Surface Cooling Due to Precipitation in the Tropics

by

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Abstract

Title: Surface Cooling Due to Precipitation in the Tropics

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Precipitation is an important component of the global hydrological cycle. It affects the upper ocean salinity by adding freshwater to the ocean. In addition, precipitation plays a role in cooling the ocean and land surfaces, when the temperature of the raindrops is lower than the temperature of the surface. However, the surface cooling due to precipitation ($Q_P$) remains an overlooked feature and is not included in the Intergovernmental Panel on Climate Change (IPCC) models. One might expect that heavy precipitation events may lead to large $Q_P$, which can impact the surface temperature, surface energy balance, and local- to regional-scale circulations.

The maximum $Q_P$ values are found following the annual march of the Intertropical Convergence Zone (ITCZ) and in the Indian Monsoon region. The mean $Q_P$ from 30°N to 30°S, for precipitating days is 1.33 W m$^{-2}$, with an annual variation of 1.0 - 1.7 W m$^{-2}$. Relatively higher $Q_P$ values are concentrated in the Bay of Bengal and Indo-Pacific Warm Pool regions. $Q_P$ values for above average rain can exceed 50% of the surface turbulent sensible heat ($Q_{SH}$) close to the tropics, and. For rain values above the 95th percentile, $Q_P$ can exceed 14 W m$^{-2}$ and have the same magnitude as the $Q_{SH}$.

To explore the role of $Q_P$, this term was included in the Weather Research and Forecasting (WRF) model, which was coupled to an ocean model. The simulation with $Q_P$ had significant impact on the upper ocean dynamics and thermodynamics, and also on the
atmosphere. As $Q_P$ lowers the surface temperature, it tends to reduce the atmospheric convection. Latent and sensible heat also decrease over the ocean. $Q_P$ was also included into the land surface model within the WRF to explore its role on the land surface and subsequent atmospheric circulations over the Maritime Continent during the passage of a Madden-Julian Oscillation (MJO) event. During the day, $Q_P$ has a cooling effect since the land surface is warmer than the air. But for precipitation during night, it can have the opposite effect in the presence of warmer air above the surface, causing surface warming and an increase in rain. The results indicate that $Q_P$ should be included in model simulations over the deep convective regions of the tropics.
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List of Abbreviations

3D-PWP: Three dimensional – Price-Weller-Pinkel ocean model coupled to WRF

COARE: Coupled Ocean-Atmosphere Response Experiment

ECMWF: European Centre for Medium-Range Weather Forecasts

GCMs: Global Climate Models

IPCC: Intergovernmental Panel on Climate Change

ITCZ: Intertropical Convergence Zone

MC: Maritime Continent

MJO: Madden-Julian oscillation

NCEP: National Center for Environmental Prediction

OAFlux: Objectively Analyzed Air-Sea Fluxes

PIRATA: Prediction and Research Moored Array in the Atlantic

RAMA: Research Moored Array for African-Asian-Australian Monsoon Analysis and Prediction

SST: Sea Surface Temperature

TAO: Tropical Atmosphere Ocean Project

TRMM: Tropical Rainfall Measuring Mission

WRF: Weather Research and Forecasting
List of Symbols

\(Q_{\text{LH}}\): Latent heat

\(Q_P\): Sensible heat due to rain

\(Q_{P,B}\): Sensible heat due to rain from buoys

\(Q_{P,R}\): Sensible heat due to rain from reanalysis

\(Q_{\text{SH}}\): Sensible heat

\(\text{RNet}\): Net radiation
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Chapter 1
Introduction

1.1 Surface Energy Budget and the Sensible Heat Due to Rain

Solar radiation plays a vital role in atmospheric motion. Incoming solar radiation can be reflected back to space by the atmosphere and clouds, can be absorbed by the atmosphere, or, can be transmitted through the atmosphere until it reaches the surface. At the surface, this radiation is reflected back, or is absorbed by the Earth’s surface. The solar radiation that is absorbed by the surface is then emitted back to the atmosphere as longwave radiation. Apart from heat loss due to longwave radiation at the surface, there is also heat loss due to turbulent heat flux, which is the sum of latent ($Q_{\text{LH}}$, due to evaporation of water) and sensible ($Q_{\text{SH}}$, due to air-sea temperature difference) heat flux (Figure 1). Another cooling term that is often not included in models is the sensible heat loss due to precipitation ($Q_P$), since the temperature of raindrops is typically lower than the temperature of the surface of the ocean/land.

As Gosnell et al. (1995) pointed out, the global averages are from 80 to 125 W m$^{-2}$ for the $Q_{\text{LH}}$ and from 100 to 200 W m$^{-2}$ for the net radiation, which are the major components of the net heat flux. Trenberth et al. (2009) calculated the averaged $Q_{\text{SH}}$ to be between 15.7 and 18.9 W m$^{-2}$ globally, from 26.3 to 27.5 W m$^{-2}$ over land and from 11.8 to 16.0 W m$^{-2}$ over the ocean. The average $Q_P$ during the TOGA-COARE experiment was found to be 2.5 W m$^{-2}$, but, with values as big as 200 W m$^{-2}$ (Gosnell et al. 1995), affecting significantly the temperature of the ocean surface. Boudreau (1965) obtained measurements showing much cooler skin temperature of the water after the occurrence of rain. As the skin temperature provides boundary conditions to the atmosphere above it, this is critical to processes within
the planetary boundary layer (PBL, Chen and Dudhia 2001). Katsaros and Buettner (1969) studied the changes in salinity and temperature profiles due to rain in a still saltwater tank and found that large raindrops can cause substantial mixing on the first few centimeters, but the temperature change effects can be carried out to greater depths by molecular transfer.

Figure 1 - Surface energy balance (W m$^{-2}$; adapted from Trenberth et al 2009).

Following that research, Katsaros (1976) also found that heat loss by precipitation was 1.6 times the loss caused by sensible heat, in a 24-hour observational case study under a low pressure disturbance in the equatorial Atlantic (August-September 1957). A time-series analysis of buoy data by Cronin and McPhaden (1997) showed prolonged precipitation events that caused cool, freshwater anomalies in the surface salinity and to the subsurface temperature. The precipitation increases the buoyancy of the upper ocean, whereas, the $Q_P$ is expected to decrease the buoyancy.

The same study by Cronin and McPhaden (1997) presented an event using the hourly data over 0°, 156°E where $Q_P$ reached 300 W m$^{-2}$ (October 1992; Table 1). Flament and Sawyer (1995) observed that $Q_P$ in a rain event in September 4 of 1990 accounted for about 40% of the net heat flux. Even though $Q_P$ has been found to be significant during heavy precipitation events, it is usually ignored and is not included in any IPCC (Intergovernmental Panel on Climate Change) models.
Table 1 - Summary of $Q_P$ values from the past studies

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<th>Author</th>
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<th>Max. $Q_P$</th>
<th>Position</th>
<th>Time</th>
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<td>213 W m$^{-2}$</td>
<td>156°E 2°S</td>
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<td>Cronin and McPhaden (1997)</td>
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<td></td>
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<td>~7 W m$^{-2}$ for ITCZ</td>
<td>400 W m$^{-2}$</td>
<td>Transit from 140°W 2°N to Hawaii</td>
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1.2 Objectives

The goals of this study are the following:

1. Provide a comprehensive documentation of the spatio-temporal variability of the heat flux from rainfall ($Q_P$) over the tropical oceans and land areas using a variety of observational datasets;

2. Implement the term into an ocean model coupled to the Weather Research and Forecasting (WRF) model, to observe ocean variations of temperature and the impact on convection;

3. Implement the term into the Noah land surface model coupled to the WRF model to observe the impact on the diurnal cycle of precipitation and temperature over land.
1.2 Organization of the Thesis

