The North American Monsoon

by

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ABSTRACT

The North American summer monsoon is documented, using precipitation data together with gridded data for outgoing long-wave radiation (OLR), geopotential height and wind at various levels. The upper level divergence field is diagnosed and compared with the precipitation field. A simple wet-dry precipitation index is used to date the monsoon onset at stations with daily precipitation data.

The analysis shows that the monsoon rains advance northward rapidly from late June to early July. The monsoon onset is accompanied by the development of a pronounced anticyclone at the jet stream level, by sea-level pressure rises over the southwestern United States, and by decreases in climatological mean rainfall over adjacent regions of the United States, Mexico and the Caribbean. This coherent pattern of rainfall changes, that covers much of North and Central America, is shown to be dynamically consistent with the circulation changes aloft. Hence, the monsoon onset is embedded within a planetary-scale pattern of circulation changes. The demise of the monsoon and the associated upper level anticyclone, which takes place around September of the year, is more gradual than the onset, and it is accompanied by an increase in rainfall throughout much of the surrounding region.

The monsoon exhibits substantial interannual variability with regard to intensity and onset date.
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CHAPTER 1. LITERATURE REVIEW

1.1 MONSOONS

The word monsoon is derived from the Arabic mausim, meaning a season, and was first applied to the seasonal shift in winds that occur over the Arabian Sea (AMS, 1959). A consensus on the definition of the monsoon has yet to be reached (Ramage, 1971). Most researchers would agree that a monsoon is characterized by two different seasonal circulations: a surface "in-flow" into a thermal low during summer; and a surface "out-flow" from an anticyclone during winter. The circulation arises due to differential heating between land and ocean.

During the summer, uneven heating occurs between adjacent land and ocean surfaces because of the higher thermal capacity of the ocean and its greater ability to redistribute the heat to greater depths by radiation and mixing, and its lower Bowen ratio. The difference in temperature gives rise to a thermally direct circulation similar to the mesoscale day-time sea breeze. During winter, the situation is reversed and the land is cooler than the surrounding oceans. This, in turn, creates a circulation similar to the night-time land breeze. However, the space and time scales of monsoon circulations are large enough such that it is affected by the rotation of the planet. The monsoon can be broken into summer and winter components. Summer (winter) monsoons reach their highest intensity near the time of the summer (winter) solstice (Das, 1986).

In order to classify a region as having a monsoon climate, Ramage (1971) proposed that it meet several criteria:

a) the prevailing wind direction must shift by at least 120° between January and July;
b) the average frequency of winds from the prevailing direction in January and July
exceeds 40%;
c) the mean resultant winds in at least one of the months (January or July) must exceed 3 m s\(^{-1}\); and
d) fewer than one cyclone-anticyclone alternation occurs every two years, in either month, in a 5° latitude-longitude rectangle.

Although the first three conditions have subjective numerical thresholds they ensure that winds are of the right direction and strength in winter and summer. The fourth criteria emphasizes that a persistent thermal cyclone in summer is replaced by an equally persistent anticyclone in winter.

Using such criteria about one half of the tropics may be defined as having a monsoon climate (Webster, 1987). Ramage's monsoon area comprises Africa, southern Asia, and northern Australia (Figure 1). This definition does not include precipitation in any manner. Therefore, areas in Africa that receive little summer rainfall would be classified as a monsoon climate, as would India, which receives over 10 m of rainfall per year. Hence, most researchers now add rainfall in some way to their definition of a monsoon climate. The most common addition is that the region should have a pronounced maximum in precipitation during the summer and a minimum during winter. It is the summer rainfall maximum, and its importance to agriculture, that has created such great research interest in monsoons. The majority of work has concentrated on the Indian, or Asian monsoon, because it is the largest of all monsoons.

