

Imperial College London
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**ANALYSIS OF THE OUTGOING
LONGWAVE RADIATION
(OLR) IN DIFFERENT
TIMESCALES OVER AFRICA
AND ATLANTIC OCEAN**

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Doctor of Philosophy in Physics of Imperial College London

Declaration of Originality

I declare this thesis is my original research work, all else has been acknowledged and referenced. The relevant references have been documented in the bibliography section of the document.

Kwun Pan Chan

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Abstract

The subject of the thesis is the study of the Outgoing Longwave Radiation (OLR) in different timescale over the African and Atlantic region. We have used geostationary satellite data (GERB) to examine some of the shorter timescale radiative variability by utilising data with much higher time resolution than in previous work.

The work begins with Fourier analysis of the OLR data to identify the most significant modes of variability within the OLR in short time scales, and observe such modes in spatial terms. These modes (semi-diurnal, diurnal, 2-3 days, 4-7 days) are then independently tested using various methods, to attribute them to the appropriate climate processes.

The diurnal cycle has been decomposed using Empirical Orthogonal Function (EOF) technique, and we have managed to separate various diurnal components according to their variance contributions in the dataset. We have found that various conditions can alter the variance contribution, the magnitude and the time phase of the diurnal cycle and its components.

We have attempted to investigate the diurnal variability over different scenes (surface types and clear/cloudy sky) using Principal Component Analysis(PCA). A time delay (2 hours) has been found between the peaks of the components over different surface types.

We have developed a simple radiation model using incident solar radiation (ISR) from GERB, surface properties and a single layer atmosphere. The results were analysed using PCA and we have found a delay of 1 hour in our model. We have investigated the possible causes for the discrepancies between the model components and the OLR data components. We have then conducted regional tests to attempt to test the possible causes.

Glossaries

ADM - Angular Distribution Models

AEJ - African Easterly Jet

AEW - African Easterly Wave

ASR - Absorbed Solar Radiation

AWD - African Wave Disturbances

CERES - Clouds and Earth's Radiant Energy System

CLAUS - (European Union) Cloud Archive User Service

COARE - Coupled Ocean Atmosphere Response Experiment

CRF - Cloud Radiative Forcing

DFT - Discrete Fourier Transform

DIV - Divergence

EOF - Empirical Orthogonal Function

ECMWF - European Centre for Medium-Range Weather Forecasts

ERA - ECMWF global atmospheric Re-Analysis

ERBE - Earth Radiation Budget Experiment

FFT - Fast Fourier Transform

FGGE - First GARP Global Experiment

FT - Fourier Transform

GARP - Global Atmospheric Research Program

GATE - GARP Atlantic Tropical Experiment

GCM - General Circulation Model

GERB - Geostationary Earth Radiation Budget (Experiment)

HC - Heat Capacity
IOP - Intensive Observing Period
IR - Infra-Red
IRIS - Infra-Red Interferometer Spectrometer
ISR - Incident Solar Radiation
ITCZ - Intertropical Convergence Zone
LST - Local Solar Time
LW - Long Wave
MODIS - Moderate Resolution Imaging Spectroradiometer
MSG - MeteoSat Second Generation satellite
MTH - Middle Tropospheric Humidity
NaN - Not a Number
NASA - National Aeronautics and Space Administration
NCAR - National Center for Atmospheric Research
NCEP - National Centers for Environmental Prediction
NOAA - National Oceanic and Atmospheric Administration
OLR - Outgoing Longwave Radiation
PC - Principal Component
PCA - Principal Component Analysis
RF - Radiative Forcing
RSR - Reflected Solar Radiation
SBDART - Santa Barbara DISORT Atmospheric Radiative Transfer
ScaRaB - Scanner for Radiation Budget
SEVIRI - Spinning Enhanced Visible and Infra-Red Imager
SW - Short Wave

SZA - Solar Zenith Angle

T_b - brightness temperature

TEJ - Tropical Easterly Jet

TOA - Top Of Atmosphere

TOGA - Tropical Ocean Global Atmospheres

TRMM - Tropical Rainfall Measuring Mission

UTC - Universal Time, Coordinated

UTH - Upper Tropospheric Humidity

U wind - westerly wind

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1. Introduction

1.1. Introduction

In the climate system, there are a lot of components and processes which contribute to the overall complex variability and raise a lot questions regarding the current state as well as the predictability of the system. Moreover, the tropical/sub-tropical region is one of the most complex regions on our planet which hosts the large scale circulation and local meteorology, extreme surface types and most intense solar radiation, drastic temperature differences and deep convection. This thesis attempts to investigate the short term variability over Africa and the Atlantic ocean, specifically to identify the various modes of variability in the region and attribute them to various climate processes. In particular, the Outgoing Longwave Radiation (OLR) of the chosen region will be studied as the OLR varies according to a combination of Incident Solar Radiation (ISR) at the top of the atmosphere (TOA), surface and atmospheric conditions. Therefore, by decomposing the OLR into various modes of variabilities, the radiative effect of the individual processes can be investigated.

The results are presented in three separate results chapters in the thesis, where the first results chapter (chapter 4) focuses on identifying the significance of the different modes of variability within the OLR. This was done by utilising Fourier Transform to decompose the frequency of the OLR variation in terms of Fourier power. This is carried out on both the selected regions of interest and over a larger region to observe the specific frequencies associated with the regions. The spatial variation of these frequencies over the African and Atlantic region are observed to attempt to provide explanation for such dominant frequencies.

The second results chapter (chapter 5) consists of analysis on diurnal timescale to attempt to isolate different components in diurnal cycles, for example, surface response to incoming solar radiation, convective events and clouds development in

diurnal timescale. The analysis utilises the Principal Component Analysis technique to statistically separate the different components within the diurnal cycles and observe the components temporally and their spatial distribution over the region. By using both temporal and spatial parts of the components, one can determine the climate processes that contributed to the specific components.

The final results chapter (chapter 6) further investigates the diurnal cycle by analysing (PCA) subsets of the data categorised by surface types. We have then used a model to attempt to described some of this variability and attribute causes to these variability. We have also used the wind data from the ERA (ECMWF Re-Analysis) ¹ interim to explain the dynamics in the African region and compare with the variation in fluxes that we observed.

1.2. Radiation balance introduction

The radiative budget is one of the most fundamental topics in climatology, since any short term imbalance of the fluxes implies a warmer/cooler overall climate. These fluxes can generally be divided into Incoming Shortwave Radiation (ISR) and Outgoing Longwave Radiation (OLR) radiation. The first of which is defined as the solar Short Wave (SW) radiation entering the Earth whereas the latter represents the Long Wave (LW) radiation emitted by the climate system into space. The ISR is then further subcategorised into Absorbed Solar Radiation (ASR) and Reflected Solar Radiation (RSR) where ASR is the amount of ISR radiation that enters and is absorbed by the climate system, and RSR is the portion of ISR which is reflected or scattered by the climate back into space.

The satellite measurements observe the OLR and RSR, then derive the ASR from subtracting the RSR from the Total ISR. The net flux is defined as the difference between the ASR absorbed and OLR emitted by the climate system at the Top Of Atmosphere (TOA). Whilst a temporary imbalance of net flux would yield an immediate warmer or cooler climate state, the climate system responds and adjusts to balance the net flux in the long run, thus providing a stable climate. The climate system adjusts itself by varying the climate components, the variability of all the components contribute towards the global climate variability. This variabil-

¹European Centre for Medium-Range Weather Forecasts

ity is seen in wind speed, pressure, atmospheric moisture, cloud types and optical depth, amount of precipitation, ocean circulation and heat absorb, land use and heat response.

Each individual climate component has its own radiative forcing (RF) characterises by its imbalance of fluxes, which collectively, creates the overall global net fluxes. Some of the major climate components, namely surface and clouds are responsible for much of the variation in upwelling LW fluxes, whereas the downwelling SW fluxes is predominantly modulated by the variation of ISR.

The individual climate components absorb and reflect a factor of the SW radiation and LW radiation, whilst certain components heat up from SW absorption and in turn emit LW radiation. The SW variability is predominantly driven by the ISR and affected by ozone, clouds and surface scattering, whereas the LW is modulated by the climate components responding to ISR in the form of radiative emission. Therefore, by observing the OLR from a remote instrument, the variabilities of these components can be studied to further our understanding in climate.

1.3. Radiation & radiative budget

1.3.1. Radiance & flux

In terms of the radiative energy in the atmosphere, the energy transfers from one component of the climate system to another in the form of convection, conduction, advection and radiation. Whereas energy transfers from the climate system into space in only in the form of radiation. For the study of Earth's radiation budget, satellite observations are utilised either in the form of broadband LW and SW, narrowband radiances or spectrally resolved measurements. The SW and LW corresponds to the measurements of reflected solar radiation and outgoing terrestrial radiation.

The intensity or radiance of the radiation, is defined by integral of the monochromatic intensity over the wavelength range of the electromagnetic spectrum. This radiance, however, only describes intensity of the radiation emitted from the point source to all solid angles covered by the observational device (see Fig.1.1).

The flux density is therefore, defined as the radiance integrated over all the solid

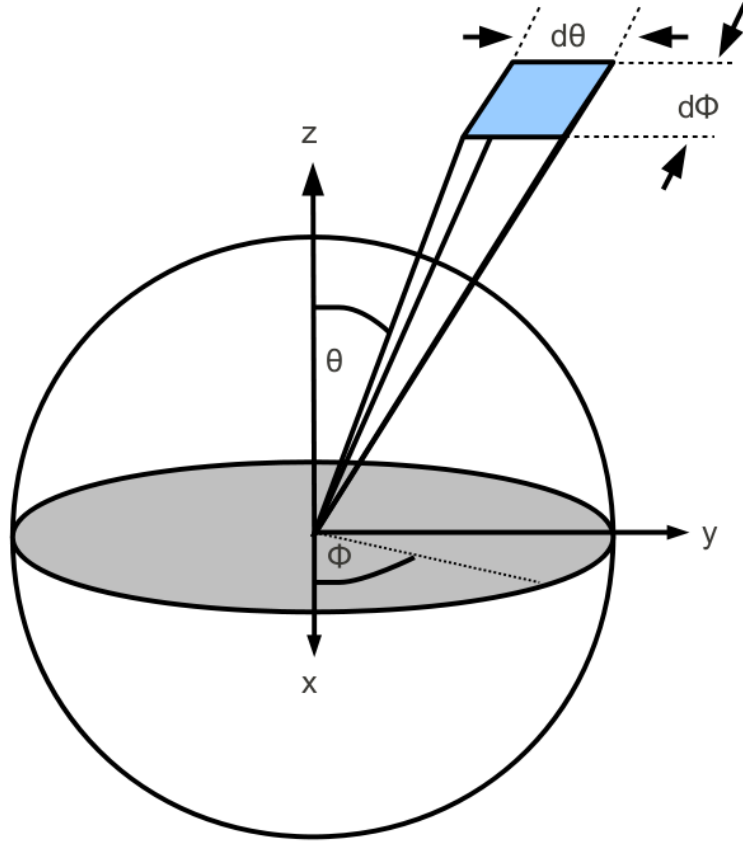


Figure 1.1.: Showing the solid angle integral conversion from radiance to flux, θ is the zenith angle and ϕ is the azimuth angle

angles that it emits into, i.e. the hemisphere:

$$F = \int_{\phi=0}^{2\pi} \int_{\theta=0}^{\pi/2} I \cos\theta \sin\theta \, d\theta \, d\phi \quad (1.1)$$

The viewing zenith angle is denoted by θ and ϕ is the azimuth angle as shown in Fig.1.1. If emitted radiation from the source is isotropic then the integral will be a constant. Following from equation 1.1:

$$F = I \left[\phi \right]_0^{2\pi} \left[-\frac{1}{4} \cos(2\theta) \right]_0^{\pi/2} \quad (1.2)$$

the conversion from radiance to flux will simply be a multiple of π since I is neither

a function of ϕ nor θ . However, in reality, the radiance function at the top of atmosphere (TOA) is not an isotropic emission. For satellite measurements, the radiance-to-flux conversion often employs the use of Angular Distribution Models (ADMs), and requires the identification of scene type to determine which conversion ratio is appropriate.

The TOA flux is the power of the radiation emitted (in the case of longwave) or scattered (SW) by the entire climate system calculated at the TOA. An ADM is a function which provides the anisotropic factors to determine the TOA flux from an observed radiance [Loeb *et al.*, 2002]. The equation is shown as follows:

$$F = \frac{\pi I(\theta, \phi)}{R(\theta, \phi)} \quad (1.3)$$

where R is the anisotropic factor and I is the radiance.

The observations made from a satellite only measure the radiance from a scene in a particular direction and the corresponding flux from the scene cannot be measured simultaneously. Therefore, the anisotropic factor ($R(\theta, \phi)$) is provided by a table of ADMs pre-defined for a number of scene types.

1.3.2. Radiation

As mentioned in section 1.2, the LW and SW together can represent the overall energy budget, whilst individually, they can provide information on the variation in the certain climate components, e.g. aerosols, clouds, surfaces. The ADM is predetermined empirically for different scene types using a large ensemble of radiance measurements that are sorted into discrete angular bins and parameters that define an ADM scene type [Loeb *et al.*, 2002].

The major climate components and their energy fluxes are represented in Fig. 1.2, where the yellow and beige denote the SW and LW radiation respectively. The SW radiation at the top of atmosphere (TOA) upon entering the climate is partially reflected and absorbed by the clouds, atmospheric gases, aerosols and the surface. A portion (α , SW albedo in equation 1.4) of the ISR is reflected back out into space, the reflected amount is given by the RSR in Fig. 1.2.

Energy escapes from the surface in the form of: a) LW radiation emission (396

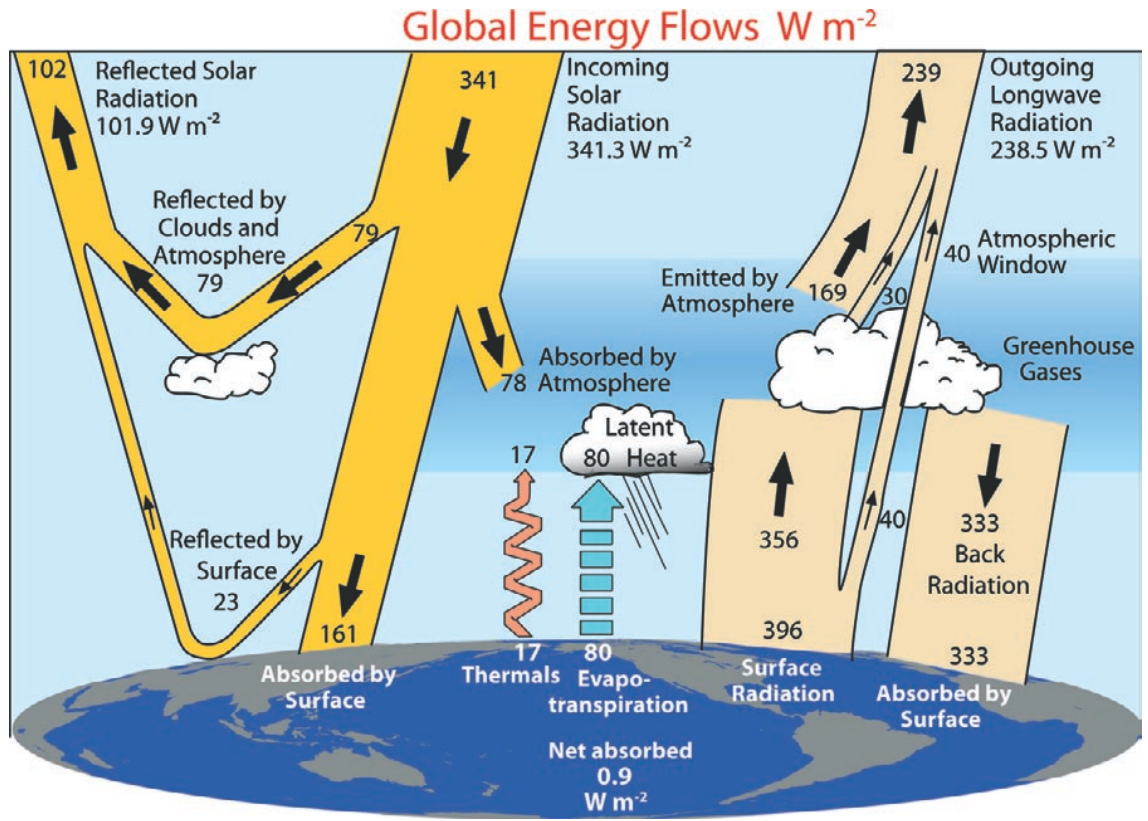


Figure 1.2.: The Earth's radiation budget ($W m^{-2}$) [Trenberth *et al.*, 2009]

$W m^{-2}$); b) latent heat fluxes via evaporation, condensation of water in the atmosphere ($80 W m^{-2}$ in fig.1.2) [Pidwirny, 2006]; c) thermal conduction between the surface and the atmosphere ($17 W m^{-2}$). A portion of the upwelling surface radiation propagates through the atmospheric window into space ($40 W m^{-2}$), whereas the rest of the energy from the surface is absorbed/reflected and re-emitted by the atmosphere and the clouds ($169 + 30 W m^{-2}$) into space.

The resultant LW radiation emitted into space is given by the OLR in Fig. 1.2 and F in equation 1.4.

The globally averaged TOA net radiative flux(H) of the climate system is represented by the net difference of the SW and LW fluxes measured at the TOA in Fig. 1.2 via the subtraction of OLR (F) and RSR(αS) from the ISR(S). The relationship is described by the equation:

$$H = (1 - \alpha)S - F \quad (1.4)$$

where α , S and F represents averaged global albedo (0.3), Incoming Solar Radiation at the top of the atmosphere ($341.3Wm^{-2}$) and Outgoing Longwave Radiation ($238.5Wm^{-2}$) respectively at the TOA [Trenberth *et al.*, 2009] represented in Fig. 1.2. The value for the net radiative flux (H) is roughly $0.9Wm^{-2}$ revised by Trenberth *et al.* (2009) on a global average.

1.3.3. Radiation balance

For a climate system to be in thermal equilibrium, the global averaged TOA net radiation over a long timescale has to be negligible. The RF defined in section 1.2 of all the individual components contribute to this global average of net flux. The most influential components would be surface, clouds, aerosols, greenhouse gases and ozone.

A variation in any of the components would lead to an overall imbalance net flux, which ultimately alters the climate as it adjusts to restore the balance. For example, a general circulation model (GCM) has been utilised to investigate the effect of greenhouse gases on global climate, Boer *et al.* (1992) have studied the effect of double CO_2 on current climate, the results were found to indicate a global warming of $3.5^\circ C$ per century.

A relevant study which investigates the long-term cloud effect is [Wang *et al.*, 2002], they have looked at the long-term changes in tropical cloud using satellite observations of the OLR. They have categorised different types of clouds (opaque, subvisual) according to their optical depth at a particular wavelength, $1.02\mu m$, and looked at the changes in the cloud occurrence, the associated cloud forcing and OLR over the period of 1985-1999. Wang *et al.* (2002) have found an increase in OLR of $3.9Wm^{-2}$ per decade during 1985-1999 which cannot be attributed to increases in greenhouse gas concentrations because an increase in the concentration would trap more OLR, rather than providing an enhanced effect. An increase in the OLR caused by a decrease in cloud optical depth should be mirrored by a decrease in the reflected shortwave radiation (RSR) in terms of radiative, and [Wong *et al.*, 2000] has reported the RSR from ERBE to have decreased at $-2.4Wm^{-2}$ per decade between 1985 and 1999. This decrease in RSR indicates a increase in ASR which almost cancels the OLR increase in terms of net radiation balance.

Amongst the different climate components, clouds and surface types are some of the most significant components which govern the energy balance. Therefore, by analysing the OLR and studying its variation, we can improve our knowledge of the different components and understand the processes associated with these variables and their variabilities over time.

1.3.4. Remote sensing measurements of the Radiation Budget

There have been various sensors designed to measure the radiation budget from remote sensing measurements. Amongst the Earth Radiation Budget(ERB) instruments, ERBE (1984-2005) [*Barkstrom and Hall, 1982*] was a scanner instrument on the ERBS satellite orbiting in a precessing orbit with a 57 degree inclination. The orbital period was 96 minutes for the ERBS, and the plane of rotation's angle was changed according to the earth's rotation around the sun (i.e. approximately 1 degree per day) plus the angle of the precessing orbit of the satellite. The period of the precessing orbit was 72 days, thus the instrument would produce measurements of all locations through all local times every 72 days. Due to the period of the orbit, it's more appropriate to analyse the data globally on a seasonal basis since a monthly dataset would involve estimation as the sampling rate of a complete set of measurements require 72 days. Both CERES [*Wielicki et al., 1996*] and ScaRaB (*Kandel et al., 1994; Desbois et al.,*) instruments are in a low earth orbit and with inclination angles of 35 and 20 degree respectively over a sun synchronous orbit, which only provide full sets of diurnal cycles over all locations every 90 days.

The limited temporal sampling of these ERB instruments is not ideal for studying the diurnal cycle or short time scale variations. However, with the GERB [*Harries et al., 2005*] instrument in a geostationary orbit over 3.5°W and 0°N, the time resolution of the observations is only limited by the instrument itself, whose sampling is roughly 17 minutes, providing roughly 85 measurements per day for all locations that it observes. This makes the instruments particularly well suited for studying short timescale radiation variability.

1.3.5. Effect of cloud on the Radiation Balance

Clouds and their development are some of the most important components of the atmosphere, where they play an important role in governing the net radiation balance as well as creating some of the more dominant modes of variabilities in both the longwave and shortwave radiation fields.

Clouds are formed by an ascent of damp air which leads to condensation at higher altitudes due to temperature differences. They can generally be divided into four categories: layer cloud formed by regular ascent such as warm air rising over frontal zones, layer cloud formed by vertical mixing, convective clouds and clouds formed as air is lifted vertically as it passes over mountains. The first two processes cause the formation of stratus clouds or cirrus clouds at higher levels, whereas convective ascent lead to cumulus clouds. On the other hand, the mountain induced vertical displacement leads to wave or lenticular clouds (*Houghton, 1986; Worthington, 2001*).

Cloud radiative forcing (CRF) is the change to the net radiation measure at the TOA due to cloud which is given by the difference in the averaged TOA net flux under clear sky (H_c) and all sky (H) conditions, to yield the equation:

$$CRF = H - H_C = (F_C - F) - S(\alpha - \alpha_C) \quad (1.5)$$

where the subscript C denotes the clear sky condition [*Cess et al., 1992*]. On average, the CRF is negative (*Ramanathan, 1988; IPCC, 2007*), therefore, indicating that the clouds induce a cooling effect to the climate in a global average.

2. Background

2.1. African/tropical meteorology

The tropical region is a very interesting and dynamically complex region due to the latitudinal temperature gradient [Holton, 1992], tropical monsoon/rainforest/savanna climate, the intertropical convergence zone(ITCZ), deep convective events and tropical cyclones. Furthermore, the continent of Africa itself provides a lot of surface variabilities which affects the tropical circulation and hence, the synoptic scale meteorology.

2.1.1. Tropical meteorology

There are several characteristic features in the tropical region which are unique and contribute to the complexity of the tropical circulations. For example, the intertropical convergence zone(ITCZ), African wave disturbances (discussed in section 2.4) and tropical monsoons.

2.1.2. Intertropical Convergence Zone

The surface on or near the equator (dependent on solar zenith angle) receives the maximum intensity of solar radiation and therefore heats up the air above it and the air rises to higher altitude, leaving a low pressure zone near the surface. The relative high pressure in the tropics north and south of the minimum SZA drives the air southward and northward respectively to equalise the pressure. The converging air experiences Coriolis force as it travels in the meridional direction and deflects towards the west at the surface pressure level in both northern and southern hemispheres. Hence, the air from the north and south in turn forms northeasterly and southeasterly trade winds which converge near the equator, thus moist air is forced

upward. This process circulates in the tropics and forms Hadley cells (see Fig.2.1) on both sides of the equator.

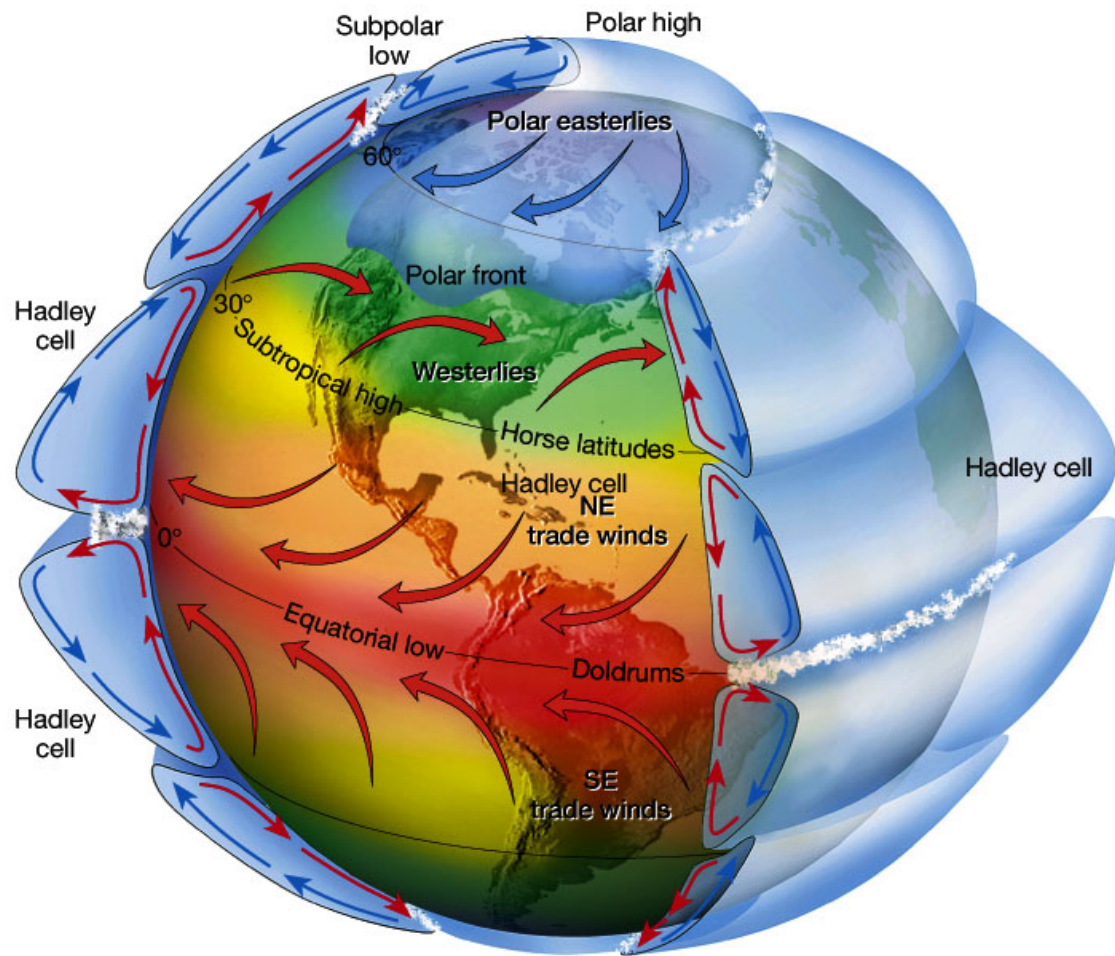


Figure 2.1.: Diagram of the Hadley cells and the corresponding wind direction, Image from NASA Earth Observatory

The area where the trade winds converge is called the Intertropical Convergence Zone (ITCZ). This is where the moist air rises and the water vapour within the air parcel condenses as the air cools during the uplift, usually resulting in cloud formation and subsequent precipitation. Over land, the position of the ITCZ changes according to solar zenith angle of the hemispheres due to variations of the latitude where surface receives maximum solar radiation. However, the seasonal positions of the ITCZ are not perfectly symmetric due to surface type and the orography

variations. This seasonal perturbation in the position of the ITCZ determines wet and dry seasons for many tropical regions, caused by the air moisture, clouds and precipitation associated with the convergence.

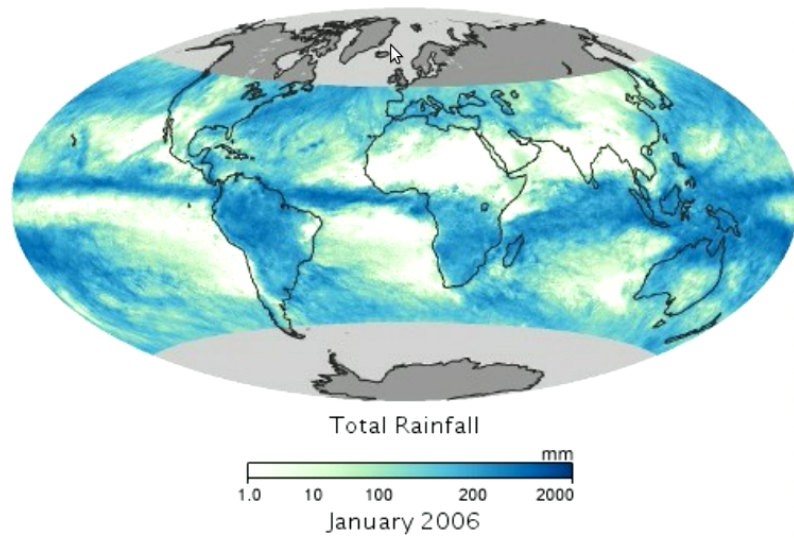
Asymmetry and seasonal variation of the ITCZ

Although the ITCZ is heavily influenced by the seasonal variation of the SZA over land, the ocean counterpart remains in the northern hemisphere all year round.

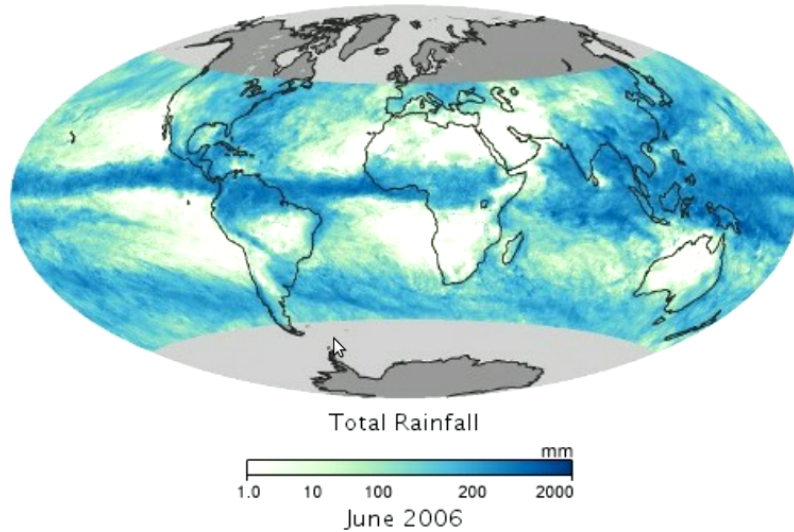
Fig. 2.2(a) and 2.2(b) show the rainfall maps for January and June 2006 respectively. The data used in the plots are from the Nasa's TRMM mission. The ITCZ is one of the main sources of precipitation in the tropics due to the convective activities and cloud associated with the convergence. Therefore, precipitation is a good proxy for the location of the ITCZ over the season. In Fig. 2.2, it can be observed that the rainfall band over land generally shifts towards the north of the equator in June and south in January. This is due to the shift in the maximum ISR associated with the minimum SZA during the months. However, the rainfall band over the ocean presents a different picture, the ITCZ remains north of the equator during both boreal winter and summer.

This has been investigated by *Philander et al.* (1996), where the asymmetries in the seasonal ITCZ are particularly prominent in the eastern Pacific and Atlantic Oceans. It was suggested in the paper that this is the effect of: 1) the interactions between the ocean and atmosphere which turns symmetry into asymmetry, 2) the geometries of the continents that determines the effectiveness of the atmosphere and ocean interaction.

The first mechanism involves physical instabilities and feedback mechanisms which magnifies any small asymmetric perturbations into a sustainable asymmetry that we observe in the atmosphere. One such mechanism proposed by [*Pike, 1970*], where the cold upwelling of the ocean at the equator prevents the tropical convergence from centering directly above the cold water, as it suppresses the uplift of the air. [*Charney, 1971*] have further propose the asymmetry of the ITCZ is dependent at a feedback mechanism between the atmosphere and ocean caused by the moist deep convection and the convergence of air and moisture feeding the convection. The convection itself creates a low pressure on the surface and thus maintains the wind convergence, this wind convergence in the northern hemisphere at the sea surface



(a) Total rainfall map over January 2006



(b) Total rainfall map over June 2006

Figure 2.2.: Total rainfall maps over months January and June 2006 [NASA, Earth Observatory website] (TRMM rainfall data) grey indicate no data present.

also prompts the downwelling of ocean water beneath the wind convergence and consequently, upwelling cold water in the south. This creates a warmer pool of water directly underneath the ITCZ relative to the colder sea surface in the other hemisphere and further maintains this asymmetry. This process, however, does not

impose any preferences on the ITCZ over the northern or southern hemisphere, it merely provides a feedback mechanism to maintain the asymmetry itself. To explain the origin of the asymmetry we consider the arguments related to the geometries of the continents.

The second mechanism considers the asymmetries of the continents in the southern and northern hemisphere and how the asymmetry of the seasonal movement of the ITCZ arises from those. In the African continent and Atlantic Ocean, the ITCZ has demonstrated a good example of the asymmetry, where the land ITCZ shifts according to the season, the ocean ITCZ, however, remains north of the equator most of the year.

In this case, the horizontal temperature gradient between the Gulf of Guinea and the land in the north is large. Therefore, creating a surface pressure difference between the two regions, and consequently, the surface air moves northward to balance the pressure. The meridional winds are cross equatorial [*Philander et al.*, 1996], thus the wind in itself is already creating an asymmetry about the equator. This is shown in fig. 2.3 c where the meridional wind remains northward over the Atlantic ocean, north of the equator, up to around 5°N. The cross equatorial meridional wind itself causes the asymmetry in two ways: first of which is forming a convergence north of the equator, thus promoting convective activities; Secondly, the wind causes Ekman transport which induces upwelling of cold water near the coast of southern Africa and the warm surface water is transported to the north of the equator, whereas, in the south, the coastal region is replaced by cold water. This will intensify the winds and amplify the asymmetries about the equator.

This asymmetry forms a convergence north of the equator and is observable via the annual averaged rainfall in fig. 2.3a where most of the precipitation over the Atlantic ocean is concentrated in the northern hemisphere.

In fig. 2.3a and 2.2 , a northern ITCZ bias appears over the tropical Pacific just as in the Atlantic. However, the asymmetries of the Pacific appear to be more complicated as there aren't any major equatorial geometry asymmetry equivalent to that of the Atlantic. [*Philander et al.*, 1996] have used an atmospheric GCM to investigate this and they have found the local coastal geometry, specifically the inclination of the coast to meridians on the west coast of Americas can contribute the asymmetry. The inclination causes the trade wind to be parallel to the coastline

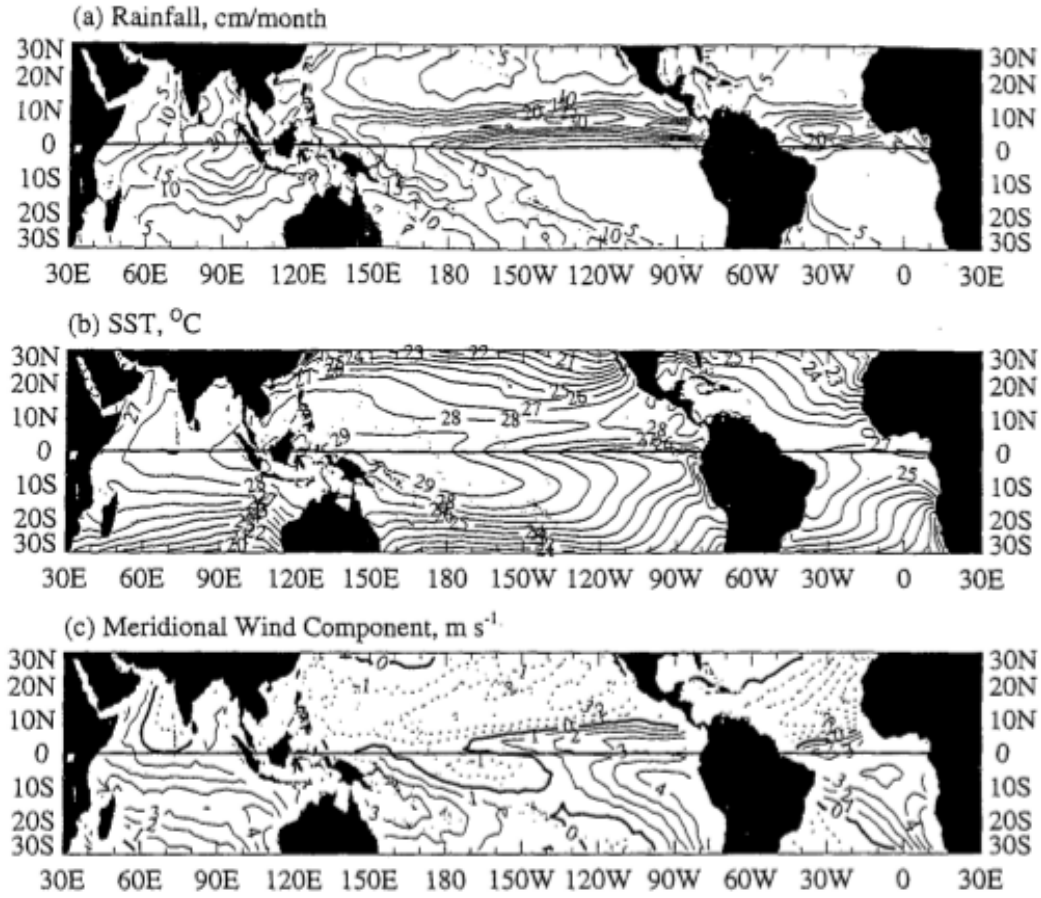


Figure 2.3.: Time-averaged distributions of (a) rainfall estimated by Spencer (1993) using measurements recorded by the Microwave Sounding Unit on NOAA polar orbiting satellites; the contour interval is 5cm/month (b) Sea surface temperatures from Reynolds and Smith (1994); The contour interval is 1°C (c) Meridional wind components at a height of 10 m from (Halpern et al. 1993) [figure taken from [Philander et al., 1996]]

in the southern hemisphere and perpendicular in the north. Therefore, the southern coast is expected to colder due to the Ekman transport which induces the upwelling of cold water in the south whilst driving the warm surface water towards the north. This asymmetry in temperature across the equator causes meridional wind and the asymmetry is intensified and maintained via the same mechanism mentioned above.

2.1.3. African meteorology

In the region of the African continent, the surface conditions can create local effects which give rise to one of the most intense circulation features over Africa [Nicholson, 2009]. During the boreal summer, due to the surface heating of the Sahara desert, there exists a strong latitudinal temperature difference between the desert and the ocean and forests near central Africa. Thus, induces a strong thermal wind which results in an easterly jet at 650mb centred near 16°N [Holton, 1992] shown in Fig. 2.4.

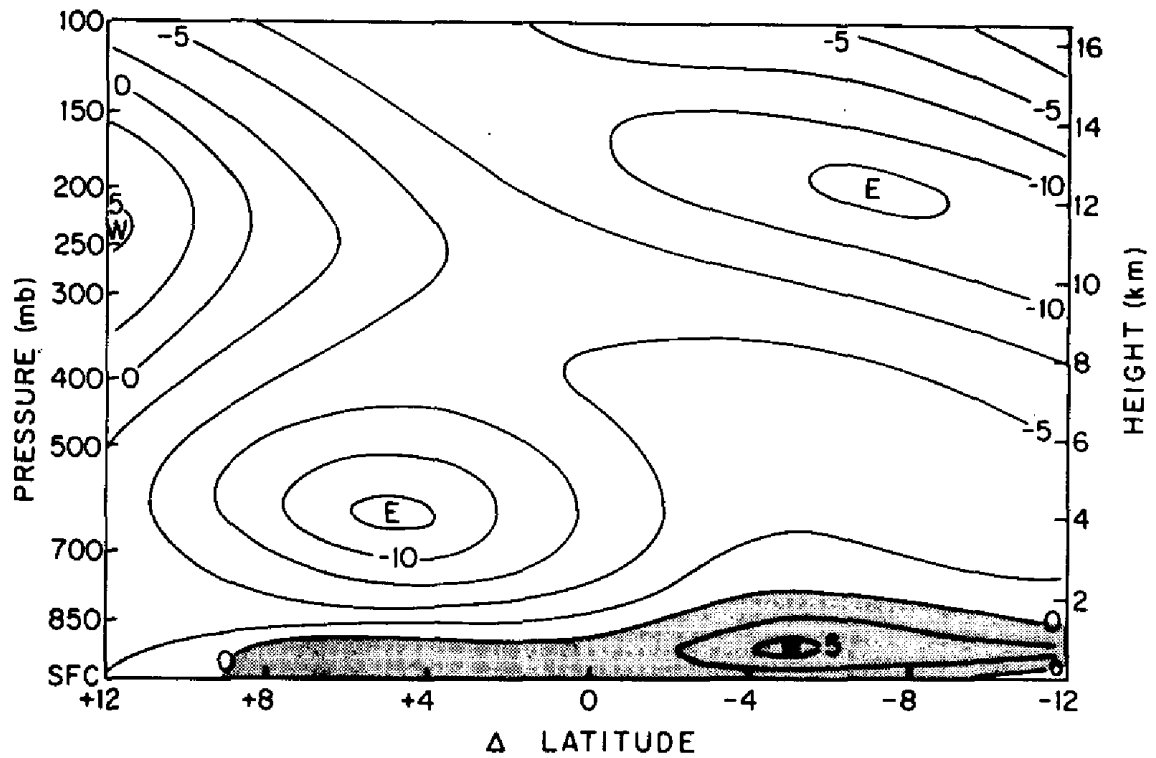


Figure 2.4.: Zonal mean wind field from 23rd August to 19 September 1974, image taken from [Reed *et al.*, 1977], Δ latitude centred at 12°N. Zonal wind field averaged over 10°E to 31°W

Synoptic-scale disturbances are observed to form and propagate westward, south of the easterly jet core shown in Fig. 2.4. These disturbances occasionally cause

storms and hurricanes in the western Atlantic. The average wavelength of the African waves have been observed to be 2500km in wavelength and possess a westward propagation speed of $8ms^{-1}$ and hence circumnavigating the earth in the timescale of 3.5 days [Holton, 1992].

The origin of such wave disturbances is induced by the perturbations along the jet stream. These perturbations can be caused by climate variability such as convection beneath the jet stream which forces the air parcels within the jet stream to ascend or topological features which could also force an uplift. Thus, it experiences a countering force by the gravitational field and Coriolis force, the air ascends and descends as it propagates along the stream, forming clouds and precipitation with the appropriate atmospheric conditions as it's ascending. The African waves are formed in the same manner, caused by the jet shown in Fig. 2.4. These wave disturbances originate over north Africa and propagate westward to the Atlantic ocean.

It has been suggested in [Mekonnen *et al.*, 2006] that the convection triggered by mountains over central and eastern Africa, near Darfur and Ethiopia are responsible for the initial development of the African waves. Therefore, one would expect to observe the wave disturbances west of the mountains. The waves are thought to occur during May-October and to have period ranging from 2-9 days, (Mekonnen *et al.*, 2006; Gu *et al.*, 2004; Grist, 2002; Diedhiou *et al.*, 1999).

The African waves are believed to distribute the rain fall in the northern part of Africa [Nicholson, 2009], where the moisture available for convection is coupled with strong uplift and this in turn is governed by African and Tropical jets rather than moisture convergence.

Studies (Mekonnen *et al.*, 2006; Gu *et al.*, 2004; Grist, 2002; Diedhiou *et al.*, 1999) have suggested a large temporal (2-9 days) and spatial (2000-5000km) range of results in the African wave disturbances. This incoherence in observations could be due to the difference in instruments and models or analysis methods, or to the innate variabilities and the complexity of the wave disturbances. By using a high temporal resolution geostationary instrument, one can expect to gain a better understanding of these phenomena and narrow the possible range of results.

2.1.4. Radiation effect

These wave disturbances are known to cause convection and are associated with precipitation variabilities. *Taleb and Druyan* (2003) have demonstrated that the rainfall associated with the African waves accounts for 30-40% of the total rainfall. This is important in local radiation balance as the convection caused by the waves usually lead to condensation and hence precipitation. The clouds absorb and reflect both LW and SW radiation to a certain degree, dependent on the cloud type (see section 1.3.5). Therefore, by studying the clouds or OLR variation, the variabilities and the effects of wave disturbances can be further understood.

The majority of the previous studies ([*Mekonnen et al.*, 2006] [*Gu et al.*, 2004] [*Grist*, 2002] [*Diedhiou et al.*, 1999], [*Nicholson*, 2009]) have utilised rainfall or cloud index in the form of OLR threshold and local observational data, wind field or pressure data, or the use of OLR as a proxy to characterise the African waves. However, there has been a lack of studies on the direct effect of OLR from the wave disturbances, which is crucial in characterising the wave and understanding the radiative variabilities of it.

The radiation variability of the wave is expected to show similarities to wind studies in terms of spatial variation, as they should essentially be observing the same disturbances. However, the temporal distribution should show some degree of phase lag behind the wind/stream studies as a change in the wind field strength/direction usually does not immediately induce convection or precipitation.

2.2. Evidence to date about the diurnal cycle of OLR

The diurnal and annual cycles of solar radiation are the most fundamental sources of variability in the climate system. The variation of downwelling SW radiation is responsible for the magnitude of the surface heating. The outgoing longwave radiation (OLR) in turn varies on a daily/annual basis in response to the variation in solar forcing, to cool the climate system. The diurnal variability of OLR is particularly important in investigating short timescale atmospheric processes which are modulated on the same time scale. There are several factors which affect the

diurnal variability of the OLR. One of the most significant components of the diurnal variability in the tropics is the diurnal cycle of convective clouds. There have been a number of studies examining the diurnal cycle of OLR observed from satellites [e.g. *Comer et al.*, 2007; *Smith and Rutan*, 2003; *Allan et al.*, 2007].

The diurnal variation of the OLR is modulated by the response of the climate system to the variation of ISR on diurnal timescale and is affected by a combination of factors: surface temperature, amount of cloud and humidity of different heights in the atmosphere, etc. *Smith and Rutan* (2003) and *Comer et al.* (2007) have decomposed the diurnal variation of OLR observed from satellite according to the covariance within the dataset using principal component analysis (PCA).

2.2.1. Diurnal timescale analysis

Smith and Rutan (2003) have used 5 years of data from the Earth Radiation Budget Experiment (ERBE) instrument, which is situated on a sun-synchronous satellite which precessed through all local times every 72 days. The satellite had an orbit with a 57° inclination angle and the ERBE instrument [*Smith and Rutan*, 2003] on board measured the broadband RSR and OLR. The OLR fluxes employed in their analysis cover the latitude range of 55°N to 55°S which accounts for 82% of the earth's surface. In their study, the data were divided according to seasons, with 15 months for each season (5 years of data in total), the analysis was carried out on seasonally averaged diurnal cycles. The temporal resolution of the mean diurnal cycles was hourly and spatial resolution was 2.5° by 2.5° .

Each of the 24 hourly means were taken away from each 24-component vector corresponding to each region k [*Smith and Rutan*, 2003]. Then, a covariance matrix was formed using the equation below:

$$\Gamma_{ij} = \frac{\sum_k w(k) h_i(k) h_j(k)}{\sum_k w(k)} \quad (2.1)$$

where $h_i(k)$ corresponds to the i th hour for the k th region, $w(k)$ is the area weighting. The eigenvalues and eigenvectors of the covariance matrix are then calculated. The eigenvalues are variances associated with eigenvectors to describe the variability of the diurnal cycles. The eigenvectors are called principal components (PC) with

the most significant one corresponding to the highest eigenvalue (or variance). The empirical orthogonal functions (EOF) are the spatial functions associated with each of the PCs, calculated by multiplying a single PC by the variance matrix $h(k)$ and then dividing by the corresponding eigenvalue.

Comer et al. (2007) employed data from the GERB instrument, with a much higher temporal resolution of 17 minutes. The instrument takes broadband radiation measurements and is on a Geostationary satellite positioned at 0° latitude, 3.5° longitude, which is especially suitable for analysis over the African region. The data used in their study is Edition 1 release of the GERB OLR data in ARG flux product. They have taken a monthly average of the diurnal cycle of OLR fluxes for July 2006, performing a similar analysis as *Smith and Rutan* (2003). However, they have considered the OLR flux data from land and ocean as a whole and suggested that the separation does not significantly affect the results.

The diurnal cycle of OLR have also been examined by *Allan et al.* (2007) in the context of evaluating a numerical weather prediction (NWP) model. They used observational LW and SW fluxes data from GERB and model data from the Met Office NWP global model. They have specifically looked at a comparison of model with GERB data by taking differences for two different scene types: total (cloud + clear) and clear sky. The differences have been taken in the OLR and RSR over the period April-September 2006 between GERB and model data, using 12 universal time, coordinated (UTC) data as well as differences for the net downwelling radiation at TOA. They have also investigated the diurnal cycle of OLR, total and convective cloud fraction averaged over the African region (7° - 45° E, 10° S- 10° N). The cloud fractions for observed data are based on SEVIRI channels [*Allan et al.*, 2007].

2.2.2. Diurnal cycle of OLR over land

Smith and Rutan (2003) have performed the analysis on the ERBE OLR fluxes data from land and ocean separately, they found that the first and second PCs contributes 72%-78% and 8%-10% to the total variance respectively. Their first PCs for all seasons over land are extremely similar as they describe the response of the surface to solar radiation heating. It was found that the curves peak at about noon and are symmetric about the peak with the minima positioned at night time.

Their corresponding first EOF dominated the diurnal cycle over land, implying that the daily (non-decomposed) signals for each region would look largely similar to the first PC. They found the strongest signal to be mostly over deserts where there is little moisture to permit solar heating to be converted into latent heat [*Smith and Rutan, 2003*].

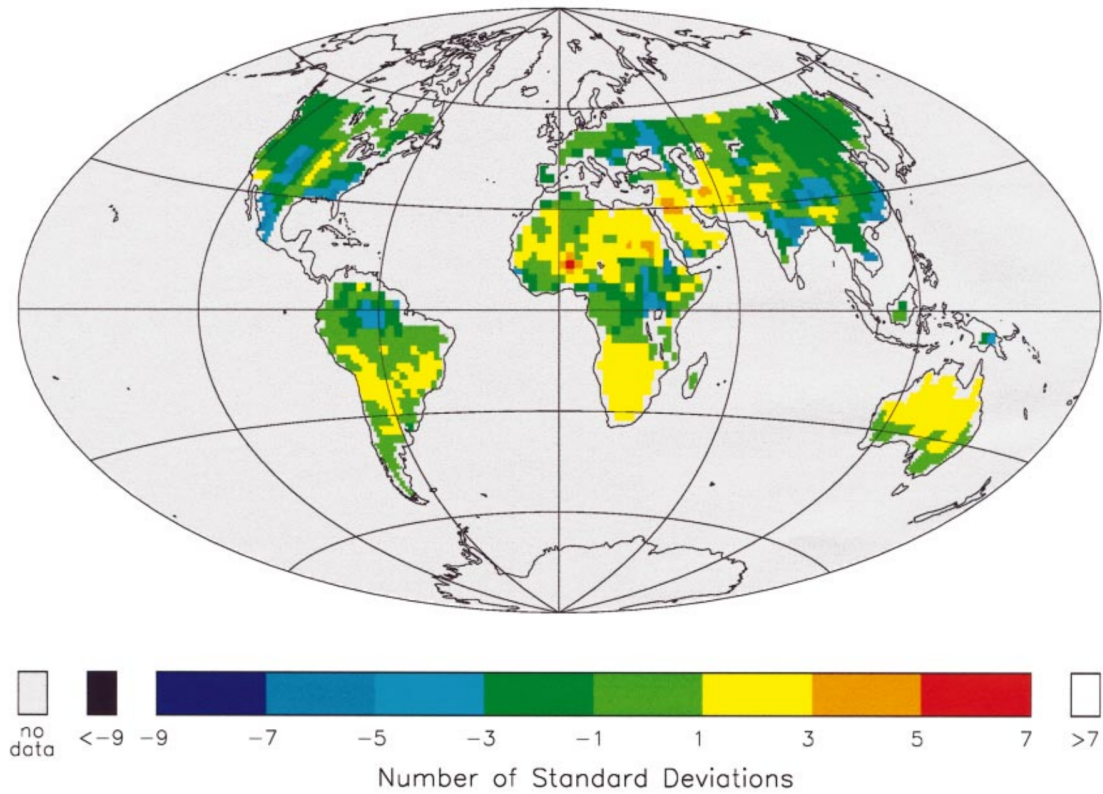


Figure 2.5.: Map of second EOF of OLR for land in boreal summer time, taken from [*Smith and Rutan, 2003*]

They have observed the second PCs for all seasons again to be quite similar to each other, taking the shape of a sinusoidal curve, peaking at approximately 1600 local solar time (LST) with a minimum at around 0800 LST which makes it about 4 hours out of phase with PC1. Their map of EOF2 (see fig.2.5) for boreal summer over land shows maxima over northern Africa, the Kalahari Desert and the Middle Eastern deserts. This signal is due to a combination of cloud development, surface and orography. Deep convective systems over regions of Brazil and Africa have

shown negative values of EOF2. This shifts the maxima of the diurnal cycle to the morning due to the effect of high cloud development during the afternoon. The PC3 for land showed two maxima during the day and the EOF3 showed that the regions with the maxima overlap with regions of deep convective activities.

In contrast to [Smith and Rutan, 2003], Comer *et al.* (2007) have found their first PC contributes 82.3% of the total variance. They found that the first PC carries an asymmetric pattern, which shifts the peak from noon [Smith and Rutan, 2003] to around 1300 LST, thought to be primarily due to the finite heat capacity of the surface but also influenced by the atmosphere and surface feedbacks to the heating. This suggests that the analysis results may be highly sensitive to the dataset temporal resolution or region observed. They have suggested that the main OLR effect in the EOF2 is mainly due to convective cloud over land triggered by surface heating and that the cloud optical depth maximises at about 1700-1800 LST. Thus, the corresponding OLR PC2 signal minimises at this time with a maximum at about 0800 LST. They also analysed the EOF2 signal with respect to the topographical features. They have found in the map of EOF2 that the strongest signals for convective clouds are over mountains (see Fig. 2.6).

The topographical features 1, 2, 4, 6 and 7 in Fig. 2.6 expressed strong signals in the EOF2 component, indicating a relationship between topography forced uplift and convective clouds. However, features 4 and 5 are more subtle in terms of size of spatial distribution, which suggests topographical induced uplift also requires the appropriate atmospheric conditions for the convective clouds to develop.

There is also a big region of strong signal just below feature 6 which overlaps with the Kilimanjaro mountain and Victoria lake. This will be investigated in chapter 5 where we discuss our diurnal analysis results.

In their results, the combined variance from the PC1 and PC2 adds up to over 95% of the variance. The PCA technique is powerful in decomposing the diurnal cycle as the physical signals of surface heating and cloud response in EOFs 1 and 2 are clearly identified. However, it would be important to observe the location at which the residual signal concentrates.

Allan *et al.* (2007) have used averaged GERB data for July 2006 over equatorial Africa (7°45'E, 10°S10°N), they have found the cycle of OLR to peak sometime near noon, a lower temporal resolution sample (3 hours) of the data (compared with

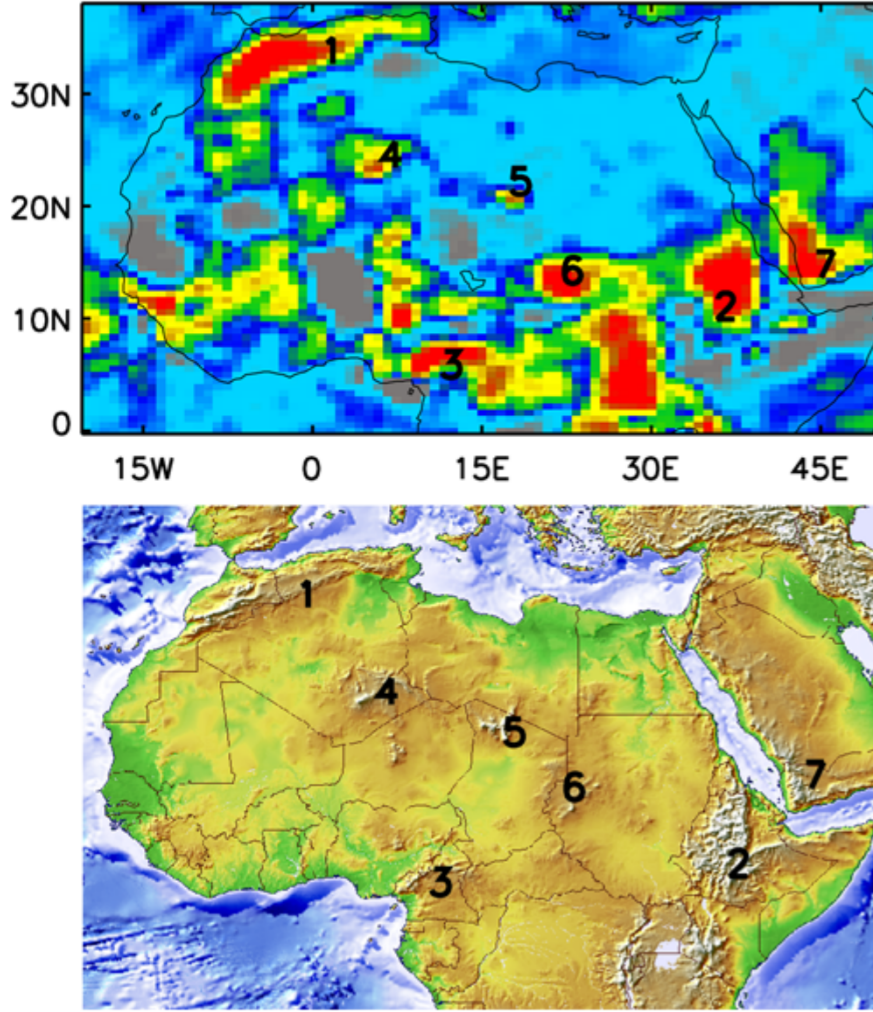


Figure 2.6.: Map of second EOF with topographical features in North Africa, taken from [Comer *et al.*, 2007]

GERB data resolution at 17 minutes) was employed for the purpose of comparison with climate models, hence the exact timing of the peak is uncertain. The diurnal cycle of OLR produced by the model peaks 3 hours earlier than observational data, where the resolution of the model output could have partially contributed to the large discrepancy.

Allan *et al.* (2007) have observed differences in their map of model and GERB data; a model overestimate of the OLR in the desert regions especially over the

Sahara and a correspondent underestimate in RSR over the sub-Saharan regions in the clear-sky comparison. The clear-sky RSR differences are likely to be due to an unrealistic spatial distribution of modelled surface albedo, whilst the OLR differences are likely to arise from the high mineral-dust optical depths which are not represented by the models. Most of these significant clear-sky differences in both RSR and OLR are also highlighted on the total OLR and RSR plots. The total plots also indicate modelling errors in higher-altitude cloud properties and the cloud fraction over the Ethiopian highlands. The net radiation comparison of GERB and the model exhibits a substantial overestimate in the model over the marine stratocumulus regions and the ITCZ.

Although, the studies reviewed in this section have analysed the diurnal cycle of OLR, some of them specifically analyse the OLR cycle over land. However, they have all treated land as a whole domain, which may not necessarily be the best selection of data to analyse, as different surface types will show various surface heating cycles, and could possibly lead to a different diurnal variation of the convective cloud signal. Thus, a separation of the data according to surface types may be appropriate for analysing both OLR and cloud cycles.

2.2.3. Diurnal cycle of OLR over ocean

Smith and Rutan (2003) have analysed the OLR data over the ocean only surface type, similar to the land only analysis in Fig. 2.5. The decomposed PCs with the highest eigenvalues were found to be much less significant in variance, with the first PC contributing from 16% to 19% of the total variance and 10% to 12% for the second PC, which indicates a greater variety of diurnal cycle forms over the ocean. Thus, the effect on OLR of the different processes are relatively stronger. Furthermore, their PC1 for ocean was found to be sinusoidal with a maximum near noon and another weaker maximum near midnight, which is quite dissimilar to the diurnal pattern to the PC1 for land. This is because of a much smaller ocean surface temperature response associated with the high heat capacity of water. Although over both land and ocean, the sunlight is absorbed partly by the clouds and then LW radiation re-emitted during the night, the effect of this is much less significant over land than ocean. Thus, the minima in PC1 correspond to the maxima in cloud

optical depth.

