

The Barrier Layer and Ocean Colour in the Amazon and Orinoco Plume: Competing for the Oceanic Control on Tropical Cyclone Intensity

PHD THESIS

Christina Newinger

Imperial College London

Department of Physics

Space and Atmospheric Physics Group

Thesis submitted for the degree of Doctor of Philosophy of
Imperial College London

January 2015

Supervisor

Prof. Ralf Toumi

Examiners

Dr. Jérôme Vialard

Prof. Joanna Haigh

Statement of originality

I declare that this thesis is my own work, except where work by others has been properly referenced.

Publications

The research presented in this thesis has been submitted for publication:

Newinger, C. and R. Toumi, 2015: Potential impact of the colored Amazon and Orinoco plume on tropical cyclone intensity, submitted to Journal of Geophysical Research - Oceans.

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Funding

This work has been funded by the Climate - Knowledge and Innovation Community (KIC) created in 2010 by the European Institute of Innovation and Technology (EIT).

Abstract

The Amazon and the Orinoco river plumes modulate ocean stratification and colour in the tropical North Atlantic. This changes air-sea interactions and may thus be important for tropical cyclones (TCs). Using a regional ocean model, the potential impact of the rivers on ocean temperatures, stability, and TC intensity is investigated.

The influence of riverine freshwater on the ocean is twofold: Firstly the freshwater plume stabilizes the water column, and secondly ocean colour in the plume modifies solar transmission. Within the Amazon and Orinoco plume, the two mechanisms have opposing and effective cancelling effects on TCs. On the one hand the freshwater plume thickens the barrier layer by up to 15 m locally, leading to increased stability and temperature inversions. The cooling inhibition index (CI) is $+2.2 \text{ (J/m}^2\text{)}^{\frac{1}{3}}$ larger when the river freshwater is present, potentially reducing SST feedbacks. Surface temperatures and upper ocean heat content are only slightly modified by the rivers. Ocean colour in the freshwater plume on the other hand, blocks the deeper ocean from sunlight, leading to moderate surface warming ($+0.1^\circ\text{C}$) and substantial sub-surface cooling (-0.3°C 100m mean temperature). As a consequence cold water is more readily available to passing storms and the CI decreases by $-2.1 \text{ (J/m}^2\text{)}^{\frac{1}{3}}$, hence increasing negative SST feedbacks. The net effect of the coloured plume on SST cooling is thus negligible. Simple, idealized relationships between expected surface cooling and TC intensity, suggest that the river-induced barrier layer enhances strong TCs by up to -5 to -12 hPa, while ocean colour may reduce intensity of strong storms by +8 hPa to +16 hPa. The net impact of the coloured plume is negligible for weak storms and a slight intensity reduction for stronger cyclones. Within the Amazon and Orinoco plume, the river freshwater effect may thus be substantially reduced or even offset by light absorbing particles.

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Acronyms

AOP	apparent optical properties
AVHRR	Advanced Very High Resolution Radiometer
BL	barrier layer
CDOM	coloured dissolved organic matter
CFSR	Climate Forecast System Reanalysis
CI	cooling inhibition index
DJF	December, January, February
GC	Guyana Current
GHR SST	Group for High Resolution Sea Surface Temperature
ITCHP	interacting tropical cyclone heat potential
ITCZ	Inter Tropical Convergence Zone
IL	isothermal layer
IOP	inherent optical properties
JJA	June, July, August
MAM	March, April, May
ML	mixed layer
MCP	minimum central pressure
MDR	main development region
MODIS	Moderate-resolution Imaging Spectroradiometer
MPI	maximum potential intensity
NBC	North Brazil Current
NBCC	North Brazil Counter Current
NCEP	National Center for Environmental Prediction

NOAA	National Oceanic and Atmospheric Administration
NODC	National Oceanographic Data Center
PIDR	plume influenced development region
QAA	quasi analytical algorithm
ROMS	Regional Ocean Modelling System
SON	September, October, November
SST	sea surface temperature
TC	tropical cyclone
TCHP	tropical cyclone heat potential
TI	temperature inversion
TRMM	Tropical Rainfall Measuring Mission
WAWP	western Atlantic warm pool
WPI	wind power index
WTNA	western tropical North Atlantic

Chapter 1

Introduction

Tropical cyclones (TCs) are among the most destructive natural disasters. They form over the tropical oceans and extract their energy from the warm waters by evaporation. The Atlantic, Caribbean, and the Gulf of Mexico witness several cyclones each year, many of which cause major damage and loss of life and property (Figure 1.1). Anything affecting their ability to extract energy is crucial for TC development. The Amazon and Orinoco rivers form the largest freshwater source into the world's oceans (Warne et al., 2002) and have been predicted to be one such possible control (Ffield, 2007; Vizy and Cook, 2010; Balaguru et al., 2012b). The freshwater plume extends several thousand kilometres into the western tropical North Atlantic (WTNA) (Hu et al., 2004) modulating oceanic temperatures and stratification (Sprintall and Tomczak, 1992; Pailler et al., 1999; Masson and Delecluse, 2001).

The rivers increase salinity stratification and enhance the barrier layer (BL) in the WTNA (Pailler et al., 1999; Masson and Delecluse, 2001; Balaguru et al., 2012b), reducing entrainment into the mixed layer (ML). Positive sea surface temperature (SST) anomalies of up to 4°C have been observed within the plume area and been associated with more intense TCs (Ffield, 2007). Furthermore BLs reduce cooling feedbacks and can thus enhance TC growth (Balaguru et al., 2012a). As TCs pass over the ocean they entrain colder subsurface waters towards the surface (Price, 1981), leaving a cool SST wake. The cooling can exceed -6°C and limits TC intensification (Bender et al., 1993). The magnitude depends on the storm strength itself, but also on the underlying ocean structure (Shay et al., 2008; Cione and Uhlhorn, 2003; Schade and Emanuel, 1999). Since BLs are commonly accompanied by temperature inversions (Vialard and Delecluse, 1998; Pailler et al., 1999), they may even impose positive feedbacks on passing storms. Balaguru et al. (2012a) find that tropical cyclone intensification is 50% higher in the presence of BLs.

The optical properties of the Amazon and Orinoco plume are another factor, that

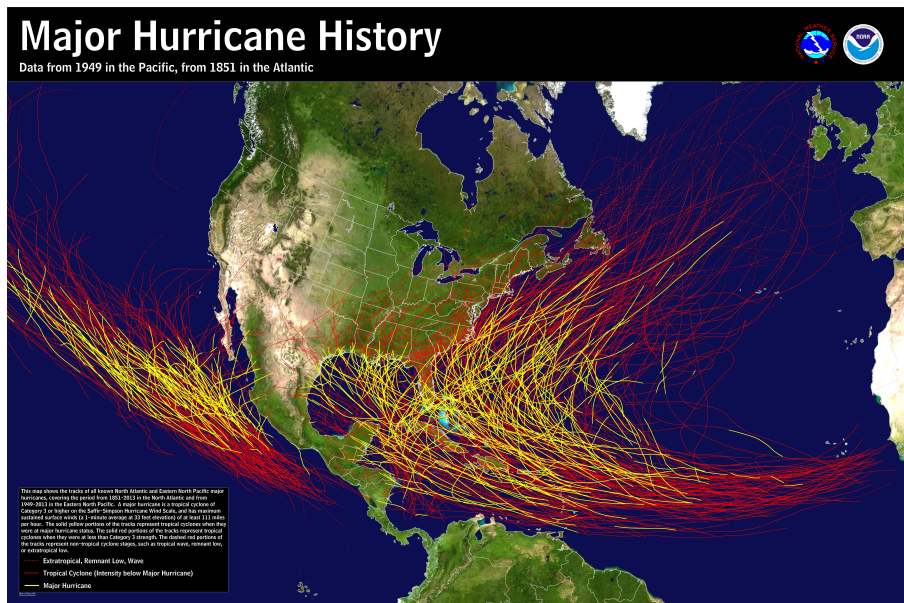


Figure 1.1: TC tracks over the tropical North Atlantic and Eastern Pacific in the period from 1851 to 2013. Yellow tracks show major hurricanes (above 3 on the Saffir-Simpson Scale), the red tracks show TCs below 3. From <http://www.nhc.noaa.gov/climo/> (accessed 07.12.14).

may alter ocean stratification and temperatures. The plume is highly turbid and contains sediments and coloured dissolved organic matter (CDOM) (Hu et al., 2004; Del Vecchio and Subramaniam, 2004). Furthermore the nutrient rich rivers also support primary production and lead to enhanced phytoplankton (i.e. chlorophyll) abundance in the WTNA (Smith and Demaster, 1996). The accumulation of light absorbing particles in the WTNA modifies the vertical heat transfer into the ocean. Solar radiation is absorbed nearer the surface, while the deeper ocean is blocked from sunlight (Denman, 1973; Ohlmann et al., 1996). Light attenuation by chlorophyll in the Amazon and Orinoco has been shown to warm SSTs and cool the ocean subsurface (e.g. Murtugudde et al. (2002)), but little effort has been put into understanding the potential impact on TC intensity. Gnanadesikan et al. (2010) suggest that ocean colour induced SST changes may amplify TCs and change storm tracks and frequency. However, the potential enhancement of SST feedbacks due to a colder ocean subsurface and the consequential weakening of TCs is not discussed.

In this work, the Regional Ocean Modelling System (ROMS) is applied to investigate the influence of the Amazon and Orinoco on ocean temperatures and stratification. For the first time the combined effect of freshwater input and light attenuation is considered. The potential impact on TC intensity is estimated using simple relationships between oceanic TC-metrics, surface cooling feedbacks and storm strength.

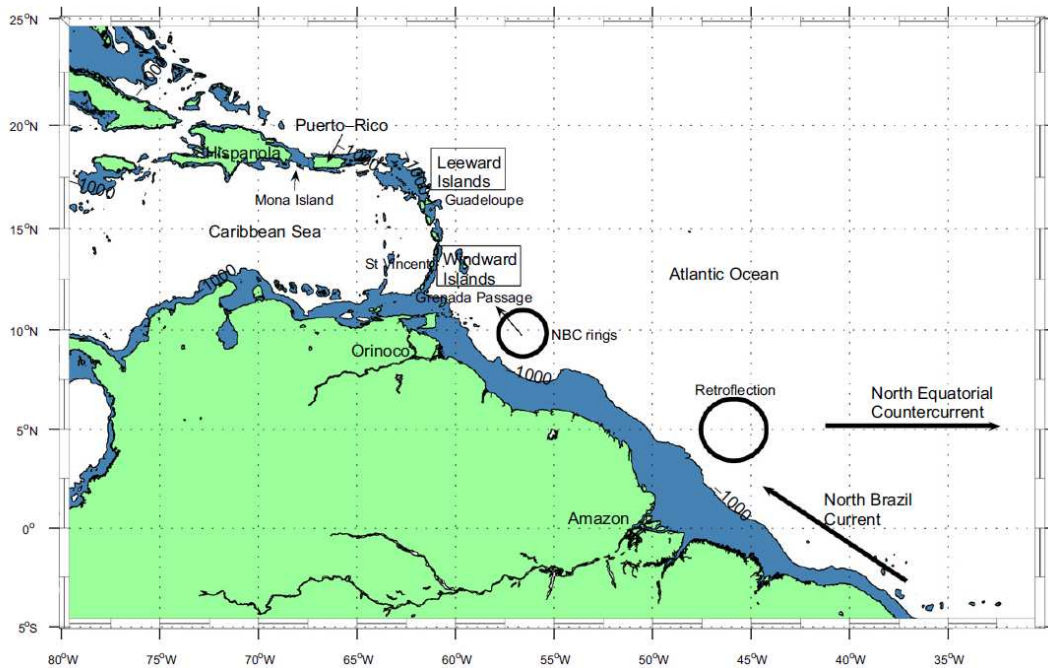


Figure 1.2: Map of the tropical North Atlantic and Caribbean Sea with the Amazon and Orinoco river, and the major currents in the area. Extracted from Cherubin and Richardson (2007).

1.1 The Amazon and Orinoco plume in the western tropical North Atlantic

1.1.1 The western tropical North Atlantic

The Amazon and Orinoco discharge into the western tropical North Atlantic (WTNA) (Figure 1.2). The climate and meteorology of this region is largely determined by the seasonal migration of the Inter Tropical Convergence Zone (ITCZ) from about 15°S in boreal winter to 5°N in boreal summer (Nieuwolt, 1977). Warm northern and southern air masses converge and rise near the ITCZ, leading to cloud formation and heavy tropical downpours. The southward position of the ITCZ in winter is associated with the dry season north of the equator, while the northern location in summer initiates the wet season. The heavy precipitation in summer is the main source of freshwater for the Amazon and Orinoco (Warne et al., 2002). The northeasterly trade winds vary seasonally, with the annual migration of the ITCZ, and are strongest in winter. Location and strength of the trade winds influence the ocean mixed layer depth in the WTNA.

With the vicinity of the equator and the western boundary, the fate of the Amazon and Orinoco plume is determined by energetic boundary currents (Figure 1.2),

namely the North Brazil Current (NBC), the Guyana Current (GC) and the North Brazil Counter Current (NBCC) (Richardson and Reverdin, 1987). The NBC carries water of South Atlantic origin northwestward along the coast of Brazil, across the equator and into the northern hemisphere. Near 6°N - 8°N a part of the NBC turns eastward, to feed the NBCC. The strength of the retroflection changes seasonally, is strongest during summer, and occasionally generates large isolated warm core rings with diameters exceeding 450 km and azimuthal velocities of up to 1 m/s (Fratantoni and Richardson, 2006). The anticyclonic rings move northwestward along the South American coastline towards the Caribbean. About 5 to 6 rings form per year, their average lifetime being 3 to 4 months, until they decompose near Barbados (Fratantoni and Richardson, 2006). The part of the NBC that continues to follow the South American coastline feeds the GC, a coastal current that flows northwestward. The currents enter the Caribbean Sea through the Antilles island channels, where they continue to flow westward (Johns et al., 2002). The seasonality in the trade winds drives large annual variations in these current patterns.

A feature that has been of some interest in the past is the western Atlantic warm pool (WAWP), a body of warm water ($\text{SST} > 28.5^{\circ}\text{C}$) (Figure 1.3). It forms in the equatorial Atlantic between March and June, in the Gulf of Mexico between July and August, and reaches its maximum in the Caribbean between August and October (Wang and Enfield, 2001). The strong seasonal cycle of the WAWP is mainly driven by the surface heat fluxes (Wang and Enfield, 2003), except in the equatorial Atlantic, where advective processes associated with the equatorial cold tongue are dominant (Lee et al., 2005b). Field (2007) attribute warm anomalies within the WAWP to the influence of the Amazon and Orinoco plume. Wang et al. (2006) assess the impact of the Atlantic warm pool on hurricanes and conclude that a larger WAWP may increase tropical cyclone activity. Besides warmer SSTs this is due lower sea level pressure, enhanced convection, and weaker wind shear in the presence of large warm pools.

1.1.2 The Amazon and Orinoco and their basins

The Amazon is the largest river in terms of freshwater discharge, and the Orinoco is the fourth largest. Together they provide about 20% of global river discharge and are the most significant freshwater source into the world oceans (Baumgartner and Reichel, 1975).

The Amazon drains about a third of the South American continent (Nittrouer and De Master, 1986). Its drainage basin covers about $5.9 \times 10^6 \text{ km}^2$ (Figure 1.4a) and it discharges $6.3 \times 10^{12} \text{ m}^3$ freshwater per year into the North Atlantic at the

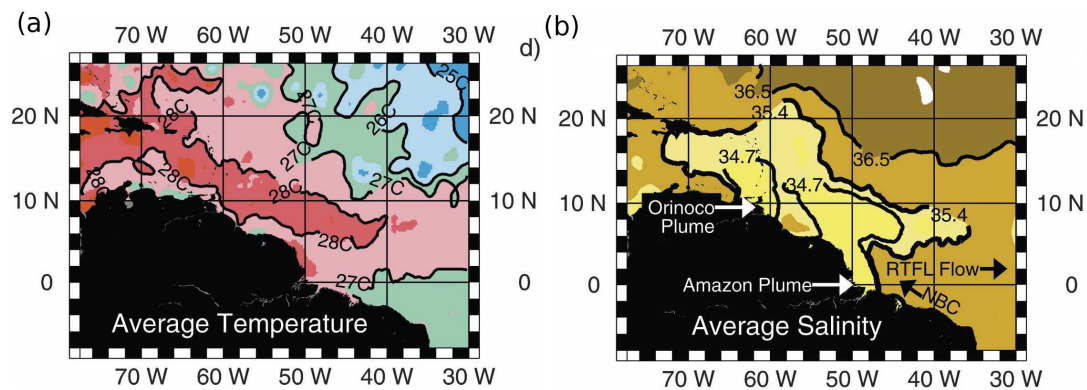


Figure 1.3: June through November hurricane season (a) average observed surface temperature ($^{\circ}\text{C}$) and (b) average observed surface salinity (PSU). Extracted from Ffield (2007)

western coast of North Brazil (Warne et al., 2002). The average channel width is about 5 km and the depth is between 20-50 m (Lewis et al., 1995). The river travels about 6400 km through South America before reaching its delta. The delta plain itself is comparatively small (25,000 km^2) and the tidal range at the coast is about 6 m (Meade, 1996). A peculiarity of the Amazon is that its discharge is so consistently large that seawater never enters the channels (Warne et al., 2002). The main stem of the Amazon river is along the equator. It is thus supplied with water from both, northern and southern tributaries. The tributaries are strongly influenced by their respective wet and dry season, and their discharge varies accordingly by a factor of up to 10 (Warne et al., 2002). Since northern and southern wet and dry season are about 6 months out of phase, the annual cycle of the Amazon river discharge is relatively smooth. It only varies by a factor of about three and peaks between May and July (Meade et al., 1991). The Tocantins, which also discharges into the WTNA slightly east of the Amazon river, is usually assigned to the Amazon river discharge. Its freshwater flow is about a third of the Orinoco (Milliman and Farnsworth, 2011).

The Orinoco releases about $1.2 \times 10^{12} \text{ m}^3$ of fresh water per year into the WTNA (Meade, 1996) and its basin covers about $1.1 \times 10^6 \text{ km}^2$ of northern South America (Figure 1.4b). The Orinoco is approximately 2000 km long, before it eventually discharges into the North Atlantic and the Gulf of Paria (Hamilton and Lewis, 1990). The mean river channel width is 2 km and the mean depth is 20-25 m in lower regions. The tidal range at the coast is 2.6 m. The Orinoco delta plain in Venezuela covers about 22,000 km^2 and is therefore almost the size of the mightier Amazon. The delta (Figure 1.4c) is a mosaic of inter-distributary swamps with five to six major channels (Warne et al., 2002). Nearly 85% of the Orinoco is discharged through the Rio Grande along the southern margin of the delta. The distribution among the northern arms of

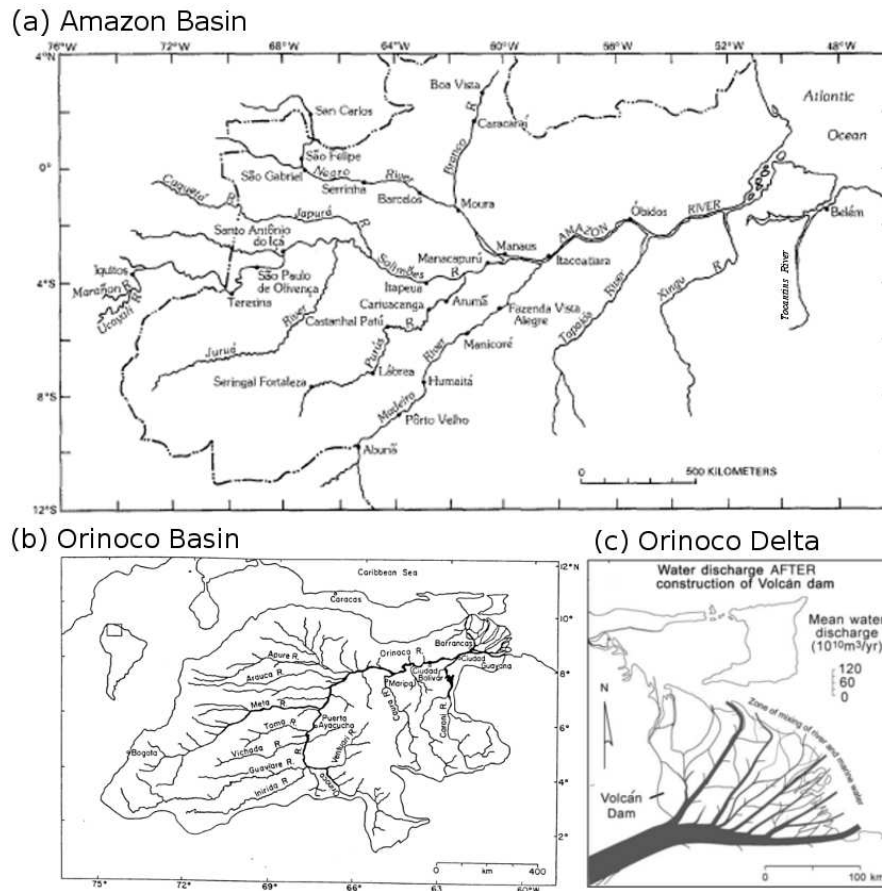


Figure 1.4: (a) Amazon basin and tributaries from Meade et al. (1991) and (b) Orinoco basin and tributaries from Hamilton and Lewis (1990) and (c) the Orinoco delta with the distribution of freshwater discharge from Warne et al. (2002).

the Orinoco delta has been modified by the Volcan dam being completed in 1965 at the Cano Manamo, the westernmost arm of the Orinoco delta. Nearly all the water of the northwestern Orinoco delta is therefore discharged into the North Atlantic near western Trinidad (Warne et al., 2002). The arrival of the ITCZ in the Orinoco basin marks the start of the wet season and generates large volumes of rainfall from June to November. The mean annual variation of Orinoco discharge is 26:1 (Nordin et al., 1994) and is one of the largest for world rivers (Warne et al., 2002).

1.1.3 Dispersal and properties of the Amazon and Orinoco plume

The dispersal of the Amazon and Orinoco into the WTNA leads to surface salinities that are several PSU fresher, than the ambient ocean. This light freshwater plume is mainly distributed horizontally across the ocean surface by winds and currents, and can extent several thousand kilometres into the tropical North Atlantic (Hu et al., 2004) (Figure 1.3b).

The plume size is largest in summer and autumn with a peak in July (about $1.2 \times 10^6 \text{ km}^2$ (Hu et al., 2004) to $1.5 \times 10^6 \text{ km}^2$ (Moller et al., 2010)) and lags the peak in river discharge by about 1 to 2 month on average. The area estimates neglect the Caribbean and only extend from the equator to 15°N (Hu et al., 2004). Since a considerable amount of freshwater can be traced into the Caribbean (Müller-Karger et al., 1989; Hellweger and Gordon, 2002; Hu et al., 2004) or can be found as far as 20°N (Hu et al., 2004; Grodsky et al., 2012), the actual plume area may be even larger.

Most of the Amazon discharge is transported northwestward by the NBC and covers most of the coastal shelf (Müller-Karger et al., 1995). From spring through fall a large proportion of the Amazon plume (up to 70%) is transported northeastward with the meandering NBCC (Lentz, 1995a; Johns et al., 1990). Some freshwater is trapped in the current rings and is re-released near the Lesser Antilles, from where the plume enters the Caribbean. With the onset of strong northeasterly trade winds in autumn, the ocean restratifies and the plume diminishes, until the minimum is reached in winter. The plume is about 3-20 m thick along the shelf (Lentz and Limebunger, 1995), and between 20-30 m thick in the open ocean (Hu et al., 2004).

The Orinoco plume area has been estimated by Müller-Karger et al. (1989) to cover about $160,000 \text{ km}^2$ in autumn in the Caribbean. The Orinoco plume is mainly transported northwestward by the strong GC and transported into the Caribbean through the Gulf of Paria and the Antilles island channels. East of the Antilles islands and in the Caribbean, the Amazon and Orinoco plume merge and are virtually indistinguishable (Hellweger and Gordon, 2002).

The spatial extent of the plume can vary considerably from year to year (Hu et al., 2004; Grodsky et al., 2014). Variability in the North Atlantic freshwater signal is driven by varying discharge, current variability and the trade winds (Lentz, 1995a; Nikiema et al., 2007), but also by precipitation (Coles et al., 2013; Grodsky et al., 2014). Furthermore wind mixing and the mixed layer depth may play an important role (Grodsky et al., 2014).

In the past the Amazon and Orinoco plume has been detected from in situ salinity measurements (e.g. Lentz (1995a); Lentz and Limebunger (1995); Del Castillo et al. (1999); Hellweger and Gordon (2002); Johns et al. (1990)) and satellite derived ocean colour (e.g. Müller-Karger et al. (1989); Hu et al. (2004); Salisbury et al. (2011)). The Amazon and Orinoco transport large amounts of sediments and detrital material into the WTNA. Furthermore the nutrient rich freshwater (mainly nitrogen) supports primary productivity and therefore leads to high phytoplankton (chlorophyll) concentrations in the plume area. As a consequence the optical proper-

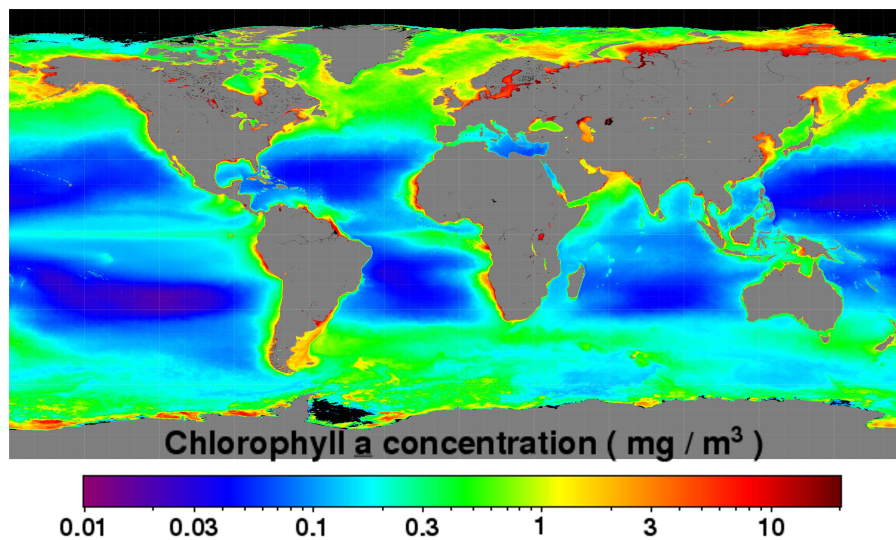


Figure 1.5: Composite mean chlorophyll concentration between 2002 and 2014 as observed from MODIS Aqua, available online at <http://oceancolor.gsfc.nasa.gov/>

ties of the plume are very different from the open ocean and are visible from space (Figure 1.5). Müller-Karger et al. (1989) detected the Orinoco plume from satellite chlorophyll concentration, arguing that primary productivity is largely dependent on the dispersal of the Orinoco. Hu et al. (2004) collected in-situ salinity data and satellite derived coloured dissolved organic matter (CDOM) to show that low surface salinity and CDOM concentration are well correlated (0.6). Salisbury et al. (2011) obtain similar results regarding the relationship of surface salinity and CDOM absorption using satellite observed salinity data. It is only recently, that Aquarius and soil moisture and salinity (SMOS) satellite data have become available, that directly derive surface salinity from microwave thermal emission frequencies in the ocean (e.g. Lagerloef (2012)).

1.2 Impact of the Amazon and Orinoco plume on the ocean

1.2.1 Barrier layer formation and ocean temperatures

The influence of the Amazon and Orinoco can be felt hundreds of miles from the river mouths. The light freshwater enhances ocean stratification and thickens the local BL (Sprintall and Tomczak, 1992; Pailler et al., 1999; de Boyer Montégut et al., 2007). BLs lie between the density mixed layer (ML) and the top of the thermocline (Lukas and Lindström, 1991) (Figure 1.6). In large parts of the ocean density stratification

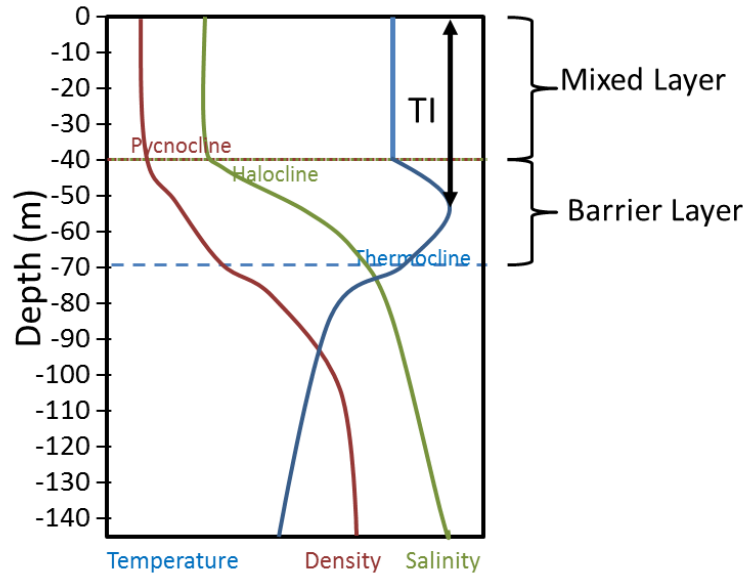


Figure 1.6: Idealized profiles of ocean temperature, salinity and density in an area of low surface salinity. The mixed layer (ML) lies above the pycnocline and a barrier layer (BL) forms between the ML and the top of the thermocline. A temperature inversion (TI) can form in the BL.

is mainly determined by temperature, but in areas of low surface salinity the ML can be shallower than the isothermal layer (IL). In the WTNA the BL is formed by surface freshening from rivers and rain (Foltz et al., 2004). Another important formation mechanism is the subduction of high salinity waters of south Atlantic origin, creating subsurface salinity maxima in the WTNA (Blanke et al., 2002; Balaguru et al., 2012a).

In this work the BL criterion suggested by de Boyer Montégut et al. (2004) is applied. The depth of the IL and the ML are given by a finite difference criterion. At the top of the thermocline the temperature has decreased by a threshold ΔT as compared to the temperature at the reference depth (10 m) (de Boyer Montégut et al., 2004). A density decrease ($\Delta \rho$) corresponding to the same temperature decrease (ΔT) at constant salinity marks the end of the ML:

$$\Delta \rho = \rho(T_{ref} - \Delta T, S_{ref}, P_0) - \rho(T_{ref}, S_{ref}, P_0) \quad (1.1)$$

Different thresholds for the temperature and density criterion have been used throughout the past ranging from 0.1°C to 1.0°C and 0.01 kg m^{-3} to 0.125 kg m^{-3} , respectively (de Boyer Montégut et al., 2004). Here the 0.2°C -threshold recommended by de Boyer Montégut et al. (2004) is applied for temperature and density.

BLs can modulate SSTs since they reduce entrainment of cold water from below the thermocline into the ML (Vialard and Delecluse, 1998). This effect is opposed by

solar penetration. With the shallowing ML, more solar radiation leaves at the ML base, potentially cooling the ocean surface. Furthermore surface cooling affects a shallower layer, potentially reducing ML temperatures. The net effect on SSTs therefore depends on the balance of these mechanisms. Pailler et al. (1999) find that low salinity in the tropical North Atlantic often occurs simultaneously with high surface temperatures. Ffield (2007) consider the observed correlation between low surface salinity and warm SST anomalies as circumstantial evidence for the Amazon and Orinoco plume warming the ocean surface (Figure 1.3). Similarly Foltz and McPhaden (2009) conclude that BLs have a considerable impact on SSTs through the modulation of vertical heat fluxes at the base of the ML. Masson and Delecluse (2001) on the other hand, do not simulate warming in the Amazon plume in their model study and Balaguru et al. (2012a) find that BL thickening in the Amazon and Orinoco plume is not able to produce significant surface warming. Vialard and Delecluse (1998) show that BLs can even lead to surface cooling in the Pacific.

The BL itself is isolated from surface cooling, but still receives sunlight. As a consequence the BL warms (Vialard and Delecluse, 1998). Locally this warming can lead to temperature inversions (TIs) (Figure 1.6). Masson and Delecluse (2001) shows that surface freshening of the Amazon warms the BL and induces TIs. Inversions of up to 1.0°C were observed in the tropical North Atlantic in winter (de Boyer Montégut et al., 2004; Mignot et al., 2012).

1.2.2 Ocean colour and solar transmission

The Amazon and Orinoco enhance the concentration of organic and inorganic compounds in the ocean. The presence of light absorbing particles affects solar transmission into the ocean and may thus alter oceanic temperatures and stratification (Denman, 1973; Simpsom and Dickey, 1981; Morel, 1988). The importance of solar transmission has also been confirmed by observations (Siegel et al., 1995; Ohlmann et al., 1996). Therefore accurate representation of light propagation is crucial to correctly simulate physical properties of the ocean.

1.2.2.1 Radiative transfer in the ocean

The optical properties of seawater can be separated into inherent optical properties (IOP) and apparent optical properties (AOP) (Mobley, 2001). IOPs are only dependent on the optical properties of seawater itself. The ocean affects solar transmission by absorption and scattering. Therefore IOPs are defined by the wavelength dependent absorption ($a(\lambda)$ [m^{-1}]) and scattering coefficients ($b(\lambda)$ [m^{-1}]). The ab-

sorption coefficient is the fraction of incident power that is absorbed within a volume of water, as the thickness becomes small. Similarly $b(\lambda)$ describes the amount of scattered radiation, and the backscattering coefficient $b_b(\lambda)$ is the backwards scattered amount. The total attenuation coefficient ($c(\lambda)$) of a light beam is then defined as the sum of absorption and scattering coefficients. The propagation of light in the ocean can be described by a radiative transfer equation, which combines IOPs with the properties of incoming radiance. It is a complex differential equation dependent on wavelength, incident light angle, wavelength, IOPs, and the physical environment (Mobley, 2001). Advanced models that solve the radiative transfer equation are available (e.g. Mobley (1995)), but are impractical for the use in three-dimensional, highly-resolved ocean models, due to their computational inefficiency (Lee et al., 2005). Therefore simplified algorithms for light transmission in the ocean are usually applied.

The AOPs depend on the structure of the light field and IOPs. A commonly used AOP is the remote sensing reflectance (R_{rs} [sr^{-1}]). R_{rs} just above the sea surface can be obtained from satellites. It is defined as the ratio of the water leaving radiance, and the downwelling irradiance and can be regarded as a measure of ocean colour (Mobley, 2001). Note that radiance is the incident power per unit angle per area, while irradiance is the incident power per unit area. R_{rs} is used in the quasi-analytical algorithm of Lee et al. (2002) to derive absorption and backscattering coefficients of the ocean.

Different particles absorb at different wavelength and lead to varying optical properties of seawater (Mobley, 2001): Seawater itself does not absorb significantly in the visible part of the spectrum, but rather in the near-ultraviolet and near-infrared. Coloured dissolved organic matter (CDOM) stems from plants and is commonly found in coastal areas. It colours ocean water yellow or brown and strongly absorbs light in the blue end of the spectrum. Chlorophyll in phytoplankton is usually the dominant absorber in the ocean and strongly absorbs in the red and blue. Other organic particles like detritus can also be important absorbers in the blue part of the visible spectrum. It is for instance produced when phytoplankton die, or when zoo-plankton graze on phytoplankton and leave cell parts or fecal pellets behind. Near river outlets inorganic particles, like sediments can also be important for light absorption. From the optical properties, the concentrations of chlorophyll or CDOM can be derived (e.g. Hu et al. (2004, 2012)) and are available as satellite products (<http://oceancolor.gsfc.nasa.gov/>).

1.2.2.2 Solar transmission algorithms in ocean models

Several absorption algorithms for the use in numerical models have been developed (Morel, 1988; Ohlmann, 2003; Lee et al., 2005). An early solution that is still popular in ocean models today is the Paulson and Simpson (1977) radiation formulation. It is based on the assumption that the downward irradiance in the ocean decreases exponentially with depth for constant attenuation coefficients, according to the Beer-Lambert law. The depth-dependent irradiance is then split into a sum of exponential terms representing different wavelengths:

$$I(z) = I_0 \left[R \cdot e^{-\frac{z}{\zeta_1}} + (1 - R) \cdot e^{-\frac{z}{\zeta_2}} \right] \quad (1.2)$$

ζ_1 and ζ_2 are the inverse of the visible and infrared attenuation coefficients, also called the attenuation depths. R represents the partition of visible and infrared light, I_0 is the incoming solar irradiance, and z is the depth in the ocean. Transmission coefficients and exponents are often given as a function of the five Jerlov (1976) water types (I, IA, IB, II, III). Water type I represents the clearest water and water type III stands for the cloudiest water. Figure 1.7a shows transmittance of different water types as a function of wavelength. Tropical waters are normally clear and therefore categorized as water type I. The transmittance peaks at around 450 nm for clear waters at an incidence level of 90°. Subtropical and midlatitude waters contain more chlorophyll and are therefore less transmittal. Nearer the coasts the abundance of light absorbers is even larger and the water therefore more turbid. For type I waters about 1% of the sunlight at 456 nm reaches a depth of 140 m (Figure 1.7b). For type III the 1% threshold is already met at about 40 m. Open oceans are normally categorized as water type I, assuming that there are no light absorbing particles contained in the ocean. This value has been widely used in heat budget studies of the ocean (McPhaden, 1982; Wang and Phaden, 1999; Foltz et al., 2003).

However, splitting ocean waters into different types is a fairly rough estimate and transitions between the different types are not smooth. Ohlmann (2003) have developed an algorithm that uses surface chlorophyll concentration to determine light absorption. Similarly Murtugudde et al. (2002) and Ballabrera-Poy et al. (2007) have determined the attenuation depth from satellite observed chlorophyll concentration. Sweeney et al. (2005) also used satellite derived chlorophyll concentration to determine solar transmission in an ocean-only model, and Gnanadesikan and Anderson (2009) used the same approach in a coupled model. Since these methods are based on chlorophyll, they may not perform as well in regions where other light absorbing particles, like CDOM are dominant. This is for example the case in the Amazon and

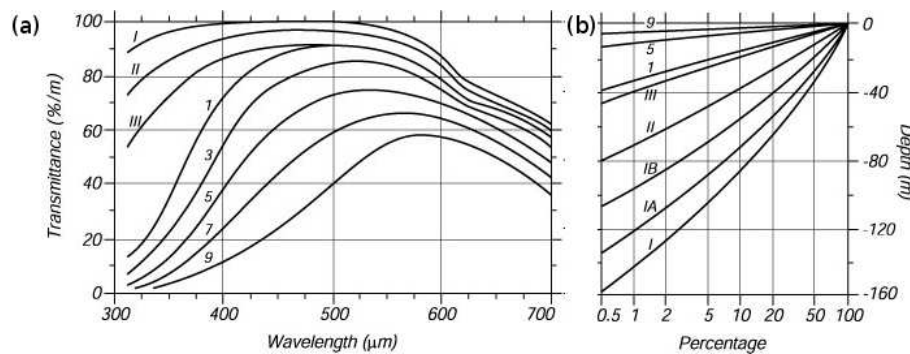


Figure 1.7: (a) Transmittance of solar radiation in the ocean in %/m as a function of wavelength. I: extremely clear water; II: turbid tropical-subtropical water; III: mid-latitude water; 1-9: coastal waters of increasing turbidity. Incidence angle is 90° for the first three cases, 45° for the other cases. (b) Percentage of 465 nm light reaching indicated depths for the same types of water. From Jerlov (1976)

Orinoco plume (Hu et al., 2004; Del Vecchio and Subramaniam, 2004). Therefore Lee et al. (2005) developed an algorithm, that derives light absorption directly from surface optical properties and is therefore able to account for all light absorbing particles in the ocean. The Lee et al. (2005) algorithm will be applied in this study to estimate the impact of the Amazon and Orinoco plume on solar transmission.

1.2.2.3 Solar transmission in the Amazon and Orinoco plume

The main light absorber in the Amazon and Orinoco plume is CDOM (Hu et al., 2004), and nearer the river mouths also sediments (Del Vecchio and Subramaniam, 2004). Further offshore - but still in the freshwater area - chlorophyll and detrital material are also important (40% each). Primary productivity is most intense at the interface of riverine and oceanic waters, where high nutrient concentrations coincide with more available sunlight (Del Vecchio and Subramaniam, 2004). Subramaniam et al. (2007) show that sunlight reaches shallower depth within the plume, compared to the ambient ocean.

A first order effect of shallower attenuation depth is a warmer ocean surface, since more radiation is absorbed in the upper layer. The deeper ocean is blocked from sunlight and can therefore be expected to cool. However, modified light absorption also changes the ocean circulation (Murtugudde et al., 2002; Sweeney et al., 2005) and ocean-atmosphere interactions (Timmermann and Jin, 2005; Shell et al., 2003), and has a feedback on the biological production in the ocean itself (Manizza et al., 2008). In Murtugudde et al. (2002) realistic solar transmission leads to warming of SSTs in the Amazon plume area of up to 0.4°C , compared to the assumption of clear water. Furthermore the ML shallows by up to 20 m in the plume region, suggesting

substantial subsurface cooling. Nakamoto et al. (2001) and Sweeney et al. (2005) found that decreasing water clarity actually cools the Atlantic and Pacific cold tongues. Sweeney et al. (2005) argue that this is due to a colder upwelling waters at the mouth of the cold tongue. In a coupled model Gnanadesikan and Anderson (2009) find that the same mechanism cools the equator, where equatorial upwelling brings cooled waters to the surface. They find subsurface cooling of up to -2.2°C in the tropics. However, their surface still warms in the Amazon and Orinoco plume region. These are only a few, among many studies, that illustrate the importance of realistic light absorption when modelling ocean stratification and temperatures in the WTNA. This aspect has not been considered in previous model studies on the impact of the Amazon and Orinoco plume on ocean temperatures and stratification (e.g. Masson and Delecluse (2001); Balaguru et al. (2012b)).

1.3 The oceanic control on tropical cyclone intensity

1.3.1 Tropical cyclone formation and intensity

TCs are large rotating storm systems that develop over the tropical oceans. They preferably form over warm waters and draw their energy from the upper ocean through latent and sensible heat fluxes. The three dimensional structure of a TC is described in Emanuel (2005) and shown in Figure 1.8. TCs are characterized by a low pressure eye, which sucks surface air towards the TC centre in a spiralling manner. A fraction of the inflowing air supplies energy to the storm system by evaporating ocean water, which is then drawn into strong updrafts. Eventually the water vapour re-condensates and is visible as spiralling clouds and rain bands surrounding the TC eye. However, the most significant energy source is near the eyewall, where the surface winds are strongest and evaporation is maximum. The water vapour is sucked upwards in the TC eye and ascends in a broadening spiral. Finally the cooled air flows outward at the TC top. Tropical cyclones can extend up to about 18 km into the atmosphere and have radii of about 100 km to 2000 km (Emanuel, 2005). The cyclone eye typically has a diameter of about 50 km, but small eyewalls of only around 3 km have also been observed. The TC winds are set into rotation by the Coriolis force and turn anticlockwise in the northern hemisphere. They are strongest just outside the eyewall, while the TC eye is practically calm. The whole system moves across the ocean surface with the ambient winds. Most tropical cyclones in the WTNA therefore travel from east to west with the trade winds. In the North Atlantic TCs are also called hurricanes, and form in the North Atlantic hurricane season (June to

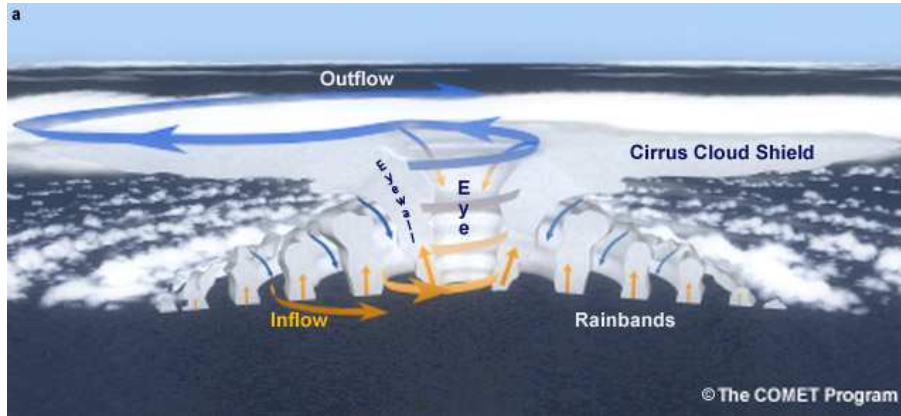


Figure 1.8: Three-dimensional structure of a tropical cyclone. Extracted from Gray and Emanuel (2010).

November). About 85% of TCs in the tropical north Atlantic are believed to originate from African easterly waves (Landsea, 1993). When TCs hit land they are deprived off their energy source and rapidly start to decay.

Several theories for the maximum obtainable energy of TCs have been developed. They can be classified into dynamic and static theories (Emanuel, 2005). Dynamic theories describe the storms as heat engines, while static theories mainly consider the eye and eyewall of the unperturbed environment. Early examples of static models are Miller (1958) or Holland (1997). The first dynamic theory has been proposed by Emanuel (1988). Treating the TC as a Carnot cycle, the maximum potential intensity (MPI) can be calculated in terms of the minimum sustainable surface pressure from the surrounding atmospheric conditions and the underlying SST. This has been further developed by Emanuel (1999) to a formulation of the maximum sustained wind speed (v_{max}):

$$v_{max}^2 = \frac{C_k}{C_D} \frac{T_s - T_0}{T_0} k_s - k_a, \quad (1.3)$$

C_k and C_D are dimensionless exchange coefficients for enthalpy and momentum, T_s and T_0 are the absolute temperatures of the sea surface and storm top, and k_s and k_a are the specific enthalpies of the air at saturation at the ocean surface and in the atmospheric boundary layer.

TCs are divided into different categories according to the Saffir-Simpson hurricane wind scale, based on maximum sustained wind speeds (Table 1.1). According to this definition, tropical storms need to have a sustained wind speed of at least 33 m/s in order to be classified as hurricanes. TCs above 50 m/s (category 3) are referred to as major hurricanes, due to their potential to cause significant loss of life and damage.

Category	Sustained Winds	Damage
1	33 m/s- 43 m/s	Minimal damage
2	43 m/s- 50 m/s	Extensive damage
3	50 m/s- 58 m/s	Devastating damage
4	58 m/s- 70 m/s	Extreme damage
5	≥ 70 m/s	Catastrophic damage

Table 1.1: Saffir-Simpson hurricane scale

Note that v_{max} and MPI describe the maximum theoretically possible intensity. The v_{max} achieved in reality is usually lower and the actual minimum central pressure (MCP) is larger than MPI, due to atmospheric and oceanic constraints (Emanuel, 1999). One of them is the self-induced surface cooling discussed below.

1.3.2 The ocean response to tropical cyclones

The strong wind stress TCs impose on the ocean surface trigger shear instability and vertical mixing. This deepens the isothermal layer (IL), entraining cold water from below the thermocline (Price, 1981). Therefore TCs often leave considerable cold wakes (Figure 1.9a). Their magnitude varies between about -1°C and up to -6°C (Black, 1983). Entrainment cooling is responsible for about 80% of the observed SST response, but can vary from about 70% to 90% for different TCs (Vincent et al., 2012a). In comparison the importance of air-sea heat fluxes is much lower (Price, 1981). In shallow regions however, latent cooling can become dominant (Shen and Ginis, 2003). The ocean response to TCs has been discussed in various case studies (e.g. Shay et al. (2008); D'Asaro (2003); D'Asaro et al. (2007)) and models (e.g. Price (1981); Sanford et al. (2007); Jullien et al. (2012)).

A pronounced rightward bias in the SST response has often been observed (e.g. Pudov et al. (1979); Shay et al. (1992); Liu et al. (2007); Mc Phaden et al. (2008)) and can be seen in Figure 1.9. On the one hand the shift is due to the anticlockwise rotation of storms (in the northern hemisphere). On the right side storm translation speed and wind speed are in the same direction and therefore amplified, while they oppose each other on the left hand side of the storm (Shay et al., 1989). However, Price (1981) argue that the main reason is the near-resonant coupling of the turning of the TC wind vector with storm-induced inertial oscillations on the right hand side of the storm. In the ocean frame, the wind vector turns clockwise on the right, as do the inertial currents. On the left hand side the wind vector turns anti-clockwise. It is therefore anti-resonant with the rotation of inertial currents, leading to less efficient

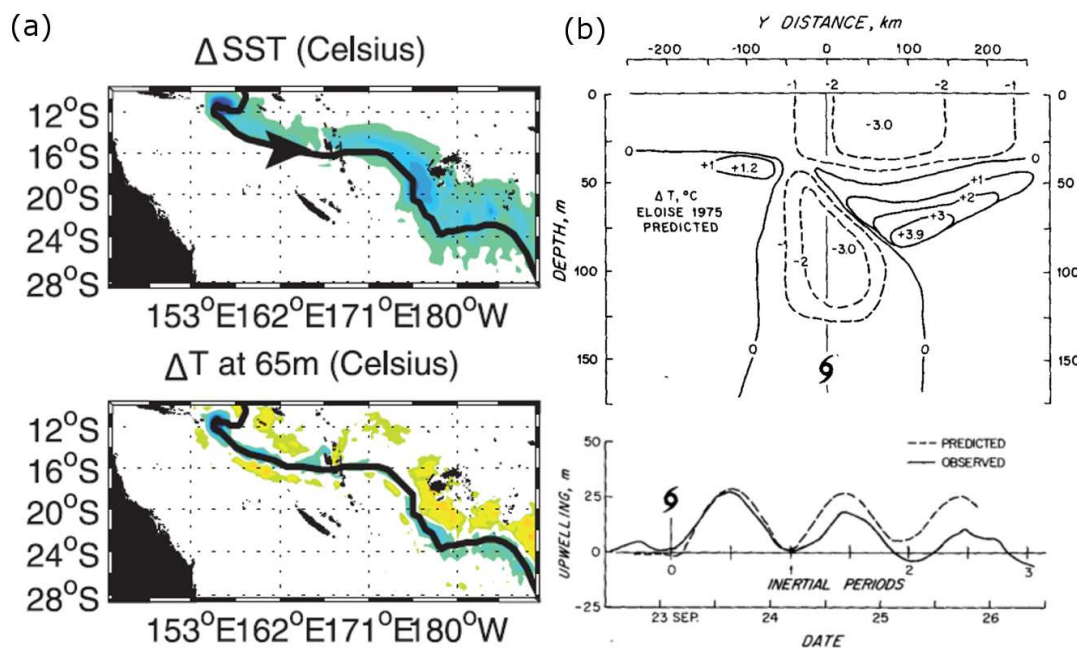


Figure 1.9: (a) Modelled surface and 65 m temperature response of a typical TC in January 1979 by obtained by comparison of a TC and no-TC experiment. Extracted from Jullien et al. (2012). (b) Predicted temperature response to TC Eloise (1975) at 1.5 inertial periods, shown as a cross track section. Contour interval is 1°C , negative values are dashed and time evolution of observed and predicted upwelling below the ML base for TC Eloise. Extracted from Price (1981).

mixing.