It is important to note that the beginning of the summer rains does not necessarily coincide with the change in surface wind direction. Thus, the definition of the monsoon onset depends on whether researchers choose the change in wind direction, or the
Figure 1 Global monsoon region as defined by Ramage
(after Ramage, 1971)
beginning of the heavy rain, to indicate the start of the monsoon season. Using the definition based on rainfall the monsoon begins on the southern tip of India near the beginning of June, with a standard deviation in arrival time of only one week (Das, 1986). The earliest recorded arrival time over the last 100 years was on May 7, 1918 and the latest on June 22, 1972 (Joseph et al, 1994). It progresses northwestwards until it reaches the northwestern region of India near the beginning of July. Its retreat from India is complete by the middle of September.

The heavy summer rains are frequently interrupted by periods of light or no rain. These periods are known as "break" monsoons. Such breaks last about one week. The distribution of rainfall varies with topography and the preferred track of rain-bearing systems. For instance, the average seasonal monsoon rainfall over northeastern India is 1650 mm while the extreme northwest receives less than 300 mm (Das, 1986). Another example is the mountain range along western India known as the Western Ghats (highest elevation 1.5 km). Its orientation is such that it lies nearly perpendicular to the southwesterly monsoon winds. Therefore, rainfall can be 5 times greater on the windward side, than the lee side.

To further our knowledge of monsoon climates there have been international research efforts. The earliest was the International Indian Ocean Expedition during 1963-65, then the Indo-Soviet Experiments of 1973, and 1977 (Das, 1986). The largest project was the Monsoon Experiment (MONEX), which began December 1, 1978, and lasted for one year, and coincided with the Global Weather Experiment (WMO, 1982). MONEX had three components:

a) the Asian summer monsoon;

b) the Asian winter monsoon; and
c) the West African monsoon.

An intense observation period took place from January 5 to March 7, 1979 to coincide with the winter monsoon. A second phase took place from May 1 to June 30, 1979 to cover the beginning of the summer monsoon. A few countries extended their special observational programmes to the end of August to cover the later part of the monsoon. Special observations came from research ships and aircraft, in addition to increased radar coverage, and real-time GOES satellite information. All data have been archived and are available at several international data centres. Most of the research literature quoted in this chapter are a direct result of these international efforts.

The large scale features of the Asian monsoon are fairly well known now (Das, 1986). Atmospheric circulation in the form of Hadley cells arise in response to major sources and sinks of heat. When the Sun is overhead of the Tibetan plateau, it acts as a major heat source, and becomes the location of the ascending branch of a Hadley cell (Koteswaram, 1958). The descending branch is over a large surface anticyclone south of the Equator, known as the Mascarene High. In winter the circulation reverses, and the ascending branch is found over Indonesia, and surrounding ocean areas. Here, the release of latent heat by moist convection is thought to be the major heat source. The descending branch during winter is to the north, over the Siberian High.

He et al (1987) analyzed First GARP Global Experiment (FGGE) data for an 80 day period near the onset of the 1979 summer monsoon, and found descent to the north and west of the area of ascent over the Tibetan plateau. Broccoli and Manabe (1992) found similar areas of descent and ascent around the Tibetan plateau using the GFDL Global Climate Model. They hypothesized that drier conditions to the north of the Tibetan plateau were due in part to the subsidence created by the Tibetan circulation.
Krishnamurti et al (1989) examined a particularly dry summer, the 1987 Indian monsoon. They found that the upper-level anticyclone shifted eastward allowing for the presence of a westerly upper wind anomaly into India which brought dry air over the region, inhibiting convection. Under normal circumstances a westerly jetstream is observed north of the upper-level anticyclone while an easterly jetstream is to the south (Soman and Kumar, 1993)

Regional or synoptic scale monsoon features are less well understood, for example:

a) the rainfall variations within the monsoon season;

b) the formation of a low-level monsoon trough along the southern border of the Himalayas, and the thermal low which develops over northwestern India;

c) the extension of the monsoon trough into the Bay of Bengal, and why it is a favoured area for formation of tropical depressions (Das, 1986); and

d) the westward propagation of monsoon depressions and equatorial waves.

Some evidence suggests that depressions and waves may be just remnants of westward moving disturbances that move into the monsoon area, nevertheless they still interact with the monsoon. It is this interaction that is not well understood.

