Atmosphere in the eye of Sumatran GPS Array : a case study of the Southwest Monsoon season

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2017

Zhang, T. F. (2017). Atmosphere in the eye of Sumatran GPS Array : a case study of the Southwest Monsoon season. Doctoral thesis, Nanyang Technological University, Singapore.

<http://hdl.handle.net/10356/69813>

<https://doi.org/10.32657/10356/69813>

ATMOSPHERE IN THE EYE OF SUMATRAN GPS ARRAY

A CASE STUDY OF THE SOUTHWEST MONSOON SEASON

NANYANG
TECHNOLOGICAL
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SUMATRAN GPS ARRAY
A CASE STUDY OF THE SOUTHWEST MONSOON SEASON

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Asian School of the Environment

A thesis submitted to the Nanyang Technological University
In fulfillment of the requirement for the degree of
Doctor of Philosophy in Earth Sciences

2016

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Acknowledgements

I owe my thanks to a number of people without whom I cannot finish the thesis.

To start with, my supervisor Koh Tieh Yong led me to the interesting world of atmospheric sciences. I am also in gratitude to Emma Hill for opening the door of the GPS technology and encouraging me to overcome the difficulties. I worked in a very friendly environment. The colleagues of the tropical atmosphere group, the Earth Observatory of Singapore and the Asian School of the Environment are all very helpful and resourceful. To name a few, Ricardo Fonseca who spared a lot of time discussing with me, Feng Lujia who helped me to process the raw GPS data, Teo Chee Kiat who taught me to think deep, Yudha Djamil who shared his knowledge and the information with no reservation. The list can go on forever, so I have to stop here. In the end, I wish to thank my dear wife Wang Yuxi, for her listening of my wordy talks, as well as her cooking of the best food of the world.

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Summary

The thesis is a study of the zenith wet delay observed by the Sumatran GPS Array. I focused on the southwest monsoon season, and conducted a thorough analysis to identify the atmospheric processes that impact the water vapor over the GPS network. The roles of the southwest monsoon and the dry air intrusion are revealed. Furthermore, GPS devices also provide a rare chance to directly observe the high-frequency variabilities of the water vapor. The diurnal variability shows an interesting propagation. The coast reaches its diurnal peak first, and then the adjacent sea follows. The semidiurnal variability has a same phase with atmospheric tides. Atmospheric tides are the periodic variation of the air mass, and their semidiurnal variability is strong. Since water vapor is a component of the air mass, atmospheric tides can cause the diurnal variability in GPS observations. In conclusion, the GPS technology is proved to be a valid tool to study the in-depth atmospheric dynamics. With the appropriate approach, the GPS observation can not only quantify the influence of various atmospheric phenomena on the water vapor, but also provide helpful guidance for the numerical simulation and the weather forecast.

Chapter 1. Introduction

1.1 An introduction to GPS meteorology

1.1.1 Global Positioning System

GPS (Global Positioning System) is developed by the United States, and managed by the Joint Program Office (Parkinson 1996). As a space-borne positioning system with a global coverage, the GPS technology is widely used in military and civilian fields. The principal of GPS is very simple. In order to deduce the position of an unknown point, we can measure the distances from the unknown point to several referential points whose positions are already known. The position of the unknown point can then be calculated. For GPS technology, several satellites work as the referential points whose positions can be known by the knowledge of the orbits. The GPS receiving device on the ground can then know its own position by measuring the distances to the GPS satellites. Other countries have developed satellite navigational systems similar to the GPS system. All the systems can be genetically termed as Global Navigation Satellite System (GNSS). In this thesis, the observational data come from GPS rather than other GNSS systems, so I will use the specific term “GPS” in the following parts.

Obviously, the GPS technology has great military potential. The first GPS

satellite was launched in 1978, with the sponsorship of the United States Department of Defense (Czerniak and Reilly 1998). Early versions could only do 2D positioning, but soon GPS devices can be used for the height measurement besides 2D positioning. This technology proved its value in the Gulf War, and helped the marching of the ground army in deserts (Anson and Cummings 1991). In 1983, President Ronald Reagan opened the system for civilian usage, as a response to the tragedy of Korean Air Lines Flight 007 (D'Roza and Bilchev 2003). In that event, the plane was shot down because it mistakenly entered prohibited airspace of Soviet Union. In late 1990s, US government stopped the selective availability, allowing the equivalent accuracy for both military and civilian devices (Clinton and Clinton 2000).

The GPS technology then embraced a boom and quickly penetrated into daily lives. The space-based service became an important aspect of the information revolution. iPhone 3G started to install GPS receivers and provide positioning function (Zandbergen, 2009). Electronic terminals such as tablets and laptops followed the trend and included the GPS receiver as a basic component. The spreading out of the hardware promoted the development of space-based software such as Google Map (Goodchild 2007). The GPS technology is now the foundation of positioning and navigation systems, and is also incorporated into electronic games and social networks.

The popularity of the technology can be attributed to several factors. The GPS network has more than 30 satellites in the orbits and any receiver can capture the signals from at least four satellites. As a result, a receiver can calculate its position with reasonable accuracy at any point of the Earth. Furthermore, the GPS system can work regardless of the weather. The frequencies of the GPS signals are selected within the so-named “atmospheric window” (Kaufman 1989). The signals are able to pass through the atmosphere without being absorbed. At last, the GPS receivers can be produced at a low cost, and do not need any registration. On the other hand, the prosperity of the application feeds the development of the GPS technology itself, closing a cycle of the positive feedback.

The GPS technology has its own weakness in some scenarios, and needs the help of other methods. Because of the reliance on the satellite signals, the GPS receivers cannot work indoor. As such, the navigation system needs other navigation systems to compensate when the vehicle is in the tunnel or the city canyon. The odometer and the gyroscope form the secondary navigating system for the vehicle. Another concern is the speed of the positioning. The calculation of the position can take a while, so the GPS technology cannot fulfill the requirement of some real-time demands. In such a case, a coarse solution

can greatly shorten the calculation time. Mobile devices such as smart phones usually make use Wi-Fi and cellular signals to get a rough position and then quickly narrow down to the accurate position with the GPS information.

The scientific community pioneers the application of the GPS technology. In the late 1980s, the GPS technology was experimented to study the tectonics. USGS (U.S. Geological Survey) used the GPS technology to monitor the seismic activities in California (Hudnut et al. 2001). GPS can provide the valuable information of crustal movements. Besides the positioning function, GPS can also observe the atmosphere. The capacities of GPS in observing the atmosphere originate from the fact that GPS signals can interact with various atmospheric components. For example, the GPS network can detect the ionosphere disturbance by measuring the total electron content (TEC) in the ionosphere (Ho et al. 1996). The GPS measurement of TEC has greatly furthered solar-terrestrial studies and contributes to the forecast of space weather (Jakowski et al. 2002). GPS satellites can also measure temperature, pressure and water vapor profiles by the detection of the occultation of radio signals (Wickert et al. 2001). GPS occultation has advanced the observation of phenomena such as gravity waves (Tsuda et al. 2000), and is proven to be able to improve the weather forecast (Huang et al. 2005).

This thesis is concerned about the measurement of the water vapor in the troposphere. The measurement of water vapor is an important aspect of the GPS meteorology (Bevis et al. 1992). Accurate and frequent measurements of the water vapor can help the in-depth exploration of atmospheric dynamics, as I will show later.

1.1.2 Basics of GPS technology

As I have mentioned, the GPS receivers rely on the knowledge of the distances to the GPS satellites. The basic idea can be easily understood in a simple case. If the distance to a referential point is known to be d , the receiver can only be located on the surface of a sphere with a radius of d . If we have three referential points and know three distances respectively, then the location of the receiver collapses to two points that are the interception points of the sphere surfaces. I can write down the relationship in the algebraic form:

$$\begin{aligned}(x - x_{s1})^2 + (y - y_{s1})^2 + (z - z_{s1})^2 &= d_1^2, \\(x - x_{s2})^2 + (y - y_{s2})^2 + (z - z_{s2})^2 &= d_2^2, \\(x - x_{s3})^2 + (y - y_{s3})^2 + (z - z_{s3})^2 &= d_3^2.\end{aligned}\tag{Equation 1.1}$$

In the equations, x, y, z are the coordinates of the receiver, while x_{sn}, y_{sn}, z_{sn} are the coordinates of the n^{th} satellite. d_n is the distance from the receiver to the n^{th} satellite. It is mathematically simple to solve the set of equations. By knowing that the receiver is “below” the satellites, it is easy to identify only

correct solution out of the two.

It is another issue of GPS to measure the distances. By analyzing the signal from a satellite, the receiver can calculate the distance from the satellite (Hofmann-Wellenhof et al. 2013). The satellite transmits sinusoidal waves, which are called "carrier waves". The satellite broadcasts two carrier waves with different frequencies. L1 wave has a frequency of 1575.42MHz, while L2 has a frequency of 1227.60MHz. Binary codes are loaded on the carrier waves. The binary codes repeat themselves after certain bits. The receiver generates its own binary code based on the equipped clock. By comparing two binary codes, the receiver can know the transmitting time and the distance.

However, the method of binary code can only provide a coarse position. In order to make accurate positioning as required by the scientific research, the receiver can use another way to measure the distance. This method is based on the phase of the carrier wave. By comparing the received carrier phase and the receiver-generated phase, the receiver can track the change of the carrier phase with time. After observing the carrier phase for some time, the receiver is able to determine the distance from the satellite. The method of carrier phase can achieve a precision of 1% of the wavelength (Xu 2007). However, the observation must be accumulated for some time, so the method is not

suitable for applications with high real-time requirements.

In the above introduction, we ignored some potential errors. Both methods are based on the assumption that GPS signals pass through the atmosphere with the speed of light in vacuum. They also assume that the clocks of the satellites and the receivers are perfectly accurate. In real life, many sources can introduce errors to the positioning. For example, the atmosphere can change the speed of the signal, and the clocks could contain errors. We need to account for the possible sources of errors as written in the following equation (Xu 2007):

$$\lambda\Phi = d - c(\delta t_r - \delta t_s) + \lambda N_r^s - \delta_{ion} + \delta_{tropo} + \delta_{tide} + \delta_{mul} + \delta_{rel} + \varepsilon,$$

Equation 1.2

where d is the d_n in Equation 1.1, indicating the real distance from the satellite to the receiver. Φ is the carrier phase measured. λ is the wavelength of the GPS signal. c is the speed of light. δt_r is the error the receiver's clock, and δt_s is the error of the satellite's clock. N_r^s is the ambiguity number that can be determined by the accumulation of GPS data. The other terms represent the necessary corrections. δ_{ion} and δ_{tropo} are the ionosphere and troposphere effects. δ_{tide} is the effect of the Earth tide and ocean loading tide. δ_{mul} arises due to the effect of multiple paths of the signal. δ_{rel} accounts for the effect of relativity. ε is the remaining error. I am mainly interested in the

tropospheric delay.

1.1.3 Zenith wet delay

The tropospheric effect in Equation 1.2 can be split into a hydrostatic (dry) and a wet delay component. The delay is measured as a distance that is equal to the product of the time delay and the speed of light. In a simplified treatment, the delays are taken as irrelevant with the horizontal direction. With a mapping function, the time delays can be converted to the zenith direction. The resultant delays are called the zenith hydrostatic delay (ZHD) and the zenith wet delay (ZWD). ZHD is proportional to the surface pressure, and is relatively stable. ZWD is mainly controlled by the precipitable water vapor (PWV), and is much more volatile. The GPS data from the SuGAR network are processed to be hourly with the procedure of Feng et al. (2015).

According to Bevis et al. (1992), ZWD can be converted to PWV with the following equation:

$$\frac{ZWD}{PWV} = \frac{\rho}{\kappa} = 10^{-6} \left(\frac{k_3}{T_m} + k_2' \right) \rho R_v \quad \text{Equation 1.3}$$

In the above equation, ρ is the density of the water, and R_v is the specific gas constant for water vapor. k_3 and k_2' are two constants. The quantity T_m is called weighted “mean temperature” of the atmosphere, defined by

$$T_m = \frac{\int (P_v/T) dz}{\int (P_v/T^2) dz} \quad \text{Equation 1.4}$$

Where P_v is the specific humidity, and T is the temperature.

PWV is a useful quantity in the meteorology. It measures the total water vapor in the vertical air column. With the hydrostatic balance, PWV can be expressed by the following integral in the pressure coordinate:

$$PWV = \int_0^{P_{surface}} q_v dp \quad \text{Equation 1.5}$$

$P_{surface}$ is the surface pressure, and q_v is the specific humidity. Since low-level atmosphere contains the major part of the water vapor, so it contributes more water vapor to PWV compared with high-level atmosphere. However, the change of PWV does not necessarily happen in the low level. The changing process is controlled by the moisture flux and the source and sink. By changing the moisture flux or the source and sink, various atmospheric processes can modify the PWV, such as the transportation by the wind, the evaporation and the condensation. In the later part of the thesis, I will analyze the atmospheric processes that modify the PWV and finally impact the GPS-observed ZWD.

1.1.4 The SuGAR network

In this thesis, I make use of the GPS data from Sumatran GPS Array (SuGAR) network. It is a dense GPS network in the deep tropics. The network was

originally designed to monitor the tectonic activities of the Sumatran plate boundary (Feng et al. 2015). It was launched in 2002 and since then has expanded considerably (Figure 1.1). Many stations are located along the west coast of Sumatra or on the offshore islands to the west of Sumatra. The eastern part of Sumatra is also covered, and one station (NTUS) is in Singapore.

Figure 1.1 The map of the GPS stations of the SuGAR network. The dots are the locations of the stations. The color indicates the installation time of the station (Feng et al. 2015)

Sumatra is the second largest island in Southeast Asia, and the Equator cuts through the middle of the island. As an island in the deep tropics, the climate near Sumatra is usually hot and humid. Besides the typical features of the

tropical climate, Sumatra's regional climate is somehow uniquely complex. The island lies in the path of the Asian-Australian monsoon (Chang 2004), and thus experiences the seasonal reversal of wind directions. The land-sea contrast also nurtures strong land-sea breeze. What is more, the Pacific and the Indian oceans can both influence the climate of Sumatra due to their proximity. In all, Sumatra has a tropical climate with its own complexity. Although the SuGAR network was designed for tectonics, i.e., d in Equation 1.2, it can also reveal the impact of different atmospheric processes by term δ_{tropo} in Equation 1.2.

The SuGAR network has accumulated 13-year data. This is a wealth of observation. My study will pick the southwest monsoon season of 2008 as the targeting period. As a result, this will be a case study rather than a thorough examination of the full dataset. I will further justify my selection of the year in section 3.2.1. However, the dataset actually contains much more information than that this thesis will explore. Long history of data allows researchers to find interannual variabilities. Atmospheric processes of different monsoonal seasons are also worth digging. I think this thesis is just a starting point, and the future work can reveal more about the tropical climate with the observation of the SuGAR network. In the next part, I will examine the basics of the tropical climate.

1.2 An introduction to tropical climate

1.2.1 Tropical Circulations

In nature, atmospheric and oceanic movements are driven by solar radiation. The Earth receives solar radiation unevenly due to the curvature of the spherical surface. The solar light is less perpendicular to the ground when it comes to polar region. Therefore, the unit area of the tropical region receives more energy than that of high latitudes. Because of the abundant supply of solar energy in the tropics, the temperature is high and the evaporation is active. The low-level air is usually hot and humid near the Equator. On the other hand, the Earth emits long wave radiation into the space to balance the incoming solar radiation. The outward long wave radiation is quite even among the latitudes. As a result, the tropical region would have net surplus in the budget of radiation, and the polar region would have net deficit. From the macro view of the Earth, the movements of air and seawater transport the heat polewards to prevent the tropical region from being too hot and the polar region too cold.

In detail, the transportation can take various forms, such as advection or eddies. For the atmosphere, the movement of the air is arranged into large-scale atmospheric circulation. It consists of circular cells along the longitudinal direction. Hadley cells are near the Equator, the polar cells around

the Poles, and the Ferrel cells in between. Strong solar heating near the tropics drives ascending branches of Hadley cells. During the process of rising, the water vapor condenses at the high level and precipitates. The hot and dry air at the high level would cool down and descend when it moves away from the Equator. Conversely, in polar cells, the air near the Poles would descend due to the lack of solar heating, and rise when the air moves away from the Poles.

The Ferrel cells are located in the mid-latitude. Unlike the Hadley and polar cells that had clear ascending and descending branches, the Ferrel cells only indicate the average motion of the air in the mid-latitude. In fact, the Ferrel cells are driven by eddies. These eddies can transport heat from low latitudes to high latitudes. The formation and the location of the Ferrel cells depend on the factors such as solar heating, rotational speed of the Earth, and the size of the Earth (Vallis 2006).

Earth's rotation also contributes to the difference between the tropical climate and the extratropical climate. The vertical component of the Earth's rotation is large in middle and high latitudes, and is small in the tropics. The vertical component of the Earth's rotation causes the so-called Coriolis effect that is responsible for the large-scale westerly system in the mid-latitude. For long-lasting weather systems in the mid-latitude, the Coriolis effect dominates

the evolution. Although the non-linear terms are still important, the strong Coriolis effect organizes the large-scale weather system into a westerly system. Since the Coriolis effect is linear, the resultant westerly system is less chaotic and more predictable. The tropics do not possess the same luck because the Coriolis effect is weak here. As a result, the middle-scale and small-scale weather systems dominate the regional climate, and they are also more chaotic and unpredictable (Hastenrath 2012).