Smith and Rutan (2003) found the EOF1 over the ocean to be mostly between -1 and 1, except for regions where there is monsoon activity which would impose a big perturbation in the diurnal variation caused by deep convection.

In *Allan et al.* (2007), the diurnal cycle of OLR was analysed by taking a monthly average of the daily cycle over the southeastern Atlantic (20°W20°E, 0°40°S). The observed OLR cycle had maximum and minimum at 1500 and 0600 LST respectively. Whereas, the model gives max/min times of 1200 and 0600 and a second minimum at 2100 which appears to be out of phase with observational data and is of a slightly different diurnal shape.

The results discussed in this section suggests the ocean OLR cycle has a small variability, due to weaker surface heating signal associated with high heat capacity, hence, the daily cloud cycle becomes a more dominant feature in influencing the diurnal OLR compared to land. However, the separation of ocean into different types is not as obvious as land type. One possible approach may be to divide regions of ocean by temperature and analyse a group of regions with similar temperatures.

2.3. Diurnal cycle of clouds

The diurnal cycle of clouds changes the cycle of OLR considerably. There are numerous studies which examine the diurnal cycle of upper tropospheric humidity (UTH), high clouds, deep convection, precipitation. by employing narrowband measurements [e.g. *Chung et al.*, 2007; *Tian et al.*, 2004; *Slingo et al.*, 2004; *Yang and Slingo*, 2001]. The majority of them define indices for different types of clouds via the application of a threshold on the readings from a particular channel or the differences of two.

Yang and Slingo (2001) have used a global archive of high-resolution window (11-12 μ m) brightness temperature data from multiple satellites developed by the European Union Cloud Archive User Service (CLAUS) project. They have analysed the diurnal cycle of brightness temperature as well as the estimated precipitation by using Fourier analysis. The amplitude and phase of the diurnal and semi-diurnal harmonics were analysed, and a threshold technique was used to estimate the precipitation rate from the window brightness temperature (see eq.2.2).

$$DCA = a(230 - T_b) \quad \text{for } T_b < 230K \quad (2.2)$$

where DCA is defined as the deep convective activity index which relates brightness temperature of the window channel to precipitation, T_b is the brightness temperature and a is the conversion constant ($0.29 \text{ mm h}^{-1} \text{ K}^{-1}$).

Fig. 2.7 demonstrates that over the oceanic regions west of the continents, the brightness temperature signal tend to peak in the mid or late afternoon due to the minimum in marine stratocumulus cloud optical depth in the late afternoon ([*Hignett*, 1991]; [*Bergman and Salby*, 1996]). However, over the rest of the ocean, the maxima tend to be vary by a large amount, due to the complex diurnal cycles caused by the deep organised convection as well as submesoscale convection which maximise in the early morning and late afternoon respectively.

Yang and Slingo (2001) have shown (see fig.2.7) that over desert regions the maximum in brightness temperature occurs just after local noon, except for regions along the ITCZ. This slight lag in time to the solar maximum heating is due to the heat capacity of the surface which was confirmed by *Comer et al.* (2007). They also observed ocean region brightness temperature outside of the convergence zone (i.e. in clear sky or marine stratocumulus conditions) to maximise in the mid to late afternoon, however, this observation may not be absolute due to the weak diurnal cycle over the ocean and the limited accuracy of the T_b data.

The phase of estimated precipitation is related to the maximum in the diurnal cycle of deep convection and the estimated precipitation tends to peak around evening to midnight over land. *Yang and Slingo* (2001) results also showed a phase variation over some locations suggesting a topographical dependence on the diurnal cycle of deep convection.

Over the oceanic regions, the timing of the maximum is more unpredictable, however, over regions where there is deep convection (i.e. along convergence zone) the brightness temperature seems to peak early in the morning.

Tian et al. (2004) have derived different indices by using more than one narrow-band channel: water vapour ($6.7\mu\text{m}$) and window ($11\mu\text{m}$) radiances from multiple geostationary satellites. In their study, as well as employing the estimated precipitation rate as defined in eq.2.2 by *Yang and Slingo* (2001), they have also applied

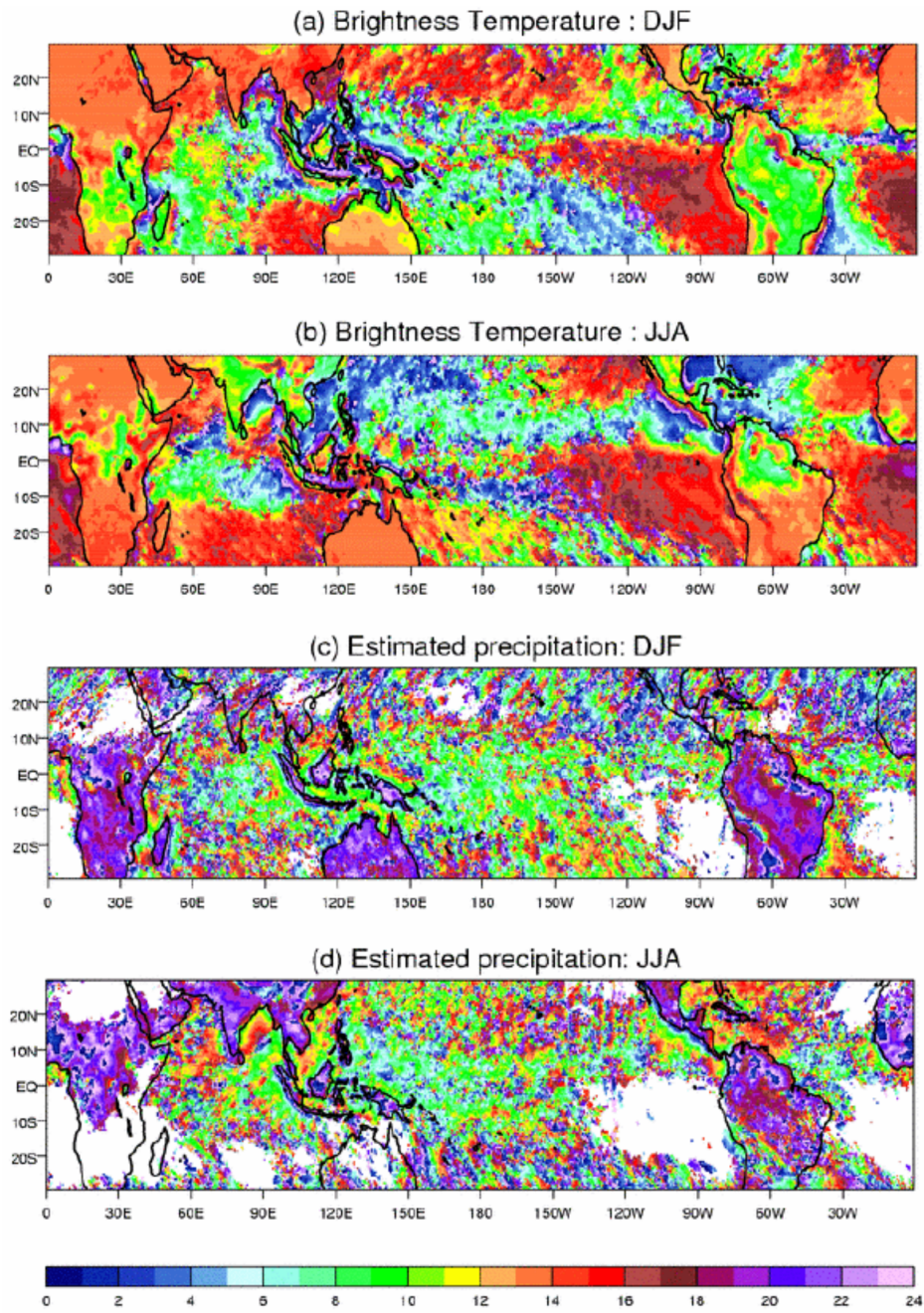


Figure 2.7.: Seasonal mean phase (*hours*) of the diurnal harmonic of brightness temperature, estimated precipitation for December-February (DJF) and June to August (JJA) , taken from [Yang and Slingo, 2001]

another parameter:

$$UTH = \frac{\cos\theta}{p_0} \exp(a + bT_{6.7}) \quad (2.3)$$

where a and b are constants, $T_{6.7}$ is the brightness temperature in the water vapour channel, θ corresponds to the observational zenith angle and p_0 accounts for the dependence of $T_{6.7}$ on atmospheric temperature. They have observed a distinct difference in amplitude and phase of precipitation, high cloud amount, upper tropospheric humidity and brightness temperature between land and ocean. Their diurnal phase plot of brightness temperature T_{11} is found to be consistent with *Yang and Slingo* (2001). In their study, a phase lag was found between UTH, high clouds and deep convection, suggesting that UTH is regulated by the diurnal cycle of deep convection and high clouds. This is consistent with the hypothesis that deep convection processes moisten the upper troposphere.

Chung et al. (2007) have carried out a similar study to *Tian et al.* (2004) using data from the SEVIRI 6.2, 7.3 and $10.8\mu\text{m}$ channels for identification of UTH and cloud classification as well as a MTH parameter using a simple linear relationship similar to eq.2.2. They have especially analysed the variation of certain humidity features in the Lagrangian perspective to account for the movement of the clouds or moisture due to atmospheric circulation.

In this study, the maxima and minima times were found to be in agreement with previous studies, with the addition of cirrus anvil cloud and MTH timings, which had not been explored in previous studies. However, the diurnal cycles of MTH and UTH are extremely similar regardless of seasons or surface types, which implies that the convective activities are less likely to be affected by advection. In their Lagrangian approach to analyse the cloud and humidity features, they have found the overland diurnal variation of high cloud (HC) to show a broad peak between 0700 and 1300. The Lagrangian analysis demonstrated that these high cloud systems could be driven away from land by upper level wind.

Convection and convective clouds are associated with higher surface temperature than the air. Therefore, in analysing the cloud cycles, the topography and heat capacity of the surface should also be considered. Pre-selecting the data according to surface features may be essential in characterising the specific diurnal cycle of OLR corresponding to different regions. It is shown in table 2.1 that in the papers

reviewed ([*Udelhofen and Hartmann, 1995*], [*Held and Soden, 2000*], [*Tian et al., 2004*], [*Chung et al., 2007*]), different types of humidity and cloud features possess different diurnal cycles, hence pre-selection of the data by these features could be a suitable process of analysis for the OLR cycle.

	Udelhofen and Hartmann (1995)	Soden (2000)	Tian et al. (2004)	This study	
Domain	15° N–15° S, 142.5° W–52.5° W	30° N–30° S, 120° W–30° W	30° N–30° S	30° N–30° S, 50° E–50° W	
Period	12–16 August 1991	June–August 1987	July 1999	July 2004	January 2005
Deep convection	–	L: 1800 (1000) O: 0600–0700 (2100)	L: 1800–2100 (0900) O: 0900 (2100)	L: 1800 (1100) O: 0700 & 1400 (2100)	L: 1700 (1000) O: 1500 & 0700 (2300)
High cloud	2000 (1000)	L: 1800 (1000) O: 1700 (0900)	L: 2100 (1200) O: 1600 (0600)	L: 2000 (1200) O: 1700 (0900)	L: 1900 (1100) O: 1800 (0900)
Cirrus anvil cloud	–	–	–	L: 0200 (1200) O: 1800 (0900)	L: 0200 (1200) O: 1800 (1000)
UTH	0600 (1800)	L: 2200 (1200) O: 2000 (1000)	L: 0300 (1500) O: 0000 (1200)	L: 0200 (1600) O: 0200 (1300)	L: 0300 (1500) O: 0400 (1500)
MTH	–	–	–	L: 0200 (1600) O: 0200 (1300)	L: 0300 (1500) O: 0400 (1700)

Table 2.1.: Table comparing time of maxima and minima (in parenthesis) of different humidity features from various studies, L and O corresponds to land and ocean, taken from [*Chung et al., 2007*]

Table 2.1 compared the maxima and minima times in various studies for a number of atmospheric clouds and humidity systems. It shows that there is a small degree of consistency in the times of the deep convection. However, there are large discrepancies between the other parameters, in particular, the UTH seems to show up to 8 hours of peak times differences.

Apart from convective cloud and tropospheric humidity, deep convective cloud is also an important component in the diurnal cycle due to the optical thickness (above 0.6) associated with these clouds and the manifestation of these clouds in the ITCZ region. *Nowicki and Merchant* (2004) have studied the diurnal cycle of the deep convection and the associated radiative forcing.

They have used OLR and outgoing SW radiation (OSR) derived from Meteosat-7, with a spatial resolution of 5km and temporal resolution of around 30 minutes during May and June 2000. They have discovered the diurnal cycle of deep convection to maximise in cloud coverage in the late afternoon (1600-1900LST) and minimises in the morning (0300-1200LST). Over equatorial Africa and east Atlantic, *Nowicki*

and Merchant (2004) have observed an averaged radiative forcing of $+13Wm^{-2}$ over land and $-39Wm^{-2}$ over the ocean. These radiative forcing are very significant when considering the global average flux of $0.9Wm^{-2}$ from 1.2. The individual events of the deep convection can cause a radiative forcing of 100 to $-300Wm^{-2}$ over the ocean and 120 to $-200Wm^{-2}$ over land. Therefore, we would expect the influence of the deep convective event to be present especially in our diurnal cycle analysis.

2.4. Two to nine days variability

In the tropical region, there exists different modes of variability which are beyond the daily timescale due to the complexity of the tropical dynamics. These variabilities have been observed in different parts of the globe. For example, Clayson *et al.* (2002) have observed a 2-3 day convective oscillation in the tropical western pacific where the OLR data from TOGA COARE [Yuter *et al.*, 1995] were bandpass filtered using a Lanczos filter [Duchon, 1979] and the 2-3 day variabilities were observed from November 1992 to February 1993, where both the amplitude and period of the variability are not consistent over the season. By studying a 11 years time period, Clayson *et al.* (2002) have also observed the strength of the 2-3 day variability to increases or decreases during El Nino/La Nina respectively.

The plot (Fig. 2.8) shows the filtered OLR from TOGA COARE where 2-3 day convective activity is distributed across the season over the tropical western pacific ocean. The sea surface temperature of the region has also been investigated by Clayson *et al.* (2002) and a similar 2-3 day variability was found during the convective events, indicating a possible thermodynamic feedback between ocean and atmosphere.

African easterly jet and the associated wave disturbances are some of the most prominent dynamic features in the local African meteorology, it has been proposed by various studies to promote cloud and rainfall in the region. The African easterly jet arises from the surface temperature gradient between the Sahara and the Gulf of Guinea which uplifts the air in the Sahara, leaving a low pressure near the surface that in turn creates thermal wind towards the Sahara and forms an easterly geostrophic wind component [Leroux, 2001]. This geostrophic wind component maximises at around 650mb to form the African easterly jet. The thermal wind

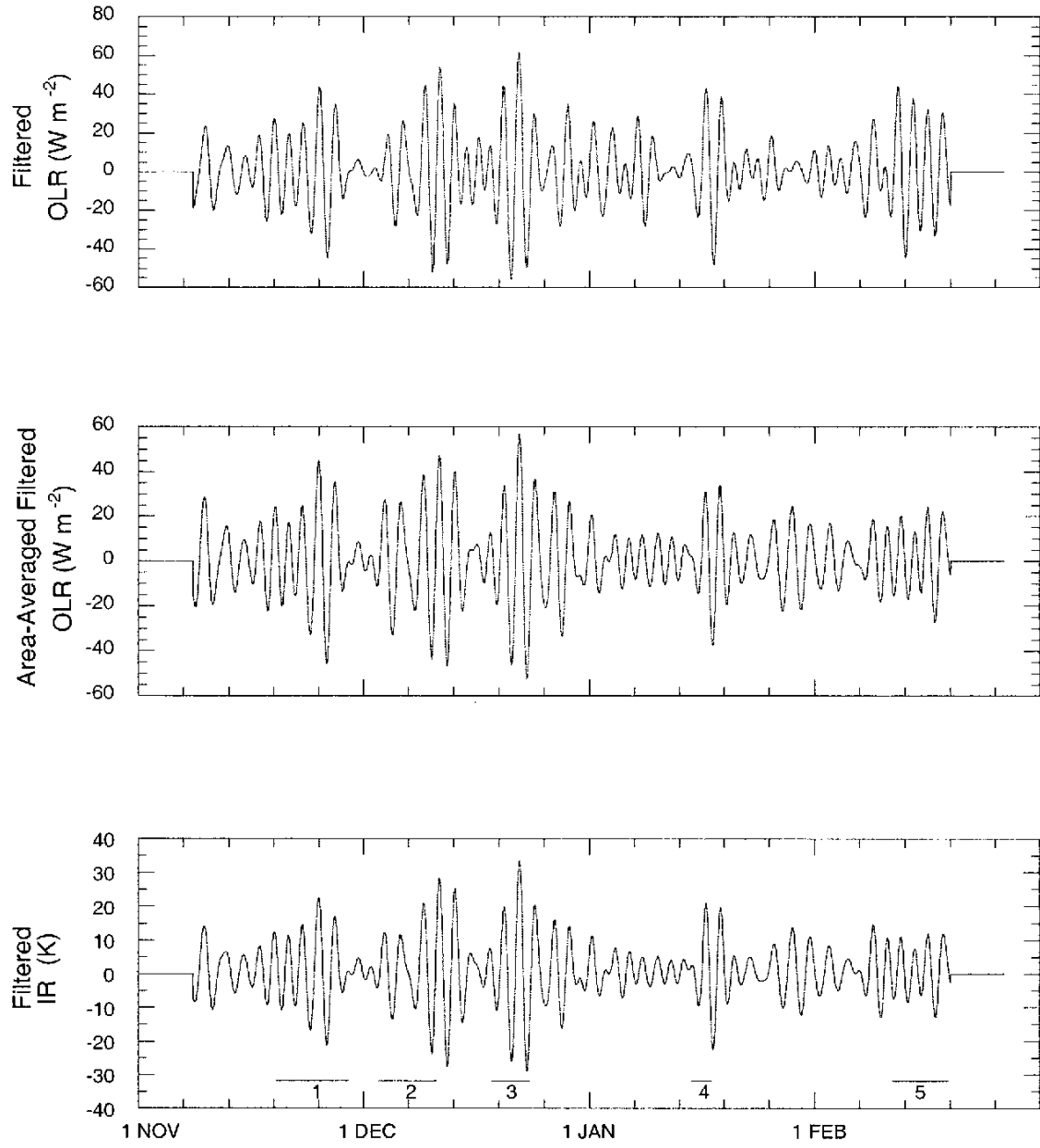


Figure 2.8.: (top) Filtered TOGA OLR, (middle) area-averaged filtered OLR, and (bottom) filtered IR. The five periods of 2-3-day convective activity during the IOP are as marked on the filtered IR. [*Clayson et al.*, 2002]

provides horizontal wind shear which creates barotropic instability and the vertical wind shear over the Sahara provides baroclinic instability. Thus, the jet creates an environment in which African wave disturbances develop through these instabilities [Cook, 1999] and affects the precipitation distribution in the region [Rowell and Milford, 1993].

Holton (1992) has used: the absolute vorticity profile for the African easterly jet to confirm the condition for barotropic instability; and the vertical wind profile to identify a strong easterly shear which causes baroclinic instability. Furthermore, Holton (1992) has also employed the observed wavelength of the African wave disturbances (2500km) and the propagation speed ($8ms^{-1}$) to deduct the period of the wave to be around 3.5 days.

Lavaysse *et al.* (2006) have studied the effect of convective activities over the west African land region (Sahel) during the monsoon season using NCEP reanalysis data, with attention to the 3-5 day African Easterly waves and convection and rainfall in wet and dry sequences. They have observed the AEW activities to increase and contribute to more of the total rainfall during the wet years. Nicholson (2009) has also studied the structure of "monsoon" and land ITCZ over West Africa using the same data, presenting an overview of the main sources of precipitation in Africa; the ITCZ and tropical rainbelt. The former being due to large scale circulation, the Hadley cell, which is only responsible for the rainfall in northern Sahel and southern Sahara. Whereas the latter is found to be the primary precipitating mechanism, governed by the African and tropical easterly jets (AEJ and TEJ respectively) and the rainfall is associated with the easterly wave disturbances associated with the jets.

There have been a number of studies on the wave disturbances specifically in Tropical Africa (e.g. Mekonnen *et al.*, 2006; Gu *et al.*, 2004; Grist, 2002 and Diedhiou *et al.*, 1999). The Africa wave disturbances have been identified in these studies to possess cycle lengths of 2-9 days, many of the studies are in slight disagreement with each other in terms of the period of the wave. However, almost all the studies have utilised wind field data to identify the jet/wave and many have associated rainfall with the presence of the wave. Therefore, it is important to also consider the radiative effect of wave disturbances since this is heavily associated with clouds and precipitation.

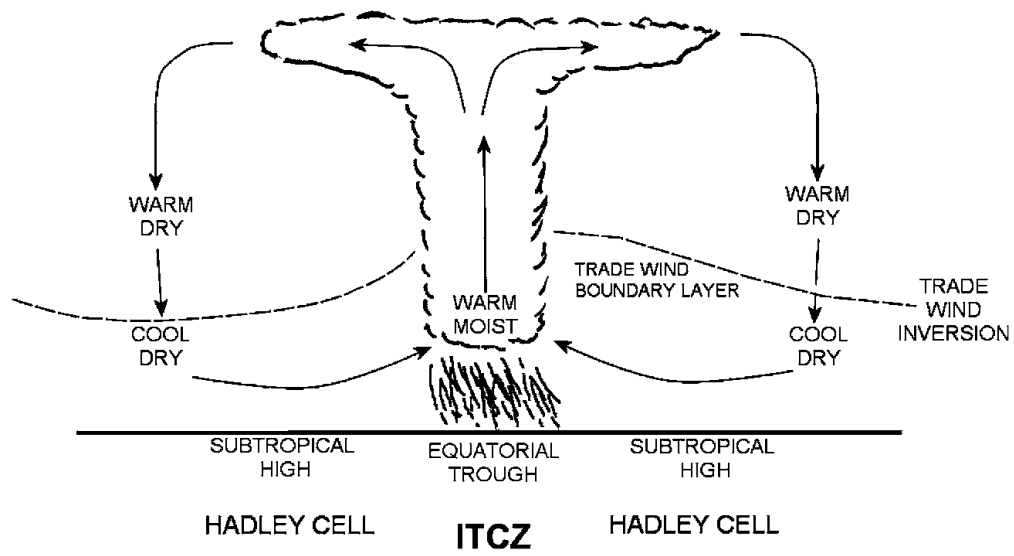
Nicholson et al. (2007) have examined the effect of wave disturbances on the tropical easterly jet over West Africa, using NCEP reanalysis data for 1983 and 1950 with a spatial resolution of $2.5 \times 2.5^\circ$ in both longitude and latitude. They have observed the waves to be of planetary scale with a period of 5-6 days and bigger than the waves associated with AEJ which are 3000-4000km.

Nicholson (2009) have studied the land ITCZ over West Africa based on NCEP reanalysis data. They assumed an ITCZ which is independent of the rainfall system and classed the association between rainfall and convergence zone to only southern Sahara and northern Sahel in abnormally wet years. They have investigated the surface pressure and wind over West Africa as well as the associated relative/specific humidity and the dew point.

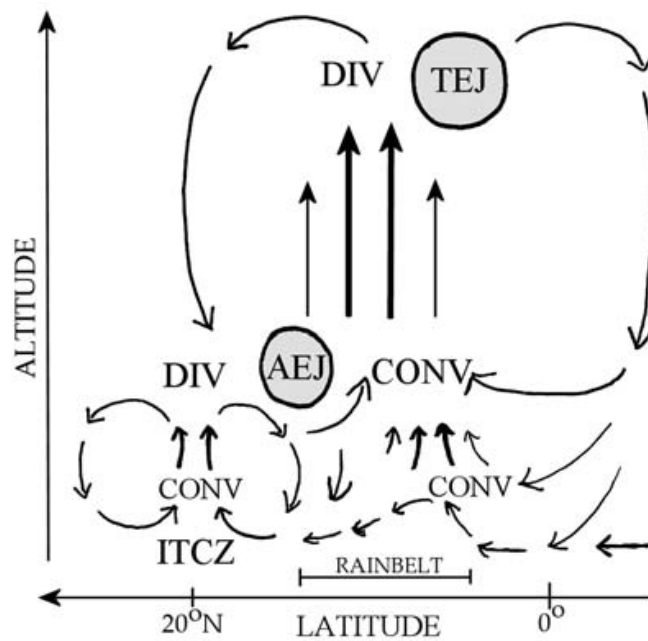
Fig. 2.9(a) demonstrates the convection solely driven by the minimum SZA at the convergence zone, where the trade winds in the subtropics flow in an equatorial direction to form the convergence. On the other hand, Fig. 2.9(b) describes a convective process which is independent of the ITCZ and is believed to be responsible for the majority of the precipitation in the West African region. This precipitation zone is known as the rainbelt which is the strong core of ascending air between the AEJ and TEJ indicated by the 4 upward arrows($\uparrow$) in Fig. 2.9(b). *Nicholson* (2009) suggested that this mechanism produces rainfall in the central and southern Sahel, where the rainfall pattern is coupled with the easterly wave disturbances observed over Central Africa and West Africa.

Nicholson (2009) have also suggested that the role of the Atlantic ocean in providing moisture for convective clouds and precipitation is a significant contribution to the overall precipitation in the region, which is consistent with previous studies (e.g. *Brubaker et al.*, 1993; *Savenije*, 1995).

An investigation on AEW has been carried out by *Diedhiou et al.* (1998) to study the frequency and amplitude of the wave over the West African region. They again utilised the NCEP reanalysis wind data to spectrally analyse the meridional wind component at 700hPa which is roughly where the AEJ is expected to be. *Diedhiou et al.* (1998) have found the power spectrum to show two peaks in the 3-5 day band period, where the most dominant power at 4.4-4.8 days exhibited by the AEW situated between 12.5°N and 20°N with a westward propagating velocity of $6-7^\circ$ longitude per day with a wavelength of roughly 3000km. Whereas, the TEJ was



(a)



(b)

Figure 2.9.: (a)Classic picture of ITCZ (b)Revised picture of ITCZ coupled with local land meteorology[Nicholson, 2009]

observed to have a 3-4 day oscillation at 5°N, and positioned at a higher altitude (200-150hPa) with a much higher propagation velocity, 11-12°.

Region	Primary data	<i>M</i>	<i>T</i>	<i>C</i>	<i>L</i>	References
Western Atlantic and Caribbean Sea	Sea level pressure, <i>u</i> and <i>v</i>	6-9	3-4	6.5	1500-2000	Riehl (1954)
	<i>v</i> (FGGE)	6-8	5.7			Nitta et al. (1985)
Eastern Atlantic and West Africa	<i>u</i> , <i>v</i> (radiosonde)	7-10	3-5	12	4000	Burpee (1972)
	<i>u</i> , <i>v</i> (radiosonde)	6-9	3.8-4.5		3100-3800	Burpee (1974)
	<i>u</i> , <i>v</i> (GATE)	8-9	3.5	8	2500	Reed et al. (1977)
	<i>u</i> , <i>v</i> (GATE)	8-9	3.5	8.4	2500	Albignat and Reed (1980)
	<i>v</i> (FGGE)	6-11	3.5		2000-3000	Nitta et al. (1985)
	<i>u</i> , <i>v</i> (ECMWF) and OLR	6-8	3-5			Lau and Lau (1990)
	<i>v</i> (ECMWF)	7-8	3-6			Druyan et al. (1997)
	<i>u</i> , <i>v</i> (ECMWF and NCEP-NCAR)	6-9	3-5		2000-4000	Diehieu et al. (1999)
	<i>v</i> (radiosonde)	6-10	2.5-6	7-9	6000	Pytharoulis and Thorncroft (1999)
	<i>u</i> , <i>v</i> (ECMWF and GCM)	5-10	3-5		2200-4400	Ceron and Gueremy (1999)

Table 2.2.: Basic characteristics of westward-propagating synoptic-scale disturbances in the African and Atlantic region [T: period (days); C: phase speed (m s⁻¹); L: wavelength (km); M: active months; u: zonal wind component; v: meridional wind component.][*Gu and Zhang, 2001*]

Gu and Zhang (2001) have presented in a table format (table 2.2), a collection of previous results on tropical wave disturbances. The table shows a large variation in the measurements where the periods, wavelengths and speeds are in slight disagreement between the published results. The disagreement could be due to the year to year variability, the different datasets employed and the different interpretation of the data.

Most of the papers reviewed in this chapter described the identification of the wave disturbances in Africa and some of the possible associations with rainfall. An article by *Taleb and Druyan* (2003) attempted to quantify the association by analysing data from Le Service de Météorologie Nationale Française, NOAA, WAMEX, and ECMWF. *Taleb and Druyan* (2003) utilised wavelet analysis to study the meridional wind field (*v*) at various pressure level (850, 700 and 600hPa) over Niamey, mainly where the AEJ is observed. The derived amplitudes are taken as evidence of periodicity divided into two wave periods of 3-5 and 5.5-9 days and they have compared the number of days with significant wavelet amplitudes (i.e. when the wave is observed) over period of a season (1950-84) to the precipitation amount recorded in those seasons measured

at ground stations. They have found the 3-5 days wave disturbances to account for only 30-40% of the total rainfall days, whereas the 5.5-9 days period accounted for 10-20%. The rest of the precipitation (30-50%) measured at the stations seem to have no correlation to the wave disturbances at 3-5 or 5.5-9 days time scale.

2.5. Chapter Summary

This chapter has reviewed the literature which investigated the short time scale (1-10 days) variability of the OLR and associated climate processes, particularly in the tropical region. The African tropical land region appeared to contain several land/atmospheric features which make the meteorology of the region unique and complicated. The main features observed in tropical Africa are the ITCZ and the rainbelt which are believed to arise from two separate mechanisms.

The diurnal cycle of the African region is interesting in that it contains some of the most extreme variability and its nature is poorly captured by models. Features include extreme day/night time temperature differences in the desert during boreal summer as well as the deep convection present in the ITCZ/rainbelt region. There are also numerous surface types which exhibit different diurnal cycles as well as the marine stratocumulus clouds off the west coast of Africa. Therefore, it's important to breakdown all the processes and investigate them individually in order to understand the local meteorology. There have been studies which investigated the diurnal cycles over the region. However, these lack direct OLR studies which characterises the short timescale cycles according to different variables, which this thesis aims to address. It is believed that by carrying out the analysis one can gain a better understanding of the processes associated with the different cycles of the OLR.

Furthermore, the ITCZ and the rainbelt have been observed to possess a longer cycle than the diurnal time scale in various literature. Some of which sub-categorised the rainbelt into AEW and tropical wave disturbances caused by the AEJ and TEJ respectively. These two types of wave disturbances are known to possess different cycle lengths along two different latitudes at different pressure levels (250hPa and 700hPa). There have been studies of the properties of the waves and the associated rainfall. In spite of the results, the direct effect of the wave disturbances on the OLR has yet to be analysed as this would show clearly a picture of the convection

associated with the wave and possibly enable better interpretation of the mechanism and the local meteorology.

The literature reviewed in this chapter have shown a lack of OLR variability analysis with respect to surface properties, especially in the short timescales due to the lack in availability of high time resolution OLR data prior to the GERB instrument. This thesis attempts to address this issue by analysing the short time scale variability over various scenario, in particular, the diurnal cycle which is one of the most fundamental modes of variability in the climate.

The overall diurnal cycle is influenced by both atmospheric processes and surface properties, thus our hypothesis is that surface properties will affect the surface diurnal cycle and the atmospheric processes related to the surface cycle, hence changing the overall diurnal variation. For example, a change in the surface heat response diurnal cycle is going to affect the associated cloud cycles across the surface type. The results chapters aim to capture the differences by utilising a high time resolution data-set, GERB, to identify the subtle differences across the different surface types and use PCA to identify sets of OLR components which better describes certain scenarios.

In the literature review, we have also looked at various studies which investigated the slightly longer timescale variability (up to 10 days). This is particularly important over the African continent as these variability are usually associated with AEJ and the corresponding wave disturbances. Most of these studies investigated the dynamics and the rainfall associated with these waves and variability, however, there appears to be a lack of studies on these waves in their manifestation on the OLR, particularly in using geostationary measurements which consistently captures the variability during the month/season. Therefore, in our first results chapter where we attempt to identify the most important modes of variability, we will also look at the manifestation of these longer modes in the OLR and how they appear geographically.

3. Data & Method

3.1. Introduction

This part of the thesis is intended to introduce the instrument/data used in the analyses as well as giving descriptions of the methods used in this study. We have primarily used data from the Geostationary Earth Radiation Budget (GERB) instrument to investigate the short timescale variability. A number of methods will be used to investigate the short time scale variation in the OLR. These methods will be described in this chapter.

3.2. GERB instrument

The GERB instruments are mounted on the MeteoSat Second Generation (MSG) satellite series, which are in geostationary orbits. Previous earth radiation budget experiments have been low earth orbit satellites which provide near global coverage but lack the temporal resolution available from geostationary orbit [Harries *et al.*, 2005].

The satellite MeteoSat-8 is positioned over 3.5°W and 0°N in a geostationary orbit with an altitude of roughly 36000km . The satellite itself hosts the GERB and Spinning Enhanced Visible and Infrared Imager (SEVIRI) instruments, where the first is a broadband instrument utilising a broadband detector and a quartz filter to distinguish between SW and total radiation, the latter one is a narrow band instrument with 4 visible/Near Infra-Red (NIR) and 8 IR channels. The GERB instrument has a spectral range of $0.32\mu\text{m}$ to $100\mu\text{m}$ which covers both the OLR and RSR from the earth TOA. Whereas, the SEVIRI instrument hosts multiple narrowband detectors which utilise the spacecraft rotation (100 rpm) to scan rapidly across the earth from east to west in lines (1249 scan lines) covering the whole earth

image.

The GERB instrument consists of a despin mirror and four further optical mirrors, a quartz filter, a black body and a linear, 256 element detector array [Harries *et al.*, 2005]. The despin mirror was designed to counteract the rotation of the satellite to produce a steady image on the detector, whereas the quartz filter is responsible for filtering out the LW to achieve a SW observation. The detector array scans across the earth to produce an image which is calibrated and converted to radiance using the on board black body radiation source (see figure 3.1 for schematics)

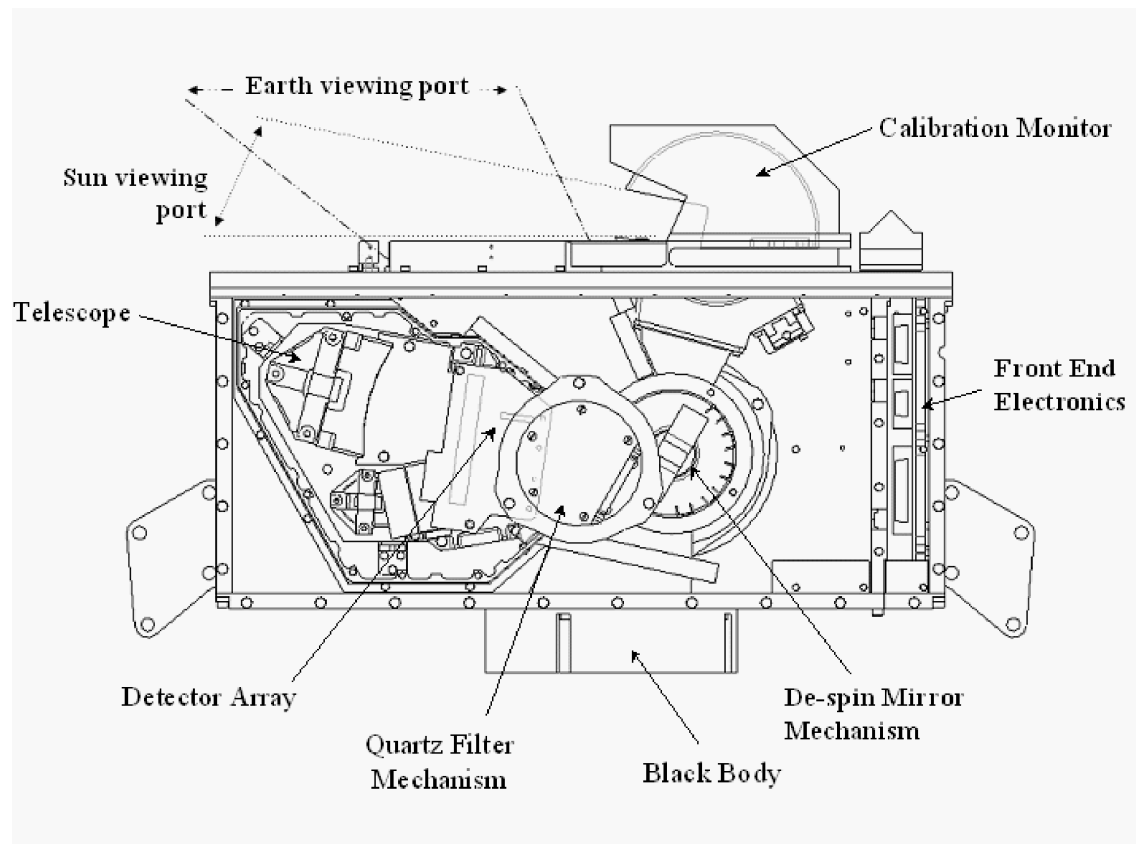


Figure 3.1.: Layout of the GERB optics unit, as seen from above [Harries *et al.*, 2005]

The GERB instrument produces a single Earth image in a little under 3 minutes, and derived products are based on 6 images (3 in both SW and total channels) acquired over 17 minutes. The scanner produces the shortwave and total TOA irradiance products roughly every 17 minutes which is comparable to the narrowband

measurements of the SEVIRI instrument at every 15 minutes. The SW measurement is obtained by using the quartz filter to eliminate the LW part of the spectrum.

The LW and SW flux product have a sampling distance of 50km producing a 256x256 image of the whole Earth disc. The combined errors of instrumental, SEVIRI inter-channel calibration, polarisation and other factors are $\pm 2.25\%$ and $\pm 0.96\%$ for GERB SW and LW radiance products respectively. These error amounts are calculated by taking the root mean square combination of the individual error percentages at one standard deviation.

The GERB-2 instrument was in operation from 28th March 2004 to 10th May 2007, since 10th May 2007, the GERB-1 instrument on MeteoSat 9 has continued to collect data. The data from GERB-2 is available for all GERB users, while pre-release GERB-1 data product is only available to validation team members [GGSPS, webpage]. Nevertheless, the GERB-2 instrument has provided several years of high temporal resolution radiation data which is invaluable for studying short term variation in the Earth's radiative budget.

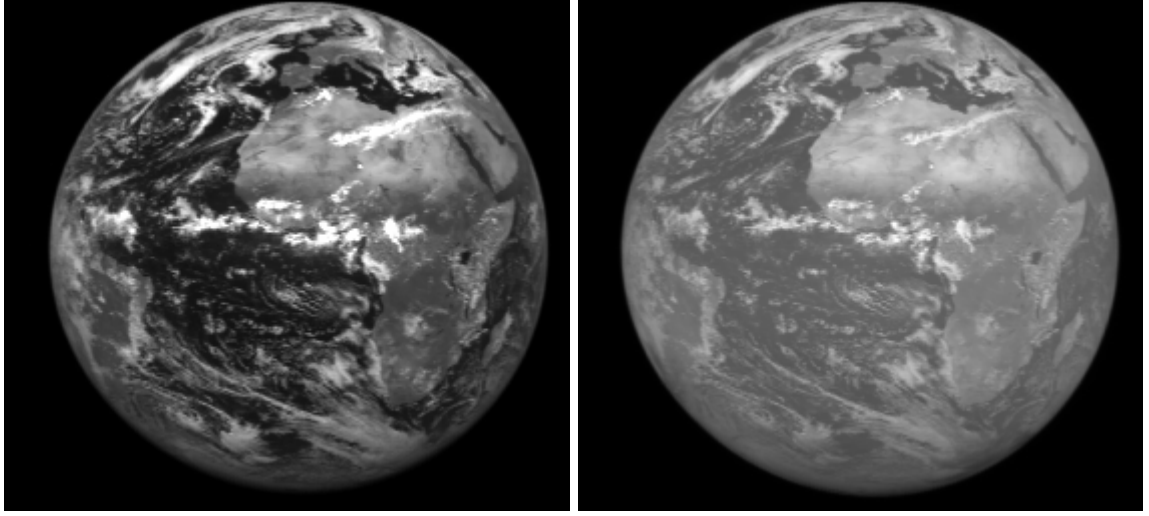
3.3. GERB data

The GERB instrument produces LW and SW products. The SW irradiance measurements are converted to SW flux using SEVIRI data for scene identification, whereas the LW irradiance is derived by subtracting SW from total measurements, and subsequently converted to flux using SEVIRI data.

The derivation of the flux product for the Geostationary Earth Radiation Budget (GERB) instrument involves the use of SEVIRI data (narrowband channels) to provide good scene identification, this is a necessary condition for applying the correct ADM. The flux product of GERB is determined using the equation [Dewitte et al.]:

$$F_{GERB/SEVIRI} = \frac{(F_{SEVIRI} L_{GERB}^f)}{L_{SEVIRI}^f} \quad (3.1)$$

Where L_{GERB}^f , $F_{GERB/SEVIRI}$ correspond to radiance and flux measurements of GERB respectively, L_{SEVIRI}^f and F_{SEVIRI} are the SEVIRI estimated broadband radiance and flux. The SEVIRI broadband radiance is estimated from spectral radiance by using the Santa Barbara DISORT Atmospheric Radiative Transfer



(a) SW channel view

(b) Total channel view

Figure 3.2.: View of Earth as seen by GERB (midday, 18th May 2004) in SW(left) and Total(right) channels[Bates *et al.*,]

(SBDART) model. This radiance is then converted into flux using eq.1.3, where the ADMs used for SEVIRI flux conversion are the ADMs scene types derived from the CERES instrument on the TRMM satellite. The GERB flux product ($F_{GERB/SEVIRI}$) is then calculated by taking the ratio between the SEVIRI estimated radiance and flux.

The LW flux product was derived in the same manner using the ratio of SEVIRI estimated broadband radiance to flux. However, prior to the flux conversion, the LW radiance is obtained using the following equation 3.2 [Dewitte *et al.*,].

$$L_{LW}^f = L_{Total}^f - A.L_{SW}^f \quad (3.2)$$

where L_{LW}^f , L_{Total}^f , A and L_{SW}^f correspond to the longwave radiance, total radiance, black body calibration factor and the shortwave radiance. The factor A is defined such that L_{LW}^f is zero when a black body of 5800K (close to the solar spectrum) is observed [Dewitte *et al.*,].

The GERB geolocation data is required to match the GERB pixel to SEVIRI pixels to enable accurate flux conversion using the equation(3.1) mentioned above [Bates *et al.*,]. Accurate measurement of the scan mirror operation is important

for geolocation accuracy, mirror positioning is repeatedly measured to an accuracy of 0.05 GERB pixels [Bates *et al.*,].

There are data gaps in GERB data due to various reasons, operational/instrumental issues, sun avoidance periods, and the sun glint region in the shortwave channel (specifically over ocean region). Sun avoidance periods refer to the eclipse of the sun at equinoxes where the GERB detector array would be directly illuminated by the sun at certain time which if occurs, would cause permanent damage to the detectors. The GERB instrument is safeguarded from these events and therefore, results in a few hours of data gap (per day) a few weeks before and after the equinoxes [Harries *et al.*, 2005].

In our analysis, we have selected the months (01, 06, 07, 12) to eliminate any sun avoidance periods due to the persistent nature of the data gap during the night time which could potentially impose a bias in our results.

We have chosen the GERB ARG product for our analysis, the data has been interpolated using a linear interpolation method (except for the data gaps). The data used in the study has suffered some of the mentioned issues, however, the data gaps have been interpolated in our analysis to minimise the impact of the missing data in the outcome of the results (i.e. setting zero/Not a Number (NaN) to the missing data would create a false "square wave" signal which would greatly affect the outcome in Fourier analysis).

3.4. Principal component analysis technique

3.4.1. Introduction

The PCA employed in previous studies have been described in section 2.2.1 where the studies have utilised the technique to decompose some of the most important components within the diurnal cycle.

Principal Component Analysis technique is a very useful statistical analysis in finding a recursive pattern in multiple dimensions [Smith, 2002]. It is a technique which transformed a set of correlated data into a set of orthogonal vectors according to their covariances. For example, the first transformed vector, known as the first principal component (PC) represents the highest percentage of variance within

the data set, the subsequent principal component will represent the next highest percentage of variance, and so on.

Principal component analysis is carried out by using eigenvalue and eigenvector decomposition of the covariance matrix which will be described in the next section (3.4.2).

3.4.2. OLR PC analysis

In this analysis, we have used GERB ARG product, which linearly interpolated to 15 minutes time resolution. We have also used the same interpolation method to address the data gaps. However, if the data gap is large than 10% of the day, then that particular is discarded from the dataset, the data is then averaged over the remaining number of days. This is done to ensure we don't introduce any significant noise into the PCs.

The analysis technique applied was similar to which *Comer et al.* (2007) used in their study: the computation of principal components (PC) and the associated empirical orthogonal functions (EOF). A month of data was used to average into mean diurnal cycles, which is done by averaging the 30 days of data in a single day according to their local solar time for each individual GERB footprint. The measurements from each pixel over time are regarded as a single column of data, and each of the columns are then subtracted from their column mean to create a deviation matrix D_{ij} , where i and j denote the rows and columns within the matrix respectively. A covariance matrix C is formed by taking the covariances between each row of data, hence the covariance matrix is 96 by 96 in dimension (see eq.3.3). The dimension 96 represents the time steps during a day with 15 minutes time gap between each measurements.

$$C = \begin{pmatrix} cov(1,1) & \dots & cov(1,i) \\ \dots & \dots & \dots \\ cov(i,1) & \dots & cov(i,i) \end{pmatrix} \quad (3.3)$$

where $cov(a,b)$ represents the covariance between rows a and b of the deviation matrix [Smith, 2002]. As shown in eq.3.3, the covariance matrix is symmetrical about the diagonal. It is then required to work out the eigenvectors and the eigenval-

ues of the covariance matrix. The eigenvectors are arranged according to value of the eigenvalues, with the eigenvector which has the highest eigenvalue to be the most significant principal component of the dataset and vice versa. The variance of each principal componenet is given by their associated eigenvalue and the PC are rearranged according to their variances.

The EOFs associated with the PCs are calculated by:

$$EOF(i) = \mathbf{e}_i \mathbf{D}_{ij} \quad (3.4)$$

where $EOF(i)$ is the empirical orthogonal function which corresponds to the i^{th} PC and $\mathbf{e}_i$ represents the i th eigenvector (PC). The eigenvectors represent the transformed factors which contribute to the most amount of variance in descending order (i.e. first PC represents most variance), whereas the eigenvalues express the variance associated with the eigenvectors.

The data can be re-derived using all/some of the PCs and corresponding EOFs: using all the PCs and EOFs will yield the original dataset, whereas, using only certain components will yield a modified dataset which only projects the dataset in the components which have been chosen [Smith, 2002]. For example, when only the most significant PC has been chosen to re-derive the dataset, the entire derived dataset will consist of functions of only the first PC in different magnitudes. The equation to re-derive the dataset is as follow:

$$D_t = PC(n)^T \times \Delta D \quad (3.5)$$

where $PC(n)$ denotes the chosen PCs to reformat data, ΔD represents the deviation matrix which is the original dataset subtracted from the mean. The transformed dataset (D_t) has the same spatial dimension as the deviation matrix and the temporal dimension n , the number of PCs chosen. To retrieve the full dataset in terms of the chosen PCs, the following equation is used:

$$D_f = PC(n) \times D_t + D_\mu \quad (3.6)$$

The transformed matrix D_t is further multiplied by the $PC(n)$ matrix to retrieve the full dataset dimension of ΔD expressed in terms of the chosen PCs. The final

step to get the final modified dataset is to add the mean of the original dataset which has been subtracted at the beginning of the PCA analysis.

In terms of application of this analysis in this study, the PC is a representation of the diurnal variation on a daily timescale. Whereas, the EOF is the spatial expression of the magnitude of that particular PC varying at each specific location (pixel).

We expect to see the first few components to correspond to some of the most significant climate processes in the chosen timescale. It is expected that by observing the PCs and the EOFs which expresses the temporal and spatial distribution of the OLR characteristics respectively, the correspondent climate processes can be understood better with the analysis. Especially, with the analysis carried out over chosen subsets of the data, the temporal and spatial variation in the climate processes regarding the different subsets can be compared and hence, we can gain a better understanding of the processes involved.

3.4.3. Uncertainty and limitations

The PCA is very powerful in identifying the repeating patterns especially when they are not a sinusoidal variation. The PCA method decomposes the dataset according to the variances and is able to represent the variances in both temporal and spatial distribution which is extremely useful in data analysis.

However, there are limitations to the method which have been previously investigated by *Smith and Rutan* (2003) in application to radiation studies. *Smith and Rutan* (2003) have done so by utilising a method proposed by *North et al.* (1982) who derived a criterion for validating an EOF. This method assumes each realisation is independent, which is unlikely to hold in radiation observations. Although this validating method is not 100% accurate, it does provide a good approximation of validating criterion. Despite the limitation to the method, it does indicate the robustness of the technique used in radiation studies.

The validating method involves the use of the standard deviation of the error of the eigenvalue, which is defined as [*Smith and Rutan*, 2003]:

$$\delta\lambda_i = \lambda_i \left(\frac{2}{N}\right)^{1/2} \quad (3.7)$$

where $\delta\lambda_i$ denotes the standard deviation of the i eigenvalue(λ_i), and N corresponds to the number of data points, which in our analysis represents the number of GERB pixels selected for the study.

North et al. (1982) have defined the criterion where if a particular EOF contains only a small contamination of variances from other EOF, this EOF is considered to be valid. This is quantified by the equation:

$$\delta\lambda_i < \Delta\lambda_i = \lambda_i - \lambda_{i+1} \quad (3.8)$$

where the criterion, is defined as the standard deviation of the eigenvalue, has to be smaller than the difference ($\Delta\lambda$) of that eigenvalue and the next for the correspondent eigenvector to be valid.

Furthermore, EOF according to LST cannot capture propagating events(such as advection). There are various ways which EOF can be modified to track propagating signals, such as complex EOFs which are based on the notion that a propagating signal should contain signals that are orthogonal to each other [*Gille, 2012*], this involves computing the Hilbert transform of the data and then calculating the complex EOFs for the transformed data. Propagating signals can also be captured by extended EOFs [*Gille, 2012*], where it's assumed that an event occurring at a single point will occur sometime later at another point, this requires creating new data sets which are time shifted and then compute the signals from these time shifted sets.

A simpler alternative method can be employed to allow the EOF analysis to track propagating features, this is done by correlating the PCs with a time lag. However, this requires prior knowledge of the time lag and the propagating speed to isolate the phase of the propagating wave. However, this is beyond the scope of this thesis and will not be included in our study.

3.5. Fourier transform analysis technique

Fourier Transform Analysis is a well known technique and it's incredibly powerful in data analysis. In our study, we utilise Fourier's technique to transform time series into frequency domain to allow us to observe some of the significant variabilities

in our data. A month of data was used in this study, each time series having $96 \times 30 = 2880$ time steps with 15 minutes between each point. The data is then spatially averaged in $2 \times 2^\circ$ boxes to attempt to remove some of the noise arising from the localised variability.

The fundamental equation of Fourier transform is as follows:

$$F = \int_{-\infty}^{\infty} f(t).e^{-2\pi\nu t.i} dt \quad (3.9)$$

where F denotes the transformed function as a function of frequency(ν), $f(t)$ represents the original function of time. Equation 3.9 demonstrates the continuous Fourier transform by integrating the original function of time with a sinusoidal function in the time domain, which essentially split the function $f(t)$ into sum of sine/cosine wave.

This transform can be reversed by a simple integral using the relations between sine and cosine expressed by Euler's formula ($e^{i\theta} = \cos\theta + i\sin\theta$):

$$f(t) = \int_{-\infty}^{\infty} F.e^{2\pi\nu t.i} dt \quad (3.10)$$

The complex exponentials produce complex Fourier coefficients which express the amplitude and phase of the waves. The Fourier transform expression described above transforms continuous functions from one domain to another, this method can be generalised to analysis of discrete numeric datasets.

Discrete Fourier Transform (DFT) is capable of transforming discrete values (time series) into the frequency domain, using the equation [ITTVIS, 2009]:

$$F_m = \sum_0^{N-1} f_n e^{\frac{-2\pi n m.i}{N}} \quad (3.11)$$

where f_n represents the n th component of the time series with N number of data points, and F_m denotes the correspondent(m th) transformed complex number in frequency domain. The DFT can be reversed similarly using the inverse DFT:

$$f_n = \frac{1}{N} \sum_0^{N-1} F_m e^{\frac{2\pi n m.i}{N}} \quad (3.12)$$

Fourier analysis is an invaluable tool in our analysis as it can be utilised to transform the time series into frequency power spectrum which is especially useful in analysing any frequencies that are significant and fundamental in the data.

3.5.1. Uncertainty and limitations

There are several limitations to this method which need to be considered carefully whilst using the technique. First is the inability to fully isolate a signal within its timescale, i.e. in theory a sinusoidal wave is required for the spectral outcome to show only one frequency response, any other wave format will be transformed into a range of frequencies. This is the nature of the technique in which the transform utilises the sine/cosine functions to represent the original function.

Second is the effect of white, pink, and red noise on the analysis. White noise is a random signal which affects all frequencies and is uncorrelated in time, whereas pink and red noise are both frequency dependent. Pink noise is also known as the $1/\nu$ noise where the frequency response is proportional to the function $1/\nu$. The $1/\nu$ noise is observed in many physical measurements and the cause is still not well understood. On the other hand, the red noise is proportional to $1/\nu^2$. Although noise is part of the error present in the measurements, both the pink and red noises are inversely proportional to frequency, which means the lowest frequencies part are most affected too. This would become a major problem if the study attempts to observe long term variation. However, our study focuses mainly on the shorter timescale which is the least affected part of the spectrum, therefore, it is believed that this noise effect would only play a minor role in our measurement errors. Nevertheless, the frequency dependent (in particular, $1/\nu$) noise is an integral part of our observation.

The $1/\nu$ noise has been discussed as part of the tropical convective variability in [Yano *et al.*, 2001], they have observed a $1/\nu$ noise in the convective variability. However the physical mechanism for which is not yet understood. In our analysis, we expect to see this $1/\nu$ noise to manifest in the Fourier analysis especially in the ITCZ. However, this $1/\nu$ signal is associated with convective variability which is part of what we are trying to observe in our analysis.

Therefore, it would be more appropriate for us to look at the strong robust signals

across the years to determine the significant modes rather than using $1/\nu$ to look for the peaks above this curve as this curve is part of the convective variability.

Third is the end affects in Fourier transform. When a finite time series is transformed, the spectral outcome is affected by the start and end of the time series. This occurs when there is an offset between the starting and ending points which create discontinuity in the function, hence affecting the frequency spectral. For example, when a simple sine wave is transformed using DFT, the spectrum outcome differs when the sine wave starts and ends at the same/different points.

This end effect is caused by the fact that DFT assumes the function $f(n)$ is periodic with the period of N [Sundararajan, 2001], essentially treating the $f(n)$ as a section of an infinite series. Hence, any offset in the begin and end values would create a discontinuity which would trigger a multiple frequency response in the power spectrum, analogous to the Fourier transformation of a square wave.

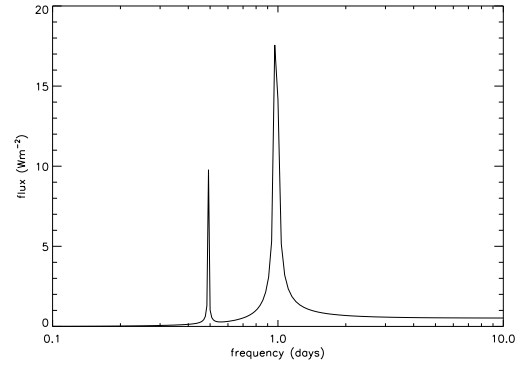
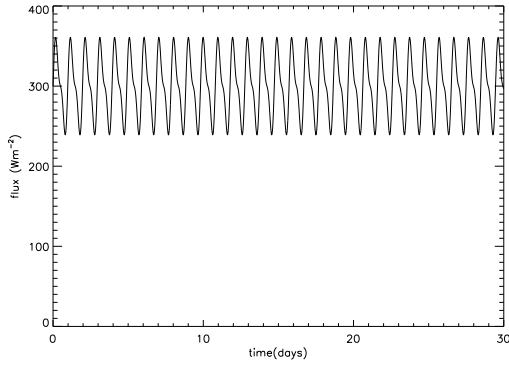
Under our investigation, the same end effect would influence our results when a discontinuity is present in the data. Therefore, in our analysis, we have attempted to interpolate the data gaps linearly to remove the noise caused by the end effect. This was found to sufficiently eliminate most of the end effect noise, particularly towards the low frequency end.

3.5.2. Interpolation of missing data

In the GERB data, there exists various gaps, for example, when the instrument is either turned off to protect it from direct solar illumination which usually lasts for a few hours per day during the whole month, or short term malfunction of the equipment. In our analysis, we have selected the months (01, 06, 07, 12) to eliminate any sun avoidance periods due to the persistent nature of the data gap during the night time which could potentially impose a bias in our results. However, there are still smaller data gaps that we have to interpolate before our analysis.

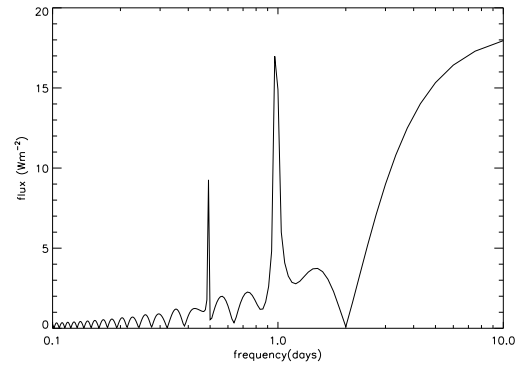
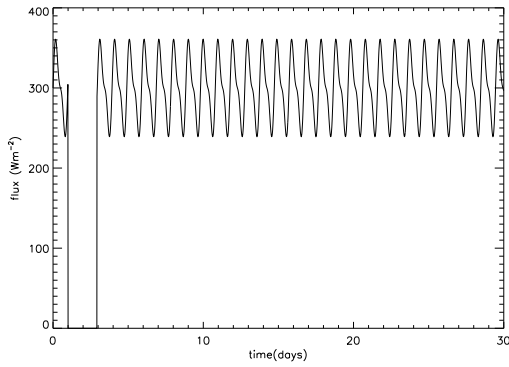
These gaps if left unattended could influence our results particularly in the longer period end of the spectrum.