The sources of moisture for the Asian monsoon have always been clear. For example, moisture can come from:

a) rain bearing systems like tropical depressions or easterly waves;

b) the monsoon trough situated along the northeastern border of India;

c) mid-level tropospheric disturbances (near 600 mb) called cyclonic vortices;

d) off-shore vortices that develop due to the enhanced topography of the southwestern coast of India (sometimes the southwest flow does not rise over the mountains but
curves back to form a mesoscale low); e) convective disturbances brought across the Arabian Sea by a low-level jet near the 1.5 km level with wind speeds of 15 m s\(^{-1}\) (Das, 1986).

An example of oceanic influence on the monsoon is the possibility that the Asian monsoon may be affected by El Ninos. During the 120 year period from 1870-1989 there were 22 years in which the onset of the monsoon was delayed by more than 8 days. In 16 of those years there was also a moderate or strong El Nino (Joseph et al, 1994).

Such interannual variability leads to interest in the possibility of monsoon prediction; in particular, the onset of the monsoon, and how much rainfall accompanies it. The focus in work thus far has been to develop regression equations with several predictors, among them the geopotential height and temperature over India in April, and the geopotential height, temperature, and wind at multiple levels over Australia from January to April (Kung and Sharif, 1980). Using such equations Kung and Sharif (1980) were usually able to predict the start of the monsoon, at the southern tip of India, to within 3 days for the period 1966-1980. The worst prediction occurred in 1979 when the monsoon was late by ten days. The prediction of the time of onset has been more successful than for rainfall amounts, but work is continuing (Prasad and Singh, 1992; Hastenrath, 1988).

The Australian northwest monsoon can be seen as an extension of the Asian winter monsoon (Joseph et al, 1991). During the northern hemisphere winter the descending branch of the Hadley cell is over Siberia and the ascending branch is over Indonesia. Thus the sections of northern Australia in close proximity to Indonesia are subject to monsoon rains during their summer season. A dry winter follows, thus satisfying Kendrew's (1961) criteria for a monsoon climate.
The African monsoon was also investigated during MONEX. The West African monsoon is the dominate feature, and it is marked with a reversal of surface winds over West Africa, from northeasterly during the winter season to southwesterly during the summer. Eastern Africa is also thought to have a monsoon climate but its climate is more associated with the movement of the Inter-Tropical Convergence Zone (ITCZ) than an onshore maritime flow (Das, 1986). The ascending branch of the Hadley cell is over the Sahel region with the descending branch off the coast of west Africa, which merges with the southwesterly low-level onshore flow (Das, 1986).

General Circulation Models (GCM) could be extremely helpful in understanding the monsoon. A variety of experiments have already been completed. The most successful have been sensitivity tests to assess the responses of monsoon characteristics to variation of several parameters. Changes in sea-surface temperature, snow cover, and soil albedo have been shown to affect the intensity of the monsoon (Meehl, 1994). A coupled ocean-atmosphere model would help because the Indian Ocean responds readily to the shift in monsoon winds. However, the ocean part of such models is still considered the weak component in such an endeavour (Meehl, 1989). The removal of the Himalayan mountains in a GCM model indicates that the abrupt northward movement of the westerly jet would no longer take place. Instead a more gradual northward movement of the jet, and a slower northward progression of the monsoon rains occurred in the model (Hahn and Manabe, 1975). By reducing the elevation of the Tibetan plateau they found that without this elevated heat source the Hadley circulation of the summer monsoon was much weaker. A serious difficulty in modelling the Asian monsoon arises when recreating the topography of the Himalayans which reach to the 500 mb level. Fennessy
et al (1994) found that sensitivity to the model's topography influenced the rainfall simulation much more than any changes to vegetation, soil moisture, or cloudiness.