This study is organized in the following chapters. Chapter 2 describes the data used to create the $Q_F$ climatology, and the data used as initial and boundary condition for the simulations. It also contains a brief description of the methods and model configurations used for the numerical simulations. Chapter 3 describes the $Q_F$ climatology. Chapter 4 is divided into two sections: the first describes the implementation of $Q_F$ into WRF, and the impacts over the ocean, and the second section describes the impact of $Q_F$ over land.
Chapter 2
Data

2.1 Observational Data

The precipitation data is from Tropical Rainfall Measuring Mission version 3B-42 (TRMM 3B-42, 0.25°x0.25°, 3-hrly, Huffman et al. 2007) from 1998 to 2013. The data for surface pressure was obtained from the National Centers for Environmental Prediction Reanalysis 2 (NCEP2, variable resolution, 6-hrly, data provided by the NOAA/OAR/ESRL PSD; Kanamitsu et al. 2011). The latent and sensible heat fluxes, the specific humidity and temperature at 2-m and the surface temperature (skin temperature) are provided by the Objectively Analyzed Air-Sea Fluxes (OAFlux, 1.0°x1.0°, daily, Yu et al. 2008) for global oceans.

The initial and boundary condition for the simulations are from the European Centre for Medium-Range Weather Forecasts (ECMWF) ERA-Interim Reanalysis (0.703° x 0.702°, 6-hrly, Dee et al. 2011). The sea surface temperatures are from the high resolution, global, sea surface temperature analysis (RTG_SST) developed at the National Centers for Environmental Prediction/Marine Modeling and Analysis Branch (NCEP/MMBA, 0.5°x0.5°, daily, Reynolds et al. 2007).

Buoy data for December 2006 corresponds to the location 0°N; 165°E and was provided by the Tropical Atmosphere Ocean (TAO) Project Office of NOAA/PMEL - Pacific Marine Environmental Laboratory (TAO-buoy, McPhaden et al. 1998). The sensible heat due to rain, when available (Figure 2), was obtained from the TAO Project Office website (http://www.pmel.noaa.gov/tao/oceansites/flux/main.html ; downloaded in May, 2016)
2.2 Model

Simulations were conducted with the version 3.7 of the Weather Research and Forecasting (WRF) model, using the advanced research WRF (ARW) dynamic core (Skamarock et al. 2008). The outer domain has a grid spacing of 30 km and the inner domain has a grid spacing of 10 km. The sensible heat due to precipitation and a fresh water input were added to a three-dimensional simplified ocean model (Price et al. 1986; Price et al. 1994) coupled to the WRF model.
The following set of physical parameterizations were used: 6-class microphysics scheme (Hong and Lim 2006), Rapid Radiative Transfer Model for GCMs (RRTMG) longwave and shortwave radiation schemes (Iacono et al. 2008), the new Kain-Fritsch cumulus parameterization (Kain 2004) and Yonsei University planetary boundary layer (Hong et al. 2006). The initial testing (2-day simulation) was conducted using the five layer thermal diffusion scheme for land surface option (Dudhia 1996), and following simulations were conducted using the Noah land surface model (Tewari et al. 2004).

2.3 Methods

The first objective of this research was to describe \( Q_P \) climatology in the tropics to identify the time of the year and areas where \( Q_P \) is important compared with other surface flux components. The OAFlux data is daily. As a result, it does not allow us to identify the critical time of day when \( Q_P \) might be larger compared with other surface heat flux components.

The \( Q_P \) is calculated following Gosnell et al. (1995):

\[
Q_P = C_W R \times (T_0 - T_r)
\]

Where \( C_W \) is the specific heat of water (4186 J kg\(^{-1}\) K), \( R \) is the rain rate, \( T_0 \) is the bulk SST approximated by the skin temperature, and \( T_r \) is the temperature of raindrops when it reaches the surface. \( T_r \) is approximated by the wet bulb temperature (Katsaros 1969) and is roughly 0.1 K cooler than the wet bulb temperature (Gosnell et al. 1995). Wet bulb temperature is obtained using air temperature and relative humidity following Stull (2011):

\[
T_r \equiv T_W = T \atan[0.151977 \times (RH\% + 8.313659)^{1/2} ] + \atan(T + RH\%)
\]

\[
- \atan(RH\% - 1.676331) + 0.00391838 \times (RH\%)^3
\]

\[
\times \atan(0.023101 \times RH\%) - 4.686035
\]
Where $RH\%$ is the relative humidity in percentage, $T_w$ is the wet bulb temperature in Celsius, and $T$ is the air temperature also in Celsius. The Eq. (2) is valid for relative humidity values ranging from 5% to 99% and air temperature between -20°C and 50°C, and therefore is suitable for the tropical region.

The total sensible heat ($Q_S$) is then defined as the sum of the turbulent sensible heat ($Q_{SH}$) and the cooling term ($Q_P$) (Gosnell et al 1995):

$$Q_S = Q_{SH} + Q_P$$  \hspace{1cm} (3).

### 2.4 Model implementation of $Q_P$

The WRF model requires SST as surface boundary conditions and provides several options to do so: a static option, a SST update option, and an ocean-atmosphere coupled option. The static option maintains the SST (provided as initial condition) as constant during the entire simulation. For longer simulations, it is recommended to use the SST update option, where, the user provides SST for the entire domain and the SST is updated to the values provided at user set intervals. The third option has two choices: a one-dimensional ocean mixed layer option where some processes like advection is not present but there’s an air-sea interaction that updates the SST. The second choice is a three-dimensional simplified ocean model, which is used for this project. This option is available since version 3.5 of WRF. This choice is more appropriate for the proposed research since precipitation has large spatio-temporal fluctuations and we are interested to find $Q_P$ and its influence in the tropics.

The three-dimensional (3D) PWP (Price, Weller and Pinkel 1986; Price et al. 1994) ocean model coupled to the WRF is a simplified ocean model, meaning that it doesn’t have an ocean bottom. The absence of ocean bottom does not interfere with the surface fluxes that are the focus of this study. This model can be initialized with up to 30 depths and different temperatures and salinities for each depth. The temperature and salinity for each layer changes throughout the simulation depending on the salt and heat fluxes. A momentum flux equation also provides ocean currents. The depths used in this case were (in meters): 1, 2, 4,
The standard code to compute salinity does not include freshwater flux (i.e., precipitation). We modified the code to include precipitation to compute salinity. The momentum flux allows vertical mixing and depends on wind stress, mixed layer depth and density. The heat flux equation, which includes both radiation and turbulent heat fluxes, and controls the surface heat balance is the focus of this project.

The shortwave radiation can penetrate to the deeper layers following this relationship (Price et al. 1986):

\[ I(z) = I(0) \times \left[ I_1 \times \exp\left(-z/\lambda_1\right) + I_2 \times \exp\left(-z/\lambda_2\right) \right] \quad (4), \]

where \( I(z) \) is the shortwave that reaches the depth \( z \) and \( I(0) \) is the irradiation at the surface, \( I_1 \) is the proportion of shortwave with respect to shortwave plus longwave, and \( I_2 \) is \( 1-I_1 \). \( \lambda_1 \) and \( \lambda_2 \) are constants with values 0.6 m and 20 m respectively. 99% of the shortwave is absorbed within the first 60 to 70 meters of the ocean.

The \( Q_P \) term was calculated as in Eq. (1) and the latent heat (\( Q_{LH} \)) was also added to the outgoing longwave at the surface, becoming:

\[ Q_1 = -1 \times (Q_{LH} + Q_{SH} + Q_P + \sigma T^4) \quad (5), \]

where \( Q_{SH} \) is the sensible heat due to the air-sea temperature difference, \( Q_{LH} \) is the latent heat, \( Q_P \) is the sensible heat due to rain, \( \sigma \) is the Stefan-Boltzmann constant (5.67051 × 10^{-8} W m^{-2} K^{-4}) and \( T \) the temperature of the top ocean layer in Kelvin.