Fig. 3.3 shows various plots of artificially created data using two modes of Sinusoidal wave over a period of 30 days, and data gaps with/without interpolation with their respective Fourier spectra. Fig. 3.3(a) has no data gaps, we have used Fourier



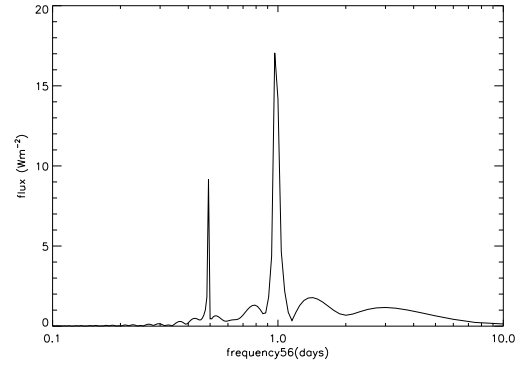
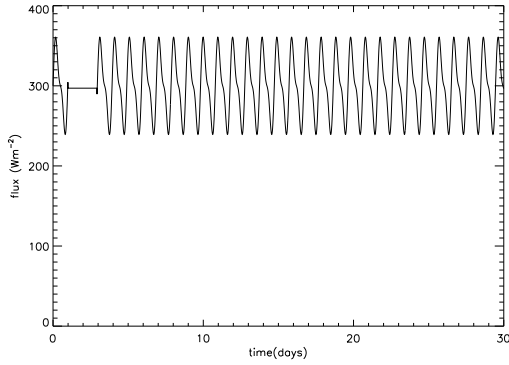
(a) Sinusoidal wave over 30 days without data gap

(b) Fourier spectrum of Sinusoidal wave over 30 days without data gap



(c) Sinusoidal wave over 30 days with data gap

(d) Fourier spectrum of Sinusoidal wave over 30 days with data gap



(e) Sinusoidal wave over 30 days with data gap + interpolation

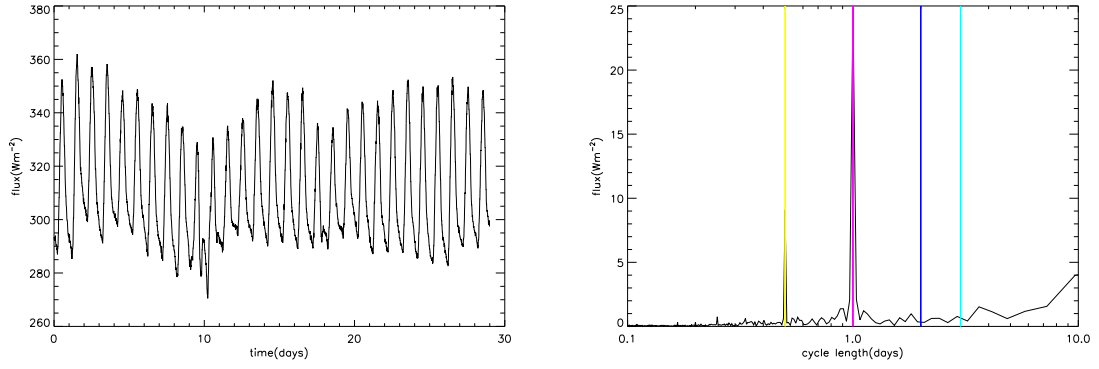
(f) Fourier spectrum of Sinusoidal wave over 30 days with data gap + interpolation

Figure 3.3.: Artificial data gap testing with and without interpolation using Fourier spectrum a,b)no data gap + spectrum; c,d) data gap + spectrum; e,f) data gap(interpolated) + spectrum

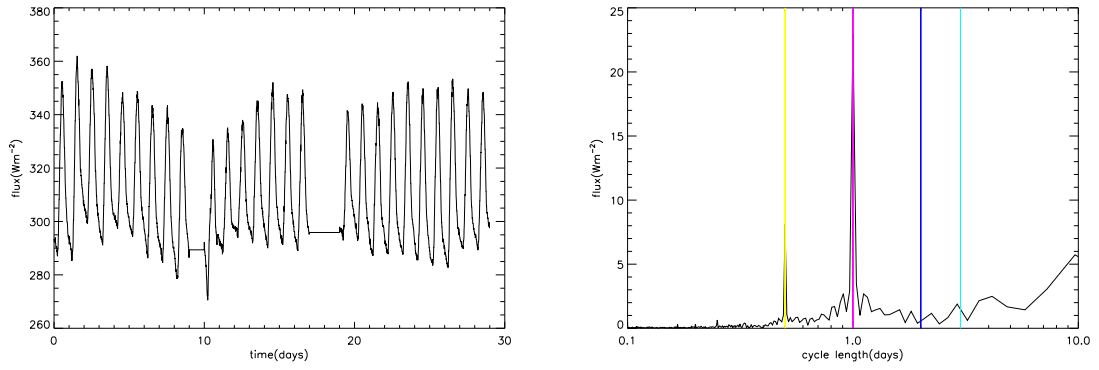
analysis to produce a control test in Fig. 3.3(b). In Fig. 3.3(c), part of the sine wave is removed to imitate the absence of data, the corresponding Fourier spectrum (see Fig. 3.3(d)) shows two peaks which corresponds to the two modes of variability that we have input. However, besides the two peaks, there exists an increasing Fourier signal with period. This is caused by the untreated data gap, the presence of the data gap is effectively the same as Fourier transform of a square wave. This creates a *sinc* function centring at 0 frequency, which is towards the higher period end of our spectrum (10 days). In our Fourier spectrum, we only observe the amplitude of the transformed product, therefore, creating the pattern that we see in Fig. 3.3(d) ($\sqrt{\text{sinc}^2}$).

This effect of data gap can be mitigated by a simple averaged interpolation of the data, as shown in Fig. 3.3(e) and 3.3(f). The data gap is interpolated simply by taking an average of before and after the data gap, the corresponding spectrum has shown the a pattern which closely resembles the spectrum with no data gap. We have also tested other interpolation methods (linear, quadratic) which haven't not been presented in this thesis. We have found in most cases, there are no discernible differences between the averaged interpolation and other methods. In certain cases, the other methods appeared to be worse than averaged, dependent on the size of the data gap.

We have extended the test further on real GERB data with more than one data gap to investigate the effect of data gap presence on the Fourier spectrum. Fig. 3.4 shows a GERB data point over the desert, with and without data gap and their respective spectra. We have observed in our analysis that the main components, diurnal and semi-diurnal to have dropped roughly $2Wm^{-2}$ and $1Wm^{-2}$ in magnitude from their original fluxes of $25Wm^{-2}$ and $9Wm^{-2}$. Furthermore, the lower frequency end of the spectrum appeared to have risen in magnitude by $1-2Wm^{-2}$, however, the important modes of variability are still observable. We have removed roughly 10% of the data in the month and observed a roughly 10% drop in Fourier spectrum magnitude. It should be noted that for the months that we have selected for our analysis, there are no data gaps which are close to this size. Therefore, we are confident our analysis will provide significant results in spite of the data gaps.



(a) GERB data (25°N, 15°E) wave over 30 days without data gap (b) Fourier spectrum of GERB data over 30 days without data gap



(c) GERB data (25°N, 15°E) wave over 30 days with data gap (d) Fourier spectrum of GERB data over 30 days with data gap

Figure 3.4.: GERB data gap testing with and without data gap using Fourier spectrum a,b)no data gap + spectrum; c,d) data gap(interpolated) + spectrum; the data gaps are 1 day and 2 days in period

3.6. Band pass filter

The band pass filter analysis is particularly useful when a signal is expected to be in a certain frequency range, by filtering the unwanted signals, the time series can be easily observed and studied.

The technique used in this study involves using a Fourier transform to transform the signal to the frequency domain and filter out the frequencies not within the chosen frequency range. The filtering function can be expressed as follows [ITTVIS,

2009]:

$$H(\nu) = \begin{cases} 1 & \text{if } D_L < D(\nu) < D_H \\ 0 & \text{otherwise} \end{cases} \quad (3.13)$$

where D_L and D_H denotes the lower and upper frequency limit respectively, $D(\nu)$ and $H(\nu)$ represents the chosen frequency and correspondent filtering component respectively. The function 3.13 essentially sets all the Fourier transformed power outside of the chosen frequency range to 0.

The filtered frequency spectrum will then be reverse transformed using the Fourier transformed method mentioned in section 3.5 to obtain a filtered time series.

It is expected that by using Fourier transform and filtering technique, we can observe the expected frequency cycles to be present in certain regions and absent in others, and therefore, confirm the corresponding climate process to be regional and the amplitude associated with the process can also be observed.

3.6.1. Uncertainty and limitations

The band pass filter technique can identify and confirm the cycles which we expect to observe by removing the unwanted signals. However, it does have some limitations which overlap with some limitations of the Fourier technique. It shares the same problem with FT which is the inability to create one frequency response to any non-sinusoidal waves, which implies a more sinusoidal biased signal would be observed when the filtered spectrum is reversely transformed.

Whilst the noises and end effects are also inherited by the band analysis, the frequency spectrum would be filtered and reversely transformed, hence the overall effect of these is minimal. The sinusoidal biased behaviour is not a major drawback in our study since it is not expected to quantify the absolute magnitude of the wave from this analysis. Instead, the band pass study aims to focus on the variation in amplitude and wavelength of the wave.

3.7. Model method

To aid in the interpretation of the results of the fourier and PC analysis a simple heat balance model has been developed. The model consists of the Incoming Solar

Radiation (ISR), a clear sky atmosphere, a heat response surface layer which radiates upwelling longwave flux, and a conductive layer that thermally conducts with the surface layer (see fig. 3.5).

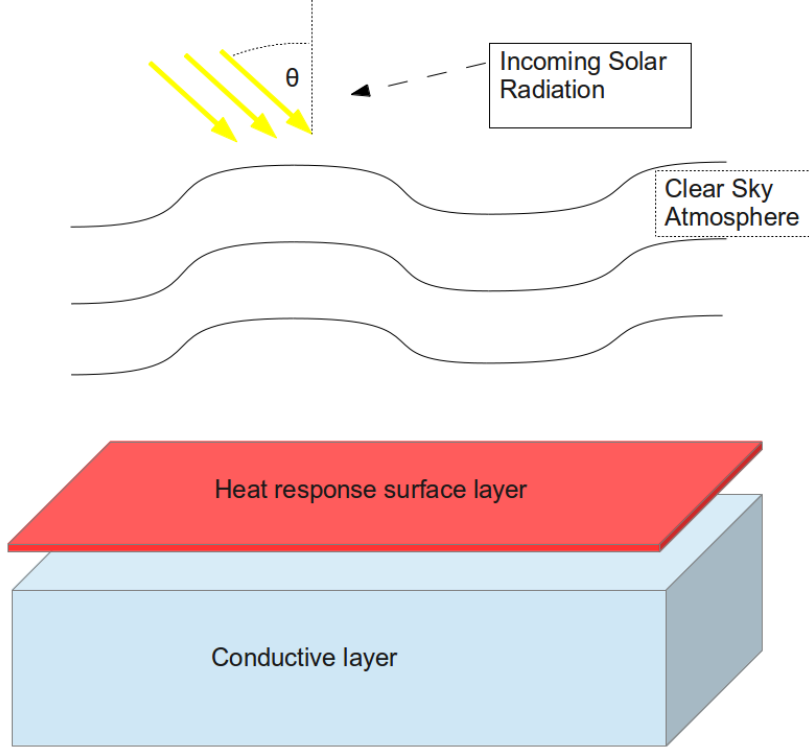


Figure 3.5.: Diagram of the model with ISR, clear sky atmosphere and a surface

The model initiates with a set arbitrary temperature ($293K$) for the surface, the heat capacity is set according to the specific heat capacities for soil, sand and other materials [Arya, 2001]. For example, a desert surface type is interpreted as sand in the model with a specific heat of $835 \text{ Jkg}^{-1}K^{-1}$. The ISR time series is taken from GERB shortwave data, a portion of which is reflected by the surface (25%) [DeLiberty, 1999].

The incoming solar flux (F) is converted into energy (E) using the equation:

$$E_{in/out} = F.A.dt \quad (3.14)$$

The area (A) is defined as $1m^2$, and the time (dt) is the model time step (15 minutes) in seconds. The ISR data is taken from the GERB file data and re-sampled to 15 minutes resolution by linear interpolation to synchronise with the model time steps.

The input ISR varies according to the solar zenith angle (θ) indicated in Fig. 3.5. This SZA is dependent on the latitude of location, the season and the time of day. This ISR variation with respect to the SZA is already integrated into the GERB ISR file. Therefore, we were only required to take this ISR data and linearly interpolated it to 15 minutes interval for our use. Note that at the illumination end points of the diurnal cycle (i.e. dawn, dusk), we have chosen to simply re-bin the points to eliminate any tailing effects from linear interpolation.

The incoming solar radiation is governed by the SZA in the following expression:

$$I = S \cos \theta \quad (3.15)$$

where I is the incoming solar radiation, S is the solar constant. The ISR (from GERB) then strikes the surface with the assumption of a clear sky scenario which in turn provide energy to the surface. In the clear sky scenario, we have assumed a fixed combine atmospheric absorption and scattering to be a factor of 0.2. We recognise the atmospheric path length changes during the day, and the atmospheric absorption is dependent on it. However, we are only modelling the OLR close to the equator (30° in latitude), and under the assumption that we do not expect the delay between peak in surface heat response and the peak in ISR to exceed 3 hours (i.e. any change in ISR outside this time limit would not influence the peak time) which is under what we have observed in our surface type PCA. This would give us equivalent of a time angle of 45° . The SZA will be around 41° , which gives roughly an extra 32% in atmospheric path length. Thus, the difference of a varying atmospheric path length will only give us a few percent more absorption/scattering in the SW around this time.

The change in temperature (ΔT , K) is then calculated using the thermal energy (Q , in J) equation:

$$Q = m.c.\Delta T \quad (3.16)$$

where m (kg) is the mass of the surface which is given by the density multiplied by unit area ($1 \times 1m$) and depth ($2.5cm$); c ($Jkg^{-1}K^{-1}$) represents the specific heat

capacity of the material. The change in temperature (ΔT , K) is added onto the original temperature for each time step. For example, the temperature after the first time step would be $293K + \Delta T_1$. The density of the surface layer for the desert type is $1600kgm^{-3}$ with a heat capacity of $800Jkg^{-1}K^{-1}$ [Arya, 2001].

This temperature of the surface and atmosphere is then used to calculate the radiative flux (F) using the Stefan-Boltzmann law:

$$F_{out} = \epsilon\sigma T^4 \quad (3.17)$$

where F_{out} signifies the upwelling shortwave flux, T is the temperature in K and ϵ is the emissivity. The flux is converted into energy using equation 3.14 and the net energy is recalculated at the beginning of every model time step. We have used a surface emissivity of 0.85 [Ogawa *et al.*, 2008] and an atmospheric emissivity of 0.68 [Herrero and Polo, 2012].

The conductive layer consists of a slab which bares the same surface properties as the surface and is responsible for conductive heat transfer with the surface layer. The heat transfer follows the Fourier's law of thermal conduction in single dimension given in equation 3.18.

$$Q = -k \frac{dT}{dx} \quad (3.18)$$

The symbols Q , k , dT and dx in eqn. 3.18 refer to heat flux exchange between the layers, conductivity of the layer, temperature difference between the two layers, and the depth of the surface layer respectively. The thermal and radiative exchanges between the layers are demonstrated in fig. 3.6. The heat responsive layer and conductive layer are $3cm$ and $1m$ respectively, where a large temperature diurnal cycle (over $20^\circ C$ of diurnal variation is observed by the soil layer model in [Arya, 2001]) at the depth of $2.5cm$ and only small diurnal fluctuations are observed at depths of $1m$.

The upwelling flux from the surface is then absorbed by the atmosphere as shown in fig. 3.6 by the factor of emissivity given by the Kirchoff's Law, where the emissivity is equal to the absorptivity. This absorption of the surface upwelling flux warms up the atmosphere and thus radiates in both directions itself. In thermal balance, the atmosphere radiates the same amount as it absorbs, leading to a direct

reach an equilibrium. The modelled outgoing flux was found to stabilise after a few days from the initial set up.

3.8. Chapter Summary

This chapter has presented information on the GERB instruments and the data products available which are relevant to our analysis. The data analysis methods used in our study have also been explained and discussed in this chapter.

The Fourier transform technique is valuable in identifying and analysing multiple timescale cycles within the time record of the data. On the other hand, the PCA and band pass filter techniques are more orientated in analysing specific timescale cycles. The PCA is especially powerful in investigating the spatial and temporal properties of various climate processes.

A simple heat balance model developed to aid in the interpretation of the results has also been introduced. The model consists of a single layer of atmosphere, a surface which radiates according to Stefan-Boltzmann law, a conductive layer beneath the surface and ISR from GERB SW data.

4. Analysing Different Modes of Variability using Fourier Analysis

4.1. Introduction

The Earth's radiation budget is forced by periodically varying solar input and exhibits an array of temporal and spatial variability. As a consequence, there are different modes of variabilities in the OLR cycle, the most fundamental ones being diurnal and annual as they are directly affected by incoming solar radiation. However, there are more subtle modes which are not as well-defined, for example, the semi-diurnal, 2-10 days cycle.

The OLR is the energy leaving the climate system and is affected by many climate processes. Hence, the variability within the OLR is an important instrument for understanding climate processes as the overall OLR consists of the superimposed OLR signals from individual climate events. Therefore, the climate processes can be better understood by separating the individual components of the OLR.

The GERB data are organised by local solar time (LST), calculated for each point as UT minus the time difference given by the longitude for the point. Fourier analysis is used in this chapter to attempt to separate the different components in the frequency domain. The chapter aims to study the variabilities of the OLR in short timescale (less than 10 days), using high time resolution data from the GERB instrument. The temporal resolution (17 minutes) is particularly crucial in studying diurnal/sub-diurnal variation in the OLR, allowing the maxima and minima in the flux to be determined to a high time accuracy, in contrast to previous diurnal broadband radiation studies.

The spatial variability associated with these different cycles will be analysed and compared with various parameters, for example , sky condition, and surface/ocean state. This should enable us to isolate the climate process for each frequency response, For example, a dominating diurnal cycle over land surface type. Different months of data will also be compared to highlight the seasonal variation in these cycles.

4.2. Chosen points/ latitude bands Fourier analysis

A simple Fourier analysis on a time series of OLR variation is powerful in identifying the important frequency modes for a particular location. Whereas an average over a region or a latitude is important in determining the fundamental modes of a more generic scenario.

This section begins with an analysis on a single time series as a test case study of the OLR variability. Three points were chosen to represent some of the variability within the OLR cycles in the tropics, one of which is situated within the ITCZ belt, whereas the other points demonstrate two different surface types outside the ITCZ region (vegetation and desert). The ITCZ is defined to be the zone where trade wind converges, and on the other hand, the rainbelt is a region close to the ITCZ driven by a large latitudinal temperature gradient which is believed to be responsible for majority of the precipitation in West Africa. In our initial point analysis, they will not be distinguished explicitly since they spatially overlap to large extent especially over central and Western Africa. The points of interest are chosen to sample a range of OLR variabilities over the land region: lots of convection/precipitation events, maximum solar radiation heating(minimum SZA) and vegetation surface type outside of the ITCZ.

Some discussion in the chapter will be focused around the surface type associated with the chosen region, therefore Fig. 4.1 presents a surface map of the African region in hope to provide visual aid for the comparisons and descriptions in the rest of the chapter.

The surface type used in the study are from GERB surface data, and is a static

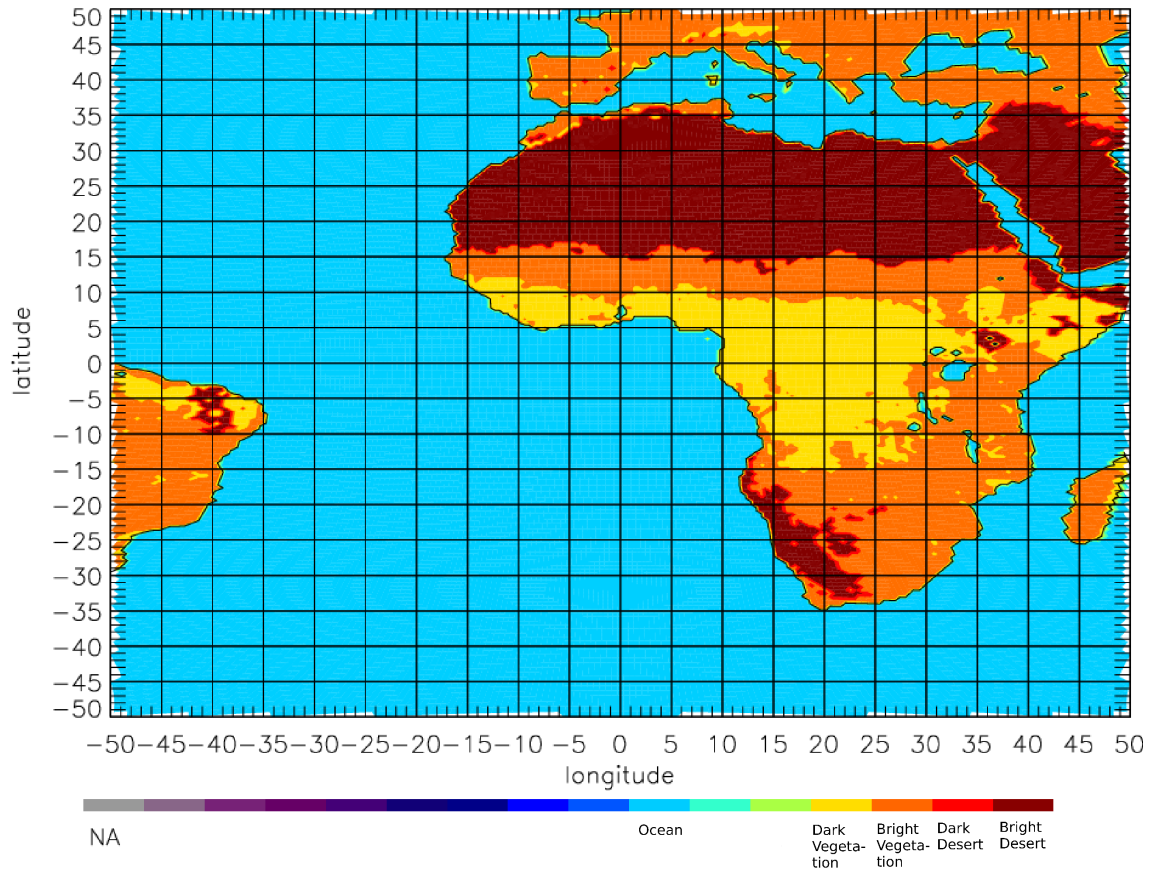


Figure 4.1.: surface type plot showing dark desert, bright desert, dark vegetation, bright vegetation and ocean, NA indicates the part of colour scale which is not applicable to the diagram

map(Fig. 4.1 of 5 surface types(Ocean, dark vegetation, bright vegetation, dark desert, bright desert). During the seasonal variation, this could potential wrongly categorized surface type data in transitional savannah, which is mostly along the Sahel in our chosen region. The Sahel region will be analysed separately in our results chapter to observe how the dominant signals manifest in the region over the season.

The time series have been transformed using FFT into frequency spectra. Fig. 4.2 shows three individual time series and frequency spectra, which demonstrate the typical OLR variabilities for three scenarios: desert, vegetation, and ITCZ. The desert spectrum shows strong diurnal and comparatively weaker semi-diurnal peaks, which indicates a strong daily OLR pattern in the desert region outside of

the ITCZ. Whereas, the other two spectra display much more noise and variabilities, the diurnal peak is still recognisable but the lower frequencies cannot be observed.

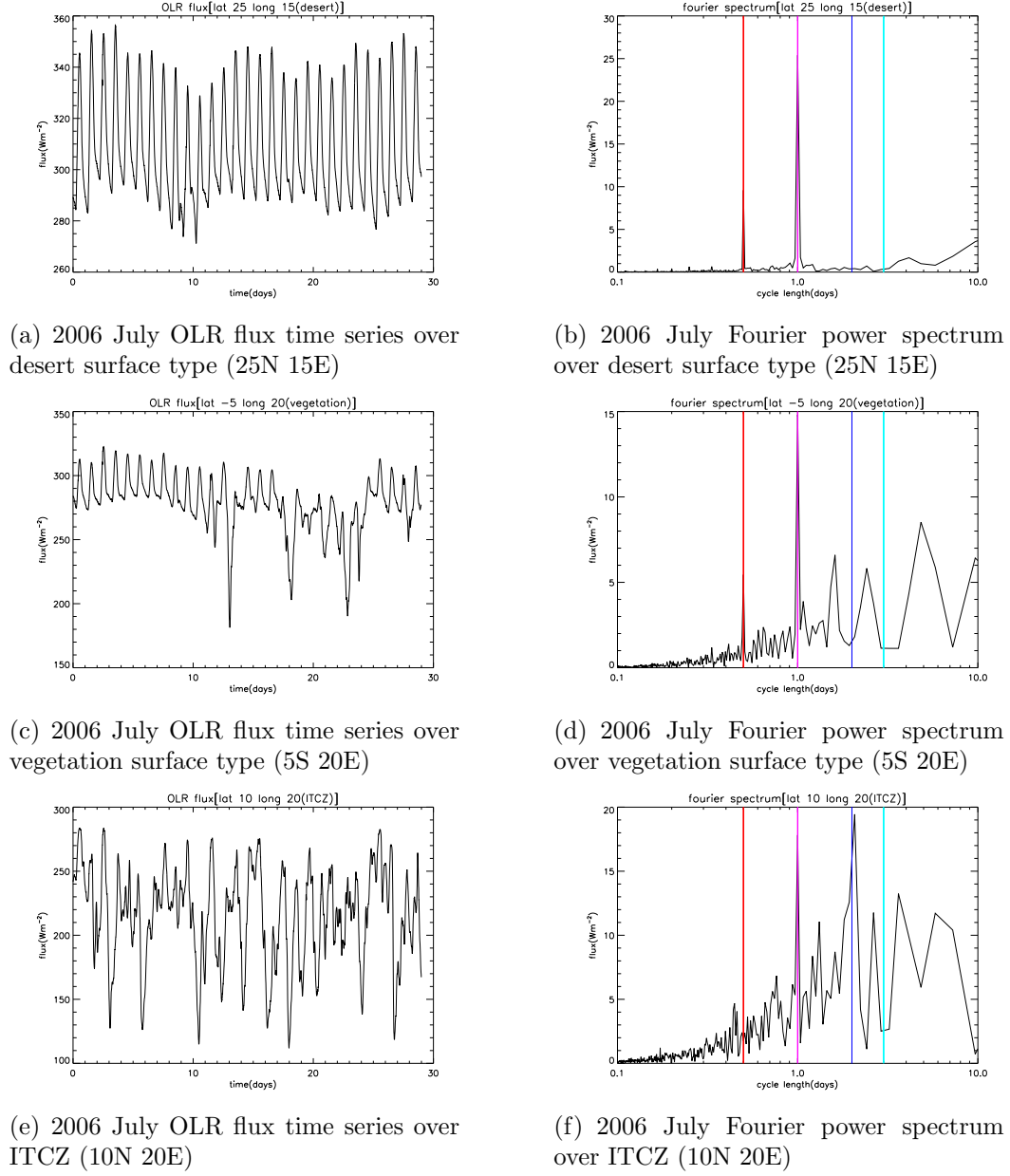


Figure 4.2.: 2006 July OLR cycles for a)desert c)vegetation e)ITCZ and their corresponding Fourier spectra b), d) and f) respectively. [latitude, longitude] a)[25,15] c)[-5, 20] e)[10, 20]. The vertical colours lines: red, purple, blue and cyan denote the timescale $\frac{1}{2}$, 1, 2 and 3 days respectively.

Whilst the semi diurnal peak in the ITCZ point plot can barely be recognised, the statistical significance of which is arguable compared with the desert and vegetation

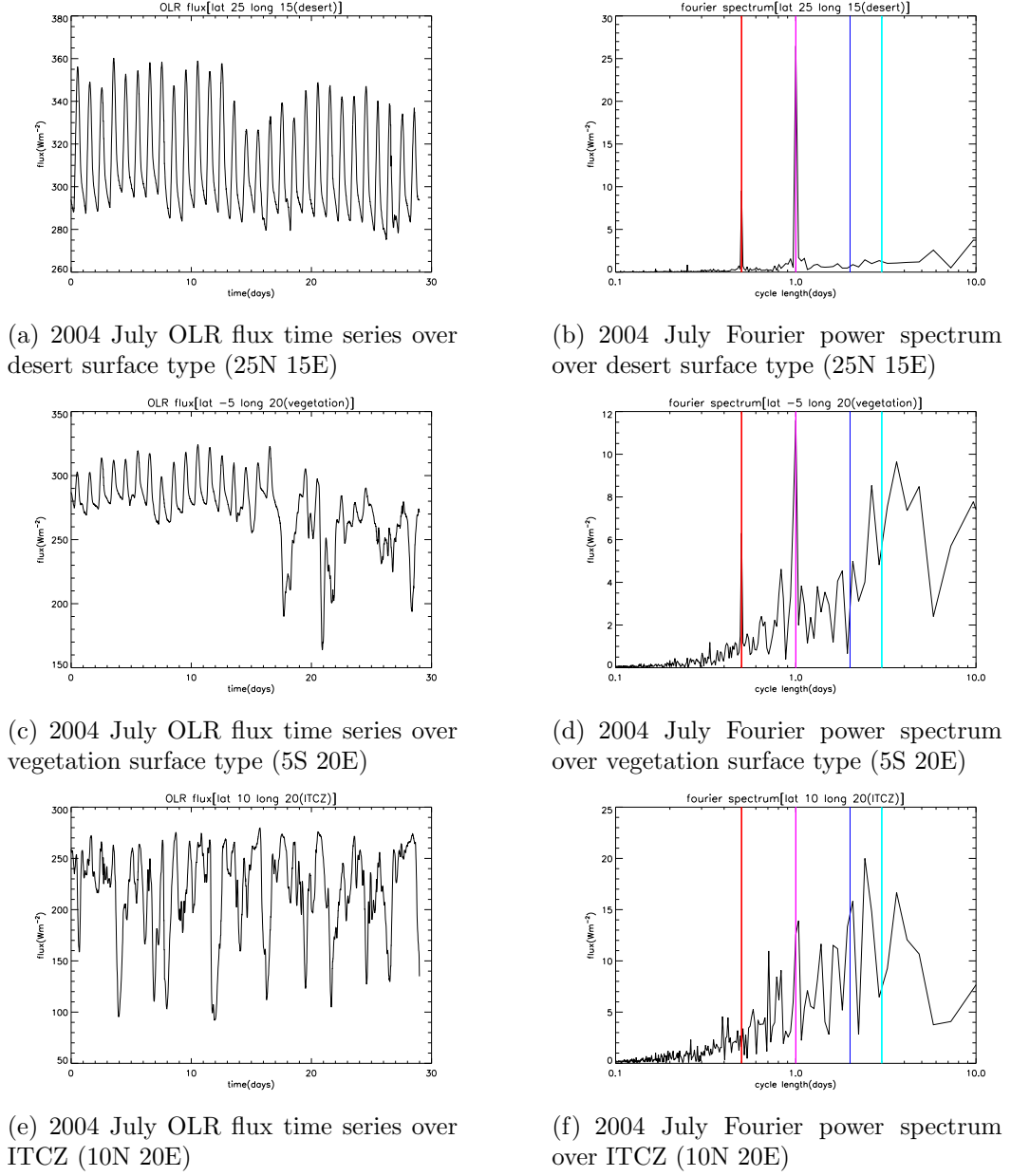


Figure 4.3.: 2004 July OLR cycles for a)desert c)vegetation e)ITCZ and their corresponding Fourier spectra b), d) and f) respectively. [latitude, longitude] a)[25,15] c)[-5, 20] e)[10, 20]. The vertical colours lines: red, purple, blue and cyan denote the timescale $\frac{1}{2}$, 1, 2 and 3 days respectively.

points observation.

In Fig. 4.4 has shown the OLR cycles and their corresponding spectra for the

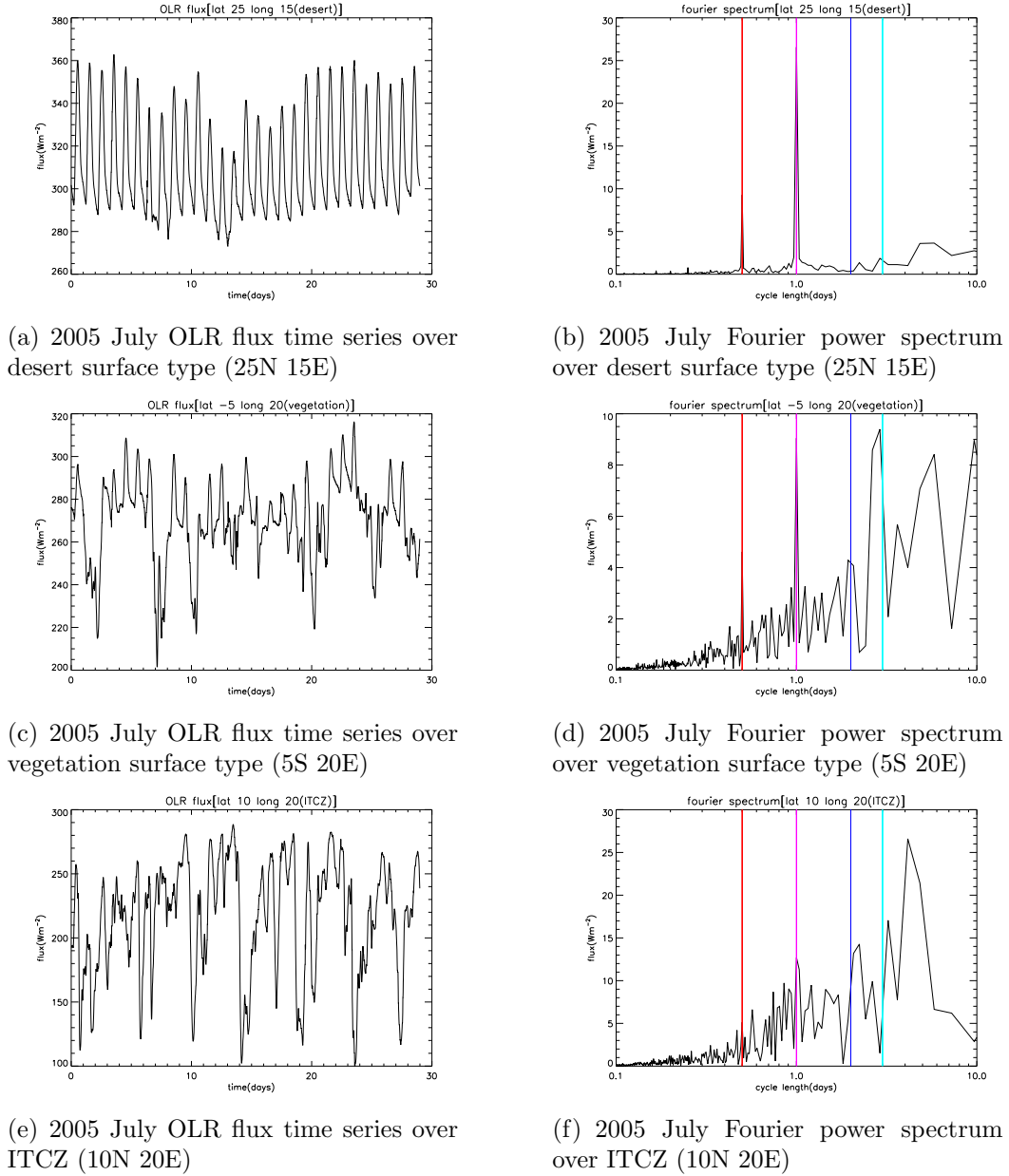


Figure 4.4.: 2005 July OLR cycles for a)desert c)vegetation e)ITCZ and their corresponding Fourier spectra b), d) and f) respectively. [latitude, longitude] a)[25,15] c)[-5, 20] e)[10, 20]. The vertical colours lines: red, purple, blue and cyan denote the timescale $\frac{1}{2}$, 1, 2 and 3 days respectively.

same points as Fig. 4.2. The desert point shows some fairly consistent diurnal variability across the month, and the spectra power of the diurnal and semi-diurnal cycles ($20Wm^{-2}$ and $10Wm^{-2}$ respectively) are very similar to the previous plot of 2006. This suggests the desert point is predominantly dry during this month and the only dominant cycle is the clear sky surface heat response.

If we compare the three years of data together in Fig. 4.2, 4.3, and 4.4, we can see the diurnal and semi-diurnal are certainly robust across the years over desert and vegetative surface types. Although the signal appears to be slightly suppressed along the ITCZ, due to the convective humidity and cloud cycles which absorb more downwelling SW flux and weakens the surface heating cycle; they are still the strongest and most consistent signals found in the tropical region. Therefore, they are determined to be significant and will be investigated further in this chapter.

Furthermore, over the vegetative surface types and the ITCZ region where there is significant moisture available for humidity and cloud convection, there appears to be higher modes of variability which are significant and dominant in those regions across the different years. Although the exact period of the variability appears to be inconsistent (from 2-7 days), their signal range from $10-20Wm^{-2}$ in Fourier power (doubles in peak to trough variation) which are significantly large and comparable with cloud signals. These could be manifestation of some of the African Wave Disturbances (see chapter.2) which are responsible for cloud convective activities and the majority of the precipitation over the Continent. These also will be investigated separately in this chapter.

4.2.1. ITCZ over the ocean

This section investigates how the ITCZ exhibits its variability over the ocean and how they compare to the land variability.

Fig. 4.5 plots the ITCZ region over the ocean surface type and their corresponding Fourier spectra for July 2004-6. The variability over the ITCZ ocean appears to be completely dominated by the lower frequency modes with less $10Wm^{-2}$ observable in the diurnal mode (c.f. $10-15Wm^{-2}$ over land ITCZ). There are two causes for this observation. First of all, the diurnal OLR cycle of the ocean heating is relatively small compared to land because of its huge heat capacity and the circulation

of the ocean. Second of all, this region is over the ITCZ, which similar to its land counterpart, would subject to convective cloud and moisture variability which means

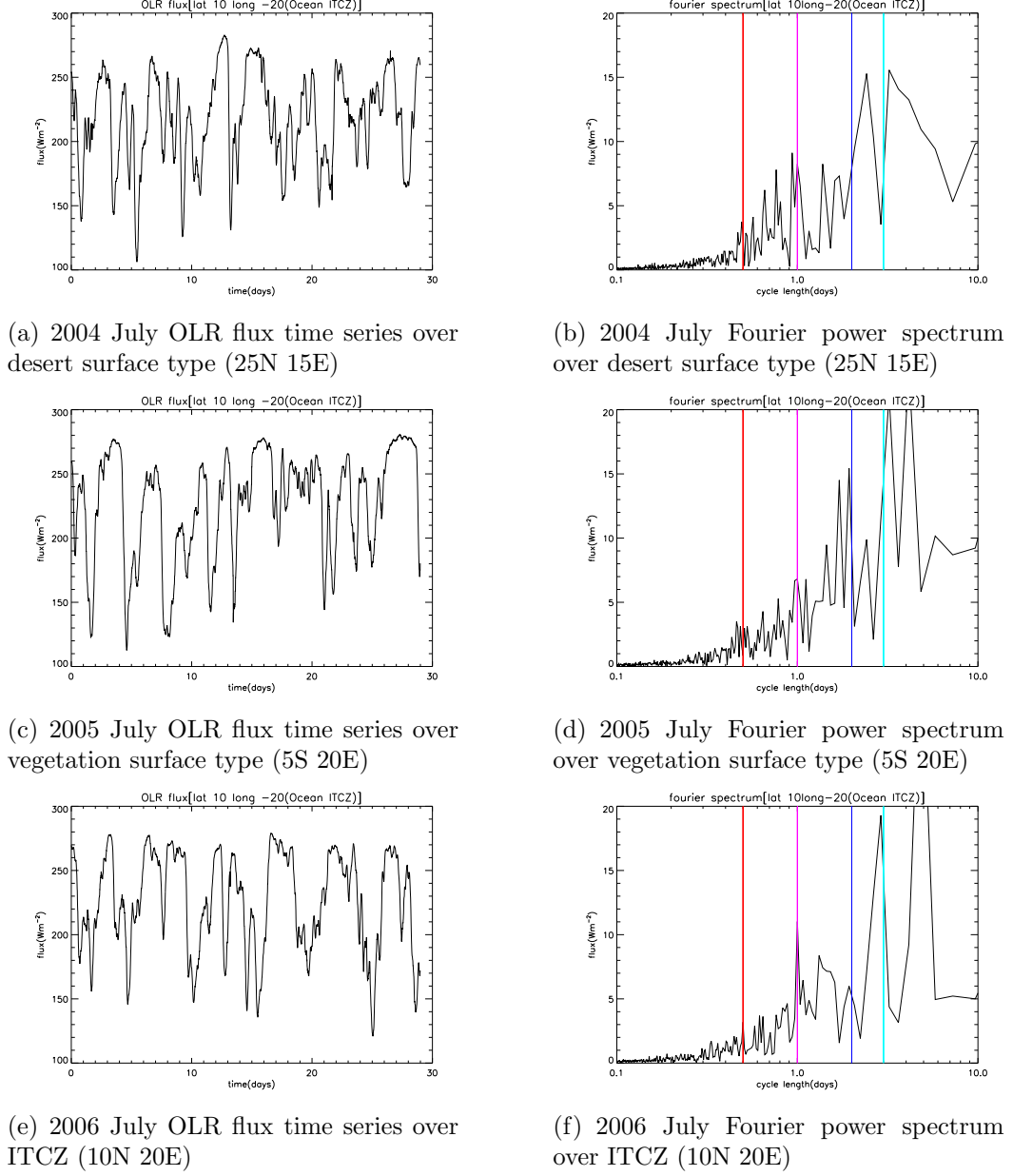


Figure 4.5.: 2004-6 July OLR cycles for ITCZ ocean and their corresponding Fourier spectra . [latitude, longitude] [10°N, 20°W]. The vertical colours lines: red, purple, blue and cyan denote the timescale $\frac{1}{2}$, 1, 2 and 3 days respectively.

much of the variability observed would be dependent on the processes associated the ITCZ and the local meteorology, namely the AEW, in this region. Therefore, we would expect the periods of these modes to be variable and generally focus towards the lower frequency end of the spectrum.

The lower frequency modes appear to be a dominant feature over the ocean ITCZ region and judging from the year to year variability, those seem to be robust signals across the years. This will be investigated further as part of the 3-7 days variability.

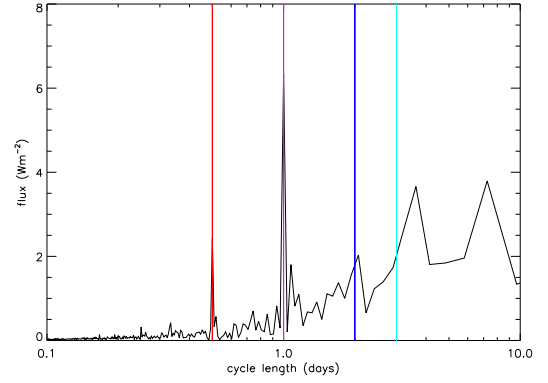
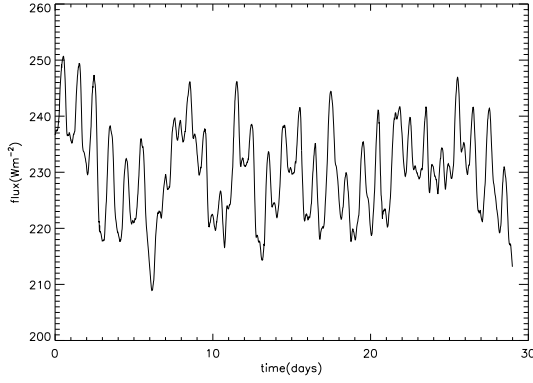
4.2.2. Latitude Band Analysis

Following on from the points analysis, we have decided to reduce some of the noise and highlight the more consistent signals by averaging the data in latitude groups. The data is averaged zonally into 5° latitude bands to attempt to analyse some of the larger scale events. The time series are Fourier transformed using the same method as before.

Fig. 4.6 shows two averaged OLR series of 30 days of data in July 2006 over two different latitudes, one of which is along the ITCZ region, whereas the other lies outside. These particular latitude bands were chosen to demonstrate the importance of different modes of variability over different regions as well as the composite of climate events associated with the variability, for example, one of which is over the ITCZ region with fair amount of convection, cloud developments and precipitation whereas the other is predominantly clear sky land and atmospheric response to solar input.

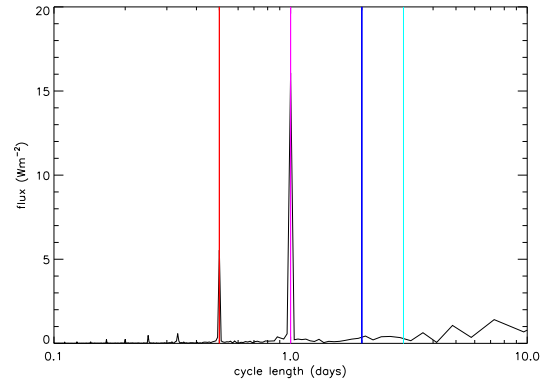
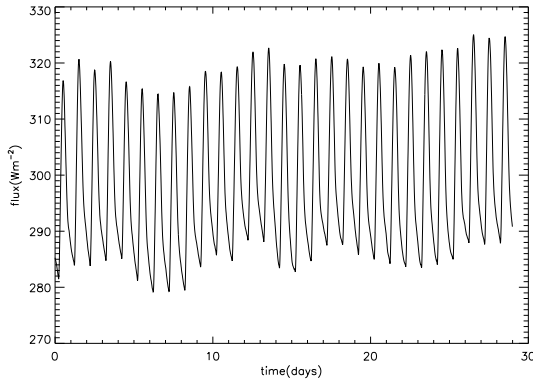
In Fig. 4.6(a) & 4.6(c), the OLR time series in the ITCZ region seem to demonstrate a much larger variability then the desert region due to convective activities and cloud formations. This cloud region can be easily defined by comparing the mean OLR flux across the African region, where average flux over the ITCZ is significantly lower than the other regions.

It can be observed from Fig. 4.6 that there exist diurnal timescale peaks in both of the spectra over the two latitude bands. However, the magnitude of the diurnal peak over the $25\text{-}30^\circ$ was observed to be more than twice the amplitude over $5\text{-}10^\circ$. This difference in magnitude could be explained by various climate conditions. Firstly, in July the SZA is a maximum in the northern hemisphere (over 20°N),



(a) 2006 July averaged OLR flux time series over latitude band 5N-10N

(b) 2006 July averaged OLR flux Fourier spectrum over latitude band 5N-10N

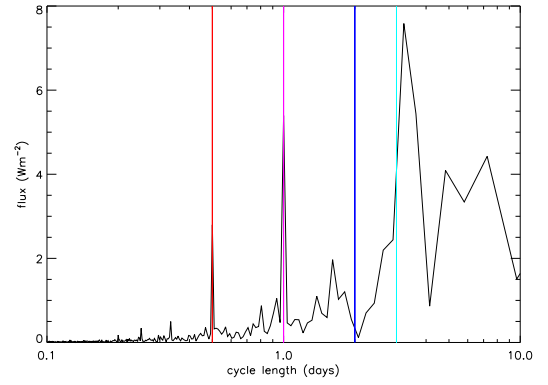
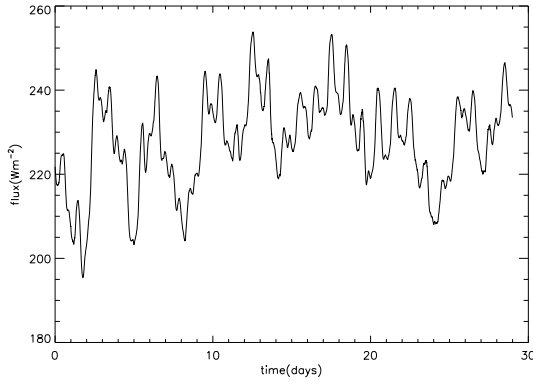


(c) 2006 July averaged OLR flux time series over latitude band 25N-30N

(d) 2006 July averaged OLR flux Fourier spectrum over latitude band 25N-30N

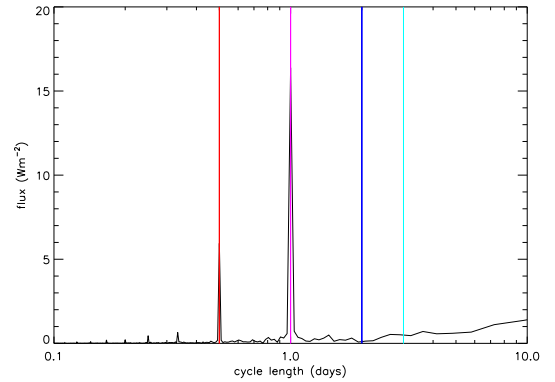
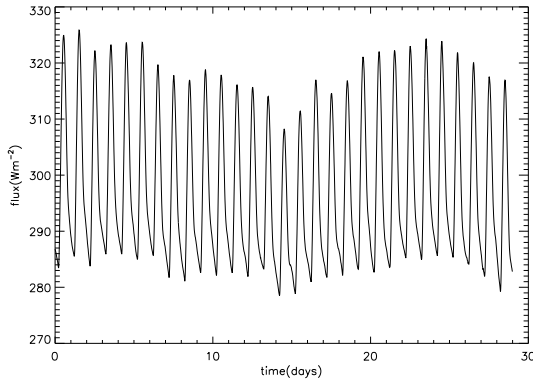
Figure 4.6.: 2006 July OLR cycles for latitude bands a) 5N-10N, c) 25N-30N and their corresponding Fourier spectra b), d) respectively. The vertical colours lines: red, purple, blue and cyan denote the timescale $\frac{1}{2}$, 1, 2 and 3 days respectively.

therefore, the 25-30° latitude band should have a slightly higher ISR than the other band. Secondly, the surface type in the 25-30° is predominantly desert, whereas 5-10° consists mainly of vegetation and heats up less during the day. Therefore, the overall diurnal OLR cycle over desert type is expected to be higher. Lastly, the two latitude bands are subjected to different atmospheric condition: the southern band is along the ITCZ region where there is a significant amount of convective events which inhibits the strength of the diurnal cycle. It can be observed that semi-diurnal components exist in both of the latitude bands, this will be investigated further in section 4.4.



(a) 2005 July averaged OLR flux time series over latitude band 5N-10N

(b) 2005 July averaged OLR flux Fourier spectrum over latitude band 5N-10N



(c) 2005 July averaged OLR flux time series over latitude band 25N-30N

(d) 2005 July averaged OLR flux Fourier spectrum over latitude band 25N-30N

Figure 4.7.: 2006 July OLR cycles for latitude bands a) 5N-10N, c) 25N-30N and their corresponding Fourier spectra b), d) respectively. The vertical colours lines: red, purple, blue and cyan denote the timescale $\frac{1}{2}$, 1, 2 and 3 days respectively.

Fig. 4.6(b) shows peaks in longer cycle frequencies which were only observed for certain latitude bands. This is believed to be caused by easterly waves triggered by the strong temperature gradient between the Sahara desert and the Gulf of Guinea, the origin of such waves is discussed in section 2.

Fig. 4.7 shows the latitude averaged flux and their spectra to compare the similarities and year to year variability with 2006. First of all, the 20°-30°N plot (Fig. 4.7 c, d) shows the same dominant diurnal and semi-diurnal variability as 2006, with diurnal component of up to $15Wm^{-2}$ in both years. This indicates this latitude is predominantly dry (i.e. desert) for this season.

Second of all, the 5°- 10°N latitude band plots (Fig. 4.7 a, b) show the same magnitude of diurnal and semi diurnal components as 2006, around $4\text{-}5Wm^{-2}$ for the diurnal component. However, comparing the longer timescale variability between the plots, we can see strong signals in 3-4 and 5-7 days, with 2005 showing a much strong peak near the 3 days variability. This could be due to the year to year variability which constitutes the wet and dry years in the continent, *Ali and Lebel* (2009) have measured such variability using standardised precipitation index (SPI). Nevertheless, in both years, there appear to be some strong longer timescale (longer than diurnal) signals present which will be investigated in section 4.5.

It is difficult to justify a significant mode of variability based on one Fourier spectrum alone since the noise could have a big influence on the variability presented in the spectra. However, by analysing the results across different years of the same months and by averaging data across latitude bands to reduce the noise and highlight the importance of the important modes. We can look at the strong signals which are persistent and determine these signals to be significant.

We have found in our points and latitude band analysis that the semi-diurnal and diurnal cycles to be a dominant and robust signals across the years, particular over land. Over the ITCZ, we have observed large variability in 2-7 days timescale, they are variable in period and strength from year to year. however, they remain one of the dominant signals across the ITCZ region in the years we observed. Therefore, these modes are determined to significant and will be investigated further in the rest of chapter.

We have used a month of data for all our points and latitude band analysis. The upper limit to the Fourier analysis technique is limited by $1/2T$. We would have to use longer sample if we need to investigate timescales beyond 10 days. The longer periods will be more susceptible to noise, however, the different months/years plots suggest we are observing a robust signals across the timescale. Therefore, the analysis up to 10 days will remain valid. Furthermore, we have found most the significant modes of variability to be up to 7 days, which is what our power spectrum map analysis will be based on. Therefore, our maximum time scale analysis (7 days) will be well below the maximum limit.

An extension of the timescale can be made for the analysis using multiple months time series, but it is beyond the scope of this study.

4.3. Diurnal variability

The diurnal cycle of OLR is one of the most fundamental variabilities associated with the solar incoming radiation. In section 4.2, the magnitude of the diurnal component was found to vary in different regions, depending on various factors: the intensity of the ISR (I.e. the SZA), the surface properties, and the atmospheric condition (whether clouds exist). Therefore, the analysis will begin by studying the spatial variation of the Fourier power which corresponds to the diurnal cycle length.

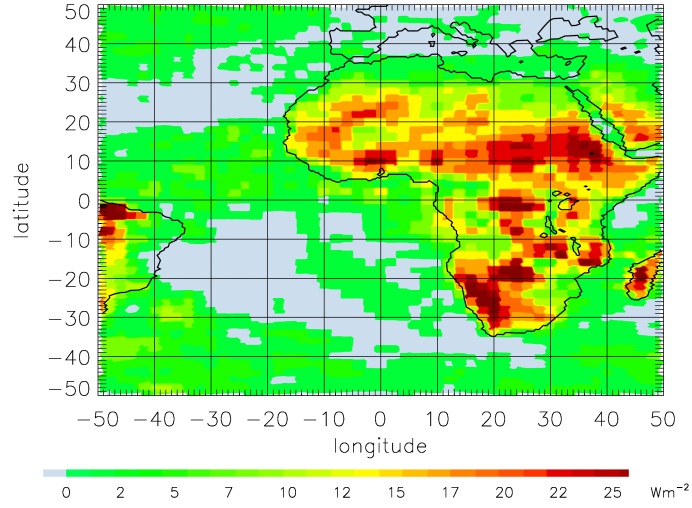
Fig. 4.8 represents the OLR variability in diurnal time scale in terms of Fourier component amplitude which corresponds to the frequency for January and July 2006. This is done by taking the Fourier component at 1 day time scale in Fig. 4.2 for each point on the $2 \times 2^\circ$ averaged grid.

It can be observed from Fig. 4.8 that strong signals are observed over land which indicates strong diurnal cycles are found over land. Whereas, the power over the ocean is much weaker in comparison, this is due to the high heat capacity of water and ocean circulation which diminished the magnitude of the diurnal surface response to the solar heating [*Smith and Rutan, 2003*].

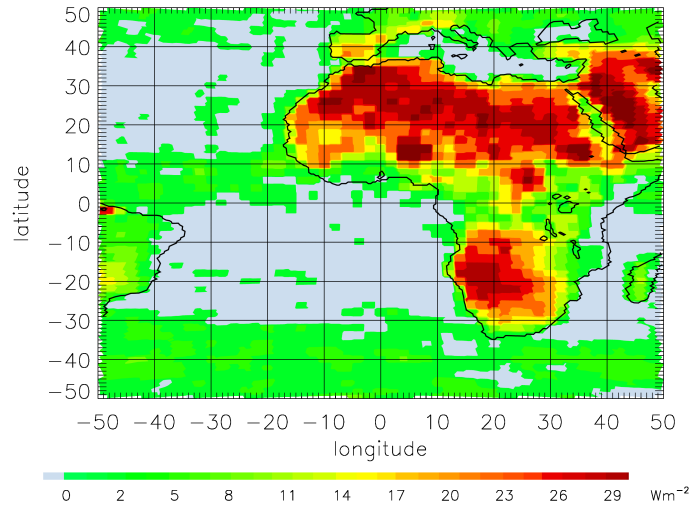
The ITCZ also has an impact on the strength of the diurnal cycle due to the convective events that it triggers as the band structure can be identified in Fig. 4.8 from Africa to Atlantic ocean. This affect is different over various surface types. For example, the presence of the ITCZ generally increases the strength of the diurnal cycle over the ocean, as shown in Fig. 4.8(a). Whereas, the ITCZ over the land region can weaken or strengthen the power of the diurnal cycle depending on the phase of the cloud cycle. This is largely due to the diurnal variability of cloud is much larger than the ocean but significantly smaller than land. Hence, the diurnal cycle of the ITCZ triggered convection and convective clouds appeared to have dominated the diurnal variability over the ocean.

This indicates that the cloud formation possesses a diurnal cycle with a power which is between land and ocean surface heat response. Dependent on the phase of the cloud cycle, it can either enhance or inhibit the overall diurnal cycle of the OLR. This can be confirmed by simply observing the diurnal OLR cycles (not demonstrated in the thesis) of two spatially near grid points with the same surface, where one is in a clear sky condition and the other in a "cloudy" region: the clear sky

point showed consistent diurnal cycles over the time period, whereas the cloudy



(a) 2006 January Fourier power (Wm^{-2}) map plot (1 day timescale)



(b) 2006 July Fourier power (Wm^{-2}) map plot (1 day timescale)

Figure 4.8.: 2006 a)January b)July Fourier power map plot for frequency equivalent to 1 day timescale

point sometimes showed a larger/smaller diurnal variability than the clear region. However, the strength of the cloud cycle is an order of magnitude higher than ocean diurnal cycle, therefore, a clear positive pattern can be observed over the ocean along the ITCZ.

Fig. 4.8(a) displays the Fourier power for January (boreal winter), where the minimum SZA is in the southern region of Africa. It can be observed that the strongest power is situated in the southern region. However, the signal is not spatially continuous in nature, due to the cloud formation introduced by the ITCZ (20°N - 5°S over ocean and 0° to 20°S over land), which weakens or strengthens the overall diurnal cycle depending on the phase of the cloud cycle. The power over the ocean region is almost zero in general, with the exception of the ITCZ band and in the mid-latitudes, where there are a significant amount of cloud activities to influence the averaged diurnal cycle of OLR.

As a comparison, a similar plot is shown in Fig. 4.8(b) for July 2006, where the ITCZ is in the central/northern region, at about 0° to 15° . The spatial distribution of the diurnal power shows SZA dependency to some degree, while the maximum power is now found in the northern region, along the Sahara desert and Ethiopian highland. The ITCZ is better defined by the diurnal power plot than in January by a band of mixed signals just below the Sahara desert due to the strong heat response exhibited by the Sahara contrasting the ITCZ region. This strong response is due to the intrinsic properties of the surface type (desert), where the moisture content in the surface is extremely low. Therefore, the heat capacity is low compare to other land surface types, causing a large day/night time temperature difference, hence the deserts express a large diurnal component in the OLR.

From the diurnal analysis, it can be observed that the diurnal component of the OLR is largely dependent on three factors, the heat capacity of the surface type, the SZA and the presence of clouds. The first two factors have a consistent impact on the magnitude of the diurnal component, whereas the effect of cloud cycle on the magnitude of the diurnal cycle of the OLR depends on an extra component, the time phase of the cloud cycle. It can enhance the magnitude of the overall diurnal cycle if clouds form after sunrise and before sunset, and vice versa.

4.4. Semi-diurnal variability

In section 4.2, the semi-diurnal component was identified to be another significant factor in the overall OLR time series. There existed a semi diurnal cycle in the power spectrum for most latitudes. Whilst the power magnitude of this frequency is significantly less than the diurnal cycle (approximately with a ratio of 1:3), it is present as one of the recognisable "peaks" in our spectrum plot along with diurnal cycle and 2-9 days variabilities. Therefore, a map plot of this power will be produced to analyse its spatial variation. This will enable us to observe the spatial variation of this particular mode and identify the climate process associated with this mode of variability.

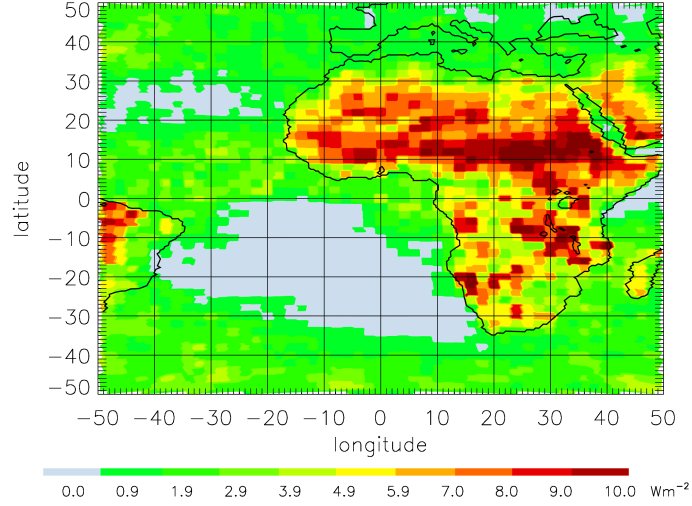
Fig. 4.9(a) shows the semi-diurnal components for January, where the strong signals are generally found over land. Some of the strong signals in the plot also overlap with the diurnal plot in certain regions, for example, from 10° to 30° , the spatial pattern between the two are quite similar. This region is out of the ITCZ during January and consists mostly of clear sky, therefore indicating that the clear sky OLR has both diurnal and semi-diurnal components. This suggests that the semi-diurnal component has surface type and SZA dependencies.