In general, monsoons are characterized by large anticyclonic circulations in the upper troposphere, which occur in both summer and winter monsoons (Das, 1986). An easterly jet is observed to the south of the anticyclone, while a westerly jet is to the north (Chen et al, 1989; Soman and Kumar, 1993). A rapid northward movement of the westerly jet is common prior to the onset of the summer monsoon. In some years the northward movement of the westerly jet is delayed, while in other years the anticyclone is not as strong. At the surface a strong thermal low (low-level monsoon trough) is present during summer monsoons and a strong anticyclone during winter monsoons. This allows for the reversal of winds found in monsoon climates. Finally, a predominantly dry winter is followed by a wet summer.

1.2 Arizona or Mexican Monsoon

Monsoons are part of the summer general circulation pattern and occur over most continents (Das, 1986). Ramage's (1971) definition of monsoon climates does not include anywhere in North America. Kendrew's (1961) monsoonal category focuses on the heavy summer rainfall rather than the seasonal shift in winds, and is described as "Heavy summer rainfall and a long dry season centred on winter." Hence, portions of Mexico and Central America satisfy Kendrew's definition. Ramage (1971) does not include Mexico and Central America as a monsoon region because the weak change in surface winds do not satisfy his criteria. On the other hand Arizona is not included in Kendrew's monsoonal category, because it does not have a dry winter. However, the term "Arizona
monsoon" has become entrenched in the research literature to describe summer precipitation in Arizona, and its use will be continued here.

Because Arizona becomes extremely hot in the summer, the consensus in the 1950s was that any moisture advected into the area would rise and condense via low-level convergence due to the intense thermal low. This in turn could produce considerable amounts of convective rainfall. Hence, the search for that source of moisture became paramount. Jurwitz (1953) found that much of Arizona's summer rainfall is restricted to the southeast, which he maintained was due to its proximity to a source of moisture, the Gulf of Mexico. Using a particular case study, Jurwitz concluded that the trajectory for moist air was directly from the Gulf of Mexico along the Texas-Mexico border at the 700 mb level. He also commented that on occasion Arizona can receive considerable precipitation associated with moisture from the remnants of tropical cyclones as they make landfall and dissipate over Mexico, if the upper flow is from the southeast. Bryson and Lowry (1955a, 1955b) analyzed several years of data and came to the conclusion that southeasterly flow around an anticyclone, which establishes itself over the middle of the United States during the height of summer, brings moisture from the Gulf of Mexico into Arizona.

The Bryson and Lowry (1955a, 1955b) hypothesis of moisture transport was accepted until Hales (1972, 1974) proposed a completely different one. Hales (1974) noted that flow from the Gulf of Mexico has to cross extensive terrain, mostly above the 800 mb level, before it arrives in Arizona. This did not converge with the work of Reitin (1960) who discovered that 50% of all precipitable water over Phoenix was located below 800 mb. Therefore, Hales could no longer ascribe to the hypothesis that the Gulf of Mexico was the source region. Based on four case studies, Hales (1972, 1974) found that
the Gulf of California acts as a perfect channel for tropical moisture surges. Under normal conditions, Hales suggested that a balance somehow existed between the hot, dry, thermal low pressure conditions in Arizona, and the cooler, moist, higher pressures across the southern part of the Gulf of California. Thus Hales implied there would be little or no wind flow between the higher pressure to the south and the lower pressure to the north. He did not provide the reasoning necessary to understand such a "balance". However, he acknowledged that a local land-sea breeze phenomenon often brings cooler moist air into the extreme southwest part of Arizona. However, it only extends a few miles inland, as shown by Dodd's (1965) analysis of dewpoint temperatures. Hales theorized that an "imbalance" can be triggered by the passage of a tropical cyclone or easterly wave across the southern part of the Gulf of California. The reasoning is as follows. The tropical disturbance is associated with considerable convective rainfall. The rain cools the airmass below it by evaporation. This cooling causes an increase in surface pressure and induces a "giant sea breeze" towards the thermally-induced lower pressure in Arizona (Brenner, 1974). This imbalance initiates a surge of moisture from the tropical air near the southern part of the Gulf of California, towards Arizona. Two types of surges were noted. A shallow surge below 1500 m (5000 feet) which typically cools southwestern Arizona by 5°C, causing stabilization of the airmass, and thereby, inhibiting thunderstorms. A deeper surge with moisture reaching up to 3600 m (12000 feet) also causes cooling, but according to Hales, increases thunderstorm activity by increasing the moisture available for convection. Either type of surge can extend as far north as the Arizona-Utah border, and fill the thermal low in less than 24 hours.