Although the Coriolis effect is not strong enough to create the large-scale weather system, but it still influence the tropical climate in other ways. One thing is the trade wind. The trade wind is easterly wind in the tropics. The Coriolis effect causes the easterly trade wind. When then low-level wind moves from the subtropical high to the equatorial low, the Coriolis force can deflect the equatorward movement of the air to westward movement. The trade wind helped the navigation in the Age of Discovery and promoted the trade by sailing. For the tropical climate, the trade wind can drive the warm water on the sea surface to move westward. The sea surface temperature (SST) thus varies in the zonal direction. The western parts of ocean basins are warmer than the eastern parts. Correspondingly, the air rises in the west and sinks in the east, forming the Walker circulation (Julian and Chervin 1978). Another phenomenon related with the Coriolis effect is the equatorial Rossby wave,

which can be attributed to the longitudinal variation of the Coriolis force (Lindzen, 1967).

For the tropical regions close to the continents, they are also influenced by the monsoon. The monsoon is caused by the different heat capacities of the land and the sea. The Asian-Australian monsoon can modify the weather of Southeast Asia, India, Australia and the east coast of Africa. Although not as strong as the Asian-Australian monsoon, the west coast of Africa and the northeast coast of South America are also influenced by the seasonally varying winds. The monsoon can perturb the Inter-Tropical Convergence Zone (ITCZ) and encourage the inter-hemispheric exchange of air. The involvement of the tropical region in the monsoonal system also creates a mechanism for the tropical-extratropical interaction. It is worth noting that although the monsoon reverses the direction during different seasons, the strength of the monsoon and its impact on the atmosphere is not symmetric. It is necessary to discern the seasons and study different seasons respectively.

Monsoon, Hadley and Walker circulations coexist and interact with each other (Saha 2009). They are the backbone of the tropical climate. For the region of the SuGAR network, all of them are important. However, the dynamics is oversimplified because I omitted temporal variabilities except the stable

seasonal change. In the following parts, I will introduce interannual, intraseasonal, diurnal and semi-diurnal variabilities. These variabilities can effectively adjust the tropical atmosphere.

1.2.2 Interannual variability

Interannual variability plays a significant role in the tropical climate. Sumatra lies between the Pacific and the Indian oceans. ENSO (El Niño Southern Oscillation) is the interannual oscillation that happens in the Pacific, while IOD (Indian Ocean Dipole) is the interannual oscillation mainly confined in the Indian Ocean. Both phenomena influence Sumatra because of their geographical proximity.

The phenomenon of ENSO was discovered in the colonial age. Bjerknes (1966) linked ENSO with the variation of Walker circulation, integrating the previous studies of El Niño and Southern Oscillation. ENSO is accompanied by the disturbance of the sea surface temperature (Rasmusson and Carpenter 1982). As I already mentioned, the trade wind can trap the warm sea surface water in the western part of the Pacific. With the air-sea interaction and the evolution of the oceanic waves, the warm seawater might flow back to the eastern Pacific every 2 to 7 years, and encourage the convections in the middle to eastern part of the Pacific. Accordingly, the SST of the western Pacific drops and

discourages the convection. As a result, Indonesia and Australia might experience severe droughts while California and Peru can face a sequence of storms. The influence of El Niño can surpass the tropical Pacific through the interaction with the monsoon, raising the possibility of droughts in Indian and northern China (Kumar et al. 2006, Lau and Weng 2001). La Niña is the opposite of El Niño, with more warm water piled up in western Pacific.

Compared with ENSO, the discovery of the Indian Ocean Dipole (Saji et al. 1999) is rather recent. As the name indicates, IOD happens in the Indian Ocean. The SST of the western Indian Ocean oscillates oppositely with that of the eastern Indian Ocean, forming a dipole mode. A year can be classified as “positive”, “negative” or “neutral” IOD phases. The positive phase would witness a positive SST anomaly in the western Indian Ocean and a negative SST anomaly in the eastern Indian Ocean. The IOD is similar with ENSO in that it is also a coupled change of the air and the sea. Easterly wind anomaly accompanies the accumulation of the warm seawater in the western Indian Ocean during the positive phase. Furthermore, the high SST can encourage the rainfall, while the low SST suppresses rainfall in the surrounding area.

The interannual variability is an interesting topic, but it is out of the scope of the thesis. We mainly use them as a standard to select the year for the study. In

order to avoid the influence of the strong ENSO or IOD events, I select a neutral year for both ENSO and IOD as the studying period, which is 2008.

1.2.3 Intraseasonal variability

Intraseasonal variability is one of the main concerns of the thesis. This topic is still under active research. Generally speaking, the intraseasonal variability includes any atmospheric variation that has a period shorter than a season and longer than a day. In the tropics, the intraseasonal variability could cause sharp changes within a season. For example, the Indian Monsoon can become very weak for weeks, i.e. the so-called “monsoon break” (Annamalai and Slingo 2001). During the monsoon break, the monsoonal rainfall would dwindle, which may lead to severe droughts in India. The study of the intraseasonal variability possesses meteorological and social importance.

Two phenomena are worth noticing. One is the Madden-Julian Oscillation (MJO; Madden and Julian 1971) that has a period of 30-60 days, and the other is the Boreal Summer Intraseasonal Oscillation (BSISO; Wang and Xie 1997). MJO is the strongest intraseasonal variability. It is active during the boreal winter. MJO is an eastward traveling pattern, characterized by a cluster of active convections that originates in the Indian Ocean. Low-level convergence and high level divergence are coupled with the enhanced convections. A dry phase

with suppressed convections follows the active phase. The eastward movement is global, although the convective features might disappear from the eastern Pacific. MJO can strongly influence the tropics, as well as some subtropical region such as North America. The discovery and the understanding of MJO have facilitated the weather forecast in the tropics, although the mechanism of MJO is still under debate.

In this study, we focus on the boreal summer rather than the boreal winter. In the boreal summer, BSISO replaces MJO to be the strongest intraseasonal variability. BSISO also includes a region of active convection that propagates eastward. A slight northward movement happens together with the eastward movement (Jiang et al. 2004). BSISO is closely related with the monsoon, especially the active and break phases of the monsoon, so BSISO can be indexed by the strength of the monsoon. As a phenomenon established much later than MJO, many aspects of BSISO needs further investigations. In this thesis, I focus on the southwest monsoon season when BSISO plays a larger role. In Chapter 3, I will show some characteristics of BSISO observed by the GPS network.

Another aspect of the intraseasonal variability receives less attention, which is the influence of the wave train of the midlatitude. Relevant discussions can be

seen in previous studies (Rodwell 1997), but usually less discussed in the scope of the intraseasonal variability. Actually, extratropical Rossby waves can impact the tropical climate and excite the intraseasonal variability of the tropical atmosphere. Rodwell (1997) has argued that the intrusion from the Southern Hemisphere drives the monsoon break. In Chapter 3, I will show that the air intrusion from the Southern Hemisphere plays a significant role for the GPS-observed intraseasonal variability. The influence of the air intrusion can be distinguished from the monsoonal influence, because it impacts the GPS network in its own special way. Whether the intrusion evolves independently, or it interacts with the monsoon, will be the question of future work.

1.2.4 Diurnal and semidiurnal variability

As another outcome of the weak Coriolis effect in the tropics, the diurnal and semidiurnal cycles of the atmosphere are prominent. The diurnal and semidiurnal cycles have periods of 24 hours and 12 hours respectively. They are caused by the daily variation of the solar heating and manifest themselves in various physical quantities, such as the surface pressure and surface temperature. However, the diurnal and semidiurnal variations are usually covered up by the large-scale disturbances in midlatitude. When it comes to the tropics, the diurnal and semidiurnal cycles become obvious because of the absence of large-scale disturbances.

The semidiurnal variability can be linked with the atmospheric tide, which is the regular change of total air mass driven by solar heating (Lindzen and Chapman 1969). Due to the hydrostatic balance, the surface pressure can measure the air mass. In the tropics, the semidiurnal component of the atmospheric tide is stronger than the diurnal component (Dai and Wang 1999). In Chapter 4, I will reveal the influence of atmospheric tides on the GPS observation of ZWD. Since the atmospheric tide also has a diurnal component, the paradigm of the “diurnal cycle” mixes with that of the atmospheric tide. Here, I am concerned with a diurnal cycle coupled with convective rainfall (Yang and Slingo 2001). The diurnal cycle of tropical convections is especially important for the tropical climate, because a large amount of convective rainfall happens in such a form. In Chapter 4, I will also discuss the aspect of water vapor in the diurnal cycle as reflected by the GPS observation.

1.3 Thesis Content and Structure

This thesis makes a case study of the GPS observation during the southwest monsoon season of 2008. I did not have enough time to extend the work for other seasons or other years during my PhD. However, by putting together different statistical and computational tools, I have explored the way further ahead. I hope that other researchers can be interested in the SuGAR

observation, and test the conclusions of the thesis based on the whole dataset.

During this season, a through analysis is conducted to understand the intraseasonal variability, as well as the diurnal and semidiurnal variabilities. We find the most important atmospheric processes that are influencing the GPS network. Furthermore, we take advantage of the high-frequency GPS observations, and compare them with the numerical model. The numerical model successfully simulates the aspect of the diurnal variability, but fails in simulating the semidiurnal variability. As a result, our model run cannot well simulate some phenomena that have a strong semidiurnal component, such as atmospheric tides. In conclusion, the study provides concrete proof for several dynamical studies and also quantifies the influence of different atmospheric processes. It can guide the future practice of the numerical simulation, and serve the weather forecast.

The thesis includes five chapters. I have concluded Chapter 1 that is the introduction to the GPS meteorology and the tropical climate. Chapter 2 specifies sources of data, as well as the quality control of GPS data and the setup of the numerical model. Chapter 3 starts the discussion of the atmospheric dynamics reflected by the GPS observation. This chapter handles the intraseasonal variability, and Chapter 4 discusses the semidiurnal and

diurnal variabilities. Chapter 5 concludes the thesis, and makes proposal for the future.

1.4 Reference

- Annamalai, H., & Slingo, J. M. (2001). Active/break cycles: Diagnosis of the intraseasonal variability of the Asian summer monsoon. *Climate Dynamics*, 18(1-2), 85-102.
- Anson, P., & Cummings, D. (1991). The First Space War: The Contribution of Satellites to the Gulf War. *The RUSI Journal*, 136(4), 45-53.
- Bevis, M., Businger, S., Herring, T., Rocken, C., Anthes, R., & Ware, R. (1992). GPS meteorology- Remote sensing of atmospheric water vapor using the Global Positioning System. *Journal of Geophysical Research*, 97(D14), 15787-15801.
- Bjerknes, J. (1966). A possible response of the atmospheric Hadley circulation to equatorial anomalies of ocean temperature. *Tellus*, 18(4), 820-829.
- Chang, C. P. (2004). *East Asian Monsoon*. World Scientific.
- Lindzen, R. S., & Chapman, S. (1969). Atmospheric tides. *Space Science Reviews*, 10(1), 3-188.
- Clinton, B., & Clinton, P. (2000). Improving the civilian global positioning system (GPS). *Media release, Office of the Press Secretary, The White House*, 1.

- Czerniak, R. J., & Reilly, J. P. (1998). *Applications of GPS for surveying and other positioning needs in departments of transportation*. Transportation Research Board.
- Dai, A., & Wang, J. (1999). Diurnal and semidiurnal tides in global surface pressure fields. *Journal of the Atmospheric Sciences*, 56(22), 3874-3891.
- D'Roza, T., & Bilchev, G. (2003). An overview of location-based services. *BT Technology Journal*, 21(1), 20-27.
- Feng, L., Hill, E. M., Banerjee, P., Hermawan, I., Tsang, L. L., Natawidjaja, D. H., ... & Sieh, K. (2015). A unified GPS-based earthquake catalog for the Sumatran plate boundary between 2002 and 2013. *Journal of Geophysical Research: Solid Earth*.
- Goodchild, M. F. (2007). Citizens as sensors: the world of volunteered geography. *GeoJournal*, 69(4), 211-221.
- Hastenrath, S. (2012). *Climate dynamics of the tropics*. Springer Science & Business Media.
- Ho, C. M., Mannucci, A. J., Lindqwister, U. J., Pi, X., & Tsurutani, B. T. (1996). Global ionosphere perturbations monitored by the worldwide GPS network. *Geophysical Research Letters*, 23(22), 3219-3222.
- Hofmann-Wellenhof, B., Lichtenegger, H., & Collins, J. (2013). *Global positioning system: theory and practice*. Springer Science & Business Media.

- Huang, C. Y., Kuo, Y. H., Chen, S. H., & Vandenberghe, F. (2005). Improvements in typhoon forecasts with assimilated GPS occultation refractivity. *Weather and Forecasting*, 20(6), 931-953.
- Hudnut, K. W., Bock, Y., Galetzka, J. E., Webb, F. H., & Young, W. H. (2001, March). The southern California integrated GPS network (SCIGN). In *The 10th FIG International Symposium on Deformation Measurements* (pp. 19-22). Orange California, USA.
- Jakowski, N., Heise, S., Wehrenpfennig, A., Schlüter, S., & Reimer, R. (2002). GPS/GLONASS-based TEC measurements as a contributor for space weather forecast. *Journal of Atmospheric and Solar-Terrestrial Physics*, 64(5), 729-735.
- Jiang, X., Li, T., & Wang, B. (2004). Structures and mechanisms of the northward propagating boreal summer intraseasonal oscillation. *Journal of climate*, 17(5), 1022-1039.
- Julian, P. R., & Chervin, R. M. (1978). A study of the Southern Oscillation and Walker Circulation phenomenon. *Monthly Weather Review*, 106(10), 1433-1451.
- Kaufman, Y. J. (1989). The atmospheric effect on remote sensing and its correction. *Theory and applications of optical remote sensing*, 336-428.

- Kumar, K. K., Rajagopalan, B., Hoerling, M., Bates, G., & Cane, M. (2006). Unraveling the mystery of Indian monsoon failure during El Niño. *Science*, 314(5796), 115-119.
- Lau, K. M., & Weng, H. (2001). Coherent modes of global SST and summer rainfall over China: An assessment of the regional impacts of the 1997-98 El Niño. *Journal of Climate*, 14(6), 1294-1308.
- Lindzen, R. D. (1967). Planetary waves on beta-planes. *Monthly Weather Review*, 95(7), 441-451.
- Madden, R. A., & Julian, P. R. (1971). Detection of a 40-50 day oscillation in the zonal wind in the tropical Pacific. *Journal of the Atmospheric Sciences*, 28(5), 702-708.
- Parkinson, B. W. (1996). *Progress in Astronautics and Aeronautics: Global Positioning System: Theory and Applications*. Aiaa.
- Rasmusson, E. M., & Carpenter, T. H. (1982). Variations in tropical sea surface temperature and surface wind fields associated with the Southern Oscillation/El Niño. *Monthly Weather Review*, 110(5), 354-384.
- Rodwell, M. J. (1997). Breaks in the Asian monsoon: The influence of Southern Hemisphere weather systems. *Journal of the Atmospheric Sciences*, 54(22), 2597-2611.
- Saha, K. (2009). *Tropical circulation systems and monsoons*. Springer Science & Business Media.

- Saji, N. H., Goswami, B. N., Vinayachandran, P. N., & Yamagata, T. (1999). A dipole mode in the tropical Indian Ocean. *Nature*, 401(6751), 360-363.
- Tsuda, T., Nishida, M., Rocken, C., & Ware, R. H. (2000). A global morphology of gravity wave activity in the stratosphere revealed by the GPS occultation data (GPS/MET). *Journal of Geophysical Research: Atmospheres* (1984–2012), 105(D6), 7257-7273.
- Vallis, G. K. (2006). *Atmospheric and oceanic fluid dynamics: fundamentals and large-scale circulation*. Cambridge University Press.
- Wang, B., & Xie, X. (1997). A model for the boreal summer intraseasonal oscillation. *Journal of the Atmospheric Sciences*, 54(1), 72-86.
- Wickert, J., Reigber, C., Beyerle, G., König, R., Marquardt, C., Schmidt, T., Grunwaldt, L., Galas, R., Meehan, T.K., Melbourne, W.G. & Hocke, K., (2001). Atmosphere sounding by GPS radio occultation: First results from CHAMP. *Geophysical Research Letters*, 28(17), 3263-3266.
- Xu, G. (2007). *GPS: theory, algorithms and applications*. Springer Science & Business Media.
- Yang, G. Y., & Slingo, J. (2001). The diurnal cycle in the tropics. *Monthly Weather Review*, 129(4), 784-801.
- Zandbergen, P. A. (2009). Accuracy of iPhone locations: A comparison of assisted GPS, WiFi and cellular positioning. *Transactions in GIS*, 13(s1), 5-25.