The semi-diurnal component was observed to be weaker over the ocean and the ITCZ region. These regions are associated and dominated by ocean circulation and convective cloud activities respectively. This suggests that the semi-diurnal component is determined mainly by ISR, and surface properties. Due to the close resemblance of this cycle with the diurnal cycle and the fact that the clear sky surface heat response diurnal has an asymmetric about its peak which cannot be fully accounted using only one mode of variability in the Fourier analysis. We believe this semi-diurnal variability is the component which gives the diurnal cycle its characteristic asymmetry and also represents the peak shift of the OLR cycle away from noon.

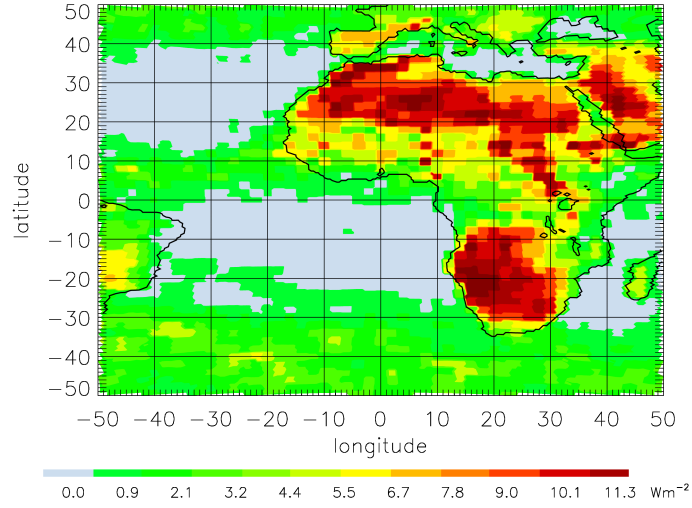
To further investigate this timescale of variability, a model was created to attempt to replicate the diurnal and semi-diurnal components using simple radiation properties of earth's surface, such as incoming radiation absorption, blackbody radiation and the heating properties of the surface.

4.4.1. Surface radiation model

As both the semi-diurnal and diurnal component of the OLR spectrum exhibit dependencies on the SZA and surface type, it was decided to create a simple surface



(a) 2006 January Fourier power (Wm^{-2}) map plot ($\frac{1}{2}$ day timescale)



(b) 2006 July Fourier power (Wm^{-2}) map plot ($\frac{1}{2}$ day timescale)

Figure 4.9.: 2006 a)January b)July Fourier power map plot for frequency equivalent to $\frac{1}{2}$ day timescale

radiation model to analyse this behaviour. The model consisted of the following parameters: specific heat, effective surface depth and the ISR, and an arbitrary surface albedo was used for this model. This simple model made the following assumptions: the entire spectrum of the ISR is absorbed by the surface after taking into account the albedo factor; the surface radiates as a blackbody. The model has been run at 15 minutes time step for the ease of comparison with GERB data.

The above assumption and the lack of atmosphere in the model, means the absolute magnitude of both the OLR and Fourier power will deviate significantly from the observed data. However, the aim of the model is to understand the relationship between semi-diurnal and diurnal components. Furthermore, the model is expected to demonstrate if there exist correlations between these components and the model parameters. Therefore, the absolute magnitude of the individual components in our model outcome is irrelevant and the two components should only be compared either with one another or with the same component for another location (i.e. with different ISR and/or heat capacity).

The model absorbs the ISR (obtained from GERB SW) according to the thermal energy equation and radiates as a blackbody according to the Stefan-Boltzmann law, please refer to appendix. A.3 for detailed description of the model. (Note. this model has been adapted and improved to investigate diurnal variability in chapter 6)

Fig. 4.10 demonstrates two sample days of modelled and observed OLR time series, where the modelled OLR has already stabilised. The two curves differ significantly in magnitude, which was expected due to the absence of atmosphere in our model. The diurnal difference, i.e. the day night OLR contrast is expected to decrease in the presence of atmosphere due to atmospheric absorption and scattering of the ISR leading to a decrease in solar radiation reaching the surface, and the absorption and re-emission of the OLR by the atmosphere. The diurnal change in the observed OLR in the desert region is around $70Wm^{-2}$, whereas the day night difference of the modelled radiation can exceed $700Wm^{-2}$.

Although the model is incapable of producing realistic OLR magnitudes, this is not crucial for our study as we are interested in comparing the phases of the cycles under clear sky conditions. The modelled OLR demonstrates a similar diurnal pattern to the observed data, which indicates the clear sky diurnal cycle can be

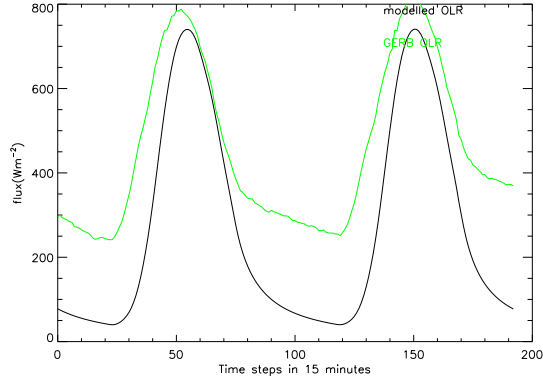


Figure 4.10.: 2006 July desert Modelled OLR & GERB observed OLR flux (Wm^{-2}) in time steps of 15 minutes, demonstrating 2 diurnal cycles over the time period (192 timesteps in total). GERB observed OLR flux has been subtracted by $250Wm^{-2}$ and then magnified by a factor of 8 for easy comparison with the modelled flux.

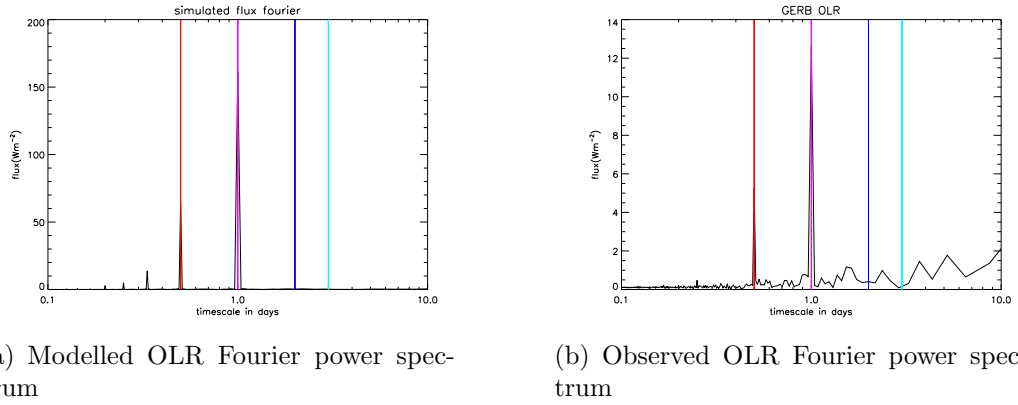


Figure 4.11.: 2006 July Fourier power spectrum (Wm^{-2}) for a)modelled and b)GERB observed fluxes in timescale of *days*. The vertical colours lines: red, purple, blue and cyan denote the timescale $\frac{1}{2}$, 1, 2 and 3 days respectively

approximated using the surface radiation model.

We have found that both the magnitude and the phase of the diurnal cycle varied with our input model parameters. In particular, the surface heat capacity has a large impact on the phase, we believe this phase affects the relative magnitude between the diurnal and semi-diurnal cycle. Therefore, the time series were transformed into frequency spectra using Fourier analysis, where the different frequencies can be

observed.

Fig. 4.11 shows the Fourier transform of the modelled OLR time series, both diurnal and semi-diurnal components were observed in the spectrum. It should be noted that due to the unrealistic magnitude of the modelled OLR, the absolute magnitude of the corresponding Fourier spectrum should not be compared with the spectrum of observed data. However, the relative magnitude between different $\frac{1}{2}$ and 1 day equivalent frequencies can still provide a valid comparison as the semi-diurnal component is believed to be a product of ISR coupled with surface properties.

The Fourier spectrum for observed data is shown in Fig. 4.11(b), by comparison with the modelled spectrum, it can be seen that the absolute magnitudes of the peaks differ by a large amount as expected. However, the relative magnitude between the diurnal to semi-diurnal of the two spectra are very similar, the power of semi-diurnal peak (indicated by the red tick marks in Fig. 4.11) is about a third of the diurnal peak (purple tick marks). This implies the strength of the semi-diurnal is dependent on: surface response to ISR and surface heat capacity, as the simple radiative model is sufficient to provide the same semi-diurnal to diurnal ratio as the clear sky observed data. Thus, indicating the semi-diurnal component is the manifestation of the diurnal cycle in the Fourier spectrum, where the diurnal timescale sine wave alone is not sufficient to describe the asymmetry of the diurnal cycle.

4.5. Two to three days variabilities

4.5.1. Introduction

This section will focus on cycle lengths longer than a day, where these modes of variability were mainly observed over the cloudy region and the mid-latitude, indicating that they are perhaps a cloud effect. These modes have been referenced as African Easterly Wave (AEW) or African Easterly Jets (AEJ) and to possess a period between 2-9 days (*Nicholson, 2009, Reed et al., 1977, Diedhiou et al., 1998, Grist, 2002, Gu et al., 2004*)

This wave like behaviour has been studied extensively in terms of change in wind speed and anomalies in precipitation. However, this subject has been rarely studied in terms of broadband OLR, which provides us an opportunity to investigate the

particular climate process in terms of outgoing energy. The broadband GERB instrument has enabled us to study the effect of this wave on the OLR by providing high time resolution measurements. The results of this study are expected to be consistent with previous studies (see chapter 2), however, we also expect certain features to be revealed under the OLR analysis: for example, the lag between the cycle of OLR and wind field, as well as the AEW cycles which do not induce precipitation, namely the moisture and cloud cycles associated with this variability. The measurements also allow for the observations of the OLR magnitudes and spatial distributions associated with these cycles, which is crucial in determining the significance of this variability, and the topological/surface variation on this variability respectively.

4.5.2. Two to three days variability analysis

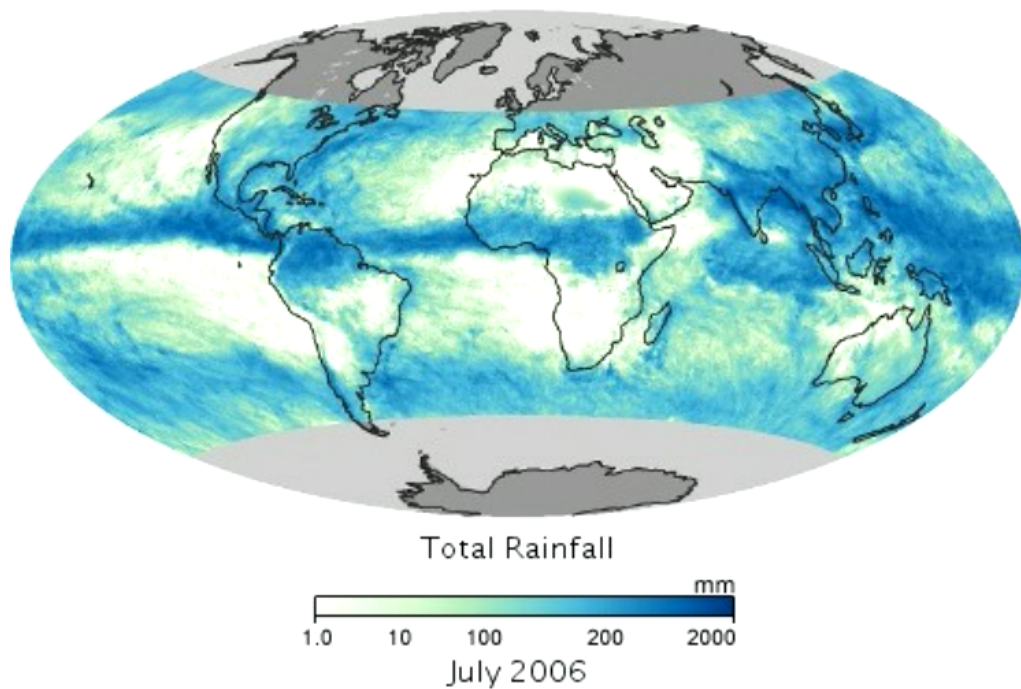


Figure 4.12.: Total rainfall maps over months July 2006 [NASA, Earth Observatory website] (TRMM rainfall data) grey indicate no data present.

From the latitude band analysis (section 4.2), it can be observed that there exist

peaks which have period over 1 day. Therefore, the analysis began by investigating the spatial distribution of the Fourier spectrum magnitude between 2 to 3 days. This was done by finding a maximum peak between a set cycle range (2-3 days) for each GERB location, the magnitudes of these peaks are then plotted onto a spatial plot, similar to the plots shown in section 4.3.

Fig. 4.12 plots the total rainfall measurement from the the TRMM observations, we have chosen to use the TRMM instead of ERA interim due to a precipitation estimation error in their forecast model which is documented in [Dee *et al.*, 2011]. The figure has shown a well defined rainfall band over the Sahel and the Atlantic, which characterises the position of the ITCZ over the region.

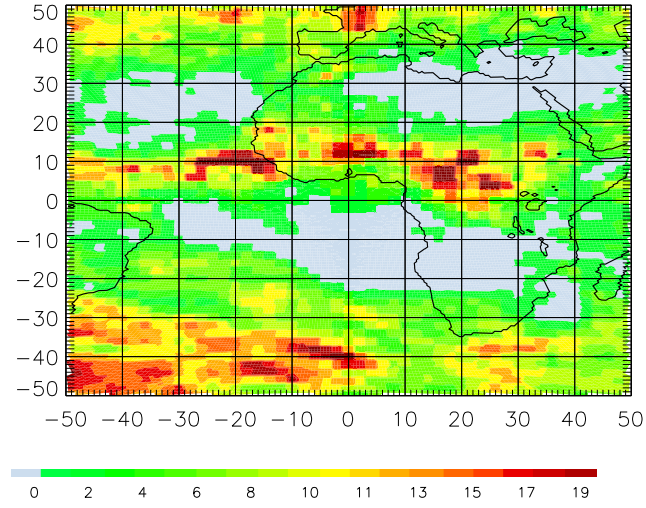
Fig. 4.13(a) shows the spatial distribution of the 2-3 days peak magnitude. The strongest signals are found in the central Africa and just off the coast of Northwest Africa. The strong signal band in the centre appears to overlap largely with the rainfall plot in Fig. 4.12, suggesting this could caused by the convective events along the ITCZ. The magnitude of the strongest signals go up to roughly $20Wm^{-2}$ which gives a peak to trough variability of around $40Wm^{-2}$.

The convective event typically possessed a diurnal cycle which maximises in cloud optical depth in the late afternoon and weakens in the morning. However, *Clayson et al.* (2002) have related the convective activity to a process named diurnal dancing by *Young et al.* (1995) where they found the atmospheric boundary layer has a recovery time of 48 hours after a convective activity. This could explain the manifestation of the convection in the 2-3 days plot along the ITCZ, as this 2-3 days variability is unlikely to be cause by the surface heat cycles which are strongly driven by the diurnal cycle of ISR.

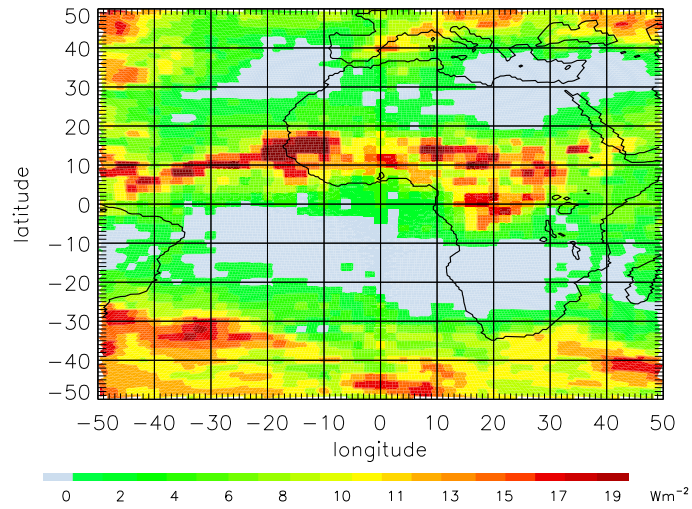
In Fig. 4.13(b), we have shown the same mode of variability in a different year (2005). The magnitude (up to $20Wm^{-2}$) and spatial distribution of the power are found to be consistent across the two years (and 2004, but not shown in Fig. 4.13), suggesting the signals are robust and above noise level and these convective activities are manifested in the ITCZ region in this month every year.

The power over most of the tropical region is almost 0 except for the area along the ITCZ, which is as expected due to the lack of convective activities outside of this region. However, outside of the tropics, there exists a region of signal extending from the subtropics to the mid-latitudes. This is believed to relate to Rossby waves which

cause low pressure system to form beneath the wave and thus induce cloud formation [Barry and Chorley, 2003], the dominant cycle length of such waves should be around



(a) 2006 July Fourier power (Wm^{-2}) map plot (2-3 days timescale)



(b) 2005 July Fourier power (Wm^{-2}) map plot (2-3 days timescale)

Figure 4.13.: a)2006 b)2005 July Fourier power map plot for frequency equivalent to 2-3 day timescale

5 days. Therefore, the signal in the mid-latitude is expected to be observable in both 3-4 and 4-7 days plots.

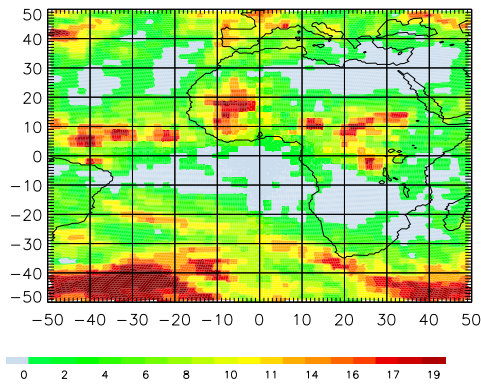
4.6. Longer cycle length variabilities(up to 7 days)

This section concerns some of the longer timescale OLR variabilities, from 3 to 10 days, as discussed previously, wave activities centred around central Africa are present over a range of frequency. Therefore, in order to study this variability, it is essential to analyse the spatial and magnitude distribution in slightly lower frequency ranges. One aim is also to compare the relative Fourier power between AEW and Rossby wave signals found in the tropics and mid-latitudes respectively, at different frequencies, by observing the spatial power plots. This will aid in identifying the dominant frequencies between the two climate processes and their respective length and time scales.

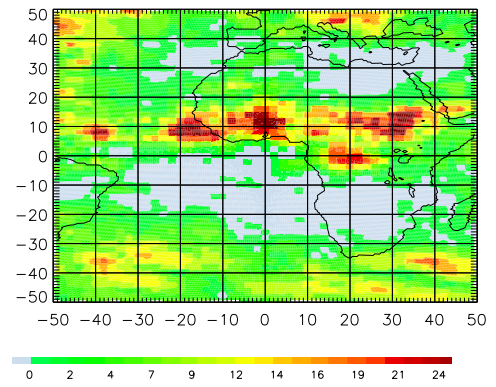
Fig. 4.14 illustrates the Fourier power plot for longer timescale, 3-4 days and 4-7 days respectively. It can be observed that the two plots share certain features with 2-3 day power plot, which are: the band of signal along the easterly jet; low or zero power in the tropics; and relatively high power found in subtropics/mid-latitudes. The easterly jet stream is defined as the band along central Africa and Atlantic ocean, although it appears to overlap largely with the ITCZ, but it's believed that the longer time scale variability is likely to be caused by the jet (described in chapter 2. However, in Fig. 4.14(a), the strongest signal was observed to be in the mid-latitudes, from 40°S to 50°S and we have observed the same magnitude of power along this region in 2005 (just under $20Wm^{-2}$). As mentioned earlier, this signal is thought to be due to Rossby waves, which have a period of around 5 days. The fact that 3-4 days and 4-7 days Fourier power were found both along the easterly jet and in the mid-latitudes, with partial coherence between the different timescale in spatial patterns. It suggests that the easterly waves and Rossby waves both have irregular wave periods. However, in 3-4 and 4-7 days timescale, the strongest signal(up to 30) is expressed in the mid-latitudes, indicating that the magnitude of the mid-latitude disturbances (Rossby waves) in 3-4 and 4-7 days frequency is higher than for the 2-3

days. Therefore, this partially supports our hypothesis that the mid-latitude waves are more dominant in lower frequencies(i.e. around 5 days).

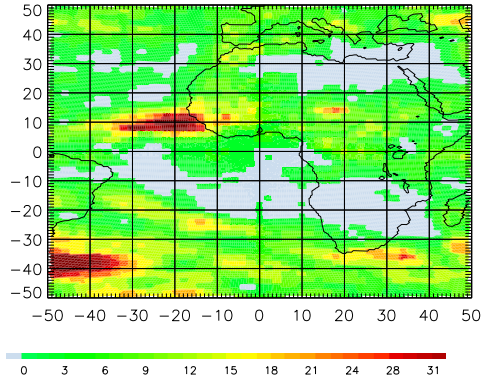
Fig. 4.14(c) shows Fourier power map of 4-7 days. In general, from the signal distribution, it is believed that the same mechanisms are observed in both 2-3 days and 3-4 days plots, along the ITCZ and in the mid-latitudes. There are two dominant regions in the plots: off the coast of Guinea/Liberia and in the mid-latitudes near the Brazilian coast. These are thought to be due to the range of frequencies which the Rossby waves and the African easterly waves have, around their dominant cycle



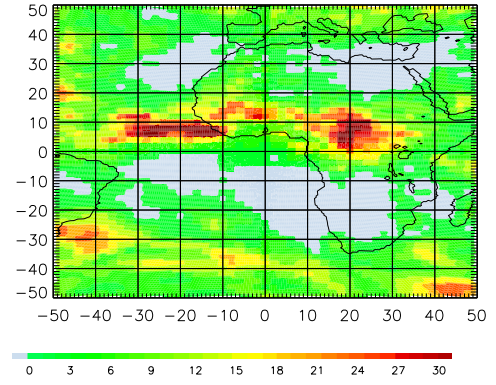
(a) 2006 July Fourier power (Wm^{-2}) map plot (3-4 days timescale)



(b) 2005 July Fourier power (Wm^{-2}) map plot (3-4 days timescale)



(c) 2006 July Fourier power (Wm^{-2}) map plot (4-7 days timescale)



(d) 2005 July Fourier power (Wm^{-2}) map plot (4-7 days timescale)

Figure 4.14.: 2005-6 July Fourier power map plot for frequency equivalent to a,b)3-4; c,d)4-7 days timescale

lengths [Holton, 1992].

Whilst the dominant frequency of the Rossby wave falls within this frequency range (4-7 days), it's uncertain whether the dominant frequency of the African wave disturbances also resides in this range. From the spatial power distribution plot (Fig. 4.14(c)), this seems to be a possible case. However, it requires further analysis to ensure the power we have detected near the equator is indeed caused by African wave disturbances. However, this is beyond the scope of this thesis.

In Fig. 4.14(b) and 4.14(d), the power maps have shown strong signals over the ITCZ region similar to the plots for 2006. However, in the 3-4 days timescale plot, the magnitude of power along the ITCZ region is higher than that of 2006, which is consistent with the high 3 to 4 days signal in our latitude band analysis, and could be attributed to year to year variability (see section. 4.2).

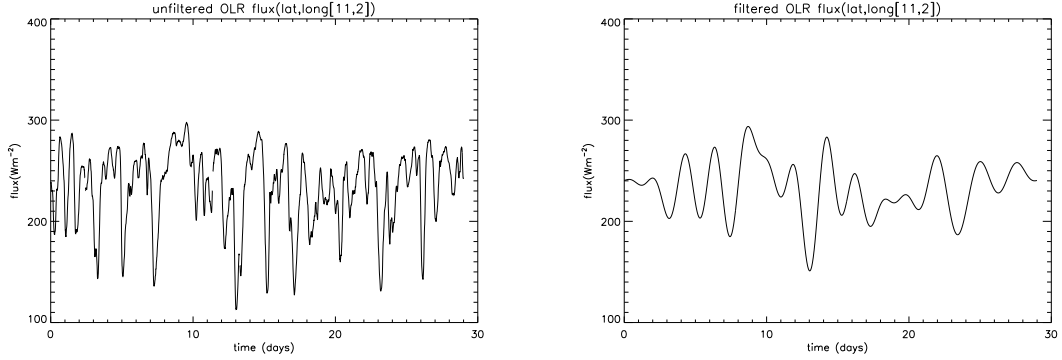
Along the mid-latitude (Fig. 4.14(d)), there exists signals in the mid-latitude (roughly $25Wm^{-2}$) which is slight weaker than in 2006 ($30Wm^{-2}$), but in both years, the 4-7 days signals are both higher than 3-4 days. This strengthens our hypothesis that there is Rossby wave presence in this latitude with a period of around 5 days.

4.6.1. Two days low-pass filter

It was decided to further analyse this longer time scale behaviour by using a bandpass filter to highlight the lower frequencies and display them as an OLR time series. The filter is designed to inhibit all the frequencies which are equivalent of 2 days cycle length or less. The results are expected to show wave like cycles in certain regions (i.e. downstream of the easterly jet and in the mid-latitudes) and changes in amplitude along the dry tropical regions are expected to be small.

Initially, we have selected Niger ($15^{\circ}N$, $5^{\circ}E$) for observation, the time series representing a month of data for the particular region averaged in $2 \times 2^{\circ}$ boxes. This region is along the AEW band during July 2006, and therefore the results can be used to test our hypothesis that the AEW should be observable in this location in July, thus, the filtered results are expected to demonstrate low frequency signal (less than 1 day) in Niger, whereas the signal is expected to diminish outside of this latitude range. As a comparison, we have chosen another region in the Sahara

desert (25°N , 20°E), which is in the dry tropical region, to demonstrate the lack of 2-10 days variabilities in the absence of the AEW. The time period chosen for this study is July 2006.



(a) 2006 July unfiltered flux(Wm^{-2}) time series (*days*) (11N 2E)

(b) 2006 July filtered flux(Wm^{-2}) time series (*days*) (11N 2E)

Figure 4.15.: 2006 July a)unfiltered b)filtered flux(Wm^{-2}) time series (*days*) over Niger (11N 2E). A low pass filter has been applied to b)

The OLR variabilities of the Niger region (see Fig. 4.15(a)) appeared to contain mixed modes in different frequencies. In particular, the day to day variability is quite large along the time series, indicating the dominant frequencies is beyond the diurnal timescale. Therefore, the time series is filtered and shown in Fig. 4.15(b); the filtered series is observed to contain a fair amount of variability, particularly in 2-3 days timescales. However, both the frequency and the magnitude of the time series were found to vary over the time period.

The change in the magnitude of the filtered OLR time series ranged from a few Wm^{-2} up to 100Wm^{-2} . The variation in magnitude implies the mechanism behind the cycle does not consistently repeat itself over the time period, unlike the diurnal cycle which is predominantly caused by the incoming solar radiation and has roughly the same diurnal magnitude change over a short period. The variation in the period of the cycle signifies the finite range of timescales of the responsible processes, where these processes do not have a well defined period.

The big magnitude differences suggest that this process is likely to be associated with cloud formation, where the clouds have a cold cloud top which inhibits the outgoing radiation measured at the top of atmosphere. For example, the TOA OLR

measures a cold cloud top event for a period of a day and subsequently a clear day, this is represented by a trough then a peak respectively in 2 days time scale. The period of the process ranged from 2 to 4 days, due to this variation in time length, it would be useful to carry out another analysis using another variable (e.g. wind speed/direction) to investigate the mechanism behind this cycle and the properties of this mechanism.

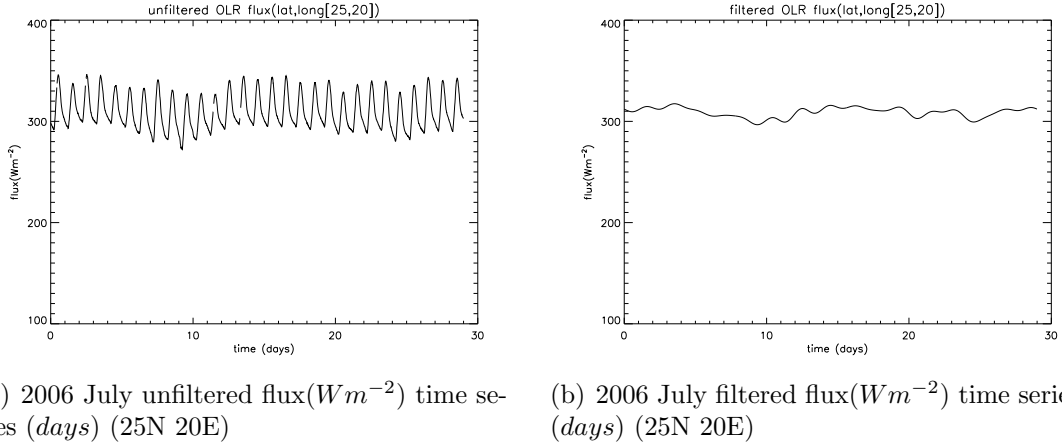


Figure 4.16.: 2006 July a)unfiltered b)filtered flux(Wm^{-2}) time series (*days*) over Sahara desert (25N 20E). A low pass filter has been applied to b)

As a comparison, we have also studied another point which is situated in the dry tropical region (25°N, 20°E) to observe how these longer cycles behave under different conditions. Fig. 4.16 displays the unfiltered and filtered fluxes over the region for July 2006, which are quite different from the results of the Niger region. The unfiltered flux plot (Fig. 4.16(a)) was observed to possess a fairly regular diurnal cycle with little variation in day to day variability. This is because the region chosen is in the dry region which indicates the main OLR component would be the surface response to the ISR.

Moreover, the filtered flux (Fig. 4.16(b)) was found to have relatively little variation over the time period, the maximum peak to peak magnitude was around $10Wm^{-2}$. The results signify the lack of longer cycles found over the dry tropical region which is as expected since the strongest longer cycle signals were found in the cloudy regions (see Fig. 4.13(a)) and therefore, only diurnal and sub-diurnal cycles would be observable over the chosen region, hence the small amplitude in filtered

flux. Furthermore, it supports the longer timescale processes being associated with cloud activities since the dry regions were found to possess no such variation.

4.7. Clear/cloudy sky analysis

These results led us to further investigate how the different cycles change with the presence or absence of clouds. Therefore, it was decided to study these effects by filtering the data for the analysis, the data went into the analysis was filtered based on the GERB SW scene ID, where if the cloud cover is found to be more than 20 than the GERB footprint is determined as cloudy. The cloud cover is the relative number of cloudy imager pixels in the GERB footprint or grid cell. A cloud cover of 0 corresponds to a clear sky footprint. A cloud cover of 100 corresponds to an overcast footprint. [*Gonzalez and Ipe,*]

The lack of night cloud data prompts the assumption of the footprint condition based on day time cloud mask where if a point is determined to be cloudy at any time during the day, then the point for the whole day is excluded from the clear sky study. Our constraint for data selection is very strict, therefore, we tend only to pick points which are predominantly clear during the month. There is a possibility that there is cloud contamination, but the effect is likely to be very small due to the transient nature of the cloud event as it's unlikely to find a point which is persistently cloudy during night time and clear during day time.

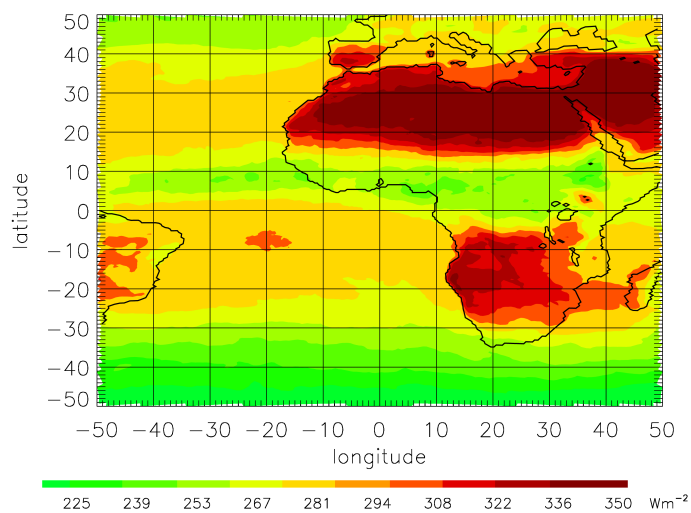
In the event where a point is cloud contaminated, it could introduce a bias where some cloudy nights are included in the clear sky analysis and vice versa, which could potentially increase day/night differences due to the cloudy night time flux measurement will be influenced by the cold cloud top temperature.

This section will analyse the Fourier power spectrum under different scenarios, where the clear sky and cloudy sky data will be filtered to highlight the spatial variation of certain cycles.

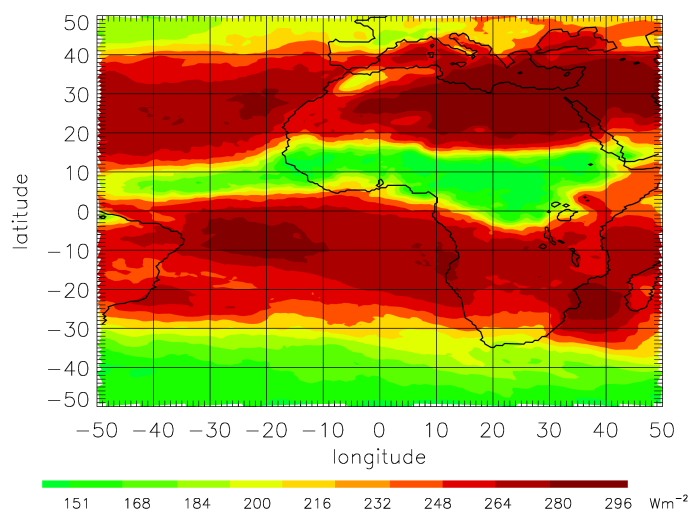
It is expected that under the two scenarios, certain parts of the power spectrum will be inhibited. For example, by filtering clear sky, it is expected that the semi diurnal and diurnal cycle to be inhibited and in terms of the Fourier map, the maximum diurnal power (I.e. over the desert region) will decrease.

The filtering method used in this study utilises GERB SW scene ID at the begin-

ning of section 4.7, except that each time series is interpolated to fill in the data gap. The interpolation operates by finding the flux value before and after the data gap and averaging the two points to interpolate the gap. However, in the interpolation



(a) 2006 July mean minimum diurnal flux (Wm^{-2})



(b) 2006 July mean maximum diurnal flux (Wm^{-2})

Figure 4.17.: Average diurnal a)minimum b)maximum fluxes (Wm^{-2}) over the month 2006 July

method, there also exists an override which linearly interpolates the gap if the time gap is found to be less than an hour to avoid creating short timescale square wave which could interfere with the results.

The analysis initially involves simple flux comparison which isolated the cloudy dominated region and the predominately clear sky region from mixed scene types. Fig. 4.17 shows the averaged diurnal minimum (Fig. 4.17(a)) and maximum fluxes (Fig. 4.17(b)), which is obtained by identifying the minimum/maximum fluxes for each day respectively, then averaging these over the month for each individual point.

By producing such a minimum flux plot, one can easily identify the region where the clouds are present/absent during the particular month, where the lowest flux reading of individual diurnal time series would be determined by the coldest cloud top temperature. It's expected that near the ITCZ region and at mid-latitudes this would yield the lowest minimum flux as: the ITCZ and rainbelt region are the predominant source of deep convection and convective clouds, where the minimum cold top temperatures occur. This is confirmed by Fig. 4.17(b) a green belt over the latitude range 5-10 degrees where the wet region is situated during the month; on the other hand, the mid-latitude receives less ISR compared to the tropical and sub-tropical regions, which is coupled with the clouds influenced by the mid-latitude Rossby waves, producing a low outgoing longwave, hence another two green regions near -50° and 50° . Both the ITCZ and southern mid-latitude regions exhibits a low mean minimum diurnal flux of $150-170Wm^{-2}$, with the northern mid-latitude's flux ranges 180 to over $200Wm^{-2}$ due to seasonal influences.

By observing the minimum and maximum fluxes, it appears that the intermittent clouds dominate the region across the ITCZ and therefore, we are likely to see this region to be excluded from our clear sky study.

The minimum flux plot also displays a moderately high flux reading along the tropical/sub-tropical ocean and land, with the highest flux reading over the Sahara, Ethiopian highland and Mediterranean. This is as expected due to the large ISR received in these regions and the relatively cloud free scenarios in the tropical region outside of the ITCZ and rainbelt. The mean minimum flux reading ranges from 250-300 Wm^{-2} over the tropical region.

The average minimum flux plot (Fig. 4.17(a)) provides us with a map of identifiable cloudy regions, 1) ITCZ, 2) mid-latitudes, which is helpful in explaining our

clear/cloudy scenario analysis.

The maximum flux plot (Fig. 4.17(b)) shows us that the tropical region outside of the ITCZ has a higher flux reading in general, with the land dominating the ocean, due to the difference in the surface temperatures and the relatively cloud-free scenario identified from the minimum flux plot. The highest fluxes are found in the Sahara desert, with a peak flux of around $350Wm^{-2}$ which is as expected due to the low surface heat capacity and the lack of moisture, which therefore, produces a high daytime peak temperature.

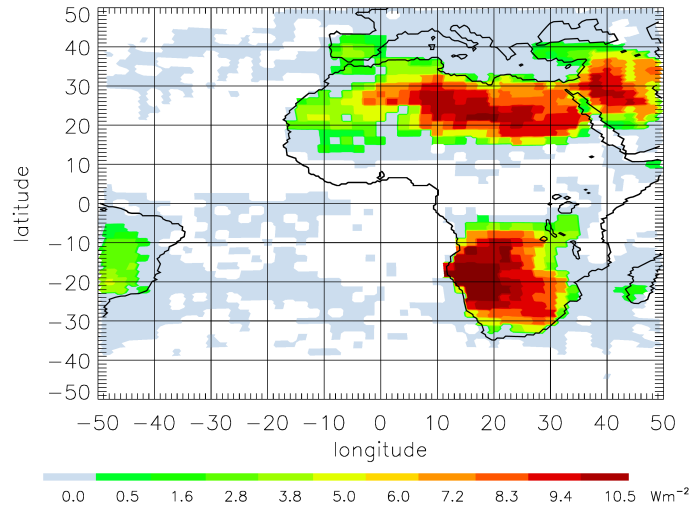
Whereas, the smallest magnitudes are found generally in high latitudes and along the ITCZ region over land, where the averaged diurnal peak could be as low as $225Wm^{-2}$, indicating those regions are under the influence of clouds, in agreement with the minimum flux plot. In particular, the large gradient between Sahara desert and the ITCZ, which suggests near the ITCZ region, the maximum fluxes are consistently inhibited by deep convective clouds.

4.7.1. Clear sky analysis

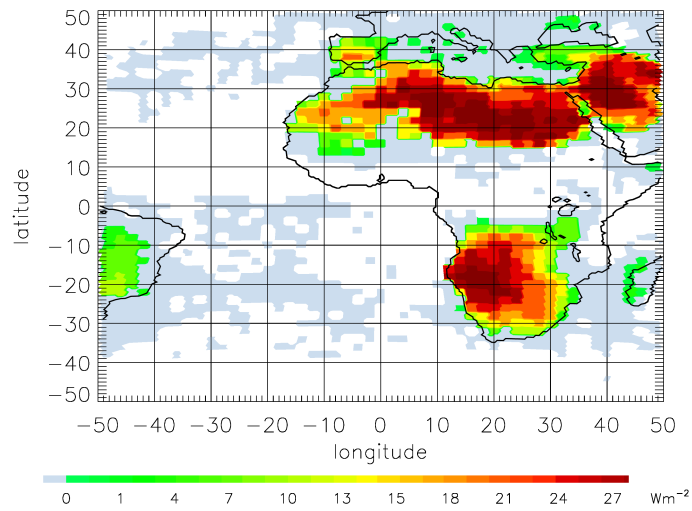
This section presents the results from the clear sky analysis, where the OLR data is pre-filtered according to the cloud cover parameter obtained from GERB scene identification in the shortwave data file. The threshold for the filter has been set to 10% which essentially filters out all the data which exceeds cloud cover ratio of 0.10 over any particular GERB grid points. This is similar to the filtering technique to be presented in chapter 5. Therefore, this technique suffers the same limitation noted there, which is the lack of any cloud data during local night time, which in our analysis, has been set to leave the data unfiltered during local night time. However, if persistent day time cloud is found to exist over a particular grid point, the point is deemed to be cloudy for the whole day and excluded from our analysis.

The filtered data is then Fourier transformed using the same technique as section 4.3 and the spatial power distribution of different frequencies are obtained to compare and highlight any changes from the all sky analysis. The semi-diurnal and diurnal spatial plots (Fig. 4.18) demonstrate the same distribution of power in the tropical land region outside of the ITCZ as the all sky analysis, whilst the spatial pattern seems to be more continuous over the land region, suggesting that the filtering

technique is fairly successful in removing the cloud cycles within the data despite the limitation mentioned. It also suggests that the tropical land diurnal/semi-diurnal cycles are fairly stable in the absence of clouds.



(a) 2006 July clear sky semi-diurnal power



(b) 2006 July clear sky diurnal power

Figure 4.18.: The clear sky Fourier spatial power (Wm^{-2}) distribution plots for timescales: a)semi-diurnal b) diurnal

In terms of distribution of the strongest signal, it can be observed that the semi-diurnal and diurnal cycle signals are strongly expressed in Southern Africa and the Sahara. The distribution of strongest diurnal signal is where a high ISR is coupled with relatively dry weather, therefore, one would expect the greatest day/night time temperature difference in the Sahara and Kalahari deserts and hence the highest diurnal power.

It can also be observed that the west coast of northern Africa has a significantly lower diurnal and semi-diurnal power than in the central northern Africa, which suggests a moderate night time cloud influence or lack of high day time readings (due to clouds) in these regions, hence the inhibited diurnal and semi-diurnal power. This is also observed in the minimum flux plot 4.17(a) where the minimum diurnal flux drops across west coast (20-30N 5-15W).

Fig. 4.19 demonstrates the 2-3 days spatial variability of the Fourier power, comparing with the all sky plot (see Fig. 4.13(a)). The filtered analysis shows lack of ITCZ and mid-latitude features. This is as expected due to the nature of this variability which largely expressed itself in the development of clouds. With a filtered data set, this mode of variability cannot be observed in this regions.

It can be observed that the tropical regions outside of the rainbelt retained much of the same spatial distribution as the all sky plot, suggesting an imperfect filter which leaves out the night time condition which contributes to this weakening of day to day variability: we are observing a clear sky signal with small 2-3 days variability due to less significant climate components in radiation, e.g. change in atmospheric humidity. Hence, the small Fourier power in northern and southern edge of the African continent.

To further investigate the longer timescale variability, the 4-7 power distribution plot for clear skies is shown in Fig. 4.20. Although the magnitude of this 4-7 days Fourier power has been dramatically reduced by filtering the data and the rainbelt/ITCZ dominance in the power plot has disappeared. It still displays a band structure just north of the rainbelt, as well as the west coast of northern Africa which has been identified as cloud contaminated in Fig. 4.18. This implies the following: the filter applied is imperfect and the 4-7 day timescale variations associate with cloud is identifiable through night time flux and under low cloud cover conditions; and the wave disturbances influence the region north of wave structure

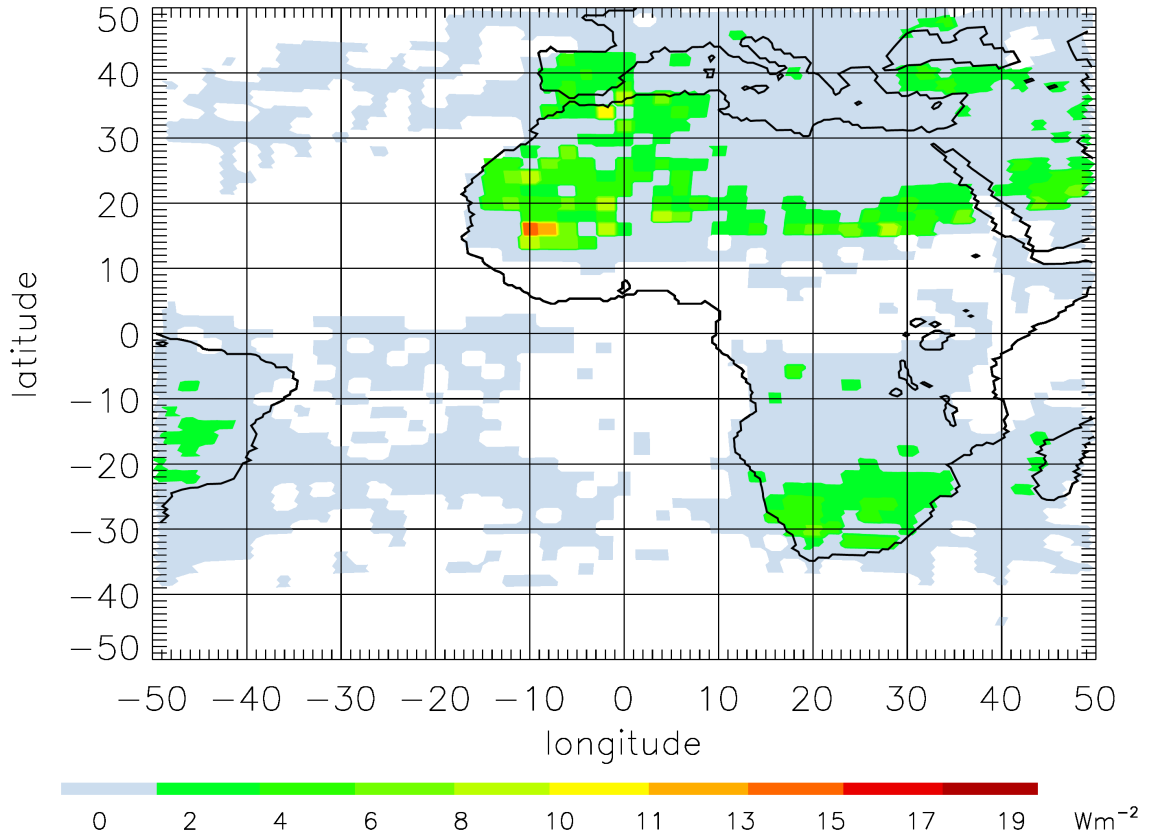


Figure 4.19.: The clear sky Fourier spatial power (Wm^{-2}) distribution plot for 2-3 days timescale (July 2006)

with a peak wave period (4-7 days) that is slightly higher than that observed in the centre of the ITCZ (2-3 days). As seen from the diurnal and semi-diurnal plots, cloud is not the only significant component in this region, there's also a mixture of moderately strong ISR induced diurnal/semi-diurnal cycle and moisture variation (i.e. convection) which partially determines the dominant period of the variability along this band.

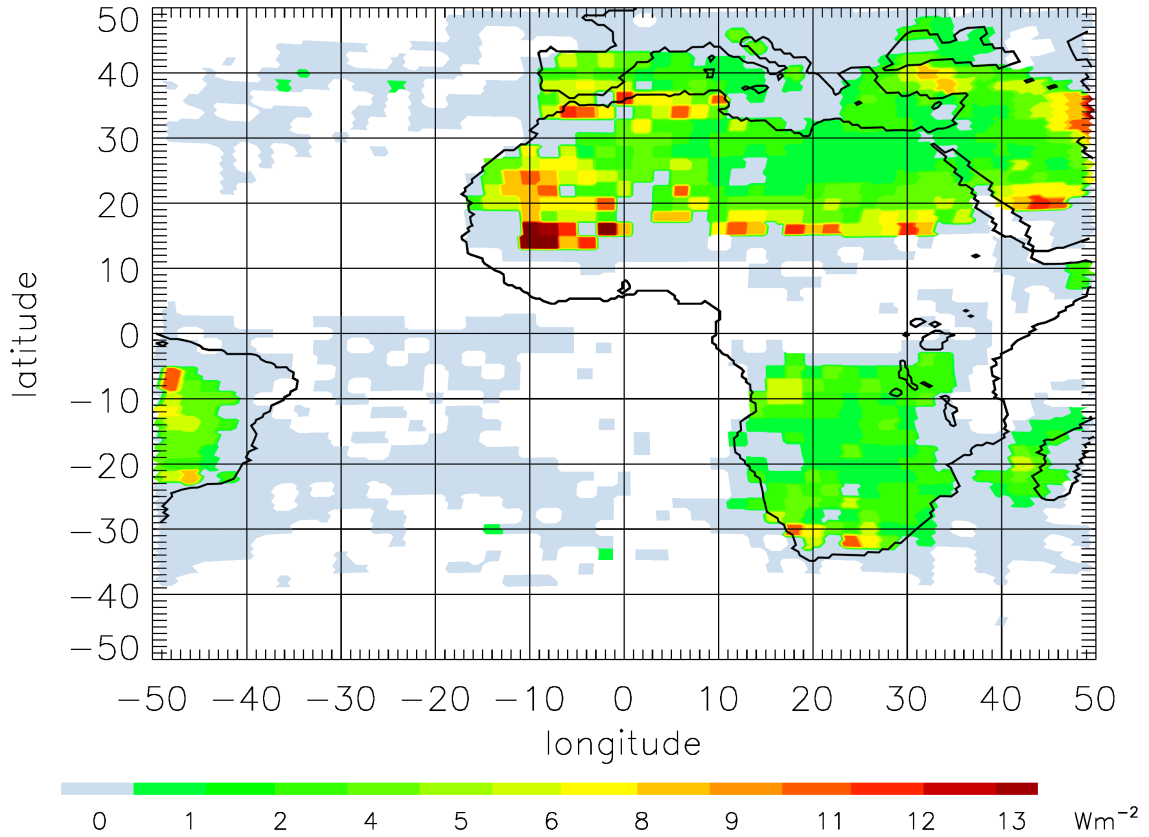


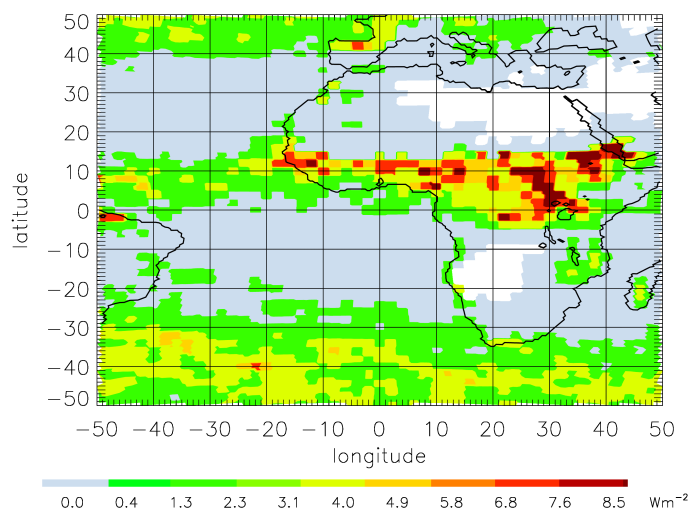
Figure 4.20.: The clear sky Fourier spatial power (Wm^{-2}) distribution plot for 4-7 days timescale (July 2006)

4.7.2. Cloudy sky analysis

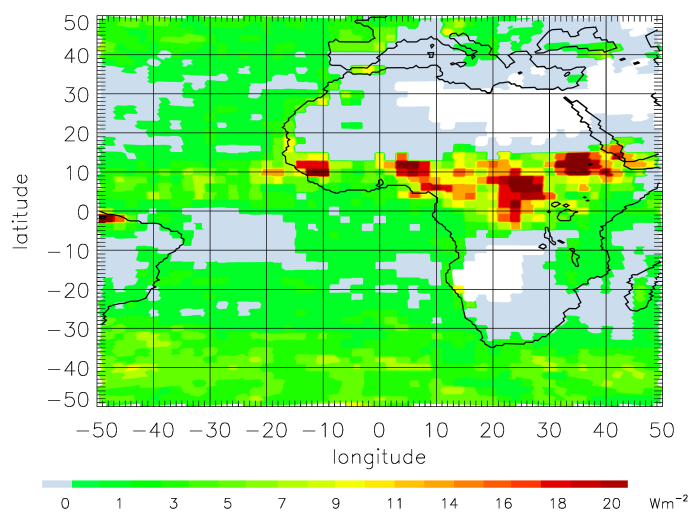
It is believed that by employing a cloudy filter instead of the clear sky filter, the cloud contribution to various modes of variability can be analysed and represented in spatial terms. The filtering method used in this analysis is largely similar to the clear sky technique, however, instead of setting a cloud threshold on 10%, the threshold was set higher at 30%. This threshold is used because of the amount of data left is still sufficient to observe a spatial distribution whilst not affecting the overall results with clear sky strong diurnal cycles. Therefore, all the data which corresponds to a cloud cover higher than 30% and all the night time data will be

accepted, and similar to previous approach, the filtered data will undergo a Fourier transform.

Fig. 4.21(a) and 4.21(b) showed the semi-diurnal and diurnal cycle of the cloudy



(a) 2006 July cloudy sky semi-diurnal power



(b) 2006 July cloudy sky diurnal power

Figure 4.21.: The cloudy sky Fourier spatial power (Wm^{-2}) distribution plots for timescales: a)semi-diurnal b) diurnal

sky Fourier power respectively.

First of all, both the power maps in Fig. 4.21 demonstrate a lack of strong diurnal/semi-diurnal signal in the tropical region outside of the ITCZ, which indicates the filter is successful in filtering the predominant clear sky tropical region observed in all sky. Secondly, the diurnal map shows the strongest signals along the ITCZ which indicates some cloudy regions exhibit a moderately strong diurnal cycle, which could be attributed to both unfiltered/partially clear sky land surface heating scenarios and diurnal convective activities. This could also explain the distribution of strong power confined by the land boundary, where the strong diurnal heating and convection or deep convection are mostly observed over the land territory.

Finally, by comparing the distribution of both the power maps, it can be observed that there is some overlapping of the stronger signals. However, unlike the clear sky and all sky plots, they seem to be less coherent with one another and note that the diurnal cycle remains a regional effect across the ITCZ (i.e. the diurnal signals appear in patches rather than just one point), whereas the semi-diurnal cycle seems to exhibit some discontinuity, where the strong signals look much more granulated. This indicates the semi-diurnal are much more localised, which perhaps suggests the rarity of the clear sky diurnal heating signal along the ITCZ. Despite the small incoherence between the diurnal and semi-diurnal patterns, we can still observe a counter part in diurnal signal for the strong semi-diurnal signal. Hence, backing up our hypothesis on the semi-diurnal component that this component can be attributed to the ISR response and the heat capacity of the surface material.

Fig. 4.22 illustrates a similar power map for period 2-3 days, where the power distribution largely resembles that of the all sky plot, with distinguishable strong signals along the ITCZ and the mid-latitudes. However, with careful observations, one can recognise the narrower band of ITCZ contribution from the cloudy sky power, 5°S - 15°N than that of all sky, 5°S - 20°N (see Fig. 4.13(a)).

This is an interesting phenomenon as we have identified the 2-3 days to be largely due to the cloud, thus we would expect the 2-3 day variabilities to be invariant from all sky to cloudy sky along the ITCZ. However, it's found that the northern part of the ITCZ to be missing under our 2-3 cloudy sky power plot. This could potentially be explained by our observation in section 4.7.1 where the "missing" stripe is where the atmospheric conditions are a mixture of clear and cloudy sky, with the wave

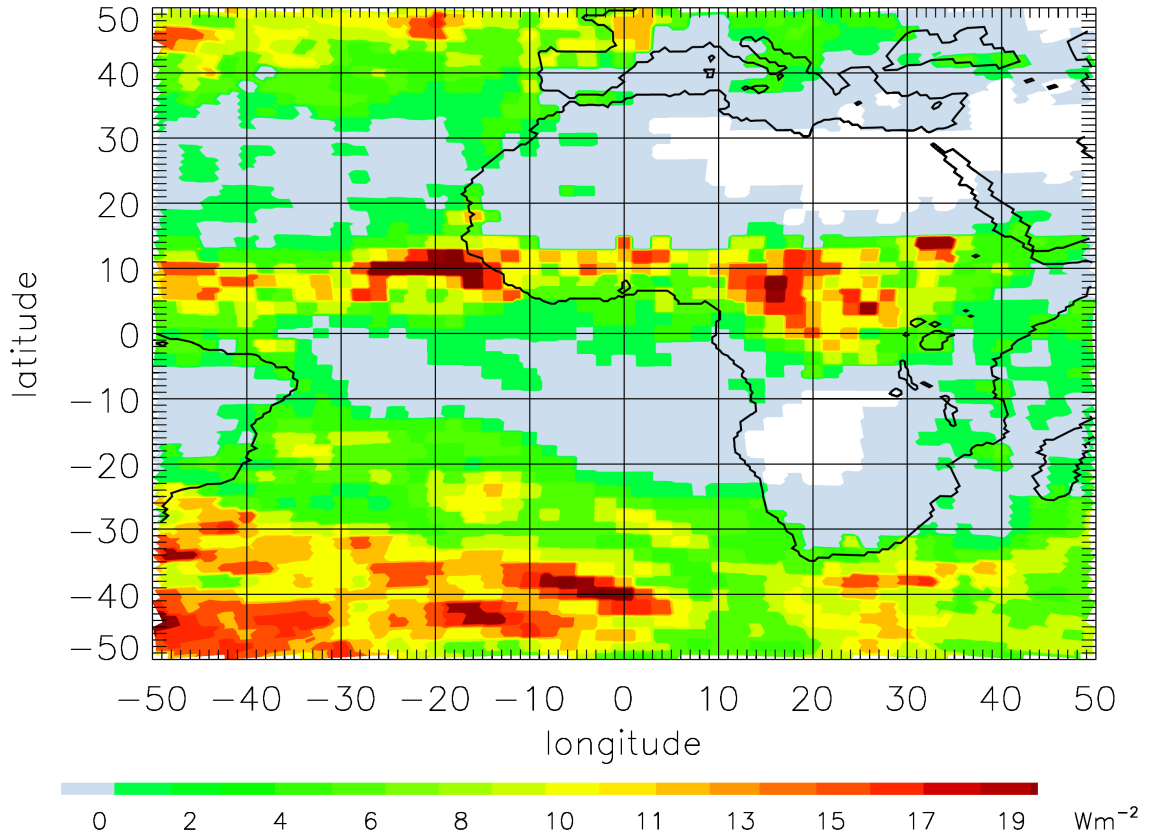


Figure 4.22.: The cloudy sky Fourier spatial power (Wm^{-2}) distribution plot for 2-3 days timescale (July 2006)

disturbances driving the moisture/clouds at a period of 4-7 days.

From the observations in Fig. 4.23, it can be seen that the stripe like pattern is not present in the power plot, but only in the clear sky 4-7 days plot, which suggests that this process is observed over a predominantly "clear" region. This prompts further investigation in the wave like disturbances in this region to advance our understanding in the rainbelt region and area around it as well as the association with African waves.

The 4-7 plot (Fig. 4.23) displays a similar spatial distribution compared to the all sky, inheriting the same strong features off the west coast of northern Africa and

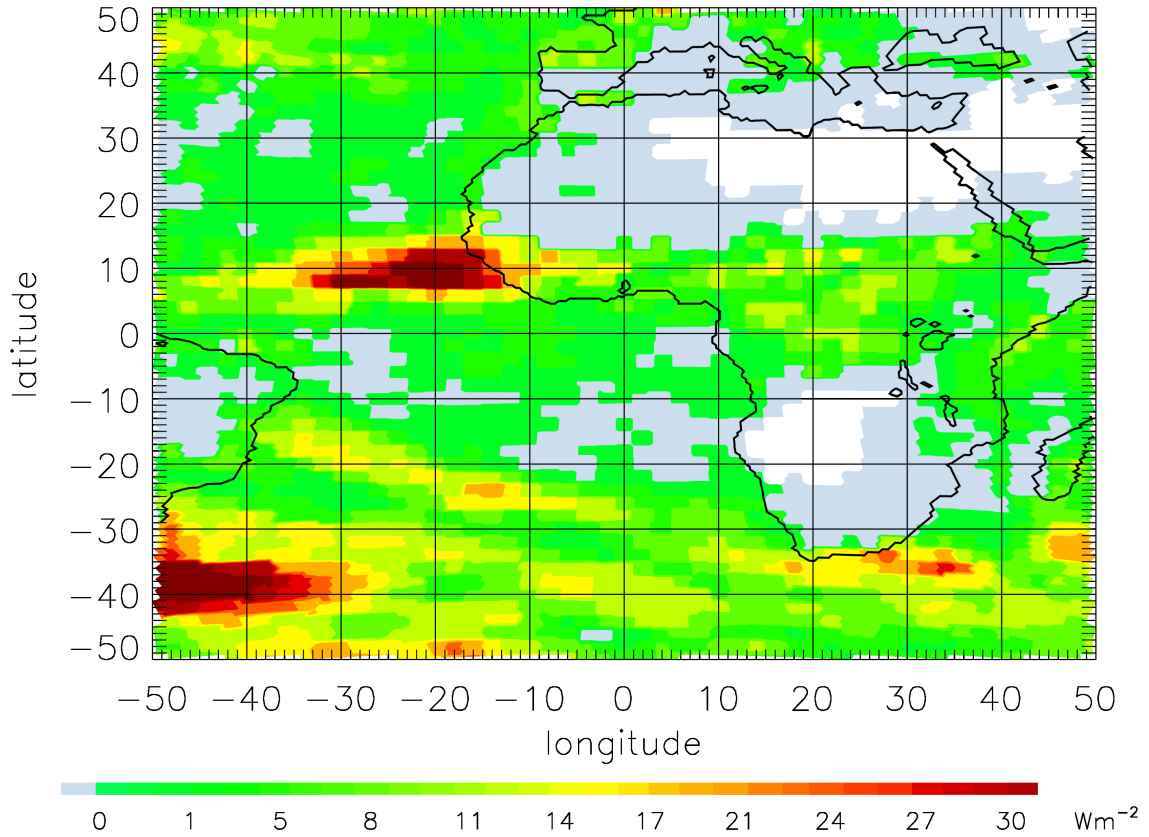


Figure 4.23.: The cloudy sky Fourier spatial power (Wm^{-2}) distribution plot for 4-7 days timescale (July 2006)

in the mid-latitude ocean just south of Brazil. Again, this power plot expressed the same narrower band structure along the ITCZ, resembling the same difference between 2-3 day cloudy to all sky plot.