By mixing the upper ocean TCs also alter subsurface temperatures and stratification (Figure 1.9). Wind mixing produces a deeper, colder isothermal layer, and consequently warms below the thermocline. Naturally this heating is biased towards the right as well. However, the rotational wind vector also induces a divergence of Ekman currents in the ML (Price, 1981), leading to Ekman pumping and upwelling of cold waters. Due to the near symmetrical structure of the wind rotation, this response is centred along the track. The upwelling magnitude oscillates at a near inertial frequency (Figure 1.9b), and the upwelling isotherms can reduce subsurface warming and ML deepening at the track centre. At some distance to both sides of the TC track, downwelling causes a small subsurface warming (Figure 1.9a).

In summary, TCs induce cooling above the IL and warming directly below, with a bias towards the right side of the track. Ekman pumping triggers oscillating upwelling (i.e. cooling) below the thermocline, centred along the TC track. The maximum cooling in the TC wake occurs about 24 hours (Mei et al., 2012; Vincent et al., 2012a) to 48 hours (Jullien et al., 2012) after cyclone passage (Figure 1.10), when entrainment cooling and upwelling amplify each other. Afterwards surface temperatures slowly restore towards their pre-storm values, but remain slightly below, even about a month after TC passage. The magnitude of the cold wake depends on the storm strength (Figure 1.10) and translation speed (Zedler, 2009), but also on the initial ocean stratification (Schade and Emanuel, 1999).

1.3.3 Oceanic control of tropical cyclone intensity

The TC induced wake cooling is important for storm intensity itself, due the sensitivity of TCs to surface temperatures. Hereby the cooling under TC eye, rather than in the wake is the relevant parameter, since this is where air-sea fluxes are maximum. While the sensitivity of MPI to ambient temperatures is about 6 hPa/°C (Schade, 2000), the sensitivity to the eye temperature can be 33 hPa/°C (Holland, 1997). The impact of self-induced cooling on TC intensity has been studied in models (e.g. Bender et al. (1993); Bender and Ginis (2000); Schade and Emanuel (1999)) and been confirmed by observations (e.g. Cione and Uhlhorn (2003); Lloyd and Vecchi (2011)). Cione and Uhlhorn (2003) conclude that a -1°C cooling can reduce heat fluxes by about 40% and find a correlation of 0.42 between cooling under the TC eye and intensity reduction. The linear relationship of surface cooling and TC intensity reduction in Schade (2000) suggests that a 2.5° temperature drop under the TC eye may be sufficient to close down a TC completely. Emanuel (1999) show that TC intensity forecasts

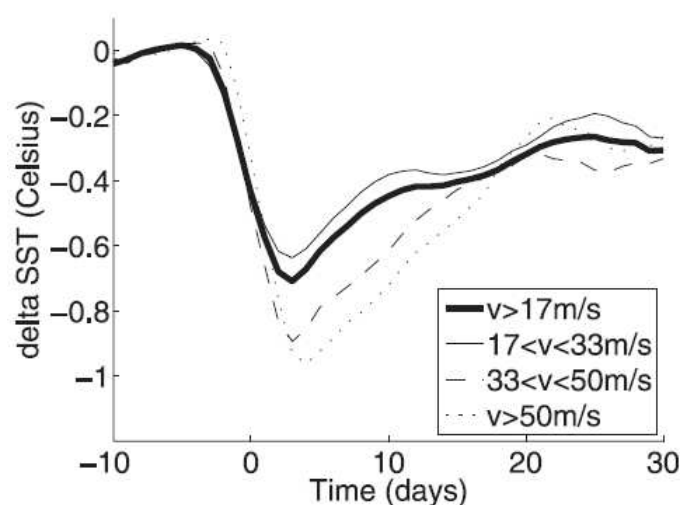


Figure 1.10: Composite of observed SST cooling anomaly ($^{\circ}\text{C}$) from 1979-2003 obtained by the TRMM Microwave Imager (TMI)-Advanced Microwave Scanning Radiometer for Earth Observing System (EOS) for various TC categories: all TC winds stronger than 17 m/s (bold solid line), TC winds between 17 and 33 m/s (thin solid line), TC winds between 33 and 50 m/s (dashed line), and TC winds stronger than 50 m/s (dotted line). Extracted from Jullien et al. (2012).

can be significantly improved when considering oceanic SST feedbacks.

According to Mei et al. (2012) the impact of the SST feedback strongly depends on the translation speed. Slow storms are likely to produce more surface cooling at a given location and also spend a longer time period above this cooling. Therefore they will be weakened to a stronger degree. Among the most important controls of SST feedback is the initial mixed layer depth, and the stratification below (Schade and Emanuel, 1999). This is intuitive, since it determines how easily the TC can entrain cold waters to the surface. The importance of ocean stratification on TC intensity has for example been observed by Lloyd and Vecchi (2011). Shay et al. (2008) found rapid intensification of Hurricane Opal whilst passing over a deep warm core ring. In a model study Vincent et al. (2014) showed the importance of interannual variations in ocean stratification on TC intensity, particularly for strong TCs. The impact of ocean stratification on SST feedbacks may be particularly relevant in the region of the Amazon and Orinoco plume, where the strong salinity gradient creates a barrier for vertical mixing and therefore reduces cooling feedbacks (Balaguru et al., 2012a; Grodsky et al., 2012). Furthermore light absorbing particles in the plume may cool the subsurface and potentially increase cooling feedbacks.

1.3.4 Oceanic metrics for hurricane ocean interactions

1.3.4.1 Ocean temperature and upper ocean heat content

The oceanic control on tropical cyclone intensity can be estimated using various different metrics. The initial SST is a first predictor for the potential intensity of a storm (Emanuel, 1987), but TCs also interact with the ocean subsurface. The tropical cyclone heat potential (TCHP) considers the heat stored in the upper ocean and therefore includes possible cooling feedbacks (Leipper and Volgenau, 1972). It is defined as the integrated ocean heat content above the 26°C-isotherm and is for instance used by the National Oceanic and Atmospheric Administration (NOAA) for hurricane forecasts:

$$TCHP = \rho_o C_p \int_{h_{26}}^0 (T(z) - 26^\circ C) dz, \quad (1.4)$$

where ρ_o is the ocean density (typically 1025 kg/m³) and C_p is the heat capacity of the ocean (typically 3985 J/kg°C). The depth h_{26} marks the position of the 26°C isotherm. The value of 26°C is chosen as the threshold above which TCs are able to form (Gray, 1968). High values of TCHP mark regions that are favourable for TC development. Regions above 60 kJ/cm² are believed to support hurricane formation and above 120 kJ/cm² they undergo rapid intensification (Goni et al., 2009b). Wada and Usui (2007) examined TCs in the Western North Pacific and find a good correlation of accumulated TCHP (0.47 correlation coefficient) and TC intensity. This is a considerable improvement compared to accumulated SSTs (0.29 correlation coefficient).

Note however, that any temperature below 26°C will not make a contribution to TCHP. As consequence TCHP is not able to represent how far below 26°C subsurface temperatures may fall. Furthermore TCHP is a depth integrated quantity, therefore it is low in coastal areas where the ocean bottom shallows. Even for a mean temperature of 30°C, TCHP would therefore be small. Price (2009) argue that the upper ocean mean temperature may be more useful outside warm deep ocean conditions. They suggest a depth of 100 m, this being a typical mixing depth under TCs. For $T_{100} > 27^\circ$ and 75 kJ/cm², the TCHP and T_{100} are highly correlated. It is only in areas < 50 kJ/cm², that T_{100} becomes the more useful metric. A metric similar to TCHP is the interacting tropical cyclone heat potential (ITCHP) introduced by Buarque et al. (2009), which is the heat contained in the isothermal layer. Lloyd and Vecchi (2011) suggest to use the thermocline depth minus 2°C as an indicator for SST feedbacks.

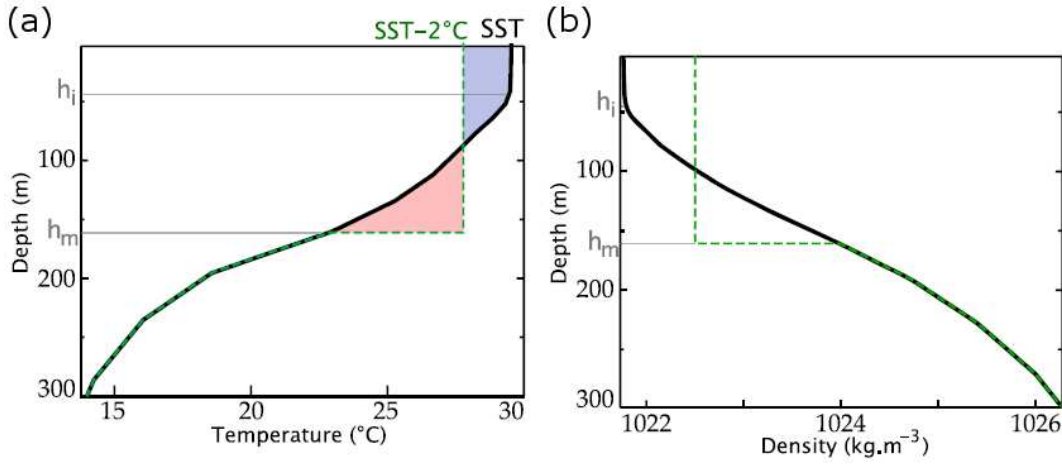


Figure 1.11: Typical tropical (a) temperature and (b) density profiles before storm passage (black) and after idealized heat and mass conserving mixing (dashed green line) used for CI calculation. Extracted from Vincent et al. (2012b)

1.3.4.2 The cooling inhibition index

Note that in reality the mixing depth is by no means fixed at 100 m or the thermocline. The depth will depend on the storm strength itself, but also on the ocean stability. In regions of high salt stratification, like the Amazon and Orinoco plume, the water column is stabilized and TCHP and T_{100} are not good representatives of the expected cooling feedback. In those cases Price (2009) suggests the use of a variable depth mean temperature (T_d), which takes ocean stratification into account by parametrizing a bulk Richardson number. In an alternative approach, Vincent et al. (2012b) develop the cooling inhibition index (CI) and the wind power index (WPI). They consider temperature and density stratification of the ocean, as well as storm strength. CI is defined as the cube root of the potential energy increase in the ocean corresponding to a 2°C surface cooling, induced by vertical mixing:

$$\Delta E_p(\Delta T) = \int_{h_m}^0 (\rho_f - \rho_i(z)) g z dz \quad (1.5)$$

$$CI = [\Delta E_p(-2^\circ C)]^{\frac{1}{3}} \quad (1.6)$$

Here the depth h_m is the depth to which the ocean needs to be mixed, in order to achieve a 2° surface cooling (Figure 1.11a). It is assumed that the the upper ocean temperature is constant after mixing and is equal to $SST + \Delta T$ in the ML. The initial unperturbed density profile is given by ρ_i , g is the acceleration of gravity, and ρ_f is the final homogeneous density profile after mixing (Figure 1.11b). CI explicitly takes

density stratification in the ocean into account and can be considered as the resistance of the ocean to vertical mixing. The larger CI, the less SST feedback is to be expected under TC passage. CI improves predicted cooling forecasts by about 45% compared to the heat content based metrics TCHP or ITCHP (Vincent et al., 2012b). CI performs similar to the thermocline depth minus 2°C suggested by Lloyd and Vecchi (2011) in a global sense, but keep in mind that this metric does not consider density stratification. It is therefore not useful within the Amazon and Orinoco plume.

Figure 1.12 shows the difference in CI between the standard CI defined in equations 1.5 and 1.6 compared to CI calculated at constant salinity stratification (33.85 PSU) derived by Neetu et al. (2012). The region of the Amazon and Orinoco plume is the area in the world, where salinity stratification is most important in altering TC induced SST feedbacks.

SST feedbacks also depend on the storm intensity. The dimensionless quantity WPI is a measure for TC strength and transfer of mixing energy to the ocean (Vincent et al., 2012b). It is formulated as the power dissipation along the TC track, normalized by a typical weak storm. The power dissipation (PD) is calculated at each track position (-3 days before (t_o) to +3 days after (t_c) TC passage):

$$PD = \int_{t_o}^{t_c} \rho C_D |\mathbf{V}|^3 dt, \quad (1.7)$$

and WPI is then defined as:

$$WPI = [PD/PD_0]^{1/3}, \quad (1.8)$$

where $\mathbf{V}$ is the local wind speed, C_D is the dimensionless surface drag coefficient and ρ is the surface air density. PD_0 is calculated for TCs with an average wind speed of 17 m/s and a translation speed of 7 m/s. Note that WPI considers cyclone wind speeds, translation speed and size, and therefore includes all TC properties that are relevant for ocean mixing. Typically tropical cyclones that reach hurricane strength have a WPI between 2.5 and 5.5 (Vincent et al., 2012b).

1.3.5 The Amazon and Orinoco plume and tropical cyclone intensity

The tropical North Atlantic hurricane season is from June to November, in which several TCs pass over the WTNA (Figure 1.1). Therefore the season of maximum

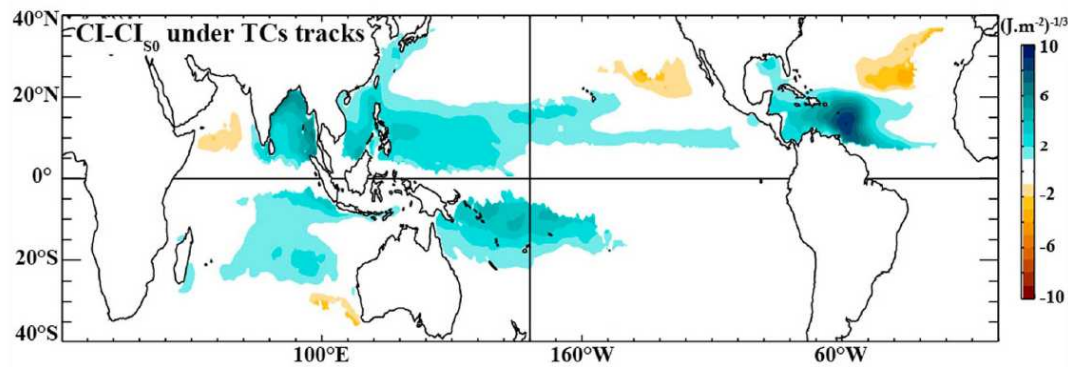


Figure 1.12: Spatial distribution of the difference between average CI underneath TC tracks minus CI at constant salinity calculated over the 1978-2007 period. Blue colours indicate areas where salinity stratification inhibits TC induced cooling. The black line shows the WPI=1 contour. Extracted from Neetu et al. (2012).

plume extent coincides with the season of maximum hurricane activity. Between 1960 and 2000 the WTNA witnessed about 225 TCs, of which about 59 crossed the plume (Ffield, 2007). Of the hurricanes that reached category 5, about 68% travelled over the plume area, raising the suspicion, that the rivers may enhance TC intensity (Ffield, 2007).

The stabilizing effect of freshwater plume along with enhanced surface temperatures has been proposed to reduce cooling feedbacks and therefore increase TC intensity (Balaguru et al., 2012b). In regions of temperature inversion (TI)s, the feedback may even turn out positive. Ffield (2007) attribute the enhanced TC activity to the warm plume anomalies (SSTAs). Using an atmosphere model, Vizy and Cook (2010) demonstrate that the warm anomalies increase the number of TCs by 60% and also increase TC intensity, but on the other hand reduce tropical storm days by about 12%. However, they neglect ocean-atmosphere interactions and potential changes in SST feedbacks. Furthermore they attribute all warm anomalies to the freshwater plume. Balaguru et al. (2012b) performed a global analysis of TC intensification in BL and non-BL regions. They find that the presence of BLs enhances TC intensification rates by 50% due to reduced cooling feedbacks. Grodsky et al. (2012) analysed the hurricane wake of Katia using satellite data, and found a reduction in SST cooling within the BL in the Amazon and Orinoco plume. On the basis of two case studies Wang et al. (2011) demonstrate that the formation of a post-monsoon BL leads to reduced surface cooling, and consequently increased TC intensity in the Northwest Pacific.

Ocean colour in the plume potentially changes ocean stratification and temperatures and may also have an impact on TCs. To this date, the only study that connected ocean colour to TC activity was performed by Gnanadesikan et al. (2010).

Based on their model study they suggest that TCs may become more intense, due to increased light attenuation warming the ocean surface. However, they neglect any colour-induced subsurface temperature changes, which may dampen the intensifying effect.

1.4 Thesis Plan

Using the Regional Ocean Modelling System (ROMS), this study investigates the impact of the clear and coloured Amazon and Orinoco river plumes on ocean stratification and temperatures. The potential effect on TC development is estimated applying simple relations between predicted cooling feedbacks and storm intensity.

The model equations and setup are introduced in Chapter 2. In Chapter 3 the freshwater effect of the Amazon and Orinoco plume on ocean stratification and temperatures is quantified. ROMS is setup for the WTNA and its performance is verified against observations. Comparing an experiment with and without river input, it will be investigated to what degree the BL and temperature inversions are caused by the freshwater plume. The freshwater effect on oceanic TC metrics, namely TCHP, 100m mean temperature, and CI, will also be assessed.

To quantify the impact of light absorbing particles, a novel light absorption algorithm (Lee et al., 2005) is implemented in ROMS in Chapter 4. Light absorption is obtained from satellite-derived surface optical properties. Thereby all absorbing particles can be considered, which is an advantage to previous chlorophyll-only based approaches. Finally the net impact of the coloured plume on the WTNA is assessed, by comparing ocean stratification, temperatures and oceanic TC metrics to a case without freshwater, or ocean colour. An alternative approach, where surface salinity serves as a proxy for light absorption, is also applied and compared to the more advanced method.

In Chapter 5 the potential impact of the coloured Amazon and Orinoco river plume on TC intensity is estimated. The expected surface cooling under TC passage is derived from ocean stability. Applying simple relations, SST cooling serves as a proxy for river-induced TC intensity changes. Thereby the freshwater, colour and the net effect of the coloured plume are assessed separately.

The results are summarized in Chapter 6.

Chapter 2

Model and data

2.1 The model equations

The Regional Ocean Modeling System (ROMS) is applied in this study (ROMS; COAWST svn 698). ROMS is a hydrostatic primitive equation ocean model and is further described in Shchepetkin and McWilliams (2005), Haidvogel et al. (2008) and Hedström (2009). ROMS has been applied in previous river and inner-shelf circulation studies (Hetland, 2005; Warner et al., 2005a; Choi and Wilkin, 2007; Denamiel et al., 2013; White and Toumi, 2014).

ROMS solves the Reynolds averaged Navier-Stokes equation using the hydrostatic and the Boussinesq approximation. The Boussinesq approximation assumes that density variations are small compared to the mean density ($\rho(x, y, z, t) = \rho_0 + \rho'(x, y, z, t)$, $\rho_0 \gg \rho'$) and may be neglected, except in the buoyancy force term of the vertical momentum equation (see Table 2.1 for the definition of the variables). Scaling of the vertical momentum equation, i.e. comparing the typical magnitudes of each term and neglecting small terms, gives the hydrostatic approximation. In the Reynolds approach, each term is split in their average part and the fluctuations around that average ($B(t) = \bar{B}(t) + B'(t)$), followed by averaging the whole equation.

The governing equations in Cartesian coordinates are the horizontal and vertical equations of motion:

$$\frac{\partial u}{\partial t} + \vec{v} \cdot \nabla u - f v = -\frac{\partial \Phi}{\partial x} - \frac{\partial}{\partial z} \left(\overline{u'w'} - v \frac{\partial u}{\partial z} \right) + F_u + D_u, \quad (2.1)$$

$$\frac{\partial v}{\partial t} + \vec{v} \cdot \nabla v - f u = -\frac{\partial \Phi}{\partial y} - \frac{\partial}{\partial z} \left(\overline{v'w'} - v \frac{\partial v}{\partial z} \right) + F_v + D_v, \quad (2.2)$$

$$\frac{\partial \Phi}{\partial z} = \frac{-\rho g}{\rho_0}, \quad (2.3)$$

with the continuity equation for an incompressible fluid,

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0. \quad (2.4)$$

The scalar transport in the model is given by

$$\frac{\partial C}{\partial t} + \vec{v} \cdot \nabla C = -\frac{\partial}{\partial z} \left(\overline{C'w'} - \nu_\theta \frac{\partial C}{\partial z} \right) + F_C + D_C, \quad (2.5)$$

and an equation of state:

$$\rho = \rho(T, S, P). \quad (2.6)$$

The equations are closed by parametrizing the Reynolds stresses and turbulent tracer fluxes as:

$$\overline{u'w'} = -K_M \frac{\partial u}{\partial z}; \overline{v'w'} = -K_M \frac{\partial v}{\partial z}; \overline{C'w'} = -K_C \frac{\partial C}{\partial z}, \quad (2.7)$$

which are constrained by vertical boundary conditions, determined by the surface wind stress and the bottom stress.

It is convenient to use stretched, terrain-following vertical σ -coordinates in ocean models. The dynamical equations and the boundary conditions are transformed according to $\sigma = \sigma(x, y, z)$, with $z = \zeta(1 + \sigma) + h_c\sigma + (h - h_c)C(\sigma)$, $-1 \leq \sigma \leq 0$, h_c is the minimum depth and $C(\sigma)$ defines the spacing and stretching of the vertical levels. It is a function of the bottom control parameter (θ_b , $(0 < \theta_b \leq 1)$) and the surface control parameter (θ_s , $0 < \theta_s \leq 20$). The larger θ_s , the higher the surface resolution, i.e. the more levels move closer to the surface.

In ROMS the Arakawa C grid is applied. The variables of state (density points) are centred in the middle of the grid box, whereas the velocities are calculated at the grid boundaries. The option to mask land areas is included in ROMS, the computations however will still be performed over all the domain. For reasons of efficiency, the ROMS time stepping is split explicit, i.e. the depth-integrated equations are treated on a shorter time-step (barotropic mode) than the full 3D-equations (baroclinic mode).

At the open boundaries the Flather (1976) conditions are used for the barotropic velocity with the corresponding Chapman (1985) conditions for the surface elevation, assuming that all outgoing signals leave at the shallow water wave speed. The currents and the tracers are imposed with the Orlanski (1976) radiation conditions, which compute a normal phase velocity to radiate out incoming and outgoing flow happening at the same boundary. This is modified by Raymond and Kuo (1984) to also account for the horizontal propagation. In addition, an eight-grid point wide nudging relaxation zone is used to relax the baroclinic structure (temperature, salin-

Variable	Definition
x, y	horizontal coordinates
z	vertical coordinate
t	time
u, v, w	the (x,y,z) components of vector velocity $\vec{v}$
$S(x, y, z, t)$	salinity
$T(x, y, z, t)$	potential temperature
$h(x, y)$	depth of sea floor below mean sea level
$\zeta(x, y, t)$	the surface elevation
$H_z(x, y, z)$	vertical grid spacing
ν, ν_Θ	molecular viscosity and diffusivity
$C(x, y, z, t)$	scalar quantity, i.e. temperature, salinity, nutrient concentration
D_u, D_v, D_C	optional horizontal diffusive terms
F_u, F_v, F_C	forcing/source terms
ν, ν_Θ	molecular viscosity and diffusivity
K_M, K_C	vertical eddy viscosity and diffusivity
P	total pressure $P \approx \rho_0 g z$
$\Phi(x, y, z, t)$	dynamic pressure $= P \rho_0$
$\rho_0 + \rho'(x, y, z, t)$	total in situ density
$f(x, y)$	Coriolis parameter
g	acceleration of gravity

Table 2.1: Variables used in the description of ROMS

ity, and velocity) towards the fields provided by the ocean climatology (Marchesiello et al., 2001). The normal velocity at the coastal wall is zero and the no slip condition is used for the tangential velocity. No flow normal to the topography is allowed. To ensure positive tracer (temperature and salinity) concentrations, the Multidimensional Positive Definite Advection Transport Algorithm (MPDATA) is used (Smolarkiewicz, 1983).

The generic length scale (GLS) scheme with the 'gen' parameters (Warner et al., 2005b) is applied for turbulence closure (Umlauf and Burchard, 2003). The scheme solves equations 2.7 by determining the turbulent kinetic energy and turbulent length scale. It has been shown to perform well in estuarine regions by Warner et al. (2005a), who obtain similar salinity distributions when applying the k- ϵ (Rodi, 1987), k- ω (Wilcox, 1988), and 'gen' scheme. Background mixing in the model is set to $10^{-5} \text{m}^2 \text{s}^{-1}$ for momentum and to $10^6 \text{m}^2 \text{s}^{-1}$ for tracers.

2.2 The model setup

The ROMS model is setup for the WTNA and the experiments are run for 11 years (2000 to 2010), of which only the last ten years will be analysed.

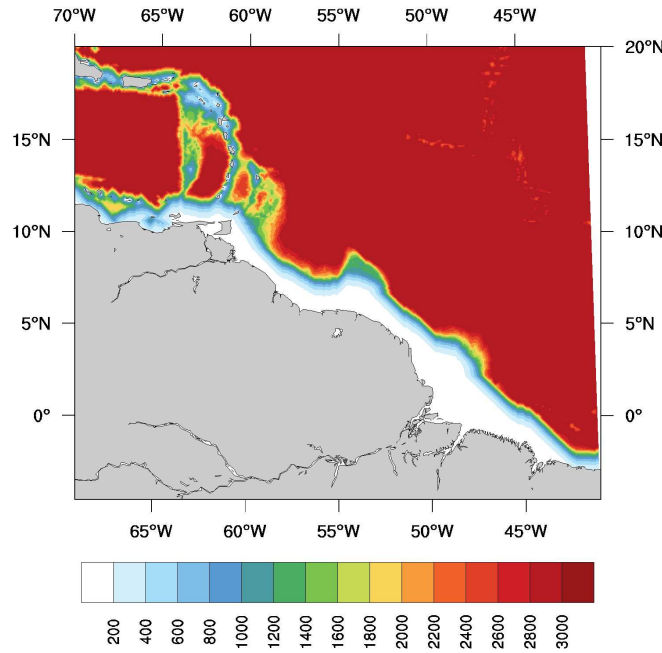


Figure 2.1: Bathymetry (m) in the western tropical North Atlantic

2.2.1 Model domain

The model domain includes the area of the Caribbean with the Lesser and the Greater Antilles as far as Puerto Rico, as well as the east coast of Venezuela and Brazil with the Amazon and Orinoco river mouths (Figure 2.1). For the present study a regular cartesian grid has been applied. The horizontal resolution of the grid is 10 km and covers the area 4.5°S - 21°N / 72.5°W - 41°W . The grid covers an area of about 2800 km x 3500 km and captures 3000 km of the northeastern South American coastline.

In the present study 50 layers have been applied in the vertical with a higher density of model levels near the ocean surface. The vertical stretching parameters are $\Theta_s=10$, $\Theta_b=0.2$, and $h_c=200\text{m}$. On average the top 26 levels are concentrated in the upper 500 meters of the ocean, 12 of which are in the upper 100 m. The topography of the ocean bottom is taken from Global Topography v14.1 (Smith and Sandwell, 1990) (Figure 2.1). The ocean bottom is shallow (< 200 m) on the South American shelf and has a very steep gradient approximately 200-400 km from the coast. In the open North Atlantic and the Caribbean the ocean is deep ($> 3000\text{m}$). In ROMS the bathymetry has been cut at 3000 m depth. The maximum depth of the original bathymetry is approximately 7000 m. Since the major currents of the area like the Caribbean current, the Guyana and the North Brazil current only extend down to 2000 m (Gyory et al., 2012a,b,c), this is not expected to affect the results.

2.2.2 Ocean boundary conditions

Open boundary and initial conditions (2000-2010) for temperature, salinity, three-dimensional velocity fields and non-tidal sea level were generated from the monthly mean $1/4^\circ$ Mercator Ocean (GLORYS2V3) global ocean reanalysis (Lellouche et al., 2013), obtained from the MyOcean database. The reanalysis is performed with the NEMOv3.1 ocean model in the ORCA025_LIM setup and assimilated with observations. Temperature and salinity are bias corrected. The assimilated observations include satellite derived SSTs and sea level, temperature and salinity from profiling floats and surface drifters, and surface velocities from the Global Drifter program.

Tides are an important processes in mixing riverine with oceanic waters. Tides are changes in the sea surface elevation caused by the gravitational force of the moon and the sun and by the rotation of the earth-moon/earth-sun system around their common centre of gravity. The variability of the earth position towards the sun and the moon is responsible for different tidal phases, with frequencies ranging from several hours to days and years (Ross, 1995). The tidal forcing in ROMS is applied at the open boundaries and affects the sea surface elevation and the ocean currents. It is derived from the 11 tidal components, extracted from the $1/12^\circ$ resolution Atlantic Ocean Atlas Solution (OSU Tidal Data Inversion, Egbert et al. (1994)). The dominating tidal constituent in the model region is the M2 component, also known as the lunar semi-diurnal tide with a period of approximately 12 h 25 min (Ross, 1995).

2.2.3 Atmospheric boundary conditions

The ocean surface is influenced by the surrounding atmosphere in form of winds, air temperature and humidity, atmospheric pressure as well as precipitation, cloud coverage and radiation. In the model setup used here bulk formulae are used to calculate the air-sea heat and momentum fluxes (Fairall et al., 1996). The National Center for Environmental Prediction (NCEP) Climate Forecast System Reanalysis (CFSR) (Saha et al., 2010) is used for the atmospheric boundary conditions (2000-2010). This dataset consists of data derived from a global high resolution, coupled ocean-atmosphere land-surface sea-ice system and contains data from 1979 to present day. The variables used here are 10 m winds, air temperature and specific humidity at 2m, the precipitation rate, downward longwave and shortwave radiation at the ground, and total cloud coverage. These variables are available at a six hourly temporal resolution and a horizontal resolution of $0.312^\circ \times 0.312^\circ$. The atmospheric pressure at mean sea level is provided in $0.5^\circ \times 0.5^\circ$ horizontal resolution.

It is important to be aware of biases in the reanalysis data, which are assessed in

Wang et al. (2010). The CFSR dataset is too wet compared to observations in the WTNA by up to 4 mm/day in the annual mean. This may lead to an overestimate of surface freshening in the WTNA and amplify the low-salinity signal of the freshwater plume. Cloudiness in CFSR is up to 20% too low compared to observations in the WTNA, leading to an overestimation of downward shortwave flux at the ocean surface (up to 20 W/m²). Consequently the upward latent heat flux is overestimated by a similar amount. The combination of too much incoming radiation and too little outgoing heat flux may lead to an excess of surface heating. Xue et al. (2010) suggest that the CFSR surface winds are too strong in winter, potentially producing too much mixing and upwelling in the ocean.

2.2.4 Amazon and Orinoco discharge

River input in ROMS is realized in the form of freshwater sources at the river mouths. The mean annual cycle of Amazon and Orinoco discharge is derived from the Global Runoff Data Center (GRDC, 2001), which provides monthly discharge measured at various stations along the rivers. River temperature has been estimated from monthly CFSR SST at the river mouths, which tends to be slightly cooler compared to the adjacent ocean. As will be shown later, the choice of river input temperature does not affect open ocean temperatures.

2.2.5 Ocean colour data

In Chapter 4 ocean colour is derived from the remote sensing reflectance at various wavelengths. Combined monthly fields of R_{rs} at 9 km horizontal resolution from Terra and Aqua MODIS at 443 nm, 488 nm, 555 nm and 667 nm from 2000 - 2010 were obtained (Feldman and McClain, 2014). MODIS is the moderate-resolution imaging spectroradiometer aboard the Terra (EOS AM) and Aqua (EOS PM) satellites, which pass over the equator at different times of day and are viewing the entire Earth's surface every 1 to 2 days, acquiring data at various spectral bands. Hu et al. (2013) estimate that the accuracy of MODIS R_{rs} is better than 5% in the blue bands, and between about 10-15% in the green part of the spectrum. The errors further grow towards red light, possibly due to imperfect atmospheric correction.

2.3 Validation data

Surface salinity in the ROMS model is compared to satellite derived surface salinity from the Aquarius instrument onboard the Aquarius/SAC-D satellite (Level 3, V2.0;

Lagerloef (2012)). Aquarius is a collaborative effort between NASA and the Argentinian Space Agency Comision Nacional de Actividades Espaciales (CONAE). The retrieved monthly SSS data is bias-adjusted and available at $1^\circ \times 1^\circ$ resolution. Aquarius salinity data are available from September 2011 to present day. The data are more prone to errors in high latitudes, than in the tropics (Lagerloef, 2013). The precision of Aquarius is 0.2 PSU. Aquarius surface salinity has been validated by Grodsky et al. (2012) against in-situ measurements in the Amazon and Orinoco plume region. They find that Aquarius salinity has a small negative bias (-0.1 PSU), with the error being slightly larger in low-salinity regimes. Due to the coarse horizontal resolution, there are no salinity data available near the coast. Therefore the data has been extrapolated in this study.

Satellite observed SSTs (Figure 3.4b) are produced daily by the Group for High Resolution Sea Surface Temperature (GHRSSST) at the National Oceanic and Atmospheric Administration (NOAA), National Oceanographic Data Center (NODC) using optimal interpolation from Advanced Very High Resolution Radiometer (AVHRR) Pathfinder Version 5 data (Reynolds et al., 2002, 2007). The horizontal resolution is $0.25^\circ \times 0.25^\circ$ and the dataset is bias adjusted using in-situ data. However, small biases may remain, but are low in the Amazon and Orinoco plume region (Reynolds et al., 2002, 2007).

Observed currents are the monthly $1/3^\circ$ resolution surface currents from the NOAA Ocean-Currents-Analyses-Real-time (OSCAR) available online at <http://www.oscar.noaa.gov/>. OSCAR estimates the ocean currents from satellite derived sea surface height, scatterometer winds and sea surface temperature (Johnson et al., 2007). The data are updated every 6 days and start in 1992. Johnson et al. (2007) validated OSCAR against in-situ measurements and conclude that OSCAR provides accurate estimates of the mean ocean circulation. In the near-equatorial regions the variability of zonal currents is also well captured.

Observational profiles have been obtained from the World Ocean Atlas 2013 (WOA13), which contains temperature and salinity profiles from 1955-2012 (Locarnini et al., 2013; Zweng et al., 2013). Analyses are produced at 102 standard depth levels between the surface and 5500 m depth and are available at $1/4^\circ$ horizontal resolution. The data sources consist of historical oceanographic temperature profile data from bottle samples, Mechanical Bathymetry-thermographs (MBT), ship-deployed Conductivity-Temperature-Depth (CTD) packages, Digital Bathythermograph (DBT), Expendable Bathythermographs (XBT), profiling floats, moored and drifting buoys, gliders, and undulating oceanographic recorder (UOR).

Model Experiments				
	No-River	River (Clear River)	Colour (Coloured River)	Water Type (Salinity- approximated ocean colour)
River Input	No freshwater input from the Amazon and Orinoco	Freshwater input from the Amazon and Orinoco	Freshwater input from the Amazon and Orinoco	Freshwater input from the Amazon and Orinoco
Light absorption	Paulson and Simpson (1977) radiation formulation with constant Jerlov (1976) water type 1 (clear water)	Paulson and Simpson (1977) radiation formulation with constant Jerlov (1976) water type 1 (clear water)	Lee et al. (2005) radiation formulation with satellite derived optical properties (colored water)	Paulson and Simpson (1977) radiation formulation with variable Jerlov (1976) water type as a function of surface salinity

Table 2.2: Description of model experiments performed in this study

2.4 Model experiments

Four different experiments have been performed in the setup described above for 11 model years (2000-2010), of which only the last 10 years are analysed, to leave the model time to adjust. The first experiment is a No-River experiment, where the Amazon and Orinoco do not discharge into the WTNA and the ocean is assumed to be clear. To assess the impact of freshwater on the ocean surface, a (Clear-)River experiment has been performed, where river discharge is included, but light absorbing particles are still neglected. In the Water-Type and Coloured-River experiments freshwater input and realistic light absorption are considered via two different methods: by deriving ocean colour from surface salinity and by using satellite obtained ocean colour and the Lee et al. (2005) algorithm. The experiments are summarized in Table 2.2.

Chapter 3

The freshwater effect of the Amazon and Orinoco plume

3.1 Introduction

The Amazon and Orinoco freshwater plume extends several thousand kilometres into the tropical North Atlantic (Hu et al., 2004) and witnesses several tropical cyclones (TCs) each year (Ffield, 2007). The low-density water reduces vertical mixing and forms a stable layer on the ocean surface, potentially increasing ocean temperatures and stratification (Sprintall and Tomczak, 1992; Pailler et al., 1999). Since TCs gain their energy from the ocean, river-induced changes have been predicted to influence storm intensities (Ffield, 2007; Vizy and Cook, 2010; Balaguru et al., 2012a)

While there is agreement on the rivers enhancing the barrier layer (BL) between the mixed layer (ML) and the top of the thermocline (Pailler et al., 1999; Masson and Delecluse, 2001; Balaguru et al., 2012b), their influence on the sea surface temperature (SST) and the upper ocean heat content is more controversial. BLs can modify ocean temperatures by isolating the surface layer from solar heating and entrainment cooling (Vialard and Delecluse, 1998). Ffield (2007) associate observed SST anomalies in the river plume with the BL (Ffield, 2007). Wang et al. (2006) and Vizy and Cook (2010) show that the elevated warm pool temperatures may increase TC activity. Masson and Delecluse (2001) and Balaguru et al. (2012b) on the other hand found no surface warming due to Amazon freshening in their model studies. However, BLs increase subsurface temperatures by decoupling the BL from surface cooling (Vialard and Delecluse, 1998). Temperature inversions of up to 1.0°C have been found to accompany the BL in the tropical North Atlantic (de Boyer Montégut et al., 2004). Elevated BL temperatures in combination with increased stratification could intensify TCs by reducing self-induced cooling feedbacks (Balaguru et al., 2012a).

In this Section the impact of the Amazon and Orinoco on the western tropical North Atlantic (WTNA) is quantified by comparing two model experiments: One containing river discharge (River experiment), and one without rivers (No-River experiment) (Table 2.2). The Amazon and Orinoco discharge applied in the model is described in Section 3.2 and the River experiment is validated against observations in Section 3.3.1. The impact of freshwater discharge on ocean temperatures and stratification is assessed in Section 3.3.2. Riverine influences on the oceanic control of TC intensity are derived from a comparison of heat content (tropical cyclone heat potential (TCHP)) and the resistance of the ocean to surface cooling (cooling inhibition index (CI)) in both experiments. The results are discussed in Section 3.4.

3.2 The Amazon and Orinoco river discharge

The mean annual cycle of Amazon discharge (Figure 3.1) has been calculated from water levels measured at Obidos from 1982-1947 and 1970-1997 (GRDC, 2001). Obidos is located about 750 km upstream the Amazon delta. The discharge is corrected for the Tapajos and Xingu, which join the Amazon 50 km and 300 km downstream Obidos, respectively (Figure 1.4a). The Tocantins is another major river with a yearly discharge that is almost a third of the Orinoco. It enters the North Atlantic close to the Amazon mouth and is usually assigned to the Amazon discharge. Amazon and Tocantins release about $200,000 \text{ m}^3/\text{s}$ freshwater into the tropical North Atlantic. The discharge is maximum in May and minimum in November. The interannual variability of the Amazon discharge can be very large. The maximum annual mean discharge was $215,648 \text{ m}^3/\text{s}$ in 1989 and the minimum $140,973 \text{ m}^3/\text{s}$ in 1997, which is a variation of up to 30% from the mean. In the model the freshwater discharge is distributed equally over five channels spread along the Amazon delta.

The mean annual cycle of the Orinoco discharge (Figure 3.1) was measured from 1923 to 1989 at the gauging station Puente Angostura, located about 350 km upstream the Orinoco delta. The discharge is corrected by about 10% due to the Rio Caroni which feeds the Orinoco from the south, approximately 100 km downstream Puente Angostura (Castellanos et al., 2010) (Figure 1.4b). Mean discharge over the 64 years of data collection has been about $30,000 \text{ m}^3/\text{s}$ the minimum annual mean discharge being $21,554 \text{ m}^3/\text{s}$ in 1924, and the maximum $37,620 \text{ m}^3/\text{s}$ in 1953. The water level is maximum in August and minimum in March. Following Warne et al. (2002), the Orinoco discharge in the model is distributed across six channels, where about 85% of freshwater flows through the northernmost arm (Figure 1.4c).

Only little human impact has happened along the Amazon and Orinoco river, so

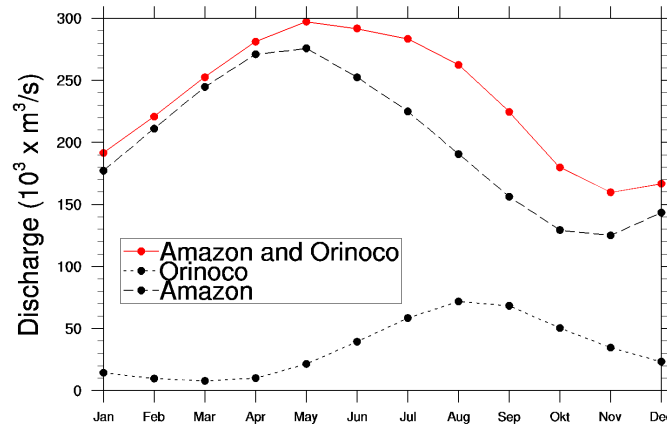


Figure 3.1: Mean annual cycle of Amazon, Orinoco, and their combined discharge derived from observations.

that there is no visible trend in the annual discharge. Amazon and Orinoco seasonal cycles are very different, since the main arm of the Amazon is along the equator, while the Orinoco flows between 5°N and 10°N. The Amazon is therefore influenced by both, northern and southern hemisphere tributaries. With the ITCZ shifting its position seasonally, the wet season in the northern and southern tributaries is out of phase by 6 months. Despite single tributaries varying in discharge by a factor of 10 seasonally, this is somewhat attenuated in the annual cycle of the Amazon discharge. The combined annual mean cycle of Amazon and Orinoco is dominated by the Amazon, but in summer and autumn the Orinoco contributes up to about 30% percent to the total.

3.3 Results

The River and No-River experiments (Table 2.2) are run for 11 years from 2000 to 2010. Only the last 10 years are analysed to leave the model time to adjust. The analysis is presented for the seasonal means, namely winter (December, January, February (DJF)), spring (March, April, May (MAM)), summer (June, July, August (JJA)), and autumn (September, October, November (SON)).

3.3.1 Validation: Modelling the western tropical North Atlantic

3.3.1.1 Surface salinity, temperature, and currents

The seasonal cycle of surface salinity in the River experiment and observations is shown in Figure 3.2. The plume (defined by the 35 PSU salinity contour) is small in

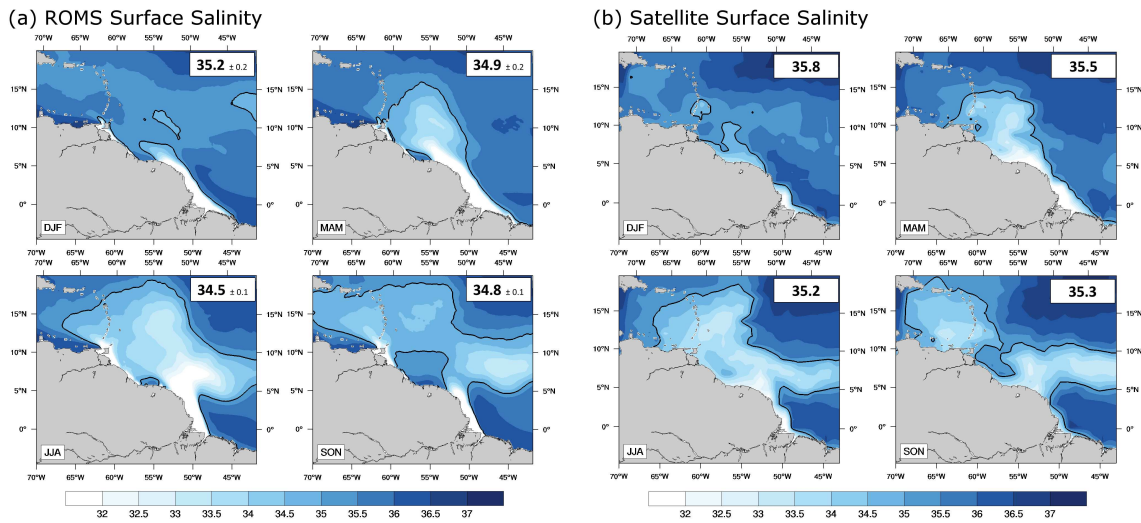


Figure 3.2: Maps of the WTNA mean seasonal cycle (DJF, MAM, JJA, SON) of (a) ROMS sea surface salinity (PSU) from 2001-2010 and (b) satellite observed surface salinity for 2012. The black line marks the 35 PSU contour. The respective area mean values are given in the upper right corner of each map.

winter and starts spreading out into the tropical North Atlantic in spring, extending to 15°N. Following the peak of freshwater discharge in May, the plume grows and extends from the coast to 18°N, spanning from 65°W to 42°W in summer. That leads to a distinct freshwater signal in the Caribbean, which peaks in autumn. A large proportion of the freshwater plume is transported east with the North Brazil Counter Current (NBCC) in summer and autumn. While there is one well defined large plume in summer, it separates into two parts in autumn, one travelling east and the other northwestward. Satellite observed surface salinity for 2012 from Aquarius shows a similar spatial extent and seasonal cycle as the model. The feature of two major plume parts in autumn is also visible in the observations. However, satellite observed mean salinities are slightly higher compared to the simulation. The modelled mean annual cycle of the plume area (defined by the 35 PSU salinity contour (Hu et al., 2004)) is shown in Figure 3.3 and is in good agreement with observed plume extent. It peaks in August, covering an area of $2.8 \times 10^6 \text{ km}^2$ in the model. The plume starts to slowly diminish in autumn until it rapidly declines from November, to reach a minimum in February ($0.9 \times 10^6 \text{ km}^2$). The maximum plume size lags maximum discharge by about 3 month, in agreement with Moller et al. (2010). The plume area is about 20% smaller in the observations from August to October.

Modelled mean SSTs are high in summer and autumn and lower in winter and spring (Figure 3.4a). Generally SSTs are higher in the southern part of the model domain and colder in the north, the difference being about 2°C to 3°C. The north-south

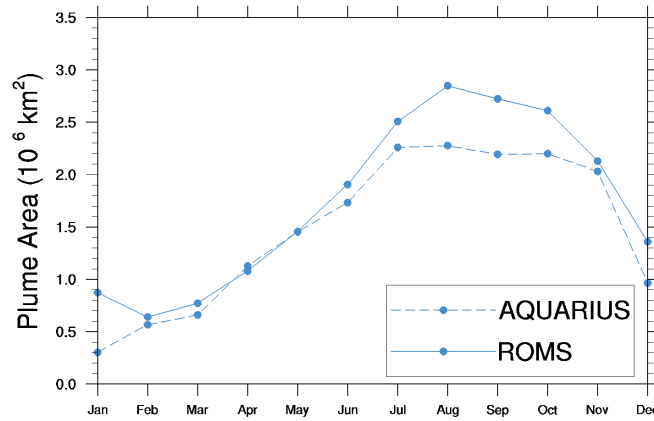


Figure 3.3: Annual mean cycle of area mean plume size (defined by the 35 PSU salinity criterion) in the WTNA in the ROMS simulation (2000-2010) and in Aquarius observations (2012).

gradient is larger in winter and spring, compared to summer and autumn. The highest temperatures ($>30^{\circ}\text{C}$) occur during autumn between 5°N and 15°N . This feature is part of the tropical western hemisphere warm pool, usually defined by temperatures above 28.5°C (Wang and Enfield, 2001). SSTs in ROMS are slightly too high compared to satellite observations (GHRSSST) throughout all seasons, the mean bias being 0.5°C . The bias is largest (up to 1°C) close to the South American coastline, within the North Brazil current. The shape and seasonal variation of the warm pool however, is very well captured by the model. The timeseries of monthly mean surface temperatures in the model area is in very good agreement with observations (Figure 3.5). Timing and magnitude of maximum and minimum SSTs are represented well in the model.

The ocean currents play a crucial role in distributing riverine freshwater across the tropical North Atlantic. The seasonal cycle of modelled and satellite-observed (OSCAR) surface currents are shown in Figure 3.6. The main oceanic features of the region are the North Brazil Current (NBC), the Guyana current, the Caribbean current and the North Brazil Counter Current (NBCC). The NBC carries warm water of South Atlantic origin northwestward along the coast of Brazil, across the equator and into the Northern hemisphere. The coastal NBC is strong throughout the whole year. Near 6°N - 8°N a part of the NBC pinches off from the South American coastline and turns to the east, to feed the NBCC. The strength of the retroflection changes seasonally, and is strongest during summer and autumn. Occasionally the retroflection generates large isolated warm core rings (not shown). The anticyclonic rings move northwestward along the South American coastline towards the Caribbean. The part of the

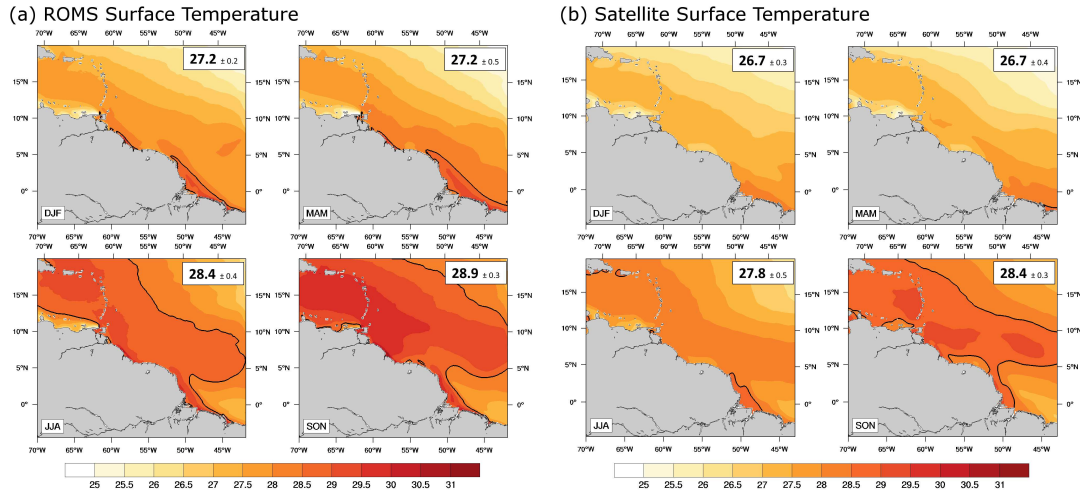


Figure 3.4: Maps of the WTNA mean seasonal cycle (DJF, MAM, JJA, SON) of (a) ROMS SST ($^{\circ}\text{C}$) from 2001-2010 and (b) satellite observed GHRSSST SST ($^{\circ}\text{C}$) for the same time period. The black line marks the warm pool defined by the 28.5°C temperature criterion (Wang and Enfield, 2001). The respective area mean values and their standard deviations are given in the top right corner of each map.