Chapter 2. Source of Data

2.1 ZWD from the SuGAR network

In Chapter 1, I briefly introduced GPS technology and the SuGAR network. I selected 2008 as the targeted period of the study, because it is both ENSO and IOD neutral. For the year 2008, there are 30 stations operating in the SuGAR network. However, the quality of the data varies. The quality of the data from certain stations is not good enough. In the end, I only keep 17 stations out of the 30 stations, and exclude the rest (Figure 2.1). The quality control is conducted over 2008, in consideration that the study of intraseasonal variability needs a longer period than the targeted monsoonal season.

Figure 2.1 Valid stations and excluded stations.

The exclusion of 13 stations is based on the subjective criterion, with the assistance of some statistics. The problems of the bad stations are erroneous values and data gaps. It is easy to identify the erroneous values of ZWD. ZWD is mainly determined by PWV. PWV should be relatively stable for a station, since it is the integral of all the water vapor in the vertical direction. PWV of CFSR reanalysis is used as a reference to find the erroneous values. Moreover, the SuGAR network is located in a small geographical area, so PWV should not change much except the influence of the terrain height. This understanding forms the basis of identifying the erroneous value. If the value deviates from the mean of the network too much, it is flagged as erroneous. Figure 2.2 shows an example of erroneous station. When a significant part of one station's observation is erroneous, I would reject the station.

Figure 2.2 Example of an erroneous station. The figure is the time series of ZWD. The beginning part of January and February is erroneously large. This station is thus excluded.

Another issue is the data gap. By right, the GPS receiver can work continuously for long time. However, the data gap might happen due to the failure of the GPS receiver or the shutdown of the device. It is easy to identify the data gaps if the time series of the whole season is plotted, such as the time series of the station in Figure 2.3.

Figure 2.3 Example of the data gap. The figure is the time series of ZWD. The period from February to April is missing.

In addition, the data gap might also take place because of the issue of undersampling. Since I am going to investigate the high-frequency variabilities such as semidiurnal and diurnal variations, so I hope that the daily observations are frequent enough to resolve the semidiurnal variability. This means at least 3-hourly output is needed. Although the raw data are processed to be hourly, some of the stations might still have gaps during the day. I count the number of the observational records for each day. If the station does not cover a large portion of the days with high-frequency observation, it is excluded from my study. Figure 2.4 shows an example of undersampling.

Figure 2.4 The count of the records for each day. The horizontal axis is the day from the beginning of the year 2008, and the vertical axis is the count of the observational records. By right, 24 records should exist for one day.

The exclusion of the 13 stations might seem strict. When studying diurnal and semidiurnal variabilities, I do not need time series longer than the season. In this case, I can possibly loosen the criteria and include more stations than in the study of intraseasonal variabilities. However, if we do this, the number of stations will change for Chapter 3 and Chapter 4, and our conclusion will suffer this inconsistency. On the other hand, Figure 2.1 shows that the stations left are good enough in covering the area of the SuGAR network, capable of

measuring the spatial variation of the ZWD signal.

After the exclusion of bad stations, the occasional data gaps are filled with method of linear interpolation. I interpolate the neighboring points to get the data in between. After the processing, I retrieve an hourly time series for each GPS station.

2.2 CFSR Reanalysis data

The GPS-observed ZWD is not enough to diagnose the complicated atmospheric processes, because the SuGAR network is limited to a small area and cannot observe other atmospheric quantities such as wind, temperature and pressure. We need other datasets besides the GPS observation. I mainly rely on the reanalysis data to provide the detailed information about the condition of the atmosphere. In order to get the reanalysis data, a global model is consistently run with the same set of parameters for a period of time. During the model run, the historical observations are assimilated into the model. The reanalysis data is thus a mixture of simulation and observation. The model can output various atmospheric fields on the model grid, providing a 3D coverage of the atmosphere. In this study, I would make use of the Climate Forecast System Reanalysis (CFSR) dataset (Saha et al. 2010). The details of the dataset will be discussed in Chapter 3.

2.3 WRF Model

CFSR can show the atmospheric condition. However, the CFSR reanalysis only outputs every 6 hour, so it is not enough to resolve the diurnal and semidiurnal variabilities. As a result, we run the Weather Research and Forecasting model (WRF; Skamarock et al. 2005) to dynamically downscale the CFSR reanalysis data to hourly output.

WRF model is a mesoscale Numerical Weather Prediction (NWP) model widely used for the simulation of regional climate. It is a non-hydrostatic model (Skamarock and Klemp 2008), and inherits from the Fifth-Generation Penn State/NCAR Mesoscale Model (Grell et al. 1994). The software system is purposely designed to be modular so that it can easily include other functional modules such as data assimilation and new physical schemes. WRF is now used in the operation of the National Center for Environmental Prediction (NCEP) and other forecasting centers. In order to downscale, CFSR is used to provide the boundary conditions and the initial condition. Since no universal guideline exists for the configuration of the model, it is worthwhile to pay special attention to the setup of WRF model for the region, where the model is less tested compared to the mid-latitude.

2.3.1 Vertical Levels

The positions of the vertical levels are generated with a new algorithm. WRF model can itself generate the vertical levels with its own version of the algorithm. Similar to the original version, the new algorithm arranges the vertical levels with intervals that change smoothly. Different to the original approach, the new algorithm resolves two regions more readily, i.e., near the surface and around the tropopause, by placing more levels in both regions (Figure 2.5). The two regions are dynamically interesting for the study of atmospheric dynamics. The transportation and mixing of heat, momentum and water vapor is intense near the surface. The atmospheric dynamics near the tropopause impacts several important atmospheric processes, from the large-scale general circulation, to the small-scale convection. In this way, the computational resource is more invested into the regions that deserve it.

Figure 2.5 Vertical Levels. Black lines indicate the levels provided by WRF model. Blue lines indicate the levels that I use.

2.3.2 Boundary Condition of the Model Top

In reality, the atmosphere stretches into the space. However, most NWP models are designed only to simulate the lower part of the atmosphere. It is a waste of computational power to include the high-level atmosphere in the model, if the main focus is the lower part of the atmosphere such as the troposphere. It is also not practical to design a NWP model for the whole atmosphere, because the dynamics change dramatically when it comes to the high level.

As a solution, the numerical model artificially imposes the top boundary at a certain height, and thus truncates the atmosphere. This artificial top boundary brings some unwanted effects, such as the reflection of gravity waves. By employing the radiative boundary condition, the model allows the gravity wave absorbed at the model top, so that the artificial reflection can be avoided. As a result, the interior of the model would not "feel" the existence of the top boundary, and the model can provide better simulation results. Among all the choices of the boundary conditions in WRF, the w-Rayleigh damping method (Klemp et al. 2008) can effectively eliminate the spurious reflection of the gravity wave. The w-Rayleigh damping is applied to all layers within 5000 meters from the model top. The damping method differs from the traditional Rayleigh damping in that it only modifies the vertical velocity in the damping layer. After damping the vertical velocity to zero, the numerical model modifies other physical fields accordingly. In this way, the w-Rayleigh damping can preserve the dynamical consistency. In our model, the damping has a time scale of 5 seconds that is considered to be strong.

2.3.3 Physical Schemes

The numerical model cannot explicitly resolve some physical processes, such as radiation, turbulence, condensation and evaporation. These physical

processes are so important that we cannot simply ignore them. Luckily, the effects of these processes can be represented by the physical schemes. Details regarding the choice of the physical scheme are discussed here.

The solar radiation drives the atmospheric dynamics. WRF model does not simulate the solar activities directly, so it has to make use of the radiative scheme. The radiative scheme defines the shortwave and longwave radiations. The Rapid Radiative Transfer Model for GCMs (RRTMG; Iacono et al. 2008) scheme is chosen. It uses correlated-k method for the radiative transfer and includes 16 spectral bands to represent major molecular absorption. Moreover, the Monte Carlo method is used to quantify the effect of sub-grid clouds. In the test, a frequent update of the radiation is necessary, in order to accurately simulate the phenomena with a shorter temporal scale, such as the diurnal cycle. If the interval of calling the radiation scheme is too long, the model risks generating unrealistic results around the sunrise and the sunset, when the real radiation changes rather rapidly. As a result, we call RRTMG scheme every 5 minutes to update the radiative profile.

The horizontal resolution of the model is not high enough to resolve the atmospheric convection directly. We thus use cumulus scheme to account for the effect of atmospheric convections. Atmospheric convections remove the

hydrostatic instability in the atmosphere and change the profiles of the temperature and the humidity. In some cases, they can also produce convective rainfall. Previous studies have shown that atmospheric convections are intense around Sumatra, so the choice of the cumulus scheme can influence the simulation result. We use the modified Betts-Miller- Janjić scheme (BMJ; Betts 1986, Betts and Miller 1986, Janjić 1994). The modification is summarized by Fonseca et al. (2015). The original BMJ tries to account for the effect of the atmospheric convections. BMJ scheme belongs to the category of adjustment schemes that are distinguished from the mass flux schemes (Arakawa and Schubert 1974; Kain and Fritsch 1990) that try to simulate the detailed processes of convections. The general idea of BMJ scheme is to adjust the profiles of the temperature and the specific humidity towards a mixing line that is typical for tropical convections. During the adjustment, the enthalpy is preserved and the convective rainfall is produced if there is surplus water vapor after the adjustment. If the scheme does not produce the convective rainfall, the scheme switches another branch that represents the effect of the shallow convection. However, the original BMJ scheme produces excessive rainfall if the scheme is used in the Indo-Pacific region. In the modification, a more moist reference profile can prevent the model from raining too readily. As noted in Fonseca et al. (2015), nudging is necessary to get the water vapor right for the tropics, and modified BMJ scheme also needs nudging. I will

discuss nudging in the next part.

2.3.4 Nudging

Nudging mechanism is implemented in WRF as a module. It allows the numerical model to adjust the state simulated according to the referential dataset, usually the observational data or the result from another model run. The main purpose is to curb the growth of the error in the numerical model (Lo et al. 2008; Bowden et al. 2012). Either the spectral nudging or the grid nudging can be used. I use the grid nudging that operates on each grid point directly. The adjustment is according to the following equation:

$$\frac{dx}{dt} = \frac{x_0 - x}{\tau} \quad \text{Equation 2.1}$$

Where x is a model field to be nudged, and x_0 is the referential quantity provided by the referential dataset, i.e. CFSR reanalysis in my case. τ is the time scale of nudging. At the grid point that is to be nudged, a tendency term is added, which is proportional to the difference between the referential value and the current value. The further the quantity drifts from the reference, the larger the nudging tendency is. The tendency helps the model come closer to the reference. Moreover, the strength of the nudging is inversely proportional to the time scale τ . We nudge the levels above 800 hpa or above the PBL. The time scale for nudging is 1 hour. Four fields are nudged, which are two components of the horizontal wind, potential temperature and water vapor

mixing ratio.

The application of nudging is debatable in some scenarios. Nudging has two weak points. First of all, nudging relies on an extra database as a reference. Secondly, nudging terms provide artificial source and sink, and make the budget unclosed. Nevertheless, the weak points of nudging do not influence my case. The aim of my model run is to downscale the CFSR reanalysis in order to get an hourly dataset to represent the atmospheric condition. CFSR will my natural choice as the reference, and no budget analysis will be conducted. As shown in previous studies, the dynamical downscaling with nudging is widely adopted (Heikkilä et al. 2011). It is able to downscale the original dataset to a fine resolution, while preserves the large-scale dynamics of the referential dataset (Otte et al. 2012).

2.3.5 Verification

The details of the model setup can refer to section 4.2. Fonseca et al. (2015) has shown that, with the same setup, the model's rainfall is similar to CFSR and the Tropical Rainfall Measuring Mission (TRMM). Here I use the observed ZWD to verify the performance of the numerical model. Two ways of model runs are conducted: with and without nudging. Other setups will be the same. In order to compare the station-based observation, we interpolate the grid of the

numerical model to get the simulated ZWD at the location of each GPS station.

As shown in Figure 2.6, the biases of the model runs are small. The model with nudging is generally wetter than the model without nudging, and the observation is in between. Anyway, the differences between the means of the simulation and the means of the observation are around 0.02m that is considered to be small. Different stations do not differ much in the means of ZWD. The comparison of the mean cannot tell us which kind model run is better, so I further look into the correlation.

Figure 2.6 The mean of ZWD for each station. The blue dots come from the model with nudging.

The green dots come from the model without nudging. The red dots are from the observation.

For the correlation, the model with nudging greatly outperforms the model without nudging. Figure 2.7 shows the correlation coefficients between the observation and the model. The model with nudging has correlation coefficients around 0.8 with the observation, which is rather plausible. In contrast, the model without nudging badly correlates with the observations. Nudging effectively reduces the growth of error, and prevents the large discrepancy as in the model without nudging, so the model with nudging can well simulate the observed ZWD.

Figure 2.7 The correlation coefficients between the observation and the numerical model. The blue squares are the correlation coefficients between the observation and the model with nudging, and

the green triangles are the correlation coefficients between the observation and the model without nudging.

Figure 2.8 The histograms of the GPS-observed ZWD. The titles record the skewness calculated from the distributions.

Besides the correlation coefficients, the skewness also shows the effectiveness of the nudging. Figure 2.8 plots the histogram the data for each GPS station, and the title indicates the skewness of the distribution. All distributions have negative skewness. The negative skewness is also shown in the histogram of the distributions. With negative skewness, the histogram has with a gentle slope on the left side and a sharp slope on the right side of the histogram. The distribution is tilted to the side of high ZWD. For the model with nudging (Figure 2.9), the simulated data also have negative skewness, similar to the

observation. However, the model without nudging (Figure 2.10) is not successful in producing the distributions with negative skewness. Some stations have positive skewness and the distributions are more symmetric. In all, the model without nudging also fails to capture the distribution of ZWD.

Figure 2.9 Same as Figure 2.8, but for the simulated ZWD from the model with nudging.

Figure 2.10 Same as Figure 2.8, but for the simulated ZWD from the model without nudging.

Whether the nudging kills the small-scale features is a question remained. As I will show in Chapter 4, the model can reproduce the diurnal cycle and its propagation very well even with strong nudging. Remembering the nudging fields only change every 6 hour, the result means that the model can produce the small-scale features. In all, the large-scale features from CFSR are well preserved. On the other hand, the model successfully generates the high-frequency features, as verified by the GPS observation. The downscaling of the tropical climate using the WRF model is a valid way to study the high-frequency phenomena.

2.4 Other Sources of Data

There are other sources of data used in this thesis. These data are used to support the dynamical interpretation. They include the surface pressure observation from Integrated Surface Global Hourly Data (dataset DS-3505, <https://data.noaa.gov/dataset/integrated-surface-global-hourly-data>) and NCEP 40-year reanalysis data (Kalnay et al. 1996). The former is the direct observation from the ground stations, and the latter is an old-generation reanalysis data. I will discuss these data sources in individual chapters.

2.5 References

- Arakawa, A., & Schubert, W. H. (1974). Interaction of a cumulus cloud ensemble with the large-scale environment, Part I. *Journal of the Atmospheric Sciences*, 31(3), 674-701.
- Betts, A. K. (1986). A new convective adjustment scheme. Part I: Observational and theoretical basis. *Quarterly Journal of the Royal Meteorological Society*, 112(473), 677-691.
- Betts, A. K., & Miller, M. J. (1986). A new convective adjustment scheme. Part II: Single column tests using GATE wave, BOMEX, ATEX and arctic air-mass data sets. *Quarterly Journal of the Royal Meteorological Society*, 112(473), 693-709.
- Bowden, J. H., Otte, T. L., Nolte, C. G., & Otte, M. J. (2012). Examining interior grid nudging techniques using two-way nesting in the WRF model for regional climate modeling. *Journal of Climate*, 25(8), 2805-2823.
- Fonseca, R., Zhang, T., & Koh, T. Y. (2015). Improved simulation of precipitation in the tropics using a modified BMJ scheme in WRF model. *Geosci. Model Dev. Discuss.*, 8, 4019-4049
- Grell, G. A., Dudhia, J., & Stauffer, D. R. (1994). A description of the fifth-generation Penn State/NCAR mesoscale model (MM5).

- Heikkilä, U., Sandvik, A., & Sorteberg, A. (2011). Dynamical downscaling of ERA-40 in complex terrain using the WRF regional climate model. *Climate dynamics*, 37(7-8), 1551-1564.
- Iacono, M. J., Delamere, J. S., Mlawer, E. J., Shephard, M. W., Clough, S. A., & Collins, W. D. (2008). Radiative forcing by long-lived greenhouse gases: Calculations with the AER radiative transfer models. *Journal of Geophysical Research: Atmospheres (1984–2012)*, 113(D13).
- Janjic, Z. I. (1994). The step-mountain eta coordinate model: Further developments of the convection, viscous sublayer, and turbulence closure schemes. *Monthly Weather Review*, 122(5), 927-945.
- Kain, J. S., & Fritsch, J. M. (1990). A one-dimensional entraining/detraining plume model and its application in convective parameterization. *Journal of the Atmospheric Sciences*, 47(23), 2784-2802.
- Kalnay, E., Kanamitsu, M., Kistler, R., Collins, W., Deaven, D., Gandin, L., ... & Joseph, D. (1996). The NCEP/NCAR 40-year reanalysis project. *Bulletin of the American Meteorological Society*, 77(3), 437-471.
- Klemp, J. B., Dudhia, J., & Hassiotis, A. D. (2008). An upper gravity-wave absorbing layer for NWP applications. *Monthly Weather Review*, 136(10), 3987-4004.
- Lo, J. C. F., Yang, Z. L., & Pielke, R. A. (2008). Assessment of three dynamical climate downscaling methods using the Weather Research and

Forecasting (WRF) model. *Journal of Geophysical Research: Atmospheres* (1984–2012), 113(D9).