4.8. Chapter Summary

This chapter has identified the significant modes of variabilities using Fourier technique. This was done by averaging the GERB grid points into $2 \times 2^\circ$ boxes to attempt to eliminate some of the noise due to localised meteorology. Furthermore, we have

analysed the semi-diurnal, diurnal, 2-3 days, 4-7 days variability in spatial terms.

In the semi-diurnal analysis, we have observed a spatial distribution of the signal which is very similar to that of diurnal timescale, suggesting that this signal could be a manifestation of the diurnal cycle which cannot be decomposed by a single mode of sine wave. There is a semi-diurnal component in the wind field variation known to be caused by atmospheric thermal tides (see [Deser and Smith, 1998]). However, this is not associated with any major surface or cloud activities which could contribute significantly to the OLR. Therefore, this cannot explain up to $10Wm^{-2}$ signal ($20Wm^{-2}$ peak to trough) that we have observed in our semi-diurnal power plots. Furthermore, this mode of variability is clearly suppressed along the ITCZ which is consistent with our hypothesis that this mode is part of the diurnal cycle.

The diurnal component was observed to be largely dominated by the land processes due to the large day and night temperature variation. We have also noticed the different spatial distributions during boreal winter and summer, which can be mostly accounted for by the change in local SZA as well as the shift in the position of the ITCZ. This suggests the significance of convection and clouds in the diurnal variation. This prompted an investigation via a principal component analysis to further decompose the diurnal cycle into different components, which is presented in chapter 5.

The 2-3 days and 4-7 day cycles are thought to be related to the ITCZ and African wave disturbances, and we have carried out a series of conditional tests which indicated that the longer period variabilities are largely due to cloud formation and tend to structure around the ITCZ.

The previous studies reviewed in chapter 2 have investigated the variability of the diurnal cycle (Slingo *et al.*, 2004; Tian *et al.*, 2004, Yang and Slingo, 2001) and this chapter has confirmed the variability of such cycles. In particular, the importance of the cycle over land and the suppression of the cycle over the convergence region has been observed in our results, this was also seen in [Yang and Slingo, 2001] in the form of brightness temperature from a narrow band window measurement. We have confirmed the same variability was observed in our broadband OLR study in this chapter.

Furthermore, Clayson *et al.* (2002) have observed 2 to 3 days variability over the tropical western pacific ocean due to convective oscillation which was also ob-

served in our 2-3 days variability plot. They have observed a filtered flux of up to $120Wm^{-2}$ peak to trough variation during the ocean convective activity. We have seen similar peak to trough magnitude of low pass filtered flux in our analysis over Niger (over $100Wm^{-2}$) in July which is subjected to African wave disturbances during the season.

Clayson et al. (2002) have proposed the moisture and temperature provided by the warm SST to be a cause for such variability, and the inertia gravity waves (IGW) to be the driving force for the convective activity to propagate eastward/westward. Although our filtered flux is measured over land, there exists similar conditions for our chosen location where the moisture and uplift force are provided by the ocean moist air from the Gulf of Guinea and the air convergence in northern Africa as part of the AEJ (see section 2.4 for description).

Their analysis has focused on the OLR variability over the Pacific ocean and linked this variability with the SST and the associated convective variability. We have identified the same mode of variability in our Fourier analysis but the signal is not limited to ocean only, it was also observable near the ITCZ which has also been suggested by *Holton* (1992) to be a result of the AEW with a period 3-4 days over the same region. Thus, our study confirms that the 2-3 days variability is observable over the region.

Our study has further confirmed the higher period variabilities to be present over the same region, which is consistent with previous studies on the AEW of 2-9 days variability (e.g. *Mekonnen et al.*, 2006; *Gu et al.*, 2004; *Grist*, 2002 and *Diedhiou et al.*, 1999). They have investigated the dynamic variability and the convective variability of the wave using wind measurements and brightness temperature as a proxy for convective activity. They have chosen such variables for observation due to the mechanism of the wave, where the temperature gradient between the Gulf of Guinea and Sahara forces a wind convergence above the desert, and the convergence itself forms a jet stream and instabilities which gives rise to the wave and the associated convective activity and precipitation. The wind measurement provides a description of the wave dynamics whereas the brightness temperature provides a mean of measuring an effect of the wave (convective activities).

However, by directly observing the variability in the OLR, we can see the overall influence (cloud, precipitation and the tropospheric humidity) of the wave has on the

upward broadband flux observed from TOA. This provides us a method to observe how these wave variability fits in the the overall variability of the OLR over the short timescale. We have observed and quantified the direct influence of such these modes of variability in the broadband OLR to peak at $20Wm^{-2}$ ($40Wm^{-2}$ peak to trough) and $30Wm^{-2}$ ($60Wm^{-2}$ peak to trough) for the 2-3 days and 4-7 days variability respectively.

Moreover, our clear/cloudy sky analysis has shown these cycles (2-9 days) to be related to clouds and are almost unobservable in our clear sky studies. This chapter has confirmed the important modes of variability with previous studies (*Slingo et al.*, 2004; *Holton*, 1992; *Nicholson*, 2009) and highlighted the diurnal mode, which is particularly important over land in general under both clear and cloudy sky conditions. Therefore, the diurnal cycle will be investigated in detail in the rest of this thesis.

5. Analysing OLR in Diurnal Timescale over Africa and Atlantic Ocean

5.1. Introduction

This chapter of the thesis aims to analyse the variability of the OLR in diurnal timescales over the region of Atlantic and Africa, where the data are first decomposed using principal component analysis (PCA) at the diurnal timescale as this is identified to be one of the most significant modes of variability from Fourier analysis (chapter 4). This is due to the fact that it is driven by the solar input and due to the climate processes associated with this mode, e.g. surface response to ISR, convection and convective clouds. The motivation of this chapter is to investigate the different components within the diurnal cycles of the OLR and to identify these processes behind the components.

The diurnal cycle and the variabilities with respect to different surface types and different atmospheric conditions will be analysed in the chapter. This analysis uses data from the GERB instrument.

The first part of the work will involve decomposing a monthly averaged diurnal dataset into principal components (PC) and associated empirical orthogonal functions (EOF). These results will be studied to investigate the time varying components within the diurnal cycle of OLR as well as the spatial distribution of the strength of this signal.

These initial results will then be used as a reference to compare with some more detailed analysis, namely, non averaged, surface types and clear/cloudy sky analysis, where the components are expected to change with respect to these variables.

5.2. All sky averaged flux analysis

5.2.1. All surface types averaged flux analysis

The first part of the analysis investigates the decomposed OLR for all sky over the region of interest, namely Africa and the Atlantic ocean with latitude range of 30°S to 30°N and longitude of 30°W to 40°E. The main objective of this section is to compare with previous studies addressing diurnal variabilities to verify the results (PC1&2) and to ensure a similar outcome with a region limited dataset. By doing this comparison, it will give confidence to the robustness of the technique.

This part of the analysis involves averaging each month of data into a single diurnal cycle for each individual GERB grid point. The data were first re-arranged according to the local solar time. Then, the monthly data were averaged into daily data, preserving the 96 timesteps (15 minutes per image) during the day. This essentially created 96 averaged images of the chosen region with the images arranged along the time axis.

The months employed primarily for this particular work are January and June of 2006. Although July and December were also analysed, the results were not shown in this thesis. This is due to the statistical similarities between the PCs of July and June; December and January. Although there are minor differences in the spatial distributions of the EOFs, these are mostly caused by the change in SZA which is minimal for consecutive months.

The averaged diurnal cycles for the region chosen were then decomposed using PCA. This gives 96 eigenvectors with a time length of 96 timesteps and corresponding 96 eigenvalues which weight the eigenvectors.

In Fig. 5.1, the decomposed signals from PCA of the OLR data are displayed. It is seen that the first PC which represents 77% of the overall variance in the dataset, has a maximum just after noon, at around 1300. This component possesses a flux strength of around $20Wm^{-2}$ when combined with its maximum EOF score of 3.0, this gives roughly $60Wm^{-2}$ of diurnal variation which is consistent with the diurnal variation of clear sky desert. This particular PC also exhibits an asymmetry about its peak. This principal component is believed to contain most of the surface and clear sky heating response from the ISR and will be referred to as surface heat response hereafter due to a few of its properties: 1) the PC peaks just after noon,

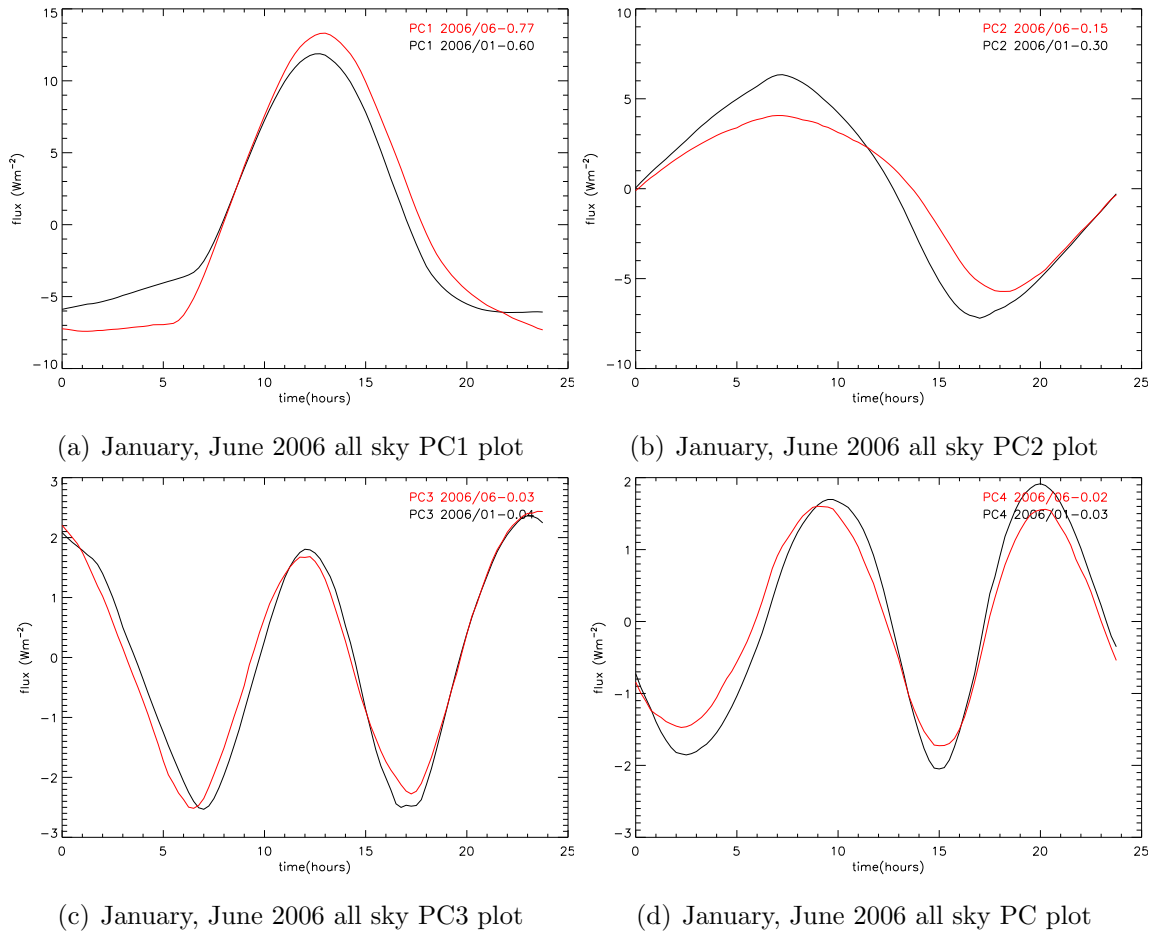


Figure 5.1.: The a)first b)second c)third d)fourth order PCs for January and June 2006, numbers on top right corner of each plot indicate the fraction of variances for the individual components. The change in Outgoing Longwave in Wm^{-2} and local time in hours are represented by y-axis and x-axis in the plots respectively.

the maximum incoming solar radiation is at noon; 2) the strongest EOF signals are present over the Sahara desert (see Fig.5.3(a)) where the day and night temperature variation is largest; 3) almost all positive signals are found only over land.

This slight lag between the maximum in ISR and peak in PC1 is due to the finite heat capacity of the surface.

The ability of PC1 to reproduce the surface heat response signal is further confirmed by recreating the OLR signal using just PC1 and EOF1. This recreated signal is subtracted from the original OLR to create a residue power plot (not shown in

this thesis). It was found that most of the residual power was focused in the ITCZ, confirming that this PC is a indeed a clear sky signal and able to describe most of the OLR signal power outside of the ITCZ.

The second PC signal for June 2006 (see Fig.5.1), the peak to trough difference is roughly $10Wm^{-2}$ when considering this either the EOF score range of -2 to 4, this gives a diurnal variability of $-20Wm^{-2}$ and $40Wm^{-2}$ respectively. This component along with the PC1 diurnal variation, provides a combined diurnal variability of around $100Wm^{-2}$ which is consistent with the OLR variability we have observed in the cloud infested ITCZ region in the points/latitude band analysis in chapter 4.

The second PC is sine-wave like, with maximum at 0600-0700 and minimum at around 1800-1900. This PC contributes to about 15% of the total variance, by comparing this with the cloud cycle and analysing the geographical distribution of this component in Fig.5.4(a), the PC is considered to be mostly due to cloud signals where the trough in this PC is thought to correspond to the maximum in cloud depth. The OLR measured at TOA for cloudy scenes are generally lower than that of the clear sky. Therefore, a minimum in OLR in this case would represent a maximum in the cloud optical depth and vice versa.

In contrast to the first two PCs, PC3&4 are found to have two peaks and contribute to much less variance overall in the OLR data. These two PCs are both sine-wave like and PC3 peaked at 1200 and 2300-0000LST. It should be noted that one of the PC3 peaks is at around 1200, about 1 hour before PC1, we will discuss this further along when we address the geographical distribution of PC3.

(a) June 2006 PCA error				(b) January 2006 PCA error			
PC	Variance %	$\Delta\lambda$	$\delta\lambda$	PC	Variance %	$\Delta\lambda$	$\delta\lambda$
1	0.77	4309.87	52.10	1	0.60	1954.18	37.76
2	0.15	815.57	10.20	2	0.30	1678.07	18.76
3	0.03	110.74	2.28	3	0.04	81.30	2.45
4	0.02	81.11	1.20	4	0.03	122.80	1.66

Table 5.1.: Principal components' (1-4) variances and error criteria for: a) June 2006 and b) January 2006; $\Delta\lambda$ = difference between eigenvalues, $\delta\lambda$ = standard deviation in the eigenvalue

Table 5.1 demonstrates the variance and error criterion for the PCs in June and January 2006 as defined in chapter 3, $\Delta\lambda$ and $\delta\lambda$ corresponds to the difference in

eigenvalue and the standard deviation of the eigenvalue respectively (see section. 3.4.2). This is a criterion derived by *North et al.* (1982) to test the validity of an EOF, the criterion requires the $\Delta\lambda > \delta\lambda$ for the associated PC/EOF pair to be valid. The tables 5.1a) and b) showed that $\Delta\lambda$ is significantly larger than $\delta\lambda$ for all the PCs demonstrated in Fig.5.1 in both January and June 2006, which confirmed the validity of the said PCs. Furthermore, the variance contribution of each PC confirmed the statistical significance of the PC in the dataset, which is greater than 1% of the total variance for all the PCs shown.

Fig.5.1 shows January decomposed signals: by comparing with June PCs, it can be seen that the peaks of the signals are roughly in phase with each other. However, the PC1 in January represents much smaller overall variance (60%) compared with June. This can be explained by variation in the location by minimum SZA at 23°S in January and at 23°N in June. The larger portion of ocean at 23°S compared with 23°N reduced the total variation in PC1 variance due to surface heat response.

By further comparing PC2 between the two months, it was observed that in January, PC2 contributes to higher overall variance (30%), which could be due to the larger overlap between African land and the ITCZ and the fact that this land is a vegetated surface. The former case would result in more GERB grid points under ITCZ cloudy condition, whereas the latter case would provide more moisture for convection and convective clouds, thus contributing to the magnitude of the cycle as well as increasing the frequency of occurrence for the convective events. Therefore, increasing overall variance in comparison to July.

Fig.5.2 displays the different surface types across Africa to enable comparison with the EOF signals and enable us to observe how the power distributes over different surface types. The surface type used in the study are from GERB surface data, and is a static map of 5 surface types (Ocean, dark vegetation, bright vegetation, dark desert, bright desert).

Fig.5.3 illustrates the geographical distribution of EOF1 in both June and January 2006. Strong positive signals are generally found over land. In particular, June data expresses strongest signals over the Sahara desert, where the minimum SZA is. However, January shows a stronger signal in central African region and the Kalahari desert in southern Africa, supporting the idea of a strong relationship between first PC and surface heat response signal. There also existed a weaker positive signal

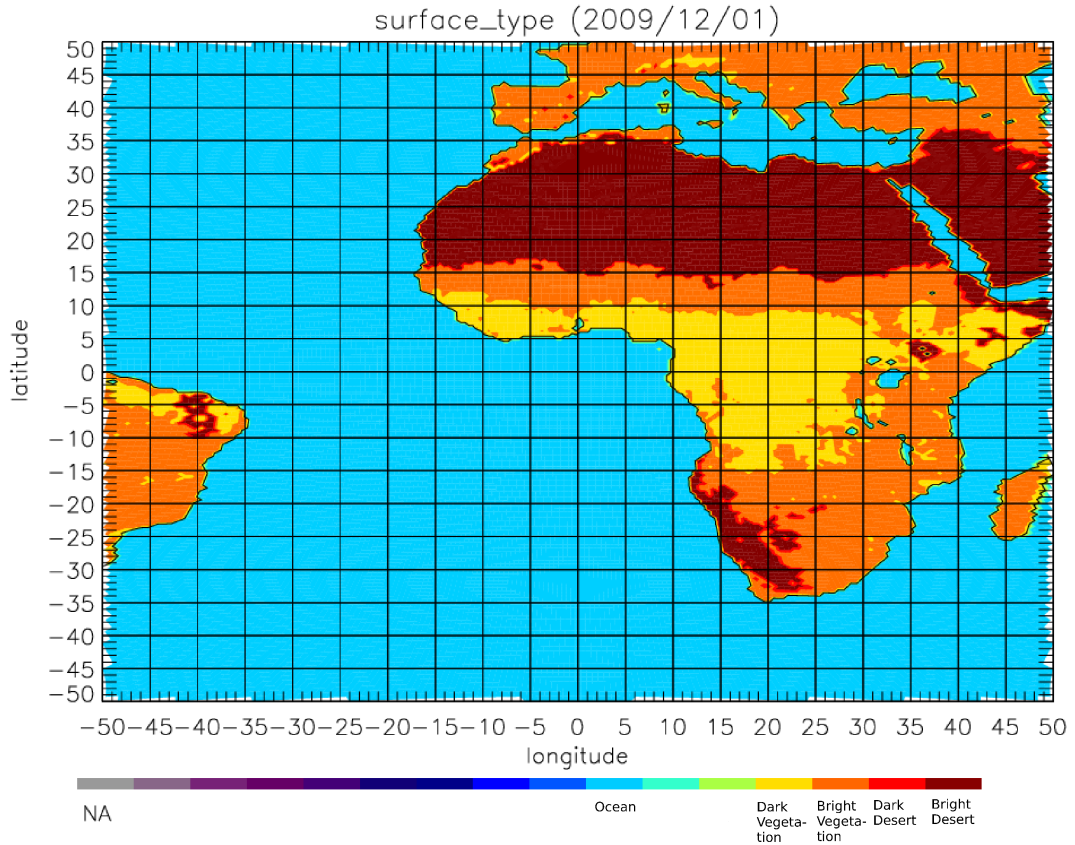
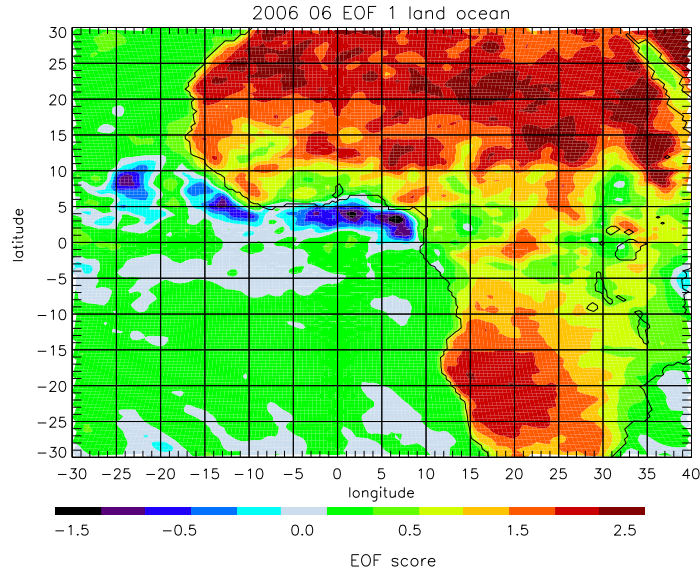


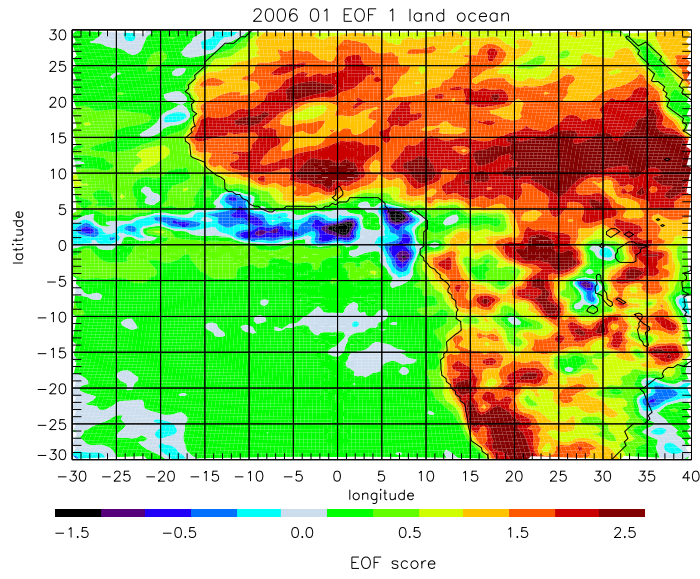
Figure 5.2.: surface type plot showing dark desert, bright desert, dark vegetation, bright vegetation and ocean, NA indicates the part of colour scale which is not applicable to the diagram

along the ITCZ band, as a consequence of the lower surface heat response over the cloudy region. It was also found that a negative signal emerged over the ocean along the ITCZ, signifying a different OLR cycle over the ocean to the land, perhaps indicating either ocean signal in anti-phase with land surface clear sky heating signal, or a cloud development signal which is in phase with the surface heat signal resulting in an OLR signal in anti phase with PC1. The latter case is more likely to be the mechanism for this PC due to the amount of cloud present along the ITCZ over the ocean as well as the relative strong diurnal signal observed in chapter 4 over this part of the ocean compared with tropical ocean outside of the ITCZ.

In January (see Fig.5.3), it's observed that the land is still the dominant factor in determining the geographical distribution of the magnitude of the signal. However,



(a) June 2006 all sky EOF1 plot

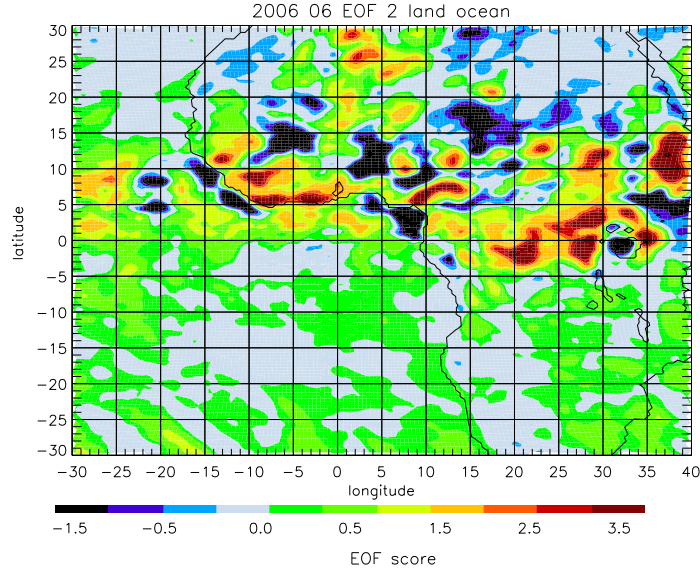


(b) January 2006 all sky EOF1 plot

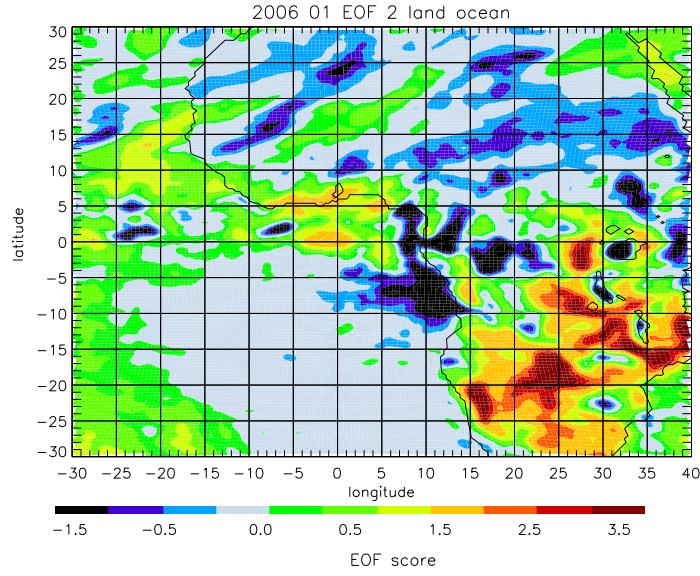
Figure 5.3.: EOF1 plots for a)2006 June b)2006 January, where the colours represent the EOF score over the region 30°S - 30°N and 30°W - 40°E , white indicates data absence. The x and y-axis denotes the longitude and latitude respectively. The EOF score is dimensionless, requires the associated PC to interpret the OLR component.

the strongest signals are now found in central and southern parts of Africa, with some strong signal regions overlapping with the ITCZ. This is expected since the minimum SZA is in the south during January, enhancing the surface heating signal should be found in these regions. It should also be noted that surface type is another important factor in determining the strength of the diurnal cycle, hence the significantly stronger diurnal signal observed in Kalahari desert compared with the southern Africa. The plot also demonstrated regions of weaker signals along ITCZ over land, confirming that clouds do indeed inhibit the magnitude of PC1, the clear sky surface heat response signal. The surface heating signal was observed along the ITCZ, this could be caused by the clouds which are not persistent in this region, and some clear sky data could still be observed. Or that some of the cloud signal resembles the surface heat signal. The former case is in disagreement with our clear sky observation in chapter 4, where we showed that there appeared to be a lack of diurnal component along the ITCZ under clear sky condition. Whereas, the latter case is supported by the cloudy sky analysis in chapter 4 where a relatively strong diurnal component was observed along the ITCZ under cloudy sky condition. Therefore, our study suggests that the PC1 component along the ITCZ is likely to be a cloud development signal in anti phase with the surface heat signal, hence resembling an overall signal which is very similar to surface heating.

By comparing the second EOFs with the first EOFs and observing the associated PC2, one can be more certain that the most dominant contributor to this PC is cloud. In Fig.5.4(a), low magnitude signals were observed in most regions over land and ocean. However, there exists a band of strong mixed positive and negative signals along the ITCZ band. A strong positive PC2 signal represents a strong cloud cycle signal. This PC is likely to be associated with a particular type of cloud activity, convective clouds, since these are focused almost entirely in the ITCZ, and are not observed over other cloudy regions, such as off the west coast of southern Africa where marine stratocumulus is present. Along the ITCZ, a negative signal can also be observed in some regions, indicating an inverted PC2 response, caused by either early morning cloud development or advection of clouds into these regions during late night or early morning. Furthermore, some strong signals are observed in mountainous regions (in some cases, just outside of ITCZ), implying a topographical dependency on the strength of this cloud signal.



(a) June 2006 all sky EOF2 plot



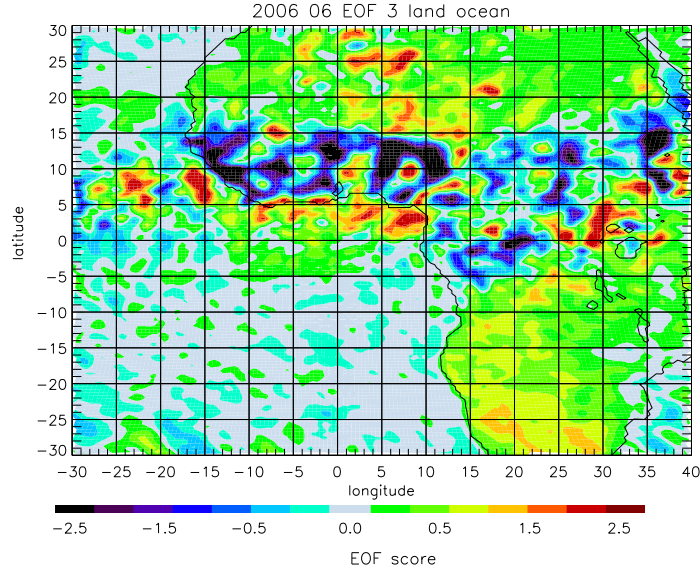
(b) January 2006 all sky EOF2 plot

Figure 5.4.: EOF2 plots for a)2006 June b)2006 January, where the colours represent the EOF score over the region 30°S - 30°N and 30°W - 40°E , white indicates data absence. The x and y-axis denotes the longitude and latitude respectively. The EOF score is dimensionless, requires associated PC to interpret the OLR component.

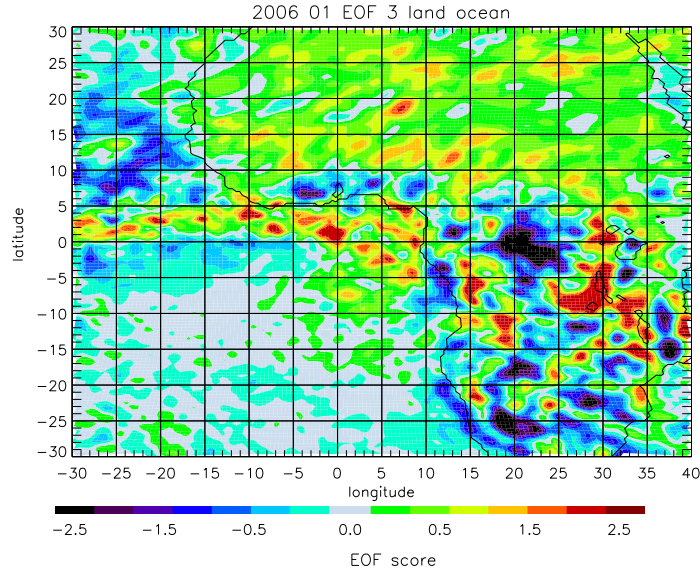
In comparison, the January EOF plot has shown that the strongest signals were present along the ITCZ which has shifted to the south in this month, confirming a strong relationship between convective clouds and PC2. The strongest negative signals were found along the coast of central Africa, suggesting advection of clouds from land to ocean in the early morning. Furthermore, the strongest positive signals are again found over the land along the ITCZ.

There appears a strong positive region neighbouring a strong negative region near 30°E in longitude and 0° in latitude over lake Victoria and the region around it, these signals are present in both January and June plots shown in Fig.5.4, and July, December (not shown due to the similarities of PCs and spatial distributions of EOFs between July and June; December January). This suggests that topography and surface type are the dominant features which affect the OLR in these regions, these surface conditions promote convective cloud development and the advection of clouds, creating invariant signals over different months, this is further supported by plots of 2005 and 2006 (not shown). *Yin et al.* (2000) has also observed a diurnal cycle of cloud and precipitation above the lake due to the warmth and moisture via evaporation provided by it which is consistent with our observation of the diurnal cloud cycle over the lake.

Fig.5.5(a) shows the geographical distribution of EOF3 for June and January 2006. In June (Fig.5.5(a)), there exists a band of strong negative and positive signals along the ITCZ and some strong positive signals also appear in regions of north africa, overlapping with strong signals of EOF2. In general, there's a positive signal over land and almost zero signal over the ocean outside the ITCZ. This suggests that PC3 contains a mixture of clear sky surface heating over land and some cloud signals. In terms of the contribution from clear sky, it was found that slightly stronger signal is found in the southern part of Africa than the north, which consisted of mostly vegetation and desert respectively. If PC3 is a clear sky surface heating term, a stronger signal should be found over the Sahara desert, but instead, it was observed in the south, implying that PC3 is not a direct surface heating signal. This PC is observed to be positive mostly over land outside of the ITCZ which implies that it is either caused by clear sky heating or surface properties such as heat capacity. The semi-diurnal nature of this PC suggests the possible causes to be atmospheric thermal tides or surface heating coupled with surface properties,

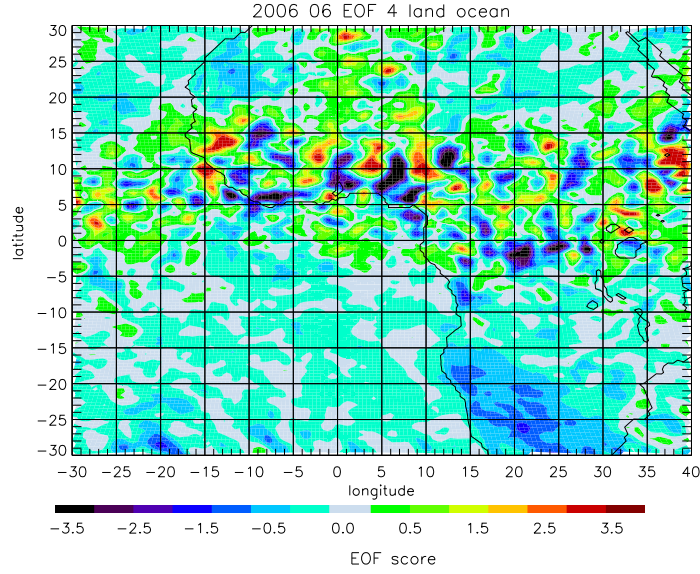


(a) June 2006 all sky EOF3 plot

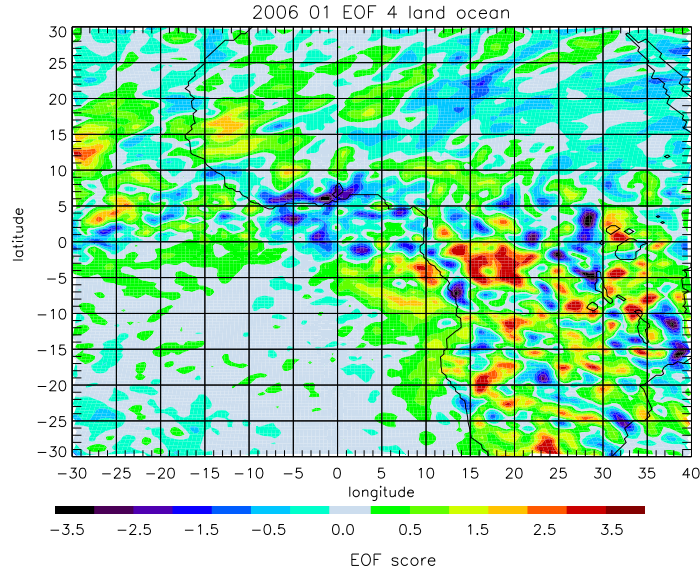


(b) January 2006 all sky EOF3 plot

Figure 5.5.: EOF3 plots for a)2006 June b)2006 January where the colours represent the EOF score over the region 30°S - 30°N and 30°W - 40°E , white indicates data absence. The x and y-axis denotes the longitude and latitude respectively. The EOF score is dimensionless, requires associated PC to interpret the OLR component.



(a) June 2006 all sky EOF4 plot



(b) January 2006 all sky EOF4 plot

Figure 5.6.: EOF4 plots for a)2006 June b)2006 January, where the colours represent the EOF score over the region 30°S-30°N and 30°W-40°E, white indicates data absence. The x and y-axis denotes the longitude and latitude respectively. The EOF score is dimensionless, requires associated PC to interpret the OLR component.

the latter case appears to be the dominant feature in this PC as this is consistent with our semi-diurnal findings in chapter 4.

Similar results were obtained for January, except for the change in latitude of the ITCZ which leads to a corresponding change in the geographical distribution of the strongest signals in EOF3.

Fig.5.6(a) shows the EOF4 plot for June 2006 where again, the strongest signals are found to overlap largely with PC2 along the ITCZ band. However, in contrast to EOF3, the EOF4 signal is found to be zero almost everywhere over both land and ocean outside of the ITCZ, which signifies that this signal is mainly due to cloud. This will be discussed further in the clear/cloudy sky section 5.3.

In January (Fig.5.6(b)), similar results were obtained, again, the distribution of the extreme signals were changed geographically due to the shift in latitude of the ITCZ. Hence, the extreme signals were found to be in southern Africa during this month. However, there appeared to be a lack of clear sky land signal in both figures 5.6(a) and 5.6(b) which indicates the associated PC is independent of surface heating which differentiates this PC from PC3.

5.2.2. Land only analysis

This subsection will focus on a decomposition of land data, where only the ocean data will be removed from the dataset. The remaining data will be treated in similar manner as section 5.2.1.

The results will include the PC and EOF plots from the decomposition, and discuss the differences in the shape of the first two PCs compared to the all sky analysis. The motivation for this part of work is to investigate the diurnal variation due only to the land, to better understand the diurnal cycles associated with specifically land surface type.

In Fig.5.7 the decomposed signals for land only data has been shown. For the June PC signals(Fig.5.7(a)), the first PC1 was observed to have a similar timing of OLR peak compared with all surfaces, but with a slightly more asymmetric curve about its peak, judging from the magnitude of this signal (around $15Wm^{-2}$) and comparing this with the all sky PC analysis, this is likely to be a surface signal. This PC has a percentage variance of 56%, much less than the contribution of PC1

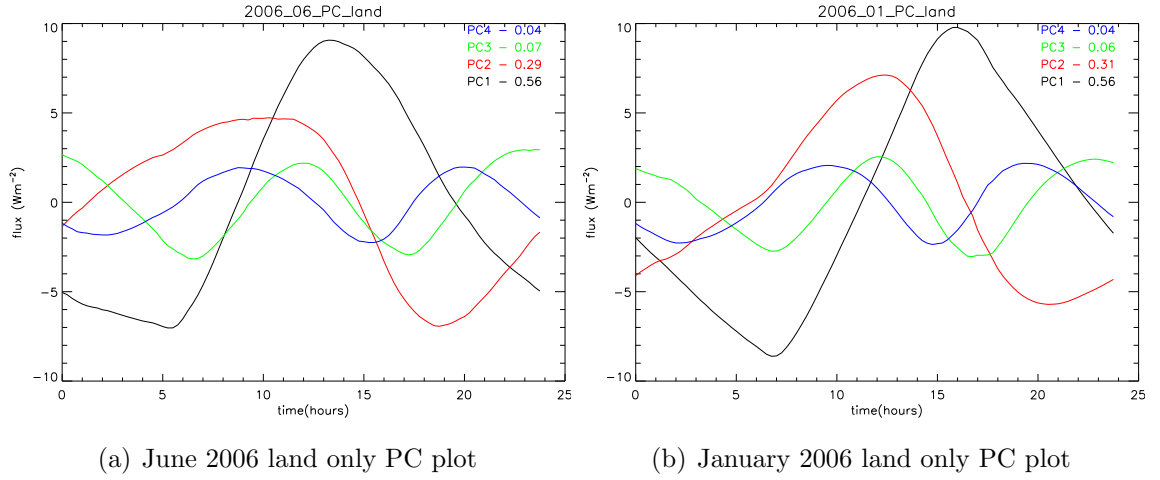


Figure 5.7.: First four PCs for land only a)2006 June b)2006 January where the fraction of variances are displayed at the top right corner of each plot. The different colours represent different order of PC.

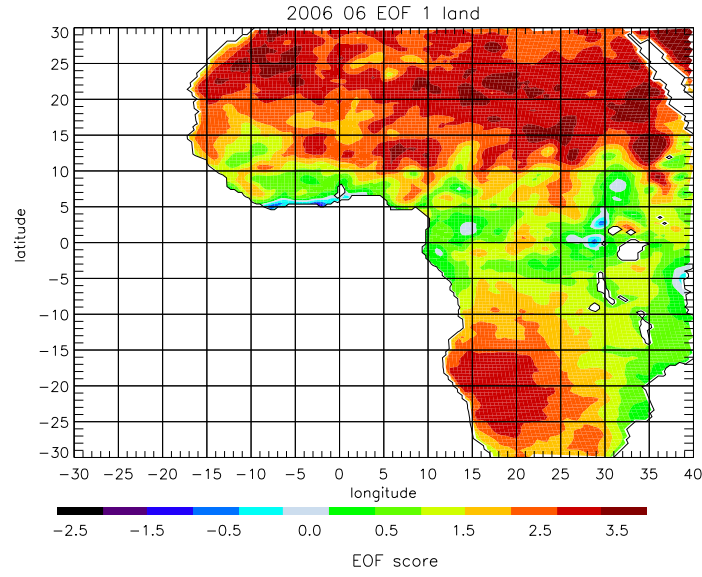
in all surface signal. This is due to the change in the input data, where the low variability ocean points have been taken out.

In June, the second PC has a magnitude of around $10Wm^{-2}$ which is similar to the PC2 all sky analysis, indicating this could be a cloud. However, the broader peak of the PC2 extending to around noon suggests that this PC could possibly contain some surface influence.

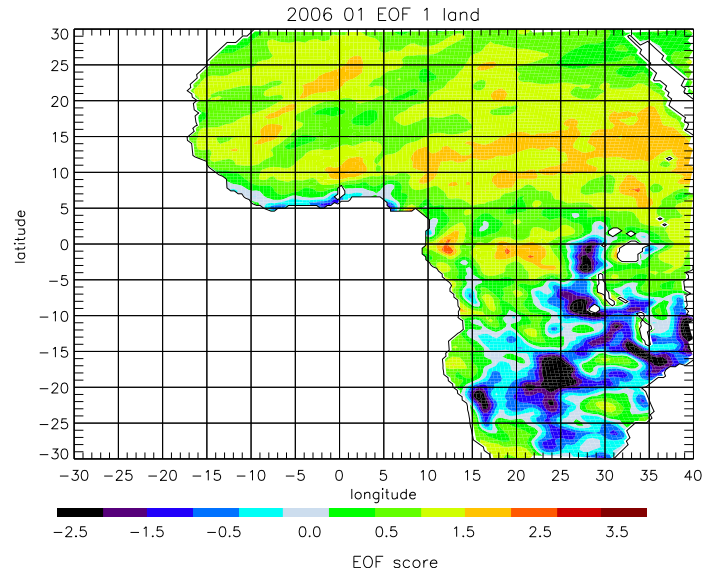
The trough of PC2 is at around 1900, whereas the peak is broader compared with all surfaces (Fig.5.4), from approximately 0800 to around 1200. This PC contributes to about 29% of the variance, much larger than all sky PC2. This could be due to lack of strong PC2 signals over the ocean, therefore, land only PC2 is capable of explaining much more variance than all sky. On the other hand, PC3&4 are similar to all surface, but they both contribute more to the overall variance.

In January 2006 (see Fig.5.7(b)), the first two PCs differ from June in both shape and peak position. The first PC peaks at around 1600 and the shape is similar to a sine wave which is roughly an inverted PC2 in the all surface analysis. Whereas the second PC peaks just after 1200 and resembles that of all surface PC1. This could be due to the ITCZ overlapping with a larger region of Africa land during this time of the year, thus causing the variance of each PC to be different for the two months. For example, the dominant component in this case is the cycle of cloud

development, which resembles the second PC of the all surface analysis.

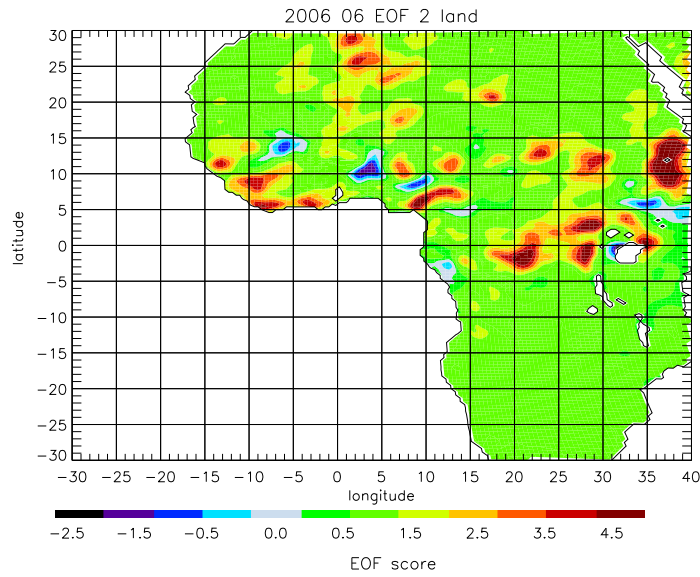


(a) June 2006 land only EOF1 plot

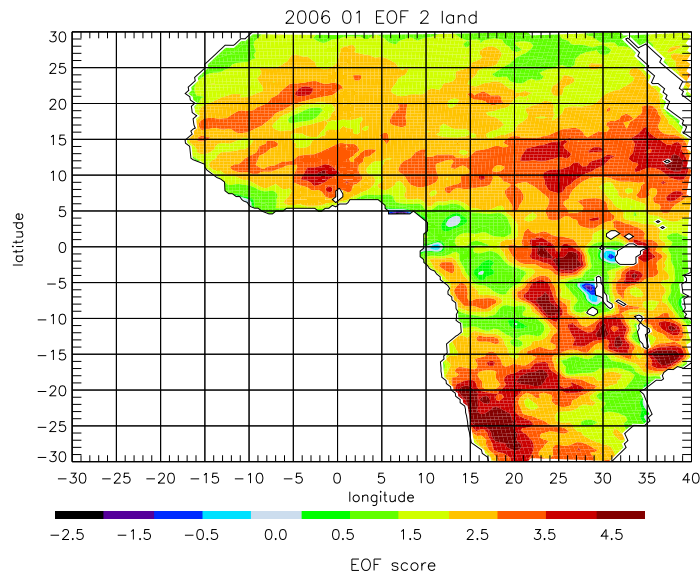


(b) January 2006 land only EOF1 plot

Figure 5.8.: Land EOF1 plots for a)2006 June b)2006 January, where the colours represent the EOF score over the region 30°S - 30°N and 30°W - 40°E , white indicates data absence. The x and y-axis denotes the longitude and latitude respectively. The EOF score is dimensionless, requires associated PC to interpret the OLR component.



(a) June 2006 land only EOF2 plot



(b) January 2006 land only EOF2 plot

Figure 5.9.: Land EOF2 plots for a)2006 January b)2006 June, where the colours represent the EOF score over the region 30°S-30°N and 30°W-40°E, white indicates data absence. The x and y-axis denotes the longitude and latitude respectively. The EOF score is dimensionless, requires associated PC to interpret the OLR component.

Fig.5.8 displays the geographical distribution of the associated first land EOF for both June, and January. The June EOF plot (see Fig.5.8(a)) shows similar geographical distribution to the all surface EOF1 plot, with some small negative signals around the coastal area which signifies the OLR has a minimum during the day and maximum in early morning. In these regions, the OLR signals are likely to be dominated by the advection of clouds into the area during day time. This is confirmed by a direct inspection of the data where a decrease in OLR is observed to shift from the coast to inland, though the movie is not shown in this thesis.

For January 2006, the EOF1 appeared slight different to the all surface plot, with the strongest positive signals over northern part and negative signals along the ITCZ. This could be due to the fact that the January land only PC1 now represents mostly inverted all surface PC2 signal (or inverted cloud development signal). This is confirmed by a strong negative signal over the ITCZ in southern part of Africa and small positive signals over most other parts. This indicates that the PC could contain both heating and clouds signal which is confirmed in Fig.5.8(b) where the EOF showed weak positive and strong negative responses over clear sky and cloudy sky regions respectively, which also implies cloud is the more dominant component in PC1.

By further looking at EOF2 (Fig.5.9), it shows how well the different PCs isolate heating and cloud signals for this particular scenario. Fig.5.9(a) shows the second EOF for June, where the plot exhibited similar features as the all surface EOF2. However, a slight positive signal is found over land in general which perhaps indicates PC2 under this scenario has been lightly "contaminated" by surface heat response signals.

Despite this small contribution from surface heat response signals, the strongest signal is still found to be along the ITCZ and overlapping with all sky EOF2 signals, which therefore, shows that the most dominant process in PC2 is cloud developments over land.

In January, Fig.5.9(b) shows a positive signal over almost all land, which is similar to EOF1 of all sky. The strongest signals are found in the south, where the minimum SZA is with some negative signals along the ITCZ region. Therefore, suggesting that PC2 is mostly dominated by surface heat signals.

It should be noted that although the land PC are roughly in phase with some of

the all sky analysis, the shapes and the exact timing of maximum OLR of the land PC are slightly different. This suggests that surface types is an important variable to how the PC express the OLR signals and will be investigated further in chapter 6.1.

5.2.3. Ocean only analysis

This is a subsection of the all sky analysis, which concentrates on the diurnal variabilities of the ocean. The fundamental variabilities will of course be different over the ocean due to the large heat capacity, the circulation beneath the surface as well as the marine stratocumulus clouds off the west coast of Africa.

The analysis will be similar to all surface analysis but limiting the data to ocean surface type only. The coastal ocean regions data have been removed to avoid the coastal effects dominating the results. The coastal signals can be influenced by the advection of cloud, it is sometimes the dominant factor in our ocean only analysis and inhibiting the visual distribution of the other signals (particular along the ITCZ), hence the coastal region was removed from our analysis.

The same averaging and analysis techniques will be applied on the data.

This part will show similar plots to the last section but limiting the dataset to ocean analysis only, the PCs in this case are expected to be quite different from both the all surface types and the land only PCs, due to a different response to the heating cycle.

Fig.5.10 shows the first four PCs for ocean only analysis in both June and January. For June, PC1 peaks at around 1500 and represents about 52% of the overall variance. The cause of this PC is not obvious since it could be due to cloud top heating, ocean heating and convective cloud events. However, by observing the associated EOF1 (see Fig.5.11) and the variance contribution, PC1 is likely to be dominated by a general ocean warming curve since positive EOF signals are found almost over all regions with the exception of the ITCZ.

The second PC was found to peak in the morning at 0800-0900 and its minimum is at 1900 with a variance of 29%. This PC is more likely to be caused by convective events over the ocean, since the associated EOF2 signal (see Fig.5.12) appears to focus over the ITCZ region. Both PC3&4 are found to have semi-diurnal periods

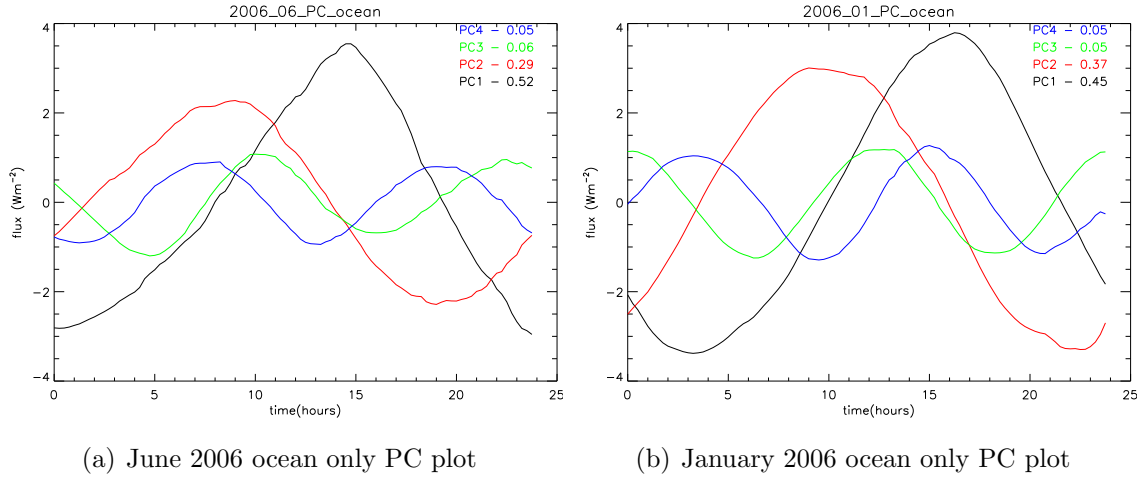


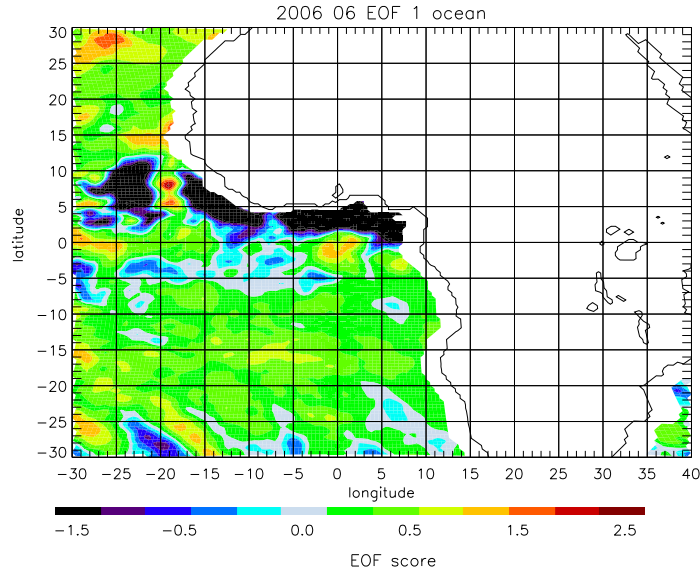
Figure 5.10.: First four PCs for ocean only a)2006 June b)2006 January numbers on top right corner of each plot indicate the fraction of variances for the individual components. The change in Outgoing Longwave in Wm^{-2} and local time in hours are represented by y-axis and x-axis in the plots respectively.

and contribute 5% to the overall variance. Similar results were obtained for June with small phase difference in the PCs, about 1 hour.

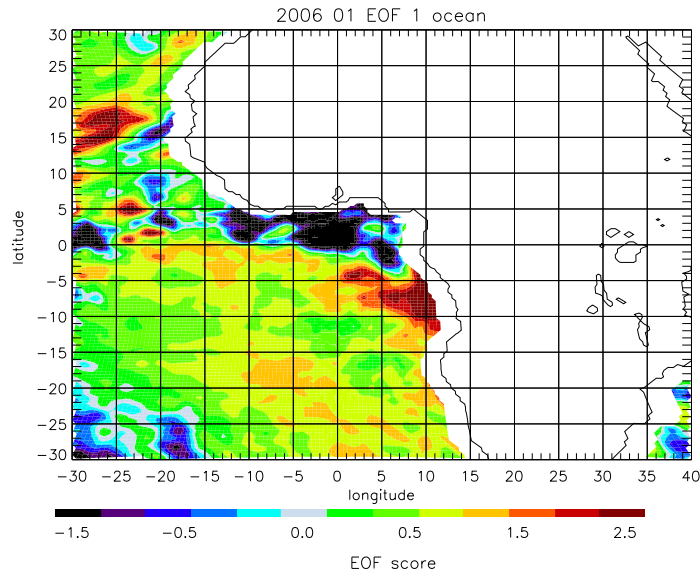
The stratocumulus can potentially influence some of these PC signals, with a diurnal cycle which peaks in cloud cover around 6am and minimises in the afternoon. It could be present in the first PC of January 2006 which peaks in the morning and minimises in the afternoon and the spatial distribution of which is shown in Fig. 5.11(b) where a strong positive signal is found in the south off the coast of western Africa. This is where marine stratocumulus are formed due to the upwelling of cold water in the coastal region.

In Fig.5.11, the first ocean only EOFs for January and June are illustrated, it should be noted that the coastal regions are removed to avoid domination in the signal by these coastal advection effects. The strongest negative signals were observed along the ITCZ, whereas strongest positive signals are present both along the ITCZ and in some tropical regions. In general, positive signals were found over the tropical ocean, indicating that this is a mixture of ocean and cloud heating signal, perhaps enhanced by cloud advection over some regions.

In contrast to the first EOF, EOF2 signals (see Fig.5.12) were only manifested

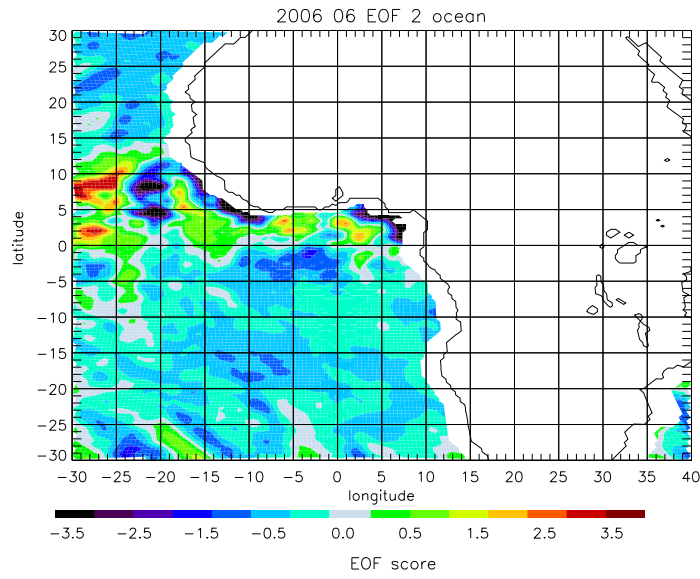


(a) June 2006 ocean only EOF1 plot

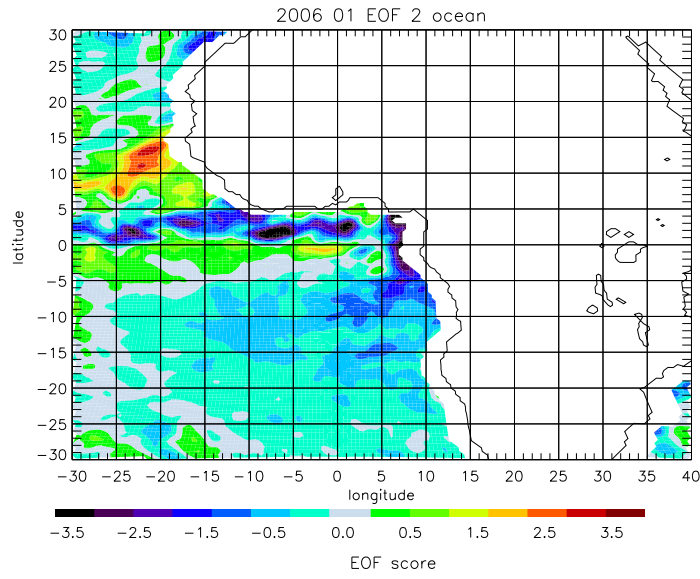


(b) January 2006 ocean only EOF1 plot

Figure 5.11.: ocean EOF1 plots for a)2006 January b)2006 June, where the colours represent the EOF score over the region 30°S - 30°N and 30°W - 40°E , white indicates data absence. The x and y-axis denotes the longitude and latitude respectively. The EOF score is dimensionless, requires associated PC to interpret the OLR component.