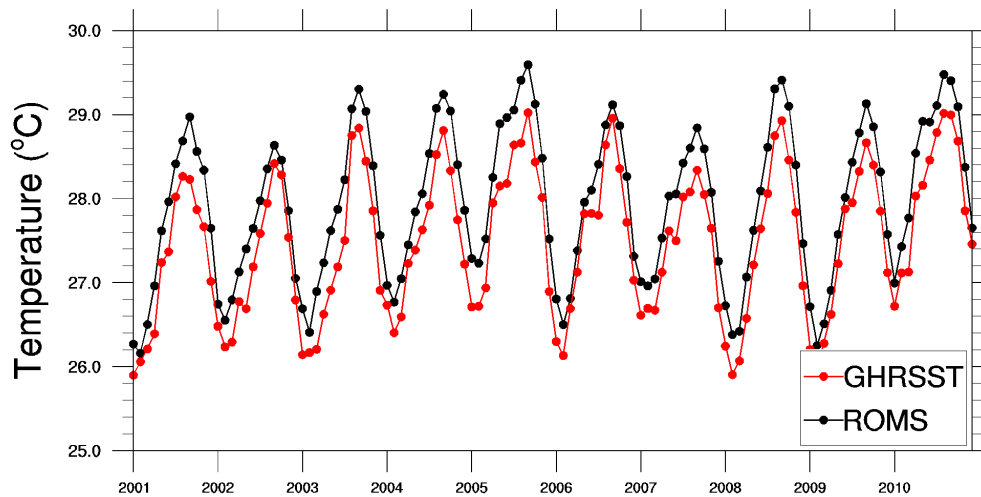


Figure 3.5: Timeseries (2001-2010) of model domain mean ROMS (black) SST ($^{\circ}\text{C}$) and satellite observed GHRSSST (red) SST ($^{\circ}\text{C}$) in the WTNA.

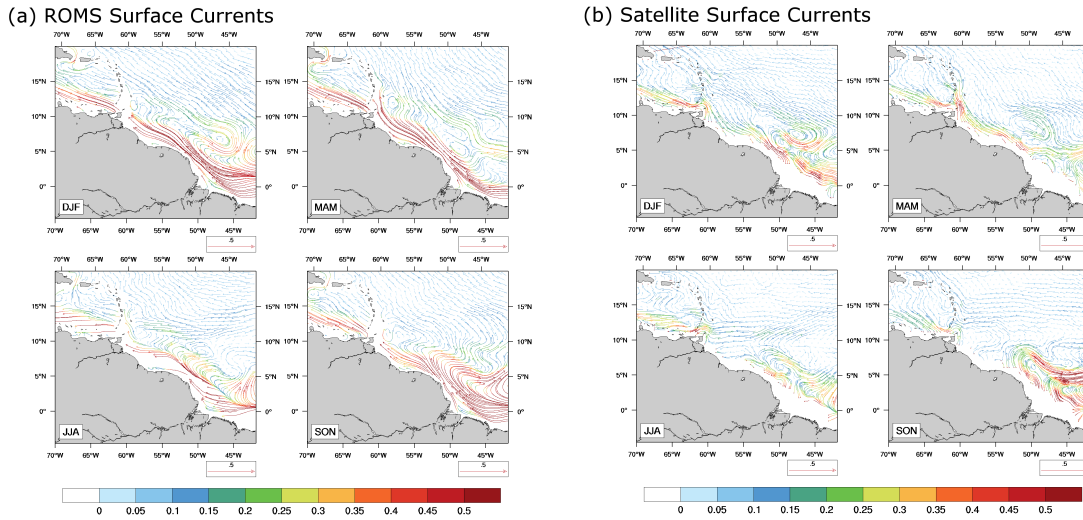


Figure 3.6: Maps of WTNA mean seasonal cycle (DJF, MAM, JJA, SON) of (a) modelled ROMS surface currents (m/s) from 2001-2010 and (b) satellite observed (OSCAR) surface currents (m/s) for the same time period.

NBC that continues to follow the South American coastline feeds the Guyana current, a coastal current that flows northwestward towards the Caribbean Sea. The water enters the Caribbean through the island channels, from where it is transported further westward by the Caribbean current along the southern coast, with the highest velocities in winter and spring. The ROMS surface currents are in good general agreement with OSCAR in terms of direction and current systems. However, the ROMS currents, especially the westward flowing coastal components, are stronger than OSCAR. There are no OSCAR data available on the shelf, where the ROMS currents are strongest. In agreement with the model, other observations indicate that these coastal regions contain the strongest currents (Gyory et al., 2012a,b,c).

3.3.1.2 Ocean stratification

The area mean seasonal cycle of salinity and temperature profiles and corresponding isothermal layer (IL) and ML depths in model and observations (WOA13) are shown in Figure 3.7. The top of the thermocline is marked by a 0.2°C temperature drop from 10 m, and the ML depth is defined as a density increase corresponding to the same temperature change (de Boyer Montégut et al., 2004). The BL lies between the ML and the top of the thermocline. The model captures the characteristic of the temperature profile fairly well. Throughout all seasons the mixed layer temperature is overestimated by about 0.5°C , in accordance with the SST bias. The model slightly underestimates salinity in the ML (by up to 0.5 PSU) and the salinity gradient is

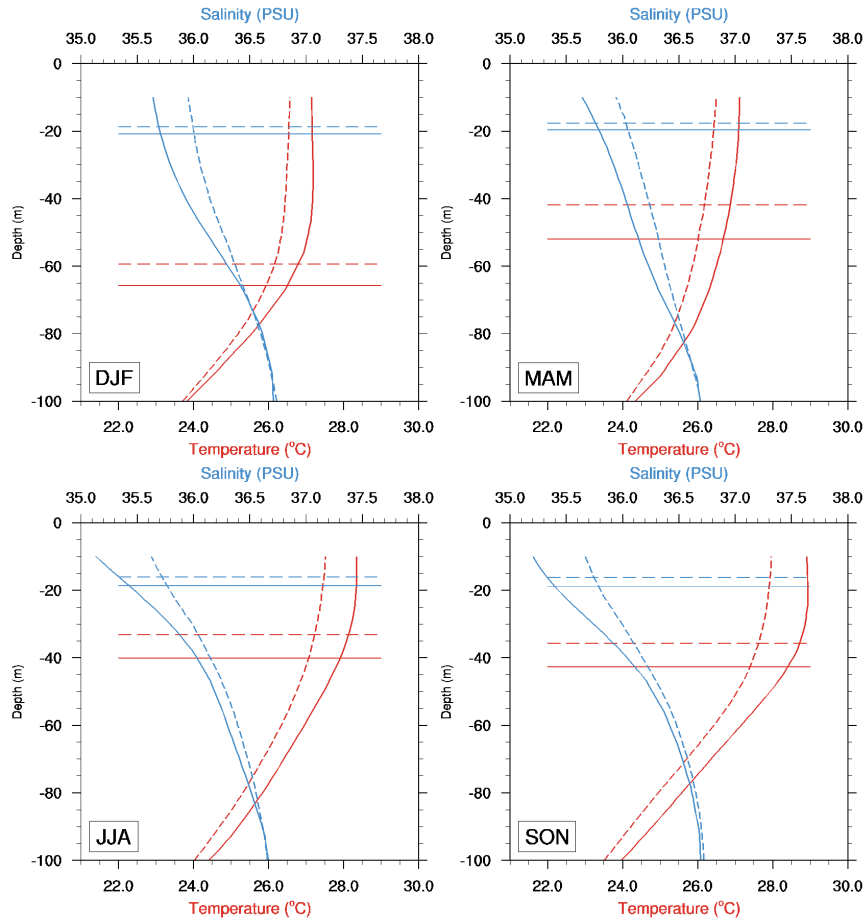


Figure 3.7: Model domain mean seasonal cycle (DJF, MAM, JJA, SON) of vertical temperature ($^{\circ}\text{C}$, red) and salinity profiles (PSU, blue) in the ROMS simulation (2001-2010, solid lines) and WOA13 (1955-2012, dashed lines) data. The horizontal lines mark the corresponding top of the thermocline (m, red) and ML depth (m, blue), defined in Section 1.2.1.

stronger compared to the observations. As a consequence salinity below the IL agrees well in model and observations. The modelled IL is deepest in winter (-66 m) and shallower in spring (-53 m), summer (-40 m) and autumn (-43 m). Compared to observations it is too deep by about 5 m in winter, summer and autumn, and by about 10 m in spring. The modelled ML depth varies hardly with the season, and is about 18 m throughout the year. The ML is about 2-3 m too deep in the model compared to WOA13. The slight overestimation of ML depth in combination with IL being too deep, leads to an overestimation of BL thickness in the model (up to 13 m in spring and about 8 m in the remaining year).

Seasonal maps of the IL depth and BL thickness in ROMS, and derived from WOA13 data are shown in Figure 3.8. The modelled IL is deepest in the Caribbean and in the southeastern WTNA, where the NBC enters the model domain. It deep-

ens substantially in winter, when the ocean surface starts to cool. In the model the isothermal layer can than be up to 100 m in the northern Caribbean and the NBC, and is up to 60-90 m deep north of 10°N. During spring the ocean restratifies, leading to a shoaling of the thermocline. The spring isothermal layer is up to 70 m in the Caribbean and up to 80 m deep in the northeastern WTNA. In summer and autumn, the isothermal layer is shallower (about 40 m on average). The IL is shallow throughout the year (below 20 m) along the upwelling regions near the coast (in the Caribbean and north-west of the Amazon mouth). Seasonal cycle and spatial distribution are fairly similar in model and observations, but the IL can be too deep in ROMS, particularly within and east of the Caribbean (up to 30 m locally). In the hurricane season (June to November) ROMS overestimates the IL depth in the entire domain by about 5 m on average.

Since the ML depth is fairly constant throughout the year, the BL thickness roughly follows the seasonal cycle of IL depth, but is smaller in magnitude. The BL is thickest in winter (in the model up to 80 m in the Caribbean, and 40-60 m north of 15°N). It shallows with the shoaling of the IL in spring and is minimum from June to November (about 25 m on average). In summer and autumn the BL can be up to 50 m thick in the Caribbean and up to 40 m east of the Caribbean. Between 50°W-40°W the ML is between 10 m and 30 m thick in the hurricane season. Along the shallow South American shelf, no BL is maintained. Due to the bias in the IL depth the modelled BL is too thick, particularly in the northern model domain in spring (up to 30 m locally). The signature of the plume is not obvious in the spatial distribution of the BL. Note that other factors like rainfall, evaporation, mixing and subduction of high salinity waters also play a role in determining BL thickness.

3.3.1.3 Tropical cyclone metrics

Of relevance for TC development is the heat contained in the upper ocean. The annual and area mean tropical cyclone heat potential (TCHP) is 48.2 kJ/cm². The spatial distribution is roughly a reflection of the warm SSTs (Figure 3.9a). Generally, TCHP is larger towards the coast where SSTs are higher. However, the heat content does not only depend on ocean temperatures, but also on the depth of the isothermal layer. The IL is deepest in the Caribbean (up to 100 m in winter, and up to 60 m in summer), leading to high TCHPs (up to 80 kJ/cm² and 120 kJ/cm²). Ocean temperatures are lower in winter and spring resulting in less heat. Since the isothermal layer is deeper in that season, the effect of the colder ocean is damped. In summer, when the ocean is warmed, TCHP increases. The spatial distribution and seasonal cycle of

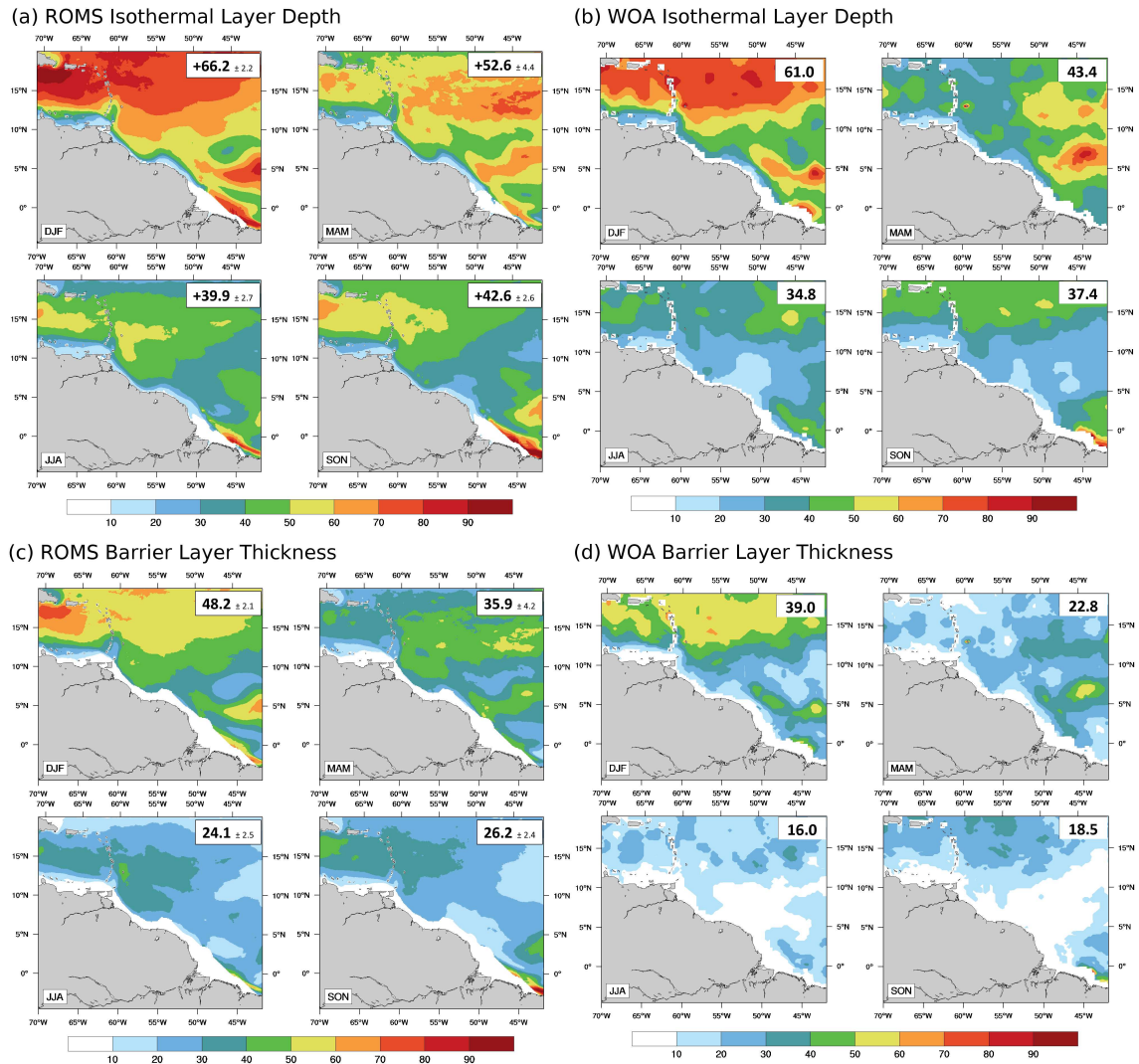


Figure 3.8: Maps of the WTNA mean seasonal cycle (DJF, MAM, JJA, SON) of the IL depth (m) in (a) the ROMS model (2001-2010) and (b) WOA13 (1955-2012) observations, and BL thickness (m) in (c) the ROMS model and (d) WOA13 observations. The definitions of IL depth and BL thickness are given in Section 1.2.1. The respective area mean values and their standard deviations are given in the top right corner of each map.

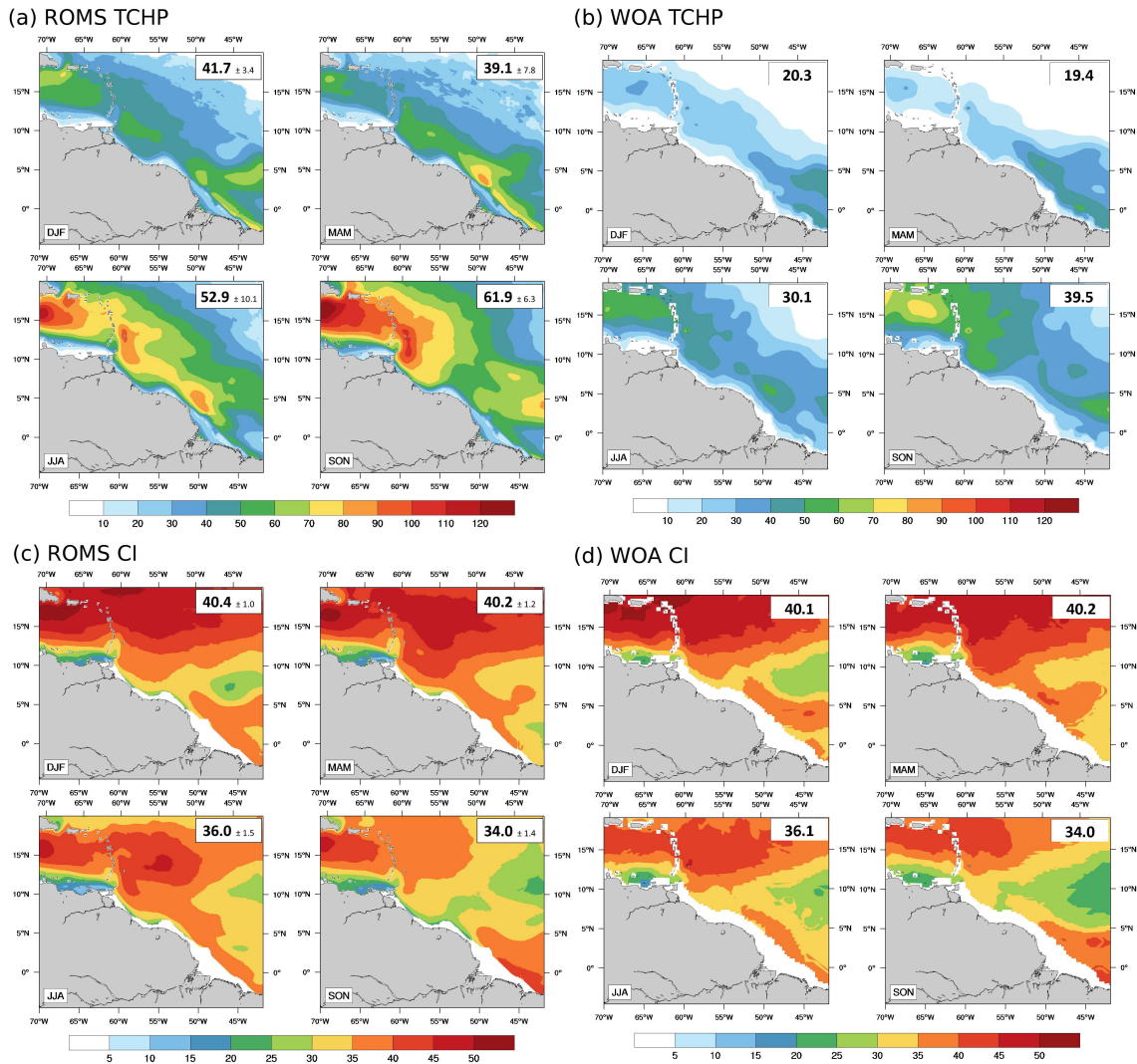


Figure 3.9: Maps of the WTNA mean seasonal cycle (DJF, MAM, JJA, SON) of tropical cyclone heat potential (TCHP) (kJ/cm^2) in (a) the ROMS model (2001-2012) and (b) derived from WOA13 (1955-2012) observations and cooling inhibition index (CI) ($(\text{J}/\text{m}^2)^{1/3}$) in (c) the ROMS model and (d) derived from WOA13 observations. The definitions of TCHP and CI are given in Section 1.3.4. The respective area mean values and their standard deviations are given in the top right corner of each map.

TCHP in ROMS is similar to TCHP maps (Figure 3.9b), derived from WOA13 data. The magnitude however, is substantially too high in the model (about $20 \text{ kJ}/\text{cm}^2$ on average throughout the entire year). Locally modelled TCHP can even be $40 \text{ kJ}/\text{cm}^2$ higher compared to observations (e.g. in the hurricane season in the Caribbean).

The cooling inhibition index (CI) measures the resistance of the ocean to vertical mixing and is a useful metric for SST feedbacks, especially where salinity stratification is important (Vincent et al., 2012b). Similar to TCHP, CI is large when the isothermal layer is very thick, since more energy is needed to mix colder water into the ML (Fig-

ure 3.9c). However, it also depends on the density stratification and grows with strong freshwater signals on the ocean surface. The annual mean CI in the model domain is $37.6 \text{ (J/m}^2\text{)}^{\frac{1}{3}}$, in the hurricane season (June-November) it is only $34.5 \text{ (J/m}^2\text{)}^{\frac{1}{3}}$. CI is lower near the South American coastline and largest in northern part of the model domain where the IL can be very deep (up to $50 \text{ (J/m}^2\text{)}^{\frac{1}{3}}$). Maximum values of CI are also reached in the western Caribbean. In the upwelling regions of the southern Caribbean and in the eastern model domain cold waters are nearer the surface, leading to low CI. The spatial distribution and seasonal cycle of modelled CI is very similar to the observed values WOA13 data (Figure 3.9d). In the hurricane season modelled CI is slightly too high in the western Caribbean (by up to $5 \text{ (J/m}^2\text{)}^{\frac{1}{3}}$), and slightly too low in the eastern model domain (by up to $-5 \text{ (J/m}^2\text{)}^{\frac{1}{3}}$).

3.3.2 The freshwater effect

To quantify the impact of the Amazon and Orinoco plume on ocean stratification and temperatures, the River experiment is compared to the No-River case in the following. Three main offshore plume areas have been identified based on Figure 3.10, since they are subject to large, seasonally varying freshwater signals. The areas are the Northern Plume (10°N - 20°N / 60°W - 50°W), the Eastern Plume (5°N - 10°N / 55°W - 45°W) and the Caribbean Plume (12°N - 17°N / 70°W - 62°W) (Figure 3.10). Mean changes for the model domain, and the three plume areas are given in Table 3.1.

3.3.2.1 Surface salinity and temperatures

As expected, surface salinity decreases significantly in the River experiment (Figure 3.10a). The plume spans almost across the entire model domain throughout the whole year. The decrease is consistently large near the river outlets (≥ -5 PSU), particularly in their respective season of maximum discharge (Figure 3.1). Away from the mouths, the salinity signal is strongest north of Brazil in spring and summer (up to -3.5 PSU). In summer and autumn low salinity water is also advected into the Caribbean. In summer and spring a large amount of freshwater is transported east by the North Brazil counter current. The northward transport of freshwater in winter leads to low salinity signals north of 15°N , within and east of the Caribbean. Plume extent and magnitude are largest from June to November, the North Atlantic hurricane season. Note that the 35 PSU salinity criterion applied in Figure 3.2 does not show a widespread freshwater plume from December to March. However, defining the plume area by a surface salinity decrease of -1 PSU, reveals that the river impact on surface salinity is still important in those months. The shape of the annual cycle

from the 35 PSU salinity criterion and the -1 PSU criterion are very similar, with the minimum in February and the maximum in August (Figure 3.11). However, under the -1 PSU threshold, the plume area is about 40% larger in winter and about 5% larger in summer, compared to an absolute salinity contour.

The freshwater input decreases SSTs significantly in the Amazon and Orinoco river mouths (Figure 3.10 b). A sensitivity experiment with variable river temperatures (+2°C, -2°C) reveals that this river outlet cooling is due to the river temperature itself (not shown). In the open ocean SST changes are not sensitive to the river input temperature. River-induced SST changes are not significant under a two-sample t-test (at 95% significance) away from the river mouths. However, some modulations are of the same sign and similar magnitude in each simulated year. Therefore a one sample t-test on the difference between the two experiments has also been performed (von Storch and Zwiers, 1999). Under this test the WTNA is significantly cooled (99% significance level) in autumn north of 7°N (north of 13°N in the Caribbean) and in winter north of 10°N. Note that the two sample method tests whether the distributions of seasonal mean values in the River-and No-River experiments are statistically different, while the one sample method tests whether the difference between the two experiments is differs significantly from zero (von Storch and Zwiers, 1999). Therefore the two-sample test represents a stricter criterion. The cooling signal is widespread and can be up to -0.3°C locally. Surface warming of up to 0.5° occurs along the South American coastline, and is most prominent in autumn. In spring and summer the coastal warming spreads far across the Caribbean. Since the strong coastal warming cancels for the widespread surface cooling, the domain mean temperature does not change on average. In the Northern Plume the mean surface temperature decreases by -0.03°C, and increases in the Eastern Plume by +0.01°C. In the Caribbean winter cooling partly cancels for warming in spring and summer, leaving a net change of +0.02°C (Table 3.1).

3.3.2.2 Ocean stratification

To illustrate the impact of the freshwater plume on ocean stratification, the December to May and June to November domain mean temperature profiles and river-induced changes are shown in Figure 3.12. The average SST remains unchanged in both seasons. The freshwater plume shallows the mixed layer (ML) by about 3 m from December to May, and by about 5 m from June to November, thickening the barrier layer (BL) by an equal amount. The isolation of the BL from surface cooling and reduced entrainment leads to BL warming. The temperature increase is strongest

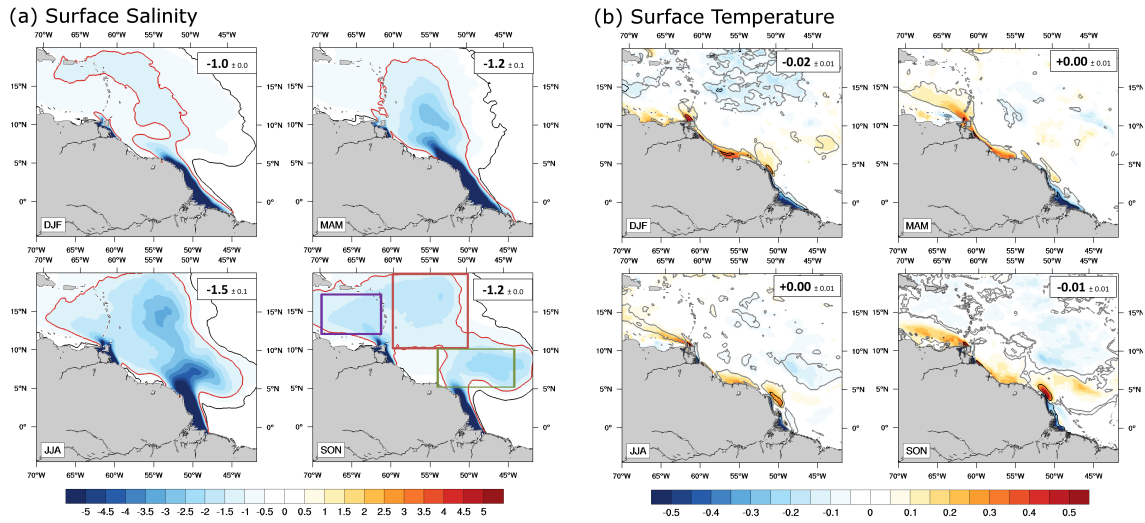


Figure 3.10: Maps of the WTNA mean seasonal cycle (DJF, MAM, JJA, SON) of the difference (River - No-River) in modelled ROMS (2001-2010) (a) surface salinity (PSU) and (b) surface temperature (°C). The red contour in (a) marks the -1 PSU salinity change. The red box in (a) SON is the Northern Plume area, the green box is the Eastern Plume area, and the purple box is the Caribbean plume area. The black contours show the 95% significance level obtained from a two-sided students t-test, and the grey contours are the 99% significance level from a one-sided students t-test (von Storch and Zwiers, 1999). The respective area mean values and their standard deviations are given in the top right corner of each map.

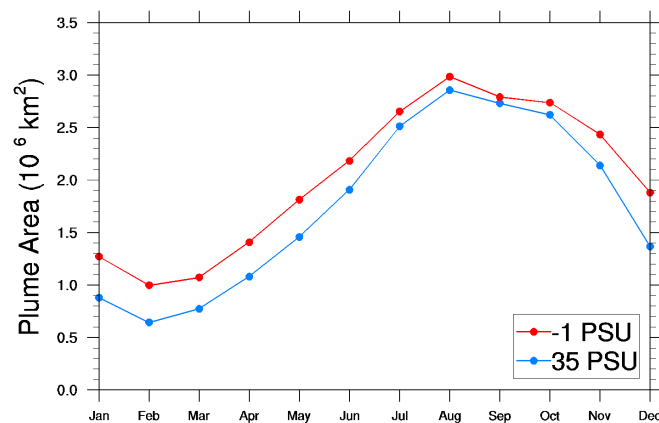


Figure 3.11: Annual mean cycle (2001-2010) of modelled plume size (10^6 km^2) in the River experiment derived with the 35 PSU salinity criterion (blue) and the -1 PSU (River - No-River) difference criterion (red).

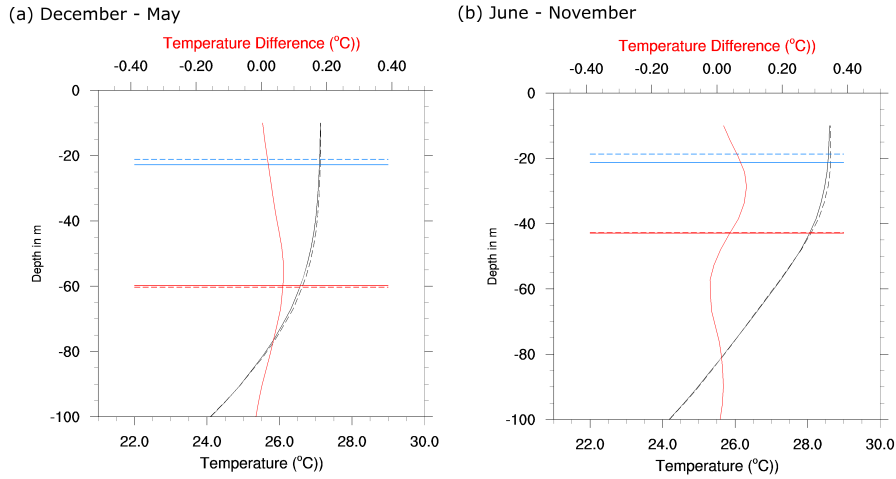


Figure 3.12: Domain mean modelled (2001-2010) vertical temperature profiles ($^{\circ}\text{C}$) in the WTNA from (a) December to May and (b) June to November in the No-River (solid black line) and River (dashed black line) experiments and the difference (River - No-River) between the experiments (red). The horizontal lines show the respective depths (m) of the top of the thermocline (red) and the ML depth (blue), defined in Section 1.2.1.

at about 30 m depth (about 0.1°C) in the hurricane season. Below the BL reduced mixing with the upper ocean cools the ocean by up to -0.05°C . From December to May BL heating is smaller in magnitude and occurs at greater depth, the average heating being up to 0.05°C at 55 m.

The uplift of the ML is widespread across the entire domain in each season (Figure 3.13a), and mirrors the low-salinity signal. The shallowing can be up to -8 m in spring where salinity is minimum in the River-experiment, and in autumn north of 15°N . The annual domain mean is a -1.8 m decrease. Annual mean shallowing is large in the Northern (-3.1 m) and Eastern Plume (-2.9 m) and slightly lower in the Caribbean (-2.2 m).

The BL thickens as a consequence of ML shallowing (Figure 3.13b). The area and annual mean BL increase is 2.2 m (7%). The magnitude in BL change is maximum in spring (up to 15 m, locally) and the spatial extent is largest in autumn. The BL increases by 4.7 m in the Northern Plume, by 3.1 m in the Eastern Plume and by 2.5 m in the Caribbean. The spatial distribution and seasonal cycle is similar to the pattern of surface salinity and ML changes. However, there are some discrepancies, i.e. in the Northern Plume. Note, that any modifications in BL thickness are a combination of the IL and ML depth. Although the IL depth does not change on average, it is subject to large natural - seasonal and regional - variations (not shown). Local changes can exceed ± 10 m, leading to additional thickening or shallowing of the BL locally. In

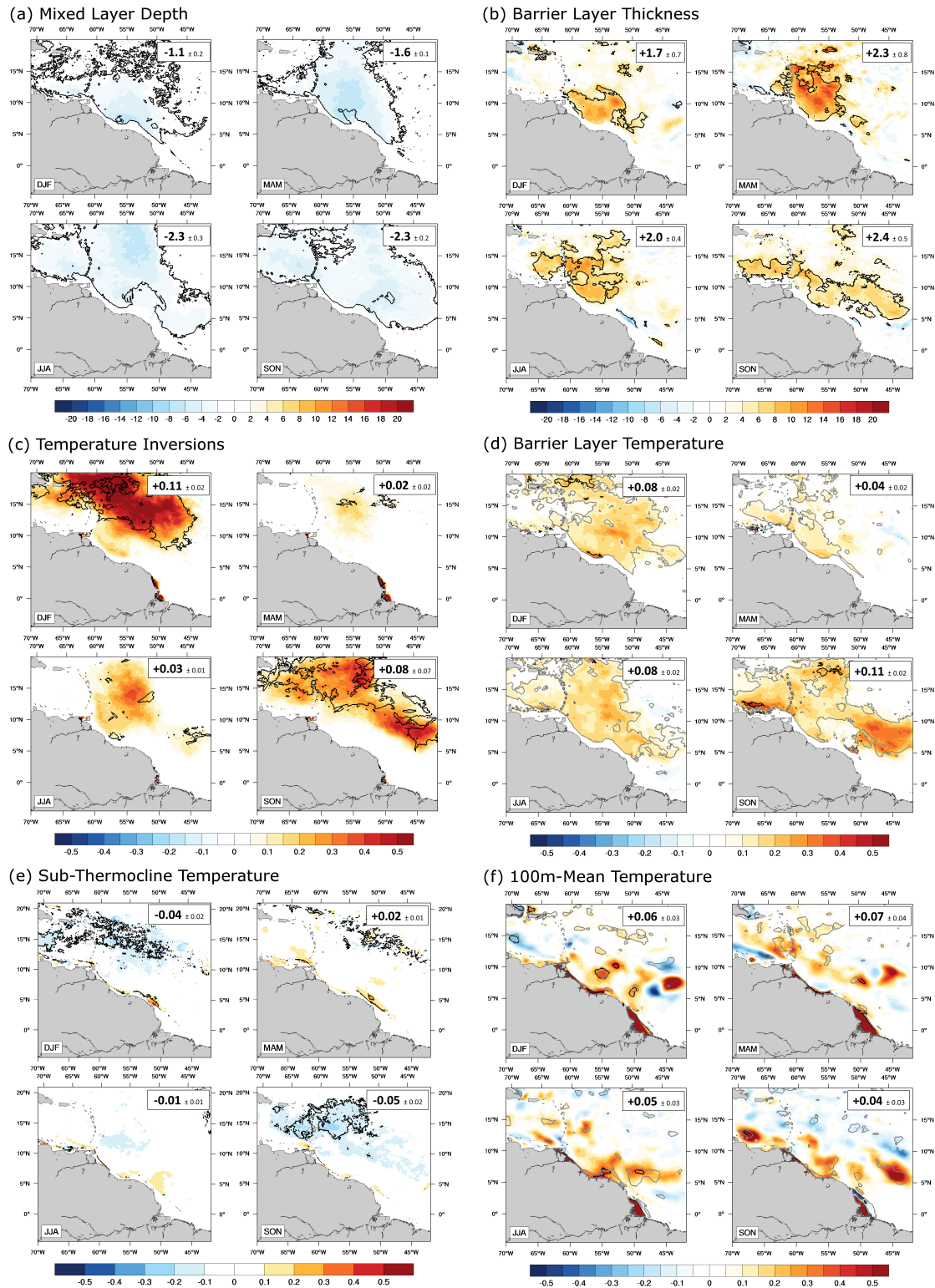


Figure 3.13: Maps of the WTNA mean seasonal cycle (DJF, MAM, JJA, SON) of differences (River - No-River) in modelled ROMS (2001-2010) (a) ML depth (m), (b) BL thickness (m), (c) mean TIs (°C), (d) mean temperature in the BL (°C), (e) mean temperature in the sub-thermocline (°C), and (f) T_{100} (°C). The black contours show the 95% significance level obtained from a two-sample students t-test, and the grey contours are the 99% significance level from a one-sample students t-test (von Storch and Zwiers, 1999). The respective area mean values and their standard deviations are given in the top right corner of each map.

the Northern Plume the IL deepens by 15 m, contributing about 30% to the change in BL thickness. The thermocline deepening in this area may be a consequence of BL heating and temperature inversions (see below). Note also, that the changes in ML depth have a small standard deviation, while the changes BL thickness are subject to larger variations (Table 3.1), due to the strong natural variability of the thermocline.

Persistent temperature inversions (TIs) can develop in the presence of BLs, due to isolation from surface cooling (Pailler et al., 1999; Vialard and Delecluse, 1998). They are here defined as the mean temperature difference compared to the reference temperature at 10 m if this difference exceeds 0.2°C (de Boyer Montégut et al., 2004). TIs occur in 31% of all modelled days and locations in the River experiment. In 8% of the plume they exceed 0.5°C . The seasonal and spatial distribution of mean TIs is shown in Figure 3.14a. They are strongest in winter north of 10°N (up to 1.0°C). These TIs diminish in spring when the ocean restratifies. They start to evolve again in summer in the Northern Plume (up to 0.35°C) and are established across the Northern and Eastern Plume (up to 0.4°C) and in the Caribbean (up to 0.3°C) in autumn. Note that domain average TIs can be lower than 0.2°C even though this is the TI threshold. This is because TIs are here assumed to be zero if they don't meet the criterion. For the heat contained in the ocean it is not only the magnitude of TIs that matters, but also their thickness. The average TI thickness is shown in Figure 3.14b. The thickness is largely a function of occurrence, stronger TIs having a greater vertical extent. TIs can stretch over up to 35 m in winter north of 10° and up to 15 m the rest of the year in the plume area. Interestingly TIs are not necessarily strongest, where the freshwater signal is maximum. However, the river plume seems to be a necessary mechanism for TI formation, since they almost entirely disappear when the Amazon and Orinoco are switched off (Figure 3.13c). They occur in only in 8% in the No-River experiment. The change in TIs mirrors almost exactly the occurrence of TIs in the River-experiment. Only a small area (north of 10°N and east of 50°W) still witnesses TIs when there are no rivers.

Despite strong TIs, the Amazon and Orinoco only induce annual mean BL heating of 0.08°C . The seasonal cycle of the temperature change is displayed in Figure 3.13d, and is maximum in autumn (0.11°C on average). Local BL heating is not significant under a two-sample t-test due to the large interannual variability of BL temperature. It is however significant at 99%, applying a one-sample test. BL heating in winter is large in the Northern Plume and north of the Brazil coast (up to 0.3°C). In spring the Caribbean experiences slight heating (up to 0.2°C), but the Northern and Eastern plume are not significantly warmer. Local summer heating can be up to 0.3°C and extends across the whole plume area, with a maximum in the Northern Plume.

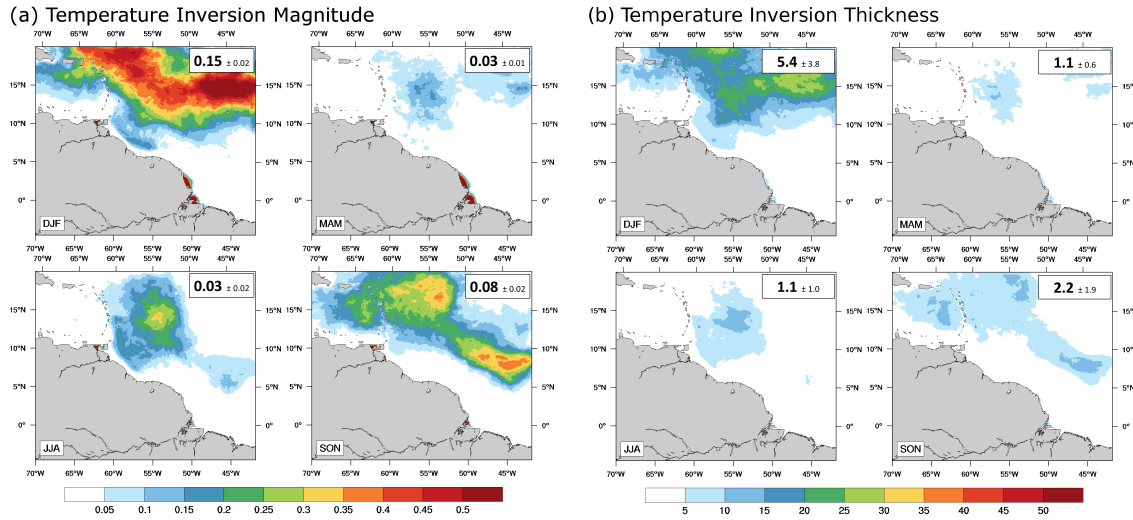


Figure 3.14: Maps of the WTNA mean seasonal cycle (DJF, MAM, JJA, SON) of modelled ROMS (2001-2010) (a) mean TI magnitude ($^{\circ}\text{C}$) and (b) TI thickness (m) in the River experiment. The respective area mean values and their standard deviations are given in the top right corner of each map.

Local heating is largest in autumn, with up to 0.4°C temperature increase in the Eastern Plume and the Caribbean. Annual mean BL heating is maximum in the Eastern Plume (0.13°C), and slightly lower in the Northern Plume (0.12°C) and the Caribbean (0.10°C).

The sub-thermocline (defined as the layer between the top of the thermocline and the 26°C isotherm) experiences widespread cooling as a consequence of reduced entrainment mixing. The annual mean sub-thermocline cooling is -0.02°C . The cooling is significant in autumn in the Northern Plume, and in winter in parts of the Caribbean and the Northern Plume (Figure 3.13e). Locally sub-thermocline cooling can reach up to -0.3°C in autumn and winter, but in spring and summer no significant cooling occurs. Annual mean sub-thermocline cooling is -0.07°C in the Northern Plume and -0.03°C in the Caribbean.

The modest BL heating leads to a slight warming of the isothermal layer ($+0.04^{\circ}\text{C}$) and an increase in 100m mean temperatures (T_{100}) ($+0.06^{\circ}\text{C}$). The spatial distribution of 100m mean temperature change hardly shows any significant features under a two-sided or one-sided t-test (Figure 3.13f). However, there are patches of significant T_{100} increase (under a one sample test) in autumn in the Eastern plume and the Southern Caribbean ($>0.5^{\circ}\text{C}$), where BL heating is strongest. Overall, despite the extensive BL heating, river-induced changes in 100 m mean temperature are patchy and no widespread warming occurs.

3.3.2.3 Tropical Cyclone metrics

River-induced modulations in the tropical cyclone heat potential (TCHP) are similar to T_{100} changes. Although the BL is slightly warmer in the River experiment, it is not able to produce any significant TCHP increase (Figure 3.15a). Apart from the BL heating not being very large, it is partly compensated for by surface and sub-thermocline cooling. The annual mean change is $+1.0 \text{ kJ/cm}^2$. The spatial distribution is very patchy and the change can even be negative. As for T_{100} , there are patches of significant TCHP increase (under a one sample test) of up to 10 kJ/cm^2 in the Eastern Plume and the southern Caribbean in autumn. BL heating in the Northern Plume is offset by cooling in the other layers. In summer TCHP increases north of the Amazon mouth, where BL heating is strong. In winter the freshwater plume lowers TCHP in the Caribbean, as a result of surface and sub-thermocline cooling combined with a lack of BL heating. The annual mean change is $+1.0 \text{ kJ/cm}^2$ in the Northern Plume, $+2.8 \text{ kJ/cm}^2$ in the Eastern Plume and $+1.1 \text{ kJ/cm}^2$ in the Caribbean. The cancelling effect of BL heating and subsurface cooling is obvious in the summer temperature profile of the Northern Plume (Figure 3.16b). The ocean is heated between about 10 m and 45 m (up to 0.18°C). Below the IL the ocean cools by up to -0.12°C , leading to a change in TCHP of only $+0.3 \text{ kJ/cm}^2$.

TCHP does not consider any stabilizing effects of the river-induced BL. However, the freshwater plume stabilizes the water column and therefore increases the energy required to cool the ocean surface. This is manifested in a widespread cooling inhibition index (CI) increase ($2.2 \text{ (J/m}^2)^{\frac{1}{3}}$, 6%). The seasonal cycle and spatial distribution of CI change is a reflection of surface salinity reduction (Figure 3.15b). The magnitude is largest in summer and spring (up to $+10 \text{ (J/m}^2)^{\frac{1}{3}}$), while the location of maximum CI growth migrates from about 10°N in spring to about 15°N in summer. River induced changes in CI are slightly lower in autumn and winter (up to $8 \text{ (J/m}^2)^{\frac{1}{3}}$), but still have a similar spatial extent. In summer and autumn the eastward transport of freshwater increases CI in the Eastern Plume (up to $6 \text{ (J/m}^2)^{\frac{1}{3}}$). Freshwater also enters the Caribbean, increasing CI by $2.8 \text{ (J/m}^2)^{\frac{1}{3}}$ on average. The average CI change is largest in the Northern Plume ($4.0 \text{ (J/m}^2)^{\frac{1}{3}}$), and slightly lower in the Eastern Plume ($2.6 \text{ (J/m}^2)^{\frac{1}{3}}$). The Amazon and Orinoco significantly enhance CI north of 10°N from June to November, when many hurricanes pass over that area (Ffield, 2007).

The modest change in TCHP suggests, that the majority of CI increase is due to increased stability, rather than subsurface warming. To estimate the temperature contribution CI is re-calculated, only considering changes in salinity stratification. Thereby

the temperature profile of the No-River experiment is used in both calculations, while the salinity profile of the respective experiment is applied. That yields an average CI increase of $+2.0 \text{ (J/m}^2\text{)}^{\frac{1}{3}}$, i.e. only $0.2 \text{ (J/m}^2\text{)}^{\frac{1}{3}}$ (9%) of the CI change is caused by subsurface heating. In the Northern and Eastern plume however, strong TIs can enhance CI considerably.

The annual cycle of CI increase and the respective contributions of temperature and salinity changes are shown in Figure 3.17 for the Northern and Eastern Plumes, and the Caribbean. The temperature contribution is consistently small compared to the salinity contribution in all three areas. In the Northern Plume, the freshwater-induced CI increase is largest in July (up to $6.3 \text{ (J/m}^2\text{)}^{\frac{1}{3}}$) and lowest in October ($4.2 \text{ (J/m}^2\text{)}^{\frac{1}{3}}$). In spring and summer, when TIs are only weak, the CI increase is almost solely determined by salinity stratification. From November to March strong TIs are able to reduce surface cooling by mixing and contribute about 20% to the change in CI. River-induced BL heating is strongest in the Eastern Plume. Here temperature contributes significantly to CI all year round. The total CI increase is largest in July and August (about $4 \text{ (J/m}^2\text{)}^{\frac{1}{3}}$) and lowest in February (about $1.5 \text{ (J/m}^2\text{)}^{\frac{1}{3}}$). The temperature contribution is large from June to December (about 20%) and lower between January and May (about 15%). In the Caribbean freshwater input decreases the heat content in winter. Therefore the temperature contribution is negative (-12%). Due to the strong salinity stratification however, the overall result is still a significant CI increase. In spring and summer the rivers increase temperatures in the Caribbean, amplifying the effect of surface freshening. The maximum CI increase in the Caribbean occurs in September ($4.5 \text{ (J/m}^2\text{)}^{\frac{1}{3}}$) and it is lowest in winter ($3 \text{ (J/m}^2\text{)}^{\frac{1}{3}}$) when ocean cooling damps the freshwater-induced CI increase.

BL formation as a consequence of surface freshening in the WTNA has two causes, river discharge and precipitation (Foltz et al., 2004). Furthermore the subduction of advected high salinity waters creates a subsurface salinity maximum, and hence increases the BL (Blanke et al., 2002; Balaguru et al., 2012a). Following Neetu et al. (2012), the impact of the total BL in the WTNA (river, precipitation and subduction induced) on CI can be obtained by comparing the standard CI in the Clear-River experiment, to a CI computed at constant salinity (33.85 PSU). According to this estimate, salinity stratification increases CI by about $8.1 \text{ (J/m}^2\text{)}^{\frac{1}{3}}$, of which only about 27% are caused by the Amazon and Orinoco.

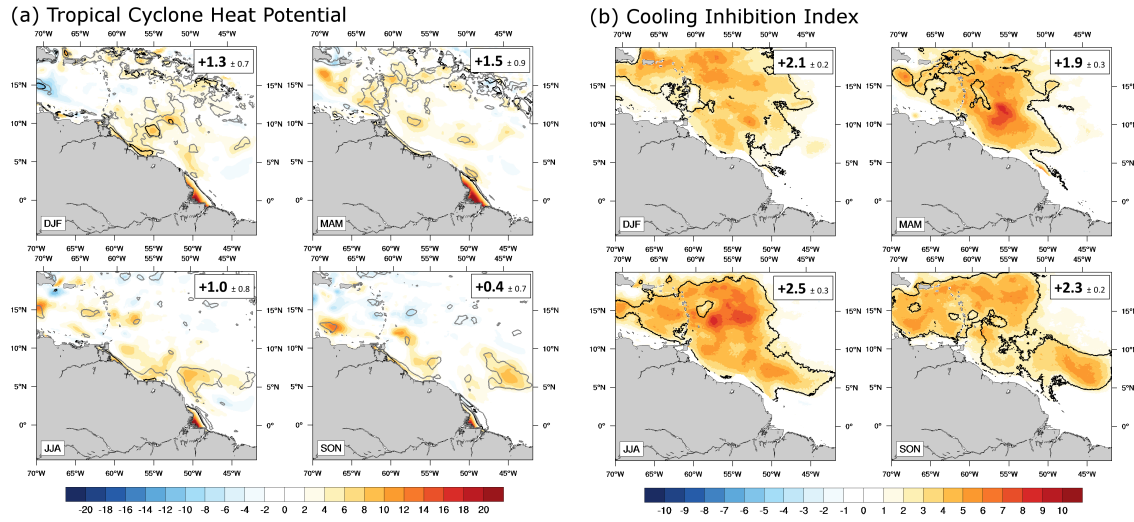


Figure 3.15: Maps of the WTNA mean seasonal cycle (DJF, MAM, JJA, SON) of differences (River - No-River) in modelled ROMS (2001-2010) (a) TCHP (kJ/cm^2) and (b) CI ($(\text{J}/\text{m}^2)^{1/3}$). The black contours show the 95% significance level obtained from a two-sample students t-test, and the grey contours in (a) are the 99% significance level from a one-sample students t-test (von Storch and Zwiers, 1999). The respective area mean values and their standard deviations are given in the top right corner of each map.

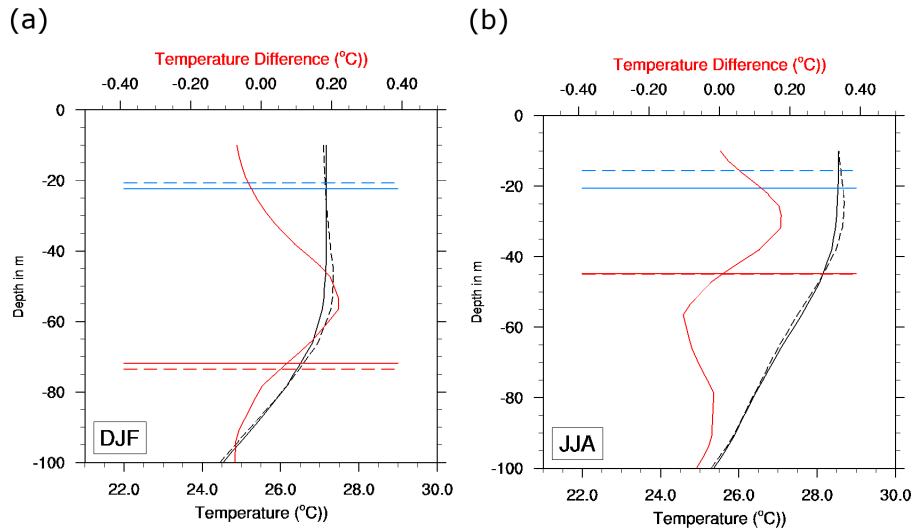


Figure 3.16: Modelled mean vertical temperature profiles ($^{\circ}\text{C}$) from the 2001-2010 ROMS simulation in the Northern Plume from (a) December to May and (b) June to November in the No-River (solid black line) and River (dashed black line) experiments and the difference (River - No-River) between the experiments (red). The horizontal lines show the respective depths (m) of the top of the thermocline (red) and the ML (blue). The area of the Northern Plume is defined in Figure 3.10.