To test the code, an area that spans from 15°S to 10°N and from 147°E to 174°E, as shown on Figure 3, was chosen for the period between 1200 UTC 13 December 2012 to 1200 UTC 15 December 2012 because of the presence of a heavy precipitating event (Figure 4).
Figure 3 - Study area. Outer domain (d01) includes Papua New Guinea and the Solomon Islands. The precipitation event is centered in the inner domain (d02).

The $Q_P$ is directly proportional to the difference between wet bulb and skin temperatures and the rain rate (Eq. (1)). However, the variation in the temperature difference is smaller compared with the variation in precipitation. For example, we can notice in Figure 4 that the maximum rain rate and the higher $Q_P$ values are aligned. At 0000 UTC 14 December 2012 (Figure 5), the peak of $Q_P$ reached ~65 W m$^{-2}$ around 3°S and 167°E, which is located south of the peak in $Q_{SH}$ (~65 W m$^{-2}$).

Figure 6 shows the difference between $Q_{SH}$ and $Q_P$. In general, $Q_{SH}$ is larger than $Q_P$, however, in many parts of the domain, $Q_P$ can be larger than $Q_{SH}$. For example, near 5°S and 163°E $Q_P$ is larger than $Q_{SH}$ by more than 30-40 W m$^{-2}$. This indicates that $Q_P$ can have substantial influence on the upper ocean dynamics and thermodynamics and needs to be included in coupled models.
Figure 4 - Rain (mm h\(^{-1}\), shaded) and 10 m winds (m s\(^{-1}\), vector) for 0000 UTC 14 December 2012. The solid black contour shows the area where \( Q_P \) is larger than 20 W m\(^{-2}\).

Figure 5 - (left) \( Q_{SH} \) (W m\(^{-2}\), shaded) and 10 m winds (m s\(^{-1}\), vector) from NCEP-2 and (right) \( Q_P \) (W m\(^{-2}\), shaded) and 10 m winds (m/s, vector) for 0000 UTC 14 December 2012. Precipitation was taken from TRMM for the \( Q_P \) calculation.
The WRF model was used to simulate the event above with and without the code modifications, combining that with and without updated sea surface temperature. The results shown are area averages for the areas depicted in Figure 7.

![Wind and Qsh minus Qp for 14/DEC/2012 - 00Z](image)

**Figure 6** - $Q_{sh}$ minus $Q_p$ (W m$^{-2}$, shaded) and winds at 10 m (m s$^{-1}$, vector) for 0000 UTC 14 December 2012.

The temperature shows a difference of 2.5 K (Figure 8) between before and after the changes at 0400 UTC of 15 December 2012. Most of this change is due to the inclusion of the latent heat in the heat balance. Almost no change is observed from using fixed or updated observational SST, due to the fact that the SST was coming from the ocean model. The salinity flux equation was inconsistent and was found to have a sign error within the code that was adjusted for in this project. Communications with the author who wrote the code revealed that the latent heat was missing from the heat balance and the sign of evaporation at the salt flux equation was wrong. We included these two corrections and new simulations
were conducted. Therefore, the differences in the results now are due to the addition of the precipitation effects in the salt and the heat balance in the model.

![Study area. Area used for the average over the two (red) and four (green) day simulations.](image)

The time evolution of precipitation and skin temperature with and without $Q_P$ can be found in Figure 9, and the difference in the precipitation is observed in Figure 10, but without a regular pattern. It shows very little difference between the control and the simulation with $Q_P$. It was a concern that the ocean model would still be entering equilibrium in first few hours of simulation so the results shown at Figure 9 and Figure 10 start 48h before the previous simulation.
For this area, the temperature and rain appear to have decreased after the addition of effects from precipitation. The subject of interest was a westward propagating precipitating event in a low-level easterly flow. The experiments failed to simulate this particular event adequately, showing it weaker and shorter lived. In the longer simulation, a cyclone in the southern hemisphere forms within the first few hours of simulation, and the subject of interest appears even weaker. For this reason, the area used to produce Figure 10 is centered in a different position than Figure 9. The simulated QP seems to be consistent with QP calculated from observation.

Figure 8 - Hourly rain (mm; bars) and skin temperature (K; lines). Updated SST (dark shaded) and fixed SST (light shades). Results represent before (green) and after (blue) the modifications. Area average (2.5°S-0.5°N; 159°W-164°W; red area Fig. 7).
Figure 9 - Hourly rain (mm; bars) and skin temperature (K; lines), with $Q_r$ (green) and without $Q_r$ (blue). Area average (7.0°S-5.0°S; 158°W-161°W).

Figure 10 - Difference between control run and the experiment. Rain (mm; bars) and skin temperature (K; line). Values above the 0 line signify that the numbers were higher before the adding of precipitation effects.
Chapter 3
Climatology of $Q_p$

The daily sensible heat due to precipitation ($Q_p$) was calculated using temperature and humidity at 2 meters and skin temperature from OAFlux dataset, pressure at the surface, from NCEP-Reanalysis 2, and the rain from the TRMM. The daily ($Q_p$) was then averaged to produce the monthly means. Due to availability of TRMM data, the study period extended from January 1998 to December 2013 and spans from 30°S to 30°N.

3.1 Results and Discussion

To understand the spatio-temporal variability of $Q_p$, it was calculated under different scenarios. In the first scenario (All), the monthly average was calculated using both precipitating and non-precipitating days. For the second (Precip), the monthly $Q_p$ was calculated using only precipitating days (rain > 0.1 mm day$^{-1}$). In the third scenario (Avg), the monthly $Q_p$ was calculated with the rain values that surpassed the average rain for that period (8.3 mm day$^{-1}$ during January 1998 - December 2013), and in the fourth scenario (p95) the monthly $Q_p$ was calculated with the rain values that surpassed the 95th percentile (37.53 mm day$^{-1}$).

The majority of the data points contain the value zero (no rain). The percentile statistics (Figure 11) consider only the values above 0.1 mm day$^{-1}$ (pluviometric uncertainty), which for this study is the threshold used to consider a day as a precipitating day. The distribution of percentiles of daily TRMM rainfall over the ocean shows that 80% of the rain events have less than 11.6 mm day$^{-1}$ (Figure 11), and that only 5% of the data points present rainfall above 37.5 mm day$^{-1}$.
Based on the above, the monthly rain climatology for precipitating days is shown in Figure 12.

**Figure 11 - TRMM ocean rainfall evolution accordingly to percentile. Unit: mm day$^{-1}$.**

**Figure 12 - $Q_p$ (orange; W m$^{-2}$) and monthly averaged precipitation (blue; mm day$^{-1}$) considering the precipitating days from the TRMM data over the ocean between 30°S and 30°N from January 1998 to December 2013.**
The driest months are March and April. The rainiest months are those of the austral spring between October and December which coincide with the Inter-Tropical Convergence Zone (ITCZ) passage from the northern to the southern hemisphere. There is also a second precipitation peak between May and July that corresponded to the months of the Indian Monsoon, and with the presence of the ITCZ north of the equator. The $Q_p$ follows the same tendency as precipitation and, as we can see in Figure 13, the $Q_p$ values follow closely the annual movement of the ITCZ, being largest between June and November in the Northern Hemisphere and going up to 2 W m$^{-2}$. The lowest values can be found around 15° N between February and April and from 20° to 15°S between mid-July to mid-September.