(a) June 2006 ocean only EOF2 plot



(b) January 2006 ocean only EOF2 plot

Figure 5.12.: ocean EOF2 plots for a)2006 January b)2006 June, where the colours represent the EOF score over the region 30°S-30°N and 30°W-40°E, white indicates data absence. The x and y-axis denotes the longitude and latitude respectively. The EOF score is dimensionless, requires associated PC to interpret the OLR component.

along the convergence region, and show 0 magnitude signal almost everywhere else. This concentration of signal in ITCZ with lack of gradual variation with respect to SZA implied that PC2 describes mostly convective cloud signals.

5.2.4. Summary

The diurnal cycle of the OLR has been studied via the decomposition of OLR signals using PCA. The initial analysis on the diurnal cycle of OLR confirms the validity of the PCA technique over a smaller region (Africa and Atlantic ocean) than [*Comer et al.*, 2007]. The findings are consistent with previous studies and has demonstrated the robustness of the technique on a subset of the data. The investigation was carried out using data from the GERB instrument for July and January 2006. The maxima of the PC1 signal were found to be at 1300 LST for both June and January. Whereas for PC2, the minimum, maximum are at 0600-0800 and 1700-1900 LST respectively in June compared with January 0600-0800 and 1600-1800 LST, suggesting perhaps a small degree of seasonal dependence on the averaged cloud cycle.

The land analysis has found a relatively larger seasonal dependence in terms of both fraction of variance explained by the PCs and the shape of the PCs. The variance is affected by the size of region which overlaps with the convergence zone. Whereas, the shapes of the PCs are influenced by the overall OLR diurnal cycles, which depend highly on the surface types. Therefore, the analysis will be extended into sub-categories of land data in section 6.1.

On the other hand, the ocean analysis shows slightly different PCs, where both the peaks and shapes are dissimilar to all sky analysis. This is largely due to the importance of the surface heat response being less dominant over ocean than land, and the large heat capacity of the ocean, creating different diurnal variations, where the first PC peaks in the late afternoon and second PC peaks in the morning.

This study is distinct from previous studies such as *Smith and Rutan* (2003) have used a low time resolution seasonal analysis which exhibits different components compared to our study. For example, *Smith and Rutan* (2003) did not find an asymmetry in the surface heating component which is noticed in [*Comer et al.*, 2007] and our study. In comparison with [*Comer et al.*, 2007], our study is regional in the tropics/subtropics and focuses on the diurnal cycles in Africa, which is heavily

influenced by convective clouds and African meteorology. For example, our land only analysis has highlighted an even more asymmetrical first component than observed in [Comer, 2007], in that a distinct minimum is observed in early morning which could be attributed to the cold temperature in the desert before sun rise.

5.3. Clear/cloudy sky

5.3.1. Clear sky analysis

To further understand the effect of clouds on the diurnal cycle of the OLR, it is essential to first analyse the cycle of clear sky flux before making comparison with cloudy/mixed data. Therefore, a cloud mask was used to produce the results shown in this section to filter out the cloudy sky data. This mask was produced using the scene ID from the GERB SW data files, which gives the cloud cover and cloud optical depth. For the purpose of including more data points in this analysis, the threshold of the cloud cover amount was relaxed to greater than 20% and the analysis is limited to land only, as previous analyses have shown that ocean possesses a different diurnal cycle to land.

Averaged diurnal flux comparison

As an initial test, the 'accepted' data were averaged into a single diurnal cycle to observe the diurnal variability on an average basis. (NB the averaged data is only used in this subsection, non-averaged data was used in Clear/cloudy sky decomposition.

It can be seen in Fig.5.13(a) that the averaged clear sky flux cycle looks very similar to the first order PC (Fig.6.1(a)) of the surface type analysis, supporting the fact that the first order PC closely resembles a surface heat reponse. In terms of peak timing of the curve, it's the closest to the curve from all land analysis (see figure5.1(a)), which is reasonable as the clear sky data would consist of different surface type. However, compared with the first order PC, the clear sky flux curve demonstrates a greater assymetry about the peak(approximately at 1200). This can be perhaps be attributed to the lack of clouds, thus providing less shielding for the OLR during night time, hence triggering this asymmetry.

Comparing figure5.13(a) with figure5.13(b)(same plot but without cloud filter),

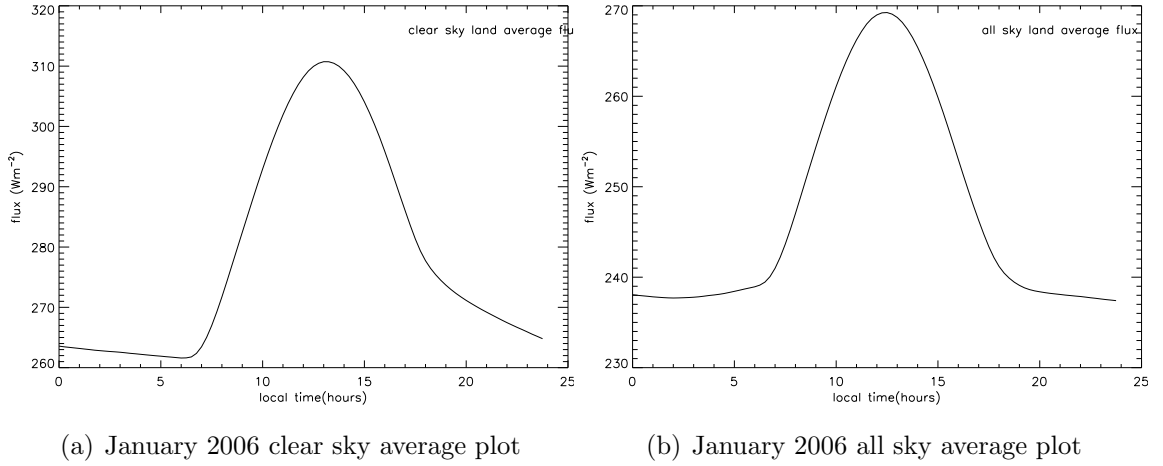


Figure 5.13.: Diurnal cycle of clear sky land average flux and all sky land average flux (Wm^{-2}) for January 2006

it is obvious that the peak of the all sky average flux occurs at an earlier time than clear sky average, which can be explained by the lack of convective events in the clear sky data. It is believed that these convective events would shift the the overall peak of OLR diurnal cycle to an earlier time, as shown in Fig.5.1(b). The PC2 in the average analysis resembles cloud development cycle which maximises in the morning and minimises in the evening, so if the contribution from this PC increases, the overall peak of the OLR cycle will be shifted earlier to the morning or around noon.

Clear sky decomposition

The clear sky decomposition employs data filtered by the cloud mask and, due to a lack of night time cloud information it is assumed a clear sky day yields clear sky night. The particular day of data for a pixel is employed only if the whole day is clear.

The remaining data are arranged as individual days of data and the data is not averaged as most of the points are not completely cloud free during the month. The data will be divided into daily data and each day of data will be treated as an extra dataset in the spatial dimension. Therefore, the number of data which enter the analysis method will be roughly 30 times the averaged analysis. Although the PC

have the same dimensions as the averaged analysis, the associated EOF will have equal spatial dimension as the input data.

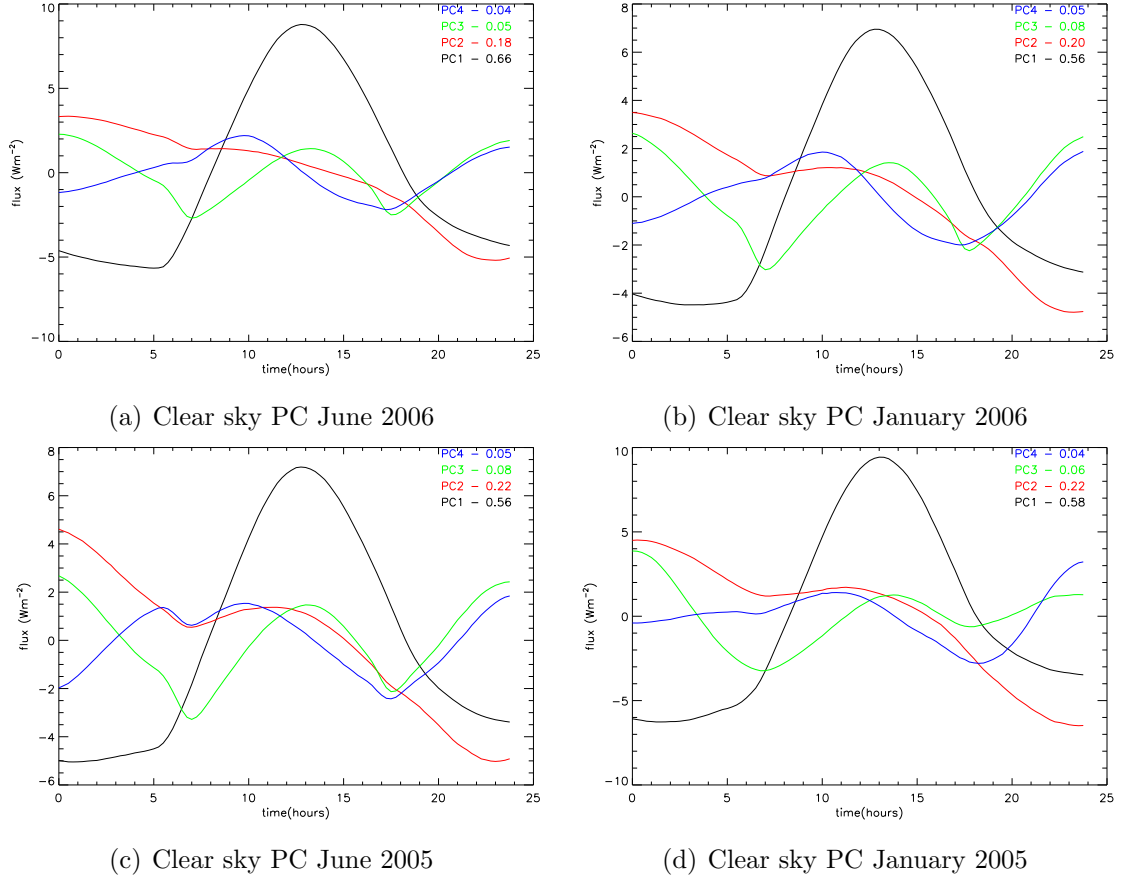


Figure 5.14.: Clear sky principal components for a) June 2006 b) January 2006 c) June 2005 d) January 2005, numbers on top right corner of each plot indicate the fraction of variances for the individual components. The average Outgoing Longwave in Wm^{-2} and local time in hours are represented by y-axis and x-axis in the plots respectively.

The first PC in Fig.5.14(a) was observed to resemble the shape of averaged all sky PC1 and also similar to PC2 of the non-averaged data, which was believed to be caused by surface and clear sky response to incoming solar radiation. Since this analysis itself is a non-averaged analysis, it would be more appropriate to draw comparison with the non-averaged all sky results. The PC1 of Fig.5.14(a) contributes to much more variance (66%) compared to the all sky non-averaged PC2 (33%) (see appendix A.2 for detailed non averaged analysis) where they both have similar

shapes and contains most of the heating signals. This is due to the fact that a major component has been filtered out in this analysis, i.e. the clouds. Hence, the variance contributions by the different PCs have changed.

On the other hand, the clear sky PC2 is quite different from the non-averaged PC1 (see appendix A.2) which consisted mostly of cloud signals. Ideally, we would expect to observe this PC2 signal disappearing altogether and be replaced by PC3. However, in Fig. 5.14(a), we see an almost linearly decreasing line and a discontinuity in beginning and end of the PC2 signal which cannot be a real physical signal on its own, this is most likely to be an artifact of removing a major component from the data set. The reason is that by removing the cloudy days, we have introduced a discontinuity on the points which are intermittently cloudy. For example, if we have an OLR time series for a month, and the second day is determined as cloudy and removed from the dataset. Now there presents a discontinuity between day 1 and 3; even though they are cloud free days, they could begin the diurnal cycle with a different level of flux given the difference in atmospheric, surface conditions preceding and following the cloud event. Therefore, imposing a discontinuity between the beginning and end of a diurnal cycle, and when correlated with another day/point of data under the same influence, the PCA could introduce this discontinuity as a signal by drawing a covariance between these data points.

In Fig. 5.14, we have shown the clear sky PCs for January, June 2005/06, they appear to be consistent in terms of their shapes (especially PC1 & 2) and their variance contribution in each of the PC. This suggests the signals we observe are invariant across the seasons and repeatable year to year, indicating that these signals are not noise induced.

Furthermore, PC4 which was semi-diurnal timescale in the previous analysis, is altered, supporting the theory that it is due to clouds. Whereas, PC3 retains its semi-diurnal pattern, with roughly the same peaks as non averaged PC4, suggesting that it is an extra component which is required along with PC1 to represent the clear sky surface heat curve.

The discontinuity on PC2 described above does not have any bearing on the lack of night time cloud data or the quality of the cloud mask and is simply a reflection on the initial condition of the day. The only method to combat the problem is to eliminate the whole time series when a cloudy signal is found at any time, which

would cause the number of available data points to drop below a useful number. Instead, we have chosen to do this analysis with this knowledge and take into the account of this effect when we analyse the patterns. For example, in Fig. 5.14(a), this discontinuity has shown up as a PC which is related to any physical process.

By extension, this effect of discontinuity would be present in all datasets in all PCA, because, for example, if we have a dataset with N number of points over a D number of days. Then, by separating all the diurnal cycles, we have $N \times D$ number of discontinuities present in $N \times D$ number of diurnal cycles. Hence, we would expect this to be observable in a non-averaged datasets which is confirmed in a non-averaged analysis (see appendix A.2).

The reason that we do not observe this effect of discontinuity in our averaged data analysis is because by averaging all days of data for a point into a single diurnal cycle, the offset between two consecutive days are neutralised. Thus, by averaging a month of data, we are essentially looking at the discontinuity between beginning of day 1 and end of day 30 divided by a factor 30 in the average, which is expected to be very small even if we assume a modest discontinuity of $100Wm^{-2}$ between day 1 and 30.

5.3.2. Cloudy sky analysis

As part of cloud cycle study, it's essential to look at both clear sky and cloudy sky to observe the components which each analysis emphasizes, and the differences compared with an all sky analysis. This section uses the same non averaged data arrangement described in section 5.3.1.

The data selection method in this analysis is slightly more relaxed than the clear sky analysis where if a pixel is classified to be cloudy for more than one third of a day, then that particular day of data is included in the analysis. The predominant clear sky data have been filtered out to highlight the components within the cloudy sky data.

The cloudy sky analysis is not intended to exclude all clear sky data, it is implemented to filter the pure clear sky (i.e. clear sky in the whole diurnal cycle) in an attempt to highlight the importance of cloud in our analysis. Therefore, the use of all sky night time is appropriate given the corresponding day time points present

a certain degree of cloudiness(1/3 of the day). The SW cloud retrievals are less reliable towards dawn and dusk due to the large atmospheric path length observed by the GERB instrument, but this effect is minimised in our study since the cloudy sky are determined upon our criteria (1/3 of the day cloudy), which eliminates any uncertainty arises from dawn, dusk measurements.

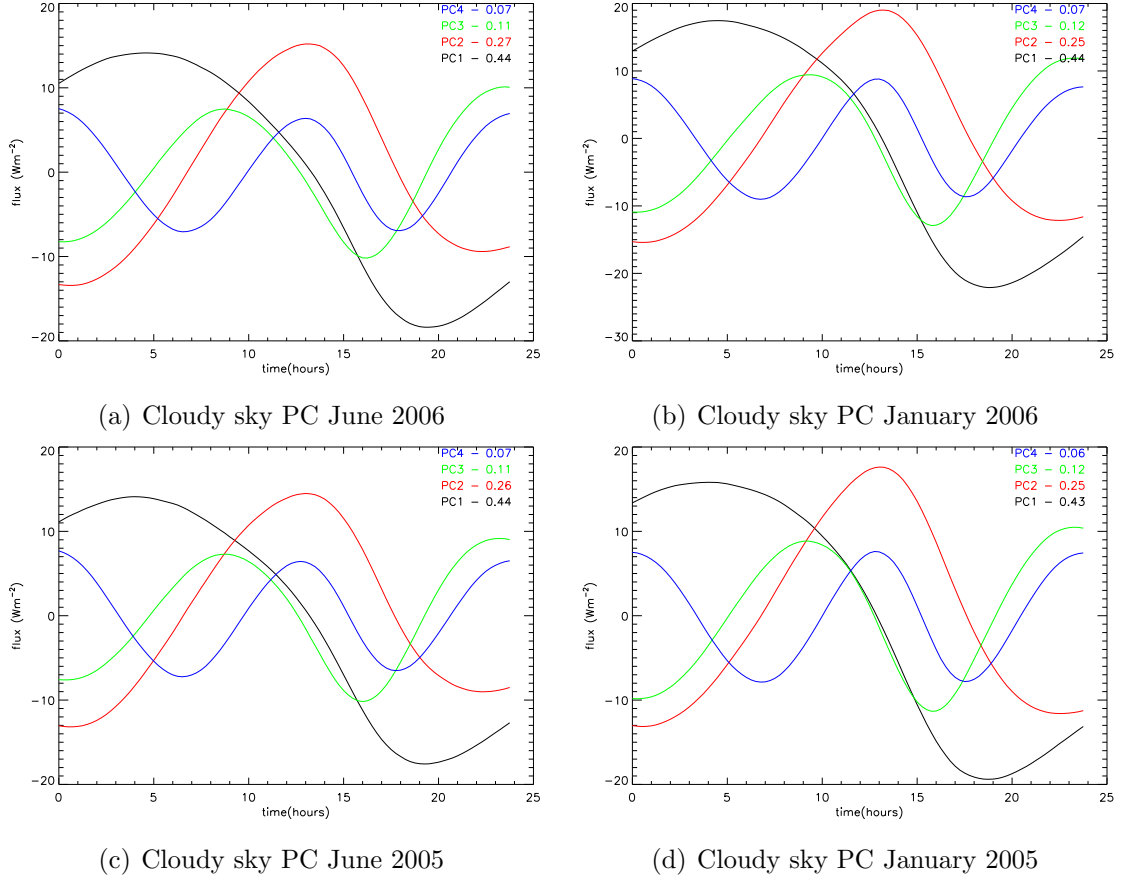


Figure 5.15.: Cloudy sky principal components for a)June 2006 b)January 2006 c)June 2005 d)January 2005, numbers on top right corner of each plot indicate the fraction of variances for the individual components. The average Outgoing Longwave in Wm^{-2} and local time in hours are represented by y-axis and x-axis in the plots respectively.

In Fig.5.15(a), the PCs are similar to the non-averaged plots (see Fig.A.1 in appendix), and the time of the peaks are roughly the same. However, the fraction of variances corresponding to each PC are different, largely due to the clear sky data being filtered in this analysis. Therefore, the first PC which was predominantly

cloud induced has gone up in variance from 42% to 44%. Conversely, the second PC that was influenced mostly by surface heating, has decreased compared with the non-averaged analysis from 33% to 26%, which was expected since this analysis excluded the clear sky data (cloudy for less than a third of the time).

Note that the first PC presents a discontinuity between the beginning and end of the diurnal cycle, this shares the mechanism as discussed in the clear sky analysis (see section. 5.3.1). PC1 clearly has both cloud and discontinuity influences, this is what gives this signal such high magnitude and contribution in variance. The high magnitude arises from the background discontinuity which can be observed in PC2 of 5.14(a) and the high variance is caused by the fact that this signal is representing two processes (clouds and discontinuity). When we take into account the discontinuity of this PC1 signal, we estimate a $10\text{-}15 W m^{-2}$ diurnal variability of PC1 contributed by clouds.

In addition, PC3&4 have retained their time of peak and semi-diurnal nature; the fraction of variances in each of these components have increased by a small amount compared to all sky.

Again, we have shown in Fig. 5.15 that the diurnal variability in these datasets are extremely consistent for the four months that we plotted.

5.3.3. Summary

This section of work analysed the clear sky and cloud sky diurnal cycles, by using a cloud filter derived from the GERB SW cloud data. The results have indicated which components are invariant and which change under different scenarios.

It was shown in the results that what we interpreted as the surface heat response remains approximately invariant in both cases, along with the phase shift component (see PC1 and 3 in section 5.2). Whereas the cloud dominated PCs (PC2&4) have altered almost completely from their all sky form. However, these PCs have not been replaced by a lower order PC, suggesting an imperfect cloud mask has been employed, leaving some residue cloud components. Nevertheless, it has demonstrated the sensitivity of the analysis technique in response to the absence of an atmospheric component and supported the hypothesis of the PCs that certain PCs are dominated by particular atmospheric processes.

We have shown that by removing certain scenes from the diurnal cycles, that certain components are robust across the scenes (i.e. surface heat and its semi-diurnal modulation in either PC3/4) and some components can be eliminated almost entirely; for example, the cloud induced PC2 in the all sky analysis is not observed from our clear sky study. We have also determined that these signals are not noise induced by demonstrating the seasonal and year to year differences in the these cycles to be small.

5.4. Chapter Summary

This chapter has investigated the outgoing longwave radiation on diurnal timescales. A number of different variables were used to divide data into subsets for the analysis. By selecting the data with these variables allowed us to explore areas which were not covered in previous similar studies (*Comer et al.*, 2007; *Smith and Rutan*, 2003). We were able to observe how these different components within the OLR changes diurnally when the analysis is focused on particular group of data.

We began our investigation by performing the PCA on the all sky all surface averaged data across Africa and Atlantic region. We have found similar results to global/near global analysis (*Smith and Rutan*, 2003, *Comer et al.*, 2007), confirming the validity and the robustness of of the technique over a smaller region. We were able to observe some of variabilities and the most dominant components over the chosen region. For example, the surface and clear sky emission in response to ISR, and the cloud development cycles. We have also observed seasonal variabilities particularly in the spatial distribution of the cloud component and how it changes according to the ITCZ.

Another variable that was considered in this chapter was clouds, for which the PCs were observed under the two scenarios. The results indicate that the analysis is fairly successful in isolating surface heat response and clouds effects. It was also shown that certain components are sensitive or invariant to different scenarios and that we were able to pick up the discontinuity in the signals introduced by the cloud mask.

This chapter has demonstrated the variability in the diurnal cycle of the OLR and shown different components of the OLR with respect to different sky scenarios. The

PCA is proven to be useful in analysing these variabilities temporally and spatially, especially in identifying the differences in the PC peaks on small timescales.

In our all sky analysis, we have observed similar PC1& 2 to [Comer *et al.*, 2007] but over a smaller region, Africa and the Atlantic. We have attributed the first and second PC/EOF to be predominantly influenced by surface heat response signal and by cloud respectively. This is consistent with the results presented in [Comer *et al.*, 2007], which implies the diurnal variability outside of our chosen region are similar to what we are seeing over the Africa and the Atlantic.

PC1/EOF1 together have shown a diurnal variability of up to $60Wm^{-2}$ which is similar to the variation in desert time series that we have looked at in 4.2. Furthermore, the timing of maximum OLR of the PC is consistent with the peak in brightness over land presented in [Yang and Slingo, 2001], suggesting the PC is a good description of the surface heat response. Our PC2/EOF2 with PC1/EOF1 gives a combined magnitude (around $100Wm^{-2}$) which is consistent with the OLR changes in the cloudy region. The PC2 presents peak and trough times which are comparable with the cloud maximum and minimum in [Chung *et al.*, 2007]. This implies this PC2 signal is predominantly influenced by clouds.

Comer (2007) has seen a 90% variance represented by the first two PCs in July but only described about 85% in January, where the rest of variability is described by the higher frequency components. However, we have found the first two PCs to contribute, in total, 90% of the variances in both June and January 2006 in our analysis. This suggests that by removing part of the data, we have removed some of the diurnal variability but we were still able to produce similar PCs which describes most of the variance across this chosen region. It also suggests that the main diurnal variability (PC1 & PC2) in this region is quite similar to the region outside.

Moreover, the PCs 3 and 4 found in [Comer, 2007] are similar to our study, though the spatial EOFs are slightly different compared with our study. Comer (2007) has found both of these components to be only significant over 3 regions in January: Madagascar, Southern Africa and Brazil. Comer (2007) has attributed these components to the persistent data gap in Madagascar during the January 2005 in the morning (0700-0800) and clouds moving slowly away their sources in Southern Africa and Brazil. However, in our study, we have found both third and fourth components to be linked with the whole ITCZ and the third component to

exhibit some surface influences in Southern Africa during June and Northern Africa during January. This suggests the third component is influenced by both clouds and surface, but the fourth component appears to be purely an adjustment of the cloud OLR cycle particularly over the ITCZ region. The PC3/EOF3 and PC4/EOF4 signals to give a diurnal variability of $10\text{-}15\text{Wm}^{-2}$ each in the ITCZ region.

In extension to [Comer, 2007], we have further analysed the components over clear/cloudy scene types using a non-averaged dataset. In our clear sky analysis, PC1 and PC3 appears to be similar to the all sky studies but we have observed a big influence in PC2 which confirms the PC2 in the all sky analysis is mainly due to cloud signals. In the clear sky PC2, we have captured the discontinuity in the signal introduced by the cloud mask (see section. 5.3.1 for discussion). Furthermore, we have also seen an influence in the PC4, that confirms the influence of clouds on all sky PC4.

In our land and ocean separation studies, we have observed components which are more localised and have more influences from clouds and strong heat response of the Sahara compared with Comer (2007). We have therefore, decided to further separate the signals in accordance to their surface types and regions (i.e.Sahel) to investigate some of the more dominant signals and observe their variability with respect to surface properties, ISR and SZA influences.

6. Surface Type Analysis and Model Comparison

6.1. Analyse separated by surface type

6.1.1. Introduction

In the previous studies shown in chapter 5, we have examined the components within the diurnal variability across the different sky scenes as well as over land and ocean. We have established that the presence of surface heat response component is fairly robust across all scene types but there are minor differences in the shape of the PC when we separated the data (i.e. months or clear sky/cloudy analysis). Since this component appears to be the most significant in the diurnal variability, we will investigate some of the factors which could give rise to these changes.

The heat capacity and emissivity are the most dominant characteristics which determine the cooling of a particular material via electromagnetic radiation. Therefore, it is expected that the diurnal cycle of OLR for different surface types would be different due to their respective chemical compositions.

The first variable examined is the OLR diurnal variation of different surface types. This is done by employing the fixed surface type dataset obtained from the GERB scene ID in the shortwave file. The surface types used in the study are from GERB surface data, and is a static map of 5 surface types (Ocean, dark vegetation, bright vegetation, dark desert, bright desert). During the seasonal variation, this could potentially wrongly categorized surface type data in the transitional savannah, which is mostly along the Sahel in our chosen region. This is expected to include some extra cloud influences from the Sahel in our desert analysis during the boreal summer where the ITCZ is in the north. However, the effect of clouds is expected to be small

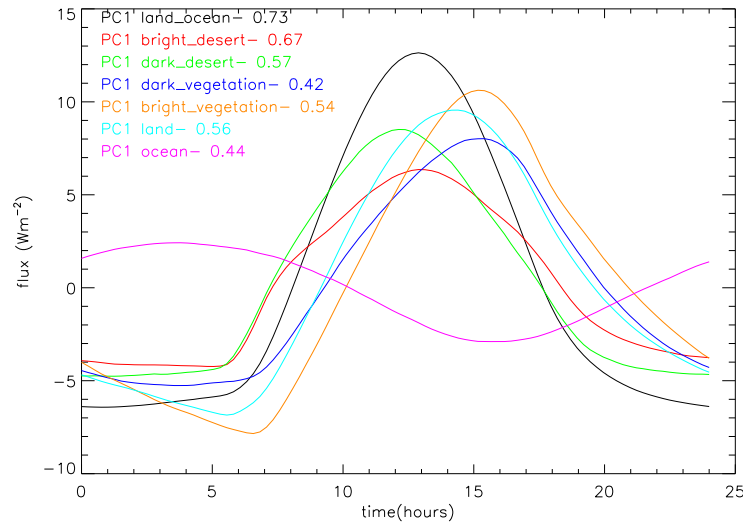
compared to the diurnal cycle of strong surface heating in the Sahara and should show up in the desert EOF2.

The OLR data used in this Chapter is arranged according to local time and separated by surface types using the data in the GERB SW scene identification. Monthly averaged diurnal cycles are produced for each point for the months: December, January, June and July over the years 2004 to 2007 inclusive. A separate PC decomposition is then carried out on these data for each surface type and the results are displayed in Fig. 6.1. Maps of the corresponding EOFs are then displayed for each month in the rest of section 6.1. From the results of the subsection, we have hypothesized the surface heat capacity to be one of the crucial components in the diurnal OLR variability which will be further investigated in section 6.2 using our model.

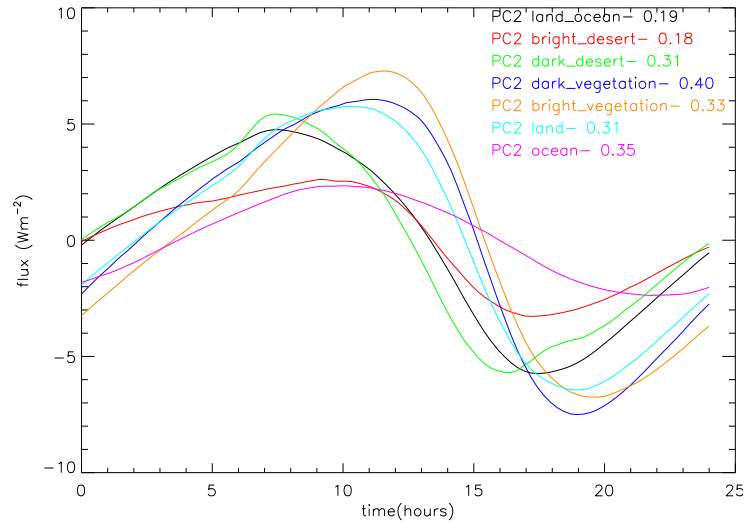
6.1.2. Bright desert

To analyse the EOF map, it is essential to consider the PC together with the associated EOFs to understand the signals. Hence, by first looking at the PC derived from the data (see Fig. 6.1(a), note that the PCs are derived from all the months mentioned above, whereas the EOFs are the corresponding spatial distributions for the particular month), it can be observed that all the PC 1 curves follow a general surface heating pattern caused by the ISR, where the peak of the curve is around 1200. Each curve peaks at a slightly different time across the different surfaces. Following from this crucial result, it is proposed in this chapter that the change in surface properties is one of major contributing factors which could influence the maximum time of OLR. For example, a surface type with a higher heat capacity tends to reach maximum later and vice versa. More specifically, the bright desert curve in Fig. 6.1(a) peaks just after noon at 1300 LST, the contribution of the first order PC to the total variance is about 67%.

Fig. 6.2(a) reveals a geographic distribution (EOF) of the PC1 signal. The EOF score is 2.5-4.5 almost everywhere in this region, which corresponds to a contribution of roughly $30\text{-}50\text{Wm}^{-2}$ in the diurnal OLR variation by the particular component. This is consistent with the OLR diurnal variation that we have observed over the clear sky desert in our points analysis in chapter 4 which suggests that this com-



(a) Different surface types all sky PC1 plot



(b) Different surface types all sky PC2 plot

Figure 6.1.: a)PC1 b)PC2 for different surface types: land, ocean, bright desert, bright vegetation, land + ocean. Numbers on top right corner of each plot indicate the fraction of variances for the individual components. The change in Outgoing Longwave in Wm^{-2} and local time in hours are represented by y-axis and x-axis in the plots respectively. Different colours represent the PC for the various surface types. months [01, 06, 07, 12] between 2004 and 2007 are all used to compute this PCs

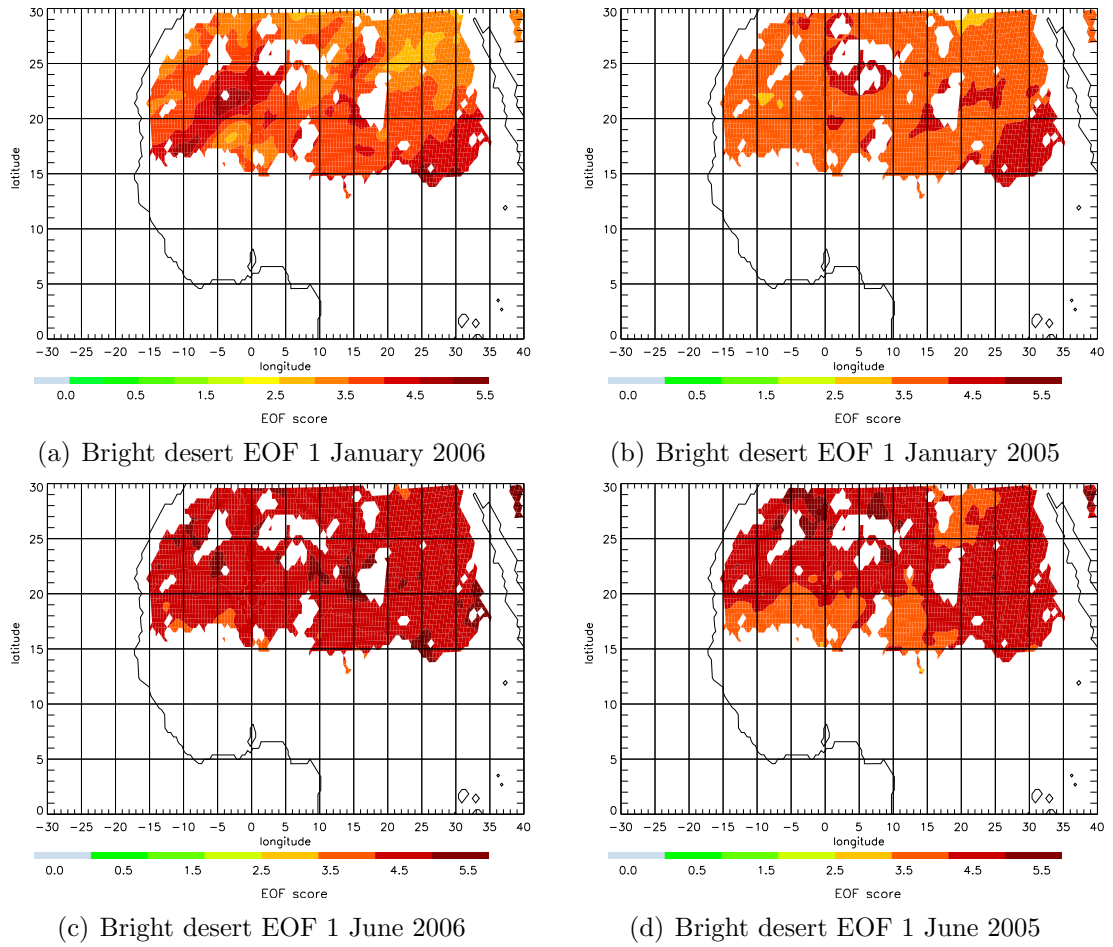


Figure 6.2.: a)January 2006 b)January 2005 a)June 2006 b)June 2005 Map of bright desert EOF1 from PCA over latitude and longitude range of 30°S-30°N and 30°W-50°E respectively, white represents an absence of data, NA indicates the part of colour scale which is not applicable to the diagram

ponent is capable of explaining most of variability in the clear sky surface heat response.

In general, the signal is positive almost in all regions which is as anticipated, since during boreal winter, the ITCZ is expected to be situated in a lower latitude of the African land. This implies the northern part of Africa, particularly over the desert, is predominantly dry, hence, we would expect to see a strong clear sky surface heat response. This is observed and confirmed in the EOF1 (see Fig. 6.2) for both 2005 and 2006 (and 2007 but not shown in figure).

In comparison to the January EOF1, the June EOF1 Fig. 6.2(c) & 6.2(d) has shown a larger EOF score which corresponds to about $40\text{-}60 Wm^{-2}$, this is due to the minimum SZA over this region during boreal summer. This increase the ISR, hence, increases the diurnal cycle of the OLR.

To further understand the diurnal cycle of OLR, PC2 for bright desert was also considered. In Fig. 6.1(b), the second order component for land and ocean is inversely related to the cloud amount in general. For example, cloud amount minimises in early morning and peaks in the late afternoon/evening as suggested in [Chung *et al.*, 2007]. This is confirmed in section 5.3 where the second component is predominantly influenced by clouds. It was observed that the second PC for the bright desert analysis contributes to about 18% of the overall variance, which has the lowest contribution out of all surface types, this is reasonable as this particular surface type is mostly distributed in/near the Sahara which has less moisture in the land for evaporation during the day due to the characteristic of the surface. Thus, it is expected that the cloud contribution to the diurnal OLR in this region to be small.

In the EOF2 plot for the desert type in January (Fig. 6.3), there exist signals in the form of stripe pattern across the desert. However, these appear to be inconsistent across the years as they show up as mostly positive signals in 2006 (Fig. 6.3(a)) and negative in 2005 (Fig. 6.3(b)). The EOF scores on these plots appear to be between -2 to 2 which corresponds to only a few Wm^{-2} , as we shall show in the following section that this is due to intermittent signals averaged over the month (see section 6.1.2).

In 2005 (Fig. 6.3(b)), although we do not observe the positive stripes of signal that we have seen in Fig. 6.3(a), we see some negative stripes signal with EOF score of -1.5 in 2005 compared with 1.5 observed in 2006 (less than $10 Wm^{-2}$ in both cases). This indicates the intermittent nature of the advection and the inconsistency in their advection time during the day. Thus, when they are averaged over months, these can give a low positive or negative overall signal, or even no signal if opposite signals are averaged.

The negative signifies where there is a 'opposite' convective cloud signal, i.e. minimum flux in the morning and peak in the evening. This could be for several reasons. For example, cloud advection onto these observed pixels in the morning; or early

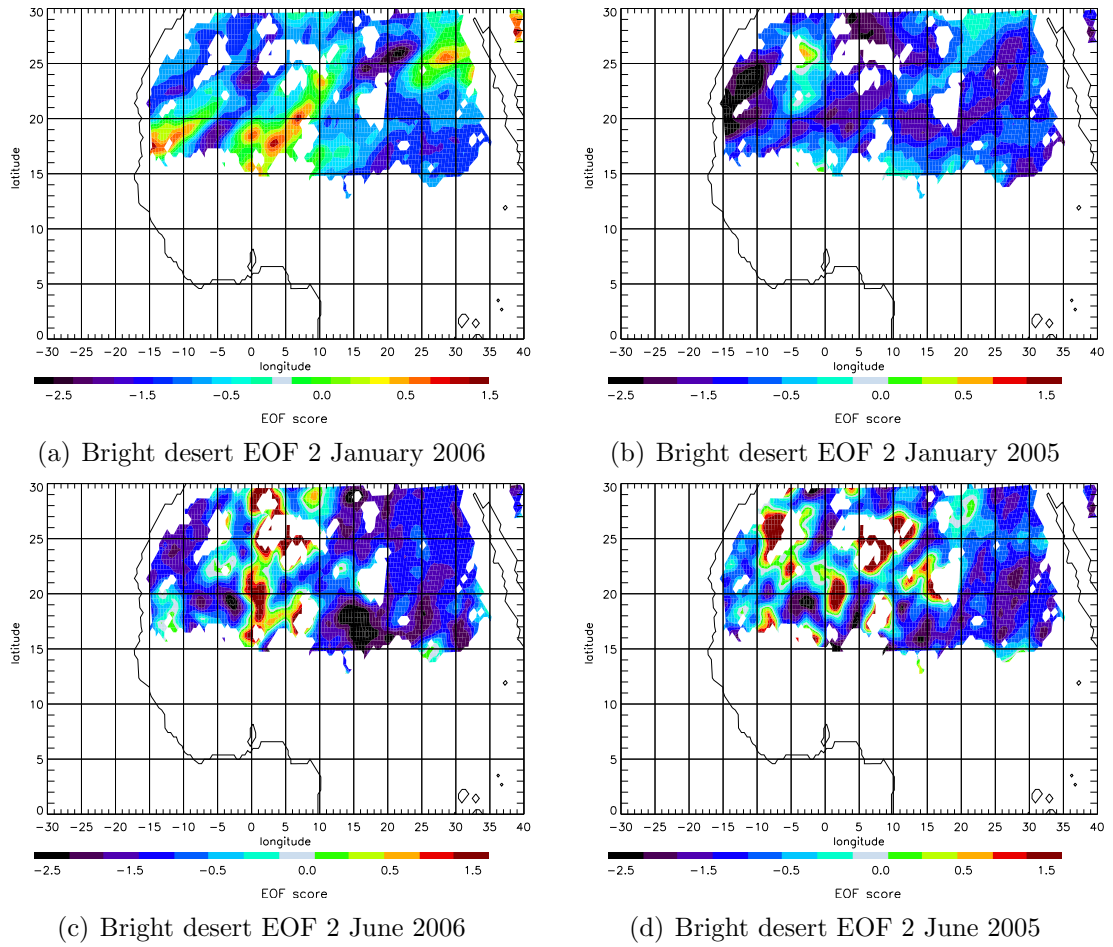


Figure 6.3.: a)January 2006 b)January 2005 a)June 2006 b)June 2005 Map of bright desert EOF2 from PCA over latitude and longitude range of 30°S-30°N and 30°W-50°E respectively, white represents an absence of data, NA indicates the part of colour scale which is not applicable to the diagram

cloud formation and advecting away from the source or dissipating later in the day. Whereas, the positive signals imply a cycle of convective cloud formation or advection in phase with this cycle, but as mentioned, these signals are not particularly strong, and could possibly be due to an averaging of the data, therefore, weakening the signals which are not consistent over particular locations. These intermittent signals will be investigated in the next subsection.

Fig. 6.3(c) & 6.3(d) show the EOF2 scores for June 2005 and 2006 respectively. The magnitude of EOF scores observed in the region are slightly higher

than January, however, the corresponding flux magnitude is still very small, less than $10Wm^{-2}$. The patterns of the EOF2 appear to be more sporadic as opposed to the stripe pattern observed in January, suggesting that these are mostly likely to be convective cloud cycles rather than pure advection of clouds. Considering the magnitude of these cloud signals, they are very likely to be just random cloud convective signals averaged over the month.

Advection over the Desert

In order to understand the stripe pattern that occurred in Fig. 6.3(a), we first need to understand the atmospheric dynamics in Africa, particularly over the Sahel and Sahara.

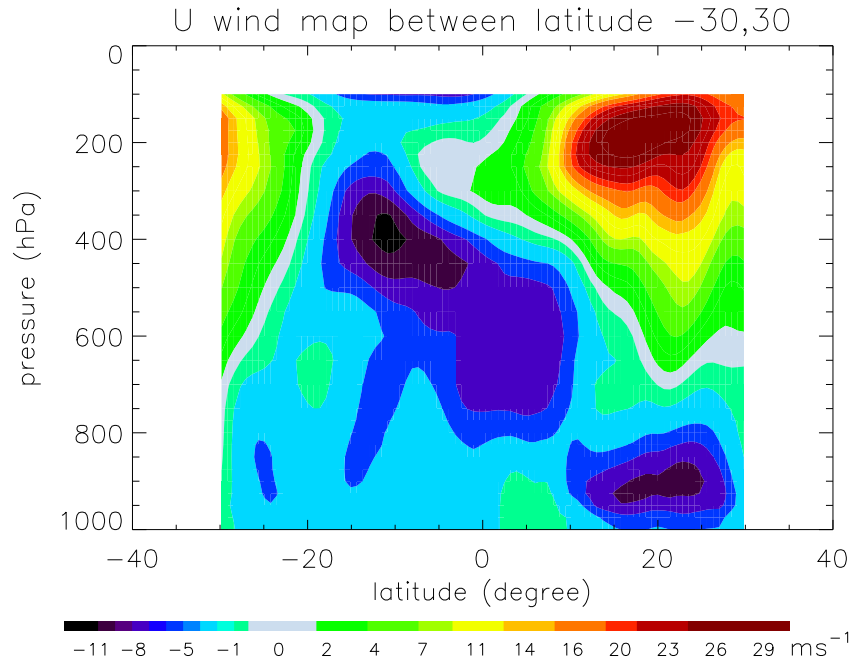


Figure 6.4.: Westerly wind map over latitude of 30°S to 30 °N averaged over longitude range between 20°E and 20°W (January 2006) from ERA data

Fig. 6.4 demonstrates the U (westerly wind) profile of the atmosphere between the latitude range of -30° to 30 ° averaged over longitude of -20° to 20° during boreal winter. In the figure, an easterly wind component can be observed at 5°N centered

at roughly 600hPa. This is formed by the temperature gradient between the Gulf of Guinea and the land, thus, causing the geostrophic wind to move from south to north and converges above the desert. Hence, creating an easterly wind at 5° .

There is another easterly component which is situated higher at approximately 400hPa which represents the equatorial easterly jet stream caused by the convergence of trade wind. The easterly wind features observed near the surface is where the trade wind is, whereas the westerly wind regions observed on both sides of the equatorial jets centered at 200hPa extending to below 400hPa are associated with the divergence of the easterly jet which inherits their westerly component from the Coriolis effect as they move towards the sub-tropics. We believe these westerly components are responsible for the advection over the desert.

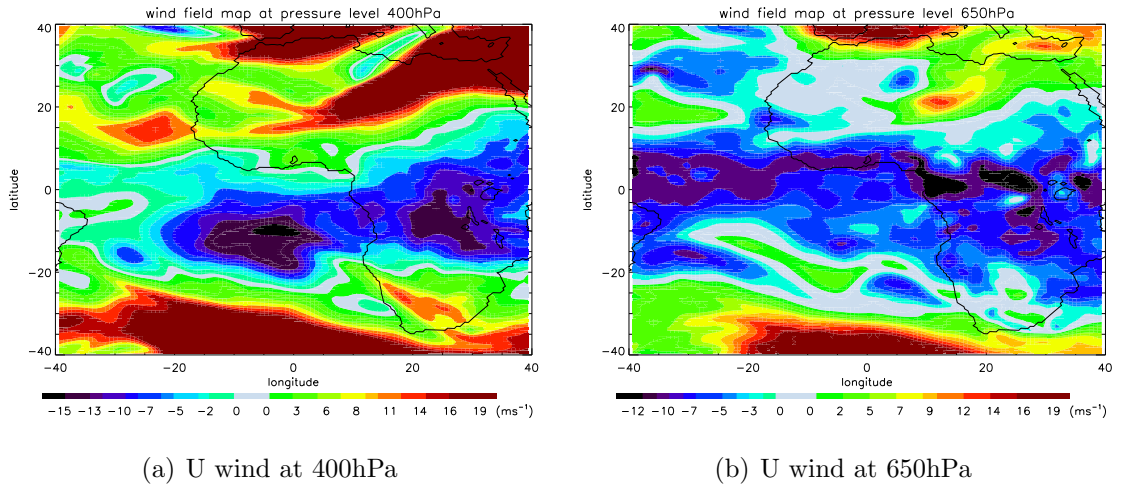


Figure 6.5.: Westerly (U) wind at pressure level a)400hPa b)650hPa for January 2006 from ERA data

Fig. 6.5(a) and 6.5(b) show two cross sections of the map in Fig. 6.4 at 400hPa and 650hPa respectively. In both figures, the easterly components in central/southern Africa represent the convergence zone over land; whereas, in Fig. 6.5(b), the easterly component over the the ocean is caused by the sea/land temperature difference which creates an easterly jet north of the equator that extends into the Atlantic and to the Americas, this forms the part of the northern hemisphere bias in ITCZ (see discussion in section 2.1.2).

In Fig. 6.5(a), there exists a westerly component over the desert (not present at

650hPa) which is caused by the divergence of air at the top of the convergence zone. This is believed to be the cause of the cloud advection which triggers the stripe signals that we see in Fig. 6.3(a) over the desert.

The magnitude of the signal that we see in Fig. 6.3(a) combined with the corresponding PC is equivalent to only less than 10 Wm^{-2} which is significantly lower than most cloud signals that we have seen in our analysis. This is due to both the intermittent nature of cloud formation and the inconsistency in the time, location of the advection.

We have conducted an analysis on the non-averaged data to investigate these advection. The data was divided into diurnal time series and each day of data is treated as an extra dataset in the spatial dimension in the PCA (refer to appendix A.2 for more detail). PC1 is found to be mainly influenced by clouds in the non-averaged analysis.

These advection events can be observed in multiple days of corresponding EOF 1 plots for January 2006 for a non-averaged analysis (see appendix A.2). Stripe like signals are observed on days 1, 3, 5, and 6 (Fig. A.2(b) and A.4) over the desert in the northern Africa; these signals correspond to a flux magnitude of $20\text{-}40 \text{ Wm}^{-2}$. These signals over the northern Africa appear to overlap with the westerly wind component observed at 400hPa, in contrast to the lack of strong U wind signal at 650hPa; this suggests the westerly wind at 400hPa to be the likely cause of the advection.

However, it can be seen that these signals are inconsistent in location, occurrences and their timings (i.e. can manifest as positive or negative EOF), which explains the low magnitude of signal observed in Fig. 6.3(a) on a monthly averaged dataset.

Furthermore, these stripe signals have been associated with the westerly wind at 400hPa, where the wind maximises at 200hPa. The clouds advecting between these two pressure levels would be considered as high level clouds, possibly cirrus or cirrostratus depending on the optical thickness.

6.1.3. Bright vegetation

As a comparison to the bright desert analysis, another surface type was selected for discussion in this section: the bright vegetation. This surface type is distributed in

two areas: mid-Africa and southern Africa.

Fig. 6.1(a) shows that the general surface heating curve is approximately the same shape as other surface types but with a peak slightly later than bright desert. This is consistent with our hypothesis on the chosen surface type as the heat capacity of vegetation is greater than desert, therefore, one would expect the heating curve of vegetation to slightly lag desert's.

There could be other factors which could influence the timing of this peak, such as the surface moisture, clear sky atmospheric moisture or the emissivity. All of which could have an impact on the OLR diurnal cycle by affecting the rate of change in temperature over the surface and the atmosphere. An increase in the surface or atmospheric moisture would delay the maximum of the OLR as they would decrease the rate of change in temperature. Similarly, an increase in emissivity would increase the emitted radiation which could decrease the rate of change in temperature, thus delaying the peak OLR.

Fig. 6.6 for January 2005/2006 illustrate the corresponding EOFs to the bright vegetation PC1 shown in Fig. 6.1(a), it shows that the band along mid-Africa exhibits a strong positive signal. Whereas, the southern region demonstrates a smaller magnitude in general where most of the regions are positive with small negative patches. The negative signals are possibly caused by clouds signals which in anti-phase with the surface heat response (i.e. late convection or advection of clouds).

The strong signals in the north indicate a flux magnitude of around $20\text{-}50\text{Wm}^{-2}$, this is consistent with diurnal variation of OLR over the clear sky vegetation surface type (see the first few days of Fig. 4.2(c) and Fig. 4.3(c) in section 4). This suggests the northern region is dominated by surface heat response.

The strong signals in mid-Africa indicate stronger heat response to the solar radiation in this region than the south. Although the minimum SZA is in the southern part of Africa, it's also where the ITCZ is positioned during this month. Hence, triggering many convective processes, which suppress the diurnal surface heat response. The OLR variability in the southern region is dominated by the cloud cycles, therefore, creating a low or zero magnitude in EOF1.

In comparison, the EOF1 for June (see Fig. 6.6(c) & 6.6(d)) have shown stronger signal in the south than in January (Fig. 6.6(a) & 6.6(b)) even though the minimum

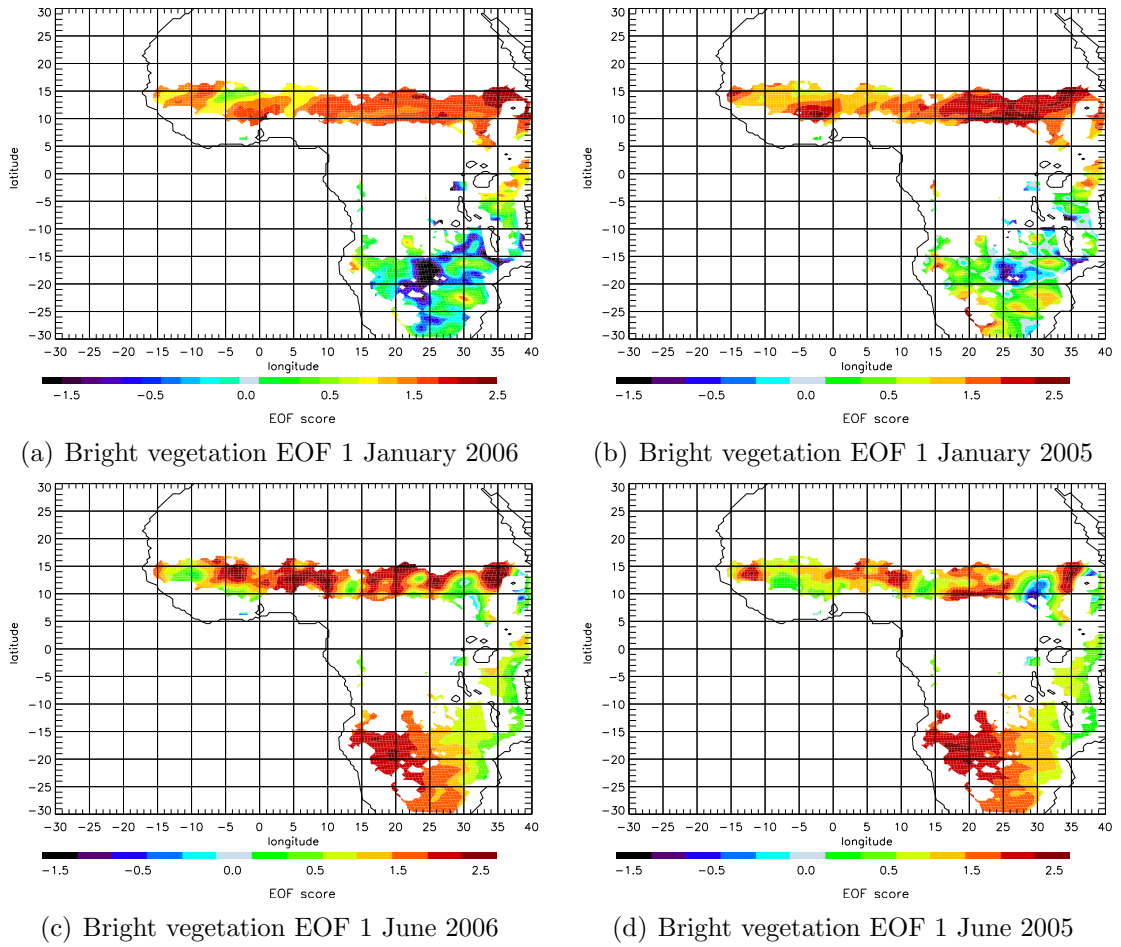


Figure 6.6.: a)January 2006 b)January 2005 a)June 2006 b)June 2005 Map of bright vegetation EOF1 from PCA over latitude and longitude range of 30°S-30°N and 30°W-50°E respectively, white represents an absence of data, NA indicates the part of colour scale which is not applicable to the diagram

SZA is now in the north, indicating that the ITCZ over the southern Africa in January significantly suppresses the surface heating signal which leads to the low magnitude of EOF1 signal observed in January over this region. However, over the Sahel region, the June EOF1 signal appears to remain strong although slightly more patchy than January. This implies the Sahel region is influenced by cloud activities in June where the ITCZ is near the region.

To further investigate the effect of cloud events on the OLR over this surface

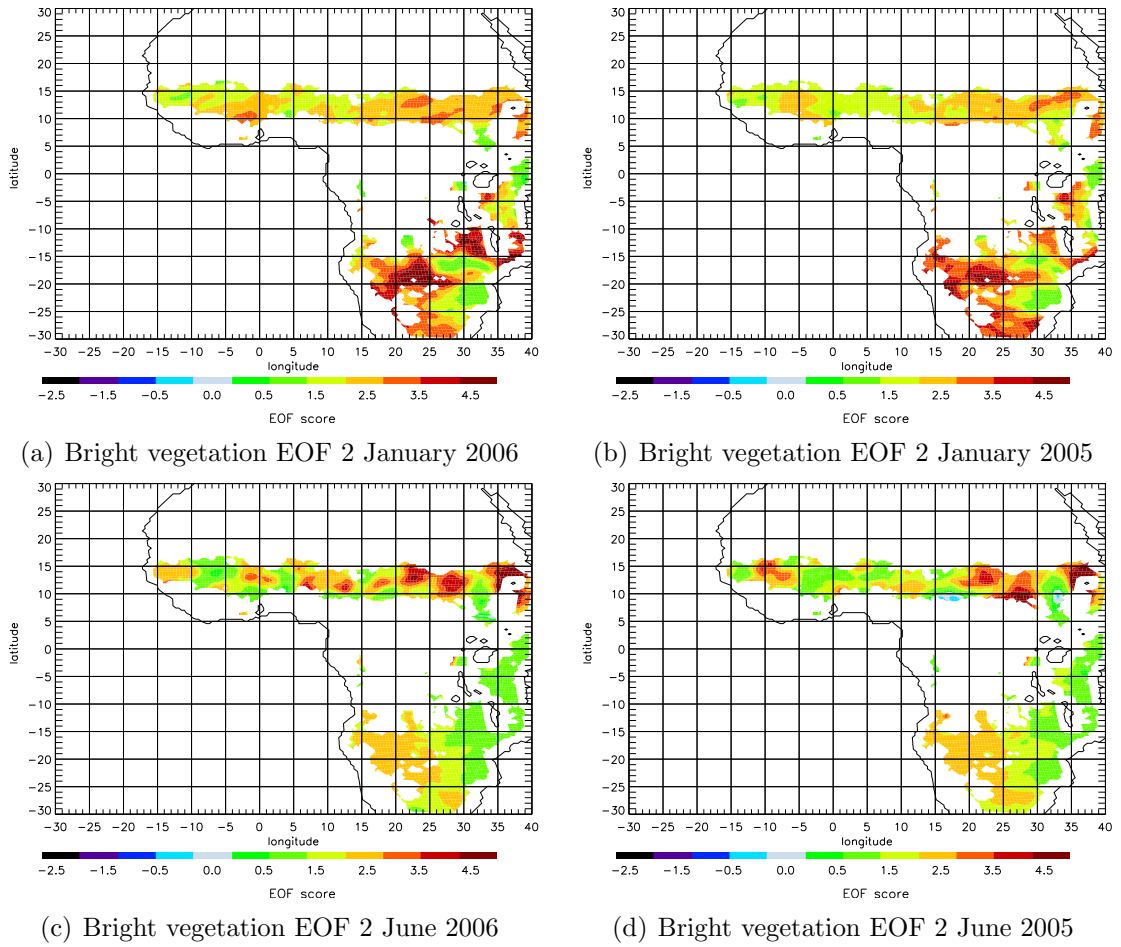


Figure 6.7.: a)January 2006 b)January 2005 a)June 2006 b)June 2005 Map of bright vegetation EOF2 from PCA over latitude and longitude range of 30°S-30°N and 30°W-50°E respectively, white represents an absence of data, NA indicates the part of colour scale which is not applicable to the diagram

type, it's essential to analyse EOF2. Firstly, by looking at the second PC for bright vegetation in Fig. 6.1(b), one can see that the curve lags the desert's PC2 by 2-3 hours, with the maximum just before noon.

The shape of the curve peaks just before noon and minimises in the evening, which is consistent with the cloud diurnal cycle [Chung *et al.*, 2007].

Fig. 6.7(a) & 6.7(b) have shown the EOF2 signals for January. The magnitude of this component is around $25Wm^{-2}$ in the north and around $30-45Wm^{-2}$ in the

south. During this particular season, most of the cloud activities are expected to occur in the south. However, there seems to be a small amount of diurnal OLR variation in the north contributed by this EOF2 component. This perhaps suggests some very different diurnal variability are present in these two regions which caused the second PC to contain some surface influences. Thus, prompted an investigation of these two separate regions in section 6.4 & 6.5.

In Fig. 6.7(c) & 6.7(d) , the June EOF2 signals appear to be strongest over the north, where the Sahel is, this is due to the ITCZ over the region during this season, generating a lot of cloud activities in the Sahel. Thus, producing a strong signal in EOF 2. However, in the southern region, there appears to be a significantly smaller but positive signal, where the region is relatively dry during this season and expected to have minimal amount of cloud activities. This again suggests there is a small surface heat influence in this signal. Therefore, we will separate the two regions and analyse them individually to further understand the diurnal variabilities.