In a more extensive analysis, Badan-Dangon et al. (1991) used instrumented aircraft and automatic weather stations to study the three-dimensional structure of the
planetary boundary layer over the Gulf of California. They found that the surges last an average of 5 days with southeasterly surface winds of approximately 10 m s⁻¹ within the Gulf of California. Their concept of these events differs from that of Hales. Unlike Hales, they did not find sudden wind shifts associated with the leading edge of the pulse, and the moisture transition is not abrupt but lasts over several days. The results of Badan-Dangon et al. (1991) are the more convincing, because of the limited number of cases in the Hales’ study.

Many have misinterpreted Hales’ study by concluding that the waters of the Gulf of California act as the moisture source for the Arizona monsoon. The surface water temperature of the Gulf of California is approximately 25°C (Badan-Dangon et al., 1991), and above the ocean is a 200-300 m deep marine layer in which dew points range from 18-25°C. Above the marine layer, the air is still relatively moist with dew points ranging from 17-21°C. Given that there is an abundance of moist air in the Gulf of California it seems reasonable that the Gulf itself could provide some of the moisture for the monsoon. However, Hales emphasized that the Gulf of California merely acts as a conduit for moisture advection. He hypothesized the transport of eastern Pacific tropical air from the waters off western central Mexico to Arizona. Both Hales (1974) and Brenner (1974) conclude that the thermal low over Arizona is filled approximately 24 hours after the surge begins, and thereafter, a pressure balance exists between the southern Gulf of California and Arizona.

Several aspects of Hales’ approach remain to be resolved:

a) the postulated initial balance between the thermal low and higher pressures to the south requires more investigation;

b) the forcing mechanism for the surge is uncertain;
c) the duration of the surge is debatable (Hales suggested that after 24 hours the surge should no longer exist, but recent observations show it lasts on average for 5 days (Badan-Dangon et al., 1991));

d) the role of initially cooler air from the Gulf of California, ahead of the warm moist tropical surge of air is uncertain (the cooler air will inhibit or destroy thunderstorm activity by destroying the thermal forcing. If thermal forcing is the only mechanism involved then it will be some time before thunderstorm activity occurs. It is obvious that the surges occur but an abundance of warm moist air is insufficient to produce monsoon type rains. Considerable vertical motion must be involved);

e) the hypothesis does not account for monsoon rainfall in New Mexico, western Texas, southern Nevada, and southern Utah;

f) the extreme southwest of Arizona should experience the monsoon rains since it borders the channel for moisture transport but it is eastern Arizona that receives most of the summer rains and lightning activity (Reap, 1986).

Thus the relationship between Gulf of California surges and precipitation in the southwestern United States is still unclear (Douglas, 1992a and Douglas et al., 1992). However, despite its shortcomings, Hales' hypothesis replaced Bryson and Lowry's (1955a, 1955b) earlier hypothesis, and has remained unchallenged since 1974.

After the debate about the source of the monsoon moisture sources was quelled, researchers turned towards more in depth studies of the Arizona monsoon. In 1975, Hales examined an intense thunderstorm using both radar and satellite imagery. A low-level moisture surge occurred the evening before the event so that most of southwestern Arizona possessed considerable low-level moisture. However, the thunderstorm formed in the east, along the Arizona-New Mexico border, over the Colorado plateau at an
elevation of over 1500 m (5000 feet). It is unlikely that the Gulf of California moisture surge reached this far eastward and to this altitude. Although Hales did not mention it, this suggests the possibility of a different source of moisture for this thunderstorm. The storm moved southwestward off the plateau into the desert and diminished in strength. The cells intensified again as they encountered the coast mountain ranges east of San Diego and broke up later as they moved westward over the mountains. Initially, the moisture surge had little if anything to do with the storm, but it probably had an effect when the storm re-intensified over the area along the coastal mountains. Even though this was a major storm only two stations in Arizona measured precipitation. One measured 80 mm (3.1 inches), equivalent to the yearly total for that station. It illustrates both the "patchiness" of the rain and the low station density in Arizona.