Figure 13 - Time evolution of climatological zonal mean of $Q_p$ (W m$^{-2}$) for precipitating days.
Figure 14 shows the variability of precipitation over the chosen time period. The rain average presents an increasing tendency for the study period. As $Q_p$ is directly proportional to precipitation, it is expected for the $Q_p$ to show a similar tendency over the years, which was apparent in Figure 15. The monthly average $Q_p$ from precipitating days over the entire period varies between 1.0 W m$^{-2}$ and 1.7 W m$^{-2}$, with the largest monthly values found at the end of the time series.

The $Q_p$ is also dependent of the difference in temperature between the surface and the wet bulb. As we can see in Figure 16, the average of that temperature difference doesn’t show the same increasing tendency. Instead, the values of surface and wet bulb temperature appear to be getting closer showing a smooth decreasing tendency on the difference value.

![Figure 14 - (Blue) Monthly evolution of the rain average for the TRMM data (mm day$^{-1}$) over the ocean between 30°S and 30°N. (Orange) Rain average for the entire period.](image-url)
Figure 15 - Monthly evolution of average $Q_P$ over the ocean between 30°S and 30°N ($Rain > 0.1 \text{ mm day}^{-1}$).

Figure 17 (a) shows maximum $Q_P$ values around the equator following the annual movement of the ITCZ. The values of $Q_P$ are very low, with higher values only going up to 2 W m$^{-2}$ but, this average considers precipitating and non-precipitating days and is a monthly average. When we consider only precipitating days (Figure 17 (b)), some areas with values up to 4 W m$^{-2}$ appear around 7°N still following the ITCZ movement.

Narrowing down the number of precipitating days to only the days where the rain values are above the average, the $Q_P$ can exceed 4 W m$^{-2}$ in the monthly mean. Comparing the $Q_P$ calculated with the rain values above average (Figure 17 (c)) with the magnitude of the sensible heat, we notice that close to the tropics the $Q_P$ value can exceed 50% of the $Q_{SH}$ (Figure 18 (c)). $Q_P$ values calculated with the 95th percentile of rain can exceed 14 W m$^{-2}$ (Figure 17 (d)) and have the same magnitude as the $Q_{SH}$ (Figure 18 (d)).
Figure 16 - Monthly evolution of average difference between wet bulb temperature and ocean surface temperature between 30°S and 30°N (Rain > 0.1 mm day$^{-1}$).
Figure 17 - Time evolution of monthly averaged $Q_p$ (zonally averaged – W m$^{-2}$) using (a) precipitating and non-precipitating days, (b) only precipitating days, (c) precipitating days above rain average (8.3 mm day$^{-1}$) and (d) rain above the 95th percentile (37.5 mm day$^{-1}$).
Figure 18 - Ratio between sensible heat due to precipitation and sensible heat for (a) precipitating and non-precipitating days, (b) only precipitating days, (c) precipitating days above rain average (8.3 mm day$^{-1}$) and (d) rain above the 95th percentile (37.5 mm day$^{-1}$).
Considering the precipitating days, the $Q_p$ values were calculated for different seasons. The areas chosen for the annual and seasonal averages are presented in the first column of Table 2 and where chosen based on the monthly climatology. The largest values for $Q_p$ are found over the Warm Pool area, reaching 2.0 W m$^{-2}$ (Table 2) in JJA (June/July/August). The sensible heat ($Q_{sh}$) is also only computed for the precipitating days and the largest values are found in the tropical areas that are farthest from the Equator, with 30.9 and 15.5 W m$^{-2}$ for the South and East China Sea and 15.2 W m$^{-2}$ for the south and east coast of Brazil. Those values, found on specific seasons, could be related to the larger temperature differences between air and surface and to higher wind speeds.
As we can see in Table 2, the $Q_P$ varies from 0.8 to 2.0 W m$^{-2}$ whereas $Q_{SH}$ has a wider range that extends from 4.3 to 30.9 W m$^{-2}$, with lower annual values in the Bay of Bengal.

Annually, if we compare the magnitudes of $Q_P$ and $Q_{SH}$ (Table 3), the largest percentages are found in the Bay of Bengal (22.5%), and in the Warm Pool (19.3%). The largest value is for JJA in the Bay of Bengal, were the area averaged $Q_P$ magnitude is 32.1% the magnitude of $Q_{SH}$.

**Table 3 - Percentage of $Q_P$ over $Q_{SH}$ for precipitating days.**

<table>
<thead>
<tr>
<th>Area</th>
<th>Annual</th>
<th>MAM</th>
<th>JJA</th>
<th>SON</th>
<th>DJF</th>
</tr>
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<tbody>
<tr>
<td>Tropics (10°S-10°N)</td>
<td>14.1</td>
<td>16.1</td>
<td>13.3</td>
<td>13.7</td>
<td>15.7</td>
</tr>
<tr>
<td>Tropics (20°S-20°N)</td>
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<td>12.0</td>
<td>10.3</td>
<td>11.4</td>
<td>11.7</td>
</tr>
<tr>
<td>Tropics (30°S-30°N)</td>
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<td>9.6</td>
<td>8.6</td>
<td>9.5</td>
<td>9.1</td>
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<td>17.2</td>
<td>14.8</td>
<td>16.9</td>
<td>18.0</td>
</tr>
<tr>
<td>Bay of Bengal</td>
<td><strong>22.5</strong></td>
<td><strong>24.8</strong></td>
<td><strong>32.1</strong></td>
<td><strong>23.2</strong></td>
<td>14.2</td>
</tr>
<tr>
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<td>18.8</td>
<td>18.7</td>
<td>18.8</td>
<td>18.9</td>
</tr>
<tr>
<td>Warm Pool</td>
<td>19.3</td>
<td>19.8</td>
<td>19.3</td>
<td>19.0</td>
<td>19.3</td>
</tr>
<tr>
<td>S. and E. China Sea</td>
<td>8.5</td>
<td>8.9</td>
<td>24.2</td>
<td>12.1</td>
<td>3.5</td>
</tr>
<tr>
<td>S. and E. coast of Brazil</td>
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<td>9.8</td>
<td>7.0</td>
<td>11.4</td>
<td><strong>21.1</strong></td>
</tr>
</tbody>
</table>

The daily values of $Q_P$ calculated from TRMM and OAFlux can be compared with $Q_P$ from buoys when available (Figure 2). Table 4 shows the latitude and longitude of buoys with more than 1000 days of available $Q_P$ data. The $Q_P$ from buoys is available from the TAO Project and, it’s calculated using the Coupled Ocean-Atmosphere Response Experiment (COARE) 3.0b algorithm whenever the complete dataset of variables are available from that particular buoy. The number of buoys with rain measurements is limited. The average $Q_P$ from TRMM and OAFlux will be referenced here as $Q_P$– Reanalysis ($Q_{P-R}$) and its average is calculated only with days where rain sensible heat from buoys ($Q_{P-B}$) is available.
Table 4 - Averaged daily $Q_P$ from buoys ($Q_{P,B} \ W \ m^{-2}$) and from Reanalysis ($Q_{P,R} \ W \ m^{-2}$). Buoy, lat. and lon. correspond to the buoy location. No. of days with $Q_P$ data represent the amount of that in the study period with available $Q_P$ data for that buoy. No. of days with $Q_{P,B} > Q_{SH-B}$ represent the amount of days in the buoy where $Q_P$ is larger than $Q_{SH}$. Largest $Q_{P,B}$ represents the largest daily value recorded in the study period.