For vegetative PC2, the contribution in variance is much greater (33%) than for desert (19%), this is due to some surface heat response influences present in this component as well as the cloud activities generated within the ITCZ region. A large portion of this surface type overlaps with the ITCZ and there is a large amount of moisture available for convection, therefore, triggering a lot of cloud events.

By examining the first and second order EOF's of the southern region together, it can be observed that a strong EOF2 signal is present, and the EOF 1 signal is relatively small. This indicates the occurrences of convection and cloud formation in this region suppress the surface heat response signal which is a reasonable interpretation; as the presence of cloud in general lowers the amount of ISR reaching the surface.

Over the bright vegetation, although the first EOF signal is dominant in the north, it's relatively weak in the south. This pattern is inverted for second EOF. This implies that the north presents a predominantly clear sky signal whereas the south is mostly influenced by clouds. We have chosen to analyse the vegetation surface type over these two regions separately as they appear to present very different diurnal variability.

6.1.4. Summary

This section of our work utilised the high time resolution data of the GERB instrument to find the small time variation in the different components within the OLR cycle of the various surface types. It's found that in general, the surface types with a higher heat capacity (i.e. vegetation) peak later in both PC1 and 2. This PC1 variability across the surface types appears to arise from the surface properties as this PC is mainly associated with clear sky surface heat response, which prompts an investigation on the effect of surface properties on the OLR.

There are a few factors which could affect the OLR diurnal cycle of clear sky surface heat response. Examples of these are the emissivity, heat capacity, SZA and atmospheric humidity. They could change the diurnal cycle by having an influence on the rate of change in temperature on the surface which in turn affect the emitted radiation.

Furthermore, energy exchange between the surface and the atmosphere can take the form of surface evaporation and sensible heat flux. These would also affect the temperature of the surface and the OLR cycle. However, these components are relatively small compared with surface emitted radiation (see fig. 1.2).

Thus, we have chosen to develop a simple model to test the effect of heat capacity(HC) on the diurnal cycle, as the HC is known to be directly proportional to the change in temperature (see eq.6.1). Also, the HC between the desert and the vegetation could differ by a factor of 2. This has a large impact on the OLR (see section 6.2).

Moreover, the small differences in the PCs have brought out other features of the EOFs which were not observed in the all sky averaged analysis. For example, cloud advection across the desert; these clouds are found to advect over the desert towards the north west direction and correspond to $20\text{-}40\text{Wm}^{-2}$ in a non-averaged dataset. However, these signals are very small over an averaged dataset (less than 10Wm^{-2}) due to the intermittent nature of the clouds, and the inconsistency in their location occurrences.

We have also found that over the vegetation surface type, the diurnal variability between the Sahel and the southern Africa is very different. Therefore, we have decided to analyse these regions separately to understand the physical mechanism behind the major OLR components in these two regions.

In this section, we have also analysed the diurnal variability over different surface types using PCA and we have found the first components to differ in the timing of the maximum OLR by about 2 hours. Since the PC1s are predominantly influenced by clear sky surface heat response, we have decided to develop a model to test this delay between the surface types' PCs.

There are a few factors which could influence the clear sky surface heat response: 1) emissivity of the surface, 2) heat capacity, 3) the SZA characterise by location and time of year, 4) atmospheric humidity. We have chosen the SZA and heat capacity to build our model on due to the fact that SZA directly influences the ISR which is the main driving force in the diurnal OLR cycle; the heat capacity is directly proportional to the change in temperature in the thermal equation, and the heat capacity can differ by a factor of 2 across the continent.

Therefore, we have chosen to investigate the influence of these causes on the diurnal cycle of the OLR by developing a single layer atmosphere surface radiation model.

6.2. Surface Radiative Model Testing

6.2.1. Model introduction

This section focuses on investigating the nature of the variation in the principal components presented in chapter 5 via the use of a surface radiation model. This model consists of a single clear sky atmospheric column with a fixed transmission, driven by a specified incoming solar flux from the GERB data, overlying a surface layer which absorbs and radiates energy, beneath the surface is a conductive layer of a variable heat capacity; full details of the model are provided in chapter 3, section 3.7.

The model will enable the study of the diurnal cycle in the OLR with respect to surface properties and ISR. It will be used to investigate the relationship between the diurnal cycle of the OLR and the values of these parameters; additionally, the output will be compared with the GERB results to provide an insight into the physical mechanisms behind the shapes of the PCs.

First, we investigate the model responses to the variations in each of its com-

ponents, and observe how the changes are manifested in the PCs derived from the model results. We then go on to use the model to understand the different OLR signatures for the different surface types that were highlighted in the previous analysis (see section 6.1). The overall aim of this investigation is to provide a physical understanding of the result previously presented in Fig. 6.1 which displays the distinct diurnal variability signatures associated with different surface types.

6.2.2. Model sensitivity

To understand the influence on the OLR of each of the parameters within the radiative model, we have first varied each parameter independently within physically reasonable ranges and the results are then compared with the control run.

The parameters include the heat capacity of the model surface and the latitude; the latitude changes the SZA, the duration of solar illumination and the magnitude of the ISR, and these are determined in conjunction with the time of year.

It is expected that by altering the heat capacity of the model, the amplitude and time variations in the temperature of the surface will be affected. In turn, the daily variation and overall magnitude of the OLR will also change along with the time of peak OLR during the day.

The fundamental concept behind this variation can be expressed in equation 6.1, where the change in temperature is directly proportional to the input energy. However, the heat capacity is inversely proportional to the change in temperature, which means the surface requires more energy to raise the temperature by the same amount with a higher heat capacity.

$$Q = m.c.\Delta T \quad (6.1)$$

m , c , ΔT , Q are defined as the mass, heat capacity, change in temperature and the thermal energy respectively (see section 3.7 for detail).

Thus, as the heat capacity goes up, the maximum in surface temperature (ΔT in eq.6.1) in response to the ISR will decrease. The modelled upwelling longwave radiation from the surface is dependent on the temperature of the surface as mentioned in section 3.7. Hence, a reduction in the overall OLR is expected as a result of higher heat capacity and vice versa.

Furthermore, an increase in heat capacity would decrease the *rate of* change in temperature, and cause a time delay for the surface to reach its maximum temperature during the day, and consequently the maximum OLR. Thus, introducing a lag in the OLR cycles between the surface types with different heat capacities.

Model Test With Varying Heat Capacity

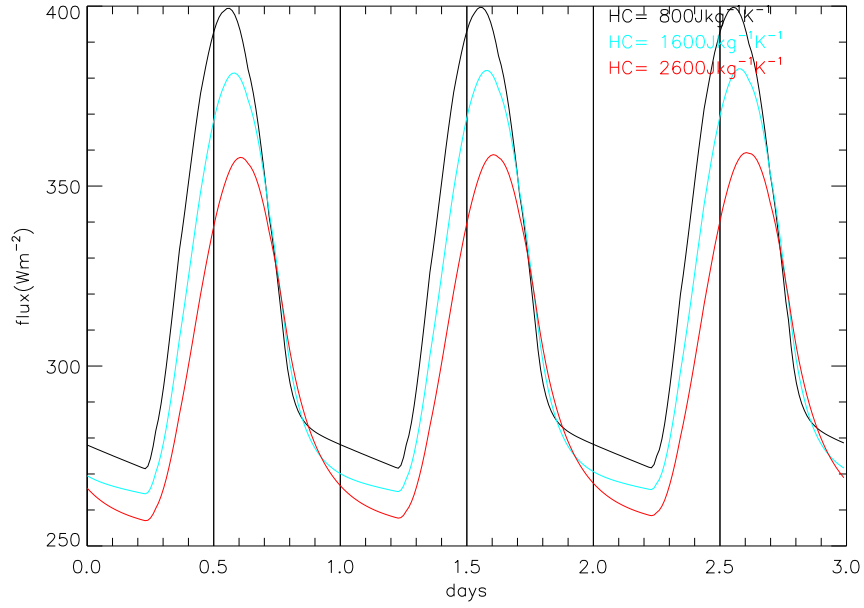


Figure 6.8.: Modelled diurnal OLR variation for various heat capacities in ISR for a point at 15°N in July

Fig. 6.8 shows the OLR variation from the model with the ISR over a GERB footprint at a latitude of 15°N in July, whilst varying between three different values of surface heat capacities. The variation in heat capacity represents one of the most important aspect of surface radiation properties, the range of which is chosen according the expected variation between different soil types and conditions. The thermal conductivity of the surface is fixed at $0.3\text{Wm}^{-2}\text{K}^{-1}$.

Different surface types have different heat capacities (HC), the modelled range of HC has been chosen to be consistent with observed surface types (sandy soil $\text{HC} = 0.80\text{Jkg}^{-1}\text{K}^{-1}$, peat soil $\text{HC} = 1.92\text{Jkg}^{-1}\text{K}^{-1}$). The magnitude of the overall OLR lowers as the model heat capacity is increased. This can be attributed to the lower

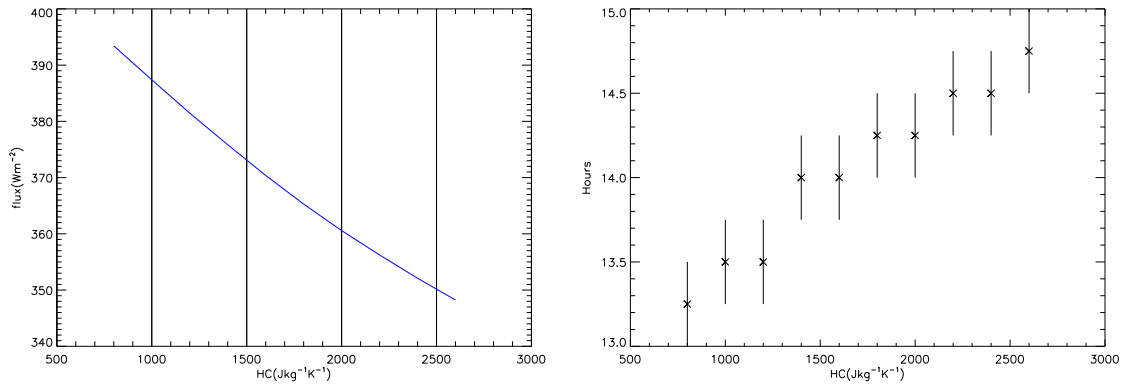
temperature change by the ISR as heat capacity increases; the resulting decrease in temperature reduces the surface radiation compare to that of a lower heat capacity surface.

This can be explained by equation 6.1, where the heat capacity is inversely proportional to the change in temperature, hence a lower heat capacity would cause the same energy input to induce a comparatively smaller temperature change. The surface in turn radiates according to the equation 3.14 at a rate proportional to T^4 .

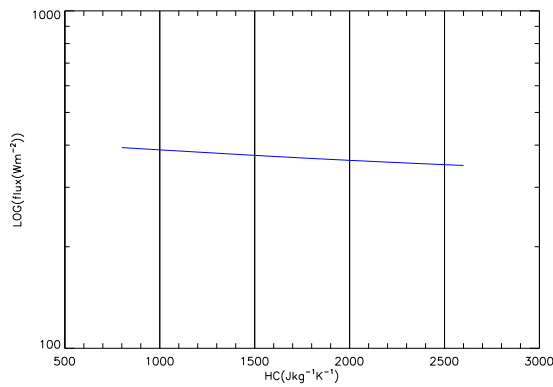
Fig. 6.9(a) shows the modelled maximum flux during the day against HC, it indicates an exponential decrease in the modelled OLR as the heat capacity increases and vice versa, which is consistent with the relationship between flux and temperature in equation 3.14. The modelled OLR has shown a diurnal variation of roughly $130Wm^{-2}$ which is slightly higher than a typical observed clear-sky OLR range of $80Wm^{-2}$ over the desert surface type at similar latitude ($15^\circ N$). However, this is expected as the model is based on an atmosphere with no aerosols or clouds, therefore no absorption in the downwelling SW and a minimal absorption in the upwelling LW with pre-set atmospheric emissivity. If we assume an aerosol scattering of a few percent in the SW (e.g. 5%), then a simple relationship between the thermal energy equation and the Stefan-Boltzmann law would reduce the radiative flux by about 20%, a rough calculation of this with the existing LW values would give a diurnal range close to $100Wm^{-2}$. However, this has not been implemented in our model as the aerosols across the whole continent is variable and the absorption effect of the aerosols would also need to be considered, which is beyond the scope of this analysis.

Fig. 6.9(c) shows a log plot of the flux against the heat capacity. The curve appears to demonstrate a linear relationship which confirms the exponential decrease of maximum flux with heat capacity.

An increase in the heat capacity shifts the diurnal peak in the OLR (or OLR maximum) to later in the day. The relationship of HC and timing of maximum OLR has been shown in fig. 6.9(b). The OLR peaks corresponding to the heat capacity have been expressed as a range since the maxima are limited by the time resolution of the model which is 15 minutes. This shift in the maxima can be explained by the change in HC, where an increase in HC causes a lower rate of change in temperature. Therefore, shifts the peak temperature of the surface to a



(a) Modelled maximum OLR flux against heat capacity (b) Modelled timing of maximum OLR flux against heat capacity



(c) Log plot of modelled timing of maximum OLR flux against heat capacity

Figure 6.9.: Modelled OLR properties with fixed ISR, attenuation and thermal conductivity whilst varying heat capacity(2006 July)

later time, and consequently delaying the maximum OLR during the day. The range of observed model timing of maximum OLR are comparable with the peaks observed in chapter 5, which is approximately an hour between dry and moist surface types.

Summary

The model study shows that as heat capacity of the surface is increased, the maximum OLR shows a $1/e$ (see fig.6.9(c)) decrease in magnitude and moves later in the day. Since the heat capacity determines the change in temperature given the same ISR time series, and this change in temperature in turn varies the radiative flux from the surface.

We have investigated the relationship between the HC and maximum in OLR using our simple radiative model, and we have seen similar variabilities that are exhibited in the PCs of the surface types. Thus, we have produced a direct comparison (in section 6.3) between the model and GERB data to observe if HC and latitude are adequate in describing the surface types variability we have observed in section 6.1.

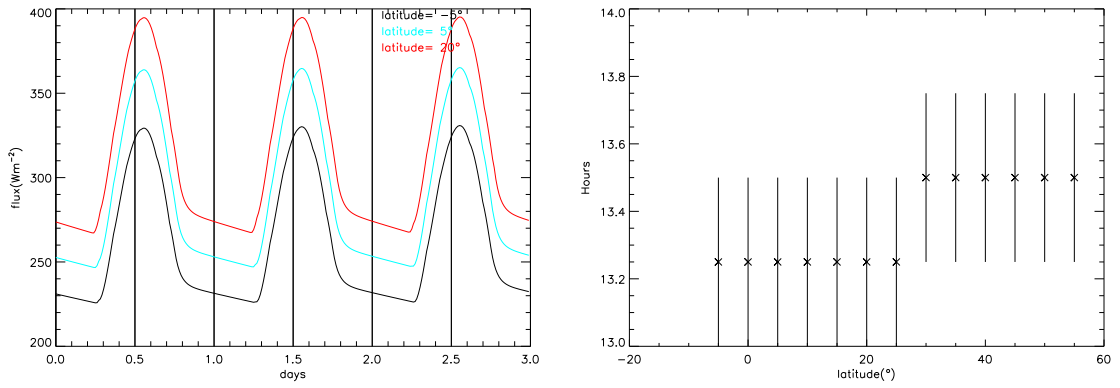
Model Test With Varying Latitude

The model sensitivity is tested with different ISR from various latitudes in this subsection. There are two degrees of variability in the ISR variation, first of which is the magnitude of the ISR that is determined by the SZA, second is the duration of daytime which is again dependent on the SZA. Since both variations in the ISR are determined by the SZA, we have therefore chosen a range of latitudes as a variable on the SZA to test the peak and magnitude of the model output OLR.

The model is expected to respond to each degree of variability in the following manners: firstly, the magnitude of the ISR will affect the temperature of the surface which will then influence the magnitude of the OLR; secondly, the illumination duration in the ISR profile determines the width of the peak, this is expected to affect the OLR curve in a similar way where the width of the peak would change accordingly. The change in the width of peak is a result of the rate of change in the ISR. For example, by lengthening the illumination duration with a fixed peak magnitude, the rate of change in ISR is decreased. Consequently, the rate of change in the OLR output is also decreased, creating a wider peak width in the diurnal cycle.

The change in the peak width indicates an adjustment in the rate of change in the ISR, thus affecting the gradient of the ISR flux around the peak. Therefore, one would expect a small change in the timing of the maximum OLR as the rate of change in the OLR cycle is dependent on the gradient of the ISR flux.

Fig. 6.10(a) indicates a change in latitude is accompanied by a change in the OLR amplitude, which confirmed our first hypothesis. Fig 6.10(b) shows a very small time variation with respect to the latitude, a temporal difference of 15 minutes is observed over a 60° change in latitude. As expected, the time profile of maximum OLR indicates that the latitude has an effect on the maxima, although clearly the



(a) Modelled diurnal OLR variation for various latitudes (b) Modelled timing of maximum flux against latitude

Figure 6.10.: Modelled OLR properties with fixed heat capacity ($0.8Wm^{-2}$) and thermal conductivity ($0.3Wm^{-2}K^{-1}$) whilst varying ISR(latitude)

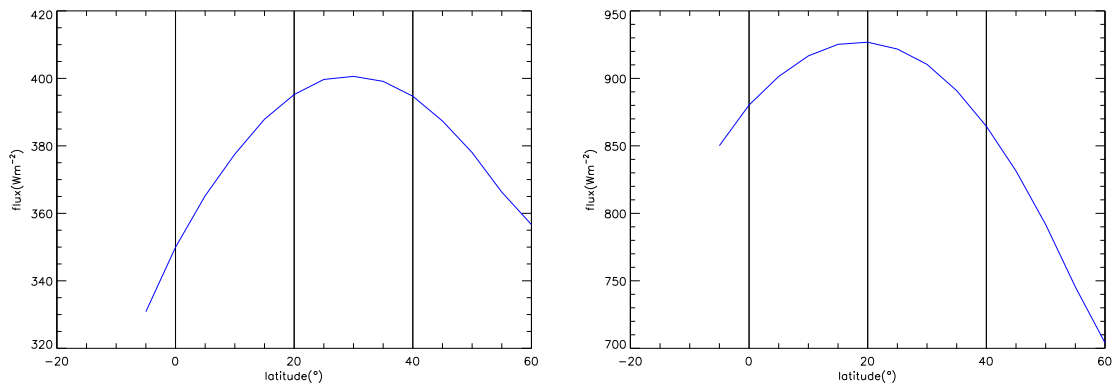
sensitivity is very small compared to the heat capacity changes.

This change in time of the maximum OLR with latitude can be explained by the change in the rate of change in the ISR during the day. The rate of change in ISR is influenced by the latitude; by changing the illumination duration during the day, we observe a corresponding change in the gradient of ISR. Therefore, changes the time of maximum surface temperature and the resultant OLR.

The influence in the peak over the 60° latitude change is roughly 15 minutes, and since our analysis is over the latitude $30^\circ S$ and $30^\circ N$, we would not expect a change in the peak to be beyond 15 minutes contributed by the latitude change alone.

The maximum modelled OLR and ISR are shown as functions of latitude in Fig. 6.11(a) and 6.11(b) for July 2006. Fig. 6.11(b) displays the variation in the maximum ISR with latitude for July 2006, where the flux function demonstrates a sinusoidal-like wave and the maximum is centered at roughly $20^\circ N$ which is consistent with the incoming flux magnitude associated with the minimum SZA during this particular month. However, fig. 6.11(a) presents a slightly different profile for the OLR.

In fig. 6.11(a), the modelled OLR flux is calculated at selected latitude range and the diurnal maximum fluxes are plotted against latitude. The modelled fluxes show a maximum near $30^\circ N$, this is approximately 10° away from the maximum ISR (minimum SZA). This indicates the magnitude of ISR is not the only factor which



(a) Modelled maximum OLR flux as a function of latitude (b) Maximum ISR flux against as a function of latitude

Figure 6.11.: Maximum a)OLR b)ISR against as a function of latitude (2006 July)

governs the strength of the modelled OLR. The duration of daylight is another factor that influences the OLR flux. The daylight duration is expected to affect the OLR in terms of the total energy input onto the surface during the day.

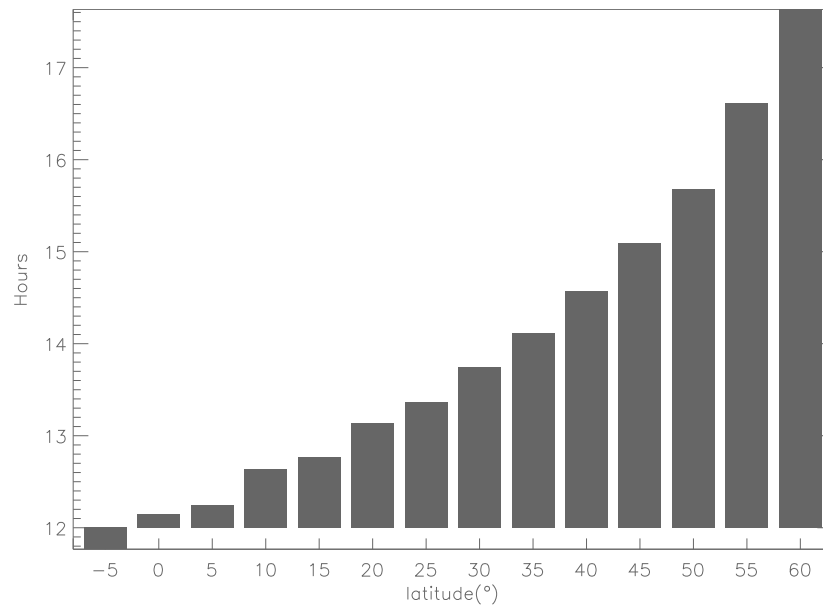


Figure 6.12.: Hours of solar illumination per day as a function of latitude, average over 2006 July

During boreal summer, the daylight duration is increased with latitude, this re-

lationship is plotted in fig. 6.12. The increased daytime duration has a positive influence on the OLR, as the illumination duration increases with latitude, the total energy incident on the surface is higher.

This positive effect combined with the negative effect from the decreasing ISR magnitude associated with the variation in SZA (with increasing latitude) to cause the total energy incident on the surface to decrease beyond 30°, shown in fig.6.13. Consequently, the surface radiates (fig. 6.11(a)) in response to the total incident energy variation shown in fig. 6.13.

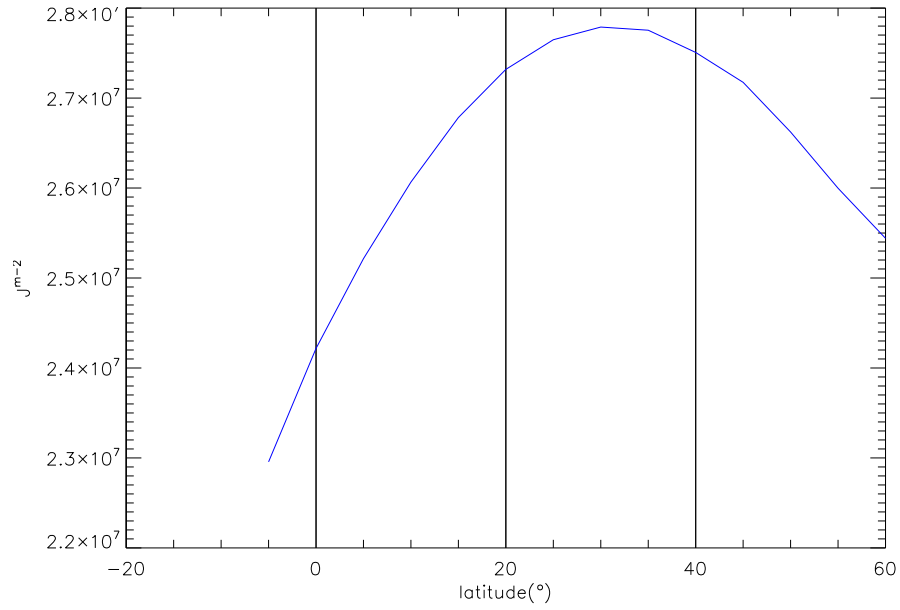


Figure 6.13.: Total energy from ISR for 1 day per square meter

Fig. 6.13 demonstrates the total energy from the ISR (an integral of the ISR fluxes during the day) against the selected latitude range. The peak in the plot centers at 30°N and is similar to the latitudinal maximum OLR distribution shown in fig. 6.11(a). This total energy from the ISR is influenced by the peak magnitude and the illumination duration, where both are determined by the latitude and the time of the year (i.e. the SZA).

We have also hypothesized a change in the peak width of the OLR in accordance with the increase in the peak width of the ISR varying with latitude, as the rate of the change in the OLR is dependent on the gradient of ISR.

This is shown in fig. 6.14, where the full width half maximum(fwhm) of the ISR/OLR peaks are plotted against the latitude range. The fwhm is calculated by measuring the time difference between the half way points between the maximum and minimum. We have used the fwhm as a mean to measure the relative rate of change in the ISR/OLR in the diurnal cycle, where a higher fwhm physically implies a longer period is required for the flux to reach the peak from its median value and vice versa.

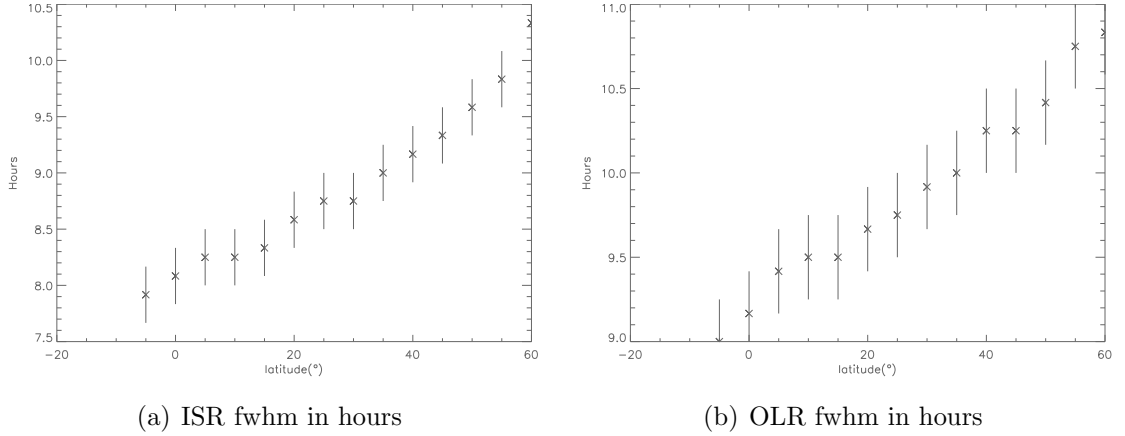


Figure 6.14.: Full width half max for incoming and outgoing fluxes

The fwhm of the ISR is expected to follow the same pattern as the time length of illumination shown in fig. 6.12 as we are measuring the widths at two different heights (greater than $0Wm^{-2}$ and half maximum) of the diurnal ISR at different latitudes, which are anticipated to be self-consistent. Furthermore, we are observing a similar increase pattern in the OLR fwhm with latitude.

The fwhm of the OLR in general are broader than the ISR which can be attributed to the dampening effect of the heat response (heat capacity) that increases the peak width of the OLR as well as decreases the amplitude of the maximum comparing with the ISR. The width of the OLR increases with latitude as expected and the time increase are well correlated between the ISR and OLR.

Summary

In this experiment, we have investigated the diurnal OLR with latitude as a variable. We have observed a change in latitude alters the peak and duration of the ISR

which produce a combined change to the total incoming energy; the maximum OLR responds to the total energy whilst the width of the OLR peak correlates with the change in the duration of illumination.

On the other hand, latitudinal variation has shown a small impact on the timing of the OLR peak, the timings of maximum OLR have been consistent over the tropical/subtropical region and shown a roughly 15 minutes shift in the OLR peak over a 60° change in latitude, which is relatively small comparing to the variation due to heat capacity especially in the latitude range we are investigating (upto 30°). This small shift of peak originates from the change in the gradient of the ISR, which in turn affects the time it takes for the surface (with a finite heat capacity) to reach its diurnal maximum temperature. Thus, shifting the peak of the OLR.

We would expect this shift in peak with latitude to vary across different surfaces as the heat capacity can also affect the time for the surface to heat up to its maximum temperature. This time shift effect is expected to play a larger role on higher heat capacity surfaces as the gradient of the temperature is lower on higher HC surface. Therefore, further delaying the surface from reaching the maximum temperature with a smaller rate of change in ISR.

We have established that the change in the timing of the maximum OLR due to the latitudinal variation across the different surface types to be small. However, the model data will further be analysed using PCA, if this change in the peak is significant, this will manifest as some sort of shifting mechanism in the lower order PCs for specific surface types, most likely to be PC2 of desert/vegetation and contribute to a significant amount of variance.

6.3. Model surface type PCA

This section is intended to use our surface radiation model to compare with our GERB data results to test if the PCs observed in the surface type analysis are influenced by surface properties, in particular the first PCs.

We have carried out this analysis to test our hypothesis that the surface properties, especially the heat capacity, are the dominant characteristics in the surface heat response across the different surface types. This was demonstrated in our GERB surface type PCs. In the first PCs, we have observed time delay for the peaks between

different surface types. Since this component is hypothesized to be predominantly caused by the surface heat response, we have devised this test to investigate how the same PCs vary in our model. Furthermore, we have already shown in our model sensitivity tests that the SZA has an influence on the magnitude of the output OLR but very little effect on the timings of the diurnal OLR peaks. Therefore, we expect any differences in time delay of the PCs to be the effect of heat capacity.

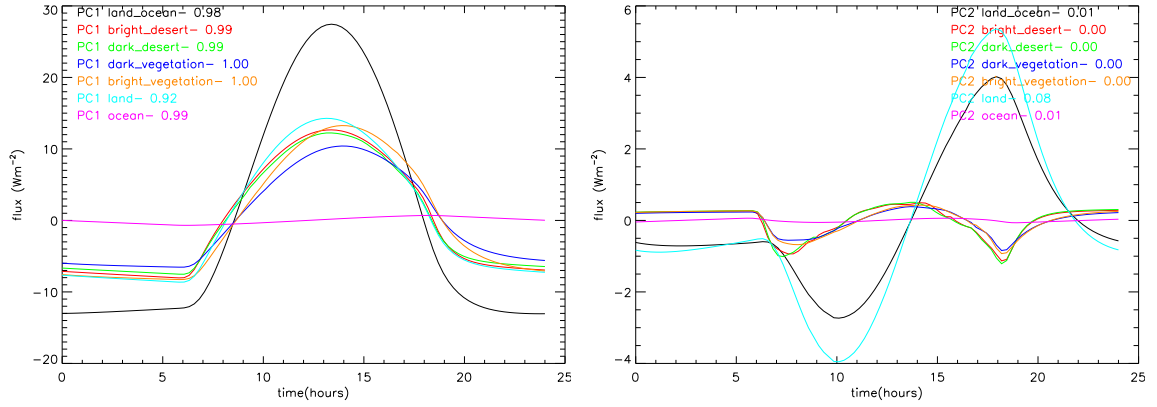
We have used the ISR data from GERB SW files and separated the data set in the same way as the surface type analysis for the model input. The heat capacities for the bright and dark desert are the same in our model ($800 J kg^{-1} K^{-1}$) and for bright and dark vegetation, the HC is $1920 J kg^{-1} K^{-1}$. The model is allowed to run for 30 days, and the last 5 days of data are averaged into diurnal OLR cycles for the PCA. The reason only to use the last few days of OLR cycles is to ensure that the model is stabilised.

Fig. 6.15(a) shows the first PCs for various surface types from the model output. Although we have plotted the ocean and land + ocean in our model PCA, it should be noted that these cycles are not intended for comparison with GERB data as we recognise our model does not capture the most crucial ocean component which is the ocean circulation.

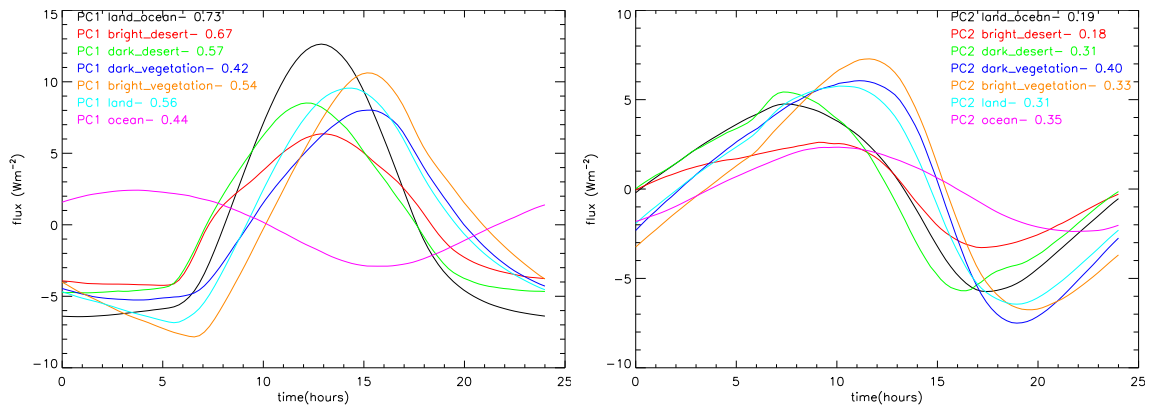
An initial comparison between the model PCs (fig. 6.15(a) & 6.15(b)) and GERB data PCs (fig. 6.15(c) & 6.15(d)) demonstrates: i) the delay between surface types in GERB is around 2 to 3 hours, compared with the 1 hour difference in the model; ii) PC2 in model has a low magnitude and low variance contribution in general. This is discussed further in the next few paragraphs.

First of all, the variances contribution from the first PCs are well over 90% across all surface types, this can be explained by the simple nature of our model, where cloud variability and convective events are not present. Therefore, the diurnal OLR cycles outputted are pure modulations of the heat response to ISR, hence the diurnal cycles are very consistent with each other with small variations in timing of maximum ISR and shape due to the surface properties and SZA.

Second of all, there appears to be a delay between the desert and vegetation surfaces, which is expected as our model sensitivity test has indicated an increasing peak delay with HC. However, the delay in fig.6.15(a) is roughly 1 hour between the two surface types, this is much smaller than the delay observed in the first PCs of the



(a) Different surface types model all sky PC1 plot (b) Different surface types model all sky PC2 plot



(c) Different surface types GERB all sky PC1 plot (d) Different surface types GERB all sky PC2 plot

Figure 6.15.: Model (July 2006) a)PC1 b)PC2 for different surface types along the tropical region (-30° - 30° N): land, ocean, bright desert, vegetation, land + ocean. Numbers on top right corner of each plot indicate the fraction of variances for the individual components. The change in Model Outgoing Longwave in Wm^{-2} and local time in hours are represented by y-axis and x-axis in the plots respectively. Different colours represent the PC for the various surface types. c)PC1 b)PC2 for GERB surface types data analysis reproduced from fig. 6.1(a) & 6.1(b)

GERB data (fig. 6.15(c)) which is between 2 to 3 hours. This cannot be attributed to the SZA as we have used the ISR data from GERB to create the model OLR in different latitudes, where the SZA is an inherent part of the ISR variability.

One possibility for the time delay differences between the GERB PCs and model PCs could be due to the surface humidity available for convective events and con-

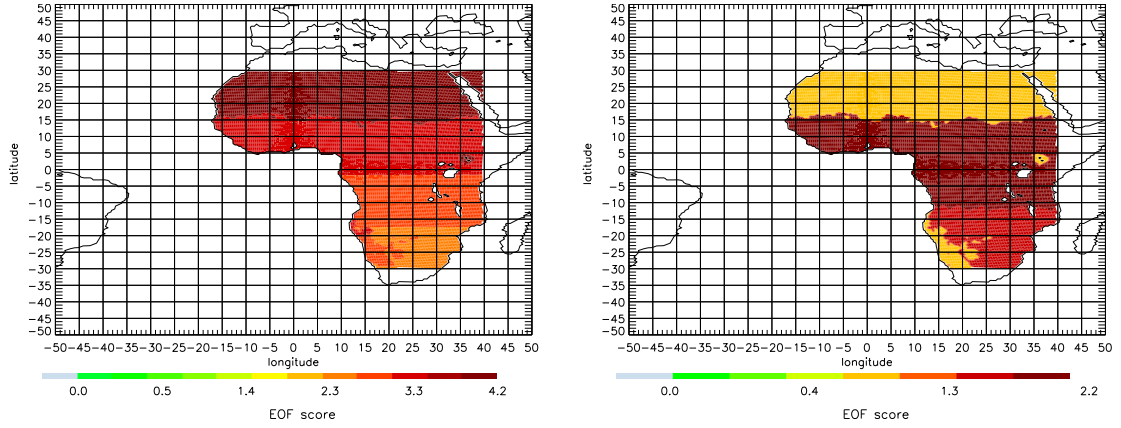
sequently affecting the moisture diurnal cycle in the atmosphere which is not accounted for in the model. In general, the vegetative surface type has more moisture available for convection which could affect the peak of the diurnal cycle by two different means: 1) increasing the moisture in the atmosphere, therefore, having a positive effect on the moisture diurnal cycle, 2) inducing cloud development cycle via moisture convection which typically peaks in the late afternoon in cloud optical depth.

The argument where the moisture diurnal cycle itself is responsible for the shift in the OLR, suggests that the warm wet surface (i.e. vegetation surface type) provides moisture and temperature gradient for the moisture to convect. The humidity in the air then absorbs part of the LW flux emitted by the ground, thus warms up the air and radiates. In general, when there's more moisture in the atmosphere, the atmosphere will absorb more LW flux from the ground. Hence, a maximum in humidity would correspond to a minimum in flux propagation from the ground and vice versa. It has been suggested by *Chung et al.* (2007) that the diurnal cycles of Mid and Upper tropospheric moisture (MTH and UTH) peaks just after midnight (0200-0300) over land and minimises at around 1600. This could explain the discrepancy in the time delays between the GERB PCs (2 to 3 hours between desert and vegetation surface types) and our model PCs (1 hour).

The second modelled PCs in fig. 6.15(b) have contributed much less to the total variance in comparison to the GERB analysis(see fig. 6.15(d)). This is expected because the model does not produce any cloud cycle which is the main contributor in GERB PC2. Instead, what we are observing in the model PC2 is the shift in the peak for the surface heat response cycle between different locations. For example, the all land and land+ocean PC2 describe the differences between one surface type to another. Thus, we observe a clear modulation of surface type signals in the corresponding EOF2 in fig. 6.16(b).

However, for the PC2s of the individual surface types, this merely represents the small latitudinal effect on the shape of the diurnal cycle (see section 6.2.2) and the uncertainties arise from the time resolution of the model and GERB ISR data (15 minutes and roughly 17 minutes respectively), which accounts for less than 1% in the desert and vegetation surface types. The PC2 for the all land model analysis has contributed to around 8% in the total variance, which suggests that the surface

properties have a considerable effect on the peak of the PC and strengthens our case for separating the surface types to analyse the individual diurnal variability.



(a) Model land EOF1 over 30S-30N, 40W-30E (b) Model land EOF2 over 30S-30N, 40W-30E

Figure 6.16.: EOFs (July 2006) corresponding to land only of the PCs demonstrated in fig. 6.15 for model OLR data

Fig. 6.16 shows the EOF1& 2 corresponding to land only PC1& 2 in fig. 6.15. In the first EOF, we can see two features which affect the magnitude of the signal. Firstly, there appears to be a latitudinal gradient (see 2°, 17°S in fig. 6.16(a)) in the EOF magnitude, this can be attributed to the SZA, which minimises in the north during boreal summer. The northern part of the continent receives the maximum ISR and this gradually decreases as the latitude decreases. The EOF1 combined with the PC1 for the land only surface shows just over $20Wm^{-2}$ in regional variability from northern Africa to the south over the vegetation surface. Secondly, we see a surface type effect in two regions: northern Africa (mostly Sahara desert) and southern Africa (Kalahari desert). The desert region shows a stronger heat response than the vegetation surface type which can be explained by the difference in heat capacity of the surfaces (see section 6.2.2). If we look near the Kalahari desert in the first EOF, we can see a change of roughly $10-20Wm^{-2}$ in diurnal cycle magnitude difference across the two surface types at the same latitude.

In the second EOF (fig. 6.16(b)), there is a low amplitude of signal over the desert, which implies the desert OLR cycle is already well represented by the first all land PC in fig. 6.15(a). However, over the vegetation surface type, there appears to be a large signal of PC2 in most region, suggesting a forward shift (in time) of

the OLR peak is required to represent the diurnal variability across this surface type due to its higher heat capacity. The second EOF score and the PC combined to just under $20Wm^{-2}$ in diurnal variation for this particular component over the vegetation surface type.

6.3.1. Desert and vegetation EOF

In this subsection, the data has been separated in desert and vegetation types according to GERB SW scene ID. The EOFs for the surface types corresponding to the PCs in fig. 6.15 have been plotted.

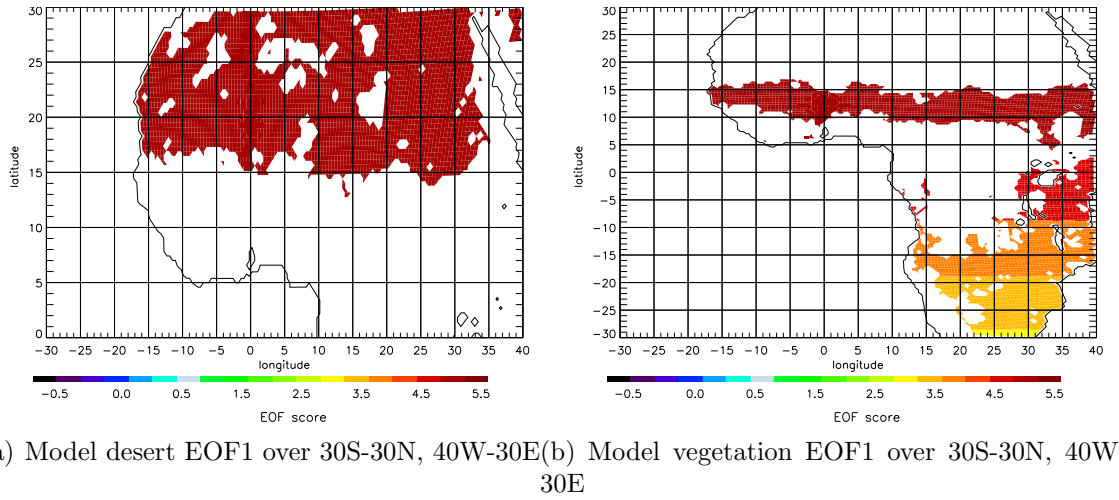


Figure 6.17.: EOFs (July 2006) corresponding to desert and vegetation of the PCs demonstrated in fig. 6.15 for model OLR data

Fig. 6.17 shows the EOF1 for desert and vegetation, EOF2 are not demonstrated for these surface types as they contribute to less than 1% of the total variance and are insignificant in changing the shape of the overall diurnal cycles. The EOF1 for desert (fig. 6.17(a)) shows a strong positive signal over the whole region which suggests the surface heat response is consistently strong across the desert and latitude has little effect over this region as the minimum SZA is in the center of the desert during this month and the ISR differences around this latitude are too small to create an observable difference in the model OLR EOF1. On the other hand, EOF1 of the vegetation surface spans across a larger latitude range further away from the minimum SZA. A latitudinal gradient can be identified in fig. 6.17(b), where the

change in the EOF score is around 2 from the Sahel region to southern Africa, indicating a difference of roughly $40Wm^{-2}$ across the latitude.

6.3.2. Summary

In this section, we have investigated and discussed the effect of heat capacity on the peaks of PC1 in the principal component analysis. We have confirmed the heat capacity to be one of the causes to the time delay between the peaks in PC1. This effect contributed to about 1 hour in peak shift across the surface types which is less than what we have observed in the GERB data analysis. In our model sensitivity test, we have investigated the latitudinal effect on the peak and found it have a minimal contribution to this (15 minutes across a 60° change in latitude).

This discrepancy between our model surface type PC maxima and the GERB's PC cannot be due to the convective cloud developments as *Chung et al.* (2007) *Held and Soden* (2000) and *Tian et al.* (2004) have suggested an afternoon/evening peak of the cloud optical depth (minimum in OLR) and this is confirmed by our PC2 study in the previous chapter (chapter 5). This cloud influence would shift the overall OLR diurnal peak to noon/late morning which is inconsistent with our observation of the PC1 over various surface types.

Therefore, we have described a humidity mechanism as the cause which could influence the OLR cycle to produce a later peak over the moist surface types, the humidity cycle itself is studied by *Chung et al.* (2007) and they have found the humidity to maximise just after midnight and minimise during the afternoon, which could explain the discrepancies between our time delay in the model PC1 and the GERB data PC1. Naturally, the moisture in the vegetation surface type is higher than desert, this could provide a source for the atmospheric humidity and hence, create the difference between our model and the observational PCs.

There are other factors which could have influenced the clear sky OLR peak for the surface types other than the atmospheric humidity described above. One of which is the emissivity of surface types, the increase of it leads to a decrease in the rate of change in temperature of the surface as the surface will be radiating at a higher rate given the same initial temperature. Consequently, the OLR peak will be delayed during the day.

Another factor which could affect the clear sky OLR is the amount of surface moisture across the different surface types. This could change the peak of the OLR cycle in two ways: 1) by changing the heat capacity of the surface, 2) by providing more atmospheric moisture via evaporation. In the former case, an increase of the surface moisture in general increases the heat capacity (due to the high heat capacity of water). This decreases the rate of change in temperature and thus, shifts the OLR peak. In the latter case, the atmospheric moisture could induce a diurnal cycle described above.

Our Sahel regional analysis in the next section serves two purposes. First of which is an extension to the surface type analysis in section 6.1.3. This is to investigate the PC contribution in this region separately as we have found the diurnal variability to be very different between the Sahel and southern Africa (see section 6.1.3); this is due to the different dominating factors in the OLR diurnal cycles between the two regions which is investigated in the next section.

Secondly, by only analysing the PCs in the Sahel region, we have eliminated some of the causes mentioned above which potentially contribute to the discrepancies between the GERB data analysis and our model analysis. The Sahel region is confined along a latitude band, which would be subjected to roughly the same atmospheric conditions and precipitation during the different seasons. Thus, bearing similar surface and atmospheric conditions which we predict to put the observational analysis much closer to our model data.

6.4. Sahel region investigation

6.4.1. Introduction

To further understand the diurnal components in the OLR along Sahel, we have analysed the Sahel region on its own to investigate any similarities/differences between the regional test and the more generic tests, thus pick out any signals which are unique in the region.

We have defined the Sahel region for this analysis to be between the latitude range of 5-20°N, this is due to the fact that this region is along the boundary between the desert and vegetation, both of which are crucial in our analysis. Moreover, this is

the region where the ITCZ is manifested during boreal summer, which enables us to observe the differences between the dry and wet seasons and how they contribute in the OLR variability.

This analysis includes only data points from the Sahel to isolate the components which are responsible for the diurnal variabilities along this region. The bright and dark vegetation types of previous results (fig. 6.1) have been combined due to their similar properties and to create a larger dataset for a more consistent analysis.

The effect of combining the bright and dark vegetation in this analysis is expected to be small because: 1) their individual PCs have shown similar timings in maxima and minima in the OLR diurnal cycle, so their combined PCs should be similar to their individual PCs, 2) of similar atmospheric conditions and precipitation distributions due to latitudinal constraint (5-20°N), 3) of similar SZA hence similar ISR cycle, again due to latitude range.

This effect of combining bright and dark surfaces is expected to be small relative to the Sahel and southern region separation as these two regions are subjected to different atmospheric conditions. They are also separated in latitude which constitute different SZAs and ISR cycles; and the presence of AEJ (boreal summer) over the Sahel region is going to effect its diurnal variability by encouraging convective events and precipitation.

6.4.2. Sahel PC analysis

Fig. 6.18 demonstrates PCs 1 & 2 for the various surface types specifically for the Sahel region from 5° to 20°N. The PC1 and PC2 combined with their corresponding EOFs to give a maximum diurnal variation of $40\text{-}50\text{Wm}^{-2}$ and around 10Wm^{-2} respectively over the desert which are consistent with our desert observation over the whole continent.

First of all, fig. 6.18(a) shows PC1s of various surface types to be close together in terms of OLR peak, with a much smaller time difference between desert and vegetation PC peaks of up to an hour compared with the two hours delay observed in the analysis over the whole continent. This PC peak time delay difference between the surfaces is very similar to what we have observed in our model PCA (see Fig. 6.15(a)) compared to the larger differences found in the larger scale analysis (see

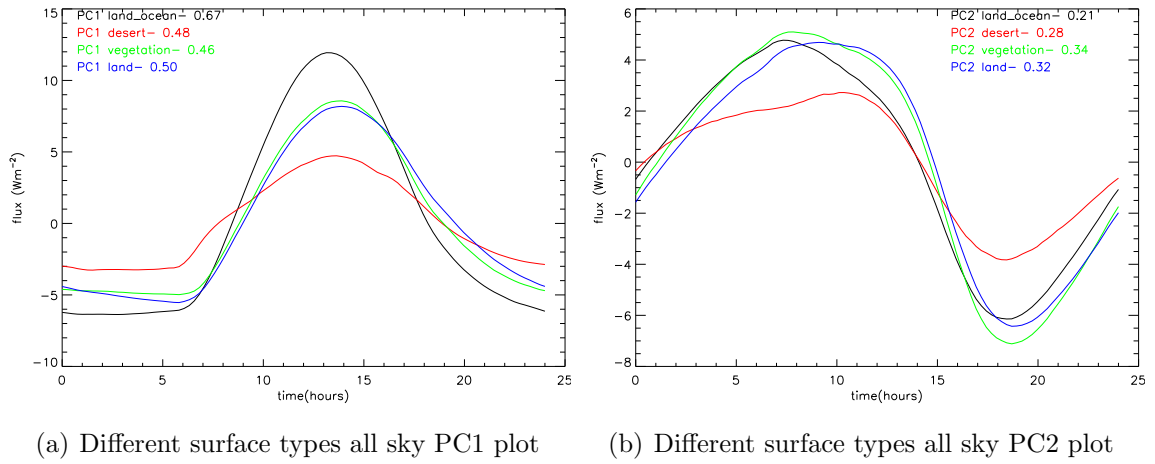


Figure 6.18.: a)PC1 b)PC2 for different surface types along the Sahel region (5-20°N): land, desert, vegetation, land + ocean. Numbers on top right corner of each plot indicate the fraction of variances for the individual components. The change in Outgoing Longwave in Wm^{-2} and local time in hours are represented by y-axis and x-axis in the plots respectively. Different colours represent the PC for the various surface types. Months [01, 06, 07, 12] between 2004 and 2007 GERB LW data are all used to compute this PCs

fig.6.1(a)). This similarity can be attributed to the consistent atmospheric and surface moisture across this latitude band (discussed in section 6.3.2).

Second of all, the land PC1 curve shows a peak which is closer in time to the vegetation PC1 than the desert's, which can be explained by the larger contribution of the vegetative surface type in the Sahel region. Therefore, the land variability is expected to be dominated by the clear sky surface heating across this surface type.

Third of all, the PC2 shown in fig. 6.18(b) exhibit similar time delay between the peaks to PC1. This is consistent with the hypothesis that the second component is a manifestation of the cloud development which is heavily linked with the surface heat response in the tropics. Therefore, we would expect a strong relationship between PC1 and PC2 in terms of timing of the peaks, given that one process is highly dependent on the other.

Furthermore, the ratio of variances contributed by each component are observed to be slightly different to the analysis over the whole Africa (Fig. 6.1(b)). In the Sahel region, the variances have shown that the emphasis on all PC1s is generally

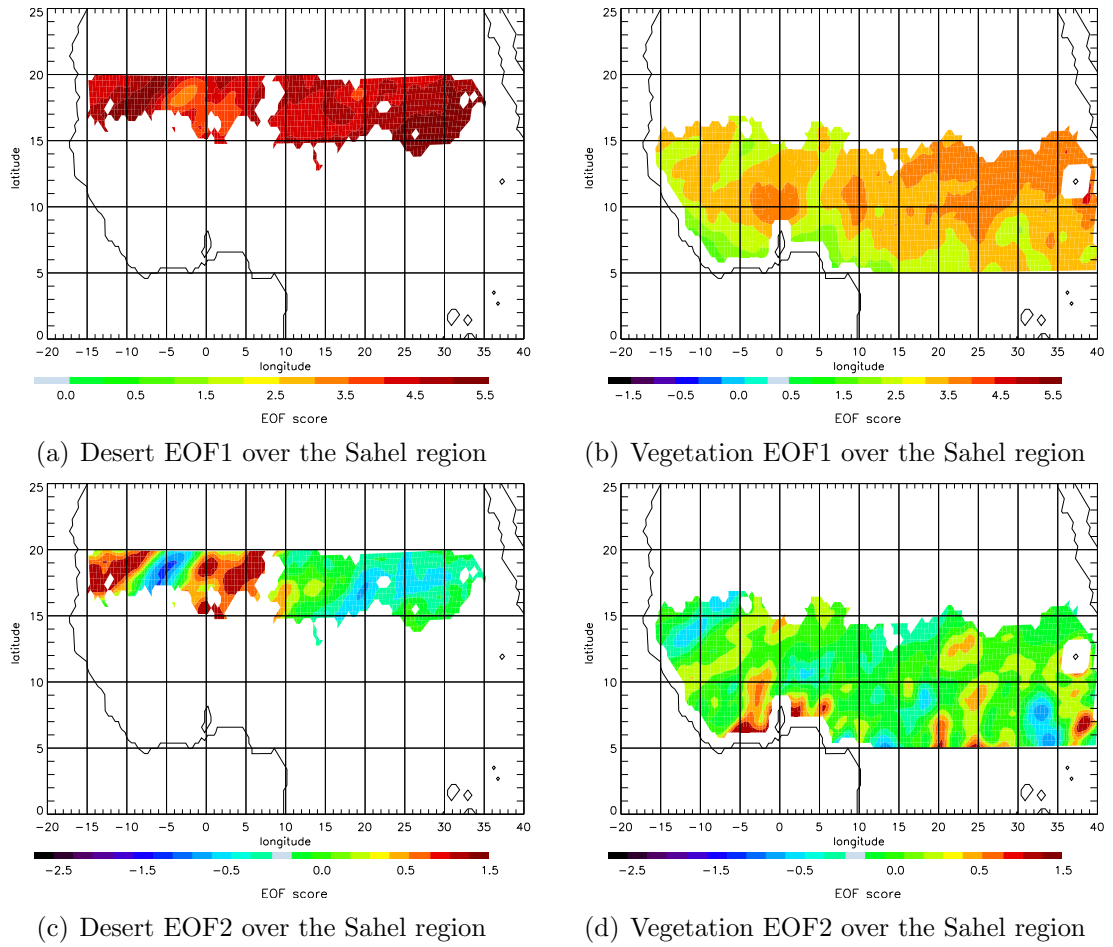


Figure 6.19.: GERB observed EOFs corresponding to some of the PCs demonstrated in fig. 6.18 in the Sahel region for the whole month January 2006

lower compared with the African continent analysis, whilst a general increase in the contribution of PC2 variances have been observed over the specific region. This is a predictable outcome as the region experiences more cloud occurrences and rainfall due to heavy influence of the ITCZ and AEW compared with the rest of the continent.

Fig. 6.19 are spatial representations (EOF) of the PC1& 2 shown in fig. 6.18 for the desert and vegetation surface types in January 2006. The pixels immediately adjacent to the land ocean boundary have been removed to eliminate any coastal processes which could potentially dominate the PCs and suppress the signals we are

interested in.

Fig. 6.19(a) and 6.19(b) both demonstrated a fairly consistent positive PC1 signals across all regions (note. the discontinuity of the colour (EOF strength) between the two surface type plots are caused by the slightly different amplitude in the associated PC1s). This indicates a strong surface heat influence across most of Sahel, this is expected as the land ITCZ is in the south during January.

Fig. 6.19(c) and 6.19(d) represent the spatial EOF2 signals associated with the PC2. First of all, it can be observed that there are alternating stripes of positive and negative signals in western Sahel in Fig. 6.19(c). Those can be attributed to the advection of clouds as discussed in section 6.1.2. However, if we shift further south into the vegetation surface type in Fig. 6.19(d), it can be observed that the most dominant signal is near the coast between 5-10°N. Again, this is likely to be part of the cloud advection that we have discussed in section 6.1.2.

By comparing the shape of the PC1 and the spatial distributions in Fig. 6.19 & 6.18(a) with previous results of the bright desert type (Fig. 6.6(a) & 6.7(a)), it can be seen that surface heat response signal (PC1) of Sahel is the dominant signal for this particular surface type, this hypothesis will be further tested in comparison with the signals of the non-Sahel region. Furthermore, the PC1 and EOF1 signals are consistent with the analysis over the whole continent, which suggests the main surface heat response within the region is generally well represented in our previous study, with some slightly differences in the percentages of variance of the components, and in the timing of the peaks

Moreover, looking at the spatial distributions of the PC2 signals in Fig. 6.19(c) & 6.19(d), it is clear that the PC2 signal in Sahel over the desert surface remains consistent with our previous desert analysis, whilst contributing to a larger percentage of variance. This is due to the fact that we have removed a large portion of all year dry desert (Sahara) from our analysis; therefore, emphasizing the importance of clouds in PC2 signal, hence, it contributes to more in variance.

6.4.3. Seasonal/yearly variability

This subsection investigates some of the seasonal variability in the Sahel region over the various surface types. The seasonal variability is expected to change the SZA

and thus the surface heat response cycle components of the analysis. It is also expected that the cloud related cycles would change as the Sahel overlaps with the ITCZ, AEJ and AEW in this particular season and consequently receives more cloud occurrences and precipitation than other seasons. The regional specific analysis is intended to highlight some of the variability associated with this change.

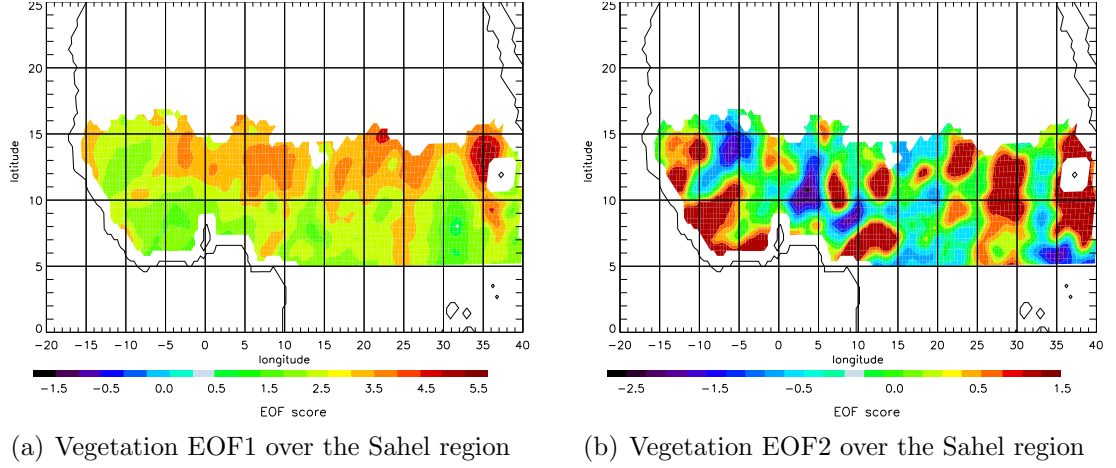


Figure 6.20.: EOFs corresponding to some of the PCs demonstrated in fig. 6.18 along Sahel region for the whole month June 2006

Fig. 6.20 show the EOF1 & 2 for June 2006 and are associated with the PCs shown in 6.18. In the EOF1 plot, a general positive signal is observed in all regions which represents the surface heat response. However, in comparison to the same plot in January 2006 (fig. 6.19), there appears to be a drop in the overall OLR PC1 magnitude in most regions even though the boreal summer provides more incoming solar radiation for the Sahel region in June. The magnitude of the diurnal variation caused by this component is around $15Wm^{-2}$ near the southern Sahel and around $30Wm^{-2}$ in the north, except for the strong signal observed over the Ethiopian highland. Over this highland, there exists a strong EOF1 and 2 signals in June 2005/2006, this is a persistent signal in boreal summer over this region. This has also been observed in [Yang and Slingo, 2001] and [Comer et al., 2007], they have described this as a topography induced convection.