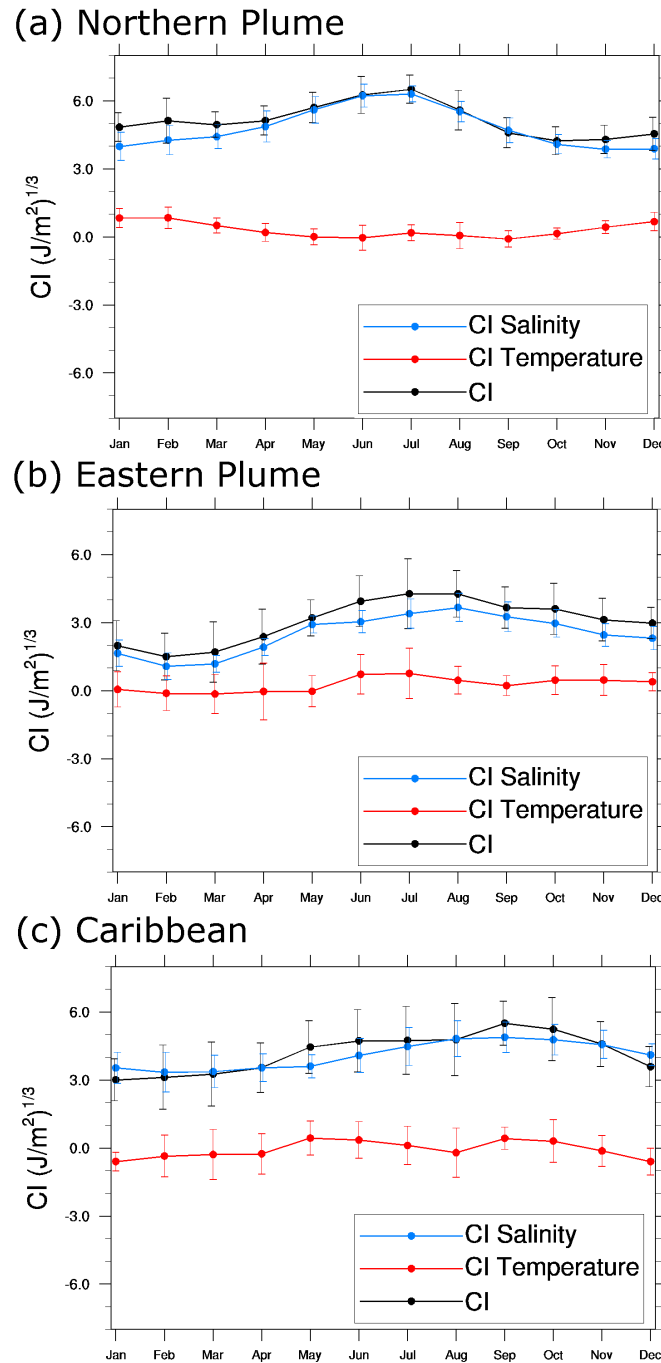


Figure 3.17: Mean annual cycle of modelled (2001-2010) differences (River - No-River) in CI in the (a) Northern Plume area, (b) Eastern Plume area, and (c) Caribbean for CI, when considering only salinity changes (blue), only temperature changes (red), and changes in both variables (black). The error bars show the standard deviation of the 10 year model average. The Northern Plume, Eastern Plume, and Caribbean are defined in Figure 3.10.

River and No-River experiments

	Whole Domain		Whole Do- main	Northern Plume	Eastern Plume	Caribbean
	No-River	River				
	River - No-River					
Salinity (PSU)						
Surface	36.1 ± 0.2	34.9 ± 0.2	−1.3 ± 0.0	−1.4 ± 0.1	−1.7 ± 0.1	−1.1 ± 0.1
Temperature (°C)						
Surface	28.0 ± 0.2	28.0 ± 0.3	−0.00 ± 0.01	−0.03 ± 0.01	+0.01 ± 0.03	+0.02 ± 0.02
Barrier Layer	27.9 ± 0.3	27.9 ± 0.3	+0.08 ± 0.02	+0.12 ± 0.02	+0.13 ± 0.05	+0.10 ± 0.02
Sub-Thermocline	27.0 ± 0.1	27.0 ± 0.1	−0.02 ± 0.01	−0.07 ± 0.02	−0.01 ± 0.02	−0.03 ± 0.02
100m-mean	26.8 ± 0.3	26.9 ± 0.3	+0.06 ± 0.03	+0.02 ± 0.03	+0.12 ± 0.14	+0.03 ± 0.07
Layer Thickness (m)						
Mixed Layer	19.5 ± 0.1	17.7 ± 0.2	−1.8 ± 0.2	−3.2 ± 0.3	−2.9 ± 0.4	−2.2 ± 0.2
Barrier Layer	31.6 ± 1.6	33.7 ± 1.5	+2.2 ± 0.5	+4.7 ± 0.8	+3.1 ± 1.2	+2.5 ± 1.0
Isothermal Layer	50.0 ± 1.6	50.1 ± 1.7	+0.2 ± 0.4	+1.5 ± 0.7	+0.0 ± 1.0	+0.3 ± 1.2
TC Metrics						
TCHP (kJ/cm²)	47.1 ± 6.5	48.2 ± 6.7	+1.0 ± 0.6	+1.0 ± 1.0	+2.8 ± 2.0	+1.1 ± 1.7
CI (J/m²) ^{1⁄3}	35.4 ± 1.0	37.6 ± 1.0	+2.2 ± 0.2	+4.0 ± 0.1	+2.6 ± 0.2	+2.8 ± 0.2

Table 3.1: Modelled (2001-2010) area and annual mean salinity, temperatures, layer thickness and TC metrics with their standard deviation in the No-River and River experiments and differences between the experiments.

3.4 Discussion

3.4.1 Model validation

There is no drift in surface temperatures over the period of the simulation (Figure 3.5), therefore the 10-year sample size does not seem to be a major limitation of this study. However, SSTs in the model are slightly too high compared to observations. This may be related to an underestimate of cloudiness (and consequently overestimate of solar radiation) in the CFSR data (Wang et al., 2010). However, the seasonal cycle and extent of the warm pool is represented well in the model. Since the relevant parameter in this study is the difference between two ROMS simulations the small bias is not expected to affect the overall conclusions.

The underestimation of surface salinity in the model may be related to an excess of surface freshening triggered by the CFSR data. Comparing the CFSR precipitation to TRMM data (Huffman et al., 2007) shows that there is about 26% too much rain in winter and about 40% in summer (Figure 3.18). This may also explain the overestimation of plume area in the model compared to Aquarius. Note however, that the Aquarius data are only for one year, but the interannual variability of the plume size can be very large (Grotsky et al., 2014). Since the shape of the plume and the seasonal variation are similar to observed characteristics, the model should be capable to simulate river-induced changes in the ocean.

The isothermal layer is too deep in the model. As will be shown in Chapter 4, this is partly related to deep solar penetration in the River-experiment leading to a warmer subsurface, and therefore deeper IL. A realistic representation of solar penetration in the plume (i.e. considering light absorbing particles) is necessary to correctly simulate temperature stratification. The overestimation of BL thickness may magnify modelled TIs. However, timing and magnitude of inversions are in good agreement with de Boyer Montégut et al. (2007) and Mignot et al. (2007) who observed TIs of up to 1.0°C in the plume area, mainly occurring in winter north of 10°N.

TCHP in the model is about 20 kJ/cm² too high on average. This is a consequence of the model warm bias in the isothermal layer. Note that TCHP only considers integrated temperatures in excess of 26°C, which artificially inflates the warm bias. The seasonal and spatial variation, however, is similar in model and observations. Daily TCHP maps are available online (<http://www.aoml.noaa.gov/phod/cyclone/data/at.html>), where TCHP is derived from satellite obtained sea surface height anomalies (Goni et al., 2009b). They show that TCHP in the Caribbean can commonly reach 100 kJ/cm² in the hurricane season and is large within the warm pool, similar to

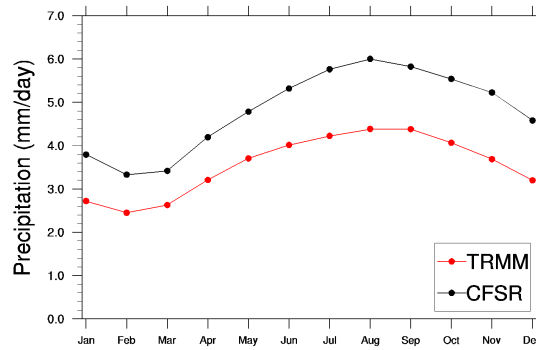


Figure 3.18: Mean Annual cycle of the 2001-2010 precipitation (mm/day) in the model domain in the CFSR forcing data and in TRMM observations.

the model results. The agreement of CI with observations is much better, compared to observed and modelled TCHP. Note that CI does not consider absolute temperatures, but the relative temperature decrease from the ocean surface. The model warm bias is therefore not visible in CI.

3.4.2 The freshwater effect

The atmospheric temperature and moisture are the same in both experiments and the ocean is not coupled to the atmosphere. This may lead to atmospheric damping and thus an underestimate of river induced SST changes. However, Wallcraft et al. (2008) conclude that atmospheric damping in bulk flux models may only have small effect. Furthermore the surface fluxes are almost identical in the River and No-River experiment, suggesting that atmospheric damping is not a major issue in this experiment.

The season of maximum plume extent coincides with the hurricane season, possibly maximizing the river impact on TC development. The maximum plume area found here ($2.8 \times 10^6 \text{ km}^2$) is considerably larger compared to Hu et al. (2004) ($1.2 \times 10^6 \text{ km}^2$). Note that they only consider freshwater signals south of 15°N and do not include the Caribbean. Commonly applied absolute salinity thresholds (e.g. Hu et al. (2004)) suggest that the Amazon and Orinoco plume vanishes in winter. The results presented here show that the rivers still have a significant influence on the salinity budget in this season.

The Amazon and Orinoco warm the coastal ocean surface. Westerly winds, blowing along the southern Caribbean and the South American coast, induce strong upwelling (Müller-Karger et al., 1989). With the rivers heating the subsurface, these upwelling waters are now warmer. Further offshore, the proposed warming of SSTs (Pailler et al., 1999; Field, 2007) does not occur, although the low-salinity stabilises the BL and therefore reduces entrainment of cold water into the ML. This is in agree-

ment with Masson and Delecluse (2001) and Balaguru et al. (2012b) who also found that the BL thickening in the Amazon plume is not able to produce a significant change in SST. Vialard and Delecluse (1998) show that the BL in the western Pacific fresh pool can lead to surface cooling. Howden and Murtugudde (2001) saw a slight surface cooling (-0.1°C) in the Bay of Bengal due to river input and White and Toumi (2014) find that the Congo plume has no effect on surface temperatures. Vinayachandran et al. (2012) and Seo et al. (2009) on the contrary report a slight warming ($+0.1^{\circ}\text{C}$) of the Bay of Bengal associated with river input and the South China Seas slightly warm due to the Changjiang river (Park et al., 2011). However, none of the modelled SST modulations seem sufficient to assign observed temperature anomalies of up to 4°C (Ffield, 2007) in the Amazon plume to river-induced BL formation. Note however, that this only applies to the Amazon and Orinoco induced part of the BL. The total effect of the BL in the WTNA (created by rivers, rain and subduction) on surface temperatures may well be different. Further analysis (Chapter 4) will show that SSTAs may partly be explained through light absorbing particles induced by the river plume.

The ML temperature modifications can be illustrated using simple considerations. As Vialard and Delecluse (1998) pointed out, the warming effect of decoupling the ML from the deep ocean is competing with cooling due to solar penetration. As the ML shallows, more radiative flux leaves at its base. Furthermore latent cooling is distributed over a shallower ML, potentially leading to surface cooling. The temperature tendency of the ML can be calculated from outgoing and incoming fluxes (e.g. Vialard and Delecluse (1998)):

$$\delta T_t = \frac{Q_n + (1 - f(\text{ML}))Q_s}{\rho_o C_p \text{ML}}, \quad (3.1)$$

where ρ_o is the density (1025 kg/m^3) and C_p the heat capacity ($3985 \text{ J/kg}^{\circ}\text{C}$) of sea water, and $f(\text{ML})$ is the fraction of solar radiation leaving at the ML base. The ML is 19.5 m in the No-River and 17.7 m in the River case. This corresponds to 82.1% (199 W/m^2) and 80.4% (195 W/m^2) solar light being absorbed in the ML for Jerlov (1976) waters of type I (equation 1.2). The annual mean solar flux (Q_s) is constant in both experiments (242 W/m^2). The mean latent, longwave, and sensible heat fluxes (Q_n) are directed upwards and are almost the same in both experiments (-0.2 W/m^2 smaller in the River experiment). The resulting temperature tendency is $-2.52^{\circ}\text{C/year}$ more cooling in the River experiment, which offsets the expected heating due to reduced entrainment fluxes. The BL on the other hand is isolated from surface cooling and at the same time receives more solar radiation (it is now closer to the

surface and thicker). Since the IL does not change, the same amount of radiation is leaving at the BL base. That means that 4 W/m^2 more radiation is absorbed in the BL, leading to a change in temperature tendency of about $+1.0^\circ\text{C}/\text{year}$.

The magnitude of BL thickening and ML shallowing due to the Amazon and Orinoco is in good agreement with Balaguru et al. (2012b). Masson and Delecluse (2001) also found local heating of the BL due to the Amazon and here it is shown that it is widespread. As the BL is isolated from surface cooling strong TIs develop. Since TIs almost entirely disappear in the No-River experiment, they can clearly be assigned to the freshwater discharge. TIs are strongest in winter north of 10°N , despite the freshwater plume being maximum in summer and autumn further south. Plume waters are being transported northward in summer and early autumn (Johns et al., 2002; Coles et al., 2013). The warm ML water in the Northern Plume can then be trapped beneath the advected freshwater (Mignot et al., 2007). To illustrate the nature of the strong TIs north of 10°N , the winter profile for the Northern Plume is shown in Figure 3.16a. While the rivers cool the surface (-0.08°C), there is deep BL heating of up to 0.22°C at about 55 m depth.

The dramatic increase in TIs in winter does not lead to large BL heating (0.08°C). Note that TIs are vertically localized features, that don't necessarily imply a heating of the whole BL. In winter they can be up to 30 m thick. Compared to the mean BL thickness of about 50 m, they are unlikely to contribute a huge amount to BL temperature. The widespread (but small) sub-thermocline cooling due to BLs, found here, has previously not been reported.

Surface freshening and BL heating does not result in a significant TCHP increase, partly due to cancellation by surface and sub-thermocline cooling. As pointed out by Price (2009) and Vincent et al. (2012b), TCHP may not be a good proxy for SST feedbacks in the Amazon and Orinoco plume region, since it neglects salinity stratification. Furthermore TCHP does not consider at which depth in the ocean a temperature change may occur (Price, 2009). CI takes this into account by weighting the density profile by depth (equation 1.5). Therefore the weakness of TCHP is two-fold: on the one hand it neglects stratification changes, and it also fails to account for the ability of temperature changes to affect TCs.

The Amazon and Orinoco plume significantly increases CI ($2.2 \text{ (J/m}^2)^{\frac{1}{3}}$). Note that this only applies to the part of the BL caused by the Amazon and Orinoco. The impact of the total BL (river, rain, and subduction induced) on CI will therefore be even larger. Idealized calculations show that only about a quarter of barrier layer induced CI increase is caused by the Amazon and Orinoco plume. Neetu et al. (2012) observe an increase of $+8.4/+3.6 \text{ (J/m}^2)^{\frac{1}{3}}$ in CI for a surface freshening of -2.2 (-

1.5) PSU in the north-eastern Bay of Bengal/East Indian Coast. Similarly, a modelled salinity reduction of -1.8 PSU corresponds to a $+4.3 \text{ (J/m}^2\text{)}^{\frac{1}{3}}$ higher CI in the June-November plume area of this study (marked by -35 PSU). The CI increase is mainly driven by increased salinity stratification (91%), rather than ocean temperatures (9%). BL heating is therefore unlikely to have a large impact on SST feedbacks. Similarly Neetu et al. (2012) find that salinity stratification contributes up to 70% to seasonal increases in CI due to monsoonal surface freshening and river discharge in the Bay of Bengal.

The increase in CI implies a reduction in SST feedback on TCs, potentially leading to a considerable increase in TC activity. Natural CI variability induces 7% changes in power dissipated by TCs in the North Atlantic (Vincent et al., 2014), and Balaguru et al. (2012a) find that BLs enhance TC intensification rates by 50%. The freshwater input of the of the Amazon and Orinoco is therefore likely to increase cyclone intensity in the area.

3.5 Summary and conclusions

The light freshwater plume of the Amazon and Orinoco stretches across the western tropical North Atlantic (WTNA) throughout the whole year. The plume extent is maximum in summer and autumn: hence the season of maximum river impact and hurricane activity occur at the same time. Although the plume does not produce any surface warming, it modifies subsurface temperatures and stratification substantially.

Plume size peaks in August ($2.8 \times 10^6 \text{ km}^2$) and lags maximum discharge by about 3 month in agreement with observations (Molleri et al., 2010). The freshwater shallows the density mixed layer (ML) and thickens the barrier layer (BL) by up to 15 m locally. The BL is isolated from surface cooling and receives more solar radiation. It is therefore warmed by up to 0.3°C locally. BL heating is widespread and occurs throughout the whole year. The warming is maximum in autumn in the Eastern Plume and the Caribbean. Strong temperature inversions (TIs) develop within the freshwater plume, particularly north of 10°N in winter. They are a consequence of warm waters being trapped below the advected freshwater plume (Mignot et al., 2012). The thicker BL inhibits entrainment mixing, leading to widespread cooling below the IL (of up to -0.3°C locally).

In contrast to several observations-based studies (Ffield, 2007; Pailler et al., 1999; Foltz and McPhaden, 2009) the Amazon and Orinoco plume does not create widespread warm anomalies in the plume area. However, SSTs are warmer near the coasts (by up to 0.5°C), where heated upwelling waters reach the surface. The rivers

even tend to cool the ocean surface in the northern part of the model domain. Note that the temperature tendency of the ML is a combination of opposing factors: Firstly the freshwater plume stabilizes the water column and therefore reduces entrainment cooling. Secondly shallowing of the ML implies that more solar radiation will penetrate below the ML, leading to surface cooling (Vialard and Delecluse, 1998). Furthermore latent cooling affects a shallower layer, potentially leading to more surface cooling. In this study the competing effects of the river-induced BL almost cancel each other on average. This is in agreement with previous explicit river studies (Masson and Delecluse, 2001; Balaguru et al., 2012a).

The modification in ocean temperatures and stratification potentially has an important feedback on TC intensity. Despite the warmer BL, the widely used tropical cyclone heat potential (TCHP) increases only slightly ($+1.0 \text{ kJ/cm}^2$ (2%)) and the 100 m mean temperature is only $+0.06^\circ\text{C}$ warmer. However, higher stability may limit SST-feedbacks and potentially lead to more intense TCs (Balaguru et al., 2012a). Since TCHP neglects any changes in density stratification it is not the appropriate metric in salinity stratified areas (Price, 2009). The cooling inhibition index (CI) introduced by Vincent et al. (2012b) is a measure for the resistance of the ocean to surface cooling and explicitly takes density stratification into account. Indeed the freshwater plume increases CI by about $2.2 \text{ (J/m}^2)^{\frac{1}{3}}$, suggesting that cooling feedbacks are reduced. Locally CI can increase by up to $10 \text{ (J/m}^2)^{\frac{1}{3}}$ in the hurricane season. The impact of the total BL (river, rain and subduction induced) on CI will be even larger. Salinity stratification increases CI by about $8.1 \text{ (J/m}^2)^{\frac{1}{3}}$, of which only 27% are caused by the freshwater plume. The fact that only 9% of the CI increase is due to temperature changes further illustrates the shortcoming of TCHP in the Amazon and Orinoco plume area. This study agrees that there may be a connection between the more intense TCs passing over the plume, as suggested by Ffield (2007). The river-freshwater contribution however, will be dominated by the increase in stratification, rather than temperatures.

Chapter 4

The ocean colour effect of the Amazon and Orinoco plume

4.1 Introduction

Solar transmission is a key factor in determining ocean temperatures and can therefore have an important impact on tropical cyclone (TC) development (Gnanadesikan et al., 2010). Subramaniam et al. (2007) measured how deep sunlight can travel through the ocean interior in the Amazon plume. They found that the euphotic zone depth (1% light level) is significantly reduced in the low-salinity waters of the Amazon plume. Figure 4.1a shows the measured light levels in low-salinity (<30 PSU), mesohaline (30 PSU - 35 PSU), and oceanic (>35 PSU) waters. The differences are striking: while solar radiation is transmitted down to about 80 m in oceanic waters, it can only reach 25 m to 33 m in plume areas. Light attenuation within the plume is due to the high abundance of light absorbing particles. These can be organic compounds or sediments being transported from the land into the ocean by rivers. Furthermore the Amazon and Orinoco are sources of nutrients (mainly nitrogen), that trigger primary productivity in the open ocean, thereby enhancing chlorophyll concentration.

The importance of ocean colour for temperatures has previously been recognized. Denman (1973) followed by Simson and Dickey (1981) were the first to mention the importance of adequate representation of light penetration in ocean models. Jerlov (1976) took optical properties into account by assigning water types as a proxy for chlorophyll concentration. However, dividing water into categories is only a qualitative estimate and introduces sharp steps between different types (Lee et al., 2005). Morel and Antoine (1994) and Ohlmann (2003) have overcome this by parametrizing attenuation coefficients as a function of surface chlorophyll concentration, available from satellite observations. Similarly, Murtugudde et al. (2002) used chlorophyll

derived attenuation depths to estimate the effect of solar transmission on ocean temperatures. This is appropriate in regions where chlorophyll is dominant, but is likely to underestimate light attenuation when other particles are present as well (Lee et al., 2005). Del Vecchio and Subramaniam (2004) investigate the optical properties in the Amazon plume and conclude that coloured dissolved organic matter (CDOM) is the main contributor to light absorption in the vicinity of the river mouth (70%). Within the higher salinity area of the plume detritus and chlorophyll become more important (40% each). Odriozola et al. (2007) found that CDOM overwhelms phytoplankton absorption in the Orinoco plume by up to a factor of 139, especially in the Gulf of Paria.

In this chapter the modulating effect of ocean colour in the Amazon and Orinoco plume on oceanic temperatures and TC metrics is investigated. To avoid the limitations of a chlorophyll-only based approach, a recently developed attenuation algorithm is applied, that derives the vertical absorption profile from satellite-obtained surface optical properties (Lee et al., 2005). This method allows to consider all light absorbing particles. Measurements show a strong anti-correlation between surface salinity and CDOM/chlorophyll concentration (Hu et al., 2004; Del Vecchio and Subramaniam, 2004; Odriozola, 2004). Therefore, light attenuation is estimated using surface salinity as a proxy for Jerlov water types, in an alternative approach. The quality of the second method is assessed by contrasting the impacts on ocean stratification and temperatures. The Coloured-River experiment is compared to a case without freshwater input and with a clear river, allowing to separate freshwater and colour-induced changes (see Table 2.2 for a summary of the model experiments). The combined impact of the river-induced BL and ocean colour on oceanic temperatures and TC controls is studied for the first time.

4.2 Light attenuation

ROMS applies the Paulson and Simpson (1977) radiation formulation, which splits solar transmission into a sum of exponential terms representing different wavelengths (eq. 1.2). Transmission coefficients and exponents are given as a function of the five Jerlov (1976) water types (I,IA,IB,II,III, numbered 1 to 5 in the following). The euphotic zone depth decreases with the water type (Figure 4.1b). In the original ROMS setup, only one fixed water type can be set for the entire region and model run. In the No-River and in the Clear-River experiment a constant Jerlov water type of 1 is applied, assuming that there are no light absorbing particles contained in the ocean.

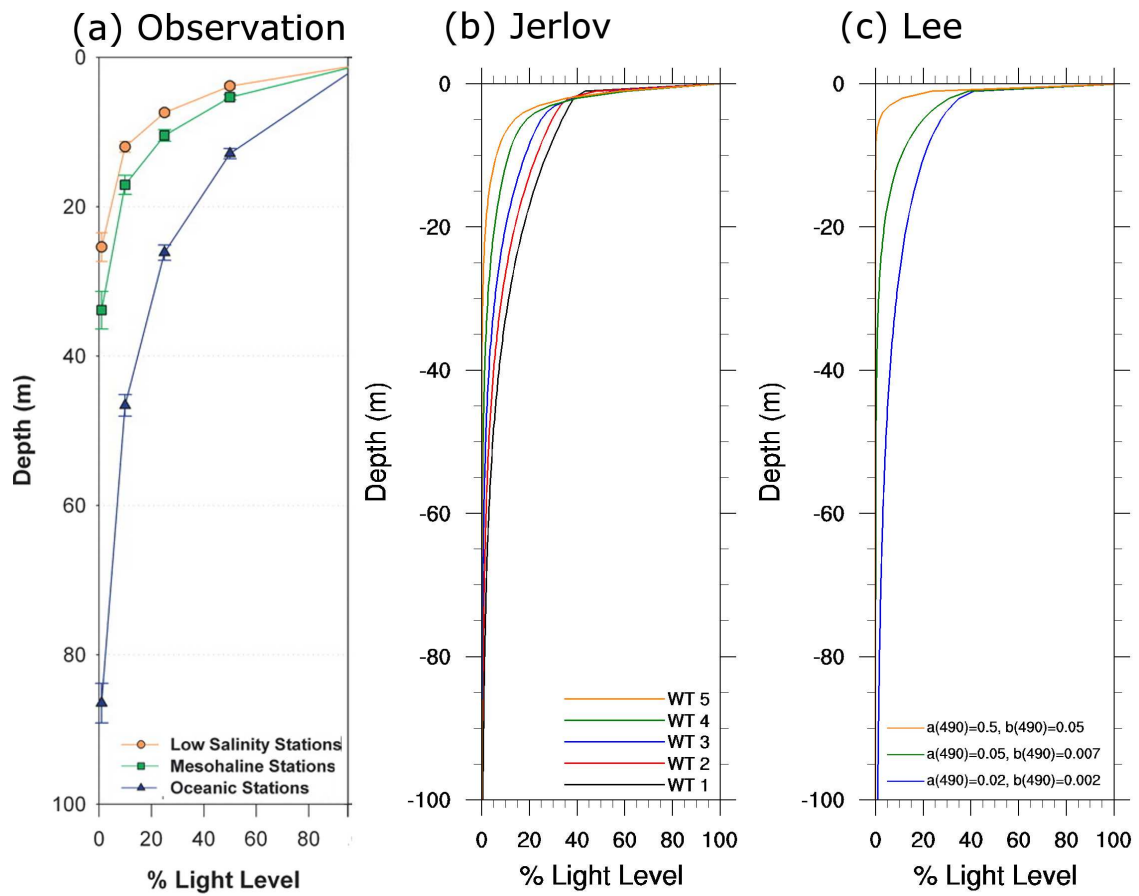


Figure 4.1: Vertical profiles of (a) observed light levels binned by station type in the WTNA from Subramaniam et al. (2007), (b) light Level as defined by Paulson and Simpson (1977) for the five different Jerlov (1976) water types, and light level derived with the Lee et al. (2005) algorithm for different values of $b_b(490)$ and $a(490)$.

Station type	Criterion	Water type
Oceanic	≥ 35 PSU	1
Mesohaline	30 - 35 PSU	4
Low Salinity	≤ 30 PSU	5

Table 4.1: Surface salinity and corresponding water type in the Water-Type experiment

4.2.1 Salinity approximation for water types

As a first order approximation, it has been attempted to estimate solar transmission from surface salinity in this study. Hu et al. (2004) found a strong correlation of CDOM and chlorophyll concentration with low-salinity waters in the Amazon and Orinoco plume, and used ocean colour as a proxy for the plume. From several ship cruises, Odriozola et al. (2007) showed a similar correlation between surface salinity and CDOM abundance in the Gulf of Paria and the Southern East Caribbean. Molleri et al. (2010) found a correlation of about 0.7 between surface salinity and absorption of dissolved and detrital material at 443 nm. Furthermore Subramaniam et al. (2007) showed that light attenuation is reduced in low-salinity regimes of the Amazon plume. Therefore it may be possible to use salinity as a proxy for light penetration in the Amazon and Orinoco plume.

The ROMS solar transmission algorithm has been slightly modified, to allow for temporal and regional variation of water types. The Paulson and Simpson (1977) radiation formulation, with five different water types determining the absorption profile, is still applied. However, surface salinity is now used as a proxy for the water types, using observational light levels inside and outside the plume. The experiment is therefore referred to as the 'Water-Type' experiment (Table 2.2). Thereby the measurements of Subramaniam et al. (2007) are used as a guidance for water type - salinity relations. Their data shows, that the euphotic zone depth for oceanic areas roughly corresponds to a Jerlov water type of 1 (Figure 4.1). In mesohaline areas light is transmitted down to only 33 m, and in low salinity areas it can only reach about 24 m. For comparison, the euphotic zone depth is at about 44 m for Jerlov 4 waters and at 24 m for Jerlov 5 waters. Oceanic areas have therefore been assigned Jerlov type 1, mesohaline areas are of type 4 and low salinity areas are water type 5 in the model (Table 4.1).

4.2.2 The Lee et al. (2005) algorithm

In a more advanced approach, solar transmission has been estimated from satellite-obtained ocean colour. To account for light absorption of all optically active particles in the Amazon and Orinoco plume, the Lee et al. (2005) algorithm has been implemented into ROMS (Coloured-River experiment). It has for example been used by Lee et al. (2007) and Shang et al. (2011) to study the euphotic zone depth and by Xiu and Chai (2014) in a study on the light attenuation in the Pacific. The method is similar to Paulson and Simpson (1977), splitting transmission into visible and ultraviolet contributions, but the attenuation coefficients depend on the depth (z) as well as the inherent optical properties. Light transmission is obtained from extensive *Hydrolight* simulations, a software package that solves the radiative transfer equation to obtain the radiance distribution, from absorption and scattering coefficients at different wavelength (Mobley, 1995). From these simulations Lee et al. (2005) deduce the vertical transmittance T :

$$T(IOP, z) = F_{VIS} e^{-K_{VIS}(IOP, z)z} + (1 - F_{VIS}) e^{-K_{IR}(z)z} \quad (4.1)$$

with F_{VIS} the ratio of visible radiation (350 nm - 700 nm) as compared to the total downwelling radiation at the ocean surface. It is about 0.424 when the solar radiation is split at 700 nm (Mobley, 1994). K_{IR} and K_{VIS} are the attenuation coefficients for visible and infrared radiation:

$$K_{IR} = \left[0.560 + \frac{2.304}{(0.001 + z)^{0.65}} \right], \quad (4.2)$$

$$K_{VIS} = K_1(IOP) + \frac{K_2(IOP)}{\sqrt{1 + z}} \quad (4.3)$$

K_1 and K_2 are model parameters that depend on the the optical properties at 490 nm, but are independent of depth. They were obtained empirically by Lee et al. (2005):

$$K_1(IOP) = \chi_0 + \chi_1 \sqrt{a(490)} + \chi_2 b_b(490) \quad (4.4)$$

$$K_2(IOP) = \zeta_0 + \zeta_1 a(490) + \zeta_2 b_b(490), \quad (4.5)$$

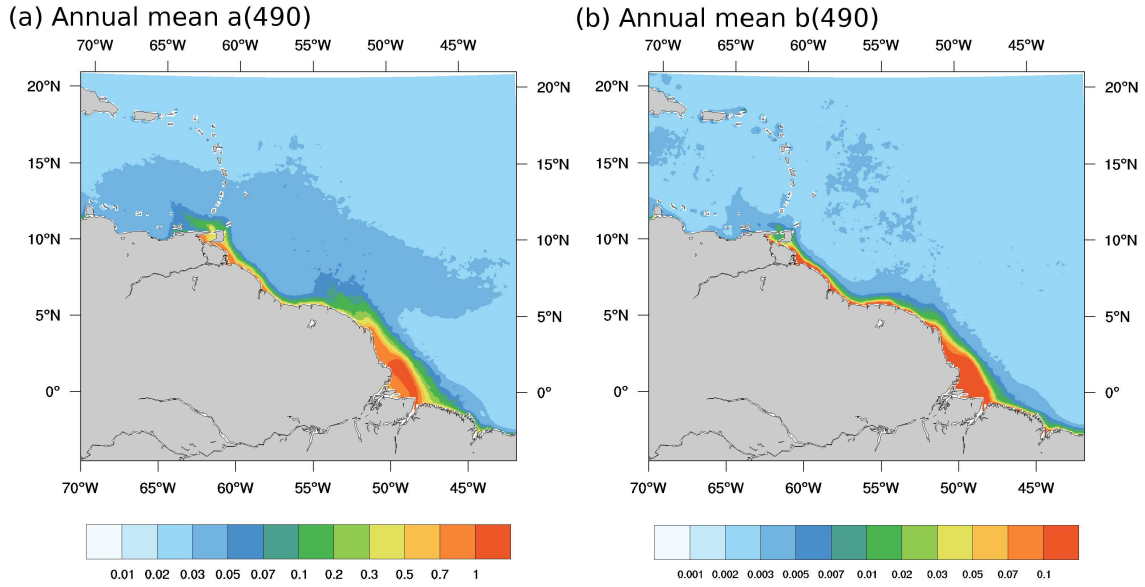


Figure 4.2: Maps of annual mean (a) absorption coefficient at 490 nm (sr^{-1}) and (b) backscattering coefficient at 490 nm (sr^{-1}) in the WTNA, derived from MODIS remote sensing reflectance using the quasi analytical algorithm of Lee et al. (2002) for the years 2001-2010.

with $\chi_0 = -0.057 \text{ m}^{-1}$, $\chi_1 = 0.482 \text{ m}^{-1}$, $\chi_2 = 4.221 \text{ m}^{-1}$ and $\zeta_0 = -0.183 \text{ m}^{-1}$, $\zeta_1 = 0.702 \text{ m}^{-1}$, $\zeta_2 = -2.567 \text{ m}^{-1}$. Note that variations of the solar zenith angle are neglected, assuming an average zenith angle of 45° . The absorption (a) and backscattering (b_b) coefficients at 490 nm can be obtained from satellite data, using a quasi analytical algorithm (QAA) developed by Lee et al. (2002) (see below).

The light profiles for different absorption and backscattering coefficients at 490 nm, applying the Lee et al. (2005) algorithm, are shown in Figure 4.1c. They represent typical low values, medium values, and high values of $a(490)$ and $b_b(490)$. Within the range of found $a(490)$ and $b_b(490)$, light absorption varies strongly with $a(490)$, but only weakly with $b_b(490)$. The low coefficients typically appear in the northern part of the model domain (north of 15°N) and the high values in the vicinity of the river mouth (Figure 4.2). The medium values are the domain averages.

4.2.2.1 The quasi analytical algorithm

The absorption ($a(490)$) and backscattering $b_b(490)$ coefficients are calculated applying the updated version (Lee et al., 2009) of the quasi analytical algorithm (QAA), developed by Lee et al. (2002). This only requires remote sensing reflectance (R_{rs}) at four different wavelengths (443 nm, 488 nm, 555 nm and 667 nm), which were obtained from MODIS Aqua and MODIS Terra. R_{rs} at a specific wavelength is converted

to the respective subsurface reflectance using:

$$r_{rs} = \frac{R_{rs}}{0.52 + 1.7R_{rs}}. \quad (4.6)$$

A threshold is applied to $R_{rs}(667 \text{ nm})$, since it could be erroneous due to imperfect atmospheric correction. If $R_{rs}(667 \text{ nm})$ is out of bounds it is approximated using Lee et al. (2009). Following Gordon et al. (1988) it is further assumed that the relationship between r_{rs} and backscattering and absorption coefficients can be described as:

$$r_{rs}(\lambda) = g_0 \frac{b_b(\lambda)}{a(\lambda) + b_b(\lambda)} + g_1 \left(\frac{b_b(\lambda)}{a(\lambda) + b_b(\lambda)} \right)^2 \quad (4.7)$$

where $g_0=0.089$ and $g_1=0.125$. The absorption coefficient at the reference wavelength $a(\lambda_0)$ is then given by:

$$a(\lambda_0) = a_w(\lambda_0) + 10^{p_0 + p_1\chi + p_3\chi^2} \quad (4.8)$$

with $p_0=-1.146$, $p_1=-1.366$, $p_3=-0.469$ and $a_w(\lambda_0)$ is the known absorption coefficient of pure water. The recommended reference wavelength for MODIS data is 555 nm, where $a_w(555 \text{ nm})=0.0596 \text{ m}^{-1}$ (Pope and Frey, 1997), and

$$\chi = \log \left[\frac{r_{rs}(443) + r_{rs}(488)}{r_{rs}(\lambda_0) + 5 \frac{r_{rs}(667)^2}{r_{rs}(488)}} \right] \quad (4.9)$$

The particle backscatter at the reference wavelength ($b_{bp}(\lambda_0)$) can be obtained by solving eq. 4.7, $b_b(\lambda)$ at other wavelength is derived via extrapolation using a power law:

$$b_b(\lambda) = b_{bw} \lambda + b_{bp}(\lambda_0) \left(\frac{\lambda_0}{\lambda} \right)^\eta \quad (4.10)$$

where

$$\eta = 2.0 \left[1 - 1.2 \exp \left(-0.9 \frac{r_{rs}(443)}{r_{rs}(\lambda_0)} \right) \right] \quad (4.11)$$

$b_{bw}(\lambda_0)$ is the backscattering coefficient of pure seawater at the reference wavelength. It has first been estimated by Morel and Antoine (1974) and has recently been updated by Zhang et al. (2009). The backscattering coefficient of pure seawater at 555nm is 0.019 m^{-1} . Substituting $b_b(\lambda)$ into eq. 4.7 gives the corresponding values of $a(\lambda)$, and using $\lambda=490 \text{ nm}$ gives the $a(490)$ and $b_b(490)$ (Figure 4.2), required to

calculate solar penetration using Lee et al. (2005).

4.3 Results

4.3.1 Satellite-derived or salinity-approximated ocean colour?

In this section solar transmission and resulting ocean stratification and temperature in the Water-Type and Coloured-River experiment are compared. The aim is to assess how good an indicator surface salinity is for light attenuation in the WTNA. The mean seasonal cycles (DJF, MAM, JJA, SON) of the model years 2001-2010 are discussed and differences in the relevant variables for the whole model domain, the Northern and Eastern Plume, and the Caribbean are summarized in Table 4.2. The Northern, Eastern, and Caribbean Plume are defined in Figure 3.10.

4.3.1.1 Euphotic zone depth and solar transmission profiles

The seasonal cycle of the euphotic zone depth in the Water-Type experiment is shown in Figure 4.3a. Note that in the No-River and Clear-River experiments the euphotic zone depth is at a constant depth of 86 m for the whole domain (Jerlov water type of 1). The average light depth in the Water-Type experiment is up to 24 m shallow near the Amazon and Orinoco river mouths (corresponding to a water type 5). In winter sunlight can reach up to 86 m in the Caribbean, but in the Northern Plume the light depth can be reduced to up to 60 m. In spring the freshwater plume spreads into the WTNA, leading to light depth below 50 m in the Northern Plume. In summer the euphotic zone depth falls below 50 m in most of the Northern Plume, the southeastern Caribbean, and the western part of the Eastern Plume. In autumn, the westward transport of freshwater into the Caribbean further increases light attenuation, and the eastward freshwater advection with the NBCC leads to shallow light levels in the Eastern Plume. In the Northern Plume the minimum light depth can still be below 50 m, but is less widespread than in summer.

Figure 4.3b shows the mean seasonal cycle of the euphotic zone depth derived from the remotely sensed ocean colour, applying Lee et al. (2005). With the Lee et al. (2005) algorithm, the outline of lower attenuation depth is similar to the shape of freshwater plume (Figure 3.2). The light depth is between 30 m to 60 m in the offshore river plume area. Outside the plume the light can reach depths of 80 m. The euphotic zone depth near the river mouths is very shallow (only about 1-20 m). This is in good agreement with Odriozola (2004), who has shown that light of certain wavelengths may not even travel below 1 m in the Orinoco plume.

Compared to Lee et al. (2005), the contour of the 50 m euphotic zone depth in the salinity approximation is fairly similar (Figure 4.3b). However, the difference between the euphotic zone depth (Figure 4.3c) reveals substantial dissimilarities. The Lee et al. (2005) light depth is shallower (< -30 m) in the Caribbean in autumn, spring and winter, and in the eastern model domain, where the NBC enters the WTNA. Satellite derived ocean colour also leads to stronger absorption near the Amazon and Orinoco mouth and in coastal upwelling regimes. Note that the salinity-approximation can not account for other nutrient sources like eddy, or equatorial upwelling. On the other, hand salinity-approximated absorption is stronger in summer in the Northern Plume area, and in autumn in the northern Caribbean, and Northern and Eastern Plumes (up to $+30$ m). Hence the salinity-approximation tends to underestimate light absorption in coastal areas and the low-plume season, but leads to stronger absorption in the high plume season.

For oceanic heating it is not necessarily the euphotic zone depth that matters, but rather the shape of the absorption profile. Figure 4.4a shows the domain mean seasonal cycle of the attenuation profiles in the Coloured-River and Water-Type experiments. In all seasons satellite ocean colour absorbs stronger, than the salinity-approximation. The differences in solar transmission are largest between the surface and about 40 m depth. In winter and spring, the light level at 20 m is about $+7\%$ larger in the Water-Type experiment, and the euphotic zone depth is about -15 m deeper. The agreement between the two methods is better in summer and autumn, when the river plume is larger. The difference at 20 m is then only about $+4\%$.

To illustrate regional characteristics, light transmission for the Northern Plume and the Caribbean are shown in Figures 4.4 b,c. In the Northern Plume, where hardly any freshwater appears in winter and spring, light is less strongly absorbed near the surface in the Water-Type experiment ($+6\%$ and $+4\%$ light level at 20 m, respectively). In summer and spring, the transmission profiles are fairly similar. Nearer the surface, attenuation is stronger in the Water-Type experiment, but the light levels are equal at 20 m. The euphotic zone depth is even shallower under the salinity approximation (by $+5$ m and by $+8$ m, respectively). The strongest deviation of Coloured-River and Water-Type absorption profiles occurs in the Caribbean, where upwelling is partly responsible for the high abundance of nutrients on the ocean surface. In winter and spring the difference in light levels is about 9% at 20 m and the euphotic zone depth is about $+20$ m shallower. The light level in summer is about 4% higher at 20 m in the Water-Type experiment. In autumn, when the freshwater signal in the Caribbean is maximum, there is good agreement between the satellite-based ocean colour method and the salinity approximation.

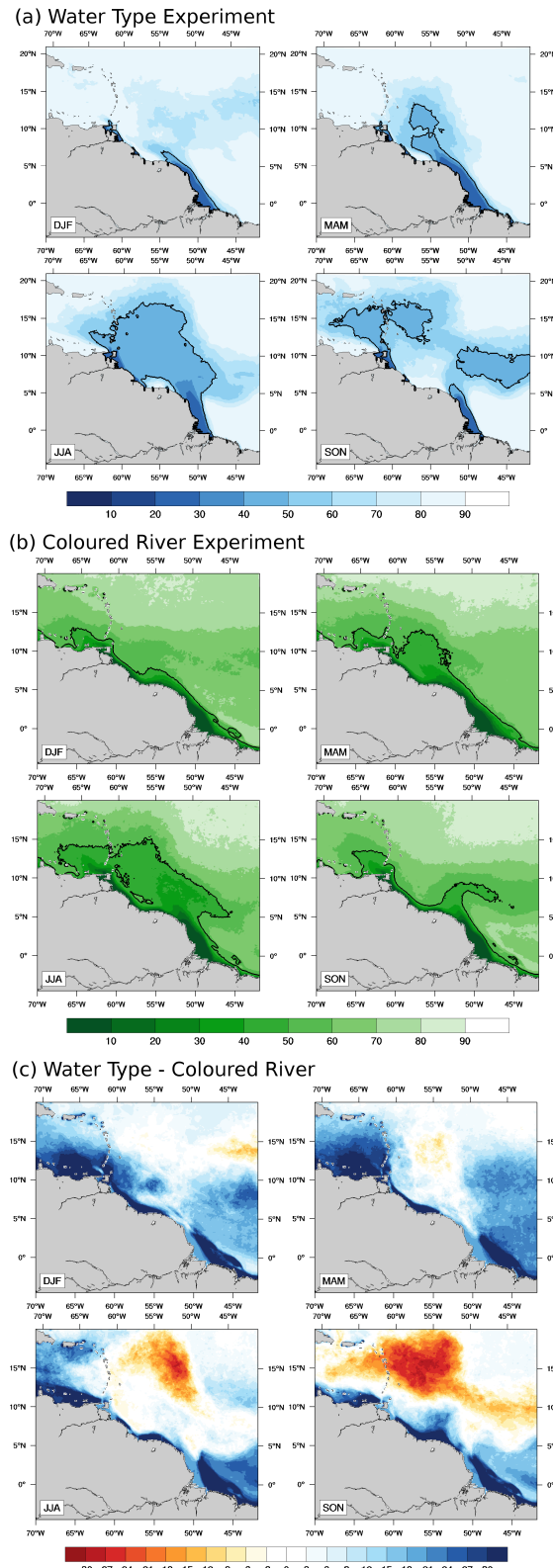


Figure 4.3: Maps of the mean seasonal cycle (DJF,MAM,JJA,SON) of the euphotic zone depth (m) in (a) the Water Type experiment and (b) the Coloured-River experiment for the years 2001-2010. The black line marks 50 m contour. (c) Difference (Water Type - Coloured-River) in euphotic zone depth.

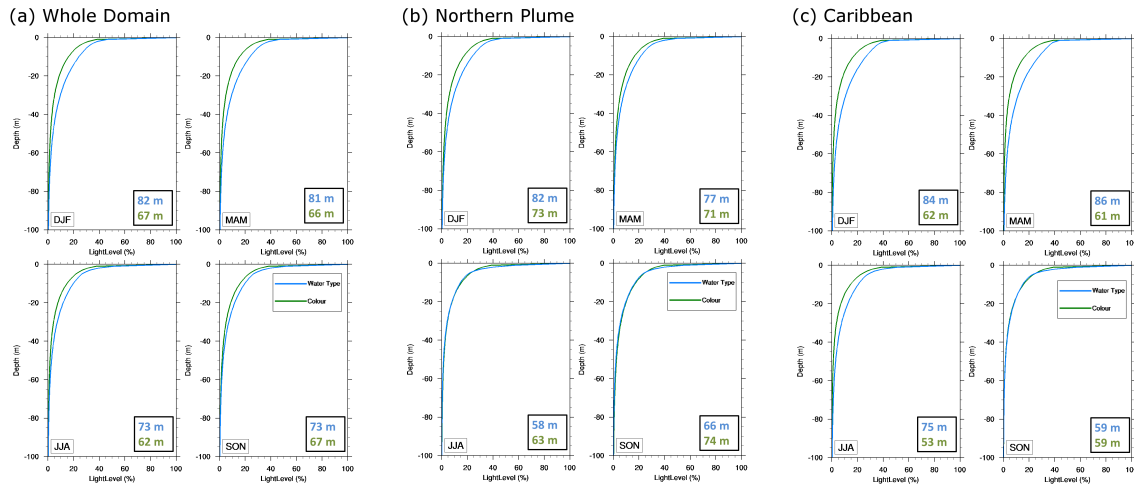


Figure 4.4: Seasonal cycles (DJF, MAM, JJA, SON) of the mean vertical light attenuation profiles (light level in %) in the ocean as a function of depth (m) in the Coloured-River (green) and Water-Type (blue) experiments for the (a) whole model domain, (b) the Northern Plume, and (c) the Caribbean for the years 2001-2010. The respective euphotic zone depth (m) is given in the bottom right corners for each season. The Northern Plume area and Caribbean are defined in Figure 3.10.

4.3.1.2 Ocean temperatures and stratification

The attenuation profiles suggest that the Water-Type approach leads to a colder ocean surface and warmer subsurface than the Lee et al. (2005) method, since less light is trapped in the upper layers. Seasonal maps of surface temperature differences between the Coloured-River and Water-Type approach are shown in Figure 4.5a. Note that the changes are only significant under a one-sample t-test (99%). The sea surface temperature (SST) is colder under the Water-Type method in coastal upwelling regions, for example in spring and autumn in the southern Caribbean (up to -0.3°C). SSTs are also lower near the river mouths, (by up to -0.5°C near the Orinoco mouth and by up to -0.4°C near the Amazon mouth). Further offshore the two algorithms do not result in significantly different SSTs in winter. In spring the salinity-approximation produces colder SSTs in the Caribbean (up to -0.3°C). In summer SSTs north of the Brazil coast and in the Caribbean are colder, when using the low-salinity plume as an indicator for light attenuation. In autumn SSTs in the Caribbean are similar with both methods. In the area between the Brazil coast and about 10°N the salinity-approximation results in a slightly colder sea surface (up to -0.15°C). On average the Water-Type method produces lower SSTs (-0.02°C). They are -0.03°C colder in the Caribbean and -0.04°C in the Eastern Plume. There is no annual mean difference in the Northern Plume.