<table>
<thead>
<tr>
<th>Buoy</th>
<th>Lat.</th>
<th>Lon.</th>
<th>No. Of days with $Q_P$ data</th>
<th>No. Of days with $Q_{P,B} &gt; Q_{SH-B}$</th>
<th>$Q_{P,B}$ ($W \ m^{-2}$)</th>
<th>$Q_{P,R}$ ($W \ m^{-2}$)</th>
<th>Largest $Q_{P,B}$</th>
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<tr>
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<tr>
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<td>1186</td>
<td>200</td>
<td>0.5</td>
<td>0.8</td>
<td>16.0</td>
</tr>
</tbody>
</table>

The study period has a total of 5844 days and out of the 40 TAO buoys with $Q_p$ available, 35 have more than 1000 days of data. For the PIRATA buoys that number is 17 out of 20 and the majority of the RAMA buoys have less than 3 years of data (only 6 out of 26). The average $Q_p$ for the TAO, RAMA and PIRATA buoys is 0.6 W m$^{-2}$, 0.6 W m$^{-2}$ and
0.3 W m$^{-2}$, respectively, and the average $Q_P$ from reanalysis for the TAO, RAMA and PIRATA buoy locations is 0.9 W m$^{-2}$, 1.1 W m$^{-2}$ and 0.5 W m$^{-2}$, respectively.

The largest $Q_{P,B}$ is 1.5 W m$^{-2}$ and is located in the Western Pacific and the second largest is 1.4 W m$^{-2}$ also from buoys in the Western Pacific. The locations with larger number of days where $Q_{P,B}$ is greater than $Q_{SH,B}$ are those where the average sensible heat is near zero and the sensible heat of rain is also extremely small.

The largest averaged $Q_P$ values for both reanalysis and buoys are located in the same region, with the largest $Q_{P,R}$ value of 2.0 W m$^{-2}$ for three locations in the West Pacific. For all three data regions (RAMA, PIRATA and TAO), the $Q_P$ from buoys appear smaller. These results can lead us to believe that the calculated $Q_P$ is overestimating the average sensible heat due to rain but it’s important to remember that the SST sensor from the buoy is located below the surface and the temperature is lower than the skin temperature, and that is a possible source of the difference. The rain from the TRMM dataset used for $Q_{P,R}$ calculations can also be larger than the measured rain at the buoy sites.

Figure 19 shows the scatter plot of $Q_{P,B}$ and $Q_{P,R}$ for all TAO buoys. The time series show good agreement with a correlation coefficient of 0.98, with larger values associated with the reanalysis dataset.

The daily and monthly values of $Q_P$ can indicate the areas where it is greater over longer periods of time. However, it does not show the highest values that this term could reach. To do so, it is necessary to go into finer time scales. Figure 20 shows the hourly values from observation at a buoy in the Western Pacific (0°N; 165°E). While the monthly and even daily averages are fairly small, the hourly values show that $Q_P$ can have the same or even higher magnitude than the sensible heat.
Figure 19 - Average of daily sensible heat of rain from all TAO buoys and from Reanalysis. Unit: W m\(^{-2}\)

Figure 20 - Hourly \(Q_P\) and \(Q_{SH}\) from buoy (December 2006 - 0°N; 165°E) Unit: W m\(^{-2}\)
3.2 Summary

The sensible heat of rain varies with the rain rate and with the temperature difference between surface temperature and the wet bulb. Monthly means show small variations in the temperature differences (~3.25 K to 3.7 K) showing a strong relativity between $e Q_p$ climatological values and precipitation. The higher values in the tropical region are associated with the presence of the ITCZ. Large rain events such as the Indian Monsoon is also reflected in the climatology. The mean $Q_p$ for precipitating days is 1.33 W m$^{-2}$, going from 1.0 to 1.7 W m$^{-2}$ on the monthly means. That value can go up to 4 W m$^{-2}$ locally. Considering only the top 5% of the rain events, the $Q_p$ magnitude can exceed 50% of the magnitude of $Q_{SH}$.

Higher relative $Q_p$ climatological values are concentrated in the Warm Pool, Indian Ocean, Bay of Bengal, and western Pacific and seasonally can have high values at the South and East China Seas and at the south east coast of Brazil.

Comparisons with monthly buoy data showed that the reanalysis had larger $Q_p$ values than buoys. It could be associated with the positioning of the SST sensor on the buoy, leading to an underestimation of the buoy values, or due to elevated TRMM rain, which would lead to an overestimation of the reanalysis values. Buoy values can be as high as 48.9 W m$^{-2}$ in a daily average, indicating that hourly values can be even higher. To evaluate the impact of $Q_p$ on the surface temperature, surface energy budget, and atmospheric conditions, it is necessary to use numerical models.
Chapter 4  
Role of $Q_p$ on ocean and atmosphere

4.1 Ocean

The hourly values of $Q_p$ can be much larger than the daily averaged $Q_p$ and might play an important role in the developing and decaying of convection. To further illustrate this, we utilize buoy data (Figure 2) for daily estimates of $Q_p$. The time series for those locations have a lot of missing days, with only 7 buoys (TAO buoys) having data for more than 50% of the days in the study period (1998-2013).

When we look at the hourly data for those buoy locations, the amount of data available drops even more. Figure 21 shows the locations with hourly $Q_p$ data available for at least one hour. Based on the climatology, and on the availability of hourly data, the period chosen to study the role of $Q_p$ over the ocean was December of 2006 to April 2007 and the area is shown in Figure 22.

The simulations follow the parameterization choices presented in section 2.2, with the grid spacing of 30 km (3 h output) for the outer domain and 10 km (1 h output) for the inner domain. Three simulations where conducted from 01 December 2006 to 29 April 2007 and the descriptions are presented below in Table 5. The model top is placed at 50 hPa and the simulations are one-way (the inner domain doesn’t feedback into the outer domain).
Figure 21 - Study focus for simulations (green box – bottom) and buoy locations with at least one hourly $Q_p$ data available (filled blue squares) between 1998-2013 for the Atlantic (top), Indian (middle) and Pacific (bottom) ocean.
Figure 22 - Model domain for ocean simulations. Outer domain (d01; 30 km spatial resolution; 3 h temporal resolution) includes some of the Solomon Islands. The inner domain is centered over a buoy location (d02; 10 km spatial resolution; 1 h temporal resolution).

Table 5 - Description of the coupled ocean-atmosphere simulations.

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Control</td>
<td>Regular WRF model, with activated PWP ocean model</td>
</tr>
<tr>
<td>Sal</td>
<td>Same as Control, but with effects of precipitation on salinity</td>
</tr>
<tr>
<td>SalQp</td>
<td>Same as Sal, but with $Q_P$</td>
</tr>
</tbody>
</table>

In order to isolate the $Q_P$ influence, all the differences, if not specified, are made with respect to the added salinity simulation (Sal).
4.1.1 Results and Discussion

The effect of rain on salinity, and the combined effect of salinity and $Q_p$, were analyzed over the buoy location ($0^\circ$N; 165°E – Buoy1) and the entire inner domain (d02).

For the Buoy1 location, the overall effect of precipitation in salinity and of $Q_p$ was to decrease the accumulated rain (Figure 23). The precipitation decrease can also be observed in the d02 domain area average, although, precipitation increased at the end of the Sal$Q_p$ simulation (April 2007 - Figure 24), what caused an overall increase in precipitation.

![Accumulated rain over 0N;165E – d02](diagram1)

Figure 23 - Accumulated rain for Control (red), Sal (green) and Sal$Q_p$ (blue) for December 2006-April 2007 for Buoy1. Unit: mm.

![Accumulated rain - Area Avg - d02](diagram2)

Figure 24 - Same as Figure 25 but for an average over d02. Unit: mm.
The average rain increase from Sal to SalQp for the Dec 2006 – April 2007 period was 0.007 mm h$^{-1}$, with an average skin temperature increase of 0.018 K with a response in the 2-m temperature of 0.005 K. The majority of the rain and temperature anomalies for the area seem to originate from a large system that was centered at 0.5°N; 168.5°E., almost at the edge of the domain (Figure 25 (a); Figure 26 (a)). That one particular event originated at the end of the 5-month simulation, and had a large impact on the diurnal cycle and can be the focus of a case study in the future. For the purpose of this study, to evaluate the average impact of $Q_P$, and due to the fact that the event is very close to the edge of the domain, the averages will be analyzed considering the period from December 2006 – March 2007.