Carleton (1985, 1986) conducted a number of statistical studies relating 3 years of Arizona monsoon rainfall to different atmospheric features, such as the strength of the Arizona thermal low, the intensity of the Bermuda anticyclone, height of the 700 mb surface, eastern Pacific sea surface temperatures, etc. No strong conclusions were reached but in later studies he found that Arizona summer rainfall is related to the position of the 500 mb anticyclone, whose position lies just to the east of Arizona during the height of summer (Carleton, 1987 and Carleton et al., 1990). During wetter Arizona summers the 500 mb anticyclone was found to be displaced farther northward. Carleton surmised that this allowed more moisture to penetrate from the south. He did not mention whether this moisture was from Gulf of California surges or transported around the 500 mb high from the Gulf of Mexico.

Moore et al. (1989) and Adang and Gall (1989) suggest a different way to examine the Arizona monsoon. During summer the Pacific and Atlantic surface
subtropical highs dominate the surface pressure patterns. The North American continent divides the two features. Mexico tends to be directly in between the two flows. On the west is a weak westerly or northwesterly flow, and on the east, an easterly or southeasterly flow often prevails. They suggest that this line of confluence be called the "Arizona monsoon boundary". Their analysis suggests it has many of the same features as a mid-latitude front with wind shear causing instability and possible baroclinic waves. This of course assumes that airmasses can transverse the Sierra Madres mountain range. It is also possible that the line of confluence is artificially imposed by the mountains.

A series of papers by Howard and Maddox (1988a, 1988b), and Maddox and Howard (1988a, 1988b), and Maddox et al. (1992a) made extensive use of satellite imagery to document the convective systems that occur over northwestern Mexico during summer, and their contribution to the Arizona monsoon. An initial finding shows that thunderstorm activity shifts from eastern Mexico to western Mexico by the middle of the summer (Howard and Maddox, 1988b). Their investigation of two case studies of mesoscale convective systems shows that convective instability is similar in strength to that associated with severe thunderstorms over the United States (Maddox and Howard, 1988a). Despite the strong convection they found no obvious large scale forcing for upward motion (Maddox and Howard, 1988b) and hypothesized that mesoscale processes such as sea breezes and valley winds may initiate the development of the convective systems. Their final paper in the series highlights the typical life cycle and motion of mesoscale convective systems over Mexico (Howard and Maddox, 1988a). The main conclusion to be drawn from their series of papers is that the extent of the convection over Mexico is clearly evident which hints that the southwest monsoon is much larger in extent than previously thought and appears to extend well into Mexico.
In order to fully explore such aspects of the monsoon a large scale project was undertaken in the summer of 1990, called the SouthWest Area Monsoon Project, or SWAMP (Meitin et al., 1991), to study:

a) central Arizona thunderstorm environments;
b) monsoon structures and moisture fluxes; and

c) convective systems in Mexico.

From July 9 to August 7, 1990, aircraft transits, radar and satellite images, frequent upper air soundings, lightning data, and continuous recording rainfall gauges were gathered to form the most extensive observational network ever provided for the area.

The SWAMP findings are summarized in a paper by Douglas et al. (1993). They examined both Mexican and United States rainfall and rawinsonde data, as well as satellite imagery to show that the Arizona monsoon is just the northern extent of a more prominent Mexican monsoon. Hales (1974) and Brenner (1974) restricted their analyses to Arizona, and therefore, did not see beyond the border, and missed the connection with Mexican summer rainfall. Examining monthly precipitation data Douglas et al. (1993) developed a monsoonal index, which is the ratio of July, August, and September rainfall totals, to the annual mean precipitation. Their monsoonal index shows a maximum in Mexico near the southern Gulf of California which continues northward along the Mexico coast. It decreases in intensity once it reaches the Mexico-United States border.