The slight underestimation of light attenuation in the Water-Type experiment lets

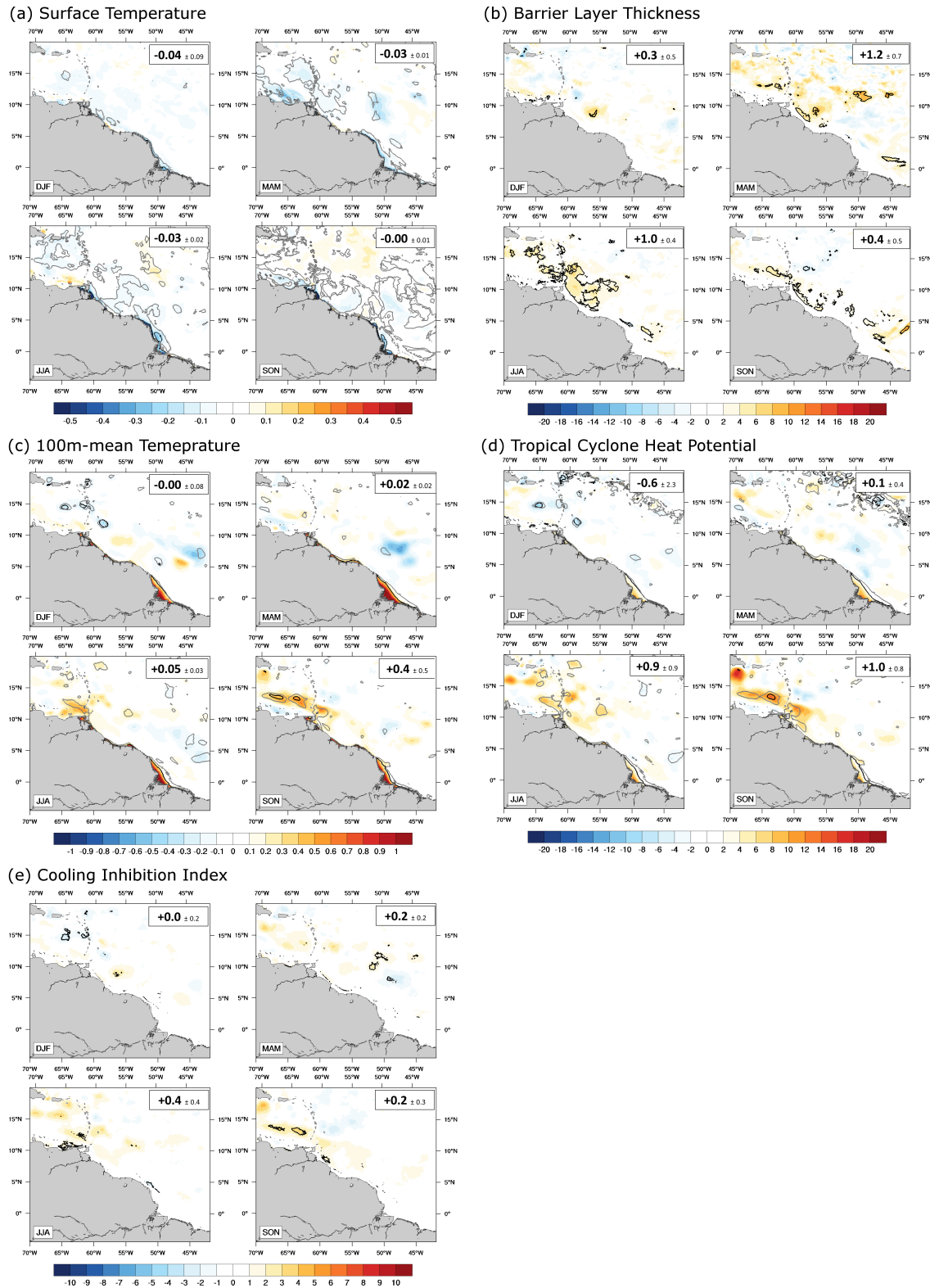


Figure 4.5: Maps of the seasonal cycle (DJF, MAM, JJA, SON) of differences resulting from salinity-approximated and satellite-derived ocean colour (Water Type - Coloured-River) in modelled ROMS (2001-2010) (a) SST ($^{\circ}\text{C}$), (b) BL thickness (m), (c) T_{100} ($^{\circ}\text{C}$), (d) TCHP (kJ/cm^2), and (e) CI ($(\text{J}/\text{m}^2)^{\frac{1}{3}}$). The black contours show the 95% significance level obtained from a two-sided students t-test, and the grey contours are the 99% significance level from a one-sided students t-test (von Storch and Zwiers, 1999). The respective area mean values and their standard deviations are given in the top right corner of each map.

more radiation penetrate into the barrier layer (BL). Therefore the subsurface is slightly warmer compared to the Lee et al. (2005) algorithm, resulting in a deeper isothermal layer (IL) and thicker BL (Figure 4.5b). Domain mean BL difference is maximum in spring (+1.2 m). Although the modelled differences can be large in magnitude (up to +/-10 m locally), the spatial distribution is patchy, and there are hardly any significant features. In summer however, the BL can be up to 8 m thicker in the Water-Type experiment. The differences are largest east of the Orinoco mouth and in the Caribbean. The modulations in BL thickness are reduced in autumn, and only show a few significant patches in coastal areas. The annual mean differences between the two approaches are maximum in the Caribbean (+1.6 m), and slightly smaller in the Northern and Eastern Plumes (+0.4 m and +0.5 m, respectively). These are mainly driven by a deeper IL, rather than differences in ML depth.

The 100m mean temperature (T_{100}) is significantly warmer near the river mouths ($>1^\circ\text{C}$), when applying the salinity approximation (Figure 4.5c). Further offshore, changes in T_{100} are patchy and largely insignificant (even under a one-sample t-test). However, there are small features of significantly warmer T_{100} (up to 0.6°C) in the southern Caribbean in summer and autumn. Results are very similar for tropical cyclone heat potential (TCHP) and cooling inhibition index (CI) (Figures 4.5d,e). While the two approaches lead to hardly any significant differences in winter and spring, there are patches of significantly larger TCHP in summer and autumn in the southern Caribbean (up to $+8 \text{ kJ/cm}^2$ and up to $+14 \text{ kJ/cm}^2$), and east of the Orinoco mouth in autumn (up to 12 kJ/cm^2). Similarly there are small features of significantly larger CI in the southern Caribbean in autumn (up to $4 \text{ (J/m}^2)^{\frac{1}{3}}$). However, as will be seen below, the dissimilarities between the two approaches are mostly marginal compared to the net impact of a coloured river plume on ocean temperatures and TC metrics. Only in areas where the net coloured river impact is small and differences in the approach hence relatively large (e.g in the hurricane season in the Caribbean), may the choice of method make a considerable difference. In those cases the salinity-approximation may lead to a slight underestimate of the ocean colour impact on CI and SST feedbacks.

The annual mean differences of relevant oceanic variables between the Water-Type and Coloured-River method are summarized in Table 4.2. They are generally smaller than the standard deviation. Exceptions are the $+0.04^\circ\text{C}$ warmer 100 m mean temperature in the Eastern plume and $+0.07^\circ\text{C}$ in the Caribbean, as well as the deeper IL (+1.8 m) and larger CI ($+0.8 \text{ (J/m}^2)^{\frac{1}{3}}$) in the Caribbean. In summary the differences in ocean stratification from the satellite and salinity-derived light attenuation are small, even though the euphotic zone depths and transmission profiles may be

Salinity as a proxy for ocean colour

	Whole main	Do- main	Northern Plume	Eastern Plume	Caribbean
Salinity (PSU)					
Surface	-0.0 ± 0.1		$+0.0 \pm 0.1$	-0.0 ± 0.1	-0.0 ± 0.0
Temperature ($^{\circ}\text{C}$)					
Surface	-0.02 ± 0.03		-0.00 ± 0.02	-0.04 ± 0.02	-0.03 ± 0.04
Barrier Layer	-0.02 ± 0.02		-0.01 ± 0.02	-0.04 ± 0.02	-0.03 ± 0.03
100m-mean	$+0.03 \pm 0.03$		$+0.01 \pm 0.04$	-0.04 ± 0.01	$+0.07 \pm 0.06$
Layer Thickness (m)					
Mixed Layer	$+0.0 \pm 0.1$		$+0.1 \pm 0.2$	-0.1 ± 0.1	$+0.1 \pm 0.2$
Barrier Layer	$+0.8 \pm 0.4$		$+0.4 \pm 0.7$	$+0.5 \pm 1.1$	$+1.6 \pm 0.9$
Isothermal Layer	$+0.8 \pm 0.4$		$+0.5 \pm 0.8$	$+0.4 \pm 1.0$	$+1.8 \pm 1.0$
TC Metrics					
TCHP (kJ/cm^2)	$+0.4 \pm 0.9$		$+0.4 \pm 0.8$	-0.3 ± 1.9	$+1.9 \pm 2.2$
CI (J/m^2) $^{\frac{1}{3}}$	$+0.2 \pm 0.2$		$+0.0 \pm 0.4$	-0.0 ± 0.5	$+0.8 \pm 0.5$

Table 4.2: Differences (Water Type - Coloured-River) between satellite and salinity derived ocean colour in annual mean salinity, temperatures, layer thicknesses and TC metrics with their standard deviation in the whole model domain, the Northern Plume, the Eastern Plume and the Caribbean.

different.

4.3.2 The ocean colour effect

In the Coloured-River and Water-Type experiments realistic light absorption in the coloured Amazon and Orinoco plume has been taken into account. The above section has shown that the two experiments produce hardly any differences in the ocean interior, and only the ocean colour effect obtained from the Lee et al. (2005) algorithm is discussed in the following. Thereby the Coloured-River experiment is compared to a freshwater plume with constant light absorption (Clear-River experiment). Note that these differences do not represent the net effect (freshwater and ocean colour) of the Amazon and Orinoco plume, but rather the additional impact light attenuation has on ocean stratification and temperatures. The analysis is performed for the seasonal means, namely winter (DJF), spring (MAM), summer (JJA), and autumn (SON) for the model years 2001-2010, and the results of relevant parameters for the whole model domain, the Northern and Eastern Plume, and the Caribbean are summarized in Table 4.3.

4.3.2.1 Ocean stratification and temperatures

Applying realistic light attenuation in the WTNA changes the vertical temperature profile substantially (Figure 4.6). From June to November the surface is about 0.1°C

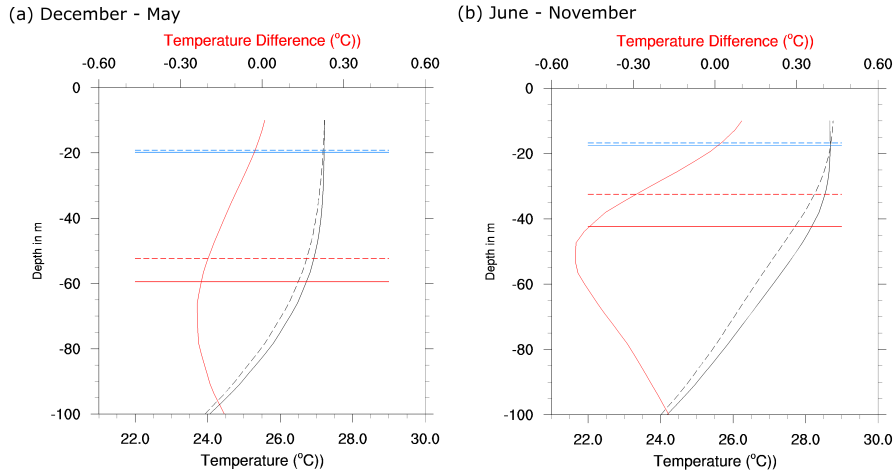


Figure 4.6: Domain mean modelled (2001-2010) vertical temperature profiles ($^{\circ}\text{C}$) as a function of ocean depth (m) in the WTNA from (a) December to May and (b) June to November in the Clear-River (solid black line) and Coloured-River (dashed black line) experiments and the difference (Coloured-River - Clear-River) between the experiments (red). The horizontal lines show the respective depths of the top of the thermocline (red) and the ML (blue).

warmer. However, isolating the deeper ocean from solar radiation induces subsurface cooling which significantly exceeds surface heating. The maximum mean temperature decrease (-0.6°C) occurs below the isothermal layer at about 50 m. Note that subsurface cooling is about 5 times larger than freshwater induced BL heating (Figure 3.12). Increased temperature stratification leads to an uplift of the top of the thermocline by about 10 m. Since the mixed layer (ML) depth hardly changes (-1 m) this implies shallowing of the barrier layer (BL) by an equal amount. The temperature changes are less distinct from December to May. The surface temperature hardly changes, while the subsurface cools by up to 0.25°C at 70 m depth, inducing shallowing of the isothermal layer by about 8 m.

More radiation is absorbed near the surface and SSTs increase by $+0.09^{\circ}\text{C}$ on average in the Coloured-River experiment. The seasonal cycle of colour induced SST changes is shown in Figure 4.7a. The coastal areas behave differently, compared to the open ocean. SSTs strongly increase close to the Amazon and Orinoco river mouths (up to $+1^{\circ}\text{C}$), where the concentration of light absorbing particles is largest. In the upwelling regions along the South American coastline on the other hand, the surface is colder. That is particularly the case in autumn and winter (up to -0.5°C). Further offshore light attenuation increases SSTs from spring to autumn. Although the warming signal seems widespread and is consistent throughout the 10 simulated years (not shown), it is only significant in a rather small patch of the Eastern plume

in summer (under a two-sample t-test), due to large natural variability. However, the temperature changes are significant under a one-sample t-test (99%). In spring the surface warms by up to $+0.3^{\circ}\text{C}$ in the Northern plume, but only marginally in other areas. In summer the SST warming is maximum and spans almost across the entire model domain. The increase is largest in the Eastern Plume (up to $+0.4^{\circ}\text{C}$) and slightly smaller in the Northern Plume (up to $+0.35^{\circ}\text{C}$) and the Caribbean (up to $+0.25^{\circ}\text{C}$). The warming magnitude drops in autumn, but is still of similar spatial extent and distribution with a maximum in the Eastern Plume (up to $+0.35^{\circ}\text{C}$). In winter, when the IL deepens, the effect of ocean colour is to slightly decrease surface temperatures (up to -0.35°C). The cooling is most prominent in the Caribbean and the northern part of the Northern Plume. Annual mean SSTs are $+0.12^{\circ}\text{C}$ warmer in the Northern and Eastern Plume. The strong winter cooling in the Caribbean reduces the annual mean SST increase to only $+0.05^{\circ}\text{C}$.

As already suggested by the temperature profiles (Figure 4.6), blocking the deeper ocean from sunlight cools the subsurface substantially. Figure 4.7b shows the temperature change at 50 m depth (note the difference in colour scales between Figures 4.7a and 4.7b). There occurs strong cooling in summer in the Caribbean and the Northern Plume (up to -1.2°C). The ocean colour impact on 50 m temperatures is maximum in autumn (up to -1.6°C), affecting the entire model domain. The 50 m cooling is weaker in winter (up to -1.0°C near the South American coast) and minimum in spring. Note that the 50 m temperature does not change substantially in winter north of 10°N . This does not necessarily mean that the subsurface is not cooled, but is rather an artifact of the arbitrary 50 m choice. In the Coloured-River experiment, the IL is up to 90 m deep at that time of the year. Since the strongest cooling would per definition occur below the isothermal layer, temperature changes may not be captured in the 50 m slice.

The average ML depth changes very slightly in the Coloured-River experiment (-0.4 m), due to an increased temperature gradient. The IL on the other hand, shallows significantly (Figure 4.7c). The annual and domain mean is about -8 m on average (Table 4.3). Since this uplift is driven by subsurface cooling, it is a good indicator for timing and location of the strongest temperature reductions. The change is widespread and is maximum in spring and summer in the Northern Plume (up to -25 m) and the Caribbean (up to -23 m). In autumn the magnitude of IL shallowing is slightly reduced, but is still widespread and can exceed -20 m locally (i.e. in the Caribbean). The effect of ocean colour on the IL depth is minimum in winter. However, local changes can be up to -18 m in the Caribbean and up to -15 m in the Northern Plume. The annual mean decrease of IL depth is largest in the Northern

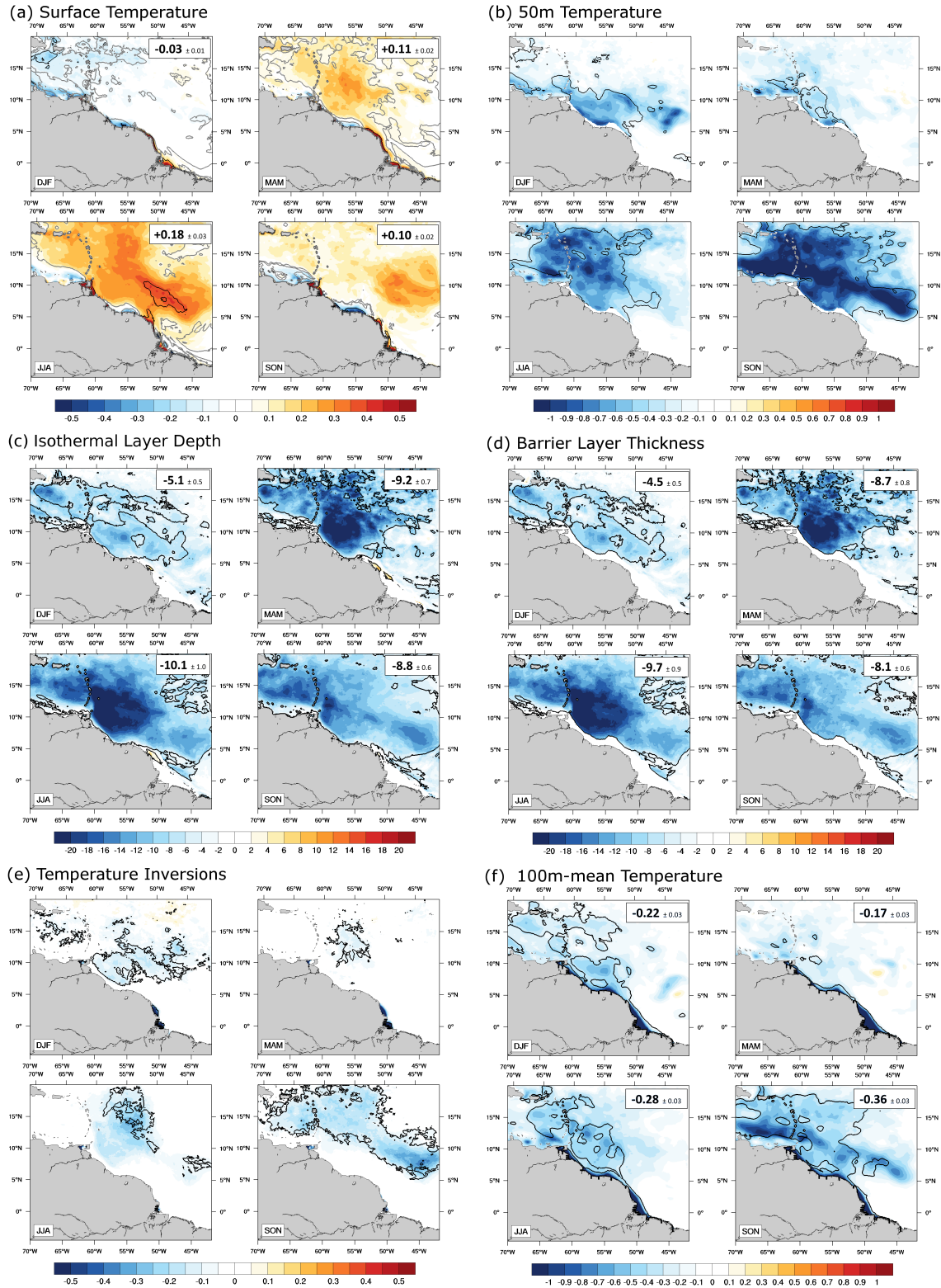


Figure 4.7: Maps of the seasonal cycle (DJF, MAM, JJA, SON) of differences (Coloured-River - Clear-River) in modelled ROMS (2001-2010) (a) SST ($^{\circ}\text{C}$), (b) 50 m temperature ($^{\circ}\text{C}$), (c) IL depth (m), (d) BL thickness (m), (e) mean TIs ($^{\circ}\text{C}$), and (f) T_{100} ($^{\circ}\text{C}$). The black contours show the 95% significance level obtained from a two-sided students t-test, and the grey contours in (a) are the 99% significance level from a one-sided students t-test (von Storch and Zwiers, 1999). The respective area mean values and their standard deviations are given in the top right corner of each map.

Plume and the Caribbean (-11.8 m each) and slightly smaller in the Eastern Plume (-8.9 m). The uplift of the top of the thermocline implies shallowing of the BL by an almost equal amount (about -8 m on average). The spatial distribution mirrors the change in IL depth almost exactly (Figure 4.7d). However, the slight thinning of the ML counteracts BL shallowing, leading to annual mean changes of -11.1 m in the Northern Plume, -9.1 m in the Eastern Plume, and -11.8 m in the Caribbean (Table 4.3).

Even though the ocean subsurface is strongly cooled, the BL temperature does not change significantly (Table 4.3). Note that by definition, the BL will always lie above the region of a 0.2°C temperature drop from the surface. Therefore the consequence of subsurface cooling at similar ML depth can only be to shallow the BL, but not to change its temperature.

Nevertheless, subsurface cooling halves the occurrence of temperature inversions (TIs). They are present in 17% of all days and places in the Coloured-River experiment, compared to 31% percent in the Clear-River experiment. TIs are particularly reduced in summer and autumn (Figure 4.7e), but are still a prominent feature in winter north of 10°N (Figure 4.8). The summer inversions entirely disappear, as do the inversions in the Eastern Plume in autumn. However, weak TIs remain in the Northern Plume in autumn (up to 0.2°C). This illustrates the different formation mechanism of TIs in the southern plume area, and winter TIs in the northern model domain. The plume inversions are a direct result of the BL being isolated from surface cooling, while entrainment cooling is reduced. The northern TIs develop as a consequence of trapping warm waters below the advected freshwater plume, and isolating them from the onset of surface cooling in winter (Mignot et al., 2012).

The impact of ocean colour on the 100 m mean temperature (T_{100}) is a combination of the warmer surface and the colder subsurface. As seen in the vertical temperature profiles, and the 50 m temperature reduction, subsurface cooling outnumbers surface heating by far (Figure 4.7f). T_{100} decreases by about -0.3°C on average. The effect is largest (> -1.0°C) along the South American coastline, where the euphotic zone depth is shallowest (see Figure 4.3a). In winter there is a substantial reduction from the Brazil coast to 10°N between 60°W to 50°W (up to -0.5°C). Winter cooling is also strong in the Caribbean (up to -0.45°C) and the Northern Plume (up to -0.3°C). In spring the modified light penetration hardly affects offshore temperatures. There is a widespread T_{100} decrease in summer in the Caribbean, north of the Brazil coast and in the Northern Plume (up to -0.5°C). Changes are maximum in autumn, affecting the whole plume area. Temperature decreases of up to -1.0°C are then found in the Caribbean. Ocean colour causes the largest annual mean cooling in the Caribbean

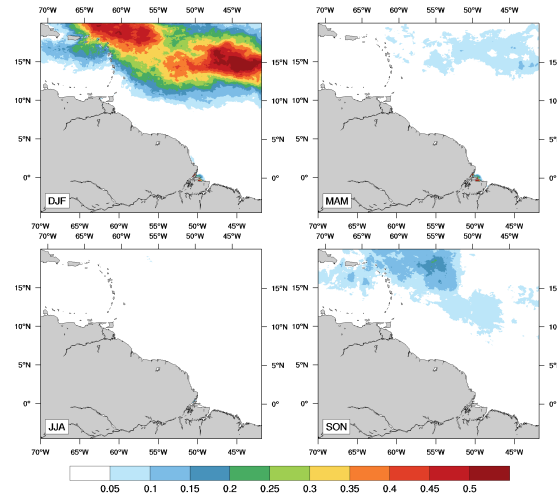


Figure 4.8: Maps of the seasonal cycle (DJF, MAM, JJA, SON) of modelled ROMS (2001-2010) TI magnitude ($^{\circ}\text{C}$) in the Coloured-River experiment.

(-0.38°C) and slightly less in the Eastern (-0.30°C) and Northern (-0.28°C) Plumes (Table 4.3). As mentioned above, the changes in IL depth are an indication for temperature changes in the ocean. Therefore the spatial distribution and seasonal cycle of T_{100} decrease and top of the thermocline lifting are fairly similar. Note however, that there are some discrepancies. For instance, strong IL shallowing just north of Brazil (from 60°W to 50°W) in spring and in the Eastern Plume in summer (Figure 4.7c) do not have a correspondingly large 100m mean temperature signal. That is because the T_{100} cooling is attenuated in areas of strong surface warming.

4.3.2.2 Tropical cyclone metrics

Subsurface cooling naturally induces a reduction in tropical cyclone heat potential (TCHP) (about $-7 \text{ kJ}/\text{cm}^2$ (-15%) on average). The seasonal cycle and spatial extent of TCHP reduction (Figure 4.9) is very similar to 100m mean temperature change (Figure 4.7f). Subsurface cooling is less distinct in winter and spring, therefore TCHP changes are small. However, in winter TCHP decreases by up to $-21 \text{ kJ}/\text{cm}^2$ in the Caribbean and by up to $-12 \text{ kJ}/\text{cm}^2$ in the Northern plume. There are hardly any significant TCHP reductions in spring. In summer TCHP decreases significantly in the Caribbean and the Northern Plume with local reductions of up to $-23 \text{ kJ}/\text{cm}^2$. Ocean colour has the biggest effect on TCHP in autumn, inducing significant reductions across the whole plume area. The decrease is then largest in the Caribbean (up to $-25 \text{ kJ}/\text{cm}^2$) and slightly lower in the Northern (up to $-23 \text{ kJ}/\text{cm}^2$) and Eastern Plume (up to $19 \text{ kJ}/\text{cm}^2$). The annual mean TCHP changes are $-8.9 \text{ kJ}/\text{cm}^2$ (-17%) in the Northern Plume, $-7.1 \text{ kJ}/\text{cm}^2$ (-13%) in the Eastern plume, and $-11.9 \text{ kJ}/\text{cm}^2$ (-19%)

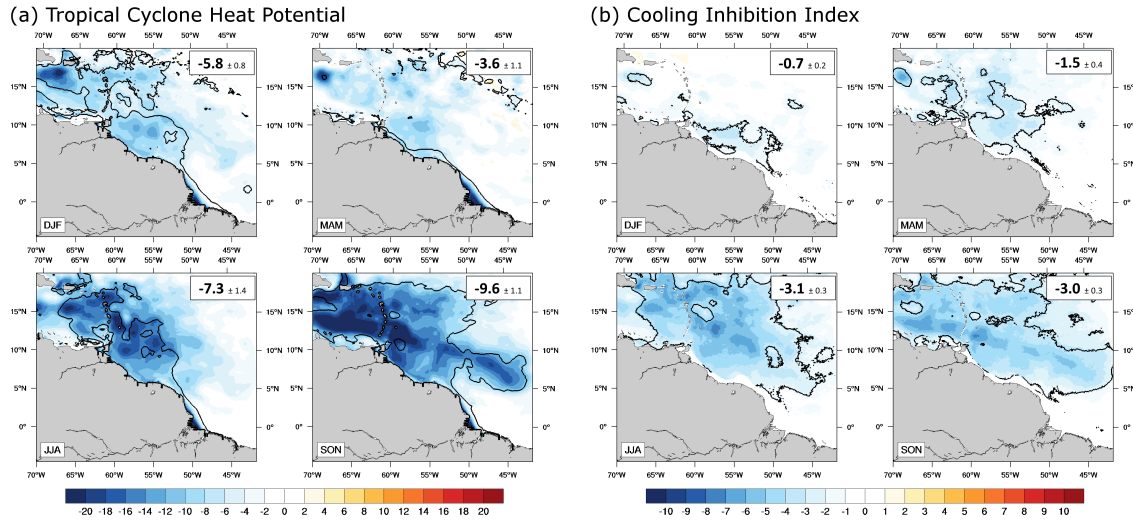


Figure 4.9: Maps of the seasonal cycle (DJF, MAM, JJA, SON) of differences (Coloured-River - Clear-River) in modelled ROMS (2001-2010) (a) TCHP (kJ/cm^2) and (b) CI ($(\text{J}/\text{m}^2)^{1/3}$). The black contours show the 95% significance level obtained from a two-sided students t-test (von Storch and Zwiers, 1999). The respective area mean values and their standard deviations are given in the top right corner of each map.

in the Caribbean. Compared to the freshwater effect discussed in Section 3.3.2, the impact of ocean colour on TCHP is about 85% larger.

Light absorbing particles reduce ocean temperatures and shift cold waters to shallower depths. The surface can therefore be cooled more easily by passing storms. This is reflected in an average cooling inhibition index (CI) decrease of $-2.1 (\text{J}/\text{m}^2)^{1/3}$ (6%). The regional distribution and annual variation of the CI reduction largely reflects the changes in TCHP and 100 m mean temperature. However, in winter there is significant TCHP decrease in the Caribbean and the Northern Plume, which is not the case for CI. Note that the surface is cooled at that time of the year. TCHP sees this change, since it is a measure for the absolute heat available for TC development. CI on the other hand is blind for the surface cooling, since it is only sensitive to sub-surface cooling relative to the SST. In spring CI decreases by up to $-6 (\text{J}/\text{m}^2)^{1/3}$ in the Caribbean and by up to $-4 (\text{J}/\text{m}^2)^{1/3}$ in the Northern Plume. The reduction in CI is most prominent in the North Atlantic hurricane season, when the concentration of light absorbing particles is maximum. In summer it stretches across the whole plume area and can be up to $-8 (\text{J}/\text{m}^2)^{1/3}$ in the Northern Plume and up to $-7 (\text{J}/\text{m}^2)^{1/3}$ in the Caribbean. Colour induced CI reductions are even more widespread in autumn, with a maximum in the Caribbean (up to $-7 (\text{J}/\text{m}^2)^{1/3}$). The annual mean CI change is largest in the Caribbean ($-3.1 (\text{J}/\text{m}^2)^{1/3}$, -8%) and is also important in the Northern

Effect of ocean colour

	Whole main	Do- main	Northern Plume	Eastern Plume	Caribbean
Salinity (PSU)					
Surface	-0.0 ± 0.0		-0.05 ± 0.05	0.05 ± 0.04	-0.03 ± 0.05
Temperature (°C)					
Surface	+0.08 ± 0.01		+0.12 ± 0.02	+0.12 ± 0.02	+0.05 ± 0.02
Barrier Layer	+0.00 ± 0.01		-0.00 ± 0.02	-0.02 ± 0.02	-0.02 ± 0.03
100m-mean	-0.26 ± 0.03		-0.28 ± 0.04	-0.30 ± 0.1	-0.38 ± 0.06
Layer Thickness (m)					
Mixed Layer	-0.4 ± 0.1		-0.7 ± 0.2	-0.3 ± 0.1	-1.0 ± 0.3
Barrier Layer	-7.9 ± 0.5		-11.1 ± 1.3	-9.1 ± 1.7	-10.9 ± 1.8
Isothermal Layer	-8.3 ± 0.6		-11.8 ± 1.3	-8.9 ± 1.6	-11.8 ± 1.7
TC Metrics					
TCHP (kJ/cm ²)	-6.6 ± 0.9		-8.9 ± 1.2	-7.1 ± 1.5	-11.9 ± 2.8
CI (J/m ²) ^{$\frac{1}{3}$}	-2.1 ± 0.2		-2.9 ± 0.3	-2.2 ± 0.6	-3.1 ± 0.5

Table 4.3: The effect of ocean colour (Coloured-River - Clear-River) on annual mean salinity, temperatures, layer thicknesses and TC metrics with their standard deviation in the whole model domain, the Northern Plume, the Eastern Plume and the Caribbean.

Plume (-2.9 (J/m²) ^{$\frac{1}{3}$} , -7%). The colour-induced absolute CI reduction is smaller in the Eastern Plume, but the relative change is larger (-2.1 (J/m²) ^{$\frac{1}{3}$} , 10%).

Note that subsurface cooling increases stratification and hence the energy required for mixing. It may therefore contribute positively to a change in CI. That effect can be estimated by calculating CI for equal mixing depths, i.e. neglecting the fact that the water column does not need to be mixed as deeply in the Coloured-River experiment. That leads to only +0.1 (J/m²) ^{$\frac{1}{3}$} larger CI in the Coloured-River experiment compared to the Clear-River experiment, and is thus negligible. In contrast to the effect of freshwater discussed in Section 3.3.2, ocean colour does not change the density stratification substantially. Therefore TCHP and CI should predict a similar SST feedback. Both metrics indicate that ocean colour acts to increase cooling feedbacks, and therefore potentially weakens TCs passing across the Amazon and Orinoco plume.

4.3.3 The combined freshwater and colour effect

The results of Section 3.3.2 suggest that the freshwater effect of the Amazon and the Orinoco leads to a more stable water column and therefore a potential reduction in surface cooling feedbacks on TCs. Light absorbing particles in the plume on the other hand, cool the ocean subsurface and are therefore likely to enhance SST feedbacks. The net effect will be a combination of the two mechanisms. The following 'Net' differences are between the No-River and Coloured-River experiments (Table 2.2).

Note that the Coloured-River case does include freshwater effects as well as light attenuation. Therefore this comparison shows the impact of the coloured freshwater plume on ocean stratification and temperatures. The analysis is performed for the seasonal means, namely winter (December, January, February (DJF)), spring (March, April, May (MAM)), summer (June, July, August (JJA)), and autumn (September, October, November (SON)) for the model years 2001-2010, and the results of relevant parameters for the whole model domain, the Northern and Eastern Plume, and the Caribbean are summarized in Table 4.4.

4.3.3.1 Ocean temperatures and stratification

The vertical distribution of temperature changes caused by the clear freshwater plume, ocean colour, and the coloured freshwater plume (satellite and salinity-derived) are displayed in Figure 4.10. From June to November the freshwater-induced subsurface warming is small compared to the colour-induced subsurface cooling. However, their maximum impacts occur at different depths. At about 30 m BL heating shifts the profile to warmer temperatures (by about $+0.1^{\circ}\text{C}$), leading to a Net cooling of -0.15°C . At 50 m the temperature change is almost solely determined by ocean colour (-0.5°C). Below the IL the BL causes a temperature reduction, increasing colour-induced cooling. At the surface the slight average warming in the Clear-River experiment adds to the warming in the Coloured-River experiment, leading to a Net of about $+0.11^{\circ}\text{C}$. Results are similar between December and May. BL induced warming is maximum at about 55 m ($+0.05^{\circ}\text{C}$), reducing the cooling ocean colour effect (-0.2°C) to a Net impact of about -0.15°C . Due to the cancellation of warming and cooling, the maximum cooling does not occur where the ocean colour effect is largest ($+0.05^{\circ}\text{C}$), but at 80 m, where BL induced temperature changes become insignificant or even turn negative.

The Net temperature change obtained from the salinity-approximation is very similar to the Lee et al. (2005) method (Figure 4.10). However, surface warming and subsurface cooling in the Water-Type experiment are slightly smaller in the hurricane season (-0.02°C less surface warming, $+0.06^{\circ}\text{C}$ less subsurface cooling at 55 m). Compared to the overall colour-induced change in ocean temperatures, the difference in the two approximations is negligible. The seasonal maps for the salinity approximation are therefore not discussed in the following, but the results are summarized in Table 4.5 for reference.

As seen in Section 3.3.2 (Figure 3.10b), the freshwater plume modifies offshore SSTs only slightly. The spatial pattern of Net SST change (Figure 4.11a) is largely a

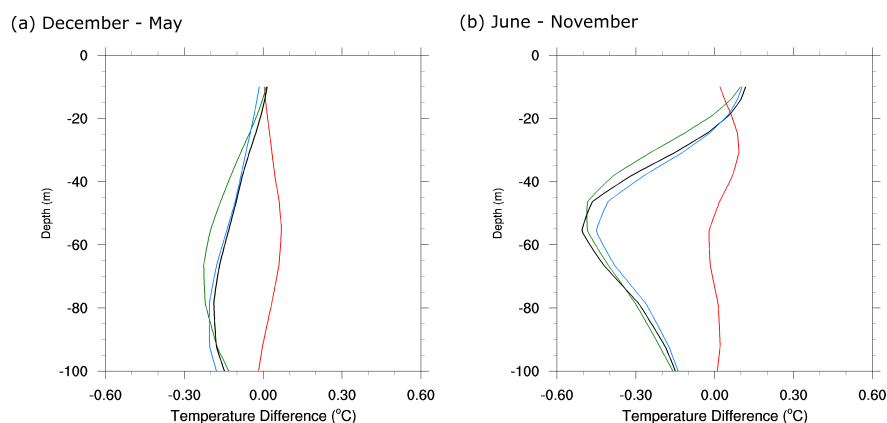


Figure 4.10: Differences in domain mean modelled (2001-2010) vertical temperature ($^{\circ}\text{C}$) profiles as a function of ocean depth (m) in (a) December to May and (b) June to November for Clear-River - No-River (red), Coloured-River - Clear-River (green), Coloured-River - No-River (black), Water Type - No-River (blue).

reflection of the Coloured-River experiment (Figure 4.7a). Most SST modifications are only significant under a one-sample t-test. The annual mean SST increase is 0.1°C . In winter surface temperatures are reduced in the Caribbean and the Northern Plume in both, the Clear-River and the Coloured-River experiment. Therefore, the Net impact of the coloured plume is an even larger cooling of up to -0.3°C . From March to November the Net SST response is an increase in the whole plume area (except near the river mouths). The increase is largest in summer (up to 0.5°C) in the Northern Plume and north of the Amazon mouth. Annual mean heating is $+0.08^{\circ}\text{C}$ in the Northern Plume, $+0.13^{\circ}\text{C}$ in the Eastern Plume and $+0.07^{\circ}\text{C}$ in the Caribbean. In the coastal upwelling regions, river-induced BL and colour effects are large, but of opposite sign. Therefore the coloured plume causes only a small Net response in this region.

As seen in Section 3.3.2 (Figure 3.13c), the clear river induces TIs (about 31% occurrence) from summer to winter, with a maximum in winter. On the other hand, ocean colour cools the subsurface and reduces these inversions to about 17% occurrence. The Net occurrence of TIs in the Coloured-River experiment is shown in Figure 4.8 and has been discussed in Section 4.3.2. However, ocean colour mainly reduces TIs in summer and autumn, south of 10°N (Figure 4.7c). Therefore the Net effect of the coloured river plume is to induce strong inversions in winter north of 10°N (Figure 4.7b). The inversions induced by the coloured plume are up to 0.5°C . Weaker TIs (up to 0.2°C) develop in autumn north of 15°N .

Ocean colour modifies the ML depth only slightly, therefore Net ML changes are mainly determined by surface freshening and are -2.6 m on average. For comparison,

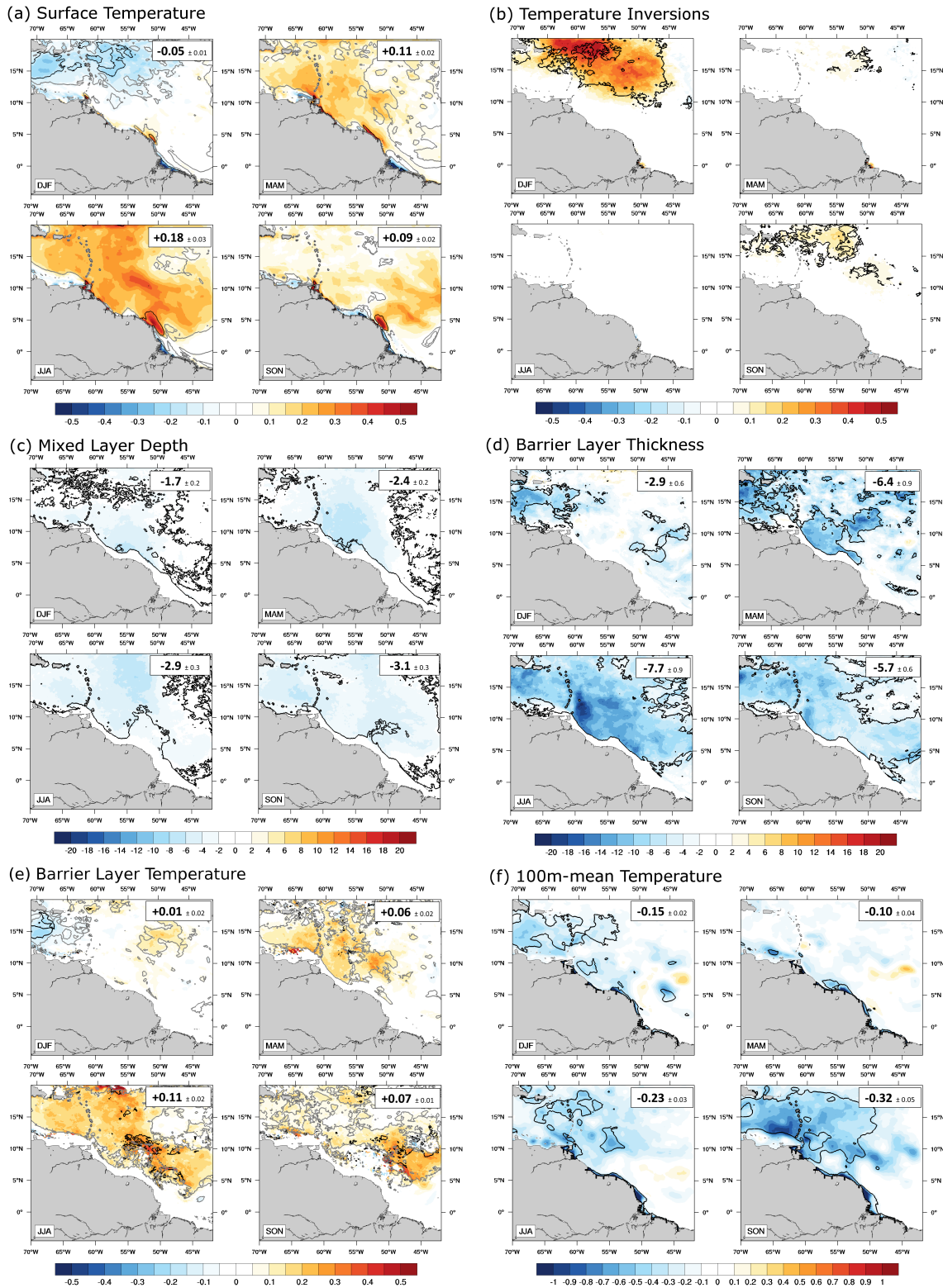


Figure 4.11: Maps of the seasonal cycle (DJF, MAM, JJA, SON) of differences (Coloured-River - No-River) in modelled ROMS (2001-2010) (a) SST ($^{\circ}\text{C}$), (b) mean TIs ($^{\circ}\text{C}$), (c) ML depth (m), (d) BL thickness (m), (e) BL temperature ($^{\circ}\text{C}$), and (f) 100m-mean temperature ($^{\circ}\text{C}$). The black contours show the 95% significance level obtained from a two-sided students t-test, and the grey contours in (a) and (e) are the 99% significance level from a one-sided students t-test (von Storch and Zwiers, 1999). The respective area mean values and their standard deviations are given in the top right corner of each map.

ML shallowing induced by the clear river was -1.8 m (Table 3.1). The seasonal and spatial modulations (Figure 4.11c) mirror freshwater-induced ML shallowing seen in Figure 3.13a. There is an extensive uplift of up to -10 m locally, throughout the entire year in the plume area.

In contrast to ML depth, changes in the isothermal layer thickness are mainly determined by the colour component of the river plume. The annual mean change is a -8.4 m uplift. The Net impact of the coloured plume on the BL is a combination of freshwater-induced thickening and colour-induced shallowing, leading to a Net thinning of the BL by -5.7 m on average. Note however, that the individual changes (colour and freshwater caused) are due to different mechanisms. In the Clear-River experiment the BL-top is lifted because of higher stability, whereas ocean colour shallows the BL from below. While this means that colder water is closer to the surface, it only leads to a modest increase in ocean stability. The seasonal and spatial Net changes (Figure 4.11d) are similar to the colour-induced BL shallowing seen in Figure 4.7d. The colour component of the plume overwhelms the freshwater effect on BL thickness throughout the whole year, across the entire model domain. It is only in areas of large freshwater-induced BL thickening, that colour-induced BL thinning is significantly reduced (e.g. in spring in the southern part of the Northern Plume). In winter the Net effect of the coloured plume is to slightly shallow the BL in the Caribbean and the Eastern Plume by up to -15 m. In spring thinning of up to -20 m occurs in the Caribbean, and there is extensive BL shallowing of up to -18 m in the remaining WTNA. In summer the coloured plume thins the BL across the entire model domain with a maximum in the Northern Plume (up to -20 m). The effect is slightly smaller, but of similar spatial extent in autumn. The annual mean impact of the coloured plume is BL thinning of -6.4 m and -6.0 m the Northern and Eastern Plume, respectively. The largest Net shallowing occurs in the Caribbean (-8.4 m).

In the Clear-River experiment the BL heats, due to isolation from surface cooling and reduced entrainment from below the IL. The annual mean Net BL warming caused by the coloured rivers is $+0.08^{\circ}\text{C}$, and is thus similar to the heating caused by the clear river. Keep in mind however, that the BL is now shallower. The overall contribution to heating of the ocean will therefore be reduced. The spatial and seasonal distribution of BL heating is very similar to the Clear-River case (Figure 3.13d), but locally BL heating in the Coloured-River experiment can be larger or smaller (Figure 4.11e). In contrast to the Clear-River experiment, the coloured river cools the Caribbean BL by up to -0.25°C in winter, when the IL deepens. The annual mean BL heating is $+0.13^{\circ}$ in the Northern Plume, $+0.11^{\circ}\text{C}$ in the Eastern Plume, and $+0.08^{\circ}\text{C}$ in the Caribbean. Hence the BL warms about -0.02°C less in those areas, compared

to the Clear-River case.

The Net effect of the coloured plume is to reduce the 100 m mean temperature (T_{100}) by -0.2°C on average (Table 4.4). This is slightly less compared to the colour-induced component (-0.26°C), due to BL heating from the clear plume. The seasonal and spatial distribution of T_{100} cooling (see Figure 4.11f) is very similar to the temperature decrease seen in the colour-only analysis above. However, in regions of distinct BL warming, T_{100} cooling is reduced. This is for example the case in the Eastern Plume in autumn. 100 m mean cooling by the coloured river is hardly significant in spring, but in winter and summer T_{100} is reduced in the Caribbean (up to -0.4°C) and the Northern Plume (up to -0.5°C). The Net cooling is most extensive in autumn and stretches across the entire plume. T_{100} can then cool by up to -1.3°C in the Caribbean. The average T_{100} decrease caused by the coloured plume is -0.26°C in the Northern Plume, -0.17°C in the Eastern Plume, and maximum in the Caribbean (-0.34°C). Hence BL heating caused by river freshwater reduces colour-induced cooling by $+0.02^{\circ}\text{C}$, $+0.03^{\circ}\text{C}$, and $+0.04^{\circ}\text{C}$, respectively.

4.3.3.2 Tropical cyclone metrics

The tropical cyclone heat potential (TCHP) and the T_{100} are both measures for the heat contained in the upper ocean. Therefore they behave similar under the influence of the coloured river. The spatial and seasonal distribution of TCHP decrease (Figure 4.12a) is almost identical to the colour-induced signal (Figure 4.9a). However, it is slightly reduced due to the modest TCHP increase caused by BL heating. The mean TCHP change is -5.6 kJ/cm^2 (-14%). In winter the coloured plume decreases TCHP by up to -23 kJ/cm^2 in the Caribbean. There is no significant TCHP change in spring, but in summer TCHP decreases by up to -19 kJ/cm^2 in the Caribbean and by up to -17 kJ/cm^2 in the Northern Plume. The reduction in TCHP is maximum in terms of magnitude and spatial extent in autumn, and can be up to -25 kJ/cm^2 in the Caribbean and up to -23 kJ/cm^2 in the Northern plume. The TCHP decrease is particularly large in areas where a freshwater induced TCHP decrease amplifies the cooling signal (i.e north of 10°N in the Northern Plume). TCHP in the Eastern plume decreases only modestly, since strong BL heating dampens the effect of subsurface cooling. The Net effect of the coloured plume is maximum in the Caribbean (-10.8 kJ/cm^2), slightly smaller in the Northern Plume (-7.8 kJ/cm^2) and minimum in the Eastern plume (-4.3 kJ/cm^2). That implies that BL heating dampens colour-induced TCHP reductions by $+1.3 \text{ kJ/cm}^2$, $+2.8 \text{ kJ/cm}^2$ and $+1.1 \text{ kJ/cm}^2$, respectively.

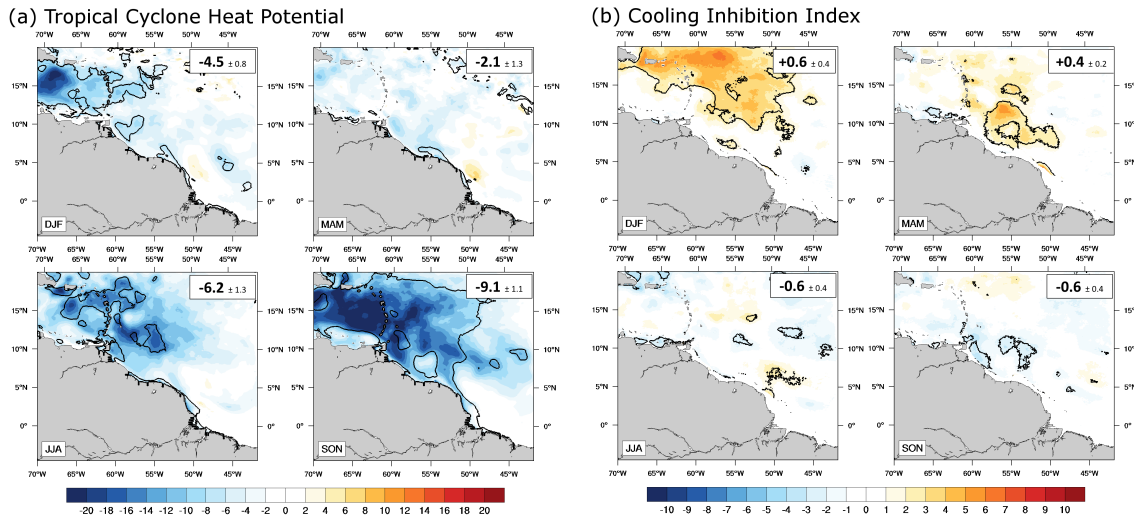


Figure 4.12: Maps of the seasonal cycle (DJF, MAM, JJA, SON) of the Net impact of the coloured river plume (Coloured-River - No-River) on modelled ROMS (2001-2010) (a) TCHP (kJ/cm^2) and (b) CI ($(\text{J}/\text{m}^2)^{\frac{1}{3}}$). The black contours show the 95% significance level obtained from a two-sided students t-test (von Storch and Zwiers, 1999). The respective area mean values and their standard deviations are given in the top right corner of each map.

As discussed in Section 3.3.2, TCHP does not consider stability changes caused by the freshwater plume. The dominance of the colour effect is therefore not surprising. The contribution of the freshwater component of the coloured plume becomes more important when considering the cooling inhibition index (CI). While the freshwater plume increases stability and therefore CI, ocean colour reduces the mixing energy required to cool the ocean surface. The Net annual mean change caused by the coloured plume is a small CI increase of only $+0.1 (\text{J}/\text{m}^2)^{\frac{1}{3}}$ (Table 4.4). Regionally CI increases by $+1.2 (\text{J}/\text{m}^2)^{\frac{1}{3}}$ in the Northern Plume, $+0.4 (\text{J}/\text{m}^2)^{\frac{1}{3}}$ in the Eastern Plume and $+0.1 (\text{J}/\text{m}^2)^{\frac{1}{3}}$ in the Caribbean. The seasonal cycle is shown in Figure 4.12b. In winter and spring, the river-freshwater effect seems to overwhelm any CI reduction caused by subsurface cooling ($+1.4$ and $+0.4 (\text{J}/\text{m}^2)^{\frac{1}{3}}$, respectively). The large Net increase between about 10°N and 20°N in winter (up to $8 (\text{J}/\text{m}^2)^{\frac{1}{3}}$ locally) is due to a combination of increased stability and persisting TIs. In the hurricane season (June-November) river-freshwater and colour effects largely cancel each other. The Net effect is even a slight reduction in CI ($-0.6 (\text{J}/\text{m}^2)^{\frac{1}{3}}$). This implies that there may only be a marginal average consequence on SST feedbacks induced by the Amazon and Orinoco.