### 4.1.1.1 Rain and Temperature Diurnal Cycles

For the December 2006 – March 2007 period, the rain had an average decrease of 0.008 mm h$^{-1}$, with an average skin temperature decrease of 0.01 K with a response in the 2-m temperature of -0.006 K. The diurnal cycle of rain for both time period can be analyzed to isolate the differences (Figure 27). The peak of precipitation occurred at 0500 LST and the minimum occurred at 1300 LST. No change was observed in the time of maximum/minimum rain, with hourly output. For the both time periods, the peak of precipitation increased but the increase was much higher in the period that accounts for the large rain event (Figure 27 (a)). The rain had a decrease for both period between 2000 and 0200 LST, but for the Dec 2006 – Mar 2007 the decrease comprises most of night and pre-dawn hours.
Figure 25 - Rain average difference. SalQp – Sal for (a) Dec 2006-Apr 2007 and (b) Dec 2006 – Mar 2007. Wet anomalies (blue) and Dry anomalies (red). Unit: mm h\(^{-1}\).

Figure 26 - Average skin temperature difference. SalQp – Sal (a) Dec 2006-Apr 2007 and (b) Dec 2006 – Mar 2007. Cold anomalies (blue) and warm anomalies (red). Unit: K.
Figure 27 - Diurnal cycle of rain for Control (gray; closed circle), Sal (black; closed circle), SalQp (black; open square) and SalQp−Sal (blue; open circle) from the domain d02 for (a) Dec 2006–Apr 2007 and (b) Dec 2006 – Mar 2007. Unit: mm h⁻¹

The skin temperature peaked at 1400 LST and also decreased with Qₚ (Figure 28). The maximum decrease also peaked at 1400 LST. The model responded as expected with the addition of a freshwater input in salinity (Sal), reducing the salinity of the upper ocean (Figure 29). The effect of Qₚ was to decrease the ocean temperature, leading to less rain and increased salinity.
The salinity increase was observed throughout the day, even during 0500-1100 LST where the precipitation increased. One possible cause for that is, as pointed out by Cronin and McPhaden (1999), due to the nighttime mixing with saltier subsurface water caused by the early day rainfall. The salinity increase is a result of the difference between the decrease in rain and the decrease in evaporation, being rain the dominant of the two terms.

Figure 29 - Same as Figure 28 but for salinity. Unit: PSU (1 PSU = 1 g kg⁻¹).
The evaporation has two peaks (Figure 30), one in the early afternoon, and another in the early morning that is counteracted by the rainfall early in the day, leading to only one salinity peak (~1400 LST).

### 4.1.1.2 Surface Fluxes

As expected, the diurnal cycle of latent heat (Figure 31) has similar shape to the diurnal cycle of evaporation, and has an average decrease of 0.78 W m\(^{-2}\). The peak in sensible heat (Figure 32) occurred at 0600 LST and corresponded to the time of maximum difference between skin temperature and 2-m temperature. The average \(Q_{SH}\) decrease was 0.06 W m\(^{-2}\).

There was very little change in the net radiation (Figure 33) during night time (2000-0600 LST) but there’s an average increase of 0.03W m\(^{-2}\) during daytime. The sign convention applied here is that positive radiation is considered upward (i.e., atmosphere gains heat). The largest increase happens at 1300-1400 LST where a decrease in incoming solar radiation was observed and could be related to the cloud fraction.

The \(Q_{P}\) estimated by the WRF model is in agreement with the results in chap 3 with respect to its dependence on the precipitation. The Rain x \(Q_P\) diagram for Buoy1 is presented in Figure 34, and shows that the \(Q_P\) varied mostly with rain. The correlation between both time series is 0.87.
Figure 31 - Diurnal cycle of latent heat for Control (gray; closed circle), Sal (black; closed circle), SalQp (black; open square) and SalQp-Sal (blue; open circle) from the domain d02. Unit: W m$^{-2}$.

Figure 32 - Same as Figure 31 but for Sensible Heat. Unit: W m$^{-2}$.

Figure 33 - Same as Figure 31 but for Net Radiation. Positive up. Unit: W m$^{-2}$.
Figure 34 - Hourly Rain (mm h⁻¹) X $Q_p$ (W m⁻²) from Buoy1 location for Dec 2006 – Mar 2007.
4.1.2 Summary

The $Q_P$ implementation on the ocean model behaved as expected. The heat loss from precipitation generated a lower skin temperature, on average the temperature was 0.01 K colder, which led to a response in the atmospheric 2-m temperature. The average $Q_P$ for Buoy1 estimated from SalQp from Dec 2006 – Mar 2007 was 1.16 W m$^{-2}$ and was larger than the observational value of 0.66 W m$^{-2}$. That is most likely connected to the overestimation of rain in our model.

The freshwater input of rain in the salinity calculation also behaved as expected, lowering the salt concentrations. As the skin temperature decrease was on average one order of magnitude larger than the decrease in the 2-m air temperature, there was an average decrease in sensible heat. The latent heat also presented lower values. Overall, the $Q_P$ lowered the surface turbulent fluxes.
4.2 Land

Following the analysis of the $Q_p$ over the ocean, this term is also added to the Noah land surface model (Noah LSM) that is coupled to the WRF model to estimate $Q_p$ over land. The term was first introduced in WRF version 3.2 to reduce large residuals in the energy budget during rain events. Its current presence in the code is diagnostic and doesn’t affect the model run. In this study, the term was accounted for during the simulation.

The Maritime Continent (MC) was chosen as a test region to estimate the magnitude of $Q_p$ and its role on the simulation. The MC is a complex area with large mountainous islands surrounded by high sea surface temperature. The time period for this study was April 2009. This particular month was within the period called “Year of Tropical Convection” (YOTC, Moncrieff et al. 2012) that spans from May 2008 to May 2010. The YOTC is a “focus period approach along with an integrated research framework tailored to exploit the vast amounts of existing observations, expanding computational resources, and the development of new, high-resolution modeling frameworks” (Waliser et al. 2012).

During the YOTC there were six cases of relatively strong Madden-Julian oscillation (MJO, Madden and Julian 1971) events. One of three strongest cases of enhanced convections associated with the MJO was during April 2009 (Waliser et al. 2012), where the MJO crosses the MC. The purpose of this study is to analyze the influence of $Q_p$ in the diurnal cycle of rain and temperature over one of the MC islands (Borneo) using high-resolution simulations.

The hypothesis is that when rain temperature is colder (warmer) than the soil, the skin temperature should decrease (increase). As a response, the lowest level air temperature and possibly evaporation would decrease (increase) leading to a decrease (increase) in precipitation. It might also influence the timing of rain over the land.

Two simulations were conducted (Table 6) for April 2009 following the parameterization choices presented in section 2.2. The simulated area includes the entire MC in the outer domain (Figure 35) with the grid spacing of 30 km (3 h output), and the second
domain covering the Borneo island with 10 km resolution (1 h output). For the period when the MJO crosses the MC (~7 April 2009 – 17 April 2009), an additional domain with 3.33 km resolution and 20 minutes output was included in the simulations. The model top is placed at 10 hPa and the inner domain feedback into the outer domains (2-way).

Table 6 - Land simulations description.