- Otte, T. L., Nolte, C. G., Otte, M. J., & Bowden, J. H. (2012). Does nudging squelch the extremes in regional climate modeling? *Journal of Climate*, 25(20), 7046-7066.
- Saha, S., Moorthi, S., Pan, H. L., Wu, X., Wang, J., Nadiga, S., ... & Reynolds, R. W. (2010). The NCEP climate forecast system reanalysis. *Bulletin of the American Meteorological Society*, 91(8), 1015-1057.
- Skamarock, W. C., Klemp, J. B., Dudhia, J., Gill, D. O., Barker, D. M., Wang, W., & Powers, J. G. (2005). *A description of the advanced research WRF version 2* (No. NCAR/TN-468+ STR). National Center For Atmospheric Research Boulder Co Mesoscale and Microscale Meteorology Div.
- Skamarock, W. C., & Klemp, J. B. (2008). A time-split nonhydrostatic atmospheric model for weather research and forecasting applications. *Journal of Computational Physics*, 227(7), 3465-3485.

Chapter 3. Intraseasonal variability of Zenith Wet Delay

3.1 Introduction

The Sumatran GPS Array (SuGAR) is a Global Positioning System (GPS) network located in Sumatra Island. Sumatra Island lies in the so-called “maritime continent” (Ramage 1968). The maritime continent includes islands and peninsulas of Malaysia, Indonesia, Philippines and Papua New Guinea. It has a tropical climate with high temperature and abundant rainfall. The maritime continent is also featured by the complex topography that can further encourage the convective rainfall (Saito et al. 2001; Yang and Slingo 2001). Strong convections in this region are a major force driving the general circulation, so the regional climate of the maritime continent can impact the global climate (Neale and Slingo 2003; Chang et al. 2004a).

The climate of the maritime continent is influenced by several factors. The interannual variability such as the El Niño Southern Oscillation (ENSO; Holton and Dmowska 1989) and the Indian Ocean Dipole (IOD; Saji et al. 1999) changes the atmospheric status year by year. The seasonal variation is mainly due to the Asian-Australian monsoon (Chang et al. 2004b). The monsoonal wind blows from Australia to Asia during the southwest monsoon season (June,

July, August, September) and reverses during the northeast monsoon season (December, January, February, March). Furthermore, within a single monsoon season, the atmospheric condition might experience dramatic changes due to the intraseasonal variability (Wang et al. 2009). The Madden Julian Oscillation (MJO; Madden and Julian 1971) is active during the northeast monsoon season, while the Boreal Summer Intraseasonal Oscillation (BSISO; Wang and Xie 1997) modulates the southwest monsoon season.

Intraseasonal variability, as reviewed by Wang et al. (2009), can lead to the activation or the suppression of atmospheric convections. Excessive rainfall or dry spells would accompany the intraseasonal variability, and strike the agriculture of the region. The intraseasonal variability of rain and cloud has been extensively studied (Fukutomi and Yasunari, 1999; Annamalai and Slingo, 2001; Chan et al., 2002). However, intraseasonal variability of water vapor is less studied, although sparse studies do exist (Cadet and Greco 1987; Chen et al. 1988). Since the atmosphere is a coupled system, the understanding of the water vapor is an inevitable step towards the full understanding of the regional climate. The limited efforts do not match the importance of water vapor in the hydrological cycle.

The lack of study partly reflects the difficulty of measuring water vapor.

Radiosonde stations can measure the vertical profile of water vapor, but radiosonde stations are quite sparse in the tropics, particularly in the maritime continent (Koh and Teo 2009). The Global Positioning System (GPS) technology can observe the atmosphere (Bevis et al. 1992). This technology was primarily designed to measure locations of ground stations by measuring the traveling time of GPS signals from a satellite to a ground station (Hofmann-Wellenhof 2012). The traveling time of GPS signals can be delayed by dry and wet components in the atmosphere. The wet delay is mainly controlled by the amount of water vapor along the travel path of GPS signals.

In this paper, we use the GPS observation of ZWD from the SuGAR network (Feng et al. 2015). This GPS network is originally designed to monitor the tectonic movement along the west coast of Sumatra (Hill et al. 2012). Besides the location information, the GPS stations can measure ZWD and thus serve the atmospheric research. Fujita et al. (2011) used some of the GPS stations to study the impact of MJO on diurnal cycles of the water vapor. The observation from the same network was also used for data assimilation in the numerical model (Shoji et al. 2011), which improves the simulation of a tropical cyclone. In our study, we want to answer the question: which physical processes are controlling intraseasonal variability of ZWD. As a result, we use an objective analysis method, i.e., rotated EOF, to let the observational data talk by itself.

The observation-based analysis and the objective approach make our study distinct with Fujita et al. (2011) and Shoji et al. (2011).

According to our results, two processes dominate the intraseasonal variability during the southwest monsoon season of 2008. First is the monsoon. The strength of the monsoon strongly modulates the observed ZWD. Active monsoon brings dry condition to Sumatra, while weak monsoon is linked with the wet condition. The result is similar to the studies of BSISO (Wang and Xie 1997). On the other hand, dry air coming from the upper troposphere of the Southern Hemisphere also excites the intraseasonal variability of ZWD. This process, which is named as dry air intrusion, was already proved important for the tropical-extratropical interactions in places such as western Pacific (Yoneyama and Parsons 1999) and western Africa (Roca et al. 2005). Our work reveals dry air intrusion plays an important role in eastern Indian Ocean.

3.2 Data

3.2.1 GPS observation

We study the GPS observation of the southwest monsoon season of 2008. We select year 2008 because it is both ENSO and IOD neutral (Lestari and Koh, in preparation; Ummenhofer et al. 2009). ENSO and IOD are the most important

interannual variabilities for the maritime continent. Besides interannual variability, the seasonal variability can also interact with the intraseasonal variability. Chang et al. (2004b) showed that the atmosphere can vary a lot during different monsoon seasons, and Wu et al. (2006) revealed the possible interaction between the intraseasonal variability and the seasonal variation. Therefore, the best way is to study seasons respectively. In this paper, we only study the intraseasonal variability during the southwest monsoon season, and leave other seasons to the future work.

There were 30 working stations during 2008. Seventeen stations are used for this study (Figure 3.1). The exclusion of the 13 stations is due to the existence of data gaps or large erroneous values. Among the 17 valid stations, one station (NTUS) is in Singapore, one station (JMBI) in East Sumatra, four stations on the west coast of the main island. The rest 11 stations are on offshore islands to the west of the main island. The network extends from 5.35°S to 5.05°N in the latitudinal direction, and 95.34°E to 103.68°E in the longitudinal direction. In all, we have a relatively dense network in a small region, and the west coast of Sumatra is well observed.

Figures 3.1 Stations used for the study of the intraseasonal variability

From the SuGAR network, we are interested in ZWD from each GPS station. This is the time delay caused by the wet component of the atmosphere when the GPS signal passes through the atmosphere. The path of the signal is usually a slant path. The slant delay can be converted into zenith delay using a mapping function (Davis et al. 1985). Compared with slant delay that can point to various directions, zenith delay gains mathematical simplicity with the loss of anisotropy. We can compensate the anisotropy with the spatial variation provided by the network, so the mathematical simplicity can justify our focus on ZWD. We thus retrieve the hourly output of ZWD for each GPS station. The details of the processing can be referred to Feng et al. (2015). Since we are mainly interested in the intraseasonal variability, we further calculate the daily

mean of ZWD. In the end, 17 time series of daily-averaged ZWD are studied.

The algorithm of converting ZWD to PWV exists, but the algorithm requires other input quantities besides ZWD, such as surface pressure and temperature. We do not have direct measurements of these quantities at the sites of GPS stations. Other datasets, such as reanalysis datasets, can provide these quantities. However, these datasets are the product of the numerical models and might introduce more errors. Since PWV is mainly decided by ZWD according to our analysis, and the two quantities correlate well, ZWD itself can represent the water vapor condition. In the end, we decide to study ZWD directly, so that our analysis is based on direct GPS observation only. Still, although we did not convert ZWD to PWV directly, we will prove their high-score correlation at the beginning of our analysis.

3.2.2 CFSR Reanalysis Data

After the direct analysis of GPS-observed ZWD, we surely want to gain some insights about the dynamics. However, the observational network limits our further study. The GPS network can only reveal the aspect of water vapor, and we need other quantities to diagnose. Although the network is considered to be dense, the network is still possible to miss some small-scale features of the atmospheric phenomena. Furthermore, the SuGAR network is confined in a

small area, but the atmospheric processes can be large-scale. In all, we need to know various atmospheric quantities in a larger area, and with a finer spatial resolution.

For such a purpose, we refer to the Climate Forecast System Reanalysis data (CFSR; Saha et al. 2010) to get a full picture of the atmosphere. This dataset is provided by the National Centers for Environmental Prediction (NCEP). A global model is consistently run from 1979 to present, providing 6-hourly global output. There are 64 model levels, and the model top is 0.266hPa. The horizontal resolution is $0.5^{\circ} \times 0.5^{\circ}$, and the surface fields are available at a higher resolution. During the model run, observational data are assimilated into the model to curb the growth of errors. The input data include such as satellite radiances, sea ice, radiosonde, surface observation, and other data sources. The CFSR dataset is evaluated by several studies (Hodges et al. 2011; Wang et al. 2011; Xue et al. 2011; Bao and Zhang 2013). The quality of the dataset is reasonably well compared with the observation and other reanalysis data.

3.3 Methods

3.3.1 Rotated EOF

There could be multiple atmospheric processes that simultaneously influence the SuGAR network. We use the rotated Empirical Orthogonal Function (rotated EOF; Richman, 1986) analysis to process the multivariate ZWD data. The aim is to retrieve the major components from the multivariate data of ZWD observation, and attribute each component to an atmospheric process. Rotated EOF is based on the Empirical Orthogonal Function (EOF; Lorenz 1956) analysis. After processing the multivariate data with the EOF analysis, we retain the leading principal components (PCs) that could explain a major part of the total variance. Each PC includes two aspects: the eigenfunction that is the spatial pattern, and the time score that is the temporal strength. An eigenfunction has several elements called “loadings”, and each loading corresponds to the contribution of one GPS station to the PC. The time score indicates the strength of the eigenfunction. It is normalized and has unit variance. Rotated EOF adds one more step to the EOF analysis. In the step of rotation, the Varimax Rotation (Kaiser 1958) is used to recombine the eigenfunctions of leading PCs. The rotation maximizes the variance of squared loadings, while preserves the total variance explained by the leading PCs. The Varimax Rotation can “localize” the signal, so that the eigenfunction are contributed by less number of stations. The time scores are rotated

accordingly. We can then get a new set of leading PCs as the final product of the rotated EOF analysis.

Both the EOF analysis and the rotated EOF analysis are used in meteorological studies (Kawamura 1994; Saji et al. 1999; Wheeler and Hendon 2004), but they have different properties. For the EOF analysis, the eigenfunctions of the PCs must be orthogonal, and the time scores are independent with each other. For the rotated EOF analysis, the eigenfunctions are not necessarily independent, while the time scores are independent with each other (Mestas-Nunez 2000). In our case, each eigenfunction is a 17 dimensional vector, representing the spatial pattern of the component. Each element of the eigenfunction is called a loading, corresponding to one GPS station. A loading represents the contribution of the GPS station to the PC. The time score is daily. It tells us how strong the spatial pattern is each day. We will use the spatial and temporal information to get the insight for each component.

3.3.2 Linear Regression

The PCs come from the ZWD data. If the PCs are physically meaningful, there should be some physical process driving each PC. The physical process must be able to explain the spatial and temporal characteristics of the PC. The atmosphere is a complex system, and the physical process involves the

evolution of various quantities. As a result, it is necessary to study the relationship between the PCs and other physical fields of the atmosphere.

As a first step, we want to find how each PC concurrently evolves with CFSR fields. To achieve the goal, we use the simple ordinary least square method (Wilks 2011) to do the linear regression. The linear regression can be used to build a linear model to predict certain variables based on the predictors, and are widely used in climate studies (Mann and Emanuel 2006; Sousa et al. 2007). The regression is always between a PC score and a physical quantity at a grid point from CFSR dataset. The PC score is taken as the only independent variable, i.e., we have the following linear model:

$$Y = aX + b$$

In our case, X is the PC score, and Y is a physical quantity at a grid point. With the linear model, we can predict the atmospheric condition according to the value of PC score.

Besides the concurrent relationship, the sequence of the evolution is also important. We can only build up the causal relationship by knowing the sequence of the events. For example, if the temperature change leads the humidity change, there is a probability for the temperature change to be the cause of the humidity change, instead of the other way around. We use the

lead-lag regression to find out the lagged relationship between the PC score and the physical fields (Peña et al. 2011). The process is similar to the simple linear regression, but the PC score is first delayed or advanced by several days. The PC score is again regressed with the CFSR data. By knowing the regression coefficient, the linear model can predict the atmospheric status several days before or after the peak of PC scores. With multiple time offsets, we can know how the atmosphere progressively evolves with certain PC.

3.4 Results

3.4.1 EOF and rotated EOF

Figure 3.2 Variance explained by the principal components of the EOF analysis.

As the first step of the rotated EOF analysis, we apply the EOF analysis to the

multivariate data, and find two leading PCs (Figure 3.2). Together, the two leading PCs explain 86.28% of the total variance. We thus focus only on the two PCs and ignore the rest. As the second step, varimax rotation is applied to two PCs and ignore the rest. After the rotation, the first PC explains 51.16% of the total variance, while the second PC explains 35.12%. The spatial patterns are shown in the eigenfunctions of the PCs (Figure 3.3). For PC1, all the stations oscillate together. For PC2, northern stations oscillate oppositely with the southern stations.

Figure 3.3 Eigenfunctions of PC1 and PC2 (From the EOF analysis). The areas of the dots are proportional to the loadings of the stations.

After the rotation, both eigenfunctions show pan-island oscillations (Figure 3.4). The eigenfunctions differs in their distributions of the loadings. For the first eigenfunction (Figure 3.4a), the loadings in the southern end of Sumatra have smaller amplitudes. In contrast, the loadings of the second eigenfunction decrease when it comes to the north (Figure 3.4b). In both cases, most stations

have negative loadings. Since the negative ZWD corresponds to low PWV, eigenfunctions represent spatial distributions of dry conditions.

Figure 3.4 Eigenfunctions of PC1 and PC2 (From the rotated EOF analysis). The areas of the dots are proportional to the loadings of the stations.

Figure 3.5 shows the time scores of the two PCs. The time scores tell us the temporal variations of the spatial patterns. We need to interpret the time scores with the eigenfunctions. In the method of rotated EOF, the physical condition represented by a principal component can be reconstructed by the multiplication of the loading and the time score. For a station with negative loading, a positive time score would mean a negative anomaly of ZWD, i.e., a dry condition. For figure 3.5, most loadings of the eigenfunctions are negative, so a positive value in the time score means a dry condition over the network. By rough counting, there are 4 peaks in PC1 score (Figure 3.5a) and 6 peaks in PC2 score (Figure 3.5b). The frequency of PC2 is higher than PC1.

Figure 3.5 Time scores of PC1 and PC2 (From the rotated EOF analysis)

3.4.2 The Choice of the rotated EOF

The linear regression is used to understand the dynamics in Section 5 of this chapter, i.e. the Discussion Part. Here we only highlight some results to support our choice of the rotated EOF method. In fact, the EOF method was more widely used than the rotated EOF analysis. As we have seen (Figure 3.3 and Figure 3.4), PC2 from the EOF analysis are very different from PC2 of the rotated EOF analysis. However, with the criterion of thermodynamic consistency, we can demonstrate that the rotated EOF analysis is better than the EOF analysis for our study.

Since water vapor is an important phase in the thermodynamic process (Rasmusson, 1968), the change of ZWD should be consistent with the cloud condition that can be reflected by the outgoing longwave radiation (OLR; Waliser et al. 1993). We linearly regress OLR with PC2 scores before and after

the rotation. Figure 3.6 shows the OLR anomaly predicted by 1 std score. For the score before the rotation and the score after the rotation, both have a patch of positive anomaly centered in the southern part of Sumatra. Remembering that PC2 of the EOF analysis actually reveals an anti-oscillation (Figure 4.2b), so it is not consistent with the spatial pattern of the OLR anomaly. On the other hand, PC2 of the rotated EOF analysis (Figure 4.3b) is consistent.

