSH wake: this shall be highly beneficial to improve our knowledge of the air-sea exchanges under extreme conditions.

Finally, given the complexity of isolating the governing physical processes in full-physics coupled models, the proposed developments can be introduced as a computational module into an atmospheric numerical model of TC evolution coupled with the ocean through the resulting surface enthalpy fluxes (e.g., Yablonsky et al., 2015). Furthermore, given its effectiveness and low computational burden, the model can be rapidly evaluated under a wide range of possible initial conditions, to best sample the whole uncertainty space to apply inverse approaches (e.g., Sraj et al., 2013).

## Appendix A

### A1. Main Equations

Solutions of equation (13) for each layer of the three-layer stratification description, satisfying dynamic and kinematic boundary conditions, read



$$\hat{w}_1(x_3)/\hat{w}_h = \cos[N_1(x_3 + h)/c] + \frac{\cos(\Sigma + \varphi) - n \cos(\delta - \varphi)}{\sin(\Sigma + \varphi) - n \sin(\delta - \varphi)} \sin[N_1(x_3 + h)/c] \quad (\text{A.1})$$

$$\hat{w}_2(x_3)/\hat{w}_h = \frac{2N_1/(N_1 + N_2)}{\sin(\Sigma + \varphi) - n \sin(\delta - \varphi)} \sin[N_2(D + x_3)/c + \varphi] \quad (\text{A.2})$$

$$\hat{w}_3(x_3)/\hat{w}_2(-D) = \frac{H + x_3}{H - D} \quad (\text{A.3})$$

where

$$\hat{w}_h = -\frac{\hat{F}}{\Omega^2 - f^2} \left[ 1 + (N_1 h/c) \frac{\cos(\Sigma + \varphi) - n \cos(\delta - \varphi)}{\sin(\Sigma + \varphi) - n \sin(\delta - \varphi)} + \frac{1}{2} (N_1 h/c)^2 \right]^{-1} \quad (\text{A.4})$$

is the vertical velocity at the ML base,  $c = \sqrt{\Omega^2 - f^2}/k$  is a variable with dimension of a velocity,  $n = (N_1 - N_2)/(N_1 + N_2)$ , and

$$\begin{aligned} \Sigma &= N_1(d - h)/c + N_2(D - d)/c \\ \delta &= N_1(d - h)/c - N_2(D - d)/c \\ \sin \varphi &= \frac{N_2(H - D)/c}{\sqrt{1 + [N_2(H - D)/c]^2}} \end{aligned} \quad (\text{A.5})$$

To derive A.4,  $g\Delta\rho/\rho$  is evaluated as  $g\Delta\rho/\rho = -N_1^2 h/2$ .

In the open ocean conditions, typical values of  $N_1$  are of order  $N_1 \propto 10^{-2} \text{ s}^{-1}$  or less, and the ML depth is of order  $h \propto 10$  to  $10^2 \text{ m}$ . If  $c$  corresponds to the phase velocity of long IWs (which is about  $c \approx 3 \text{ m/s}$ ), then the magnitude of the parameter  $N_1 h/c$  is in the range  $N_1 h/c = 3 \times (10^{-2} \text{ to } 10^{-1})$ , that is, can be considered small. To the smallness accuracy of the term  $1/2(N_1 h/c)^2 \ll 1$ , relationships A1 and A.2 can be rewritten as

$$\hat{w}_1(x_3)/\hat{w}_0 = \cos[N_1 x_3/c] + \frac{\cos(\Sigma_0 + \varphi) - n \cos(\delta_0 - \varphi)}{\Delta} \sin[N_1 x_3/c] \quad (\text{A.6})$$

$$\hat{w}_2(x_3)/\hat{w}_0 = \frac{2N_1/(N_1 + N_2)}{\Delta} \sin[N_2(D + x_3)/c + \varphi] \quad (\text{A.7})$$

where

$$\hat{w}_0 = -\frac{\hat{F}}{\Omega^2 - f^2} \quad (\text{A.8})$$

and

$$\begin{aligned} \Sigma_0 &= N_1 d/c + N_2(D - d)/c \\ \delta_0 &= N_1 d/c - N_2(D - d)/c \\ \Delta &= \sin(\Sigma_0 + \varphi) - n \sin(\delta_0 - \varphi) \end{aligned} \quad (\text{A.9})$$

These relations do not explicitly depend on either  $h$  or  $g\Delta\rho/\rho$ , demonstrating that for the considered conditions, the ML does not impact the baroclinic motions in the interior layer.