However, there are significant month to month variations (Figure 4.13). The June

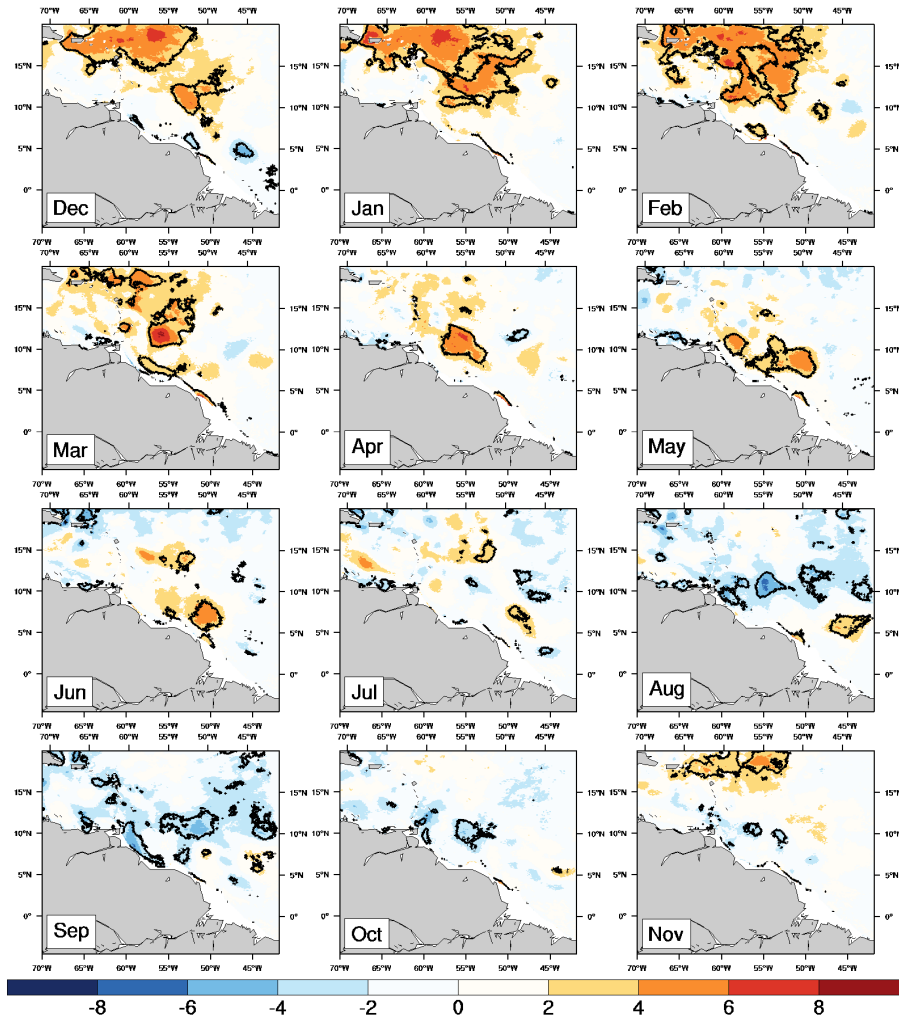


Figure 4.13: Monthly mean maps of the Net impact of the coloured river plume (Coloured-River - No-River) on modelled ROMS (2001-2010) CI ($(\text{J/m}^2)^{\frac{1}{3}}$). The black contours show the 95% significance level obtained from a two-sided students t-test.

and July maps show small areas of increasing/decreasing CI of up to $+6 (\text{J/m}^2)^{\frac{1}{3}} / -4 (\text{J/m}^2)^{\frac{1}{3}}$. In August, September and October, the Net is mostly negative. There is a significant decrease of up to $-6 (\text{J/m}^2)^{\frac{1}{3}}$ between about 10°N and 15°N . North of 15°N CI remains virtually unchanged. In November CI increases significantly north of 15°N (up to $+6 (\text{J/m}^2)^{\frac{1}{3}}$).

4.4 Discussion

4.4.1 Light attenuation algorithms

To account for a more realistic heat transfer into the upper ocean, a recently developed light attenuation algorithm (Lee et al., 2005) has been used. The advantage of

Net effect of the coloured plume (Lee et al. (2005))

	Whole main	Do- main	Northern Plume	Eastern Plume	Caribbean
Salinity (PSU)					
Surface	-1.3 ± 0.0		-1.5 ± 0.1	-1.7 ± 0.1	-1.1 ± 0.1
Temperature (°C)					
Surface	+0.08 ± 0.01		+0.08 ± 0.02	+0.13 ± 0.03	+0.07 ± 0.02
Barrier Layer	+0.08 ± 0.01		+0.13 ± 0.02	+0.11 ± 0.04	+0.08 ± 0.02
100m-mean	-0.20 ± 0.03		-0.26 ± 0.04	-0.17 ± 0.10	-0.34 ± 0.10
Layer Thickness (m)					
Mixed Layer	-2.6 ± 0.2		-4.0 ± 0.4	-3.1 ± 0.4	-3.1 ± 0.4
Barrier Layer	-5.7 ± 0.6		-6.4 ± 1.1	-6.0 ± 1.4	-8.3 ± 1.7
Isothermal Layer	-8.1 ± 0.6		-10.4 ± 1.1	-8.9 ± 1.4	-11.4 ± 1.8
TC Metrics					
TCHP (kJ/cm ²)	-5.6 ± 1.0		-7.8 ± 1.4	-4.3 ± 1.6	-10.8 ± 3.1
CI (J/m ²) ^{$\frac{1}{3}$}	+0.1 ± 0.2		+1.2 ± 0.4	+0.4 ± 0.6	+0.1 ± 0.8

Table 4.4: Net effect of the coloured plume derived with Lee et al. (2005) (Coloured-River - No-River) on annual mean salinity, temperatures, layer thicknesses and TC metrics with their standard deviation in the whole model domain, the Northern Plume, the Eastern Plume and the Caribbean.

Net effect of the coloured plume (salinity-approximation)

	Whole main	Do- main	Northern Plume	Eastern Plume	Caribbean
Salinity (PSU)					
Surface	-1.3 ± 0.05		-1.4 ± +0.1	-1.7 ± 0.1	-1.1 ± 0.1
Temperature (°C)					
Surface	+0.06 ± 0.03		+0.07 ± 0.03	+0.10 ± 0.04	+0.04 ± 0.04
Barrier Layer	+0.06 ± 0.03		+0.11 ± 0.03	+0.07 ± 0.05	+0.04 ± 0.04
100m-mean	-0.18 ± 0.04		-0.25 ± 0.04	-0.21 ± 0.15	-0.27 ± 0.09
Layer Thickness (m)					
Mixed Layer	-2.5 ± 0.2		-3.9 ± 0.4	-3.2 ± 0.3	-3.0 ± 0.3
Barrier Layer	-5.0 ± 0.8		-6.0 ± 1.4	-5.4 ± 1.2	-6.7 ± 1.7
Isothermal Layer	-7.4 ± 0.7		-9.9 ± 1.4	-8.4 ± 1.1	-9.7 ± 1.8
TC Metrics					
TCHP (kJ/cm ²)	-5.4 ± 1.2		-7.4 ± 1.4	-4.6 ± 1.9	-8.8 ± 2.7
CI (J/m ²) ^{$\frac{1}{3}$}	+0.4 ± 0.2		+1.4 ± 0.2	+0.4 ± 0.7	+0.9 ± 0.6

Table 4.5: Net effect of the coloured plume under the salinity-approximation (No-River - Water Type) on annual mean salinity, temperatures, layer thicknesses and TC metrics with their standard deviation in the whole model domain, the Northern Plume, the Eastern Plume and the Caribbean.

it being, that it considers all light absorbing particles, rather than just chlorophyll. However, only backscattering and absorption coefficients at 490 nm are used in this approach. Lee et al. (2005) argue that in terms of solar heating positive and negative deviations from the 490 nm properties cancel on average. Light attenuation is solely derived from surface values, subsurface properties are inherently prescribed. This may turn out problematic when applied below the well mixed surface layer. However, there is no satellite product which would reveal the subsurface distribution of light absorbing substances to this date. Diurnal and seasonal variations in the solar zenith angle are not considered in this study, assuming an average angle of 45° for the tropics. Ohlmann (2003) estimate that the mean error of neglecting variations in the solar zenith angle is only a few percent, and conclude that the sun angle is of little importance in regulating solar transmission on climate model scales.

While CDOM commonly has terrestrial sources (eg. Del Vecchio and Subramaniam (2004)), chlorophyll in the WTNA should not solely be assigned to the Amazon and Orinoco. Eddy, wind-driven or equatorial upwelling can also act as nutrient sources and therefore support primary production (Pennington et al., 2006), as do the upwelling regions along the Caribbean coastline (Hu et al., 2004). Del Vecchio and Subramaniam (2004) find that chlorophyll abundance is the dominant contributor to light absorption outside the plume. The ocean colour effect should therefore rather be seen as the combined effect of all particle and nutrient sources in the tropical North Atlantic. However, Hu et al. (2004) find a negative correlation between chlorophyll concentration and surface salinity in the main plume season. Similarly, measurements of Subramaniam et al. (2007) suggest that light absorption is significantly enhanced in low salinity regimes. Furthermore the salinity-approximation used here produces similar results to the satellite-derived ocean colour, suggesting the the main part of light-absorbing particles in the WTNA is delivered by the Amazon and Orinoco.

Surprisingly satellite- and salinity-derived light absorption produce the largest temperature and stratification differences from June to November, the season where light attenuation profiles are fairly similar. Simson and Dickey (1981) and Sweeney et al. (2005) suggest that light penetration is less important in areas of deep ILs, since the radiation is distributed across a thicker layer. Note, that the IL in the WTNA is deep in winter (about 60 m), while the isothermal layer depth is only about 30 m in summer. Keeping this in mind, it is easy to understand, why the choice of absorption algorithm may have a bigger impact in the hurricane season. Note that the good agreement in ocean temperatures and stratification may partly be a result of cancelling over- and underestimation of euphotic zone depth in the Water-Type experiment (Figure 4.3). However, the biggest changes in temperatures do not occur,

where the difference in light depth is maximum.

The largest temperature differences between satellite- and salinity-derived ocean colour occurs near the river mouths. This may be due to the limitation of the salinity method to a maximum water type of 5. Note that the light can still travel to 25 m for the muddiest Jerlov waters. The satellite derived algorithm suggests euphotic zone depths of only a few meters near the river outlets. This is in agreement with measurements of Odriozola (2004), who found that the penetration level in the Gulf of Paria can be below 1 m. This extremely low penetration depth is mainly due to the vast amount sediments being transported into the ocean (Warne et al., 2002).

Apart from the river outlets, the difference in the two algorithms is large in regions where additional nutrient sources enhance light attenuation (e.g. in the southern Caribbean), for which the salinity-approximation can not account. Does this mean that the the salinity-approximation may be interpreted as the river-induced component of ocean colour? The Water-Type solution slightly overestimates light attenuation in the in northern high plume season (Figure 4.3c). This suggests, that assigning water type 4 to all mesohaline areas may be too extreme. Attributing the entire impact derived with the Water Type experiment to the Amazon and Orinoco may therefore be an overestimation. A more gradual decrease of water types with salinity increase may be desirable for such an estimate. Compared to the overall impact of ocean colour, the difference in satellite and salinity derived light absorption is marginal. Therefore the salinity approximation seems a viable route in the Amazon and Orinoco plume region to estimate the effect of ocean colour on ocean stratification and temperatures.

4.4.2 The ocean colour effect

Realistic solar transmission tends to warm the ocean surface. In this study any contribution of the Amazon and Orinoco to the western hemisphere warm pool will therefore be through light attenuation, rather than surface freshening. Murtugudde et al. (2002) and Frouin et al. (2007) modelled an annual mean SST increase of up to $+0.2^{\circ}\text{C}$ and $+0.8^{\circ}\text{C}$, respectively, in the Amazon and Orinoco plume area when accounting for chlorophyll absorption. Similarly Lee et al. (2005b) find that decreasing the attenuation depth from water type 1 to realistic attenuation depths increases simulated SSTs by up to 0.2°C .

The SST increase in the Coloured-River and Water-Type experiments further amplifies the warm SST bias in the model simulation to about $+0.6^{\circ}\text{C}$ on average. As mentioned above atmospheric damping may lead to an underestimate of the SST response to ocean colour. The SST increase may therefore be even larger in a coupled

environment. On the other hand, Lengaigne et al. (2007) suggest that atmospheric feedback may modify colour-induced SST anomalies due to a Bjerknes (1969) feedback. In the present study colour induced warming increases the east-west SST gradient, thus potentially enhancing the trade winds and reducing surface temperatures. Furthermore the change in SSTs is likely to modify atmospheric temperatures and circulation (Shell et al., 2003; Gnanadesikan et al., 2010). This study focuses on the oceanic part of TC intensity control and does not include any atmospheric feedbacks.

Subsurface cooling due to modified light absorption in model studies has previously been reported (Lin et al., 2008; Gnanadesikan and Anderson, 2009). Murtugudde et al. (2002) found subsurface cooling of about -1.0°C at 50 m in the Atlantic ocean, when accounting for realistic light attenuation. The surface cooling in the upwelling regions near the coast is a result of cooled subsurface waters reaching the ocean surface and cancelling the effect of surface heating (Lin et al., 2008; Ma et al., 2012). The same effect explains surface cooling in winter when the IL deepens and the cooled subsurface waters are mixed into the ML.

Light absorbing particles in the coloured plume increase the amount of sunlight that is absorbed in the ML. In the Coloured-River experiment, 91% solar radiation are absorbed in the ML, and only 9% penetrate below. For a clear river, the ML lets 20% radiation pass through. That means that the subsurface receives about 48 W/m^2 in the Clear-River experiment, but only 21 W/m^2 for a coloured plume. That is equivalent to a reduction of -56%, and is responsible for the substantial subsurface cooling in the Coloured-River experiment. The cooling outnumbers surface heating, leading to a significant decrease of the upper-ocean heat content. Since the solar flux is constant in both experiment (242 W/m^2), this heat must leave the ocean in the form of enhanced surface fluxes (when neglecting changes in horizontal and vertical advection). Indeed the latent and sensible heat losses are -3.4 W/m^2 and -0.8 W/m^2 larger, respectively. The overall radiation balance of the ML is -25 W/m^2 in the Clear-River experiment, and -3.4 W/m^2 in the Coloured-River experiment, i.e. a heat gain of 80%. However, vertical and horizontal heat transports will modify this balance. Entrained water will for example be colder in the Coloured-River experiment, but stronger stratification may reduce vertical entrainment. Furthermore, dynamical responses to ocean colour may also play an important role for the heat budget (Ballabrera-Poy et al., 2007).

It has been implicitly assumed in this analysis, that ocean colour and freshwater effects are independent. However, ocean colour may influence plume dispersal and thus alter the freshwater effect. Similarly the dispersal of freshwater into the WTNA may impact ocean colour. However, the changes in plume size and surface salinity in the Coloured-River experiment are small. Furthermore the combination of the indi-

vidual freshwater and colour effects almost reproduces the results of the net coloured river impact (see Tables 3.1, 4.3, 4.4). This suggests that such non-linearities in the ocean response may be small.

Earlier studies also found a shallowing of the IL when applying more realistic light absorption profiles. Murtugudde et al. (2002) showed local shoaling of the IL by up to -18 m in the Amazon plume area, which is in agreement with the findings presented here. Similarly Sweeney et al. (2005) find that a 10-18% increase in the euphotic zone depth increases the IL depth by 3-20m in tropical regions.

Ocean colour induced modulations in ocean temperatures are largest from June to November. On the one hand this is due to the shallower attenuation depth in the maximum plume season. Note also, that the isothermal layer is deeper in winter. Changes in the vertical absorption profile thus have less impact on upper ocean temperatures, since the heat will be redistributed across the thicker isothermal layer. Therefore the impact of modified light penetration will be more important, the shallower the IL (Simspon and Dickey, 1981; Sweeney et al., 2005).

The absorption profile in the Coloured-River experiment is more realistic, than the simple water type 1 assumption in the Clear-River experiment. Agreement of modelled ocean temperatures and stratification with observations may therefore be improved. It was shown in Section 3.3.1, that the modelled IL is too deep in the Clear-River experiment by 5-10 m, compared to WOA13 observations. The coloured plume shallows the isothermal layer. As a result the IL is only about 2 m too deep in winter and spring and about 4 m and 2 m too shallow in summer and autumn in the Coloured-River experiment, compared to WOA13 data (Figure 4.14). While the ML is about 2-3 m too deep in the Clear-River experiment, it is slightly shallower (by about -1 m) in the Coloured-River experiment, due to stronger stratification. The improvement in IL and ML depth also leads to a more realistic BL thickness. Overall including realistic light attenuation substantially improves the agreement of modelled and observed ocean stratification.

Increased stratification of the Amazon and Orinoco may inhibit vertical mixing and therefore reduce SST feedbacks. On the other hand, ocean colour counteracts this effect by cooling the subsurface, potentially increasing surface cooling feedbacks. While increased stratification dominates the balance in winter and spring, CI changes are negligible in the North Atlantic hurricane season. Therefore the average impact of the Amazon and Orinoco rivers may be marginal. However, the balance of freshwater and colour effects varies in time and space, and plume size and ocean colour are subject to regional and seasonal variations (e.g. Grodsky et al. (2014); Hu et al. (2004)). There may therefore be river-induced intensifying/de-intensifying effects on individ-

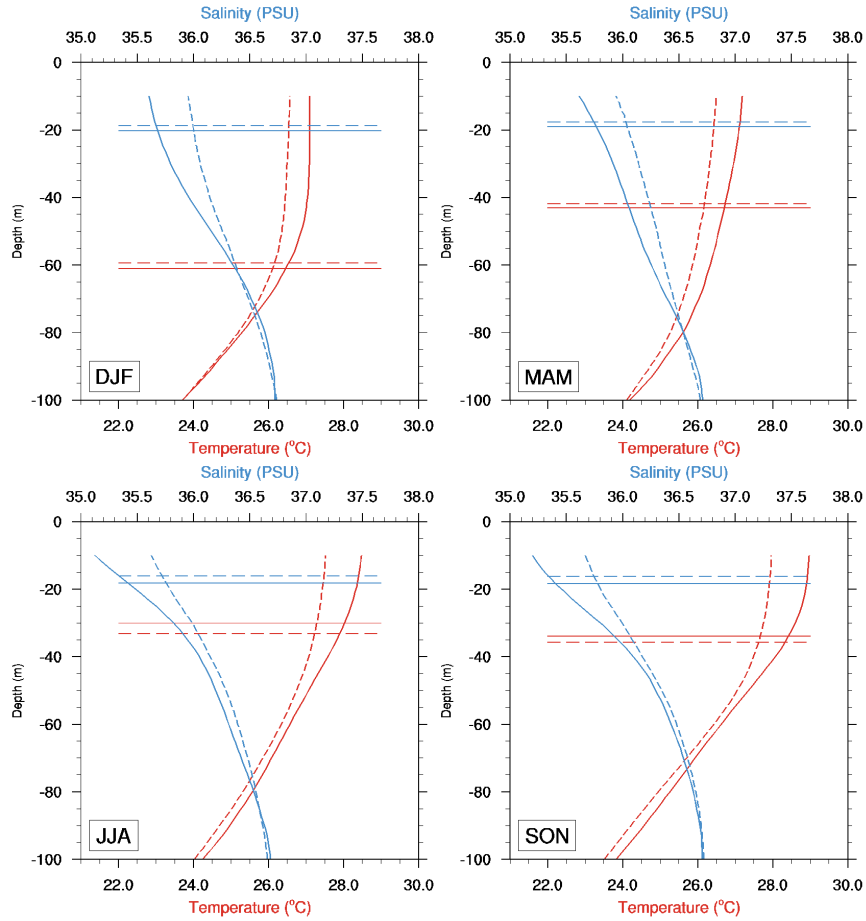


Figure 4.14: Model domain mean seasonal cycle (DJF, MAM, JJA, SON) of vertical temperature ($^{\circ}\text{C}$, red) and salinity profiles (PSU, blue) in the Coloured-River experiment (2001-2010, solid lines) and WOA13 (1955-2012, dashed lines) data. The horizontal lines mark the corresponding top of the thermocline (m, red) and ML depth (m, blue), defined in Section 1.2.1.

ual TCs, depending on the environmental conditions along their path. Shay et al. (2008) have shown rapid intensification of TCs over warm core eddies. Ocean colour and the freshwater plume may change such features regionally and could therefore have an important impact on single TCs.

TCHP neglects changes in density stratification and is therefore not a good metric to assess potential freshwater impacts on TCs. However, ocean colour hardly affects density stratification. Therefore the colour induced differences in TCHP and CI are very similar. Nevertheless, when quantifying the impact of the coloured plume (freshwater and ocean colour), TCHP is misleading. Using TCHP as an indicator for the potential impact of the coloured Amazon and Orinoco on SST feedbacks, would lead to the false conclusion that SST feedbacks will be enhanced and TC intensity hence reduced. As CI measures the ocean resistance to surface cooling, it is able to account for stabilization and subsurface cooling. Note that CI is only a measure for cooling feedbacks, the variable to which TC development is most sensitive (Schade and Emanuel, 1999). But ocean colour warms the ocean surface, which increases the maximum obtainable energy of a storm (Emanuel, 1987). Field (2007) relate strong hurricanes to a passage over the warm Amazon and Orinoco plume. Ocean colour heats the surface and may therefore partly be responsible for the temperature anomalies. The mean temperature increase is small and would only lead to marginal changes in maximum intensity. Locally however, surface temperatures increase by up to $+0.5^{\circ}\text{C}$, potentially increasing TC intensity significantly (Emanuel, 1987). Therefore both aspects should be taken into account when assessing the potential impact of the Amazon and Orinoco on TC intensity: Surface temperature and SST feedbacks.

4.5 Summary and conclusions

In this chapter, the impact of ocean colour on ocean stratification and temperatures has been explored using two different methods. Firstly by applying an absorption algorithm that uses satellite derived surface optical properties, and secondly by approximating light absorption by the low salinity signal of the freshwater plume. Although euphotic zone depths are quite different in the salinity-approximated and satellite-derived light absorption, they produce very similar results regarding ocean temperatures and stratification. However, the salinity-approximation may underestimate light attenuation in regions with additional nutrient sources, like coastal or equatorial upwelling regions. Note that the good performance of the salinity approximation may be a specific characteristic of the model region and not be valid in other parts of the world.

Ocean colour in the Amazon and Orinoco plume slightly enhances surface temperatures by absorbing more radiation near the surface. Since the deeper ocean is isolated from solar light it cools by a considerable amount, in agreement with previous studies (Murtugudde et al., 2002; Gnanadesikan and Anderson, 2009). Local cooling can easily exceed -1°C . The 100m mean temperature is reduced by about -0.3° on average, resulting in a shoaling of the isothermal layer and barrier layer thickening by about 8 m on average. This leads to a substantial improvement of modelled and observed ocean stratification, and emphasizes the need to correctly represent solar transmission in ocean models.

The impact of ocean colour on tropical cyclone (TC) metrics is a substantial decrease of the tropical cyclone heat potential (TCHP) ($-5.6\text{kJ}/\text{cm}^2$) and the cooling inhibition index (CI) ($-2.1\text{ (J/m}^2)^{\frac{1}{3}}$). Therefore SST feedbacks may be increased by ocean colour, potentially leading to weaker TCs. The impact of the coloured plume (freshwater and ocean colour) on SST feedbacks is a combination of colour-induced cooling, and the stabilizing freshwater plume. The two conflicting mechanisms result in a negligible Net effect on CI in the annual mean ($+0.1\text{ (J/m}^2)^{\frac{1}{3}}$), and a small decrease in the hurricane season ($-0.6\text{ (J/m}^2)^{\frac{1}{3}}$). This suggests that the coloured Amazon and Orinoco rivers may have a small negative influence on cooling feedbacks, potentially reducing TC intensity.

The cancellation of freshwater and ocean colour effects on cooling inhibition is based on a multi-annual mean, but the balance varies seasonally and regionally. Therefore certain cyclones could experience enhanced or reduced CI along their path across the WTNA. Note further that the cancellation of river-induced changes is a specific characteristic of the Amazon and Orinoco plume, and may not be valid for other river plumes. Furthermore the cancellation only applies to the river-induced component of the BL. Since the BL is also formed by rain and subduction, the stabilizing effect of the total BL may dominate potential impacts on SST feedbacks. However, ocean colour will always act to reduce cooling inhibition. The effect of light attenuation on the oceanic TC control is not only important in river plumes, but in all areas with high surface chlorophyll concentration that experience TC activity.

Chapter 5

Potential impact of the Amazon and Orinoco plume on tropical cyclone intensity

5.1 Introduction

Several tropical cyclones (TCs) pass over the Amazon and Orinoco plume each year (Ffield, 2007), many of which make landfall in densely populated areas of Mexico and the southeastern United States (Weinkle et al., 2012). Ocean stratification affects TC development, since it modifies self imposed sea surface temperature (SST) feedbacks (Schade and Emanuel, 1999; Vincent et al., 2012b). The Amazon and Orinoco plume modifies ocean stratification and can therefore be crucial for TC development. On the one hand the freshwater plume acts as a barrier to surface cooling and may therefore reduce cooling feedbacks (Balaguru et al., 2012a). Furthermore barrier layers (BLs) have been proposed to increase ocean temperatures, thereby providing more energy to support TC development (Pailler et al., 1999; Ffield, 2007). On the other hand ocean colour in the Amazon and Orinoco slightly enhances surface temperatures and cools the subsurface substantially. The potential impact of reduced solar penetration depth on self-induced cyclone feedbacks has not yet been studied.

TCs mix the water column while passing over the ocean, thereby cooling the ocean surface (Price, 1981). Cold anomalies of up to 6°C have been observed in TC wakes (Black, 1983). SST feedbacks reduce TC intensity by limiting the TC's energy source (Bender et al., 1993; Schade and Emanuel, 1999). The analysis of Schade (2000) suggests that a 2.5°C temperature drop under the TC eye may be sufficient to shut down a cyclone completely. The magnitude of SST feedbacks depends on the ocean stratification and storm strength (Schade and Emanuel, 1999). Furthermore slow TCs cause and experience more cooling and are therefore more affected (Bender et al.,

1993; Mei et al., 2012).

The importance of ocean stratification for SST feedbacks and TC intensity has long been recognized in various case studies (e.g. Cione and Uhlhorn (2003); Lloyd and Vecchi (2011)). Shay et al. (2008) observed rapid intensification of hurricane Opal in 1995 as it passed over a warm core ring in the Gulf of Mexico. Vincent et al. (2014) show that natural variability in ocean stratification leads to variations in TC intensity, where the sensitivity to stratification increases with storm strength. Similarly Emanuel (2006) conclude that the isothermal layer (IL) depth can alter TC intensity substantially, applying a downscaling approach.

BLs received particular attention in regard of their influence on SST feedbacks. Neetu et al. (2012) show that SST cooling is reduced in the presence of the post-monsoonal BL and Wang et al. (2010) conclude that this will lead to less intense typhoons. Ffield (2007) observed that 68% of category 5 TCs pass over the Amazon and Orinoco plume. While they attribute this phenomenon to warm anomalies caused by the river plume, Balaguru et al. (2012a) suggests that the BL in the plume area may be responsible for the TC intensity increase, due to reduced SST feedbacks. They performed a global analysis comparing TC intensification rates within and outside BL regions and find that BLs may enhance intensification rates by 50%.

To this date there has been only one study that connects ocean colour effects to TC intensity. Gnanadesikan et al. (2010) propose that ocean colour changes the track and frequency of TCs in the Pacific due to SST changes. However, they do not discuss the potential effect of the colder subsurface to increase SST feedbacks and consequently reduce TC intensity.

In this chapter river-induced changes in ocean stratification and temperatures are linked to surface cooling and associated TC intensity reduction. The cooling inhibition index (CI) measures the resistance of the ocean to surface cooling and can therefore serve as an indicator for SST feedbacks under storm passage (Vincent et al., 2012b). Idealized relations between surface cooling, intensity reduction, and intensification rates are used to estimate the potential impact of the Amazon and Orinoco plume on TC intensity.

5.2 Relating ocean stratification to SST feedbacks

The oceanic control of TC intensity consists of the initial SST and surface cooling under the TC eye. The maximum potential intensity (MPI) is a thermodynamic estimate for the minimum sustainable TC pressure and is linearly related the SST ($-6 \text{ hPa}/^{\circ}\text{C}$) (Schade and Emanuel, 1999). The actual minimum central pressure (MCP) is usually

larger due to atmospheric and oceanic constraints, e.g. the self-induced surface cooling (Emanuel, 1999). The dimensionless intensity reduction (ΔI) is a measure for a TC's relative deviation from MPI. For steady state TCs it is defined as the reduction in MPI compared to the difference in ambient pressure (p_a) and MPI (Schade, 2000):

$$\Delta I = \frac{MCP - MPI}{p_a - MPI}. \quad (5.1)$$

The range of ΔI is $[0, 1]$, where $\Delta I=1$ implies that the central pressure (MCP) is equal to the ambient pressure, and hence the TC has been shut down. $\Delta I=0$ on the other hand means that the central pressure is equal to the minimum obtainable pressure. A linear relationship of ΔI to cooling under the TC eye was found by Schade (2000) and is shown in Figure 5.1a. Based on this relation, the TC intensity reduction is given by the eye cooling (ΔT_{eye}) normalised by 2.5°C :

$$\Delta I = \frac{\Delta T_{Eye}}{2.5^\circ\text{C}}. \quad (5.2)$$

Although the wake cooling (ΔT_{wake}) is of greater magnitude (Price, 1981), the eye cooling is the relevant parameter for TC development, since this is where ocean-hurricane fluxes are maximum (Emanuel, 2005).

TC intensification rates are also affected by surface cooling. From observations, Mei et al. (2012) find that intensification-rates decrease almost linearly with the magnitude of surface cooling. Similar results were obtained by Vincent et al. (2014) from a modelling approach (Figure 5.1b). TCs experiencing up to -1°C cooling intensify, while any stronger cooling initiates the cyclones' decay. From the surface cooling, the corresponding intensification rate can be inferred via the least squares fit of Vincent et al. (2014), which has a slope of $4.1 \text{ (m/s/day)}/^\circ\text{C}$.

To deduce TC intensity reduction or changes in intensification rates, the knowledge of surface cooling under the TC eye is required. The eye cooling is the SST cooling +12 h after to -12 h before TC passage averaged over a $200 \text{ km} \times 200 \text{ km}$ box around the TC position compared to pre-storm values (Vincent et al., 2012a). It has been found to be roughly half of the wake cooling (ΔT_{wake}) (Vincent et al., 2012a). Following Vincent et al. (2012b), ΔT_{wake} is defined as the SST +1 day to +3 days after TC passage as compared to SSTs -10 days to -3 days before TC passage in the same $200 \text{ km} \times 200 \text{ km}$ box.

The magnitude of surface cooling depends on the strength of the TC, i.e. the amount of mixing energy transferred into the ocean (described by the wind power index (WPI)), and the resistance of the ocean to vertical mixing (described by the cooling inhibition index (CI)) (Section 1.3.4). The increase of ΔT_{wake} with WPI is

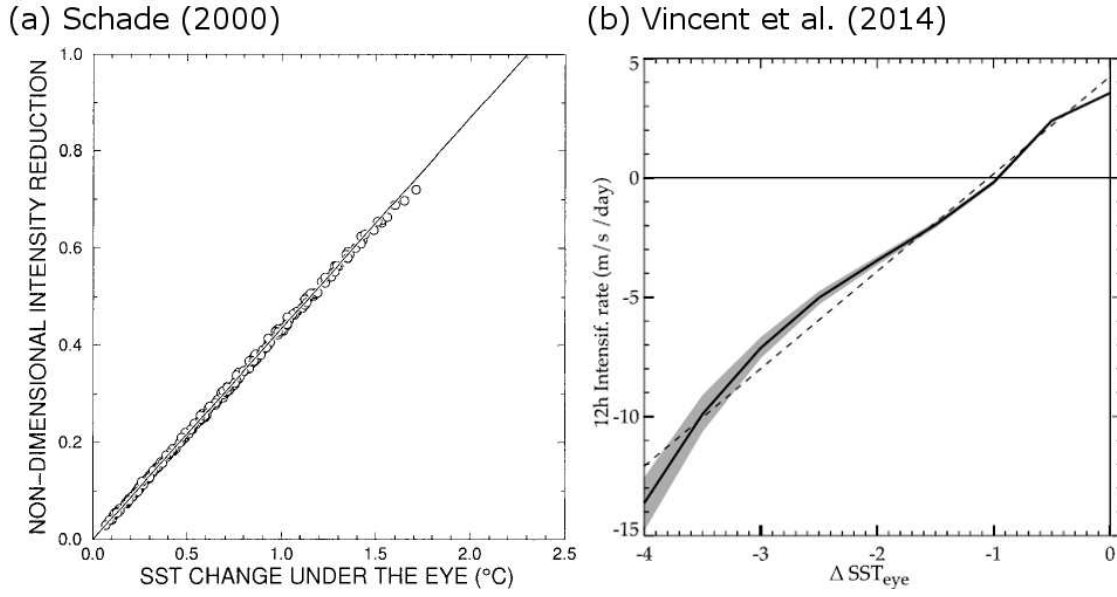


Figure 5.1: (a) Nondimensional intensity reduction as function of the ΔT_{Eye} for an ambient SST of 28°C and a relative humidity of 81% from Schade (2000). (b) Sensitivity of a TCs intensification rate as function of ΔT_{Eye} from Vincent et al. (2014). The thick line is the average intensification rate (between TC passage and 12 h after) and the dashed line is the least squares linear fit with a slope of 4.1 (m/s/day)/°C.

almost linear, except for strong TCs above $WPI=4$ (WPI_4), where the slope is larger (Figure 5.2a). The cooling magnitude increases with decreasing CI, while the range of cooling for varying CI increases with WPI . Vincent et al. (2012b) applied a first order fit to the relationship of modelled cooling to WPI and CI:

$$\Delta T_{Wake} = c_0 + c_1 \times WPI + c_2 \times CI + c_3 \times WPI \times CI, \quad (5.3)$$

with c_0, c_1, c_2 , and c_3 being the fit parameters.

Assuming that ΔT_{Eye} is half of ΔT_{Wake} equations 5.2 and 5.3 can be combined to estimate the TC intensity reduction caused by the Amazon and Orinoco rivers. Thereby the respective CIs obtained in Chapters 3 and 4 for the No-River, Clear-River and Coloured-River experiments will be applied (Table 2.2). From equation 5.1, river-induced changes in MCP will also be inferred.

5.3 Results

5.3.1 Tropical cyclones in the western tropical North Atlantic

To gain an idea about tropical cyclone strength and frequency in the WTNA, TC track, intensity and translation speed are derived from the International Best Track Archive

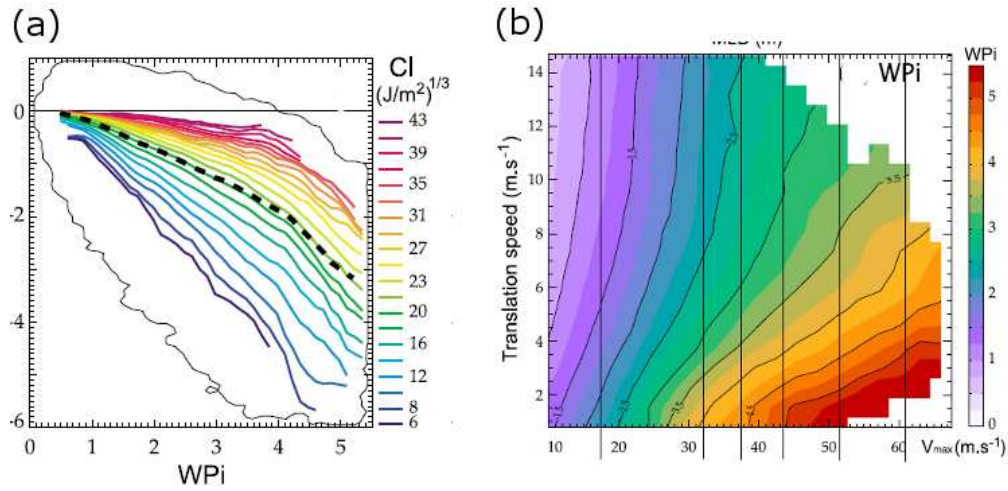


Figure 5.2: (a) Average modelled wake cooling (°C) as a function of WPi for 20 regularly spaced values of CI. (b) Average WPi as a function of the maximum 10-min sustained wind speed (v_{max}) and TC translation speed. Extracted from Vincent et al. (2012b)

for Climate Stewardship (IBTrACS) (Knapp et al., 2010). Between 1960 and 2012 about 300 TCs crossed the model domain (Figure 5.3), of which about 130 reached major hurricane strength at some point in their life cycle (at least category 3 on the Saffir-Simpson scale). About 52 were already of hurricane strength when passing through the model domain. However, the relevant parameter for SST-cooling is the Wind Power index (WPi), rather than the hurricane category. From the TC data, the translation speed and maximum sustained wind speed can be obtained, and the corresponding WPi can be estimated from Vincent et al. (2012b) (Figure 5.2b). According to this estimate, at least 30 hurricanes between 1960 and 2012 were of, of which about 19 reached WPi4, while passing the WTNA. None of the observed TCs were of WPi5. In the following WPi4 will therefore be treated as an upper bound for TC occurrence in the WTNA.

Most TCs don't actually travel over the core plume area, but stay north of 10°N (Figure 5.3a). This is part of the main development region (MDR) (Goldenberg et al., 2001). Therefore only the oceanic values north of 10°N will be considered in the following, which will be referred to as the plume influenced development region (PIDR).

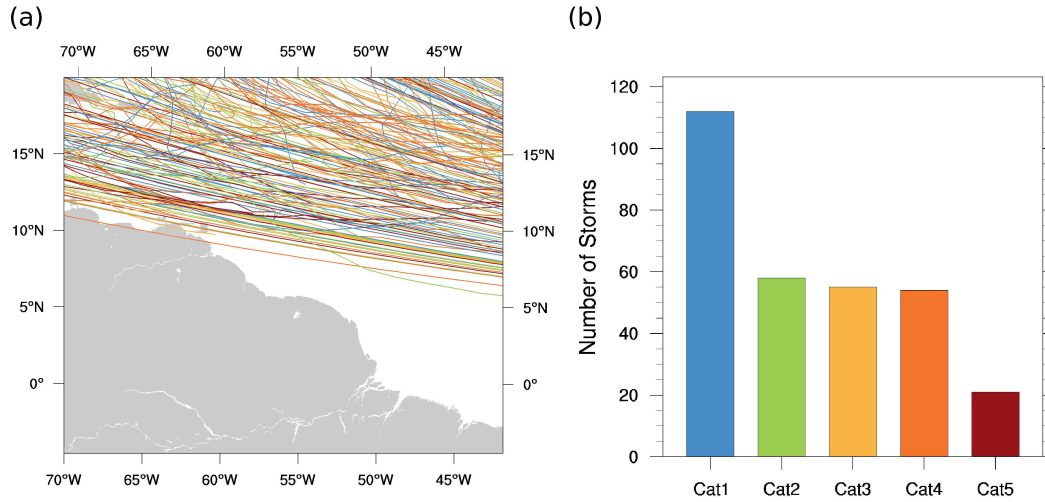


Figure 5.3: (a) Tropical Cyclone tracks in the WTNA between 1960 and 2012 derived from the International Best Track Archive for Climate Stewardship (IBTrACS) (Knapp et al., 2010). Colours indicate the hurricane category on the Saffir-Simpson scale corresponding to (b). (b) Frequency of Category 1, Category 2, Category 3, Category 4 and Category 5 TCs in the WTNA derived from the same dataset.

5.3.2 Riverine impact on TC metrics in the season and area relevant for TC development

Surface temperatures in the PIDR are maximum in September (29.2°C) and minimum in June (27.8°C) in the Clear-River experiment (Figure 5.4), the average being 28.7°C. At this surface temperature Emanuel (1987) finds a maximum potential intensity (MPI) of about 890 hPa, for typical TC conditions of 1015 hPa ambient pressure, and 80% relative humidity. Average SSTs in the PIDR are the same in the No-River and the Clear-River case (Table 5.1). Hence MPI is equal in both experiments. In the Coloured-River experiment mean SSTs increase marginally (+0.1°C). According to Schade and Emanuel (1999) that would lead to about -0.6 hPa stronger TCs. The SST increase is slightly larger (up to +0.2°C) from June to September (Figure 5.4a), corresponding to -1.2 hPa stronger TCs. In October and November, colour induced SST changes in the PIDR are marginal. As seen in Figure 4.11a, the coloured river plume can increase SSTs by up to 0.4°C locally in summer, which would lead to -2.4 hPa lower MPI.

Riverine influences on CI are more distinct than SST changes (Figure 5.4b, Table 5.1). While CI is $32.9 \text{ (J/m}^2\text{)}^{\frac{1}{3}}$ when there are no rivers, it increases by $2.5 \text{ (J/m}^2\text{)}^{\frac{1}{3}}$ (to $35.4 \text{ (J/m}^2\text{)}^{\frac{1}{3}}$) when freshwater is discharged into the WTNA. On the other hand, considering the optical properties of the WTNA decreases CI by -

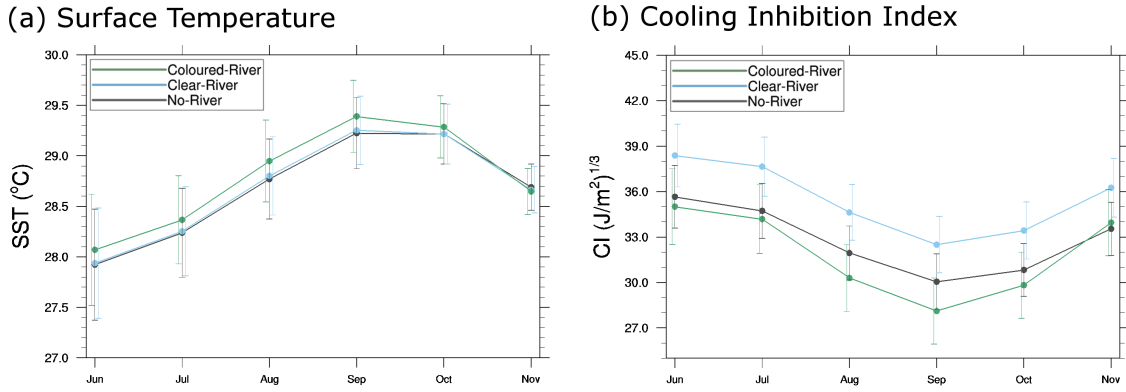


Figure 5.4: Mean seasonal cycle (June to November) averaged over the PIDR (north of $10^\circ N$) of (a) SST ($^\circ C$) and (b) CI ($(J/m^2)^{1/3}$) for the No-River, Clear-River and Coloured-River experiment (2001-2010). The error bars are the standard deviations for the 10-year average.

$3.3 (J/m^2)^{1/3}$ (to $32.1 (J/m^2)^{1/3}$). In all three experiments CI is minimum in September ($28.5 (J/m^2)^{1/3}$ in the Coloured-River experiment) and larger in June and November (about $35.5 (J/m^2)^{1/3}$ in the Coloured-River experiment). The CI increase induced by the clear freshwater plume is of similar magnitude (about $+2.0 (J/m^2)^{1/3}$) throughout the whole hurricane season. The net impact of the coloured plume varies more strongly in time. There is a maximum CI decrease (between the No-River and Coloured-River experiment) of up to $-1.8 (J/m^2)^{1/3}$ in September. The net change turns into an increase of $+0.2 (J/m^2)^{1/3}$ in November. This suggests that the net impact of the coloured plume on TC intensity may have a strong seasonality. Furthermore changes in stratification are most important for TC intensity at low CIs (Vincent et al., 2012b, 2014). The maximum net CI change in September coincides with minimum absolute CI, potentially amplifying the impact of the coloured plume.

5.3.3 Riverine impact on surface cooling

The expected surface cooling in TC wakes can be calculated for particular storm strengths, using equation 5.3. To illustrate the seasonal and spatial characteristics of surface cooling in the PIDR, Figure 5.5 shows monthly maps of the predicted June to November SST response for a WPi4 TC. Without the Amazon and Orinoco WPi4 TCs would cool the surface throughout the whole hurricane season (Figure 5.5a). The cooling is larger in August, September and October, when CI is low. The storm-induced temperature reduction is maximum in the upwelling regions of the southern Caribbean (up to $-4^\circ C$ in August), where the IL is shallow and cold waters are near

	No-River	Clear-River	Coloured-River
SST ($^{\circ}\text{C}$)	28.7 ± 0.3	28.7 ± 0.3	28.8 ± 0.3
CI (J/m^2) ^{$\frac{1}{3}$}	32.9 ± 1.0	35.4 ± 1.0	32.1 ± 1.1

Table 5.1: June-November mean SSTs and CIs and their standard deviation averaged over the PIDR (north of 10°N).

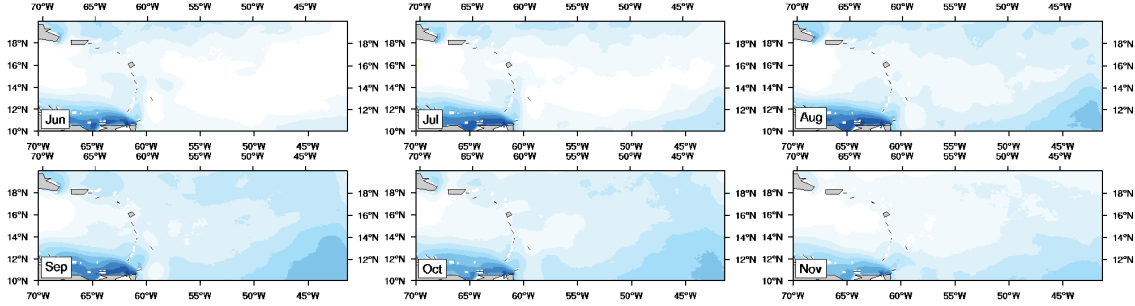
the ocean surface (see Figure 3.8). The SST response is also large from August to October in the southeastern part of the PIDR (up to -2.8°C). In the northern Caribbean, where the IL can be up to 70 m deep, the surface cools only slightly ($< -0.4^{\circ}\text{C}$). The same applies to the area east of the Caribbean in June and July.

When the Amazon and Orinoco enhance the barrier layer in the WTNA, the SST feedback decreases (Figure 5.5b). The magnitude of SST cooling is reduced throughout the whole hurricane season, compared to the saline case. Locally equation 5.3 even predicts surface warming in the course of passing storms. For instance in the western Caribbean, or east of the Caribbean in June and July (up to 1.2°C). Despite the river-induced reduction in SST response, the predicted SST cooling in the Clear-River experiment is still large in the Southern Caribbean (up to -3.2°C) and the southeastern part of the PIDR (up to -2.8°C).

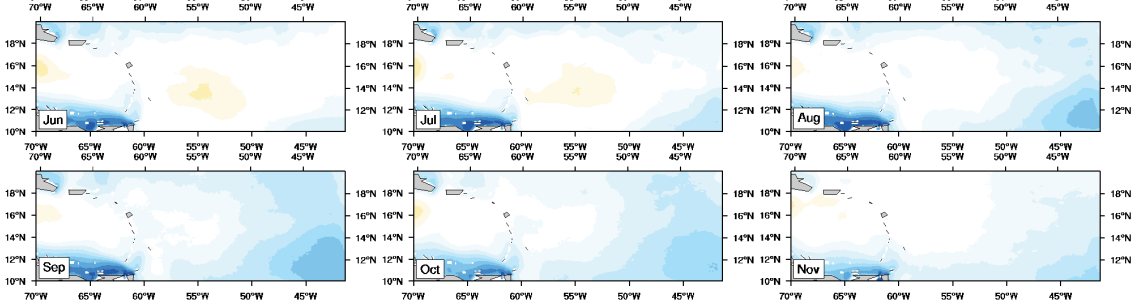
A coloured plume reverses the stabilising freshwater effect due to the colder subsurface. The predicted surface cooling produced by WPi4 storms is similar to the scenario without the Amazon and Orinoco. Positive SST feedbacks are not to be expected from WPi4 TCs. TC induced SST cooling is small in the northern Caribbean and east of the Caribbean ($< -0.4^{\circ}\text{C}$), and large in the Southern Caribbean (up to -4°C) and southeastern PIDR (up to -2.8°C).

The net impact of the coloured plume (Coloured-River - No-River) on WPi4 wake cooling (Figure 5.5d) is very similar to the net CI change (see Figure 4.13). Freshwater and colour effects largely cancel each other, leaving only small patches of modulated SST feedback. The coloured river plume produces hardly any significant changes in June and July. In August, September and October the SST feedback increases (by up to -0.6°C) between about 10°N and 14°N . In November, when TIs start to develop in the northern PIDR, the SST response is likely to decrease by up to $+0.6^{\circ}\text{C}$ north of 15°N . Keep in mind that the wake cooling shown here, is only for one category of TCs (WPi4), and would be of different magnitude for weaker or stronger TCs. However, the spatial pattern and seasonal cycle are likely to be similar.

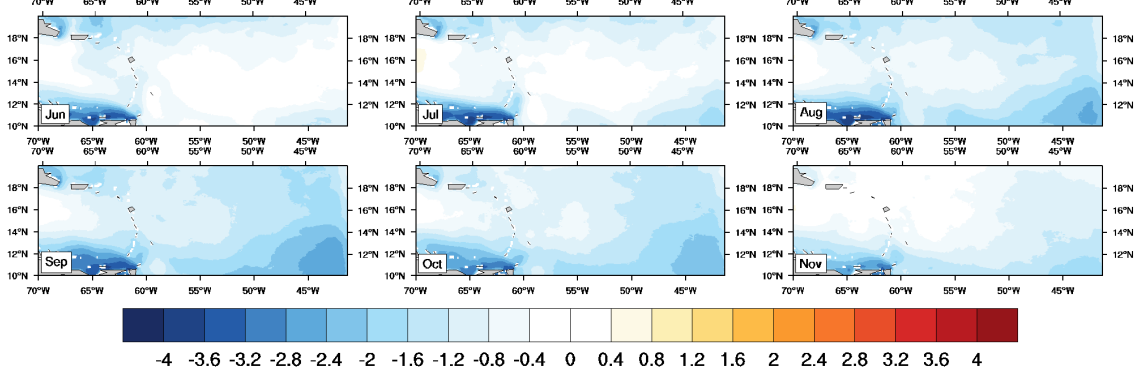
(a) No-River



(b) Clear-River



(c) Coloured River



(d) Coloured-River - No-River

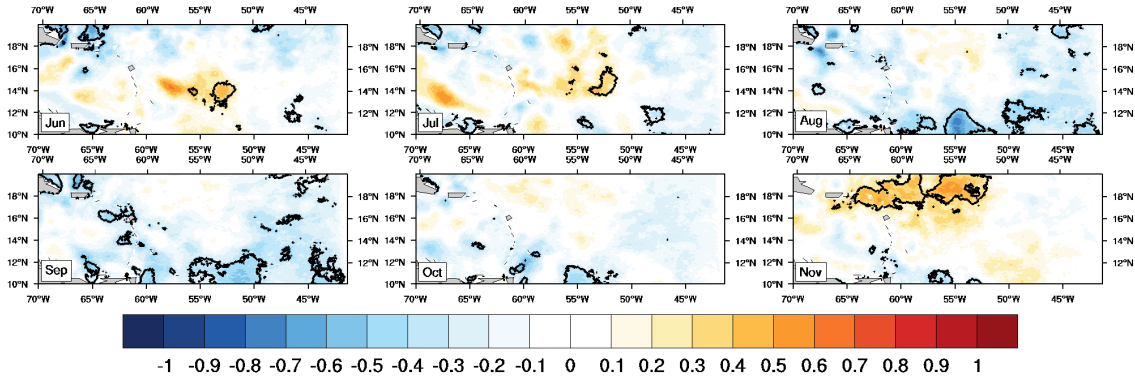


Figure 5.5: Monthly maps of predicted wake cooling (ΔT_{Wake}) for a WPi4 TC in the PIDR derived with equation 5.3 in (a) the No-River experiment, (b) the Clear-River experiment, (c) the Coloured-River experiment, and (d) the net change induced by the coloured plume (Coloured-River - No-River) from June to November for the model years 2001-2010. Black contours in (d) mark the 95% significance level from a two-sided students t-test.