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Control</td>
<td>Regular WRF model, with updated SST</td>
</tr>
<tr>
<td>Flx2Qp</td>
<td>Same as Control, but with effects of Q_p in the Noah LSM</td>
</tr>
</tbody>
</table>

Figure 35 - Model domain for land simulations. Outer domain (d01; 30 km spatial resolution; 3 h temporal resolution; 01-30 April 2009) includes the entire Maritime Continent. Second domain (d02; 10 km spatial resolution; 1 h temporal resolution; 01-30 April 2009) is centered at Borneo Island as is the inner most domain (d03; 3.33 km spatial resolution; 20 min temporal resolution; 07-17 April 2009).
4.2.1 Results and Discussion

The time evolution of rain anomaly averaged between 5°N and 5°S of TRMM presented an area of larger precipitation during the month of April 2009, eastward propagating across the MC (Figure 36). The climatology used to create the anomalies is the average April TRMM rain (1998-2012).

![Figure 36 - Time evolution of TRMM rain anomaly (5°N – 5°S). Solid black line indicates the longitudes for Borneo. Unit: mm h⁻¹.](image)

The control simulation captures an eastward propagating feature associated with the MJO but with larger precipitation. Smaller scale rain systems can be observed propagating eastward from 9 – 15 April 2009 over Borneo. The hourly results over the Borneo Island will be presented as a monthly average and as a 10 day average over 7-16 April 2009 time period to focus on the period of the MJO passage.
Figure 37 - Time evolution of anomalous rain for Control simulation (5ºN – 5ºS; d01; 3 h). Solid black line indicates the longitudes for Borneo. Unit: mm h⁻¹.

The time evolution for the Flx2Qp simulation (Figure 38) shows the same general features as the control simulation, with larger differences from 120-145°E after 16 April (Figure 39 – green box).

Over the Borneo Island, the average precipitation for April 2009 was 0.99 mm h⁻¹ for the Control and it decreased to 0.98 mm h⁻¹ for the Flx2Qp simulation (Table 7). The majority of the rain was concentrated over the center-southwest areas of Borneo (Figure 40). If we consider only the 10 day period from 7-16 April, the Control average is 0.98 mm h⁻¹ and it shows a larger decrease to 0.92 mm h⁻¹ (Figure 41 - b).
The Borneo Island has a large mountainous area, NE-SW orientated. For 7-16 April period, the majority of wet anomalies were concentrated on the north side of the mountain and the dry anomalies on the south side.

Figure 38 - Time evolution of rain for Flx2Qp simulation (5°N – 5°S; d01; 3 h). Solid black line indicates the longitudes for Borneo. Unit: mm h\(^{-1}\)
Figure 39 - Time evolution of Flx2Qp minus Control averaged rain difference (5°N – 5°S; d01; 3 h). Dry (red) and wet anomalies (blue). Unit: mm h⁻¹

Table 7 - Averaged rain (mm day⁻¹) for Control and Flx2Qp. Rain difference (mm day⁻¹): Flx2Qp minus Control and decrease from control (%).

<table>
<thead>
<tr>
<th>Averaging Period</th>
<th>Control Rain (mm day⁻¹)</th>
<th>Flx2Qp Rain (mm day⁻¹)</th>
<th>Flx2Qp - Control (mm day⁻¹)</th>
<th>Rain Decrease (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>April 2009</td>
<td>23.9</td>
<td>23.6</td>
<td>-0.3</td>
<td>-1.2</td>
</tr>
<tr>
<td>7-16 April 2009</td>
<td>23.6</td>
<td>22.2</td>
<td>-1.4</td>
<td>-5.9</td>
</tr>
</tbody>
</table>
Figure 40 - Monthly average rain (d02) over Borneo for (a) Control and (b) Flx2Qp simulations. Area average over land on the bottom. Unit: mm h⁻¹.

Figure 41 - 7-16 April 2009 average rain (d02) over Borneo for (a) Control and (b) Flx2Qp simulations. Area average over land on the bottom. Unit: mm h⁻¹.
Our initial hypotheses holds true for the ten day average (7-16 April 2009), where not only the rain decreased (-0.06 mm h^-1) but the average skin temperature also decreased (-0.04 K; Figure 43). The monthly mean, however, had a very small decrease in average of rain (~ -0.01 mm h^-1), and no significant change in skin temperature.
4.2.1.1 Rain and Temperature Diurnal Cycles

The diurnal cycle of precipitation and temperature can help understand the influences of the $Q_p$ in the simulations. The peak of rain didn’t change over the monthly mean, but the amplitude of the peak decreased by $\sim 0.05 \text{ mm h}^{-1}$ (Figure 44).
Figure 44 - Diurnal cycle of rain for Control (black; closed circle), Flx2Q_p (black; open square) and Flx2Q_p-Control (blue; open circle) for April 2009 from the domain d02. Unit: mm h^{-1}.

Figure 45 - Diurnal cycle of skin temperature (K) for Control (black; closed circle), Flx2Q_p (black; open square) and Flx2Q_p-Control rain difference (mm h^{-1}; blue; open circle) for April 2009 from the d02 domain.

The peak difference in skin temperature was positive (~0.04 K) and occurred around 0900 LST (Figure 46). During the night and at the first hours of the day (before sunrise), the 2-m air temperature was higher than the skin temperature. During that period, Q_p can have a warming effect that causes the skin temperature and rain to increase, including the first few hours after sunrise (0800-0900 LST; Figure 45). As soon as the sun rises and starts warming the earth, the skin temperature rises rapidly and becomes warmer than the 2-m temperature.
When that happens, the rain begins to have a cooling effect (1100-1400 LST), which eventually causes a decrease in the peak of rain.

In the monthly average, the difference in skin temperature values are only ~0.035 K but for the 7-14 April average is up to ~ -0.12 K.

![Diurnal Cycle Difference over Borneo](image)

Figure 46 - Diurnal cycle difference (Flx2Qp- Control) for skin (K; green), 2-m temperature (K; pink) and for rain (mm h⁻¹; black) for April 2009 (d02).

During the period of intense rain (7-16 April 2009; 10 day average) the cooling effect is more pronounced. The rain was smaller throughout most of the day (except for the first hours after sunrise 0800-1000 LST) and more stable, with a less pronounced peak (Figure 47). The skin temperature average also decreased (Figure 48). Although the hourly output (d02) showed no difference in the time of maximum rain, the finer scale (d03; 3.33
km; 20 min) presents a smaller peak with the addition of $Q_p$ and it is captured in the 20 minutes output (Figure 49).

Figure 47 - 7-16 April 2009 diurnal cycle of rain for Control (black; closed circle), Flx2$Q_p$ (black; open square) and Flx2$Q_p$-Control rain difference (blue; open circle) for d02. Unit: mm h$^{-1}$.

Figure 48 - 7-16 April 2009 diurnal cycle of skin temperature for Control (black; closed circle), Flx2$Q_p$ (black; open square) and Flx2$Q_p$-Control difference (blue; open circle) for d02. Unit: K.
4.2.1.2 Surface Fluxes

The addition of $Q_P$ over land also has an impact on the surface energy budget. The magnitude of the net radiation (net shortwave plus net longwave or $R_{Net}$) and the latent heat decreased (Figure 50). The latent heat ($Q_{LH}$) depends on the wind speed at the surface and the humidity difference between the surface and air. It seems that the humidity difference in this case was the source of the decrease since the 10 meters wind speed had no significant change. The decrease in rain caused the surface moisture ($Q_s$) to decrease generating lower surface minus air moisture contrast which leads to smaller latent heat values (Figure 51).

In the ocean, we observed a decrease of sensible heat during most of the day. However, the surface temperature had much smaller diurnal variation, and the rain peaked during the night (when surface temperature ($T_s$) > air temperature ($T_a$)) so the $T_s-T_a$ was positive over the entire diurnal cycle.

Over land, the surface temperature was only larger than air temperature over daylight hours, and nighttime rain caused the opposite $Q_P$ effect, or, surface warming. That can explain the difference in average skin temperature between the April 2009 average and the large rain event (7-16 April 2009) average. The average April rain, as we can see in Figure 40 was more uniform than the large rain event and over areas very close the shore, that could...
behave as ocean precipitation (night peak) causing the average warming. The 7-16 April 2009 rain, however, is more concentrated on the inland areas and the lower skin temperature is better observed.