## A2. Solutions

Relations (A.6) and (A.7) display a singularity in the vicinity of  $C_n$ , subsequent solution of equation

$$\Delta \equiv \sin \left[ \frac{N_1 d}{C_n} + \frac{N_2(D - d)}{C_n} + \varphi \right] - \frac{N_1 - N_2}{N_1 + N_2} \sin \left[ \frac{N_1 d}{C_n} - \frac{N_2(D - d)}{C_n} - \varphi \right] = 0 \quad (\text{A.10})$$

Solution of this equation defines the phase velocity of IWs,  $C_n$ , for the different modes,  $n$ , for given parameters of the ocean stratification.

Once equation (A.10) is solved and the phase velocity of IW defined, denominator in A.6 and A.7 can be expanded to the first order of  $(\Omega - \Omega_0)$ , that is, around the resonance curve  $\Omega_0 = \sqrt{f^2 + k^2 C_n^2}$ :

$$\begin{aligned} \Delta(\Omega) &= \frac{\partial \Delta}{\partial C_n} \frac{\partial C_n}{\partial \Omega} (\Omega - \Omega_0) \\ &= -\frac{1}{2(\Omega^2 - f^2)} [\Sigma_0 \cos(\Sigma_0 + \varphi_n) - n \delta_0 \cos(\delta_0 - \varphi_n)] (\Omega^2 - \Omega_0^2) \end{aligned} \quad (\text{A.11})$$

In this case, expressions (A.6) and (A.7) for the vertical velocity read

$$w_n(\mathbf{x}, z) = \begin{cases} a_{1n} W_n(\mathbf{x}) \sin[N_1 x_3 / C_n], & \text{at } -d < x_3 < 0 \\ a_{2n} W_n(\mathbf{x}) \sin[N_2 (D + x_3) / C_n + \varphi_n], & \text{at } -D < x_3 < -d \end{cases} \quad (\text{A.12})$$

where  $a_{1n}$  and  $a_{2n}$  are dimensionless vertical velocity amplitudes depending on stratification as

$$\begin{aligned} a_{1n} &= -\frac{2C_n}{\Sigma_0} \left[ \frac{\cos(\Sigma_0 + \varphi_n) - n \cos(\delta_0 - \varphi_n)}{\cos(\Sigma_0 + \varphi) - n(\delta_0 / \Sigma_0) \cos(\delta_0 - \varphi)} \right] \\ a_{2n} &= -\frac{2C_n}{\Sigma_0} \left[ \frac{2N_1 / (N_1 + N_2)}{\cos(\Sigma_0 + \varphi_n) - n(\delta_0 / \Sigma_0) \cos(\delta_0 - \varphi_n)} \right] \end{aligned} \quad (\text{A.13})$$

Parameters  $\Sigma_0$  and  $\delta_0$  are given by A.9 with  $c$  equal to IW phase velocity,  $C_n$ , and  $W_n(\mathbf{x})$  is a function defining 2D field of vertical velocity for  $n$  mode, which in Fourier space reads

$$\widehat{W}_n(k_1, k_2) = -\frac{\widehat{F}(k_1, k_2)}{\Omega^2 - \Omega_0^2} = -\frac{\widehat{F}(k_1, k_2)}{C_n^2} \left[ \frac{1}{k_1^2 (V^2 / C_n^2 - 1) - \kappa^2 - k_2^2} \right] \quad (\text{A.14})$$

where  $\kappa = f / C_n$  is the inverse baroclinic radius of deformation. Expression (A.14) corresponds to the classical equation for vertical velocity in a two-layer model with constant density at each of the layer (Geisler, 1970, his equation (17)). Inverse Fourier transformation of A.14 gives the vertical velocity in physical space that appeared in A.12

$$W_n(x_1, x_2) = -\frac{1}{4\pi^2 C_n^2} \iint \frac{\widehat{F}(k_1, k_2)}{(V^2 / C_n^2 - 1) k_1^2 - k_2^2 - \kappa^2} \exp(ik_\beta x_\beta) dk_1 dk_2 \quad (\text{A.15})$$