5.3.4 Riverine impact on steady state intensity

5.3.4.1 Eye cooling and intensity reduction

Figure 5.6a shows ΔT_{Eye} as a function of WPi for the mean CI in the No-River, Clear-River and Coloured-River experiments, respectively. The cooling is derived for the June-November mean values in the PIDR using equation 5.3, assuming that ΔT_{Eye} is half of ΔT_{Wake} . Based on the linear relationship between ΔT_{Eye} and ΔI (equation 5.2), the normalised intensity reduction is also given. Note that a larger negative magnitude of ΔT_{Eye} and ΔI implies a stronger response. Therefore referring to an increase of these quantities means more negative values and vice versa in the following.

Surface cooling increases linearly with cyclone strength in each of the three experiments. This is a consequence of the definition of ΔT_{Wake} as a linear function of WPi. Predicted surface cooling and ΔI in the No-River and Coloured-River experiments are very similar. Hence the effects of BL and light attenuation almost cancel each other on average in the season and area relevant for TC development. A TC of WPi4, for instance, would induce SST cooling of -0.47°C , and intensity would be reduced by -0.19 in the No-River case. In the Clear-River experiment the SST feedback is smaller (-0.33°C) and the resulting intensity reduction is -0.13 . The presence of light absorbing material in the ocean however, does cancel and even overwhelm the river-induced BL effect on average, leading to a stronger SST-feedback of -0.52°C (-0.21 intensity reduction) for the WPi4 TC. The differences between the experiments grow with WPi, suggesting that modulations in ocean stratification become more important for stronger TCs. The river impact on TC intensity ranges from a change in ΔI of $+0.01$ (WPi1) to $+0.06$ (WPi4) for a clear river, which is opposed by an intensity decrease from -0.01 (WPi1) to -0.08 (WPi4) when comparing the coloured to a clear plume. This leaves only a small net river impact (Coloured River - No-River) ranging from 0.0 (WPi1) to -0.02 (WPi4) intensity reduction. Therefore the Amazon and Orinoco plume may slightly reduce the intensity of strong storms on average. River-induced changes in SST cooling and intensity reduction are summarized in Table 5.2.

River-induced CI changes vary seasonally (Figure 5.4). Therefore monthly relations between storm strength and ΔI are shown in Figure 5.6b. In June and July CI is high and therefore the intensity reduction is small. The intensity of a WPi4 storm can be reduced by up to -0.16 by the coloured river, and by up to -0.08 for the clear river. ΔI for the No-River and Coloured-River experiments are very similar, with slightly larger ΔI for the coloured plume (up to -0.02 for WPi4). The impact of the Clear-River is of opposite sign and is up to $+0.06$ (WPi4) less intensity reduction.

CI is lower and the predicted ΔI therefore larger in August, September and October. The response is maximum in September, ranging up to about -0.32 (WPI4) for the coloured plume and up to -0.22 (WPI4) for the clear plume. While the difference between the No-River and the Clear-River experiments is similar compared to June and July (up to +0.06 in September for WPI4), the colour effect becomes more important (up to -0.03 in September for WPI4). In November CI grows again and ΔI is lower. It can be up to -0.22 (WPI4) for the coloured plume and -0.10 (WPI4) for the clear plume. As seen in Figure 5.4, the barrier layer overwhelms the colour effect in November. Therefore the net coloured river impact is a slight ΔI reduction (up to +0.02 (WPI4)), compared to a scenario without rivers.

5.3.4.2 The minimum central pressure

It may be hard to grasp what changes in the dimensionless, normalised intensity reduction (ΔI) mean for the actual intensity. Therefore TC intensity changes are expressed in terms of the minimum central pressure (MCP) using equation 5.1 in the following. At a typical atmospheric pressure of 1015 hPa and a relative humidity of 80%, MPI ranges from 850 hPa to 910 hPa for typical SSTs (27°C to 31°C) seen in the PIDR in the hurricane season (Emanuel, 1987). Figure 5.7 shows the difference in MCP ($\Delta MCP = MCP_1 - MCP_2$) between the three experiments as a function of WPI and MPI derived from the respective ΔI s shown in Figure 5.6a.

Weaker SST feedback in the Clear-River experiment leads to a reduction in MCP (i.e. higher intensity) compared to the No-River scenario (Figure 5.7a). The higher WPI, the bigger the intensity gain. The dependence on MPI, however is weak. WPI1 cyclones are about -1 hPa stronger, while WPI4 intensity can increase by up to -9 hPa. The corresponding relative intensity gain ($\Delta MCP_{rel} = \frac{\Delta MCP}{p_a - MCP_{Control}}$) is even less dependent on MPI and ranges from 1% for WPI1 TCs to 7% for WPI4 (Figure 5.7b).

The ocean colour effect opposes the river-induced BL effect, the consequence being an increase in MCP (Figure 5.7c,d). Note that the SST increase in the Coloured-River experiment is taken into account by assuming a -0.6 hPa lower MPI. The colour induced MCP increase overpowers the decrease seen previously and ranges from +2 hPa (-2%) for weak storms to 12 hPa (-8%) for strong storms. As a consequence the net effect of the coloured plume is a small intensity increase of -0.2 hPa (+0.2%) for weak TCs, but can mean a reduction of up to +2.5 hPa (-2%) for strong TCs (Figure 5.7e,f). Relative and absolute river-induced changes in MCP for a typical TC of 890 hPa MPI at 1015 hPa ambient pressure (Emanuel, 1987) are summarized in Table 5.2 for various WPis.

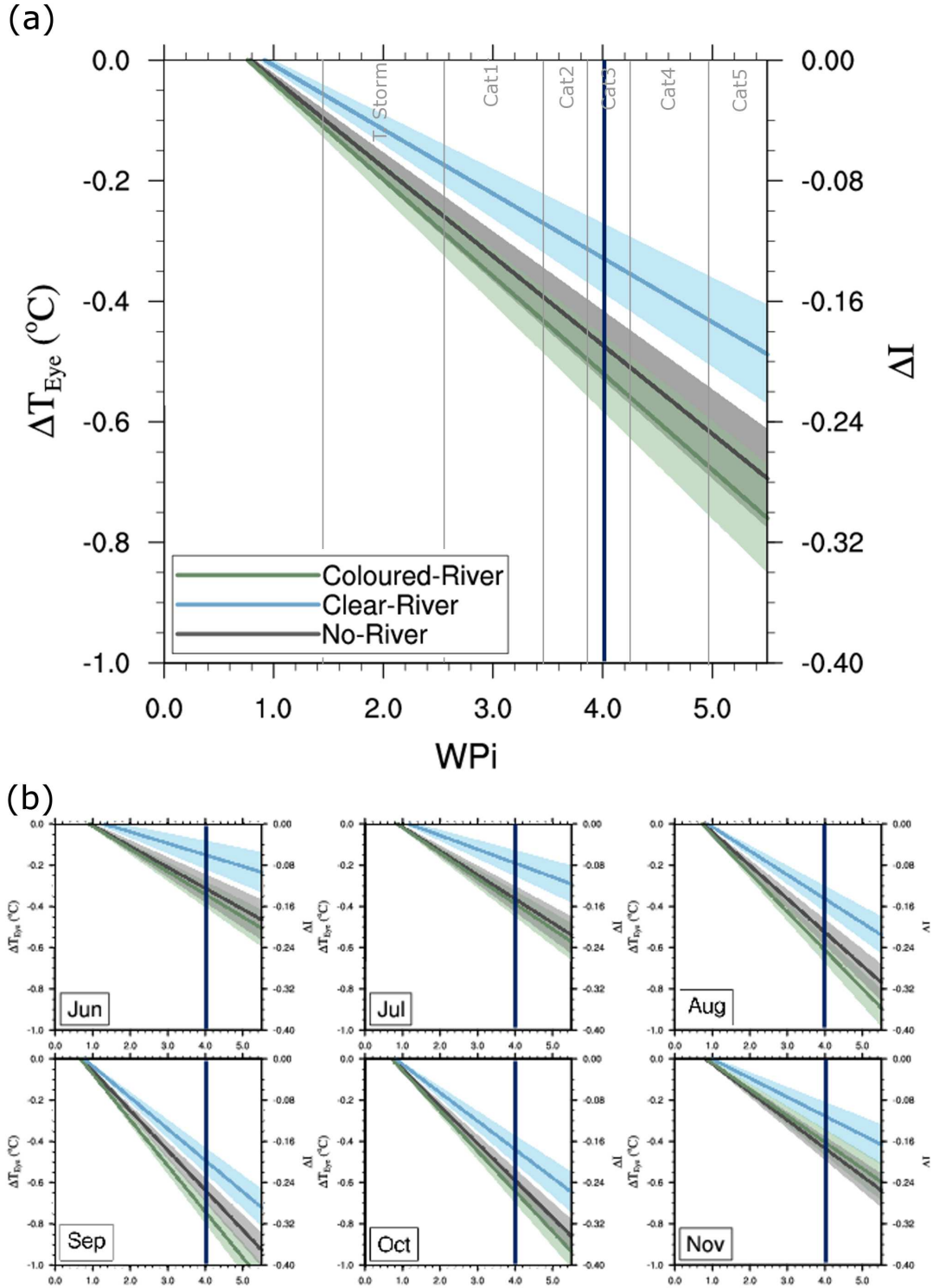


Figure 5.6: (a) June to November mean Eye-cooling (ΔT_{Eye} , $^{\circ}\text{C}$) and associated dimensionless intensity reduction (ΔI) as a function of the dimensionless WPI for the mean CI averaged over the PIDR (north of 10°N) in the No-River, Clear-River and Coloured-River experiments (2001-2010). The shaded areas mark the respective standard deviations. The black vertical line marks WPI4 TCs. Corresponding TC categories on the Saffir-Simpson scale for a TC moving at 5 m/s are given for reference and are derived from Vincent et al. (2012b) (their Figure 4b) (b) The same as (a), but derived for the monthly mean CI values from June to November.

River Induced TC Intensity Changes

	WPI	Cl-Riv - No-Riv	Col-Riv - Cl-Riv	Col-Riv - No-Riv
Surface Cooling $\Delta (\Delta T_{Eye}) (^{\circ}\text{C})$	1	+0.02	−0.03	−0.01
	2	+0.06	−0.08	−0.02
	3	+0.10	−0.14	−0.03
	4	+0.14	−0.19	−0.05
	5	+0.19	−0.24	−0.06
Intensity reduction $\Delta (\Delta I)$	1	+0.01	−0.01	−0.00
	2	+0.03	−0.03	−0.01
	3	+0.04	−0.05	−0.01
	4	+0.06	−0.08	−0.02
	5	+0.07	−0.10	−0.02
Minimum Central Pressure $\Delta \text{MCP (hPa)}$	1	−1.1	+0.9	−0.2
	2	−3.1	+3.6	+0.5
	3	−5.2	+6.5	+1.1
	4	−7.2	+9.1	+1.8
	5	−9.3	+11.8	+2.5
Minimum Central $\Delta \text{MCP}_{rel} (\%)$	1	+1	−1	+0
	2	+3	−3	−0
	3	+5	−6	−1
	4	+7	−9	−2
	5	+10	−12	−3
Intensification Rate $\Delta I_{Rate} (\text{m/s/day})$	1	+0.09	−0.12	−0.03
	2	+0.26	−0.34	−0.08
	3	+0.43	−0.56	−0.14
	4	+0.59	−0.78	−0.19
	5	+0.76	−1.00	−0.24
Intensification Rate $\Delta I_{Rate,rel} (\%)$	1	+2	−3	−1
	2	+8	−9	−3
	3	+15	−18	−5
	4	+28	−29	−9
	5	+49	−43	−16

Table 5.2: Mean June to November river-induced changes in the PIDR (north of 10°N) in eye cooling ($\Delta(\Delta T_{Eye})$ in °C), dimensionless intensity reduction ($\Delta(\Delta I)$), minimum central pressure (ΔMCP in hPa), relative change in minimum central pressure (ΔMCP_{rel} in hPa) intensification rate (ΔI_{Rate} in m/s/day), and relative change in intensification rate ($\Delta I_{Rate,rel}$ in %) for WPI1, WPI2, WPI3, WPI4 and WPI5 TCs. MCP is derived for an ambient pressure of 1015 hPa and an MPI of 890 hPa. The values for WPI5 TCs are shown in grey since they are unlikely to occur in the PIDR.

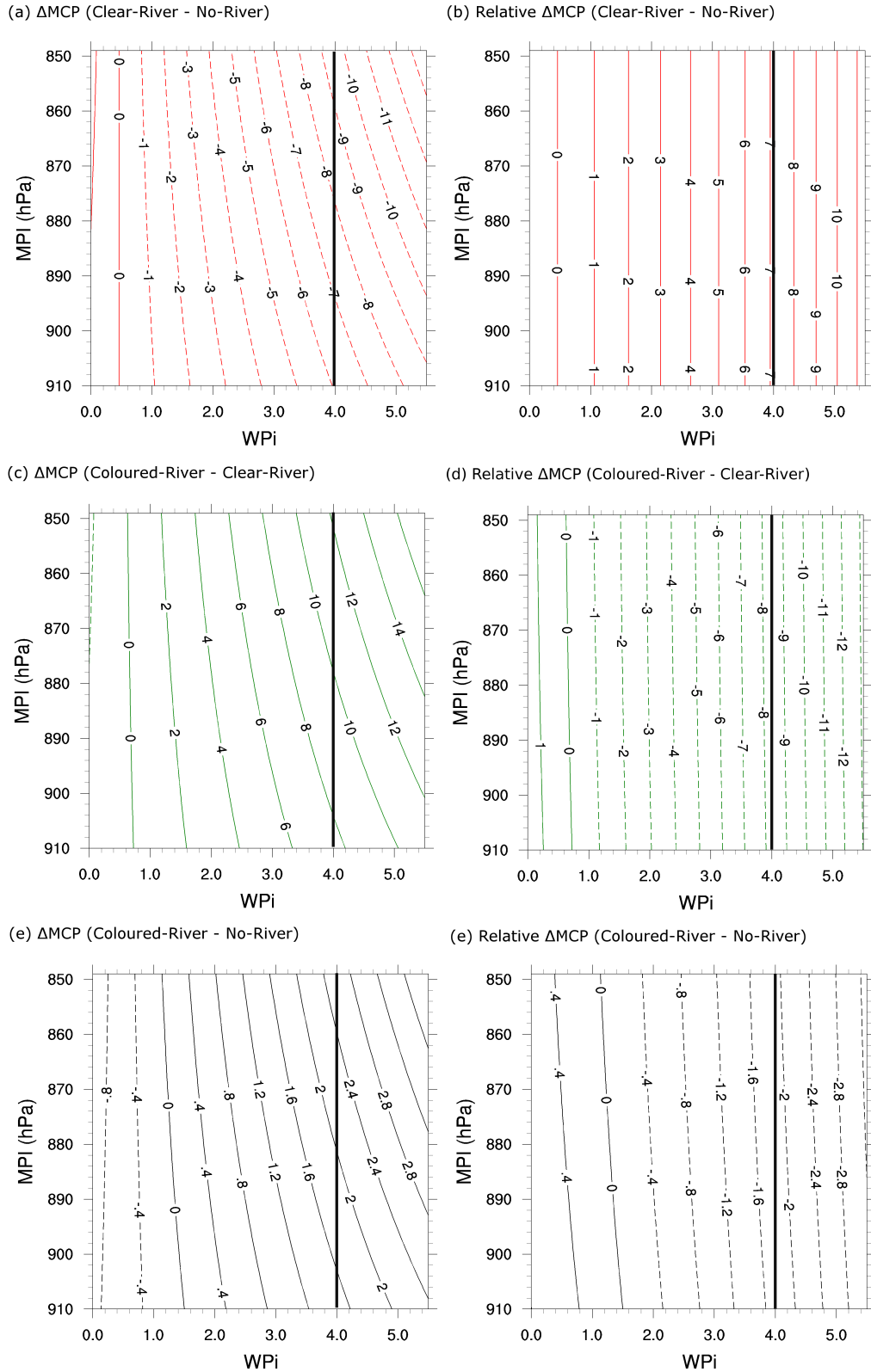


Figure 5.7: Mean June to November (2001-2010) changes averaged over the PIDR (north of 10°N) in MCP (ΔMCP in hPa) and corresponding relative change (ΔMCP_{rel} in %) between (a),(b) the No-River and Clear-River, (c),(d) the Clear-River and Coloured-River and (e),(f) the No-River and Coloured-River (net) experiments as a function of WPI and $\text{MPI}_{Control}$ at an ambient pressure of 1015 hPa. The vertical black lines mark WPI4 TCs.

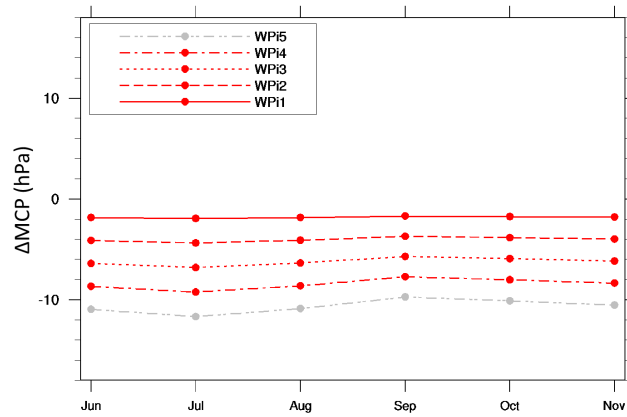
The seasonal cycle of river-induced changes in MCP is shown in Figure 5.8 under typical conditions (890 hPa MPI and 1015 hPa ambient pressure) for WPI1, WPI2, WPI3, WPI4, and WPI5 TCs. As the impact of ocean stratification on TCs depends only weakly on MPI, this is also representative for other environmental conditions. The intensifying effect of the Clear-River is fairly constant throughout the hurricane season with intensification of about -2 hPa for weak TCs (Figure 5.8a). WPI4 TCs can become up to -9 hPa more intense in the Clear-River scenario. The intensity increase is about -2 hPa stronger in July compared to September. The impact of ocean colour shows a stronger seasonality (Figure 5.8b), with a maximum in August and September (about +1 hPa for WPI1 TCs to +12 hPa for WPI4), and a minimum in November (+0.2 hPa for WPI1 to +6 hPa for WPI4). The seasonal cycle of the net impact of the coloured river (Figure 5.8c) is a combination of the nearly constant clear river effect and the stronger seasonality of ocean colour. Potential TC modulations are very weak in June and November (up to +0.6 hPa for WPI4). In September the potential intensity loss is maximum and is up to +5 hPa for WPI4. As mentioned above, the freshwater effect dominates ocean colour in November, leading to an intensity increase of up to -2.0 hPa (WPI4). Note that the net effect on weak TCs (WPI1) is a slight intensity increase of about -0.2 hPa throughout the whole hurricane season.

5.3.4.3 Sensitivity to atmospheric and oceanic parameters

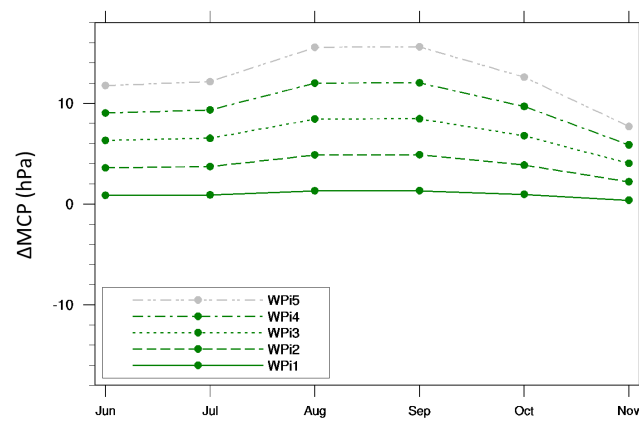
The above results have been obtained assuming an ambient pressure of 1015 hPa. From the CFSR data it can be seen, that the mean sea level pressure over the WTNA varies between about 1012 hPa and 1018 hPa (not shown). Applying a different ambient pressure slightly shifts the results in the sense, that a lower ambient pressure (1012 hPa) leads to an about -0.4 hPa larger impact on WPI4 TCs in the Clear-River and Coloured-River experiments, while a higher ambient pressure (1018 hPa) has the opposite effect (not shown).

The dimensionless intensity reduction (ΔI) is here normalized to 2.5°C. In reality this value will vary with environmental factors, such as air humidity and ambient SSTs (Schade, 2000). The analysis of Schade (2000) shows that the temperature threshold roughly varies between 1.5°C and 3.5°C for varying ambient relative humidity and SSTs. Applying different thresholds does not change the overall answer regarding near-cancellation of freshwater and colour effects, but the magnitude of freshwater and ocean colour components changes (Table 5.3). A lower threshold (1.5°C) leads to an even larger impact of river-induced changes (up to -12.2 hPa intensity increase induced by the clear plume). The impact of the clear and coloured river plumes is

(a) Clear-River - No-River



(b) Coloured-River - Clear-River



(c) Net (Coloured-River - No-River)

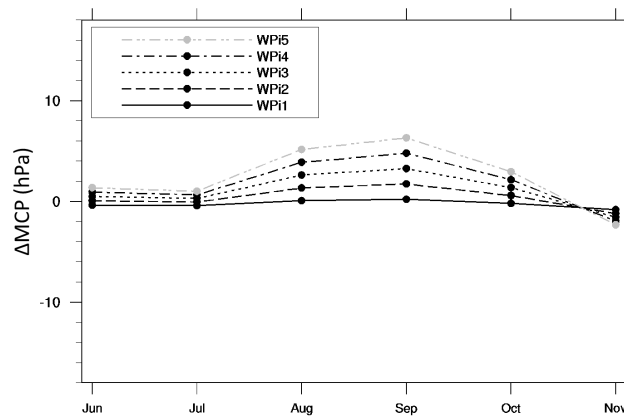


Figure 5.8: Monthly mean (2001-2010) river-induced MCP changes (ΔMCP in hPa) in the PIDR (north of 10°N) between the (a) No-River and River, (b) the Clear-River and Coloured-River and (c) the No-River and Coloured-River (net effect of the coloured plume) experiments for WPI1, WPI2, WPI3, WPI4 and WPI5 TCs from June to November for a typical TC of 890 hPa MPI at 1015 hPa ambient pressure. WPI5 TCs are shown in grey since they are unlikely to occur in the PIDR.

	1.5°C		2.5°C		3.5°C	
	WPI1	WPI4	WPI1	WPI4	WPI1	WPI4
Clear-River - No-River	-1.8 hPa	-12.1 hPa	-1.1 hPa	-7.2 hPa	-0.8 hPa	-5.1 hPa
Coloured-River - Clear-River	+1.9 hPa	+15.5 hPa	+0.9 hPa	+9.1 hPa	+0.5 hPa	+8.3 hPa
Coloured River - No-River	+0.0 hPa	+3.5 hPa	-0.2 hPa	+1.8 hPa	-0.3 hPa	+1.6 hPa

Table 5.3: Range of June to November (2001-2010) river-induced changes in MCP (Δ MCP in hPa) averaged over the PIDR (north of 10°N) for WPI1 to WPI4 TCs for Clear-River - No-River, Coloured-River - Clear-River, and net Coloured-River - No-River experiments, applying different thresholds for TC decline (1.5°C, 2.5°C, 3.5°C). The range is given for an ambient pressure of 1015 hPa and MPI of 890 hPa.

reduced, when assuming that TCs are shut down at a surface cooling of 3.5°C (up to -5.1 hPa intensity increase induced by the clear plume). Note that the impact of ocean stratification increases exponentially with threshold decrease.

5.3.5 Riverine impact on intensification rates

The above calculations were for steady state TCs. However, TCs passing over the plume are within the tropical North Atlantic main development region, frequently affected by developing storms (Goldenberg et al., 2001). Figure 5.9 shows the mean June to November intensification rates as a function of WPI for the different experiments. They are derived for the average CI given in Table 5.1, using equation 5.3 and applying the least square linear fit of Vincent et al. (2014) shown in Figure 5.1. For WPI4 TCs the resulting intensification rates are about +2.2 m/s/day (No-River), +2.8 m/s/day (Clear-River) and +2.0 m/s/day (Coloured-River). Intensification rates decrease with wind power. For the coloured plume they range from about 3.9 m/s/day (WPI1) to 2.0 m/s/day (WPI4).

The impact of ocean stratification on intensification rates ($\Delta I_{Rate} = I_{Rate1} - I_{Rate2}$) increases with wind power from WPI1 to WPI4 TCs: +0.09 to +0.6 m/s/day (2% to 28%) for the river-induced BL effect, -0.12 to -0.8 m/s/day (-3% to -29%) for the colour effect and from -0.03 to -0.2 m/s/day (-1% to -9%) for the net effect. Light attenuation effects on TC intensification rates in the river plume may therefore be of similar size, but opposite sign to river freshwater effects, leading to a small net impact of the coloured plume for weak TCs, but possibly large impacts for stronger storms.

Relative and absolute river-induced changes on intensification rates for a typical TC of 890 hPa MPI at 1015 hPa ambient pressure (Emanuel, 1987) are summarized in Table 5.2 for various WPis.

The seasonal cycle of absolute intensification rates in the Coloured-River experiment is shown in Figure 5.10. Since the average SST feedback is fairly weak in the PIDR, mean intensification rates are positive throughout the whole hurricane season. They are up to 4 m/s/day for weak TCs and vary only weakly with the season. Stronger TCs are more sensitive to the seasonal cycle of CI and induce more SST cooling. Therefore they have weaker intensification rates. The lowest intensification experienced by WPi4 TCs occurs from August to October (as low as 1.0 m/s/day in September). Regionally, the strong surface cooling in the southern Caribbean (up to -2.0°C eye cooling for WPi4) may lead to negative intensification rates (or rather de-intensification) of -4 m/s/day.

The seasonal variation of absolute and relative changes in intensification rates (ΔI_{Rate} , $\Delta I_{\text{Rate,rel}}$) induced by the clear and the coloured Amazon and Orinoco are shown in Figure 5.11. The shape of the seasonal cycle is very similar to riverine influences on steady state TC intensity seen previously. Increased stability in the Clear-River experiment enhances intensification rates by about 0.1 m/s/day for WPi1 TCs and by up to 0.7 m/s/day for WPi4 TCs (Figure 5.11a). The freshwater induced change in intensification rate is only weakly dependent on the month, but is slightly larger in July compared to September. Since absolute intensification rates are lowest in September, the relative impact of the river-induced BL peaks at that time of the year (110%). The colour-induced changes show a more pronounced seasonality (Figure 5.11 b). Intensification rates decrease by up to -1.0 m/s/day (WPi4) in August and September, compared to a decrease of only -0.5 m/s/day in November. Again the relative impact of ocean colour peaks in September and is a decrease of up to -50% for WPi4 TCs. In June and July the decrease is only about -20%. The net effect of the coloured plume is small in June and July (up to -0.2 m/s/day (-4%) for WPi4) and maximum in September (-0.4 m/s/day (-30%) for WPi4). In November, the intensification rate slightly increases (+0.1 m/s/day, 8% for WPi4 TCs). Note also that river induced changes in intensification rates increase linearly with wind power, while the relative impact increases exponentially with WPi.

5.4 Discussion

The impact of the coloured Amazon and Orinoco plume on TC intensity is a combination of the freshwater-induced intensity increase and the colour-induced weakening.

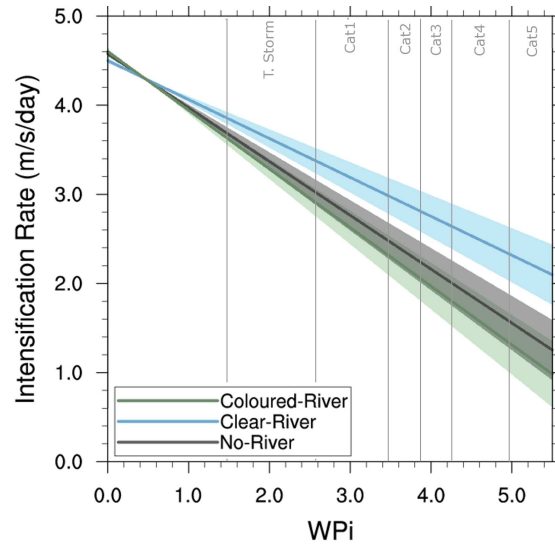


Figure 5.9: Mean June to November (2001-2010) intensification rate (m/s/day) as a function of WPI for the mean CI averaged over the PIDR (north of 10°N) in the No-River, Clear-River and Coloured-River experiments. Intensification Rates are derived using the relation of Vincent et al. (2014). The shaded areas mark the respective standard deviations. Corresponding TC categories on the Saffir-Simpson scale for a TC moving at 5 m/s are given for reference and are derived from Vincent et al. (2012b) (their Figure 4b)

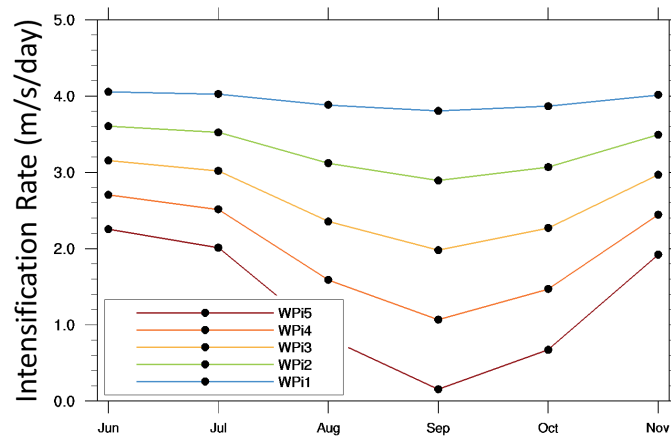
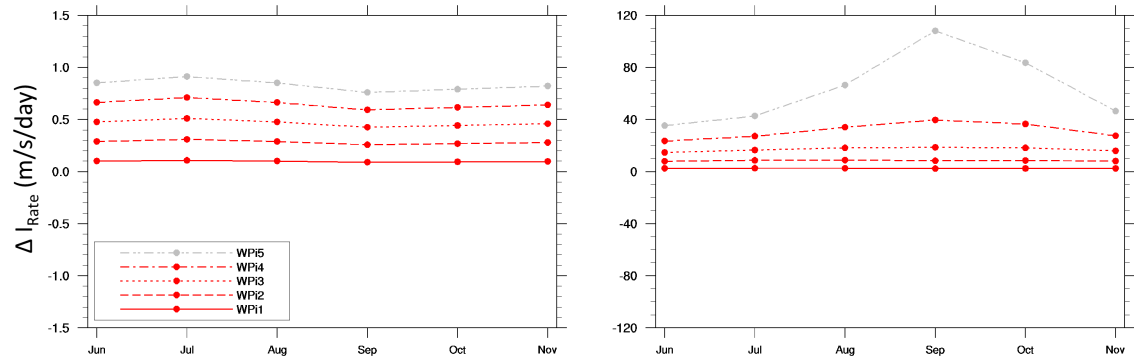
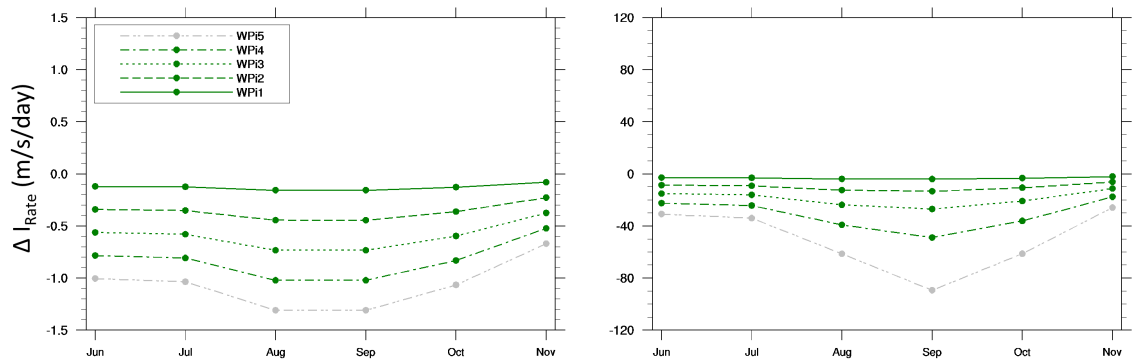


Figure 5.10: Monthly mean (2001-2010) intensification rate (m/s/day) in the hurricane season in the PIDR (north of 10°N) in the presence of the coloured plume for WPI1, WPI2, WPI3, WPI4, and WPI5 TCs, derived from the average relation of intensification rates and ΔT_{Eye} obtained by Vincent et al. (2014).

(a) Clear-River - No-River



(b) Coloured-River - Clear-River



(c) Coloured-River - No-River

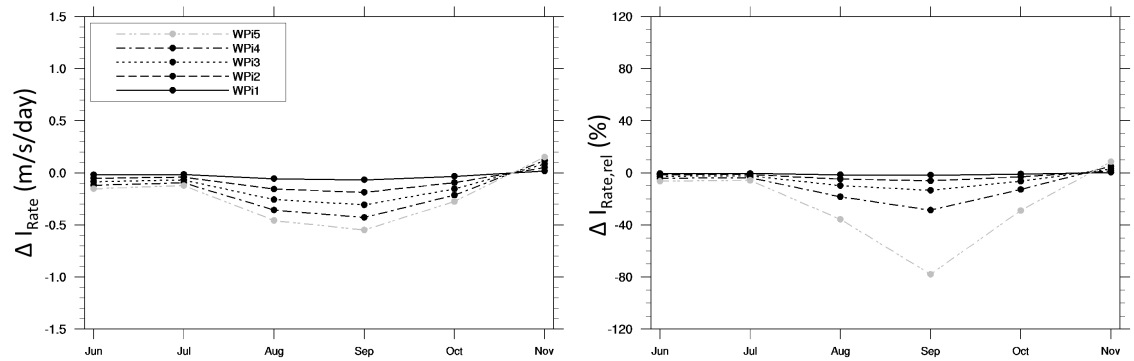


Figure 5.11: Monthly mean river-induced absolute (left) and relative (right) changes in intensification rate in the PIDR between the (a) No-River and Clear-River, (b) the Clear-River and Coloured-River and (c) the No-River and Coloured-River (net effect of the coloured plume) experiments for WPi1, WPi2, WPi3, WPi4 and WPi5 TCs from June to November. WPi5 TCs are shown in grey since they are unlikely to occur in the PIDR.

The two effects are of similar magnitude, therefore leading to a near-cancellation of changes in TC intensity caused by the rivers. The sign and magnitude of the balance depends on the storm strength. Weak storms are slightly more intense in the presence of a coloured plume. On the one hand this is due to enhanced surface temperatures in the Coloured-River experiment. Note also that weaker storms induce shallower mixing and may not reach the cooled subsurface, which is maximum at about 50 m. Enhanced stability and BL warming may therefore be dominant, leading to reduced SST feedbacks. For more energetic storms, the heat reduction caused by ocean colour outweighs the enhanced stability effect. On average strong TCs are therefore likely to be slightly less intense, when the coloured Amazon and Orinoco discharge into the WTNA.

While the intensifying freshwater effect is fairly constant throughout the year, the impact of ocean colour varies with the season. This leads to a strong seasonal dependence of the net effect of the coloured river plume. Ocean stratification is more important for SST feedbacks at low CIs (Vincent et al., 2012b, 2014). The maximum net CI reduction in September occurs at the same time as the annual CI minimum, potentially amplifying river effects. In November on the contrary, the freshwater effect dominates ocean colour.

The near-cancellation is the product of a regional average. River-induced changes in TC intensity vary with location, and depend on the local balance of freshwater and colour effects. According to Vincent et al. (2012b), the impact of stratification on surface cooling is relatively small for CIs exceeding $30 \text{ (J/m}^2\text{)}^{\frac{1}{3}}$, and more important for low CIs. The mean CI in the hurricane season can for example be up to $45 \text{ (J/m}^2\text{)}^{\frac{1}{3}}$ in the western Caribbean, suggesting a small effect of the freshwater plume. CI is lower near the upwelling regions in the southern Caribbean and in the eastern WTNA, potentially resulting in a larger impact of river-induced changes. The riverine impact on individual TCs therefore depends on the season and exact TC path and the near-cancellation only represents an average property.

It may be tempting to conclude that the surface warming predicted by equation 5.3 for WPi4 TCs (Figure 5.5b) is a consequence of temperature inversions induced by the freshwater plume (see chapter 3). Note however, that in this framework it is only due to the characteristic of the linear fit applied here. Equation 5.3 predicts surface warming for any CIs above about $42 \text{ (J/m}^2\text{)}^{\frac{1}{3}}$ and does not account for the actual subsurface temperature stratification. Although it is not a coincidence that the predicted surface warming is co-located with TIs (since they are partly responsible for the large CI value), high CIs may mainly result from deep ILs or very stable stratification in other regions or seasons. However, Balaguru et al. (2012a) observed a positive SST

response for hurricane Omar, as it passed over an inversion layer in the Amazon and Orinoco plume in October 2008. In the real world the potential for surface warming does not only depend on the TI magnitude, but also on their depth and the storm strength. In summer TIs in the WTNA occur at shallower depth. Therefore less energy is needed to mix them to the surface. As soon as a storm becomes strong enough to lift colder water from below the inversion layer into the mixed layer (ML), surface warming will be reduced. In the presence of TIs, the resulting feedback will therefore be a complex combination of the inversion depth and magnitude, ocean stability, stratification below the inversion layer, and the storm strength. However, TIs will always act to reduce SST feedbacks.

Strong TCs entrain more water into the ML and therefore cause more surface cooling than weak storms. Following Vincent et al. (2012b) and Schade (2000) the SST feedback and associated intensity reduction increases linearly with wind power. Similarly, the impact of ocean stratification on cyclones increases with TC strength. Since stronger storms mix more vigorously, they are more exposed to river-induced changes. This is in agreement with Vincent et al. (2014), who studied the impact of natural variability in ocean stratification on TCs. Therefore the individual components of river-induced TC forcing (freshwater and ocean colour) become more important for more energetic storms. Note also, that slow storms induce more mixing at a given location and experience more of their own surface cooling, since they leave the site slower than quick storms (Mei et al., 2012). Therefore they become less intense. This effect is included in WPI, which is a measure of cyclone winds, translation speed and size. The changes in TC intensity are only weakly dependent on the initial atmospheric and oceanic conditions, represented in MPI.

Jourdain et al. (2012) find that cyclonic rainfall can also reduce SST-feedbacks since it increases ocean stability. However, they conclude that this effect is weak in the Bay of Bengal, due to the local BL. Similarly cyclonic rain effects may be negligible in the Amazon and Orinoco plume, where freshwater enhances ocean stability.

The linear relationship of surface cooling to wind power and CI applied here, is based on an average result and can be different for various TCs. Similarly the linear relations of surface cooling to TC intensity reduction and TC intensification rates are average properties, and may not apply exactly to individual TCs. Furthermore the eye cooling in Vincent et al. (2012b) and Schade (2000) may be defined differently. The actual response depends on the exact pass of TCs and the oceanic and atmospheric conditions along their track. The results presented here should be seen as an estimate of the magnitude and tendency of TC responses to river-induced changes in ocean-stratification. The CI used in this study is the potential energy increase in the ocean,

corresponding to a 2°C surface cooling. This value has been used by Vincent et al. (2012b), as a typical cooling under TCs. They show that CI under the assumption of 1°C or 3°C cooling is highly correlated to the 2°C CI. The choice should therefore not change the answers regarding river-induced changes in SST feedback.

Vincent et al. (2012b) find that the average WPI in the plume area is low (< 1), which suggests a minor role of ocean stratification on TC intensity. However, TC track data shows that, despite many weak (or still developing) storms, the plume area is frequently affected by major hurricanes, which are more sensitive to ocean stratification. It has been found here that no WPI5 TCs have passed the Amazon and Orinoco plume in the past 50 years, and WPI4 has been treated as an upper bound. Note that WPI has been inferred from the relationship of TC wind and translation speeds shown in Vincent et al. (2012b), and binned to groups of WPI1 to WPI5. This is a very qualitative estimate, and TCs between WPI4 and WPI5 may have occurred. WPI4 does not represent a natural threshold for TCs over the plume. As the plume area is frequently affected by developing storms which mainly travel east to west, WPI may be higher in the western part of the PIDR. Therefore changes in ocean stratification may be more important for TC intensity in this area.

Previous studies found TC-intensification in the presence of BLs and the Amazon and Orinoco plume. However, this is not necessarily a contradiction to the findings presented here. Ffield (2007) observed that 68% of all WTNA category 5 TCs pass over the plume area and thus suggest circumstantial evidence of a relationship, assuming that the plume is responsible for warm SST anomalies. Vizzy and Cook (2010) demonstrate that the warm anomalies enhance the number of TCs by 60%. However, neither study provides direct evidence that the river is causing the warm anomalies. In this and other explicit river studies (e.g. Masson and Delecluse (2001); Balaguru et al. (2012b)) the Amazon and Orinoco plume does not increase SSTs substantially. There may therefore be a connection between the more intense TCs passing over the warm pool region, but the role of the coloured river plume may only be marginal. Balaguru et al. (2012a) compared global intensification rates in barrier layer and non-barrier layer regions and conclude that the overall effect of BLs is to increase TC intensification rates by 50%. Neetu et al. (2012) observe and model reduced average SST cooling in the presence of barrier layers in the Bay of Bengal. Grodsky et al. (2012) also showed reduced cooling in the Amazon and Orinoco plume in the wake of Hurricane Katia. None of these three, or other studies distinguish between turbid and non-turbid regions. Unlike presented here, possible effects of ocean/river colour and barrier layers have not been separated. Another reason the presented findings may appear different to BL intensification arguments is that only the river-induced part of

the BL is considered. The BL in the WTNA is however formed by several mechanisms: the Amazon and Orinoco plume, precipitation (Foltz et al., 2004) and subduction of high salinity waters (Blanke et al., 2002; Balaguru et al., 2012b). Salinity stratification in the WTNA increases CI by about $8.1 \text{ (J/m}^2\text{)}^{\frac{1}{3}}$, of which only about a quarter are caused by the Amazon and Orinoco plume (Section 3.3.2). The near-cancellation of freshwater and colour effects on TC-intensity only refers to river-induced changes, but in any case the ocean colour will reduce TC intensification.

Gnanadesikan et al. (2010) report that the presence of light absorbing material causes Pacific TCs to develop in higher latitudes. Their focus is on the frequency of TCs, but they also suggest that the intensity of typhoons may be increased. This appears to contradict the findings presented here. However, their Genesis Potential Intensity index does not consider changes in subsurface temperatures and IL depths. Applying a downscaling approach Emanuel (2006) has shown, that a 10% increase in IL depth results in 4% more intense TCs. The IL depth in this study even decreases by 30%. Gnanadesikan et al. (2010) consider feedbacks of SST changes to the atmosphere like a strengthening of the Hadley circulation, while this study focuses on the oceanic feedback part. It may thus be possible to reconcile both studies, recognizing that the impact of light absorption is likely to be a complex combination of changing track, frequency and intensity.

5.5 Summary and conclusion

The impact of the Amazon and Orinoco plume on tropical cyclone intensity consists of two components. Firstly the intensifying freshwater effect, which reduces SST feedback due to higher stability and enhanced barrier layer temperatures. Secondly the ocean colour effect increases SST feedbacks, due to cooling of the subsurface by isolating the deeper ocean from sunlight. Interestingly, the two effects almost cancel each other on average. Applying idealized relationships between ocean stratification, SST feedback and storm strength, the net impact of the coloured plume is a slight intensity gain (about -0.2 hPa) for weak storms, and a small intensity reduction for strong storms (1.6 hPa to 3.5 hPa). The average impact of the Amazon and Orinoco plume on TC development may therefore be negligible. However, the balance of the two effects varies locally and seasonally, so that individual TCs may be altered by river-induced changes in ocean stratification.

The potential impact of the Clear-River on steady state TCs ranges from about -1 hPa (weak TCs) to -5 to -12 hPa (strongest TCs occurring in the area). Light attenuation opposes this effect and is slightly larger in magnitude, ranging from +1 hPa

(weak TCs) to +6 to +16 hPa (strong TCs) weakening. Stronger storms are more affected by the changes in ocean stratification since they induce more mixing (Vincent et al., 2014). Results are similar for developing storms with 2% (weak TCs) to 28% (strongest TCs occurring over the plume area) higher intensification rates induced by the clear river. This is comparable to the 50% increase in the presence of BLs found by Balaguru et al. (2012a) on the basis of a global analysis. Light attenuation on the contrary slows down intensification rates by a similar amount.

The river-induced changes in ocean-stratification increase with storm strength, in agreement with Vincent et al. (2014). While the river-induced barrier layer effect on TC intensity is fairly constant throughout the hurricane season, the colour effect is weaker in the early season and strong from August to October. Hurricanes occurring in October may therefore experience even stronger intensity reduction by the plume of up to about +5 hPa. While the impact of ocean stratification depends strongly on the storm strength it is only weakly dependent on the maximum potential intensity (MPI). The impact of freshwater and colour effects depends on atmospheric and oceanic conditions, which determine the threshold of surface cooling causing TC collapse. A lower threshold (1.5°C) significantly enhances the importance of river-induced stratification changes and a higher threshold (3.5°C) slightly reduces the river impact. Since the net change remains a balance of enhanced or reduced components of river impact (freshwater and colour), the net change is less affected by the threshold. The range of possible changes in TC intensity, presented here, includes these environmental variations.

Note that the cancellation of BL and colour effects only applies to the river-induced part of the BL. The total BL (river, rain and subduction induced) is likely to overwhelm colour effects. The overall impact of salinity stratification in the WTNA is therefore likely to be an intensification of TCs. However, ocean colour will reduce this effect.

Chapter 6

Conclusions and further work

6.1 Conclusions

The impact of the Amazon and Orinoco rivers on ocean stratification and temperatures has been examined with a regional ocean model, and been related to the potential impact on tropical cyclone (TC) intensity. Two mechanisms determine the influence of riverine freshwater on the ocean. One is the stabilizing effect of the river freshwater, the other is light absorption by ocean colour. Within the Amazon and Orinoco plume, the two mechanisms have opposing and effective cancelling effects on TCs. On the one hand, sea surface temperature (SST) feedbacks are likely to decrease due to vertical stabilization. On the other hand, this effect is outweighed by subsurface cooling as a consequence of light attenuation.

The freshwater shallows the mixed layer (ML) and thickens the barrier layer (BL) by up to 15 m locally. Isolating the surface from entrainment cooling does not change SSTs in the model, since the warming effect is cancelled by solar penetration (Vialard and Delecluse, 1998). Reduced surface cooling of the BL, on other hand, leads to modest BL warming ($+0.08^{\circ}\text{C}$) and induces temperature inversions. This heating does not significantly increase the heat available for TC development. However, the stable low-salinity plume inhibits vertical mixing by TCs and may therefore reduce SST feedbacks. The cooling inhibition index (CI) is a measure for the resistance of the ocean to vertical mixing and is $+2.2 \text{ (J/m}^2\text{)}^{\frac{1}{3}}$ higher in the presence of a plume. As this mechanism is not represented in the tropical cyclone heat potential (TCHP), CI appears to be the more skilful TC metric in the Amazon and Orinoco plume region. 90% of the increase in CI are due to increased salinity stratification. Therefore freshwater induced changes in TC intensity are likely to be caused by increased stability, rather than modified temperatures.

Light absorbing particles in the river plume change the ocean state considerably.

While the sea surface is only slightly warmed ($+0.1^{\circ}\text{C}$), the 100m mean temperature is cooled by -0.3°C . This is likely to increase the surface cooling magnitude induced by TCs. CI is $-2.1 (\text{J}/\text{m}^2)^{\frac{1}{3}}$ lower when accounting for realistic light absorption, leading to a net river impact of only $+0.1 (\text{J}/\text{m}^2)^{\frac{1}{3}}$ CI increase. The light absorption algorithm applied here (Lee et al., 2005) considers all light absorbing particles in the ocean. Since CDOM is the main absorber in the Amazon and Orinoco plume (Hu et al., 2004; Del Vecchio and Subramaniam, 2004), this is a major advantage to chlorophyll-only based methods (e.g. Ohlmann (2003)). Using salinity as a proxy for light attenuation leads to very similar results, suggesting that the rivers are the main control of solar transmission. However, the salinity approximation underestimates ocean colour in the southern Caribbean, where nutrients are also supplied by cold upwelling waters.

Using simple relations between CI, surface cooling and TC intensity, it has been found that stronger TCs are more affected by the changes in ocean stratification since they induce more mixing, in agreement with Vincent et al. (2014). The potential impact of a clear river on steady state TCs ranges from about -1 hPa (weak TCs) to -5 to -12 hPa (strongest TCs occurring in the area). Light attenuation opposes this effect and is slightly larger in magnitude, ranging from up to +6 to +16 hPa weakening for strong TCs. That leaves a small range of net coloured river impact from +1.6 to +3.5 hPa for the strongest storms. Results are similar for developing storms with 2% (weak TCs) to 28% (strongest TCs occurring over the plume area) higher intensification rates induced by the clear river. This is comparable to the 50% increase in the presence of BLs found by Balaguru et al. (2012a) on the basis of a global analysis. Light attenuation on the contrary slows down intensification rates by a similar amount.

The cancellation of stability and ocean colour effects is based on a multi-annual mean in the area and season relevant for TC development. The balance of the two mechanisms however, is complex and varies in time and space. The plume size for example has a large interannual variability, depending on precipitation, discharge, surface winds and ocean currents (Grodsky et al., 2014). This may lead to important interannual variations in the impact of the rivers on TC intensity. Chlorophyll concentrations vary with the euphotic zone depth, the amount of sunlight, and nutrient abundance. Primary production is also sensitive to iron fertilization (Martin et al., 1993; Mills et al., 2004), which is transported from the Sahara desert into the WTNA by the trade winds (Gao et al., 2001). Certain cyclones could thus be intensified or de-intensified depending on the season and the oceanic conditions on their exact path. This will be different for each cyclone.