Figure 3.6 Linear regression's prediction of OLR anomaly by 1 std PC2 score. a) PC2 score from the EOF analysis. b) PC2 score from the rotated EOF analysis

The problem of the EOF analysis may be attributed to its mathematical rigidity. As Mestas-Nunez (2000) has proved, the EOF analysis requires the eigenfunctions to be orthogonal, which is purely mathematical. Due to the requirement of spatial orthogonality, if PC1 is a pan-island oscillation pattern, the loadings of PC2 cannot be the pan-island pattern again. As a result, the EOF analysis could exclude physically meaningful components with the mathematical rigidity. The rotated EOF analysis allows the possibility of two

pan-island patterns, because it does not require the orthogonality of the eigenfunctions. In the following discussion, we will further explore the physical meaning of the rotated PCs.

3.5 Discussion about PC1

We have retrieved two components from the rotated EOF analysis. In this section, we discuss the PC1 first. As we have mentioned, the coverage of the network might be too small to capture a large-scale phenomenon. For example, the dryness revealed by PC1 (Figure 3.4a) might cover a larger region. In order to know the horizontal scale of the phenomenon, we regress PWV from the CFSR dataset against the PC1 score. Figure 3.7 shows the PWV anomaly predicted by 1 std PC1 score.

From Figure 3.7, Sumatra Island is covered by negative PWV anomaly. This is consistent with the first eigenfunction of ZWD (Figure 3.4a). The correlation between ZWD and PWV is significant in Sumatra. The map further tells us the scale of the phenomenon. The dry anomaly is actually much larger than the coverage of the SuGAR network, stretching westward into the Indian Ocean. At the same time, there is a patch of positive PWV anomaly in the northern hemisphere, especially around northern India, the Bay of Bengal and South China Sea. In conclusion, PC1 of the ZWD is in essence an oscillation of PWV in

the eastern part of the Indian Ocean, accompanied by an opposite response in northern India.

Figure 3.7 Linear regression's prediction of PWV by 1 std PC1 score

We can regress the specific humidity with PC1 score. The result can tell us the vertical distribution of the dry anomaly. We pick a vertical cross-section along the latitude 3°N, which cuts through the center of the dry anomaly. Figure 3.8 shows the specific humidity anomaly predicted by 1 std PC1 score. The dry patch extends from about 900hpa to 500hpa, which is the low-middle troposphere. Although the lowest level of the atmosphere contains a large portion of the water vapor, the low levels do not contribute much to the dry anomaly. As a result, the mechanism we are searching for should not depend on the dynamics of the planet boundary layer.

Figure 3.8 Linear regression's prediction of the specific humidity anomaly by 1 std PC1 score. The vertical cross section cuts through 3°N latitude.

We then look into the 850hpa wind that plays a significant role in low-level advection. We still apply the linear regression method. Zonal and longitudinal wind components are regressed with the PC1 score separately. With 1 std PC1 score, we can get the predicted values for two components of the wind anomaly. In Figure 3.9, arrows show the 850hpa wind anomaly predicted by 1 std PC1 score. From the arrows, we can see a band of wind anomaly from Arabian Sea to East Asia. The band of wind anomaly is located around the monsoon trough. The wind anomaly is mainly westerly, and follows the direction of southwest monsoon (Lau and Yang 1997). Negative PWV anomaly is located on the southern bank of the wind anomaly, and positive PWV anomaly on the northern bank. A strengthened monsoon is thus

accompanying the dry condition around Sumatra.

Figure 3.9 Lead-lag linear regression and its prediction of wind and PWV anomaly by 1 std PC1 score. The wind anomaly is at 850hPa level.

Interestingly, the coupling between the wind anomaly and the PWV anomaly is

closer than what we described, if we look into the whole evolution process. As shown in Figure 3.9, with 6 days leading, the wind anomaly is very strong near Arabian Sea. This section of the strong wind anomaly is flanked by negative PWV anomaly to the south and positive PWV anomaly to the north. With the evolution of time, the section of the strong wind anomaly propagates eastward, and PWV anomalies propagate as well. A slight northward shift also takes place. The eastward propagation and the slight northward offset are consistent with previous studies of the intraseasonal variability of the summer monsoon, as reviewed by Wang et al. (2009).

The dipole oscillation of dry and wet anomalies might be a response to the strengthened monsoon, as pointed out by several studies (Joseph and Sijikumar 2004; Yoon and Huang 2012). Strengthened monsoon can transport water vapor from Southeastern Indian Ocean to the Bay of Bengal. To confirm Joseph and Sijikumar's argument, moisture flux is linearly regressed with PC1 score. As shown in Figure 3.10, strengthened monsoon can increase the eastward transportation of the water vapor. Furthermore, three days before the dry peak, a southerly surge of moisture flux appears near the east side of the Bay of Bengal. It would encourage the transportation of water vapor from Sumatra to the Bay of Bengal. As a result, a dipole of PWV anomaly (Figure 3.9) takes place on the two sides of the monsoon.

Figure 3.10 Lead-lag linear regression and its prediction of moisture flux anomaly by 1 std PC1 score.

The Indian Monsoon Index (Wang et al. 2001) can further confirm the link between PC1 score and the monsoon. The index was devised to measure the strength of Indian Monsoon. The definition of the index is based on the low

level wind. It is defined as the difference of 850hpa zonal wind of two regions. Region 1 covers Arabian Sea, and Region 2 covers northwestern part of Indian subcontinent (Figure 2 in Wang et al. 2001). With data from 1948 to 1997, the Indian Monsoon Index has a correlation coefficient of 0.72 (Wang et al. 2001) with the all-Indian summer rainfall (Parthasarathy et al. 1994). Although all-Indian summer rainfall is more widely used as an index of the Indian Monsoon, we think the Indian Monsoon Index is more suitable for our purpose because it is based on the 850hpa wind that we are interested in.

Figure 3.11 Indian Monsoon Index and PC2 score with 5-day running mean.

Since we link PC1 with the monsoon strength, the PC1 score should match IMI. In order to get rid of the high frequency noise, 5-day running mean is applied to the PC1 score and IMI, as shown in Figure 3.11. The curve of PC1 score

follows the curve of IMI quite well. The Cross-correlation of PC1 and the Indian Monsoon Index is also calculated. When the IMI leads the PC1 score by five days, the correlation coefficient reaches the maximum which is as high as 0.724, so the PC1 score closely follows IMI. Above analysis is consistent with the hypothesized relationship between PC1 of ZWD and the strength of the monsoon.

In all, PC1 is associated with the strength of monsoon. When the monsoon strengthens, water vapor flux would transport the water vapor to the north and cause the dry anomaly from 500hPa to 800hPa in Sumatra. ZWD of the GPS network is strongly influenced by such a process.

3.6 Discussion about PC2

PC2 explains 35.12% of the total variance. Following the same philosophy, the simple linear regression and the lead-lag regression are again used to explore the physical meaning of PC2.

We can linearly regress PWV against PC2, to find out the horizontal distribution of wet and dry anomalies. Figure 3.12 shows a patch of dry anomaly in the southern part of Sumatra, as predicted by 1 std PC2 score. The dry anomaly is centered near the southern end of Sumatra, and barely

influences northern Sumatra. The magnitude of the dry anomaly is about 5 mm, comparable to that of PC1. If we recall the spatial pattern of PC2 (Figure 3.4b), the eigenfunction agrees well with the regression map of Figure 3.12, as the loadings are large in southern Sumatra and decrease when coming to the north.

Figure 3.12 Linear regression's prediction of PWV by 1 std PC2 score.

In order to show the vertical profile of the humidity, we take a cross-section along 5°S. This latitude cuts through the center of the dry anomaly. The specific humidity along the cross-section is regressed against the PC2 score. The linear model shows that the dry anomaly spreads from 800hPa to 500hPa (Figure 6.2), with the maximum around 600hPa.

Figure 3.13 Linear regression's prediction of the specific humidity anomaly by 1 std PC2 score.

The vertical cross section cuts through 5°S latitude.

By knowing the center of dry anomaly, we can trace dry events backward to see where the air parcel comes from. Back-trajectory analysis can be used to study the transportation of air parcels. From PC2 score, two strong dry events happen on 21 June 2008 and 23 July 2008. The end points of the trajectories are picked around the center of dry anomaly, i.e., 102.5°E, 5°S, at the altitude of 4000m, around the level of 600hPa. Nine points are selected, with 2.5° of spacing.

As shown in Figure 3.14, air parcels can be traced back to the Southern Hemisphere around the latitude of 30°S. They start from a relatively high level, and descent when they come to the tropics. During the whole process, air

parcels are almost isentropic with 325K potential temperatures, if we ignore the slight cooling trend. The cooling can be attributed to the radiative cooling in the upper troposphere (Liou 2002). In all, the movements of the air parcels are similar to the process of extratropical air intrusions (Parsons et al. 2000; Roca et al. 2005; Zuidema et al. 2006). Since the upper troposphere of extratropical regions is much drier than middle troposphere, the intrusion of the extratropical air can bring dry anomaly to Sumatra.

Figure 3.14 Back-trajectory of two positive peaks in the PC2 score.

The back-trajectory analysis provides the Lagrangian view following the movement of the air parcel. We can use the linear regression to reveal the Eulerian view of the dynamics. We first study the evolution of wind anomaly at the 325K isentropic level. From the above analysis, we already know that the intrusion generally follows the potential temperature of 325K. The horizontal

wind on this isentropic level is regressed with the PC2 score. The arrows in Figure 3.15 show the wind anomaly predicted by 1 std of PC2 score. To the south of 15°S, a wave train can be identified. The wind anomaly is arranged into a row of rotational cells. A clockwise cell follows an anticlockwise cell. The colored contours indicate the anomaly of potential vorticity (PV). The anticlockwise centers have positive PV anomaly, and the clockwise centers have negative PV anomaly. The coupling of the rotational wind with the PV is the typical characteristic of Rossby wave.

The lead-lag regression indicates that the wave propagates eastward, as shown in Figure 3.15. The eastward propagation happens when the westerly background flow surpasses the westward movement of the Rossby wave. When the number of lagging days reduces, a negative PV cell stretches and extends into the tropical region. The wind anomaly changes accordingly, with a southeasterly wind anomaly pointing towards Sumatra, which can encourage the southerly air mass transportation.

Figure 3.16 Lead-lag linear regression and its prediction of wind anomaly and PV anomaly at 325K isentropic level by 1 std PC2 score.

The dry air also intrudes into the tropics. Figure 3.17 shows the specific humidity predicted by the linear model. Note that we plot the absolute value instead of the anomaly as in other figures. On the level of 325K, the specific humidity is low in the subtropical region of the Southern Hemisphere. With time, a tongue of dry air from the Southern Hemisphere moves northward and reaches Sumatra Island. The movement mimics that of the PV anomaly. It is not surprising, considering that specific humidity is conserved in the adiabatic process. Thus, during the evolution of the subtropical Rossby wave, the wave can influence the tropical region, bringing air parcels with low PV and low humidity.

Figure 3.17 Lead-lag linear regression and its prediction of wind anomaly and the specific humidity at 325K isentropic level by 1 std PC2 score.

The role of Rossby wave in exchanging the tropical-extratropical air was studied before. Newell et al. (1992) found a narrow filament of water vapor flux from the tropics to the extratropical region. They termed the phenomenon as the “atmospheric river”, and it can influence the weather at mid-latitude. There are usually 4 to 5 such rivers in a latitude circle, and the rivers couple with the mid-latitude wave activity. As a counterpart of the “atmospheric river”, the dry-air intrusion can transport the dry air back to the tropics. This reversed transportation is in a belt wider than the typical atmospheric river (Figure 10-21 in Koh 2001). Our study of PC2 shows that the dry-air intrusion is important for the tropical climate and it deserves more attention in the future study.

3.7 Comparison of Two Mechanisms

In the above discussions, I proposed two mechanisms to explain PC1 and PC2 of the SuGAR network. The first mechanism is the monsoon, and the second one is the dry air intrusion. The discussions are independent. It is interesting to ask whether two mechanisms can contribute to each other.

Figure 3.18 IMI index and PC2 Score

Figure 3.18 put the IMI index and the PC2 score together. PC2 correlates badly with the IMI index. If we look into the lead-lag correlation of the IMI index and the PC2 score, two curves' correlation score is always low regardless of the time offset (Figure 3.19). Given the low correlation score, it is hard to believe that the monsoon plays a role in the evolution of PC2.

I also apply the trajectory analysis to the peak of PC1, as what I did in PC2. From PC1 score (Figure 3.5), a strong dry event happens on 29 July 2008. The end points of the trajectories are picked around the center of dry anomaly, i.e.,

92.5°E, 1.25°N, at the altitude of 3500m. Nine points are selected, with 2.5° of spacing. As shown by the trajectory (Figure 3.19), the air parcel came from the west, rather than the Southern Hemisphere. The mechanism of dry air intrusion in PC2 fails to explain what happens in PC1.

Figure 3.14 Back-trajectory of one positive peak in the PC1 score.

3.8 Conclusion

We studied the southwest monsoon season of 2008 with the ZWD observations from the SuGAR network. We discovered two leading components in the intraseasonal variability of ZWD. The two components explain a major part of the total variance.

The first component is attributed to monsoon strength. The strengthened monsoon causes the change of water vapor flux and encourages the northward transportation of the water vapor. As a result, Sumatra becomes drier and the Bay of Bengal gets humidified. As a representation of the monsoon strength, the Indian Monsoon Index can correlate well with PC1 score if lagged by 5 days. This indicates the close relationship between the monsoon and the Sumatran humidity. The above conclusion is consistent with the study of BSISO, and reveals the aspect of the water vapor with the direct observation of GPS network.

The second component mainly influences the southern part of Sumatra. The dry air intrudes from the subtropical region in the Southern Hemisphere and descends from the upper troposphere to the middle troposphere. Subtropical Rossby wave is linked with dry air intrusion events. Actually the dry air intrusion is the counterpart of the atmospheric river under the framework of subtropical eddies, although the dry air intrusion drew less attention before. Future study should pay more attention to the dry air intrusion since it can cause a significant intraseasonal oscillation as observed by the SuGAR network.

This paper has proved that a dense GPS network like SuGAR can reveal the

large-scale dynamics. Since the SuGAR network is expanding and continuously working, we expect the network to accumulate abundant observational data in the following years. The data collected will be a wealth for the study of the regional climate of the maritime continent.

3.9 References

- Annamalai, H., & Slingo, J. M. (2001). Active/break cycles: diagnosis of the intraseasonal variability of the Asian summer monsoon. *Climate Dynamics*, 18(1-2), 85-102.
- Bao, X., & Zhang, F. (2013). Evaluation of NCEP–CFSR, NCEP–NCAR, ERA-Interim, and ERA-40 reanalysis datasets against independent sounding observations over the Tibetan Plateau. *Journal of Climate*, 26(1), 206-214.
- Bevis, M., Businger, S., Herring, T. A., Rocken, C., Anthes, R. A., & Ware, R. H. (1992). GPS meteorology: Remote sensing of atmospheric water vapor using the Global Positioning System. *Journal of Geophysical Research: Atmospheres*, 97(D14), 15787-15801.
- Cadet, D. L., & Greco, S. (1987). Water vapor transport over the Indian ocean during the 1979 summer monsoon. Part II: water vapor budgets. *Monthly Weather Review*, 115(10), 2358-2366.

- Chan, J. C., Ai, W., & Xu, J. (2002). Mechanisms responsible for the maintenance of the 1998 South China Sea summer monsoon. *Journal of the Meteorological Society of Japan, Ser. II*, 80(5), 1103-1113.
- Chang, C. P., Wang, Z., Ju, J., & Li, T. (2004a). On the relationship between western maritime continent monsoon rainfall and ENSO during northern winter. *Journal of Climate*, 17(3), 665-672.
- Chang, C. P., Harr, P. A., McBride, J., & Hsu, H. H. (2004b). Maritime continent monsoon: annual cycle and boreal winter variability (pp. 107-152). World Scientific.
- Chen, T. C., Yen, M. C., & Murakami, M. (1988). The water vapor transport associated with the 30-50 day oscillation over the Asian monsoon regions during 1979 summer. *Monthly Weather Review*, 116(10), 1983-2002.
- Davis, J. L., Herring, T. A., Shapiro, I. I., Rogers, A. E. E., & Elgered, G. (1985). Geodesy by radio interferometry: Effects of atmospheric modeling errors on estimates of baseline length. *Radio science*, 20(6), 1593-1607.
- Feng, L., Hill, E. M., Banerjee, P., Hermawan, I., Tsang, L. L., Natawidjaja, D. H., ... & Sieh, K. (2015). A unified GPS-based earthquake catalog for the Sumatran plate boundary between 2002 and 2013. *Journal of Geophysical Research: Solid Earth*.
- Fujita, M., Yoneyama, K., Mori, S., Nasuno, T., & Satoh, M. (2011). Diurnal convection peaks over the eastern Indian Ocean off Sumatra during different

- MJO phases. *Journal of the Meteorological Society of Japan, Ser. II*, 89, 317-330.
- Fukutomi, Y., & Yasunari, T. (1999). 10-25 day Intraseasonal Variations of Convection and Circulation over East Asia and Western North Pacific during Early Summer. *Journal of the Meteorological Society of Japan*, 77(3), 753-769.
 - Hill, E. M., Borrero, J. C., Huang, Z., Qiu, Q., Banerjee, P., Natawidjaja, D. H., ... & Sieh, K. (2012). The 2010 Mw 7.8 Mentawai earthquake: Very shallow source of a rare tsunami earthquake determined from tsunami field survey and near-field GPS data. *Journal of Geophysical Research: Solid Earth (1978-2012)*, 117(B6).
 - Hodges, K. I., Lee, R. W., & Bengtsson, L. (2011). A comparison of extratropical cyclones in recent reanalyses ERA-Interim, NASA MERRA, NCEP CFSR, and JRA-25. *Journal of Climate*, 24(18), 4888-4906.
 - Hofmann-Wellenhof, B., Lichtenegger, H., & Collins, J. (2012). *Global positioning system: theory and practice*. Springer Science & Business Media.
 - Holton, J. R., & Dmowska, R. (1989). *El Niño, La Niña, and the southern oscillation*. S. G. Philander (Ed.). Academic press.
 - Joseph, P. V., & Sijikumar, S. (2004). Intraseasonal variability of the low-level jet stream of the Asian summer monsoon. *Journal of Climate*, 17(7), 1449-1458.