If the translation velocity of TC is fast enough and satisfies the condition  $V > C_n$ , then A.15 has a singularity around the resonant curve in the wave number space

$$k_1^2 (V^2 / C_n^2 - 1) - k_2^2 - \kappa^2 = 0. \quad (\text{A.16})$$

Analytical solution for this case is given by Geisler (1970, his equation (30)), which can be adopted. However, we suggest a slightly different form that can easily be implemented for numerical calculations using fast Fourier transform.

To simplify A.15, we suppose the main contribution to integral in A.15 to arise in the vicinity of a resonant curve in the wave number space:

$$\begin{aligned} k_1 &= \pm k_{10} \\ k_{10} &= \sqrt{(k_2^2 + \kappa^2) / (V^2 / C_n^2 - 1)} \end{aligned} \quad (\text{A.17})$$

After substitution of  $k_1 = \pm k_{10} + \Delta k_1$  in A.15 and accounting for

$$\int [\exp(i\Delta k_1 x_1) / \Delta k_1] d\Delta k_1 = i\pi \quad (\text{A.18})$$

we arrive at



$$W(x_1, x_2) = -\frac{i}{4\pi(V^2 - C_n^2)} \int \left( \frac{\hat{F}(k_{01}, k_2) e^{ik_{01}x_1}}{k_{01}} \right) e^{ik_2x_2} dk_2 \quad (\text{A.19})$$

Using by definition

$$\hat{F}(k_{01}, k_2) = \int \hat{F}(x_1, k_2) e^{-ik_{01}x_1} dx_1' \quad (\text{A.20})$$

and introducing the radiation conditions (Lighthill, 1967), stating in our case (TC moves in direction opposite to the  $-x$  axis) that disturbances generated by a point source located at  $x_1 = x_1'$  can exclusively be observed at  $x_1 > x_1'$ , we finally get

$$W(x_1, x_2) = \frac{1}{2\pi(V^2 - C_n^2)} \int_{-\infty}^x dx_1' \int_{-\infty}^{\infty} \hat{F}(x_1', k_2) \frac{\sin(k_{01}(x_1 - x_1'))}{k_{01}} e^{ik_2x_2} dk_2 \quad (\text{A.21})$$

where  $\hat{F}(x_1', k_2)$  is  $k_2$  Fourier transform of wind stress source for given  $x_1'$ , and the factor 2 is introduced to preserve the total energy of the source. Only the real part of A.21 must be taken into account.

If the TC is slow, that is,  $V < C_n$ , then relation (A.15) is not singular, and hence, the vertical velocity can be found directly as Fourier transform of A.15.

Solutions A.21 for fast TC,  $V > C_n$ , and A.15 for slow TC,  $V < C_n$ , provide, together with A.10 and A.12, the full description of baroclinic vertical motions caused by the TC.

#### Acknowledgments

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# Supplement material for the wake analysis: chapter 6.

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Table F.1: time and space variability of the parameters from the ocean structure, with  $\Delta \overline{N}_1 = |\overline{N}_{1argo} - \overline{N}_{1ISAS}|$ .

| temporal variability                              | AT   | EP   | WP   | SH   |
|---------------------------------------------------|------|------|------|------|
| $\Delta \overline{N}_1 : (\times 10^{-3} s^{-1})$ | 1.7  | 1.2  | 2.3  | 0.8  |
| spatial variability                               | AT   | EP   | WP   | SH   |
| $\sigma_{N_1} : (\times 10^{-3} s^{-1})$          | 2.7  | 4.3  | 2.7  | 1.9  |
| $\sigma_{MLD} : (m)$                              | 14.0 | 16.3 | 16.1 | 14.0 |
| $\sigma_{c_1} : (m.s^{-1})$                       | 0.47 | 0.81 | 0.80 | 0.60 |

Table F.2: occurrence of particular mixed layer features

|                     | AT | EP | WP | SH |
|---------------------|----|----|----|----|
| Barrier Layers      | 3  | 6  | 3  | 1  |
| Slim ML ( $< 20m$ ) | 5  | 8  | 16 | 5  |
| Deep ML ( $> 40m$ ) | 8  | 16 | 11 | 4  |
| Medium ML           | 21 | 22 | 18 | 5  |