The small net effect of the coloured river plume is a specific characteristic of this

region. Results may be different for other river plumes. In BL regions of low chlorophyll concentration, the net effect will still be an intensification of TCs as found in Balaguru et al. (2012a). Furthermore BL-induced TC intensification can be caused by river and precipitation freshening. Idealized calculations suggest, that only a quarter of barrier layer induced CI increase is caused by the Amazon and Orinoco. The overall effect of the total BL (river, rain, and subduction induced) in the WTNA is therefore likely to dominate the colour effect, even though the river-induced components cancel each other. Light attenuation is likely to reduce TC intensification not only in river plumes, but in all areas with high surface chlorophyll concentration like the northern and south-eastern Indian Ocean or the tropical East Pacific. To show the effect of ocean colour on TC intensity in observations, a study similar to Balaguru et al. (2012b), comparing TC intensification in colour-affected and clear regions, could be useful.

6.2 Further work

SST changes induced by ocean colour and river freshwater are likely to modify atmospheric temperatures and circulation (Shell et al., 2003; Timmermann and Jin, 2005; Gnanadesikan et al., 2010; Balaguru et al., 2012b). In this study a forced ocean model is applied and the focus is on the oceanic part of TC intensity control. Hence atmospheric feedbacks are not considered. Furthermore chlorophyll populations induce dynamical and thermal changes in the ocean which may modify the plume dispersal itself, or feed back on primary productivity. Marzeion et al. (2005) coupled their ocean-atmosphere model to a simple biology module and Lengaigne et al. (2007) studied the impact of ocean biology using a fully coupled model with interactive biology. They conclude that interactive chlorophyll is important to correctly represent the seasonal cycle or ENSO variability. Since light absorption in this study is derived from surface optical properties, possible feedbacks between modified ocean temperatures and primary productivity can not be assessed. Note however, that monthly optical properties derived from satellites are applied. Therefore interannual and seasonal variations in ocean colour are represented.

Light attenuation is solely derived from surface values, subsurface properties are inherently prescribed. This may turn out problematic when applied below the well mixed surface layer (Lee et al., 2005). However, there is no satellite product which would reveal the subsurface distribution of light absorbing substances to this date. The absorption algorithm applied here does not distinguish between river-induced ocean colour and other sources. To assess atmospheric, oceanic and biological feed-

backs, and to evaluate the river-induced impact on the WTNA a coupled ocean-ecosystem model with and without river input, that explicitly simulates riverine nutrient and sediment delivery could be applied (e.g. Fennel et al. (2006)).

TC mixing of cold nutrient-rich waters into the ML often results in algae blooms (e.g. Liu et al. (2009); Han et al. (2012); Merritt-Takeuchi and Sen (2013)). Those have the potential to modify ocean temperatures in cyclone wakes, with possible impacts on following storms. Furthermore chlorophyll in the Amazon and Orinoco plume is an important contributor to carbon sequestration (Subramaniam et al., 2007). Algae blooms in TC wakes may enhance the capacity of the river plume for carbon storage. Processes setting the large inter-annual variability of the plume size can thus act supporting or limiting to carbon storage. The plume size varies with discharge and wind mixing (Hu et al., 2004; Nikiema et al., 2007), but plume advection may also be altered by precipitation (Grodsky et al., 2014). Seasonal and interannual variations in the ocean currents may also play a crucial role in the distribution of the river plume across the WTNA. Note that the light ocean water remains on the surface. Higher discharge levels may thus lead to a larger plume via two mechanisms: Firstly, directly by releasing more water to the ocean, and secondly due to enhanced stabilisation as a consequence of more freshwater. The plume size may thus increase non-linearly with river discharge.

TCs deepen the mixed layer, and hence cool the surface and warm the subsurface. As the surface temperature is restored to prestorm values, there is a net warming of the upper ocean (Emanuel, 2001). Since the Amazon and Orinoco change ocean stability and thus the amount of mixing, they may limit the ocean heat uptake in the plume. This may be of relevance for subsequent storms. TC-induced ocean heat uptake has also been suggested to contribute to the Meridional Overturning Circulation (MOC) (Emanuel, 2001). However, based on a model study, Vincent et al. (2012c) conclude that only one tenth of the ocean heat uptake is transported poleward and that the contribution of TCs to the MOC is thus marginal.

Human actions and global warming may change tropical rivers and floodplains in the future. Temperatures in the tropics are predicted to warm by 2-4°C by 2100 (Meehl et al., 2007), which will cause enhanced evaporation and precipitation and also lead to warmer ocean and river temperatures (Hamilton, 2010). Altered river discharge will mainly be a consequence of precipitation changes (Hamilton, 2010). While some tropical rivers are expected to deliver 10-40% more freshwater into the oceans only minor modifications are expected for the Amazon and Orinoco. Over the next 100 years model simulations show a substantial precipitation decrease (20% or more) in June, July and August in the Amazon Basin, but a slight increase (ap-

proximately 5%) in December, January and February (Bates et al., 2008). Milly et al. (2005) find runoff reductions to the order of 5-20% in eastern Amazonia, and an equal increase in western Amazonia. Apart from global warming, deforestation of the Amazonian rainforest may alter the Amazon discharge due to changes in evapotranspiration (Davidson et al., 2012). Altered river discharge and temperatures may affect the temperature gradient of oceanic and riverine waters and possibly modify the plume dispersal across the WTNA.

Ocean colour may also change in a warmer climate. In general higher temperatures enhance primary productivity, but organism may not be able to survive if temperatures reach a critical point (Talling and Lemoalle, 1998). While algae growth is enhanced in the temperature range between 15°C and 30°C, growth rates may decline above 35°C. However, in most tropical waters, primary production rates are often more constrained by nutrient limitation rather than temperature (Lewis, 1987). Day et al. (2008) conclude that climate change mainly influences coastal ecology via sea level rise, changes in storm frequency and intensity, and modified freshwater, sediment, and nutrient inputs. Sea level-rise and increasing storm surges have a direct impact since they modify the freshwater-seawater interface. Note that riverine sediment and nutrient supply to the ocean may also change due to human activity along the river channels in the future (Hamilton, 2010). Seitzinger et al. (2002) estimate that rivers may transport about 250% more nitrogen into the WTNA by 2050 (compared to 1990), due to modified land use (i.e deposition of fertilizers and manure into the river systems).

Future changes in river-induced barrier layer and ocean colour are a complex combination of changing discharge levels and temperatures, nutrient and sediment delivery, and dynamical changes at the ocean-river interface. The individual components may induce feedbacks which makes future predictions an interesting but challenging task (Hamilton, 2010). Models indicate that intense tropical cyclones are likely to become more frequent in a warmer climate (Bender et al., 2010). Since ocean stratification has larger impacts on strong storms, Amazon and Orinoco discharge may become even more important for tropical cyclone development in the future.

Bibliography

- Balaguru, K., Chang, P., Saravanan, R., Leung, L. R., Xu, Z., Li, M., and Hsieh, J., 2012a: Ocean Barrier layers' effect on tropical cyclone intensification, *PNAS*, 109, 14 343–14 347.
- Balaguru, K., Chang, P., Saravanan, R., and Jang, C. J., 2012b: The Barrier Layer of the Atlantic warm pool: Formation mechanism and influence on the mean climate, *TELLUS SERIES A-DYNAMIC METEOROLOGY AND OCEANOGRAPHY*, 64, 18 162.
- Ballabrera-Poy, J., Murtugudde, R., Zhang, R.-H., and Busalacchi, A. J., 2007: Coupled Ocean-Atmosphere Response to Seasonal Modulation of Ocean Color: Impact on Interannual Climate Simulations in the Tropical Pacific, *J. Climate*, 20(2), 353–374.
- Bates, B. C., Kundzewicz, Z. W., Wu, S., and Palutikof, J. P., 2008: Climate Change and Water, Technical Paper of the Intergovernmental Panel on Climate Change, IPCC Secretariat, Geneva.
- Baumgartner, A. and Reichel, E., 1975: *The World Water Balance*, Elsevier Sci., New York.
- Bender, M. A. and Ginis, I., 2000: Real-case simulations of hurricane-ocean interaction using a high-resolution coupled model: Effects on hurricane intensity, *Monthly Weather Review*, 128, 917–946.
- Bender, M. A., Ginis, I., and Kurihara, Y., 1993: Numerical Simulations of Tropical Cyclone-Ocean Interaction With a High-Resolution Coupled Model, *J. Geophys. Res.*, 93, 245–263.
- Bender, M. A., Knutson, T. R., Tuleya, R. E., Sirutis, J. J., Vecchi, G. A., Garner, S. T., and Held, I. M., 2010: Modeled impact of anthropogenic warming on the frequency of intense Atlantic hurricanes, *Science*, 327(5964), 454–458.

- Bjerknes, J., 1969: Atmospheric teleconnections from the equatorial pacific, *Mon.Weather Rev.*, 97, 163–172.
- Black, P. G., 1983: Ocean temperature change induced by tropical cyclones, Ph.D. dissertation, Pennsylvania State University, p. 278.
- Blanke, B., Arhan, M., Lazar, A., and Prevost, G., 2002: A Lagrangian numerical investigation of the origins and fates of the salinity maximum water in the Atlantic, *J. Geophys. Res.*, 107(C10), 3163.
- Buarque, S. R., Vanroyen, C., and Agier, C., 2009: Tropical cyclone heat potential index revisited, *Mercator Ocean Q. Newsl.*, 33, 24–30.
- Castellanos, B., Laraque, A., Aparicio, E., Perez, J., Michel, L., Rodriguez, M., and Fraizy, P., 2010: 9^a Campaña de aforos muestreo de agua y sedimentos en la cuenca del rio Orinoco, *Hidrogeodynamica de la Cuenca del Rio Orinoco*.
- Chapman, D., 1985: Numerical treatment of cross-shelf open boundaries in a barotropic coastal ocean model, *J. Phys. Oceanogr.*, 54, 1451–1473.
- Cherubin, L. and Richardson, P., 2007: Caribbean current variability and the influence of the Amazon and Orinoco freshwater plumes, *Deep Sea Research I*, 54, 1451–1473.
- Choi, B. and Wilkin, J., 2007: The Effect of Wind on the Dispersal of the Hudson River Plume, *J. Phys. Oceanogr.*, 37, 1878–1897.
- Cione, J. J. and Uhlhorn, E. W., 2003: Sea surface temperature variability in hurricanes: Implications with respect to intensity change, *Monthly Weather Review*, 131(8), 1783–1796.
- Coles, V. J., Brooks, M. T., Hopkins, J., Stukel, M. R., Yager, P. L., and Hood, R. R., 2013: The pathways and properties of the Amazon river plume in the tropical North Atlantic Ocean, *J. Geophys. Res.*, 118, 6894–6913.
- D’Asaro, E. A., 2003: The Ocean Boundary Layer below Hurricane Dennis, *J. Phys. Oceanogr.*, 33, 561–579.
- D’Asaro, E. A., Sanford, T. B., Niiler, P. P., and Terrill, E. J., 2007: Cold wake of Hurricane Frances, *Geophys. Res. Lett.*, 34(15), L15 609.

- Davidson, E. A., de Araujo, A. C., Artaxo, P., Balch, J. K., Brown, I. F., Bustamante, M. M. C., Coe, M. T., DeFries, R. S., Keller, M., Longo, M., Munger, J. W., Schroeder, W., Soares-Filho, B. S., Souza, C. M., and Wofsy, S. C., 2012: The Amazon basin in transition, *Nature*, 481(7381), 321–328.
- Day, J. W., Christian, R. R., Boesch, D. M., Yanez-Arancibia, A., Morris, J., Twilley, R. R., Naylor, L., Schaffner, L., and Stevenson, C., 2008: Consequences of climate change on the ecogeomorphology of coastal wetlands, *Estuaries and Coasts*, 31, 477–491.
- de Boyer Montégut, C., Madec, G., Fischer, A. S., and Lazar, A., 2004: Mixed layer depth over the global ocean: An examination of profile data and profile-based climatology, *J. Geophys. Res.*, 109, C12 003.
- de Boyer Montégut, C., Mignot, J., Lazar, A., and Cravatte, S., 2007: Control of salinity on the mixed layer depth in the world ocean: 1. General description, *J. Geophys. Res.*, 112, C06 011.
- Del Castillo, E., Coble, P. G., Morell, J. M., Lopez, J. M., and Corredor, J. E., 1999: Analysis of the optical properties of the Orinoco River plume by absorption and fluorescence spectroscopy, *Marine Chemistry*, 66, 35–51.
- Del Vecchio, R. and Subramaniam, S., 2004: Influence of the Amazon River on the surface optical properties of the western tropical North Atlantic Ocean, *J. Geophys. Res.*, 109, C11 001.
- Denamiel, C., Budgell, W. P., and Toumi, R., 2013: The Congo River plume: Impact of the forcing on the far-field and near-field dynamics, *J. Geophys. Res. - Oceans*, 118(2), 964–989.
- Denman, K. L., 1973: A Time-dependent Model of the Upper Ocean, *J. Phys. Oceanogr.*, 3, 173–184.
- Egbert, G., Bennett, A., and Foreman, M., 1994: TOPEX/POSEIDON tides estimated using a global inverse model, *J. Geophys. Res. - Oceans*, 99(C12), 28,821–24,852.
- Emanuel, K. A., 1987: The dependence of hurricane intensity on climate, *Nature*, 326, 483–485.
- Emanuel, K. A., 1988: The maximum intensity of hurricanes, *J. Atmos. Sci.*, 45, 1143–1155.

- Emanuel, K. A., 1999: Thermodynamic control of hurricane intensity, *Nature*, 401(6754), 665–669.
- Emanuel, K. A., 2001: Contribution of tropical cyclones to meridional heat transport by the oceans, *J. Geophys. Res.*, 106(14), 771–781.
- Emanuel, K. A., 2005: *Divine Wind - The history and science of hurricanes*, Oxford University Press.
- Emanuel, K. A., 2006: Climate and tropical cyclone activity: A new model downscaling approach, *J. Clim.*, 19, 4797–4802.
- Fairall, C. W., Bradley, E. F., Rogers, D. P., Edson, J. B., and Young, G. S., 1996: Bulk parameterization of air-sea fluxes for Tropical Ocean-Global Atmosphere Coupled Ocean-Atmosphere Response Experiment, *J. Geophys. Res.*, 101(C2), 3747–3746.
- Feldman, G. C. and McClain, C. R., 2014: Ocean Color Web, MODIS Aqua and MODIS Terra, L3 ,NASA Goddard Space Flight Center. Eds. Kuring, N., Bailey, S. W. 29.07.2014., <http://oceancolor.gsfc.nasa.gov/>.
- Fennel, K., Wilkin, J., Levin, J., Moisan, J., O'Reilly, J., and Haidvogel, D., 2006: Nitrogen cycling in the Middle Atlantic Bight: Results from a three-dimensional model and implications for the North Atlantic nitrogen budget, *Global Biogeochemical Cycles*, 20(3), GB3007.
- Ffield, A., 2007: Amazon and Orinoco river plumes and NBC rings: Bystanders or participants in hurricane events, *J. Clim.*, 20, 316–333.
- Flather, R., 1976: A tidal model of the northwest European continental shelf, *Memoires de la Societe Royale de Liege*, 6, 141–164.
- Foltz, G. R. and McPhaden, M. J. M., 2009: Impact of Barrier Layer Thickness on SST in the Central Tropical North Atlantic, *Journal of Climate*, 22, 285 – 299.
- Foltz, G. R., Grodsky, S. A., and Carton, J. A., 2003: Seasonal mixed layer heat budget of the tropical Atlantic ocean, *J. Geophys. Res.*, 108(C5), 3146.
- Foltz, G. R., Grodsky, S. A., Carton, J. A., and Phaden, M. J. M. M., 2004: Seasonal salt budget of the northwestern tropical Atlantic Ocean along 38°W, *J. Geophys. Res.*, 109, C03 052.
- Fratantoni, D. and Richardson, P., 2006: The Evolution and Demise of North Brazil Current Rings, *Journal of Physical Oceanography*, 36, 1241–1264.

- Frouin, R. K., Ueyoshio, K., and Kampel, M., 2007: Influence of solar radiation absorbed by phytoplankton on the thermal structure and circulation of the tropical Atlantic ocean, *Proceedings of the Society of Photo-Optical Instrumentation Engineers (SPIE)*, 6680, 68 015–68 015.
- Gao, Y., Kaufman, Y. J., Tanre, D., Kolber, D., and Falkowski, P. G., 2001: Seasonal distributions of aeolian iron fluxes to the global ocean, *Geophys. Res. Lett.*, 28(1), 29–32.
- Gnanadesikan, A. and Anderson, W. G., 2009: Ocean water clarity and the general circulation in a coupled climate model, *J. Phys. Oceanogr.*, 39, 314–332.
- Gnanadesikan, A., Emanuel, K., Vecchi, G. A., Anderson, W. G., and Hallberg, R., 2010: How ocean color can steer Pacific tropical cyclones, *Geophys. Res. Lett.*, 37, L18 802.
- Goldenberg, S. B., Landsea, C. W., Mestas-Nunez, A. M., and Gray, W. M., 2001: The recent increase in Atlantic hurricane activity: causes and implications, *Science*, 293, 474–479.
- Goni, G. J., DeMaria, M., Knaff, J., Sampson, C., Ginis, I., Bringas, F., Mavume, A., Lauer, C., Lin, I.-I., Ali, M. M., Sandery, P., Ramos-Buarque, S., Kang, K., Mehra, A., Chassignet, E., and Halliwell, G., 2009b: Applications of satellite-derived ocean measurements to tropical cyclone intensity forecasting, *Oceanography*, 22(3), 176–183.
- Gordon, H. R., Brown, O. B., Evans, R. H., Brown, J. W., and Baker, K. S., 1988: A semi-analytical radiance model of ocean color, *J. Geophys. Res.*, 93, 10 909–10 924.
- Gray, W. M., 1968: Global view of the origin of tropical disturbances and storms, *Mon. Wea. Rev.*, 96, 669–700.
- Gray, W. M. and Emanuel, K. A., 2010: Introduction to tropical cyclone meteorology, COMET.
- GRDC, 2001: (Global Runoff Data Center)/Federal Institute of Hydrology/Germany, and International Hydrological Programme/United Nations Educational/Scientific and Cultural Organization, <http://rda.ucar.edu/datasets/ds552.1/>, Accessed 20 Mar 2014.
- Grodsky, S. A., Reul, N., Lagerloef, G., Reverdin, G., Carton, J. A., Chapron, B., Quilfen, Y., Kudryavtsev, V. N., and Kao, H.-Y., 2012: Haline hurricane wake in the

- Amazon/Orinoco plume: AQUARIUS/SACD and SMOS observations, *Geophys. Res. Lett.*, 39, L20 603.
- Grodsky, S. A., Reverdin, G., Carton, J. A., and Coles, V. J., 2014: Year-to-year salinity changes in the Amazon plume: Contrasting 2011 and 2012 Aquarius/SACD and SMOS satellite data, *Remote Sensing of Environment*, 140, 14–22.
- Gyory, J., Mariano, A. J., and H.Ryan, E., 2012a: The North Brazil Current, *Ocean Surface Currents*, <http://oceancurrents.rsmas.miami.edu/atlantic/north-brazil.html>, accessed:12.12.2012.
- Gyory, J., Mariano, A. J., and Ryan, E. H., 2012b: The Caribbean Current, *Ocean Surface Currents* <http://oceancurrents.rsmas.miami.edu/caribbean/caribbean.html>, accessed:12.12.2012.
- Gyory, J., Mariano, A., and E.H.Ryan, 2012c: The Guiana Current, *Ocean Surface Currents*, <http://oceancurrents.rsmas.miami.edu/atlantic/guyana.html>, accessed:12.12.2012.
- Haidvogel, D. B., Arango, H., Budgell, W. P., Cornuelle, B. D., Curchitser, E., DiLorenzo, E., Fennel, K., Geyer, W. R., Hermann, A. J., Lanerolle, L., Levin, J., McWilliams, J. C., Miller, A. J., Moore, A. M., Powell, T. M., Shchepetkin, A. F., Sherwood, C. R., Signell, R. P., Warner, J. C., and Wilkin, J., 2008: Ocean forecasting in terrain-following coordinates: Formulation and skill assessment of the regional ocean modeling system, *Journal of Computational Physics*, 227(7), 3595–3624.
- Hamilton, S., 2010: Biogeochemical implications of climate change for tropical rivers and floodplains, *Hydrobiologia*, 657, 19–35.
- Hamilton, S. and Lewis, W. J., 1990: Physical characteristics of the fringing floodplain of the Orinoco River, Venezuela, *Interciencia*, 15, 491–500.
- Han, G., Ma, Z., and Chen, N., 2012: Hurricane Igor impacts on the stratification and phytoplankton bloom over the Grand Banks, *Journal of Marine Systems*, 100-101, 19–25.
- Hedström, K., 2009: Technical Manual for a coupled Sea-Ice/Ocean Model (Version 3), available online: <https://www.myroms.org>.
- Hellweger, F. and Gordon, A., 2002: Tracing Amazon River water into the Caribbean Sea, *Journal of Marine Research*, 60, 537–549.

- Hetland, R. D., 2005: Relating river plume structure to vertical mixing, *Journal of Physical Oceanography*, 35(9), 1667–1688.
- Holland, G. J., 1997: The maximum potential intensity of tropical cyclones, *J. Atmos. Sci.*, 54, 2519–2541.
- Howden, A. D. and Murtugudde, R., 2001: Effects of river inputs into the Bay of Bengal, *J. Geophys. Res.*, 106(C9), 19,825–19,843.
- Hu, C., Montgomery, E., Schmitt, R., and Müller-Karger, F., 2004: The dispersal of the Amazon and Orinoco river water in the tropical Atlantic and Caribbean Sea: Observations from space and S-PLACE floats, *Deep-Sea-Research II*, 51, 1151–1171.
- Hu, C., Lee, Z., and Franz, B., 2012: Chlorophyll a algorithms for oligotrophic oceans: A novel approach based on three-band reflectance difference, *J. Geophys. Res.*, 117, C01 001.
- Hu, C., Feng, L., and Lee, Z., 2013: Uncertainties of SeaWiFS and MODIS remote sensing reflectance: Implications from clear water measurements, *Remote Sensing of Environment*, 133, 168–182.
- Huffman, G., Adler, R., Bolvin, D., Gu, G., Nelkin, E., Bowman, K., Hong, Y., Stocker, E., and Wolff, D., 2007: The TRMM Multi-satellite Precipitation Analysis: Quasi-Global, Multi-Year, Combined-Sensor Precipitation Estimates at Fine Scale, *J. Hydrometeor.*, 8, 38–55.
- Jerlov, N. G., 1976: *Marine Optics*, Elsevier.
- Johns, W., Lee, T., Schott, F., Zantopp, R., and Evans, R., 1990: The North Brazil Current Retroflexion: Seasonal Structure and Eddy Variability, *Journal of Geophysical Research*, 95, 22 103–22 120.
- Johns, W., Townsend, T., Fratantoni, D., and Wilson, W., 2002: On the Atlantic inflow to the Caribbean Sea, *Deep Sea Research I*, 49, 211–243.
- Johnson, E. S., Bonjean, F., Lagerloef, G. S. E., Gunn, J. T., and Mitchum, G. T., 2007: Validation and Error Analysis of OSCAR Sea Surface Currents, *J. Atmos. Oceanic Technol.*, 24, 688–701.
- Jourdain, N., Lengaigne, M., Vialard, J., Madec, G., Menkes, C. E., Vincent, E. M., Jullien, S., and Barnier, B., 2012: Observation-Based Estimates of Surface Cooling Inhibition by Heavy Rainfall under Tropical Cyclones, *J. Clim.*, 43, 205–221.

- Jullien, S., Menkes, C., Marchesiello, P., Jourdain, N. C., Lengaigne, M., Koch-Larrouy, A., Lefevre, J., Vincent, E. M., and Faure, V., 2012: Impact of Tropical Cyclones on the Heat Budget of the South Pacific Ocean, *J. Phys. Oceanogr.*, 42(11), 1882–1906.
- Knapp, K. R., Kruk, M. C., Levinson, D. H., Diamond, H. J., and Neumann, C. J., 2010: The International Best Track Archive for Climate Stewardship (IBTrACS): Unifying tropical cyclone data, *Bull. Am. Meteor. Soc.*, 91, 363–376.
- Lagerloef, G., 2012: Satellite mission monitors ocean surface salinity, *Eos. Trans. AGU*, 83(25), 233.
- Lagerloef, G., 2013: Aquarius salinity validation analysis, Data Version 2.0, Aquarius Proj. Doc. AQ-014-PS-0016, National Aeronautics and Space Administration (NASA).
- Landsea, C. W., 1993: A climatology of intense (or major) Atlantic hurricanes, *Mon. Wea. Rev.*, 121, 1703–1713.
- Lee, S.-K., Enfield, D. B., and Wang, C., 2005b: Ocean general circulation model sensitivity experiments on the annual cycle of Western Hemisphere Warm Pool, *J. Geophys. Res.*, 110, C09004.
- Lee, Z., Carder, K. L., and Arnone, R. A., 2002: Deriving Inherent Optical Properties from Water Color: a Multiband Quasi-Analytical Algorithm for Optically Deep Waters, *Applied Optics*, 41(27), 5755–5772.
- Lee, Z., Du, K., Arnone, R. A., Liew, S., and Penta, B., 2005: Penetration of solar radiation in the upper ocean: A numerical model for oceanic and coastal waters, *J. Geophys. Res.*, 110, C09019.
- Lee, Z., Weidemann, A., Kindle, J., Arnone, R., Carder, K. L., Kendall, L., and Davis, C., 2007: Euphotic zone depth: Its derivation and implication to ocean-color remote sensing, *J. Geophys. Res.*, 112(3), C03009.
- Lee, Z., Lubac, B., Werdell, J., and Arnone, R., 2009: An update of the Quasi-Analytical Algorithm (QAA_v5), available online at: http://www.ioccg.org/groups/Software_OCA/QAA_v5.pdf.
- Leipper, D. and Volgenau, D., 1972: Hurricane Heat Potential of the Gulf of Mexico, *Journal of Physical Oceanography*, 2, 218–224.

- Lellouche, J., Galloudec, O. L., Dré villon, M., Régnier, C., Greiner, E., Garric, G., Ferry, N., Desportes, C., Testut, C. E., Bricaud, C., Bourdallé-Badie, R., Tranchant, B., Benkiran, M., Drillet, Y., Daudin, A., C., and Nicola, D., 2013: Evaluation of global monitoring and forecasting systems at Mercator Océan, *Ocean Sci.*, 9, 57–81.
- Lengaigne, M., Menkes, C., Aumont, O., Gorgues, T., Bopp, L., Andre, J. M., and Madec, G., 2007: Influence of the oceanic biology on the tropical Pacific climate in a coupled general circulation model, *Climate Dynamics*, 28(5), 503–516.
- Lentz, S., 1995a: The Amazon River plume during AMASSEDs: subtidal current variability and the importance of wind forcing, *J. Geophys. Res.*, 100(C2), 2377–2390.
- Lentz, S. and Limebunger, R., 1995: The Amazon River Plume During AMASSEDs - Spatial Characteristics and Salinity Variability, *J. Phys. Res. - Oceans*, 100, 2355–2375.
- Lewis, W. M., 1987: Tropical limnology, *Annual Review of Ecology and Systematics*, 18, 159–184.
- Lewis, W. M., Hamilton, S. K., and Saunders, J. F., 1995: Rivers of Northern South America, *Ecosystems of the World 22. River and Stream Ecosystems*. Elsevier, Amsterdam, pp. 219–256.
- Lin, P., Liu, H. L., and Zhang, X. H., 2008: Effect of Chlorophyll-a Spatial Distribution on Upper Ocean Temperature in the Central and Eastern Equatorial Pacific, *Adv. Atm. Sci.*, 25(4), 585–596.
- Liu, X., Wang, M., and Shi, W., 2009: A study of a Hurricane Katrina-induced phytoplankton bloom using satellite observations and model simulations, *J. Geophys. Res.*, 114(C3), C03 023.
- Liu, Z., Xu, J., Zhu, B., Sun, C., and Zhang, L., 2007: The upper ocean response to tropical cyclones in the northwestern Pacific analyzed with Argo data, *Chinese Journal of Oceanology and Limnology*, 25(2), 123–131.
- Lloyd, I. and Vecchi, G., 2011: Observational Evidence for Oceanic Controls on Hurricane Intensity, *J. Climate*, 24(4), 1183–1153.
- Locarnini, R. A., Mishonov, A. V., Antonov, J. I., Boyer, T. P., Garcia, H. E., Baranova, O. K., Zweng, M. M., Paver, C. R., Reagan, J. R., Johnson, D. R., Hamilton, M., and Seidov, D., 2013: World Ocean Atlas 2013, Volume 1: Temperature. S. Levitus, A. Mishonov Technical Ed.; NOAA Atlas NESDIS, 73.

- Lukas, R. and Lindström, E., 1991: The mixed layer of the western equatorial Pacific Ocean, *J. Geophys. Res.*, 96, 3343–3358.
- Ma, J., Liu, H., Zhan, H., Lin, P., and Du, Y., 2012: Effects of chlorophyll on upper ocean temperature and circulation in the upwelling regions of the South China Sea, *Aquatic Ecosystem Health and Management*, 15(2), 127–134.
- Manizza, M., Quéré, C. L., Watson, A. J., and Buitenhuis, E. T., 2008: Ocean biogeochemical response to phytoplankton-light feedback in a global model, *J. Geophys. Res.*, 113(C10), C10 010.
- Marchesiello, P., McWilliams, J. C., and Shchepetkin, A. F., 2001: Open boundary conditions for long-term integration of regional ocean models, *Ocean Modelling*, 3, 1–20.
- Martin, J. H., Fitzwater, S. E., Gordon, R. M., Hunter, C. N., and Tanner, S. J., 1993: Iron, primary production and carbon-nitrogen flux studies during the JGOFS North Atlantic bloom experiment, *Deep Sea Research Part II: Topical Studies in Oceanography*, 40, 115–134.
- Marzeion, B., Timmermann, A., Murtugudde, R., and Jin, F.-F., 2005: Biophysical Feedbacks in the Tropical Pacific, *Clim. Dyn.*, 18(1), 58–70.
- Masson, S. and Delecluse, P., 2001: Influence of the Amazon River Runoff on the Tropical Atlantic, *Phys. Chem. Earth (B)*, 26, 137–142.
- McPhaden, M. J., Foltz, G. R., Lee, T., Murty, V. S. N., Ravichandran, M., Vecchi, G. A., Vialard, J., Wiggert, J. D., and Yu, L., 2008: Ocean-atmosphere interactions during Cyclone Nargis, *Eos: Transactions, American Geophysical Union*, 90(7), 53–60.
- McPhaden, M. J. M., 1982: Variability in the central equatorial Indian Ocean, part II: Oceanic heat and turbulent energy balance, *J. Mar. Res.*, 4440, 403–419.
- Meade, R. H., 1996: River-sediment inputs to major deltas, in: Milliman, J. D., Haq, B.U. (Eds.), *Sea-level Rise and Coastal Subsidence*. Kluwer Academic Publishers, Dordrecht, Netherlands., pp. 63–85.
- Meade, R. H., Rayol, J. M., and Conceicao, S. C. D., 1991: Backwater Effects in the Amazon River Basin of Brazil, *Environ. Geol. Water Sci.*, 18(2), 105–114.
- Meehl, G. A., Stocker, T. F., Collins, W. D., Friedlingstein, P., Gaye, A. T., Gregory, J. M., Kitoh, A., Knutti, R., Murphy, J. M., Noda, A., Raper, S. C. B., Watterson, I. G.,

- Weaver, A. J., and Zhao, Z.-C., 2007: Global climate projections. In Solomon, S., D. Qin, M. Manning, Z. Chen, M. Marquis, K. B. Averyt, M. Tignor and H. L. Miller (eds), *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*, Cambridge University Press, Cambridge, UK.
- Mei, W., Pasquero, C., and Primeau, F., 2012: The effect of translation speed upon the intensity of tropical cyclones over the tropical ocean, *Geophys. Res. Lett.*, 39(7), L07801.
- Merritt-Takeuchi, A. M. and Sen, C., 2013: Case Studies of Tropical Cyclones and Phytoplankton Blooms over Atlantic and Pacific Regions, *Earth Interactions*, 17, 1–19.
- Mignot, J., de Boyer Montégut, C., Lazar, A., and Cravatte, S., 2007: Control of salinity on the mixed layer depth in the world ocean: 2. Tropical areas, *J. Geophys. Res.*, 112, C10010.
- Mignot, J., Lazar, A., and Lacarra, M., 2012: On the formation of barrier layers and associated vertical temperature inversions: A focus on the northwestern tropical Atlantic, *J. Geophys. Res.*, 117, C02010.
- Miller, B. I., 1958: On the maximum intensity of hurricanes, *J. Meteor.*, 15, 184–195.
- Milliman, J. and Farnsworth, K. L., 2011: *River Discharge to the Coastal Ocean, A Global Synthesis*, Cambridge University Press.
- Mills, M. M., Ridame, C., Davey, M., Roche, J. L., and Geider, R. J., 2004: Iron and phosphorus co-limit nitrogen fixation in the eastern tropical North Atlantic, *Nature*, 429, 292–294.
- Milly, P. C. D., Dunne, K. A., and Vecchia, V., 2005: Global pattern of trends in stream-flow and water availability in a changing climate, *Nature*, 438, 347–350.
- Mobley, C. D., 1994: *Light and Water: Radiative transfer in natural waters*, Academic Press, New York.
- Mobley, C. D., 1995: *Hydrolight 3.0 users's guide*, SRI Int., Menlo Park, Calif.
- Mobley, C. D., 2001: Radiative transfer in the ocean, in *Encyclopedia of Ocean Sciences*, edited by J. H. Steele, S. Thorpe, and K. Turekian, Elsevier, New York, pp. 2321–2330.

- Moller, G. S. F., Novo, E., and Kampel, M., 2010: Space-time variability of the Amazon River plume based on satellite ocean color, *Cont. Shelf Res.*, 30, 3442–3452.
- Morel, A., 1988: Optical modeling of the upper ocean in relation to its biogenous matter content (case I waters), *J. Phys. Oceanogr.*, 93(C9), 10,749.
- Morel, A. and Antoine, D., 1974: Optical properties of pure water and pure seawater, In N. G. J. E. Steeman-Nielsen [ed.], *Optical aspects of oceanography*. Academic, pp. 1–24.
- Morel, A. and Antoine, D., 1994: Heating rate within the upper ocean in relation to its bio-optical state, *J. Phys. Oceanogr.*, 24, 1652–1665.
- Müller-Karger, F., McClain, C., Fisher, T., Essias, W., and Varela, R., 1989: Pigment distribution in the Caribbean Sea Observations from Space, *Prog. Oceanogr.*, 23, 23–64.
- Müller-Karger, F., Richardson, P. L., and McGillicuddy, D., 1995: On the offshore dispersal of the Amazon plume in the North Atlantic, *Deep-Sea Research I*, 42, 2127–2137.
- Murtugudde, R., Beauchamp, J., McClain, C., Lewis, M., and Busalacchi, A., 2002: Effects of penetrative radiation on the upper tropical ocean circulation, *J. Clim.*, 15, 470–486.
- Nakamoto, S., Kumar, S. P., Oberhuber, J. M., Ishizaka, J., Muneyama, K., and Frouin, R., 2001: Response of the equatorial Pacific to chlorophyll pigment in a mixed layer isopycnal ocean general circulation model, *Geophys. Res. Lett.*, 28, 2021–2024.
- Neetu, S., Lengaigne, M., Vincent, E. M., Vialard, J., Madec, G., Samson, G., Kumar, M. R. R., and Durand, F., 2012: Influence of upper-ocean stratification on tropical cyclone-induced surface cooling in the Bay of Bengal, *J. Geophys. Res.*, 117, C12020.
- Nieuwolt, S., 1977: *Tropical Climatology - An Introduction to the Climates of the Low Latitudes*, John Wiley and Sons, London.
- Nikiema, O., Devenon, J., and Baklouti, M., 2007: Numerical modeling of the Amazon River plume, *Continental Shelf Research*, 27, 873–899.
- Nittrouer, C. and De Master, D., 1986: Sedimentary processes on the Amazon continental shelf: past, present and future research, *Continental Shelf Research*, 6(1-2), 5–30.

- Nordin, C., Meija, A., and Delgado, C., 1994: Sediment studies of the Orinoco River, Venezuela, The variability of large rivers. ASCE Press, New York, pp. 243–265.
- Odriozola, A. L., 2004: On the color of the Orinoco River plume, PhD Thesis, University of South Florida.
- Odriozola, A. L., Varela, R., Hu, C., Astor, Y., Lorenzoni, L., and Müller-Karger, F. E., 2007: On the absorption of Light in the Orinoco River plume, *Continental Shelf Research*, 27(10-11), 1447–1464.
- Ohlmann, J. C., 2003: Ocean radiant heating in climate models, *J. Clim.*, 16, 1337–1351.
- Ohlmann, J. C., Siegel, D., and Gautier, C., 1996: Ocean mixed layer heating and solar penetration: A global analysis, *J. Geophys. Res.*, 9, 2265–2280.
- Orlanski, I., 1976: A simple boundary condition for unbounded hyperbolic flows, *Comput. Phys.*, 21, 251–269.
- Pailler, K., Bourles, B., and Gouriou, Y., 1999: The barrier layer in the western tropical Atlantic Ocean, *Geophys. Res. Lett.*, 26, 2069–2072.
- Park, T., Jang, C. J., Jungclaus, J. H., Haak, H., Park, W., and Oh, I. S., 2011: Effects of the Changjiang river discharge on sea surface warming in the Yellow and East China seas in summer, *Continental Shelf Research*, 31(1), 15–22.
- Paulson, C. A. and Simpson, J. J., 1977: Irradiance measurements in the upper ocean, *J. Phys. Oceanogr.*, 7, 953–956.
- Pennington, J. T., Mahoney, K. L., Kuwahara, V. S., Kolber, D. D., Calienes, R., and Chavez, F. P., 2006: Primary production in the eastern tropical Pacific: A review, *Progress in Oceanography*, 69(2-4), 285–317.
- Pope, R. M. and Frey, E. S., 1997: Absorption spectrum (380-700 nm) of pure water II, integrating cavity measurements, *Appl. Opt.*, 36, 8710–8723.
- Price, J., 1981: Upper Ocean response to a Hurricane, *J. of Phys. Oceanogr.*, 11, 153–175.
- Price, J., 2009: Metrics of hurricane-ocean interaction: vertically-integrated or vertically-averaged ocean temperature?, *Ocean Science*, 5, 351–368.

- Pudov, V. D., Varfolomeev, A. A., and Fedorov, N., 1979: Vertical structure of the wake of a typhoon in the upper ocean, *Okeanolokiya*, 21, 142–146.
- Raymond, W. and Kuo, H., 1984: A radiation boundary-condition for multi-dimensional flows, *Q. J. R. Meteorol. Society*, 110, 535–551.
- Reynolds, R., Rayner, N., Smith, T., Stokes, D., and Wang, W., 2002: An improved in situ and satellite SST analysis for climate, *J. Climate*, 15, 1609–1625.
- Reynolds, R. W., Smith, T. M., Liu, C., Chelton, D. B., Casey, K. S., and Schlax, M. G., 2007: Daily high-resolution-blended analyses for sea surface temperature, *J. Clim.*, 20(22), 5473–1496.
- Richardson, P. and Reverdin, G., 1987: Seasonal cycle of velocity in the Atlantic North Equatorial Countercurrent as measured by surface drifters, current meters, and ship drifts, *J. Geophys. Res.*, 92(C4), 3691–3708.
- Rodi, W., 1987: Examples of calculation methods for flow and mixing in stratified fluids, *J. Geophys. Res.*, 92 (C5), 5305–5328.
- Ross, D. A., 1995: *Introduction to Oceanography*, New York, Harper Collins.
- Saha, S., Moorthi, S., Pan, H., Wu, X., Wang, J., Nadiga, S., Tripp, P., Kistler, R., Woollen, J., Behringer, D., Liu, H., Stokes, D., Grumbine, R., Gayno, G., Wang, J., Hou, Y., Chuang, H., Juang, H. H., Sela, J., Iredell, M., Treadon, R., Kleist, D., van Delst, P., Keyser, D., Derber, J., Ek, M., Meng, J., Wei, H., Yang, R., Lord, S., van den Dool, H., Kumar, A., Wang, W., Long, C., Chelliah, M., Xue, Y., Huang, B., Schemm, J., Ebisuzaki, W., Lin, R., Xie, P., Chen, M., Zhou, S., Higgins, W., Zou, C., Liu, Q., Chen, Y., Han, Y., Cucurull, L., Reynolds, R., Rutledge, G., and Goldberg, M., 2010: The NCEP Climate Forecast System Reanalysis, *Bull. Amer. Meteor. Soc.*, 91, 1015–1057.
- Salisbury, J., Vandemark, D., Campbell, J., Hunt, C., Wisser, D., Reul, N., and Chapron, B., 2011: Spatial and temporal coherence between Amazon river discharge, salinity and light absorption by colored organic carbon in western tropical Atlantic surface waters, *J. Geophys. Res.*, 116, C00H02.
- Sanford, T. B., Price, J. F., Girton, J. B., and Webb, D. C., 2007: Highly resolved observations and simulations of the ocean response to a hurricane, *Geophys. Res. Lett.*, 34(13), L13 604.

- Schade, L. R., 2000: Tropical Cyclone Intensity and Sea Surface Temperature, *J. Atmos. Sci.*, 57(18), 3122–3130.
- Schade, L. R. and Emanuel, K. A., 1999: The ocean's effect on the intensity of tropical cyclones: Results from a simple coupled atmosphere-ocean model, *J. Atmos. Sci.*, 56, 642–651.
- Seitzinger, S. P., Kroeze, C., Bouwman, A. F., Caraco, N., Dentener, F., and Styles, R. V., 2002: Global patterns of dissolved inorganic and particulate nitrogen inputs to coastal systems: Recent conditions and future projections, *Estuaries*, 25, 640–655.
- Seo, H., Xie, S.-P., Murtugudde, R., Jochum, M., and Miller, A. J., 2009: Seasonal effects of Indian Ocean freshwater forcing in a regional coupled model, *J. Clim.*, 22, 6577–6596.
- Shang, S., Lee, Z., and Wei, G., 2011: Characterization of MODIS-derived euphotic zone depth: Results for the China Sea, *Remote Sensing of Environment*, 115(1), 180–186.
- Shay, L. K., Elsberry, R. L., and Black, P. G., 1989: Vertical structure of the ocean current response to a hurricane, *J. Phys. Oceanogr.*, 19, 649–669.
- Shay, L. K., Black, P., Mariano, A. J., Hawkins, J. D., and Elsberry, R. L., 1992: Upper ocean response to Hurricane Gilbert, *J. Geophys. Res.*, 97, 20,227–20,248.
- Shay, L. K., Goni, G. J., and Black, P., 2008: Effects of a warm oceanic feature on hurricane Opal, *Mon. Wea. Rev.*, 128, 1366–1383.
- Shchepetkin, A. and McWilliams, J., 2005: The regional oceanic modeling system (ROMS): a split-explicit, free-surface, topography-following-coordinate oceanic model, *Ocean Modeling*, 9(4), 347–404.
- Shell, K., Frouin, R., Nakamoto, S., and Somerville, R. C. J., 2003: Atmospheric response to solar radiation absorbed by phytoplankton, *J. Geophys. Res.*, 108(D15), 4445.
- Shen, W. and Ginis, I., 2003: Effects of surface heat flux-induced sea surface temperature changes on tropical cyclone intensity, *Geophys. Res. Lett.*, 30(18), 1933.
- Siegel, D. A., Ohlmann, J. C., Washburn, L., Bidigare, R. R., Nasse, C. T., Fields, E., and Zhou, Y. M., 1995: Solar-radiation, phytoplankton pigments and the radiant heating of the equatorial Pacific warm pool, *J. Geophys. Res.*, 100, 4885–4891.

- Simspon, J. J. and Dickey, T. D., 1981: The relationship between downward irradiance and upper ocean structure, *J. Phys. Oceanogr.*, 11, 309–323.
- Smith, W. and Sandwell, D., 1990: Global seafloor topography from satellite altimetry and ship depth soundings, *Science*, 277, 1957–1962.
- Smith, W. O. J. and Demaster, D. J., 1996: Phytoplankton biomass and productivity in the Amazon River plume: correlation with seasonal river discharge, *Continental Shelf Research*, 16(3), 291–391.
- Smolarkiewicz, P. K., 1983: A simple positive definite advection scheme with small implicit diffusion, *Mon. Weather Rev.*, 111(3), 479–486.
- Sprintall, J. and Tomczak, M., 1992: Evidence of the barrier layer in the surface layer of the tropics, *Journal of Geophysical Research*, 97, 7305–7316.
- Subramaniam, A., Yager, P. L., Carpenter, E. J., Mahaffey, C., Bjorkman, K. M., Cooley, S. R., Kustka, A., Montoya, J. P., Sanudo-Wilhelmy, S. A., Shipe, R. F., and Capone, D., 2007: Amazon River enhances diazotrophy and carbon sequestration in the tropical North Atlantic Ocean, *Proc. Natl. Acad. Sci. USA*, 105, 10 460–10 465.
- Sweeney, C., Gnanadesikan, A., Griffies, S. M., Harrison, M. J., Rosati, A. J., and Samuels, B. L., 2005: Impacts of Shortwave Penetration Depth on Large-Scale Ocean Circulation and Heat Transport, *J. Phys. Oceanogr.*, 35, 1103–1119.
- Talling, J. F. and Lemoalle, J., 1998: *Ecological dynamics of tropical inland waters*, Cambridge University Press, Cambridge, UK.
- Timmermann, A. and Jin, F.-F., 2005: Phytoplankton influences on tropical climate, *Geophys. Res. Lett.*, 29(23), 2104.
- Umlauf, L. and Burchard, H., 2003: A generic length-scale equation for geophysical turbulence models, *Journal of Marine Research*, 61, 235–265.
- Vialard, J. and Delecluse, P., 1998: An OGCM study for the TOGA decade. Part I: Role of salinity in the physics of the Western Pacific fresh pool, *Journal of Physical Oceanography*, 28, 1071–1088.
- Vinayachandran, P. N., Neema, C. P., Mathew, S., and Remya, R., 2012: Mechanisms of summer intraseasonal sea surface temperature oscillations in the Bay of Bengal, *J. Geophys. Res.*, 117, C01 005.

- Vincent, E., Lengaigne, M., Madec, G., Vialard, J., Samson, G., Jourdain, N., Menkes, C. E., and S.Jullien, 2012a: Processes setting the characteristics of sea surface cooling induced by tropical cyclones, *J. Geophys. Res.*, 117, C02 020.
- Vincent, E., Lengaigne, M., Vialard, J., Madec, G., Jourdain, N., and Masson, S., 2012b: Assessing the Oceanic Control on the Amplitude of Sea Surface Cooling induced by Tropical Cyclones, *J. Geophys. Res.*, 117, C05 023.
- Vincent, E., Madec, G. M., Lengaigne, M., Vialard, J., and Koch-Larrouy, A., 2012c: Influence of tropical cyclones on sea surface temperature seasonal cycle and ocean heat transport, *Clim. Dyn.*, 41, 2019–2038.
- Vincent, E., Emanuel, K., Lengaigne, M., Vialard, J., and Madec, G., 2014: Influence of upper ocean stratification interannual variability on tropical cyclones, *J. Adv. Model. Earth Syst.*, 6.
- Vizy, E. and Cook, K., 2010: Influence of the Amazon/Orinoco Plume on the summer-time Atlantic climate, *Journal of Geophysical Research*, 115, D21 112.
- von Storch, H. and Zwiers, F. W.: *Statistical Analysis in Climate Research*, Cambridge University Press, 1999.
- Wada, A. and Usui, N., 2007: Importance of tropical cyclone heat potential for tropical cyclone intensity and intensification in the western North Pacific, *J. Oceanogr.*, 63, 427 447.
- Wallcraft, A. J., Kara, A. B., Hurlburt, H. E., Chassignet, E. P., and Halliwell, G. H., 2008: Value of bulk heat flux parameterizations for ocean SST prediction, *J. Marine Systems*, 74, 241–258.
- Wang, C. and Enfield, D. B., 2001: The tropical western hemisphere warm pool, *Geophys. Res. Lett.*, 28, 1635–1638.
- Wang, C. and Enfield, D. B., 2003: A further study the Tropical Western Hemisphere Warm Pool, *J. Clim.*, 16(10), 1476–1493.
- Wang, C., Enfield, D. B., Lee, S., and Landsea, C. W., 2006: Influences of the Atlantic Warm Pool on Western Hemisphere Summer Rainfall and Atlantic hurricanes, *J. Clim.*, 19, 3011–3028.
- Wang, W. and Phaden, M. J. M., 1999: the surface-layer heat balance in the equatorial Pacific Ocean, part I: Mean seasonal cycle, *J. Phys. Oceanogr.*, 29, 1812–1831.

- Wang, W., Xie, P., Yoo, S.-H., Xue, Y., Kumar, A., and Wu, X., 2010: An assessment of the surface climate in the NCEP climate forecast system reanalysis, *Clim. Dyn.*, 37, 1601–1620.
- Wang, X., Xie, G., Qi, Y., and Li, W., 2011: Impact of barrier layer on typhoon-induced sea surface cooling, *Dynamics of Atmospheres and Oceans*, 52, 367–385.
- Warne, A., Meade, R., White, A., Guevara, E., Gibeaut, J., Smyth, R.C. Aslan, A., and Trembley, T., 2002: Regional controls on geomorphology, hydrology, and ecosystem integrity in the Orinoco Delta, Venezuela, *Geomorphology*, 44, 273–307.
- Warner, J. C., Geyer, W. R., and Lerczak, J. A., 2005a: Numerical modeling of an estuary: A comprehensive skill assessment, *Journal of Geophysical Research*, 110.
- Warner, J. C., Sherwood, C. R., Arango, H. G., and Signell, R. P., 2005b: Performance of four turbulence closure models implemented using a generic length scale method, *Ocean Model.*, 8(1-2), 81–113.
- Weinkle, J., Maue, R., and Pielke, R., 2012: Historical global tropical cyclone land-falls, *J. Clim.*, 25, 4729–4735.
- White, R. and Toumi, R., 2014: River flow and ocean temperatures: The Congo River, *J. Geophys. Res. - Oceans*, 119(4), 2501–2517.
- Wilcox, D. C., 1988: Reassessment of the scaling-determining equation for advanced turbulence models, *AIAA Journal*, 26, 1299–1310.
- Xiu, P. and Chai, F., 2014: Connections between physical, optical and biogeochemical processes in the Pacific Ocean, *Progress in Oceanography*, 122, 30–53.
- Xue, Y., Huang, B., Hu, Z.-Z., Kumar, A., Wen, C., Behringer, D., and Nadiga, S., 2010: An assessment of oceanic variability in the NCEP climate forecast system reanalysis, *Clim. Dyn.*, 37(11–12), 2511–2539.
- Zedler, S., 2009: Simulations of the Ocean Response to a Hurricane: Nonlinear Processes, *J. Phys. Oceanogr.*, 39(10), 2618–2634.
- Zhang, X., Hu, L., and He, M.-X., 2009: Scattering by pure seawater: Effect of salinity, *Optics Express*, 7, 5698–5710.
- Zweng, M. M., Reagan, J. R., Antonov, J. I., Locarnini, R. A., Mishonov, A. V., Boyer, T. P., Garcia, H. E., Baranova, O. K., Johnson, D. R., Seidov, D., and Biddle, M. M.,

2013: World Ocean Atlas 2013, Volume 2: Salinity. S. Levitus, Ed., A. Mishonov Technical Ed.; NOAA Atlas NESDIS, 74.