The sensible heat had a small decrease in the first few hours of the day due to nighttime precipitation, but sensible heat increased throughout most of the day (Figure 52). As sensible heat is a function of the difference between surface and air temperature, it showed that, over daylight hours, that gradient increased.

![Figure 50 - Diurnal cycle of the net radiation (RNet) for Control (black; closed circle), Flx2QP (black; open square) and Flx2QP-Control RNet difference (blue; open circle) for April 2009 over d02. Unit: W m⁻².](image-url)
Figure 51 - Diurnal cycle of latent heat for Control (black; closed circle), Flx2Q_p (black; open square) and Flx2Q_p-Control Q_LH difference (blue; open circle) for April 2009 over d02. Unit: W m\(^{-2}\).

Figure 52 - Diurnal cycle of sensible heat for Control (black; closed circle), Flx2Q_p (black; open square) and Flx2Q_p-Control Q_SH difference (blue; open circle) for April 2009 over d02. Unit: W m\(^{-2}\).
4.2.2 Summary

The addition of $Q_p$ over land changed some rain features associated with the MJO over the MC during April 2009. Large rain events trigger larger differences with the addition of $Q_p$. For both, the monthly and 10 day average during the MJO passage, the rain over Borneo decreased but the 10 day average showed a decrease of ~6% in the total rain as the monthly average only has a 1% decrease. With the rain decrease, the $Q_p$ values also decreased.

The diurnal cycle of skin and 2-meter temperatures were influenced by the timing of rain. For the 10-day average, the rain peak decreased. The skin temperature also showed a decrease of up to -0.12 K in the diurnal cycle. The larger skin/air temperature contrast led to higher sensible heat values.

For the April average, shore areas acted like ocean areas, with nighttime/early morning precipitation causing a warming effect. The afternoon rain cooling effect, and consequent rain decrease was also present. It is possible to isolate the warming effect over land averaging only the coastal areas.

When the rain decreases, soil moisture also decreases. This leads to a decrease in latent heat flux of up to 13 Wm$^{-2}$ (14 LST). The net radiation was also affected by the inclusion of $Q_p$. The lower surface temperature caused lower outgoing longwave radiation values, but an upward net radiation anomaly was observed with the inclusion of $Q_p$. Analysis of the net radiation revealed less incoming shortwave radiation. This decrease in shortwave radiation can be connected to the increase in low and mid-level cloud fraction, with a decrease in high level cloud coverage. The precipitation acts do diminish cloud fraction and the decrease of rain lowers that process.
Chapter 5
Summary and Conclusions

Sensible heat due to rain ($Q_P$) is often neglected in global climate models and for estimation of surface energy budget due to its climatological low values that could be comparable to the error from each of the terms on the surface budget. The observations of energy budget components on a global scale, obtained specially from satellites are in constant improvement and relatively small terms could be the cause of uncertainties. The monthly averaged $Q_P$ was calculated from January 1998 to December 2013 in the tropics (30°S to 30°N). The lowest values of $Q_P$ corresponded to the driest months (March and April). The highest $Q_P$ occurred during the rainiest months (October to December) that coincide with the ITCZ movement from the northern to the southern hemisphere. There was also a secondary peak of $Q_P$ (May to July) that corresponded to the Indian Monsoon and the ITCZ.

The maximum $Q_P$ values were around the equator following the annual march of the ITCZ. The values of $Q_P$ were very low, with higher values only going up to 2 W m$^{-2}$. But, this average considered precipitating and non-precipitating days, and was a monthly average. When we considered only precipitating days, some areas with values up to 4 W m$^{-2}$ appeared in the monthly means around 7°N, still following the ITCZ movement.

The monthly average $Q_P$ from precipitating days over the entire period varied between 1.0 W m$^{-2}$ and 1.7 W m$^{-2}$, and it followed closely the monthly average precipitation. The $Q_P$ average for the entire period was 1.33 W m$^{-2}$ with average precipitation of 8.3 mm day$^{-1}$. The largest year-round values were found in the western Pacific Ocean and Warm Pool. Also considering only precipitating days, sensible heat was calculated for the same period, and the largest values of $Q_P$ were found in the Bay of Bengal (22.5% of $Q_{SH}$), and on the Warm Pool (19.3% of $Q_{SH}$). When compared with the $Q_P$ estimated from buoys, the reanalysis had larger values. This is possibly because the SST sensor from the buoy was located below the surface and the temperature is lower than the skin temperature. The other
possibility is due to larger precipitation in reanalysis data compared to buoy precipitation which is thought to underestimate precipitation for heavier precipitating events.

The calculated $Q_P$, based on coarse resolution observation/reanalysis, showed that climatological values of $Q_P$ were small. However, on specific locations, under certain circumstances, $Q_P$ can have the same, or larger magnitudes than surface sensible heat flux (depending on rain rate and to difference between SST and raindrop temperature) and even latent heat flux. Those larger values may play an important role in the developing and decaying of convection. But, the amount of hourly $Q_P$ observations is rare and sporadic. As a result, to evaluate the impact of $Q_P$ over the ocean, the term was introduced into the ocean model coupled to the WRF.

For the tropical ocean, where surface temperature was warmer than the rain droplet temperature, $Q_P$ lowers the surface temperature and reduces the atmospheric convection. From December 2006 to March 2007, the average decrease in rain was 0.008 mm h$^{-1}$ over the simulated domain (4°S - 4°N; 161°E - 169°E), and the decrease in skin temperature was 0.01 K. No change was observed in the time of maximum/minimum rain, with hourly output. The sensible heat over the period also had a decrease of 0.06W m$^{-2}$ and it was related to the larger decrease in surface temperature in comparison with the decrease in 2-m temperature. For the surface net radiation, the incoming shortwave during daylight hours was found to be smaller with $Q_P$ than without $Q_P$ due to the increase in mid and low-level clouds.

The $Q_P$ was also added to the Noah LSM to verify if the same impacts were found over land. The time of precipitation in this case plays an important role since the surface temperature was typically lower than the air temperature during the nighttime. Coastal areas were found to have rain diurnal cycles similar to the ocean, which led to an early morning warming and consequently to an increase in rain. The inland areas with afternoon rain behave similarly to the ocean, with surface cooling and a decrease in rain. During April 7-16 over Borneo, the average decrease in rain was ~0.06 mm h$^{-1}$, with an average skin temperature decrease of 0.04 K.
As in the ocean, there was no observed change in the time of rain, in the hourly output. With a decrease in afternoon rain, the cloud fraction of low and mid-level clouds was higher, diminishing the downward shortwave radiation, which had an impact on the net radiation. The drier surface and lower atmosphere and surface moisture difference led to a decrease in latent heat. For the sensible heat however, there were differences between the ocean and land simulations. In the ocean, the sensible heat decreased, but over land, the predawn coastal precipitation and consequent surface warming increased sensible heat. The afternoon rain over land, however, leads to the sensible heat decrease after dawn.

The combined effect of $Q_P$ over land and ocean was not analyzed in this study, but it may be important, especially for the MC, which has a lot of islands. The cooler ocean surface temperature during daytime would increase the land/sea temperature contrast, which could lead to more rain in the afternoon over land. More rain over land could create larger land surface temperature decrease and possibly, near land areas, increase nighttime rain over the ocean, but that effect would propagate to far over ocean, which should still have, on average, less skin temperature and less rain.

Although climatologically small, the sensible heat of rain can have significant impact on the upper ocean thermodynamics and on the atmosphere over land and ocean, especially during large rain events. The results indicate that model simulations over deep convective regions over the tropics would have significant differences with the inclusion of $Q_P$. 