The general magnitude of EOF1 over this region is lower in June than January; this can be explained by the associated clouds and precipitation occurrences in the region which inhibits the strength of pure surface heat cycle. This is observed and

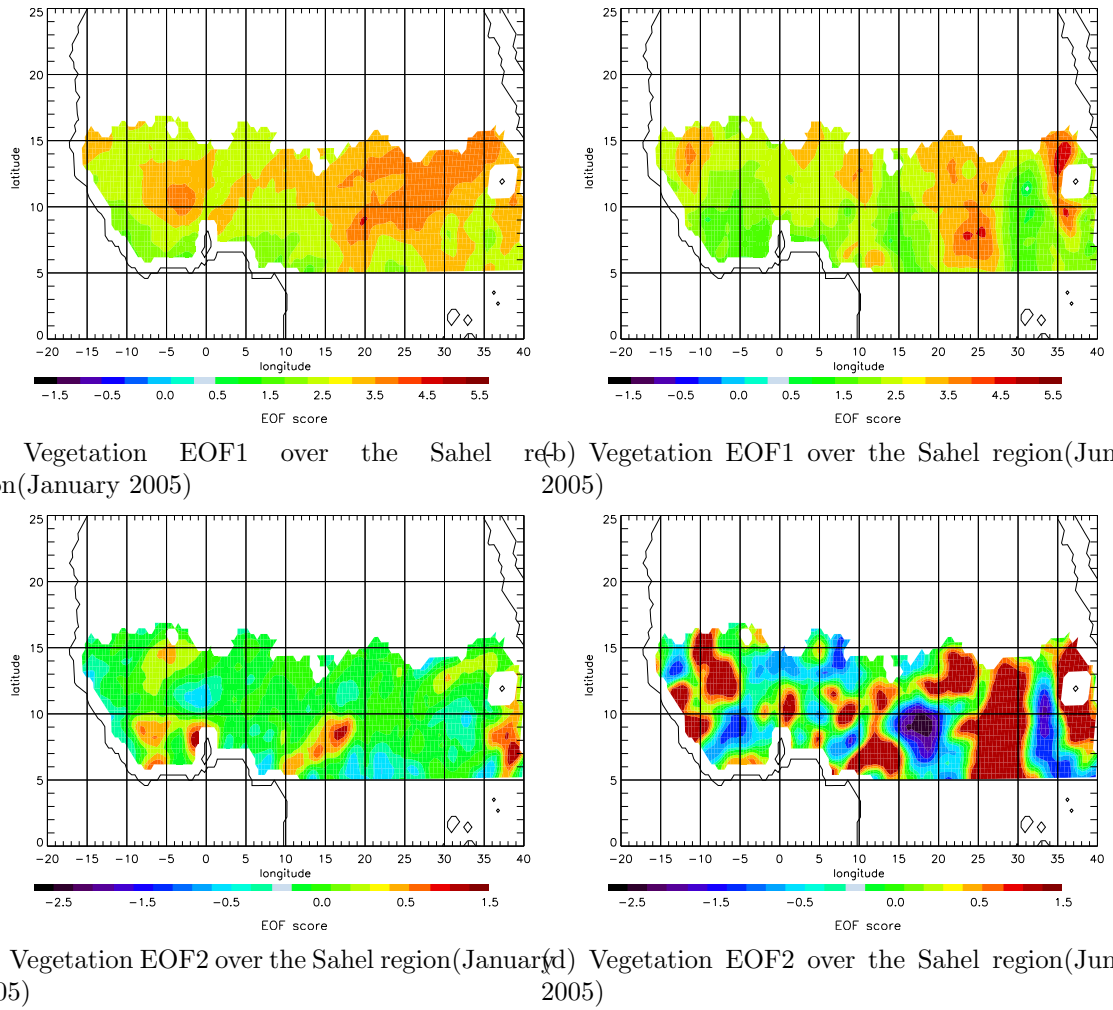


Figure 6.21.: EOFs corresponding to some of the PCs demonstrated in fig. 6.18 in the Sahel region for January and June 2005

confirmed in EOF2 (fig. 6.20(b)) where there appears to be some strong signals, particularly in comparison to the EOF2 in January (fig. 6.20(b)). This again suggests that cloud signals are dominant during the Sahel in this particular season.

Fig. 6.21 demonstrates the EOFs associated with fig. 6.18 for 2005 in comparison with the 2006 EOFs we have analysed in this chapter. In Fig. 6.21(a) and Fig. 6.21(c), the EOFs shown are fairly similar to 2006, for the January 2005 EOF1, the vegetative region in Sahel demonstrated mostly positive signal which indicates a strong surface heat response in most of the region. In contrast, the first EOF for June 2005 shows a weaker signal in general, compared with January 2005; this

combined with a much more sporadic pattern in the EOF2 signal, confirmed the presence of clouds in the region due to the ITCZ during boreal summer.

6.4.4. Summary

In this section, we have analysed the Sahel region exclusively to observe some of its diurnal variability. We have looked at both the desert and vegetation surface types and found the desert analysis to be very similar to the analysis over the whole continent. Furthermore, we have seen the same advection pattern in the desert (northern Sahel) as described in section 6.1.2.

We have also found a set of PCs for this region which are much more similar to our model PCs compared to the analysis over the continent. The difference in the OLR peaks between the vegetation and desert surfaces were found to be around 1 hour in both the Sahel and our model analyses.

We have also observed the spatial distribution of the associated EOF patterns across this region and their variation in boreal summer and winter. We have found that during June (summer), the ITCZ overlaps with the Sahel and creates a lot of convective activities that were observed in EOF2; this in turn suppresses the magnitude of the clear sky surface heat response signal which is represented by a weakening of EOF1 in this region. In January, this region is mostly dominated by the clear sky heat response signals (PC1).

In comparison with the analysis over the whole continent, this regional analysis performed much better in isolating the surface heat and cloud components individually, i.e. we see low/no signal of PC2 over the Sahel vegetative surface in January. It suggests that this set of PCs give a much better description of this region in terms of diurnal variability.

6.5. Analysis on vegetation outside the Sahel

The analysis on the vegetative surface type outside the Sahel is intended to highlight some of the variability in this particular region and to confirm the hypothesis that the dominant PC1 signal in the bright vegetation analysis originates from the Sahel region. Thus, we expect to find a first component which would be quite different

from the analysis over the whole continent.

We have already found a set of PCs which are better suited to describe the variability in the Sahel and are able to isolate the individual components. Therefore, by doing this analysis, we hope to achieve the same for the vegetation surface outside of the Sahel.

We have removed the points close to the coast to avoid any coastal effects from dominating our PCs. We have also eliminated the desert surface type from this regional analysis as there aren't enough data points left in this region to obtain a significant set of PCs.

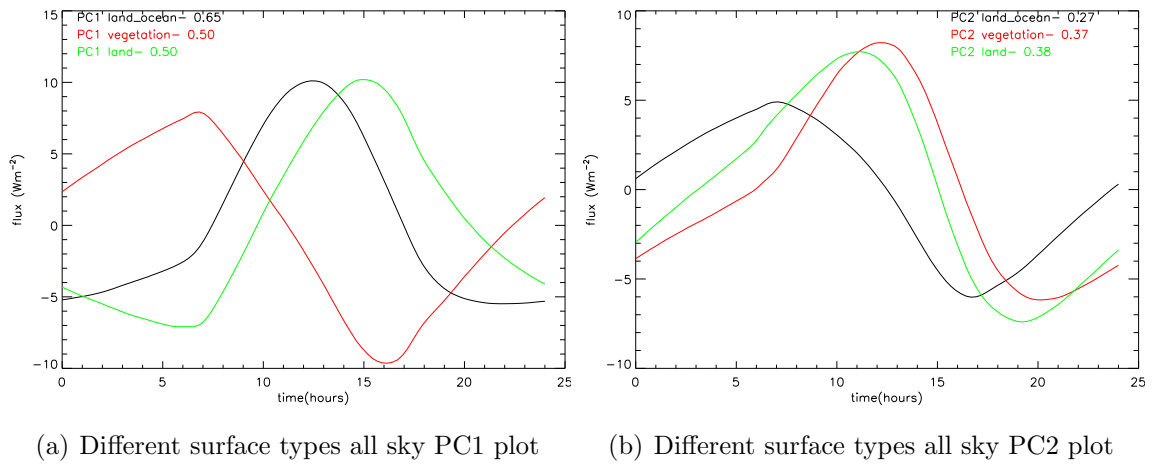


Figure 6.22.: a)PC1 b)PC2 for different surface types outside the Sahel (5°N-30°S): land, vegetation, land + ocean. Numbers on top right corner of each plot indicate the fraction of variances for the individual components. The change in Outgoing Longwave in Wm^{-2} and local time in hours are represented by y-axis and x-axis in the plots respectively. Different colours represent the PC for the various surface types. Months [01, 06, 07, 12] between 2004 and 2007 from GERB LW data are all used to compute this PCs

Fig. 6.22 demonstrates the PCs associated with various surface types outside the Sahel region (5°N-30°S). The desert PC in this particular region should be disregarded as there are insufficient desert points due to the fact that most desert points in the region come from the Kalahari desert neighbouring the coast and are therefore eliminated from our analysis.

In fig. 6.22, the first PCs in the region are looking very different from our Sahel

and our larger scale analysis, which confirms our hypothesis which is the Sahel region dominates the first PC signals over the bright vegetation in the African region.

The first vegetation PC shows a diurnal cycle which resembles a cloud cycle similar that of PC2 in the Africa and Sahel analyses. It indicates this surface type in this region is dominated by cloud influences. The second vegetation PC is unlikely to be a pure surface heat response signal judging by the asymmetry of the curve, suggesting this PC is also influenced by clouds.

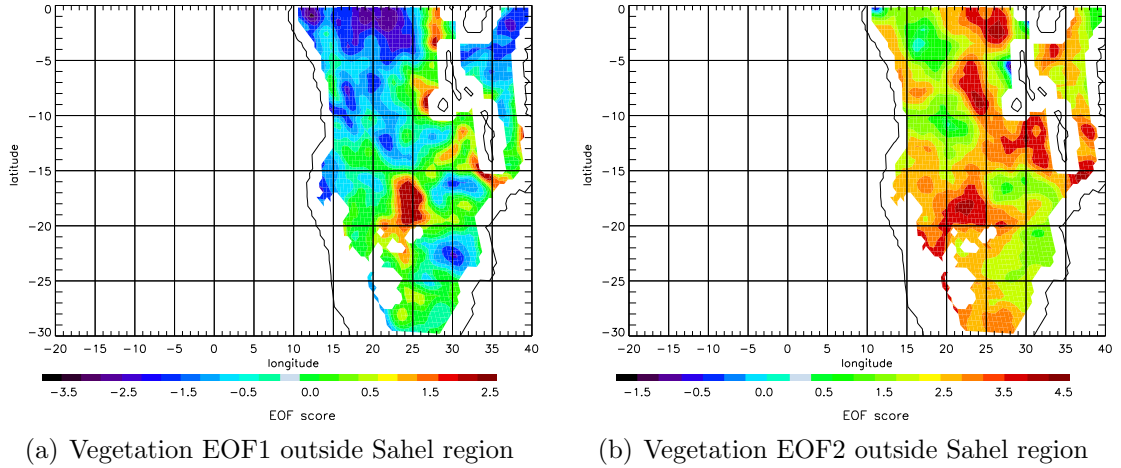


Figure 6.23.: EOFs corresponding to some of the PCs demonstrated in fig. 6.22 outside the Sahel region for the whole month January 2006

Fig. 6.23 demonstrates the EOF spatial distribution of the PCs in fig. 6.22. The first EOF has shown a large region of negative signal in the west whilst the eastern region shows positive EOF1 signal.

If we compare the EOF2 plot from all sky (corresponds to the PC2 in section 5, fig. 5.1) and the EOF1 plot from fig. 6.23(a), we can observe a similar distribution of powers with smaller area of positive signals in the eastern region. If we consider: 1) the positions of these EOF1 signals in fig. 6.23(a), 2) the overlapping of such signals with the EOF2 in fig. 6.23(b), 3) the shapes of the associated PCs, their combined shape to form an OLR cycle which peaks in the morning and minimises in the afternoon; we can attribute such signals to cloud dominated OLR cycles. Whereas the combined signal of the western region of EOF1 & 2 indicates that these signals are more likely to be surface heat responses interlaced with occasional cloud occurrences.

6.6. Chapter Summary

In this chapter, we have investigated some of the diurnal variability over tropical Africa and separated some of the signals by surface types and region (Sahel). We have also developed a simple single layer atmosphere surface radiation model and used our model to describe some of the variability observed in the PCs.

We have analysed diurnal variability separated by surface types using PCA in the beginning of this chapter, and we have found the first components of these surfaces to be mainly effected by clear sky surface heat response. However, we have observed differences in the timing of the OLR peaks with respect to different surface types. We have seen similar differences in the peaks of PC2 over various surface types, which is reasonable given the relationship between surface heat response and convective activities; i.e. when surface temperature peaks later in the day, the corresponding convective cycle also delays by the same amount. We have observed a set of PCs which describe the diurnal variability across the different surface types, this has not been previous explored by *Comer et al.* (2007) or *Smith and Rutan* (2003). It provided an insight into how the diurnal variabilities differ across the different surface types and how they are manifested in the spatial distributions.

For example, we have seen advection over the desert from our OLR PCA which has not been previous observed. This is further demonstrated using the non-averaged PCA dataset and confirmed with the wind dynamics in the tropical Africa. (see section 6.1.2)

We have hypothesized a few factors which could influence the clear sky heat response: the emissivity, surface moisture, atmospheric humidity, and the heat capacity. We have determined the heat capacity to be the most viable candidate for the delay due to its direct proportionality with rate of change in temperature and the fact that HC could double across different surface types over the continent.

A simple radiation model was developed with a single layer of atmosphere and a radiative surface (see description in section 3.7). We have found from our surface type analysis and model study that the surface heat capacity is one of the key components in explaining the time delay between first PCs across different surface types. However, the surface alone only accounted for roughly 1 hour of delay, whereas, in the GERB data analysis, we have observed up to 2 hours of delay. According to

our model sensitivity test, it would require a heat capacity well over $3000 Jkg^{-1}K^{-1}$ (c.f. water $HC = 4190 Jkg^{-1}K^{-1}$) to produce the delay that we have observed in the surface type analysis. This would require all the vegetative surface to be completely saturated in water for this range of HC. This is an unrealistic assumption, therefore, we explored the possibility of other causes mentioned above.

One possibility for this difference can be explained by the available moisture of the surface types convecting into the troposphere to produce a MTH and UTH cycle described by *Chung et al.* (2007). The UTH and MTH cycles maximise just after midnight and minimise in the late afternoon, which increase the OLR in the late afternoon observed from the TOA. In our model, we have also found that the latitude has an influence on the magnitude of the OLR but very little effect on the timing of maximum OLR as shown in section 6.2.2.

This difference could also be caused by emissivity of the surface, which has a positive influence on the time delay. For example, an increase in emissivity decreases the surface emitted radiation and consequently increases the rate of change in temperature. Another possibility for the discrepancies between the GERB data time delay and our model time delay, could be due to the surface moisture which could influence the surface heat capacity and evaporate to induce the humidity mechanism described above. As discussed above, surface moisture alone is not enough to provide the range of HC ($3000 Jkg^{-1}K^{-1}$) that is required for the change in the delay. Therefore, we expect the atmospheric moisture to have a role in this delay.

Thus, we have chosen to separate the Sahel for a regional analysis due to the latitudinal constraint that could provide us with a dataset which has atmospheric conditions and precipitation across the whole region. In our Sahel analysis, we have found a set of PC1s which have much closer peaks, roughly 1 hour between the desert and vegetation surface types. This is very similar to our model interpretation of the surface heat response. This serves as an initial test for our hypothesis where the atmospheric humidity cycles and the surface humidity are the possible causes for the delay between the timing in the OLR peaks across the surface types. This test had shown consistent delay in OLR peaks between our model and the Sahel PCs.

Following on from this initial test, it would be logical for us to introduce atmospheric humidity and surface humidity into our model, perhaps by drawing a simple

relationship between the average monthly rainfall and surface humidity. Then, introduce either a fixed or parametrised atmospheric humidity cycle in our one layer atmosphere by following the diurnal cycle of UTH and MTH presented in [*Chung et al.*, 2007]. This will be discussed further as part of the future work in chapter 7.

As a comparison, we have investigated the diurnal variability in the vegetative region outside of Sahel (in southern Africa), we have found both the first and second PCs to be heavily influenced by clouds and convection, due to the ITCZ over this region. This implies the cloud activities are the dominant signal in the region, thus manifesting itself as the first component in the PCA. This predominant cloud effect suppresses the surface heat signal over this region in general, imposing an influence on the second PC.

7. Conclusion & Future Work

7.1. Conclusion

In this thesis, we have identified and explained some of short term OLR variability associated with tropical climate in the African and Atlantic region. We have used geostationary satellite data (GERB) to examine some of the shorter timescale radiative variability. This instrument provides a high time resolution dataset in the broadband OLR channel which is an improvement over other studies (except [*Comer et al.*, 2007], same data).

We have identified some of the most significant modes of variabilities in chapter 4 using Fourier analysis, the data was averaged from GERB resolution to $2\times 2^\circ$, then the averaged time series is Fourier transformed into the frequency domain. The frequencies are then used to determine the dominant modes, we have found the semi-diurnal and diurnal modes to be dominant over the land regions, whilst the longer timescale (2-10 days) variability were expressed near the ITCZ and is partially contributed by the wave disturbances caused by the African Easterly Jet which is formed by the large latitudinal temperature gradient between the Sahara desert and the Gulf of Guinea.

We have observed the semi-diurnal cycle to be primarily distributed over the land, to resemble the pattern of the diurnal cycle. Thus, we created a simple surface radiation model. We found the ratio of semi-diurnal to diurnal for the observed and modelled data were found to be very similar, at a ratio of 1:3. This suggests that the semi-diurnal mode can be created with ISR absorption coupled with surface radiation and is a manifestation of the diurnal component in the semi-diurnal Fourier timescale where a single frequency of sine wave is insufficient in describing the asymmetry in the diurnal cycle.

In chapter 4, the diurnal component was observed to be dominated by land pro-

cesses largely due to the surface response to the incoming solar radiation. We have noticed that during boreal winter and summer the spatial distributions of the strongest diurnal signal changes according to the position of the SZA and the signal strength was suppressed in the ITCZ. This is consistent with previous understanding (*Slingo et al.*, 2004; *Yang and Slingo*, 2001) of this particular mode of variability.

On the other hand, the longer timescale variability appears to be predominantly a cloud/convection mode according to our clear/cloudy sky conditional tests. We have also confirmed the presence of the 2-9 days variability in our Fourier analysis. These modes have also been observed and analysed by various studies (*Mekonnen et al.*, 2006; *Gu et al.*, 2004; *Grist*, 2002 and *Diedhiou et al.*, 1999) and they have attributed such variability to the AEW. Although we cannot confirm the wave to be the cause of our observations purely from a Fourier analysis, we have observed magnitudes of these modes to be consistent with cloud manifestation and the modulation of these modes in spatial dimension are in agreement with these previous studies.

In chapter 5, the diurnal component was further investigated by decomposing the OLR into subsets of diurnal variation according to the percentage variance of the eigenvectors by utilising Principal Component Analysis. A similar experiment has been conducted by *Comer et al.* (2007) and *Smith and Rutan* (2003) on a large spatial scale and our results demonstrated very similar results which confirms the validity and robustness of the technique over a smaller region. The results have shown that the first PC is dominated by the surface heating response in clear sky condition which contributed to 77% , 60% of the overall variances in June and January 2006 respectively. Whereas the second PC can be attributed mainly to the diurnal cloud/convective events which contributed to 15% and 30% of the total variances in June and January 2006 respectively.

We have taken the analysis further from previous work (*Comer et al.*, 2007 and *Smith and Rutan*, 2003) by a separation of data according sky conditions (clear and cloudy sky) in chapter 5 and shown that surface heat and its semi-diurnal modulation are robust across different scenes. Whilst, the cloud induced all sky PC2 can be eliminated by the removal of cloudy sky conditions.

These results are novel and not only do they show the robustness of the observations and how well the PCA isolate certain diurnal signals. They also demonstrate that within the selected data, the components and their variance contributions are

extremely robust from month to month and year to year. This tells us that the main diurnal variations (i.e surface heat and clouds) themselves are invariant across seasons in this region, instead the season variability in the diurnal cycles arise from the manifestation of these components in spatial dimension (e.g. the ITCZ shifts in latitude, creating more or less cloudy points depending on the amount of land/ocean it covers).

In our last results chapter (6), we have extended the analyses presented in (*Comer et al.*, 2007; *Comer*, 2007; *Smith and Rutan*, 2003) by investigating the diurnal variability across different surface types and developed a simple model to try to describe some of these differences. We have discovered a time delay in the OLR peaks between the different surface types in our OLR PCA analysis, which has not been discussed in previous studies.

In general, the PC1/EOF1 gave a flux variation of around $30\text{-}60\text{Wm}^{-2}$ which is consistent with diurnal variation of clear sky signal that we have previously observed. On the other hand, PC2 has shown a lot of variance across the surface types, for example, over the desert surface, this component only constitute to about 10Wm^{-2} at maximum due to: 1) this region is predominantly dry which has minimal cloud development 2) intermittent cloud advection which averages over the month to a low magnitude. Whereas, over the vegetation, regions of $30\text{-}45\text{Wm}^{-2}$ was observed, suggesting a strong cloud influence over this surface type.

Moreover, the delay between the OLR maxima across different surface types was observed to be up to 2 hours between the desert and vegetated surface types. This novelty result is the key motivation in the development/improvement of our simple surface model from chapter 4. Our original model consisted of a surface with a heat capacity receiving the ISR and radiating according to the Stefan-Boltzmann law. We have improved this model to suit our study in our Chapter 6 by introduce a 1 layer atmosphere, a conductive layer below the surface, variable ISR according to the SZA, a fixed emissivity for the layer and the atmosphere (see details in section 3.7).

We have analysed our model data with the same PCA method and the results were found to be slightly different from the our PCA over the African continent. We have only observed a delay of around 1 hour in our model and this is mainly caused by the heat capacity of the surface; as we have found through our model sensitivity

test that the SZA has small influences on the timing of the maximum OLR (15 minutes over 60° in latitude). Through the model and the African continent surface types analysis, we have hypothesized the discrepancies to be influenced by surface and atmospheric moisture (see discussion in chapter 6 and summary 6.6)

Therefore, we have analysed the sahel region on its own as a latitude constraint to minimise the differences in atmospheric conditions and precipitation across the region and we have found the PCs across the desert and vegetation surface in this region to be similar to our model findings. We recognise that this test alone does not confirm our hypothesis on surface and atmospheric moisture effect on the OLR, but it does provide reasonable results for us to devise a series of test to further examine and attribute this effect on the OLR cycle in our future work.

In this thesis, we have investigated some of the most significant short term modes of variability in the OLR channel and associated these modes to some of the most important components in the tropical climate. Studying the shorter timescale variability helps us to further our understanding in the local meteorology as well as larger scale tropical climate events, and it is important to test the climate models under the same mechanism in terms of radiation to ensure not only the radiative outcome is similar but also to ensure the dynamical process associated with the variability is correct.

7.2. Future Work

We have attributed the surface and atmospheric moisture to be a possible cause for the delay in the PC peaks that we have observed in our surface types PCA test. To confirm this hypothesis, we could adapt our model and introduce the atmospheric humidity and surface humidity into it.

This could be done by using precipitation data from other sources, such as ERA, and drawing a simple relationship between the averaged monthly data and the surface humidity. Thus, introducing a fixed surface humidity in our model across the region. Furthermore, we can set a criteria for wet surfaces and assume a part of the moisture to evaporate into the atmosphere; then, either introduce a fixed decrease in the transmissivity as a representation of moisture in the atmosphere, or have a variable diurnal transmissivity as described by [Chung *et al.*, 2007] as the UTH and

MTH.

We can then use the improved model to compare with the surface type analysis over the African continent to check if the discrepancies between model and OLR data diurnal cycle can be described by this effect.

We have tested in our model sensitivity that the latitude effect on the ISR alone does not have a big influence in the timing of the maximum OLR. However, we have also discovered in our Sahel analysis that latitude constraint appears to be an adequate way to group data which are subjected to similar atmospheric conditions and precipitation. A better approach to further classify the atmospheric conditions for the purpose of grouping the data could be to use narrowband instrument such as SEVIRI to categorise the moisture using the water vapour channels ($6.2\mu\text{m}$, $7.3\mu\text{m}$).

Furthermore, we have shown robust diurnal components under clear/cloudy sky conditions over different years and months. We could use the same method to conduct tests on GCMs to compare these important components within the diurnal variability to the observed OLR. This could give an indication of how well model represents these surface heat response and cloud cycles variability across the same region.

A. Appendix

A.1. PCA error criterion

(a) Land PCA error				(b) Ocean PCA error			
PC	Variance %	$\Delta\lambda$	$\delta\lambda$	PC	Variance %	$\Delta\lambda$	$\delta\lambda$
1	0.56	1363.50	11.98	1	0.44	73.91	1.44
2	0.31	1384.07	6.63	2	0.35	215.27	1.13
3	0.06	100.70	1.20	3	0.07	10.36	0.22
4	0.04	146.10	0.81	4	0.05	25.77	0.17
(c) Dark vegetation PCA error				(d) Bright vegetation PCA error			
PC	Variance %	$\Delta\lambda$	$\delta\lambda$	PC	Variance %	$\Delta\lambda$	$\delta\lambda$
1	0.42	97.20	15.91	1	0.54	1352.56	23.05
2	0.40	1693.75	15.19	2	0.33	1748.13	14.10
3	0.07	105.94	2.73	3	0.06	158.71	2.54
4	0.05	184.57	1.95	4	0.03	158.69	1.49
(e) Dark desert PCA error				(f) Bright desert PCA error			
PC	Variance %	$\Delta\lambda$	$\delta\lambda$	PC	Variance %	$\Delta\lambda$	$\delta\lambda$
1	0.57	1049.71	31.53	1	0.67	1050.88	10.63
2	0.31	1028.27	16.94	2	0.18	285.52	2.90
3	0.05	53.56	2.64	3	0.05	32.48	0.80
4	0.03	100.62	1.91	4	0.04	47.66	0.56

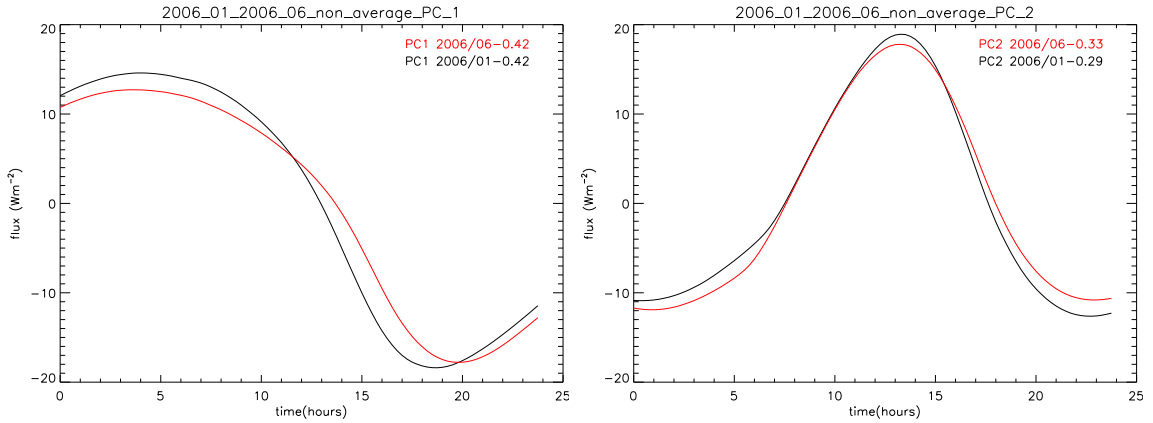
Table A.1.: Principal component(1-4) variances and error criteria for different surface types

A.2. All sky(non averaged) PCA analysis

This section is intended to investigate the decomposition of the all sky data without averaging the diurnal cycles over a month. This part of the work is intended to

highlight the distribution of the variance of the data in a non-averaged manner, especially in events which are suppressed by averaging the data over time.

The data employed in this analysis was all sky non averaged data. However, the data will be divided into daily data and each day of data will be treated as an extra dataset in the spatial dimension. Therefore, the number of data which enter the analysis method will be roughly 30 times the averaged analysis, though the PC have the same dimensions as the averaged analysis. The associated EOF will have equal spatial dimension as the input data.



(a) January, June 2006 all sky non average PC1 (b) January, June 2006 all sky non average PC2

Figure A.1.: First two PCs for a)2006 June b)2006 January non average data, numbers on top right corner of each plot indicate the fraction of variances for the individual components. The change in Outgoing Longwave in Wm^{-2} and local time in hours are represented by y-axis and x-axis in the plots respectively.

In figure A.1, the PCs are shown for non averaged data of January and June 2006. The first PCs in figure A.1(a) are sine wave like with peaks sometime early in the morning at around 0400, the shape resembles that of PC2 in the average all land and ocean plot (see figure 5.1(b)). This indicates the dominant factor in this PC is likely to be cloud development.

On the other hand, the second PCs (figure A.1(b)) peaks at about 1300 and resembles quite closely to the shape of PC1 of the average plot (see figure 5.1(a)), except the shape of non average PC2 is slightly more symmetric, suggesting that the PC consists mostly of the surface heat response. The PC1s in two different months

represent an equal amount of variance, 0.42. This is much higher compared with the PC2 in averaged analysis (0.15-0.30), meaning that cloud events contribute to much more variance in non averaged data.

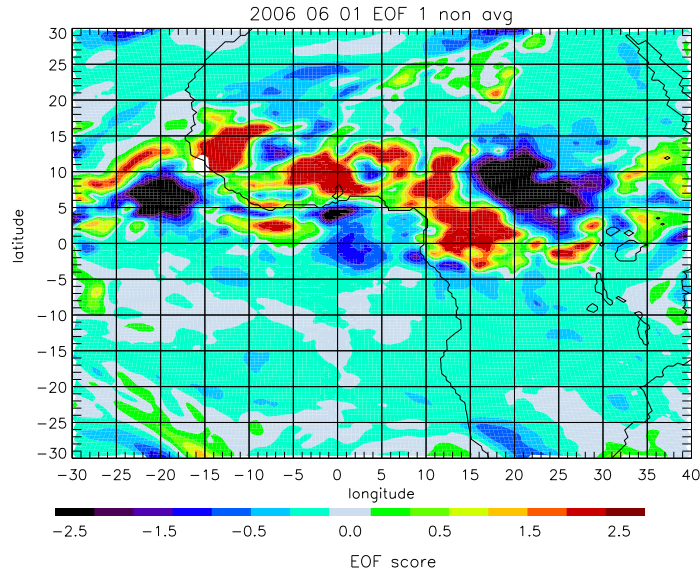
These 2 components are preserved when the analysis shifted from average to non average data, but the variance of these components are different in the two scenarios: non average (0.42, 0.29-0.33), average (0.60-0.77, 0.15-0.30). This suggests the variability caused by clouds along the ITCZ (see figure A.2(b)) is the dominating factor in OLR variability for the region chosen at diurnal timescale, due to possible cloud development events over a large portion of the chosen region.

However, it also implies the cloud events are not as consistent from one day to another over a particular region as surface heat response. Therefore, its affect is diminished in the averaged dataset. It should also be noted that both PC3&4 are semi-diurnal timescale (not shown), and have switched in their importance in variance compared to all sky analysis (figure 5.1) as one is dominated by solar heating response whereas the other is by cloud events as discussed in section 5.2.

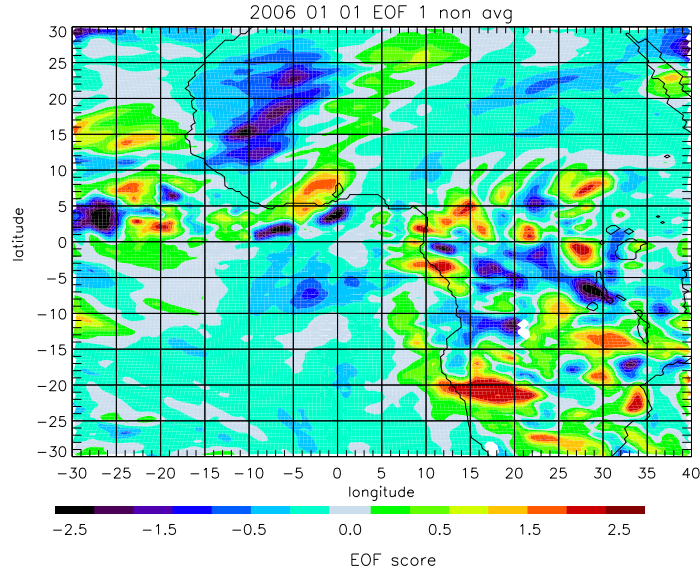
In figure A.2(a), it can be observed from the first EOFs that the extreme signals are concentrated in the ITCZ region. The regions outside of the ITCZ are mostly zero, it should also be noted that the strong positive signals are usually accompanied by neighbouring negative signals, perhaps indicating the advection of clouds across these regions in early morning, creating a minimum in OLR. Hence, is interpreted as a negative signal. Whereas, in figure A.2(b), the strongest signals are found in southern Africa and off the coast of central Africa, which is mostly along the ITCZ. However, the spatial distribution density of the signal is much lower compared with January.

The second EOFs are positive over land in general with the strongest positive and negative signals found over the ITCZ, suggesting that the second PC consisted of a mixture of surface heat response and cloud signals, enhancing the signal by the advection of clouds into these regions in late night or early morning, and vice versa.

The seasonal effect was mostly found to be a difference in geographical distribution due to the change in latitude of the ITCZ and the minimum in SZA.

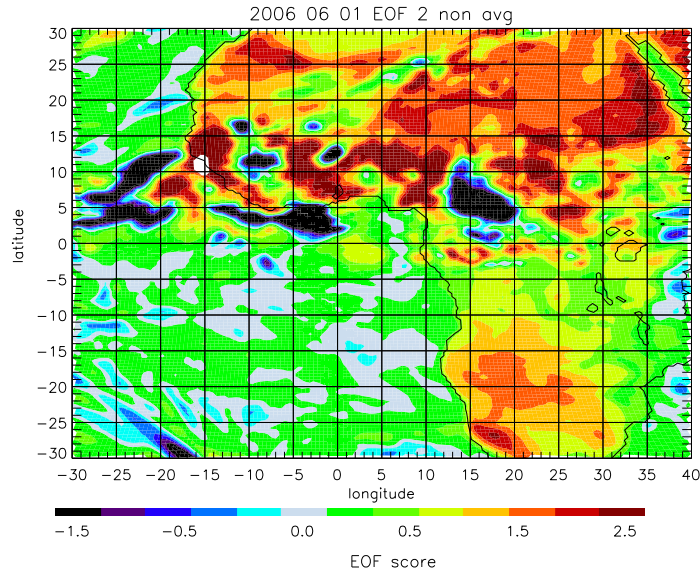


(a) 1st June 2006 all sky non average EOF1 plot

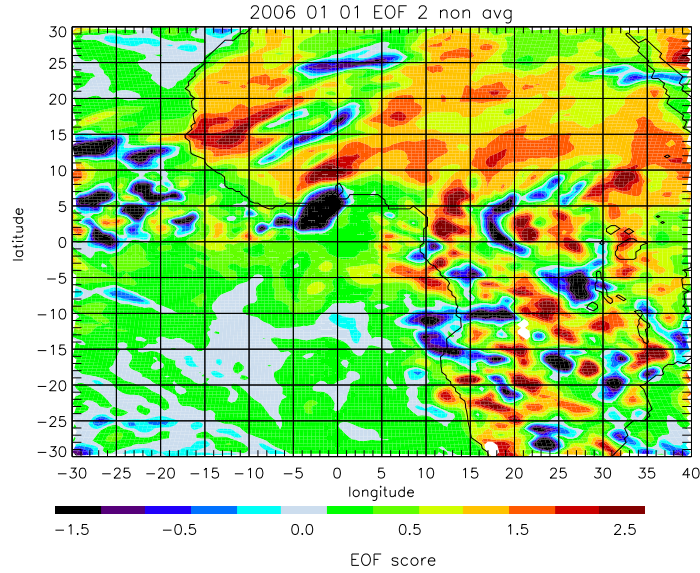


(b) 1st January 2006 all sky non average EOF1 plot

Figure A.2.: EOF1 for a)1st 2006 June b)1st 2006 January non average data, where the colours represent the EOF score over the region 30°S - 30°N and 30°W - 40°E , white indicates data absence. The x and y-axis denotes the longitude and latitude respectively. The EOF score is dimensionless, requires associated PC to interpret the OLR component.

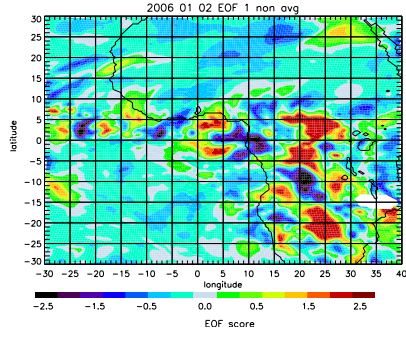


(a) 01 June 2006 all sky non average EOF2 plot

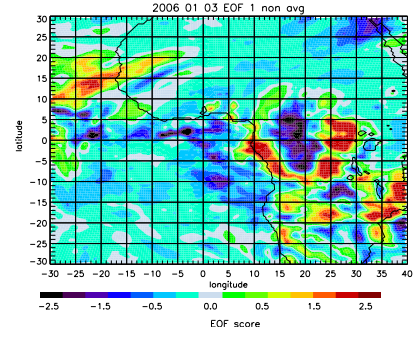


(b) 01 January 2006 all sky non average EOF2 plot

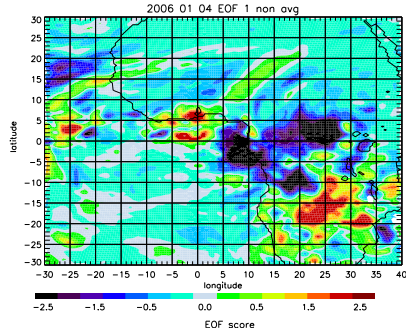
Figure A.3.: EOF2 for a)1st 2006 June b)1st 2006 January non average data, where the colours represent the EOF score over the region 30°S - 30°N and 30°W - 40°E , white indicates data absence. The x and y-axis denotes the longitude and latitude respectively. The EOF score is dimensionless, requires associated PC to interpret the OLR component.



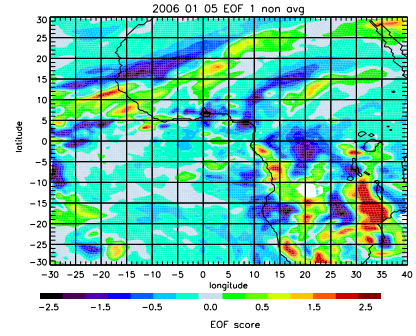
(a) 02 January 2006 all sky non average EOF1 plot



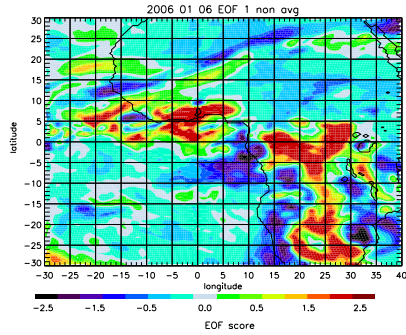
(b) 03 January 2006 all sky non average EOF1 plot



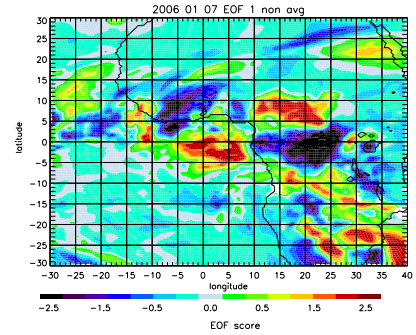
(c) 04 January 2006 all sky non average EOF1 plot



(d) 05 January 2006 all sky non average EOF1 plot



(e) 06 January 2006 all sky non average EOF1 plot



(f) 07 January 2006 all sky non average EOF1 plot

Figure A.4.: EOF1 for 02-07 January 2006(NB. day 1 shown in Fig. A.2(b))

A.3. Model Method for diurnal cycle analysis

The model initiates with a set arbitrary temperature ($293K$) for the surface, the heat capacity is set according to the specific heat capacities for soil, sand and other materials [Arya, 2001]. For example, a desert surface type is interpreted as sand in

the model with a specific heat of $835 \text{ Jkg}^{-1}\text{K}^{-1}$. The ISR time series is taken from GERB shortwave data, the total is then calculated using an averaged earth albedo (0.3) [Houghton, 1986].

The incoming solar flux (F) is converted into energy (E) using the equation:

$$E_{in/out} = F.A.dt \quad (\text{A.1})$$

The area (A) is defined as 1m^2 , and the time (dt) is the model time step (15 minutes) in seconds. The change in temperature (ΔT) is then calculated using the thermal energy (Q) equation:

$$Q = m.c.\Delta T \quad (\text{A.2})$$

where m is the mass of the surface, and c represents the specific heat capacity of the material. The change in temperature (ΔT) is added onto the original temperature for each time step. For example, the temperature after the first time step would be $293\text{K} + \Delta T_1$.

This temperature is then used to calculate the radiative flux (j) using the Stefan-Boltzmann law:

$$F_{out} = \sigma T^4 \quad (\text{A.3})$$

where F_{out} signifies the outgoing radiative flux similar to OLR flux from GERB data and T is the temperature in K . The flux is converted into energy using equation 3.14 and the net energy is recalculated at the beginning of every model time step.

The ISR time series contains one month of data, thus the model was allowed to run during the same length of time. The modelled outgoing flux was found to stabilise after 2/3 days from the initial set up.

Bibliography

- A. Ali and T. Lebel. The sahelian standardized rainfall index revisited. *International Journal of Climatology*, 29(12):1705–1714, 2009.
- R. P. Allan, A. Slingo, S. F. Milton, and M. E. Brooks. Evaluation of the met office global forecast model using geostationary earth radiation budget (gerb) data. *Quarterly Journal of the Royal Meteorological Society*, 133(629):1993–2010, 2007. Part B 255BP Times Cited:7 Cited References Count:34.
- P. S. Arya. *Introduction to micrometeorology*, volume 79. Academic press, 2001.
- B. Barkstrom and J. Hall. Earth radiation budget experiment (erbe): An overview. *Journal of Energy*, 6(2):141–146, 1982.
- R. G. Barry and R. J. Chorley. *Atmosphere, weather, and climate*. Psychology Press, 2003.
- M. Bates, A. Smith, P. Allan, B. Coan, D. Lepine, M. Ricketts, and B. Stewart. Geolocation accuracy of gerb data.
- J. W. Bergman and M. L. Salby. Diurnal variations of cloud cover and their relationship to climatological conditions. *Journal of Climate*, 9(11):2802–2820, 1996.
- G. Berry, C. Thorncroft, and T. Hewson. African easterly waves during 2004 - analysis using objective techniques. *Monthly Weather Review*, 135(4):1251–1267, 2007. 160ER Times Cited:21 Cited References Count:24.
- G. J. Boer, N. A. Mcfarlane, and M. Lazare. Greenhouse gas induced climate change simulated with the ccc 2nd-generation general-circulation model. *Journal of Climate*, 5(10):1045–1077, 1992. Kf433 Times Cited:198 Cited References Count:39.

- K. L. Brubaker, D. Entekhabi, and P. S. Eagleson. Estimation of continental precipitation recycling. *Journal of Climate*, 6(6):1077–1089, 1993. Lk397 Times Cited:163 Cited References Count:48.
- R. D. Cess, E. F. Harrison, P. Minnis, B. R. Barkstrom, V. Ramanathan, and T. Y. Kwon. Interpretation of seasonal cloud-climate interactions using earth radiation budget experiment data. *Journal of Geophysical Research-Atmospheres*, 97(D7):7613–7617, 1992. Hv487 Times Cited:22 Cited References Count:6.
- D. S. Chang-Seng and M. R. Jury. Tropical cyclones in the sw indian ocean. part 1: inter-annual variability and statistical prediction. *Meteorology and Atmospheric Physics*, 106(3-4):149–162, 2010a. 555NR Times Cited:0 Cited References Count:57.
- D. S. Chang-Seng and M. R. Jury. Tropical cyclones in the sw indian ocean. part 2: structure and impacts at the event scale. *Meteorology and Atmospheric Physics*, 106(3-4):163–178, 2010b. 555NR Times Cited:0 Cited References Count:32.
- J. Charney. Tropical cyclogenesis and the formation of the intertropical convergence zone. *Mathematical problems of geophysical fluid dynamics*, 13:355–368, 1971.
- S. S. Chen and R. A. Houze. Diurnal variation and lifecycle of deep convective systems over the tropical pacific warm pool. *Quarterly Journal of the Royal Meteorological Society*, 123(538):357–388, 1997.
- T. C. Chen. Characteristics of african easterly waves depicted by ecmwf reanalyses for 1991-2000. *Monthly Weather Review*, 134(12):3539–3566, 2006. 118AB Times Cited:6 Cited References Count:44.
- E. S. Chung, B. J. Sohn, J. Schmetz, and M. Koenig. Diurnal variation of upper tropospheric humidity and its relations to convective activities over tropical africa. *Atmospheric Chemistry and Physics*, 7(10):2489–2502, 2007. 173WT Times Cited:8 Cited References Count:38.
- C. A. Clayson, B. Strahl, and J. Schrage. 2-3-day convective variability in the tropical western pacific. *Monthly Weather Review*, 130(3):529–548, 2002. 523VA Times Cited:9 Cited References Count:41.

- R. E. Comer. Understanding the diurnal cycle in clouds and water vapour. 2007.
- R. E. Comer, A. Slingo, and R. P. Allan. Observations of the diurnal cycle of outgoing longwave radiation from the geostationary earth radiation budget instrument. *Geophysical Research Letters*, 34(2):–, 2007. 133VA Times Cited:6 Cited References Count:19.
- K. H. Cook. Generation of the african easterly jet and its role in determining west african precipitation. *Journal of Climate*, 12(5):1165–1184, 1999.
- D. Dee, S. Uppala, A. Simmons, P. Berrisford, P. Poli, S. Kobayashi, U. Andrae, M. Balmaseda, G. Balsamo, and P. Bauer. The erainterim reanalysis: Configuration and performance of the data assimilation system. *Quarterly Journal of the Royal Meteorological Society*, 137(656):553–597, 2011.
- T. DeLiberty. Solar and terrestrial radiation, 1999.
- M. Desbois, R. R. L. Eymard, N. Viltarda, M. Violliera, J. Srinivasanb, and S. Narayananc. The megha-tropiques mission. In *Proc. of SPIE Vol*, volume 4899, page 173.
- C. Deser and C. A. Smith. Diurnal and semidiurnal variations of the surface wind field over the tropical pacific ocean. *Journal of Climate*, 11(7):1730–1748, 1998.
- S. Dewitte, N. Clerbaux, L. Gonzalez, A. Hermans, A. Ipe, A. Joukoff, and G. Sadowski. Generation of gerb unfiltered radiances and fluxes.
- A. Diedhiou, S. Janicot, A. Viltard, P. de Felice, and H. Laurent. A fast moving easterly wave of the west africa troposphere. *Meteorology and Atmospheric Physics*, 69(1-2):39–47, 1998. 152KM Times Cited:5 Cited References Count:21.
- A. Diedhiou, S. Janicot, A. Viltard, P. de Felice, and H. Laurent. Easterly wave regimes and associated convection over west africa and tropical atlantic: results from the ncep/ncar and ecmwf reanalyses. *Climate Dynamics*, 15(11):795–822, 1999. 258JX Times Cited:93 Cited References Count:59.
- C. E. Duchon. Lanczos filtering in one and 2 dimensions. *Journal of Applied Meteorology*, 18(8):1016–1022, 1979. Hj974 Times Cited:278 Cited References Count:10.

- GGSPS. Gerb ground segment processing system team webpage. webpage.
- Gille. Details and variations on empirical orthogonal functions. *Lecture*, 2012.
- L. Gonzalez and A. Ipe. Rmib gerb products user guide.
- J. P. Grist. Easterly waves over africa. part i: The seasonal cycle and contrasts between wet and dry years. *Monthly Weather Review*, 130(2):197–211, 2002. 509YL Times Cited:22 Cited References Count:45.
- J. P. Grist, S. E. Nicholson, and A. I. Barcilon. Easterly waves over africa. part ii: Observed and modeled contrasts between wet and dry years. *Monthly Weather Review*, 130(2):212–225, 2002. 509YL Times Cited:20 Cited References Count:46.
- G. J. Gu and C. D. Zhang. A spectrum analysis of synoptic-scale disturbances in the itcz. *Journal of Climate*, 14(12):2725–2739, 2001. 441PQ Times Cited:18 Cited References Count:60.
- G. J. Gu, R. F. Adler, G. J. Huffman, and S. Curtis. African easterly waves and their association with precipitation. *Journal of Geophysical Research-Atmospheres*, 109 (D4):–, 2004. 780HG Times Cited:11 Cited References Count:33.
- J. E. Harries and C. Belotti. On the variability of the global net radiative energy balance of the nonequilibrium earth. *Journal of Climate*, 23(6):1277–1290, 2010. 572KH Times Cited:0 Cited References Count:36.
- J. E. Harries, J. E. Russell, J. A. Hanafin, H. Brindley, J. Futyan, J. Rufus, S. Kellock, G. Matthews, R. Wrigley, A. Last, J. Mueller, R. Mossavati, J. Ashmall, E. Sawyer, D. Parker, M. Caldwell, P. M. Allan, A. Smith, M. J. Bates, B. Coan, B. C. Stewart, D. R. Lepine, L. A. Cornwall, D. R. Corney, M. J. Ricketts, D. Drummond, D. Smart, R. Cutler, S. Dewitte, N. Clerbaux, L. Gonzalez, A. Ipe, C. Bertrand, A. Joukoff, D. Crommelynck, N. Nelms, D. T. Llewellyn-Jones, G. Butcher, G. L. Smith, Z. P. Szewczyk, P. E. Mlynchak, A. Slingo, R. P. Allan, and M. A. Ringer. The geostationary earth radiation budget project. *Bulletin of the American Meteorological Society*, 86(7):945–+, 2005. 946PN Times Cited:61 Cited References Count:41.

- I. M. Held and B. J. Soden. Water vapor feedback and global warming 1. *Annual review of energy and the environment*, 25(1):441–475, 2000.
- J. Herrero and M. Polo. Parameterization of atmospheric longwave emissivity in a mountainous site for all sky conditions. *Hydrology and Earth System Sciences*, 16(9):3139, 2012.
- P. Hignett. Observations of diurnal variation in a cloud-capped marine boundary layer. *Journal of Atmospheric Sciences*, 48:1474–1482, 1991.
- J. R. Holton. *An introduction to dynamic meteorology*. International geophysics series. Academic Press, San Diego, 3rd edition, 1992. 91040568 James R. Holton. ill. ; 24 cm. Includes bibliographical references (p. 493-497) and index.
- J. T. Houghton. *The physics of atmospheres*. Cambridge University Press, Cambridge Cambridgeshire ; New York, 2nd edition, 1986. 86006875 John T. Houghton. ill. ; 26 cm. Bibliography: p. [252]-254. Includes index.
- IPCC. Ipcc report 2007. *Tcec*, page 790, 2007.
- ITTVIS. Idl (interactive data language) help file by itt visual information solutions in version 7.1. 2009.
- R. Kandel, J.-L. Monge, M. Viollier, L. Pakhomov, V. Adasko, R. Reitenbach, E. Raschke, and R. Stuhlmann. The scarab project: Earth radiation budget observations from the meteor satellites. *Advances in Space Research*, 14(1):47–54, 1994.
- J. T. Kiehl and K. E. Trenberth. Earth’s annual global mean energy budget. *Bulletin of the American Meteorological Society*, 78(2):197–208, 1997. Wq051 Times Cited:251 Cited References Count:47.
- A. G. Laing, R. Carbone, V. Levizzani, and J. Tuttle. The propagation and diurnal cycles of deep convection in northern tropical africa. *Quarterly Journal of the Royal Meteorological Society*, 134(630):93–109, 2008. Part A 349PF Times Cited:20 Cited References Count:71.

- C. Lavaysse, A. Diedhiou, H. Laurent, and T. Lebel. African easterly waves and convective activity in wet and dry sequences of the west african monsoon. *Climate Dynamics*, 27(2-3):319–332, 2006. 070CB Times Cited:9 Cited References Count:36.
- M. Leroux. *The meteorology and climate of tropical Africa*. Springer, 2001.
- R. S. Lindzen. The interaction of waves and convection in the tropics. *Journal of the Atmospheric Sciences*, 60(24):3009–3020, 2003. 750FY Times Cited:13 Cited References Count:43.
- N. G. Loeb, S. Kato, and B. A. Wielicki. Defining top-of-the-atmosphere flux reference level for earth radiation budget studies. *Journal of Climate*, 15(22):3301–3309, 2002. 609CJ Times Cited:13 Cited References Count:16.
- B. Mapes, S. Tulich, J. Lin, and P. Zuidema. The mesoscale convection life cycle: Building block or prototype for large-scale tropical waves? *Dynamics of Atmospheres and Oceans*, 42(1-4):3–29, 2006. Sp. Iss. SI 108XJ Times Cited:44 Cited References Count:69.
- A. Mekonnen, C. D. Thorncroft, and A. R. Aiyer. Analysis of convection and its association with african easterly waves. *Journal of Climate*, 19(20):5405–5421, 2006. 101QQ Times Cited:25 Cited References Count:34.
- S. E. Nicholson. A revised picture of the structure of the "monsoon" and land itcz over west africa. *Climate Dynamics*, 32(7-8):1155–1171, 2009. 435ZN Times Cited:1 Cited References Count:83.
- S. E. Nicholson, A. I. Barcion, M. Challa, and J. Baum. Wave activity on the tropical easterly jet. *Journal of the Atmospheric Sciences*, 64(7):2756–2763, 2007. 194HN Times Cited:5 Cited References Count:26.
- G. R. North, T. L. Bell, R. F. Cahalan, and F. J. Moeng. Sampling errors in the estimation of empirical orthogonal functions. *Monthly Weather Review*, 110(7):699–706, 1982.

- S. M. Nowicki and C. J. Merchant. Observations of diurnal and spatial variability of radiative forcing by equatorial deep convective clouds. *Journal of Geophysical Research: Atmospheres (1984-2012)*, 109(D11), 2004.
- K. Ogawa, T. Schmugge, and S. Rokugawa. Estimating broadband emissivity of arid regions and its seasonal variations using thermal infrared remote sensing. *Geoscience and Remote Sensing, IEEE Transactions on*, 46(2):334–343, 2008.
- S. Philander, D. Gu, D. Halpern, G. Lambert, N. Lau, T. Li, and R. Pacanowski. Why the itcz is mostly north of the equator. *Journal of Climate*, 9(12):2958–2972, 1996.
- M. Pidwirny. fundamentals of physical geography, 1996.
- M. Pidwirny. Net radiation and the planetary energy balance. 2006.
- A. C. Pike. The inter-tropical convergence zone studied with an interacting atmosphere and ocean model. Technical report, DTIC Document, 1970.
- V. Ramanathan. The radiative and climatic consequences of the changing atmospheric composition of trace gases. *The changing atmosphere*, pages 159–186, 1988.
- R. J. Reed, D. C. Norquist, and E. E. Recker. Structure and properties of african wave disturbances as observed during phase iii of gate. *Monthly Weather Review*, 105(3):317–333, 1977. Cz395 Times Cited:333 Cited References Count:32.
- P. B. Rhines. Encyclopedia of atmospheric sciences - rossby waves. 2002.
- D. P. Rowell and J. R. Milford. On the generation of african squall lines. *Journal of Climate*, 6(6):1181–1193, 1993. Lk397 Times Cited:65 Cited References Count:43.
- H. H. G. Savenije. New definitions for moisture recycling and the relationship with land-use changes in the sahel. *Journal of Hydrology*, 167(1-4):57–78, 1995. Qq227 Times Cited:51 Cited References Count:20.
- J. Schmid. The seviri instrument.

- J. M. Schrage, C. A. Clayson, and B. Strahl. Statistical properties of episodes of enhanced 2-3-day convection in the indian and pacific oceans. *Journal of Climate*, 14(16):3482–3494, 2001. 464WF Times Cited:3 Cited References Count:24.
- A. Slingo, K. I. Hodges, and G. J. Robinson. Simulation of the diurnal cycle in a climate model and its evaluation using data from meteosat 7. *Quarterly Journal of the Royal Meteorological Society*, 130(599):1449–1467, 2004. Part B 832XF Times Cited:22 Cited References Count:24.
- G. L. Smith and D. A. Rutan. The diurnal cycle of outgoing longwave radiation from earth radiation budget experiment measurements. *Journal of the Atmospheric Sciences*, 60(13):1529–1542, 2003. 685UD Times Cited:10 Cited References Count:23.
- L. I. c. Smith. A tutorial on principal components analysis. 2002.
- D. Sundararajan. The discrete fourier transform: theory, algorithms and applications. 2001.
- E. H. Taleb and L. M. Druyan. Relationships between rainfall and west african wave disturbances in station observations. *International Journal of Climatology*, 23(3):305–313, 2003. 654HV Times Cited:8 Cited References Count:12.
- C. M. Taylor and D. B. Clark. The diurnal cycle and african easterly waves: A land surface perspective. *Quarterly Journal of the Royal Meteorological Society*, 127(573):845–867, 2001. Part A 429DD Times Cited:19 Cited References Count:57.
- C. D. Thorncroft, N. M. J. Hall, and G. N. Kiladis. Three-dimensional structure and dynamics of african easterly waves. part iii: Genesis. *Journal of the Atmospheric Sciences*, 65(11):3596–3607, 2008. 374VZ Times Cited:14 Cited References Count:37.
- B. J. Tian, B. J. Soden, and X. Q. Wu. Diurnal cycle of convection, clouds, and water vapor in the tropical upper troposphere: Satellites versus a general circulation model. *Journal of Geophysical Research-Atmospheres*, 109(D10):–, 2004. 823KQ Times Cited:48 Cited References Count:47.

- K. E. Trenberth, J. T. Fasullo, and J. Kiehl. Earth's global energy budget. *Bulletin of the American Meteorological Society*, 90(3):311–323, 2009.
- P. M. Udelhofen and D. L. Hartmann. Influence of tropical cloud systems on the relative humidity in the upper troposphere. *Journal of Geophysical Research: Atmospheres* (1984-2012), 100(D4):7423–7440, 1995.
- P. H. Wang, P. Minnis, B. A. Wielicki, T. M. Wong, and L. B. Vann. Satellite observations of long-term changes in tropical cloud and outgoing longwave radiation from 1985 to 1998. *Geophysical Research Letters*, 29(10):–, 2002. 609DL Times Cited:9 Cited References Count:23.
- B. A. Wielicki, B. R. Barkstrom, E. F. Harrison, R. B. Lee III, G. Louis Smith, and J. E. Cooper. Clouds and the earth's radiant energy system (ceres): An earth observing system experiment. *Bulletin of the American Meteorological Society*, 77(5):853–868, 1996.
- T. Wong, D. F. Young, M. Haeffelin, and S. Weckmann. Validation of the ceres/trmm erbe-like monthly mean clear-sky longwave dataset and the effects of the 1998 enso event. *Journal of Climate*, 13(24):4256–4267, 2000.
- R. Worthington. Alignment of mountain lee waves viewed using noaa avhrr imagery, mst radar, and sar. *International Journal of Remote Sensing*, 22(7):1361–1374, 2001.
- G. Y. Yang and J. Slingo. The diurnal cycle in the tropics. *Monthly Weather Review*, 129(4):784–801, 2001. 424VD Times Cited:185 Cited References Count:50.
- J. Yano, R. Blender, C. Zhang, and K. Fraedrich. $1/f$ noise and pulselike events in the tropical atmospheric surface variabilities. *Quarterly Journal of the Royal Meteorological Society*, 130(600):1697–1721, 2004.
- J.-I. Yano, K. Fraedrich, and R. Blender. Tropical convective variability as $1/f$ noise. *Journal of Climate*, 14(17):3608–3616, 2001.
- X. Yin, S. E. Nicholson, and M. B. Ba. On the diurnal cycle of cloudiness over lake victoria and its influence on evaporation from the lake. *Hydrological sciences journal*, 45(3):407–424, 2000.

- G. S. Young, S. M. Perugini, and C. Fairall. Convective wakes of the equatorial western pacific during toga. *Monthly Weather Review*, 123(1):110–123, 1995.
- S. Yuter, R. Houze Jr, B. Smull, F. Marks Jr, J. Daugherty, and S. Brodzik. Toga coare aircraft mission summary images: An electronic atlas. *Bulletin of the American Meteorological Society*, 76(3):319–328, 1995.