Table F.3: Evolution of SSTA along latitude bands.

|                   | [0°-10°] N/S | [10°-20°] N/S | [20°-30°] N/S | [30°-40°] N/S |
|-------------------|--------------|---------------|---------------|---------------|
| $\overline{SSTA}$ | 0.93°C       | 1.47°C        | 2.12°C        | 1.9°C         |
| $\overline{MLD}$  | 47.0 m       | 33.0 m        | 29.5 m        | 22.4 m        |



Table F.4: SAR-derived maximum wind speed averaged for each mixed layer depth band and for barrier layer cases.

|                                                  | 0m-20m | 20m-40m | 40m-60m | 60m-80m | 80m-100m | BL |
|--------------------------------------------------|--------|---------|---------|---------|----------|----|
| $\overline{V_{max}} \text{ (} m.s^{-1} \text{)}$ | 42     | 44      | 48      | 55      | 63       | 59 |



**Titre :** Mélange et anomalies de surface de la mer dans le sillage inertiel des cyclones tropicaux : processus et contribution des données satellites micro-ondes.

**Mots clés :** Cyclones Tropicaux, sillage froid inertiel, lois d'échelles, satellites MW, Argo.

**Résumé :** Le sillage cyclonique d'ondes proche-inertielles est un traceur clé du couplage océan atmosphère, symptomatique du brassage de l'océan par les vents intenses du cyclone, et se manifeste en surface par un refroidissement de plusieurs degrés et un creux d'une dizaine de centimètres. L'analyse de ces signatures nécessite la synergie entre la constellation satellitaire micro-onde et les flotteurs Argo, pour permettre un suivi complet des processus de mélange dans le sillage. Guidée par une approche semi-empirique, l'intégration de ces observations au sein de lois d'échelles permet, pour la première fois, de décrire avec grande précision ces anomalies de surface (température et hauteur dynamique), et d'étudier quantitativement leur dynamique et co-variabilité.

Les résultats les plus probants sont obtenus lorsque les observations hautes résolutions SAR du forçage sont combinées aux mesures argo et altimétriques, pour interpréter l'évolution des anomalies de niveau de mer. Afin de pallier l'intermittence temporelle des mesures SAR, un nouveau type de modèle paramétrique est aussi proposé, fondé non pas sur le rayon des vents maximums mais sur l'information d'un rayon extérieur, facilement déductible grâce aux observations plus répétitives des instruments micro-onde de basse et moyenne résolution. Enfin, l'utilisation de cette base semi-empirique offre l'opportunité, via les anomalies de niveau de mer, d'étudier le comportement du coefficient de poussée lors du passage d'un extrême.

**Title :** Mixing and sea surface anomalies in the inertial wake of tropical cyclones: processes and contribution of microwave satellite data

**Keywords :** Tropical Cyclones, inertial cold wake, scaling laws, MW remote sensing, Argo.

**Abstract :** The Tropical Cyclone wake of near-inertial waves is a key tracer of ocean-atmosphere coupling, symptomatic of ocean mixing by intense cyclonic winds, and is evidenced at the surface by a cooling of several degrees and a trough of about ten centimetres. The analysis of these signatures requires synergy between the constellation of microwave satellites and the Argo floats, to allow a complete monitoring of the mixing processes in the wake. Guided by a semi-empirical approach, the integration of these observations within scaling laws allows, for the first time, to describe these surface anomalies (temperature and dynamic height) with a high degree of accuracy, and to quantitatively study their dynamics and co-variability.

The most significant results are obtained when the high-resolution SAR observations of the forcing are combined with Argo and altimetry measurements to interpret the evolution of sea level anomalies. To overcome the temporal intermittency of the SAR measurements, a new type of parametric model is also proposed, based not on the radius of the maximum winds but on the information of an external radius, easily deduced from the more repetitive observations of low-to-medium resolution microwave instruments. Finally, the use of this semi-empirical approach offers the opportunity, via the sea level anomalies, to investigate the behaviour of the drag coefficient during the passage of an extreme event.