- Kaiser, H. F. (1958). The varimax criterion for analytic rotation in factor analysis. *Psychometrika*, 23(3), 187-200.
- Kawamura, R. (1994). A rotated EOF analysis of global sea surface temperature variability with interannual and interdecadal scales. *Journal of Physical Oceanography*, 24(3), 707-715.
- Koh, T. Y. (2001). Isentropic diagnostics of mid-latitude circulation and transport (Doctoral dissertation, Massachusetts Institute of Technology).
- Koh, T. Y., & Teo, C. K. (2009). Toward a mesoscale observation network in Southeast Asia. *Bulletin of the American Meteorological Society*, 90(4), 481-488.
- Lau, K. M., & Yang, S. (1997). Climatology and interannual variability of the Southeast Asian summer monsoon. *Advances in Atmospheric Sciences*, 14(2), 141-162.
- Lestari, R.K. and Tieh-Yong Koh: Statistical Evidence for ENSO - IOD Interaction. In Preparation.
- Liou, K. N. (2002). *An introduction to atmospheric radiation*. Academic press.
- Lorenz, E. N. (1956). Empirical orthogonal functions and statistical weather prediction. Scientific Report No. 1, Statistical Forecasting Project. Air Force Research Laboratories.
- Madden, R. A., & Julian, P. R. (1971). Detection of a 40-50 day oscillation in the zonal wind in the tropical Pacific. *Journal of the Atmospheric*

Sciences, 28(5), 702-708.

- Mann, M. E., & Emanuel, K. A. (2006). Atlantic hurricane trends linked to climate change. *Eos, Transactions American Geophysical Union*, 87(24), 233-241.
- Mestas-Nunez, A. M. (2000). Orthogonality properties of rotated empirical modes. *International Journal of Climatology*, 20(12), 1509-1516.
- Neale, R., & Slingo, J. (2003). The maritime continent and its role in the global climate: A GCM study. *Journal of Climate*, 16(5), 834-848.
- Newell, R. E., Newell, N. E., Zhu, Y., & Scott, C. (1992). Tropospheric rivers?—A pilot study. *Geophysical Research Letters*, 19(24), 2401-2404.
- Parsons, D. B., Redelsperger, J. L., & Yoneyama, K. (2000). The evolution of the tropical western Pacific atmosphere-ocean system following the arrival of a dry intrusion. *Quarterly Journal of the Royal Meteorological Society*, 126(563), 517-548.
- Parthasarathy, B., Munot, A. A., & Kothawale, D. R. (1994). All-India monthly and seasonal rainfall series: 1871–1993. *Theoretical and Applied Climatology*, 49(4), 217-224.
- Peña, D., Tiao, G. C., & Tsay, R. S. (2011). *A course in time series analysis*. John Wiley & Sons.
- Ramage, C. S. (1968). Role of a tropical “maritime continent” in the atmospheric circulation. *Monthly Weather Review*, 96(6), 365-370.

- Rasmusson, E. M. (1968). Atmospheric water vapor transport and the water balance of North America: II. Large-scale water balance investigations. *Monthly Weather Review*, 96(10), 720-734.
- Richman, M. B. (1986). Rotation of principal components. *Journal of climatology*, 6(3), 293-335.
- Roca, R., Lafore, J. P., Piriou, C., & Redelsperger, J. L. (2005). Extratropical dry-air intrusions into the West African monsoon midtroposphere: An important factor for the convective activity over the Sahel. *Journal of the Atmospheric Sciences*, 62(2), 390-407.
- Saha, S., Moorthi, S., Pan, H. L., Wu, X., Wang, J., Nadiga, S., ... & Reynolds, R. W. (2010). The NCEP climate forecast system reanalysis. *Bulletin of the American Meteorological Society*, 91(8), 1015-1057.
- Saji, N. H., Goswami, B. N., Vinayachandran, P. N., & Yamagata, T. (1999). A dipole mode in the tropical Indian Ocean. *Nature*, 401(6751), 360-363.
- Saito, K., Keenan, T., Holland, G., & Puri, K. (2001). Numerical simulation of the diurnal evolution of tropical island convection over the Maritime Continent. *Monthly Weather Review*, 129(3), 378-400.
- Sousa, S. I. V., Martins, F. G., Alvim-Ferraz, M. C. M., & Pereira, M. C. (2007). Multiple linear regression and artificial neural networks based on principal components to predict ozone concentrations. *Environmental Modelling & Software*, 22(1), 97-103.

- Shoji, Y., Kunii, M., & Saito, K. (2011). Mesoscale data assimilation of Myanmar cyclone Nargis Part II: assimilation of GPS-derived precipitable water vapor. *Journal of the Meteorological Society of Japan, Ser. II*, 89(1), 67-88.
- Ummenhofer, C. C., England, M. H., McIntosh, P. C., Meyers, G. A., Pook, M. J., Risbey, J. S., ... & Taschetto, A. S. (2009). What causes southeast Australia's worst droughts? *Geophysical Research Letters*, 36(4).
- Waliser, D. E., Graham, N. E., & Gautier, C. (1993). Comparison of the highly reflective cloud and outgoing longwave radiation datasets for use in estimating tropical deep convection. *Journal of Climate*, 6(2), 331-353.
- Wang, B., & Xie, X. (1997). A model for the boreal summer intraseasonal oscillation. *Journal of the Atmospheric Sciences*, 54(1), 72-86.
- Wang, B., Wu, R., & Lau, K. M. (2001). Interannual variability of the Asian summer monsoon: contrasts between the Indian and the Western North Pacific-East Asian Monsoons. *Journal of Climate*, 14(20), 4073-4090.
- Wang, B., Huang, F., Wu, Z., Yang, J., Fu, X., & Kikuchi, K. (2009). Multi-scale climate variability of the South China Sea monsoon: A review. *Dynamics of Atmospheres and Oceans*, 47(1), 15-37.
- Wang, W., Xie, P., Yoo, S. H., Xue, Y., Kumar, A., & Wu, X. (2011). An assessment of the surface climate in the NCEP climate forecast system reanalysis. *Climate Dynamics*, 37(7-8), 1601-1620.

- Wheeler, M. C., & Hendon, H. H. (2004). An all-season real-time multivariate MJO index: Development of an index for monitoring and prediction. *Monthly Weather Review*, 132(8), 1917-1932.
- Wilks, D. S. (2011). *Statistical methods in the atmospheric sciences*. Academic press.
- Wu, M. L. C., Schubert, S. D., Suarez, M. J., Pegion, P. J., & Waliser, D. E. (2006). Seasonality and meridional propagation of the MJO. *Journal of climate*, 19(10), 1901-1921.
- Xue, Y., Huang, B., Hu, Z. Z., Kumar, A., Wen, C., Behringer, D., & Nadiga, S. (2011). An assessment of oceanic variability in the NCEP climate forecast system reanalysis. *Climate Dynamics*, 37(11-12), 2511-2539.
- Yang, G. Y., & Slingo, J. (2001). The diurnal cycle in the tropics. *Monthly Weather Review*, 129(4), 784-801.
- Yoneyama, K., & Parsons, D. B. (1999). A proposed mechanism for the intrusion of dry air into the tropical western Pacific region. *Journal of the Atmospheric Sciences*, 56(11), 1524-1546.
- Yoon, J. H., & Huang, W. R. (2012). *02 Indian Monsoon Depression: Climatology and Variability*. InTech.
- Zuidema, P., Mapes, B., Lin, J., Fairall, C., & Wick, G. (2006). The interaction of clouds and dry air in the eastern tropical Pacific. *Journal of Climate*, 19(18), 4531-4544.

Chapter 4. Semidiurnal and diurnal variabilities of Zenith Wet Delay

4.1 Introduction

In Chapter 3, we investigated the intraseasonal variability of Zenith Wet Delay (ZWD) observed by Sumatran GPS Array (SuGAR) network. Intraseasonal variability of ZWD was attributed to two factors: monsoon strength and dry air intrusion. Together, two factors could explain 86.28% of the total variance. In this paper, we examine the same observational data, but shift to high-frequency variabilities: semidiurnal and diurnal variabilities of ZWD. The semidiurnal variability has a period of 12 hours, and the diurnal variability has a period of 24 hours. Semidiurnal and diurnal variabilities exist in various physical fields, such as the surface pressure and the rainfall. Since GPS ZWD is mainly influenced by the precipitable water vapor (PWV) in the atmosphere (Bevis et al. 1992), any physical phenomena that can modify the water vapor could also influence ZWD. We investigate the semidiurnal and diurnal variabilities of ZWD from the GPS observation in this paper.

It is long known that the semidiurnal variability is strong in the tropics. The semidiurnal variability can be linked with the atmospheric tide, which is the

regular change of total air mass driven by solar heating (Chapman and Lindzen 1970). According to the hydrostatic balance, the surface pressure can measure the change of the air mass. In the tropics, the semidiurnal component of the atmospheric tide is stronger than the diurnal component (Dai and Wang 1999). Based on the theory of atmospheric tides, the heating of ozone accounts for about two thirds of semidiurnal component, and the heating of water vapor and ground accounts for the rest (Butler and Small 1963). The atmospheric tide modifies the total air mass. Water vapor, as a component of the air, could also be modified, and so is ZWD. In this study, we will reveal the influence of atmospheric tides on the GPS observation of ZWD.

Diurnal cycles are also strong in the tropics, especially in the maritime continent where the SuGAR network is located. The maritime continent includes Indonesia, Philippines and Papua New Guinea, and is featured by many islands and peninsulas (Ramage, 1968). Due to the land-sea contrast, a land-sea breeze system can form in coastal regions (Simpson 1994; Houze et al. 1981). This local circulation changes direction during a day. In the afternoon, the sea breeze is mature and brings rainfall to the land. Later in the early morning, the land breeze well develops and the rainfall is over the sea. As revealed by several studies (Yang and Slingo 2001; Teo et.al. 2011), the convective system can further propagate into the ocean. This phenomenon is

evident in the observation, but debatable in the theory. Houze et al. (1981) linked it with the downdraft outflow, while Mapes et al. (2003) attributed the offshore propagation to the gravity wave. Since convections are able to modify PWV in the air column, the diurnal variation of ZWD could also manifest a propagation mode, which we will show in this paper.

Semidiurnal and diurnal variabilities also interact with other variabilities such as interannual and intraseasonal variabilities. It is important to exclude the influence of interannual and intraseasonal variabilities, so that we can focus on the semidiurnal and diurnal variabilities. ENSO (Barnett 1991) and IOD (Saji et al. 1999) are the most important interannual variabilities in the maritime continent. Year 2008 is thus picked as the sample study, because this year is both ENSO neutral and IOD neutral. The intraseasonal variability, which was studied in our last paper, would be removed by a filtering method. Semidiurnal and diurnal variabilities can also have seasonal variations. The maritime continent is the pathway of Australian-Asian monsoons (Chang et al. 2004; Chang et al. 2005). The monsoon can interact with the local circulation, and creates different forms of diurnal cycle. We only focus on the southwest monsoon season (including June, July, August, and September, according to Chase et al. (2003)), and leave the northeast monsoon season to the future study. This chapter, together with the last chapter, composes a thorough study

for the southwest monsoon season of 2008.

Traditionally, the study of semidiurnal and diurnal variabilities is confined by the observational platform. To resolve both variabilities, the frequency of the observation should at least be 3-hourly. Only limited number of observational methods can achieve such a high temporal resolution. For example, the study of atmospheric tides is traditionally based on the high-frequency observation of surface pressure. However, surface observations could hardly reflect all the atmospheric conditions beyond the ground. Radiosonde stations, which can provide the vertical profile of the atmosphere, are sparse in the maritime continent (Koh and Teo 2009). What is more, the balloons are usually released only twice a day, and thus cannot resolve the semidiurnal and diurnal variability. The Tropical Rainfall Measuring Mission (TRMM) could provide 3-hourly observations of tropical rainfall. TRMM satellite has greatly facilitated the research (e.g. Nesbitt and Zipser 2002), but other satellites could rarely be as qualified as TRMM for the purpose of studying semidiurnal and diurnal variabilities. In summary, new observational tools are needed to compensate the traditional platforms of the observation. In this paper, we present the GPS technology as a valid tool to study the semidiurnal and diurnal variabilities in the maritime continent.

In recent years, numerical models are increasingly used in the study of semidiurnal and diurnal variabilities (Clark et al. 2007; Koo and Hong 2010). Numerical models can provide high-resolution output of various 3D atmospheric fields. The performances of numerical models are not satisfactory regarding semidiurnal and diurnal variabilities (Sato, et al., 2009), and the complex atmospheric dynamics in the maritime continent poses a further challenge (Neale and Slingo 2003). The improvement of the simulation is hindered partly because of the lack of observational data for verification and diagnosis. New observational methods are needed to solve the modeling issues. We compare the GPS observation with the simulation from a regional model, in order to tell the strength and weakness of the numerical model. In our simulation, the numerical model could reasonably represent the diurnal cycle, but has a difficulty in simulating the semidiurnal variability. The reasons are discussed, and the conclusion may help the future development and setup of numerical modeling.

4.2 Data

In this study, we use the ZWD observation from the SuGAR network. Further details about the SuGAR network can be referred to Chapter 3. We focus on Southwest Monsoon season of 2008 (June, July, August, September), which is the same as the last paper. Out of 30 working GPS stations in 2008, 17 stations

are selected in consideration of data quality (Figure 4.1). The ZWD observation is processed to be hourly, and the details of the processing can be referred to Feng et al. (2015). With an hourly resolution, semidiurnal and diurnal variabilities can be fully resolved.

Figure 4.1 Stations used for the study of the semidiurnal and diurnal variabilities

Besides the observation of the SuGAr network, we also include the surface pressure observation from ground stations. The data are 3-hourly, and come from Integrated Surface Global Hourly Data (dataset DS-3505, <https://data.noaa.gov/dataset/integrated-surface-global-hourly-data>) provided by the National Climate Data Center (NCDC). The surface pressure can be used to study the atmospheric tide. Four stations near Sumatra are selected (Figure 4.2). According to Table 4.1, although there are gaps in the data, but the missing part is only a small fraction of the whole 3-hourly time

series. We fill the data gaps with linear interpolation.

Figure 4.2 Surface stations that measure the 3-hourly surface pressure

	No. of Measurements	Ratio of Observational Gap
KIJANG	940	3.69%
BATAM/HANG NADIM	891	8.71%
TABING	946	3.07%
SULTAN SYARIF KASIM II	881	9.73%

Table 4.1 Observation counts of the 3-hourly surface pressure observation.

Weather Research and Forecasting Model (WRF; Skamarock et al. 2005) is used to simulate Southwest monsoon season of 2008. WRF model is a Numerical Weather Prediction model widely used for regional climate

simulation. Two domains are nested (Figure 4.3). The outer domain has a horizontal resolution of 36 km, and the inner domain is 7.2 km. There are 37 vertical levels, with denser levels near the ground and the tropopause (Figure 2.5). The NCEP Climate Forecast System Reanalysis dataset (Saha et al. 2010) provides the initial condition and the lateral boundary condition for the outer domain. The model top is set to be 30hpa. The layers within 5 km from the model top are damped to avoid the reflection of gravity wave at the model top. In order to curb the growth of error for a long simulation, nudging is applied to the levels above 800hPa and excluding the planet boundary layer. We analyze the hourly output from the inner domain. For the inner domain, cumulus schemes are not used because they are generally designed for the lower resolution. WRF model outputs every hour, so it is able to resolve the semidiurnal and diurnal variabilities. ZWD is not a direct product of the numerical model, but we can calculate from the model outputs based on Bevis et al. (1992). Other setups of the simulation can be referred to Table 4.2.

Figure 4.3 Domains of the model. The red colored patch indicates the inner domain.

Setup	Option	Reference
Cumulus Scheme (no cumulus scheme in the inner domain)	Modified Betts-Miller-Janjic	Fonseca et al. (2015)
Microphysics Scheme	WRF double-moment five-class microphysics	Lim and Hong (2010)
Radiation Scheme	Rapid Radiative Transfer model for Global	Iacono et al. (2008)
Planetary Boundary Layer Scheme	Yonsei University PBL	Hong et al. (2006)
Surface Scheme	Monin-Obukhov surface layer	Monin and Obukhov (1954)
Land Surface Scheme	Four-layer Noah land surface model	Chen and Dudhia (2001)

Table 4.2 Model Setup

4.3 Method

Since this paper aims at studying the semidiurnal and diurnal variabilities, we use a high-pass Lanczos filter (Lanczos, 1956) to preprocess all the data. The

high-pass Lanczos filter can remove the low frequency variability. Compared with other filters, Lanczos filter has the advantage of suppressing the artificial Gibbs oscillation (Duchon, 1979). The cut-off frequency of the Lanczos filter is 48 hours, and the window size of the Lanczos kernel is 241, equivalent to 10 days. The filtered result is equal to the convolution of the filter's kernel and the original time series. After the filtering, the multiple-day variability is damped, and semidiurnal and diurnal variabilities are preserved, as shown in the response function (Figure 4.4).

Figure 4.4 The response function of the Lanczos filter used. I use the filter to retain the semidiurnal and diurnal variabilities.

We first analyze the semidiurnal and diurnal variabilities in the time series.

The most important information of certain variability is its amplitude and phase. The Fourier analysis is used to transform the time series to a set of trigonometric basis (Wilks, 2011). We can then retrieve the amplitudes and the phases of the trigonometric components. We use the UTC time of the first peak in the day to denote the phases of diurnal and semidiurnal variability. Although local solar time is often used in the study of diurnal and semidiurnal variability, the choice of UTC time allows us to conveniently compare with other data sources that are recorded with UTC time, such as the surface pressure.

The SuGAR network contains 17 stations around Sumatra, and thus is able to capture the spatial variation of ZWD. We use the rotated EOF analysis to identify the leading components of the multivariate data. Rotated EOF is a variant EOF analysis (Jackson 1991), which is used to identify the components based on the spatial coherence. The rotated EOF analysis would further rotate the leading principal components (PCs). The rotation is based on the varimax criterion that maximizes the sum of the spatial variances of the eigenfunctions (Richman 1986). The same method was already used in Chapter 3.

After the rotated EOF analysis, each PC can be studied from two aspects: spatially through the eigenfunctions, and temporally through the time scores. Our ultimate goal is to find the physical meaning of the leading PCs.

Cross-correlation (Chatfield 2013) is used to find the time lag between the time series. The daily composite works as a filter to preserve the diurnal variability and its harmonics. We use Pearson's correlation coefficient to measure the similarity between the eigenfunctions.

4.4 Results and Discussion

4.4.1 Semidiurnal Variability

Figure 4.5 Power spectrums of station PBLI. Above: the power spectrum before filtering. Below: the power spectrum after filtering

As an example, Figure 4.5 shows the power spectrum of station PBLI. We can see that the Lanczos filter damps the low-frequency signal. The power spectrum shows two peaks, corresponding to semidiurnal and diurnal frequencies, indicating the importance of the semidiurnal and diurnal variabilities. In order to reveal the general information of all the stations, we plot the amplitudes and phases of all stations in a polar coordinate. Figure 4.6 shows the distribution of diurnal components. Each dot is one station, with the radius as the amplitude and the angle as the phase. Each dashed line links corresponding dots from the observation and the simulation. The amplitudes and the phases of the diurnal variability vary among the stations, but the simulation could generally agree with the observation. Two stations, LAIS and MNNA, have exceptionally high amplitudes of the diurnal variabilities. These two stations are located in the Southern coast of Sumatra.

Figure 4.6 The distribution of the phases and the amplitudes of the diurnal components. The red colored dots are the stations in the model, while the blue dots came from the direct observation of the SuGAR network. A dashed line connects each corresponding pair.

If we turn to the distribution of the semidiurnal variability in Figure 4.7, the simulation is very different with the observation. In the simulation, the amplitudes vary a lot among the stations, and the phases are scattered in all directions. In contrast, most observational stations own similar amplitudes of the semidiurnal variability, and the amplitudes are stronger than the simulation.

Figure 4.7 Same as Figure 4.6, but for the semidiurnal components.

Except two stations that peak around 0UTC, the phases of the other stations are confined in a narrow range from 3UTC to 5UTC. The phases of the semidiurnal variabilities are consistent with previous studies of PWV (Pramualsakdikul et al. 2007). It is worth noting that those two abnormal stations, i.e. LAIS and MNNA, have a strong diurnal variability. Since diurnal variation could include a semidiurnal component as a higher harmonics, it is possible that higher harmonics of diurnal variability is not clearly separated in the power spectrum, and cause the discrepancy in semidiurnal variability.

The uniform distribution of the semidiurnal variability could be attributed to

the large-scale effect of a migrating atmospheric tide. The atmospheric tide is a function of the local solar time, and migrating atmospheric tides are not related with the longitude (Chapman and Lindzen 1970). Since our stations only varies 1 hour in solar time, we can expect a uniform impact from the semidiurnal atmospheric tide. Jin et al. (2008) have demonstrated that the atmospheric tide can cause a semidiurnal variation in GPS observation, although their study did not differentiate the neutral component and the wet component of the atmosphere. The atmospheric tide essentially changes the air mass in the air column. Water vapor is a component of the air. If there is more air mass in the air column, we can expect more water vapor in the air column too. If the above hypothesis is true, the observed ZWD should have a synchronized semidiurnal variation with the semidiurnal atmospheric tide. In another word, the phase of the semidiurnal ZWD should be similar to the phase of the semidiurnal atmospheric tide.

In order to test, we retrieve the phase of the semidiurnal atmospheric tide from the observation of the surface pressure, and compare it with the phase of the semidiurnal ZWD. The stations for the surface pressure do not overlap the GPS stations, but they are within the spatial area of the SuGAR network. As argued before, the network spanned a short distance in the longitudinal direction, so the effect of the semidiurnal atmospheric tide should be uniform

over the network. As a result, the semidiurnal phase of the surface pressure should be similar to the phase of the GPS ZWD. We apply the Fourier analysis to the surface pressure. The semidiurnal oscillations reach their peaks around 3:30UTC (Table 4.3). The peaking time agrees with the observation of surface pressure (Dai and Wang 1999). Furthermore, the semidiurnal atmospheric tide is in phase with the semidiurnal ZWD variability because they all peak around 3:30UTC.

Stations	Phase in observation (UTC)	Amplitude in observation (hpa)	Phase in simulation (UTC)	Amplitude in simulation (hpa)
KIJANG	3.23	0.42	0.96	0.15
BATAM/HANG NADIM	3.20	0.55	1.09	0.15
TABING	3.42	0.59	0.62	0.17
SULTAN SYARIF KASIM II	3.26	0.58	0.62	0.25

Table 4.3 Amplitudes and phases of the semidiurnal surface pressure

Figure 4.8 Amplitudes and phases of the semidiurnal surface pressure (as the data in Table 4.3)

As noted before, the numerical model could not reproduce the semidiurnal ZWD variability very well. Actually, the numerical model also performs badly in simulating the semidiurnal atmospheric tide. As shown in Table 4.3, the simulated amplitudes are weaker than those from the observation, and the phases are also quite different. I speculate that the reason of the difference could root from the truncation of the model top. It is known that the ozone heating is a major driving force for the semidiurnal atmospheric tide (Lindzen, 1967). The ozone layer mainly extends from 30hPa to 0.3hPa (Paul, et al., 1998), and the ozone heating maximizes near the stratopause (Hargreaves,

1992). The top of our numerical model is only 30hPa, much below the stratopause. Our simulation thus cannot account for a large portion of the ozone heating. The stratospheric contribution to the atmospheric tide is not resolved, and thus the semidiurnal ZWD from the model may contain some error. I hope to confirm this speculation in the future study, with another model run that accounts for the ozone layer.

Although the numerical model fails, we can statistically model the influence of the semidiurnal atmospheric tide on the GPS ZWD. First, we calculate the mean amplitude of the semidiurnal components. Secondly, the phases of semidiurnal components are adjusted according to the time difference with 100°E longitude, and then the mean phase is calculated. During the last two steps, we exclude the abnormal stations, i.e., MNNA and LAIS. The mean amplitude is 0.00168m, and the mean phase is 4.11UTC. Again, this phase is quite close to the phase of the semidiurnal atmospheric tide. A semidiurnal oscillation with the above calculated amplitude and phase would represent the effect of atmospheric tide on ZWD at 100°E longitude. We then reconstruct the semidiurnal component for different longitudes by taking into account of the local solar time. The modeled semidiurnal ZWD can be expressed by the following function:

$$ZWD_{semidiurnal} = 0.00168m \times \cos \left[\frac{\pi}{6} \times \left(t + \frac{x}{15} - 10.78 \right) \right] \quad \text{Equation 4.1}$$

Where t is the UTC time, x is the longitude.

4.4.2 Diurnal Variability

The advantage of a dense GPS network is its spatial resolution. By knowing the spatial variation, we can gain further insight into the physics process. We still use the rotated EOF analysis method, which is already used in our last section. We do the rotated EOF analysis to the multivariate data of ZWD from observation and the simulation. We have already known the discrepancy of the semidiurnal variability between the observation and the simulation. In order to prevent the discrepancy interfere the following analysis, I remove the modeled semidiurnal component (Equation 4.1) from the observed ZWD. At the same time, we keep the simulated ZWD untouched. For both observation and simulation, we get two leading PCs, as shown in Table 4.4. The percentages of variance explained by the leading PCs are around 20%, and are much larger than the following PCs. According to the similarity between the observation and the simulation, we pair the PCs into two groups, and name them as XPC and YPC.

	Observation	Simulation
XPC	18.8%	22.9%

YPC	18.8%	17.1%
PC3 (which explains the largest variance among the following PCs)	11.4%	7.4%

Table 4.4 Percentage of total variance explained by XPC and YPC

We first look into the XPC. The eigenfunction of the observation is much the same with that of the simulation (Figure 4.9). The Pearson's correlation coefficient of the loadings is 0.96, indicating the strong similarity between the observation and the simulation. Stations on the main island of Sumatra share the same sign. These stations have high loadings, especially for those located in the southern part of Sumatra. Offshore stations can be separated into two groups and oscillated oppositely. Northern stations have a weak and negative loading. The time scores of XPC are also in phase. If we calculate the cross-correlation (Figure 4.10), the observation only leads the simulation by 1 hour. Weighing the fact that the data frequency is hourly, the time difference of 1 hour is trivial.

Figure 4.9 The eigenfunctions of XPC. The left figure is from the observation, and the right figure from the simulation.

Figure 4.10 The cross-correlation of the time scores of XPC.

We then investigate the YPC. For the eigenfunctions (Figure 4.11), the Pearson's correlation coefficient is 0.85, although not as high as XPC, but still significant enough. For the observation, most stations have the same sign, and most offshore stations have high loadings. For the simulation, several stations

flip the sign, but these stations do not have high loadings anyway. In the simulation, most offshore stations also own high loadings, and these offshore stations keep the same sign as those in the observation. The time score of the observation is again in phase with the simulation's time score. The cross-correlation again demonstrates 1 hour leading by the observation (Figure 4.12). In all, the simulation reproduces the two leading PCs very well.

Figure 4.11 Same as Figure 4.9, but for YPC

Figure 4.12 Same as Figure 4.10, but for the time scores of YPC

The daily composite of the time score can roughly show the variability of the data. Figure 4.13 shows the daily composite of the time scores from the observation. The curves corresponding to XPC and YPC have a diurnal variation with only one peak in a day, so they are strongly influenced by the diurnal variability. The amplitudes of the daily composite are around 1. Remember the fact the variance of the whole time score is 1, the amplitude of the daily composite is on the same order as the time score itself. This reveals the dominant role played by diurnal variability in XPC and YPC.

Figure 4.13 Daily composite of the time scores for the observation

The conclusion, that diurnal variabilities are mainly explained by XPC and YPC, can also apply to daily composites of the simulated time scores (Figure 4.14). In fact, the daily composite is equivalent to a filter that preserves the diurnal variability and its harmonics (Bernhardt and Schlapp 1976). If the time score is not dominated by a diurnal oscillation or a harmonic of 24 hours, the daily composite will suppress the signal and result in a much weaker variance. This is what happens in the non-leading PCs, as shown in Figure 4.13 and Figure 4.14. The amplitudes of daily composites are much suppressed for the time scores of the non-leading PCs, so the diurnal variability does not contribute

much to the following PCs. In conclusion, XPC and YPC reveal main aspects of the diurnal cycle, and the simulation agrees the observation quite well.

Figure 4.14 Same as Figure 4.13, but for the simulation

4.4.3 Propagation of Diurnal ZWD Cycle

Since the numerical model can well capture the diurnal cycle, we can rely on the numerical model to understand the observation. At this point, we migrate to a continuous region along the west coast of Sumatra, as the color-shaded regions in in Figure 4.15. This region contains most stations of the SuGAR network. The numerical model can simulate ZWD and other atmospheric fields

for the region. Since the horizontal resolution of the numerical model is 7.2km, the model output provides a much denser sampling of the region. In this way, the physical interpretation would not be subject to the specific locations of GPS stations.

Figure 4.15 The eigenfunctions of the leading PCs

We get the ZWD output from the numerical model, and conduct the rotated EOF analysis again. There are two leading PCs. The percentages of the total variances explained by two leading PCs are similar, which are 15.2% and 13.5% respectively. The spatial patterns of the eigenfunctions (Figure 4.15) generally orient along the coastline, hinting the impact of the topography. XPC has positive anomaly on west coast of the main island (Figure 4.15a), and it has negative anomaly further west. YPC has the positive anomaly along the adjacent sea (Figure 4.15b). The positive stripe of YPC lies in between the positive and negative stripes of XPC. From the daily composites of the time

scores (Figure 4.16), the diurnal variability is prominent in both PCs. XPC time score peaks around 10UTC, corresponding the local afternoon, and YPC peaks in the mid-night.

Figure 4.16 The daily composites of the time scores

According to cross-correlation, the time scores for the leading PCs are in a quadrature relationship. XPC leads YPC by 6 hours. If we combine the two leading PCs, they show a propagating signal. At the peaking time of XPC's time score, the west coast of Sumatra first gains high ZWD in local afternoon as shown by XPC's eigenfunction. Then 6 hours later, the time score of XPC becomes close to zero, and YPC's time score reaches its peak. The positive

anomaly is now over the sea adjacent to the coast, as shown by YPC's eigenfunction. Another 6 hours later, XPC's time score reaches its valley, and YPC's time score is around zero. The negative patch in XPC's eigenfunction, i.e. the region near the west boundary of the domain, is where the positive anomaly is. Put different phases together, XPC and YPC show a gradual propagation of ZWD anomaly. Because ZWD is proportional to PWV, what we see is also a propagation of positive PWV anomaly. The timing is consistent with previous GPS studies, which showed the diurnal cycle of PWV peaks in the afternoon on the land (Wu 2003; Jin and Luo 2009; Sato et al. 2009). Interestingly, the leading PCs of ZWD are very similar to the rainfall PCs in Teo et al. (2011), so the propagation of diurnal ZWD is probably coupled with the propagation of the convective system. The diurnal rainfall and ZWD both peak in the afternoon along the coast of Sumatra, and then propagate offshore.

Several mechanisms can lead to the variation of water vapor, and cause the change of ZWD, but some of them cannot explain the propagation characteristics. Evaporation can supply the water vapor (Dai et al. 1999a; Santanello and Friedl 2003), but the diurnal variation of evaporation is very weak over the ocean. Another mechanism, i.e. large-scale vertical motion mentioned by Dai et al. (1999b), should also be excluded because XPC and YPC show a sharp variation in a small area. With the advection of the low-level

wind, the diurnal PWV variation can propagate, as shown by Sato and Kimura (2003). However, the monsoon wind is southwesterly in the northern part of Sumatra. The southwesterly wind should transport PWV inland, which contradicts the offshore transportation in our study. As a result, the advection is not the reason for the propagation.

In the end, low-level moisture convergence (Sato and Kimura 2005) is probably the mechanism for the diurnal ZWD. We already mentioned, ZWD is coupled with the convective system. The low-level convergence in the convective system can enhance the PWV (Sato and Kimura 2005). As we have mentioned, the land-sea contrast of Sumatra can nurture a land-sea breeze system and produce the diurnal rainfall (Yang and Slingo 2001). The convections take place in the coastal region in the afternoon and then propagate offshore (Mori et al. 2004). This is different with the traditional view of land-sea breeze that the convergence zone switches its location between the coast and the sea. Instead, the convergence zone would form along the coast first, and then gradually propagate into the sea and further away. The low-level convergence of the convective system can then produce a propagating signal of ZWD. This mechanism provides a possible explanation for propagating diurnal signals in the observation and the simulation.

4.5 Conclusion

We study the observation of ZWD from the SuGAR network. Semidiurnal and diurnal variability dominates the high-frequency spectrum. Semidiurnal variability is relatively uniform in space and varies simultaneously with the atmospheric tide, indicating the influence of the atmospheric tide on the wet component of the air. The numerical model failed to simulate the semidiurnal variability. I speculate that the missing ozone layer play a role in the failure of the simulation, but further work need to be done to confirm this. The observation and the numerical model agree well on the diurnal variability. A propagating pattern is revealed, which can be linked to the propagation of convective diurnal cycle.

4.6 References

- Barnett, T. P. (1991). The interaction of multiple time scales in the tropical climate system. *Journal of Climate*, 4(3), 269-285.
- Bernhardt, P. A., & Schlapp, D. M. (1976). Suppression of the unwanted variations in data. *Journal of Geophysical Research*, 81(19), 3428-3430.
- Bevis, M., Businger, S., Herring, T. A., Rocken, C., Anthes, R. A., & Ware, R. H. (1992). GPS meteorology: Remote sensing of atmospheric water vapor using the Global Positioning System. *Journal of Geophysical Research*:

Atmospheres (1984–2012), 97(D14), 15787-15801.

- Butler, S. T., & Small, K. A. (1963). The excitation of atmospheric oscillations. *Proceedings of the Royal Society of London. Series A. Mathematical and Physical Sciences*, 274(1356), 91-121.
- Chang, C. P., Harr, P. A., McBride, J., & Hsu, H. H. (2004). Maritime continent monsoon: annual cycle and boreal winter variability. *East Asian Monsoon*. World Scientific.
- Chang, C. P., Wang, Z., McBride, J., & Liu, C. H. (2005). Annual cycle of Southeast Asia-Maritime Continent rainfall and the asymmetric monsoon transition. *Journal of climate*, 18(2), 287-301.
- Chapman, S., & Lindzen, R. S. (1970). *Atmospheric tides: thermal and gravitational*. Springer Science & Business Media.
- Chase, T. N., Knaff, J. A., Pielke Sr, R. A., & Kalnay, E. (2003). Changes in global monsoon circulations since 1950. *Natural Hazards*, 29(2), 229-254.
- Chatfield, C. (2013). *The analysis of time series: an introduction*. CRC press.
- Chen, F., & Dudhia, J. (2001). Coupling an advanced land surface-hydrology model with the Penn State-NCAR MM5 modeling system. Part I: Model implementation and sensitivity. *Monthly Weather Review*, 129(4), 569-585.
- Clark, A. J., Gallus Jr, W. A., & Chen, T. C. (2007). Comparison of the diurnal precipitation cycle in convection-resolving and non-convection-resolving mesoscale models. *Monthly Weather Review*, 135(10), 3456-3473.

- Dai, A., & Wang, J. (1999). Diurnal and semidiurnal tides in global surface pressure fields. *Journal of the Atmospheric Sciences*, 56(22), 3874-3891.
- Dai, A., Trenberth, K. E., & Karl, T. R. (1999a). Effects of clouds, soil moisture, precipitation, and water vapor on diurnal temperature range. *Journal of Climate*, 12(8), 2451-2473.
- Dai, A., Giorgi, F., & Trenberth, K. (1999b). Observed and model simulated precipitation diurnal cycles over the continental United States. *Journal of Geophysical Research*, 104, 6377-6402.
- Duchon, C. E. (1979). Lanczos filtering in one and two dimensions. *Journal of Applied Meteorology*, 18(8), 1016-1022.
- Feng, L., Hill, E. M., Banerjee, P., Hermawan, I., Tsang, L. L., Natawidjaja, D. H., ... & Sieh, K. (2015). A unified GPS-based earthquake catalog for the Sumatran plate boundary between 2002 and 2013. *Journal of Geophysical Research: Solid Earth*.
- Fonseca, R., Zhang, T., & Koh, T. Y. (2015). Improved simulation of precipitation in the tropics using a modified BMJ scheme in WRF model. *Geosci. Model Dev. Discuss.*, 8, 4019-4049
- Hargreaves, J. K. (1992). *The solar-terrestrial environment*. Cambridge Press.
- Houze Jr, R. A., Geotis, S. G., Marks Jr, F. D., & West, A. K. (1981). Winter monsoon convection in the vicinity of North Borneo. Part I: Structure and time variation of the clouds and precipitation. *Monthly Weather Review*, 109(8),

1595-1614.

- Hong, S. Y., Noh, Y., & Dudhia, J. (2006). A new vertical diffusion package with an explicit treatment of entrainment processes. *Monthly Weather Review*, 134(9), 2318-2341.
- Iacono, M. J., Delamere, J. S., Mlawer, E. J., Shephard, M. W., Clough, S. A., & Collins, W. D. (2008). Radiative forcing by long-lived greenhouse gases: Calculations with the AER radiative transfer models. *Journal of Geophysical Research: Atmospheres*, 113(D13).
- Jackson, J.E. (1991). *A user's guide to principal components*. John Wiley & Sons, New York.
- Jin, S., Wu, Y., Heinkelmann, R., & Park, J. (2008). Diurnal and semidiurnal atmospheric tides observed by co-located GPS and VLBI measurements. *Journal of Atmospheric and Solar-Terrestrial Physics*, 70(10), 1366-1372.
- Jin, S., & Luo, O. F. (2009). Variability and climatology of PWV from global 13-year GPS observations. *Geoscience and Remote Sensing, IEEE Transactions on*, 47(7), 1918-1924.
- Koh, T. Y., & Teo, C. K. (2009). Toward a mesoscale observation network in Southeast Asia. *Bulletin of the American Meteorological Society*, 90(4), 481-488.
- Koo, M. S., & Hong, S. Y. (2010). Diurnal variations of simulated precipitation over East Asia in two regional climate models. *Journal of*

Geophysical Research: Atmospheres (1984–2012), 115(D5).

- Lanczos, C. Applied analysis. 1956.
- Lim, K. S. S., & Hong, S. Y. (2010). Development of an effective double-moment cloud microphysics scheme with prognostic cloud condensation nuclei (CCN) for weather and climate models. *Monthly Weather Review*, 138(5), 1587-1612.
- Lindzen, R. S. (1967). Thermally driven diurnal tide in the atmosphere. *Quarterly Journal of the Royal Meteorological Society*, 93(395), 18-42.
- Mapes, B. E., Warner, T. T., & Xu, M. (2003). Diurnal patterns of rainfall in northwestern South America. Part III: Diurnal gravity waves and nocturnal convection offshore. *Monthly Weather Review*, 131(5), 830-844.
- Monin, A. S., & Obukhov, A. (1954). Basic laws of turbulent mixing in the surface layer of the atmosphere. *Contrib. Geophys. Inst. Acad. Sci. USSR*, 151, 163-187.
- Mori, S., Jun-Ichi, H., Tauhid, Y. I., Yamanaka, M. D., Okamoto, N., Murata, F., ... & Sribimawati, T. (2004). Diurnal land-sea rainfall peak migration over Sumatera Island, Indonesian maritime continent, observed by TRMM satellite and intensive rawinsonde soundings. *Monthly Weather Review*, 132(8), 2021-2039.
- Neale, R., & Slingo, J. (2003). The maritime continent and its role in the global climate: A GCM study. *Journal of Climate*, 16(5), 834-848.

- Nesbitt, S. W., & Zipser, E. J. (2003). The diurnal cycle of rainfall and convective intensity according to three years of TRMM measurements. *Journal of Climate*, 16(10), 1456-1475.
- Paul, J., Fortuin, F., & Kelder, H. (1998). An ozone climatology based on ozonesonde and satellite measurements. *Journal of Geophysical Research: Atmospheres*, 103(D24), 31709-31734.
- Pramualsakdikul, S., Haas, R., Elgered, G., & Scherneck, H. G. (2007). Sensing of diurnal and semi-diurnal variability in the water vapour content in the tropics using GPS measurements. *Meteorological Applications*, 14(4), 403-412.
- Ramage, C. S. (1968). Role of a tropical “maritime continent” in the atmospheric circulation. *Monthly Weather Review*, 96(6), 365-370.
- Richman, M. B. (1986). Rotation of principal components. *Journal of Climatology*, 6(3), 293-335.
- Saha, S., Moorthi, S., Pan, H. L., Wu, X., Wang, J., Nadiga, S., ... & Reynolds, R. W. (2010). The NCEP climate forecast system reanalysis. *Bulletin of the American Meteorological Society*, 91(8), 1015-1057.
- Saji, N. H., Goswami, B. N., Vinayachandran, P. N., & Yamagata, T. (1999). A dipole mode in the tropical Indian Ocean. *Nature*, 401(6751), 360-363.
- Santanello Jr, J. A., & Friedl, M. A. (2003). Diurnal covariation in soil heat flux and net radiation. *Journal of Applied Meteorology*, 42(6), 851-862.

- Sato, T., & Kimura, F. (2003). A two-dimensional numerical study on diurnal cycle of mountain lee precipitation. *Journal of the Atmospheric Sciences*, 60(16), 1992-2003.
- Sato, T., & Kimura, F. (2005). Diurnal cycle of convective instability around the central mountains in Japan during the warm season. *Journal of the Atmospheric Sciences*, 62(5), 1626-1636.
- Sato, T., Miura, H., Satoh, M., Takayabu, Y. N., & Wang, Y. (2009). Diurnal cycle of precipitation in the tropics simulated in a global cloud-resolving model. *Journal of Climate*, 22(18), 4809-4826.
- Simpson, J. E. (1994). *Sea breeze and local winds*. Cambridge University Press.
- Skamarock, W. C., Klemp, J. B., Dudhia, J., Gill, D. O., Barker, D. M., Wang, W., & Powers, J. G. (2005). *A description of the advanced research WRF version 2* (No. NCAR/TN-468+ STR). National Center For Atmospheric Research Boulder Co Mesoscale and Microscale Meteorology Div.
- Teo, C. K., Koh, T. Y., Chun-Fung Lo, J., & Chandra Bhatt, B. (2011). Principal component analysis of observed and modeled diurnal rainfall in the Maritime Continent. *Journal of Climate*, 24(17), 4662-4675.
- Wilks, D. S. (2011). *Statistical methods in the atmospheric sciences*. Academic press.
- Wu, P., Hamada, J. I., Mori, S., Tauhid, Y. I., Yamanaka, M. D., & Kimura, F.

- (2003). Diurnal variation of precipitable water over a mountainous area of Sumatra Island. *Journal of Applied Meteorology*, 42(8), 1107-1115.
- Yang, G. Y., & Slingo, J. (2001). The diurnal cycle in the tropics. *Monthly Weather Review*, 129(4), 784-801.

Chapter 5. Conclusions

In this thesis, I focus on the study of GPS observation. As a new technology, the GPS device can frequently observe ZWD that is proportional to the total water vapor in the air column. Our observational data come from the SuGAR network located in Sumatra, and contain valuable information about the tropical climate. The southwest monsoon season of 2008 is targeted.

The variabilities of two ranges are studies. I separated the variabilities with the filtering method. The intraseasonal variability is studied in Chapter 3. With the method of the rotated EOF analysis, I objectively retrieved two components that can be attributed to different dynamical processes. The first component explains a major part of the total variance. It has a close relationship with the strength of the monsoon, and shares common characteristics with BSISO. Moreover, the temporal evolution of this component can correlate well with the IMI. It is shown that the southerly anomaly of moisture flux caused by the strengthened monsoon encourages the transportation of water vapor from Sumatra to the Bay of Bengal, which conclusion is also reached by previous studies. My work not only confirms the impact of the monsoon on the water vapor, but also highlights the quantified importance of the process. The other component explains a minor but still significant part of the total variance. The source of the variation is the intrusion of dry air from the Southern

Hemisphere. The intrusion is closely related with the Rossby wave in the subtropical region. When it arrives in the tropics, the dry air from the high level descends to the middle level. The dynamical process was discovered before in the study of tropical-extratropical air exchange, but the equatorward transportation received less attention. The equatorward transportation of the dry air should receive more attention in the future studies because of its role in the intraseasonal variability, as pointed out by my work.

I also studied the variability with high frequency. The semidiurnal and diurnal variabilities are prominent in the tropical climate. Limited observational platforms can resolve the semidiurnal and diurnal variabilities. The GPS device is an exception because of its capacity to observe the atmosphere. The SuGAR network has made frequent and continuous observations of ZWD, which can be used for the study of the high-frequency variabilities of water vapor. In addition, I ran the WRF model as the supplement. The purpose is to downscale the low-frequency CFSR data to provide hourly output. Special efforts are taken to make sure the model can get the simulation result similar to the CFSR data in a large scale, including the careful choices of the physical schemes and the innovation of the cumulus scheme. Furthermore, nudging is also used to control the growth of the error. The simulation result correlates well with the GPS observation, and the propagating process of the diurnal cycle is also

captured, which give me the confidence to explain the GPS observation with the downscaling of the numerical model.

However, the numerical model failed to simulate the semidiurnal variability. The strong and uniform signal of the semidiurnal variability is related with the semidiurnal atmospheric tide, and the semidiurnal variability of the observed ZWD synchronizes with that of the surface pressure. Considering that the heating of the ozone is the main cause of the semidiurnal mode, the exclusion of a large part of the ozone layer in the model setup is probably the reason for the failure. This finding highlights the necessity of the ozone layer for successful simulation of atmospheric tides. After removing the semidiurnal oscillations from the GPS observation, the simulation result is then consistent with the observation. Two components are captured, and they are orthogonal both in space and time. Based on the continuous domain of the numerical model, the two components actually indicate a seaward propagation of the diurnal ZWD cycle. Similar observations of the propagation were made, but mostly regarding the cloud and rainfall. This is the first direct observation of the propagation of the water vapor in a convective system of the diurnal cycle.

The SuGAR network has actually accumulated a wealth for the study of tropical atmosphere. My study is only the starting point of the digging. The GPS

network can be strengthened if appropriate meteorological devices are installed at the stations. Anyway, I have found useful methods to tackle the ZWD observation in a dense network, including the lead-lag correlation and multivariate analysis. The reanalysis data and the numerical model are also able to compensate the GPS observation. Similar procedures can be applied to other seasons and years. Moreover, the atmospheric processes highlighted in the study can be further pursued, since they have shown their importance in the GPS observation.

How to make use of the GPS observation in the weather forecast is still an open question, but at least the GPS device can be used for the monitoring purpose. I have found the typical patterns of various processes, so reversely it is possible to measure the strength of each atmospheric process solely based on the SuGAR network. This will not only help the study of regional climate, but can be far-reaching because phenomena such the monsoon and the dry air intrusion can influence a large area. Since the GPS technology has special advantages over other traditional observational methods and the technology is developing fast, the integration of the GPS observation has great potential in improving the weather study and the weather forecast in the tropics.

Acronym

BMJ: Betts-Miller-Janjić (scheme)

BSISO: Boreal Summer Intraseasonal Oscillation

ENSO: El Niño Southern Oscillation

EOF: Empirical Orthogonal Function

GPS: Global Positioning System

IOD: Indian Ocean Dipole

MJO: Meddan-Julian Oscillation

NCDC: National Climate Data Center

NCEP: National Centers for Environmental Prediction

PC: Principal Component

PWV: Precipitable Water Vapor

RRTMG: Rapid Radiative Transfer Model for GCMs

SuGAR: Sumatran GPS Array

TEC: Total Electron Content

TRMM: Tropical Rainfall Measuring Mission

WRF: Weather Research and Forecasting (model)

ZHD: Zenith Hydrostatic Delay

ZWD: Zenith Wet Delay

Appendix. Orthogonality of EOF Analysis and Rotated EOF Analysis

In this thesis, I rely on the rotated EOF analysis to find out the patterns in the data. The rotated EOF analysis is a variant of the EOF analysis. Both methods reexpress the data to be the linear combinations of principal components. For the EOF analysis, the eigenfunctions of the PCs must be orthogonal, and the time scores are independent with each other. For the rotated EOF analysis, the eigenfunctions are not necessarily independent, while the time scores are independent with each other. In the Appendix, I will go to the mathematical details of the orthogonality.

We start from the EOF analysis. EOF analysis represents the data matrix as a linear combination of principal components. A data matrix can be written as:

$$Z = XA$$

X is the data matrix, and it has a dimension of $N \times P$. N is the length of the time series, corresponding to the length of GPS observations. P is the number of the locations, corresponding to the number of GPS stations. A is a $P \times P$ matrix, including the spatial patterns of the principal components. Z is a $N \times P$ matrix, including the time scores of the principal components.

In my analysis, I add the constraint that the variances of time scores to be 1. As

a result, A is orthogonal, and Z are orthonormal:

$$A^T A = S^{-2}, Z^T Z = I$$

where S^2 is the diagonalized matrix of $X^T X$. Obviously, without rotation, the EOF analysis would generate orthogonal time scores and orthogonal spatial patterns.

The rotation can be represented by matrix T , so $T^T T = I$. After the rotation, $Z^* = ZT$, $A^* = AT$. A^* are the rotated spatial patterns, and Z^* are the rotated time scores.

With the above condition, we have:

$$A^{*T} A^* = T^T S^{-2} T, Z^{*T} Z^* = I$$

In another word, after the rotation, the time scores are still orthogonal, and each time score has a variance of 1. However, the orthogonality is no more guaranteed.

Reference

- Mestas-Nunez, A. M. (2000). Orthogonality properties of rotated empirical modes. *International Journal of Climatology*, 20(12), 1509-1516.