Douglas et al. (1993) also used satellite imagery to better define the spatial extent of the Mexican monsoon. Using eight years of data (1985-1992) they created colour composite maps of the frequency of occurrence of infrared cloud top temperatures colder than -38°C. This surrogate for deep convection shows that the convection is prominent
over western Mexico in June, but is not as evident over northwestern Mexico until July, thus corroborating their rainfall analysis.

Douglas et al. (1993) went on to use upper air data to determine the source of moisture for the Mexican monsoon. Rawinsonde data along the northwestern Mexico coast shows the mid-level flow switching from westerly in June to easterly in July. This would support the earliest hypothesis of advection from the Gulf of Mexico (Bryson and Lowry, 1955a, 1955b), but Douglas et al. (1993) is quick to point out that a dew point analysis at both the 700 mb and 500 mb levels shows that eastern Mexico has much drier conditions than western Mexico. They suggest that the higher moisture content at upper levels over western Mexico may be due to ascent associated with the easterly flow over topography, and/or vertical mixing of boundary level moisture by convective scale motion. The lack of rawinsonde data makes it difficult to examine the topographic forcing hypothesis. The hypothesis of low level moisture vertically transported by convection then poses the question of where the low level moisture comes from initially. The Gulf of California could provide the low level source of moisture. Douglas et al. (1993) suggest that local evaporation from the Gulf of California may or may not play an important role but acknowledge that Hales' (1974) hypothesis that the Gulf of California acts as a conduit for moisture transport from the tropical Pacific ocean likely plays an important role. However, as discussed previously, Hales (1974) hypothesis still has a number of deficiencies.

Although Douglas et al. (1993) did not provide an answer to the moisture source of the monsoon, they did identify that the Arizona monsoon has a larger spatial extent with the strongest changes occurring in northwestern Mexico. Given their evidence they
suggest that the Arizona monsoon should be more appropriately named the Mexican monsoon.

A second project occurred in the summer of 1993 called the Experimento Meteorologico del Verano (EMVER) (Douglas, 1993) to supplement the spatial density of upper air observations obtained in the SWAMP project, especially over Arizona.

The most important discovery thus far to arise from these projects was the exact nature of the trajectory of low level flow in the Gulf of California during the monsoon season (Douglas et al., 1991, 1992 and Douglas, 1992a, 1993). A distinct southeasterly flow was evident up the Gulf of California. Surface winds near 9 m s\(^{-1}\) were apparent over the Gulf but just inland along the coast winds were nearly calm. This indicates that in past studies the southeasterly flow may have been missed by the network of coastal weather stations. EMVER balloon observations also noted a strong diurnal change in the winds. In the early morning hours the flow was parallel to the coast but by afternoon shifted to an onshore and upslope flow, presumably due to a strong sea breeze development.

EMVER observations also captured many "Gulf surges" of air up the Gulf of California which are critical to Hales (1974) moisture source hypothesis. The two projects also showed the variability of the monsoon. The observation period of the SWAMP project was unusually wet while the EMVER project was characterized by several extended dry spells. Analysis of the extensive data sets from both the SWAMP and EMVER projects is continuing and there is considerable hope that they will answer many questions about the mesoscale aspects of the monsoon.
CHAPTER 2. OBJECTIVES

2.1 RATIONALE FOR THE RESEARCH

In his monograph "The Climates of the Continents", Kendrew (1922) described various precipitation regimes\(^1\) that exist in North America, but offered little, or no reasons as to why they existed. Since that time several researchers have quantified the various regimes using different statistical techniques, but there is still little if any work done to show why they occur. This thesis was originally undertaken to better understand the summer precipitation regimes that exist in North America, in particular, the initial focus was on the Arizona (or Southwest) monsoon. The research expanded beyond Arizona, as the spatial extent of the monsoon became evident.

2.2 OBJECTIVES OF THE RESEARCH

People living in Arizona and New Mexico are quite familiar with the dangerous lightning storms, strong winds, heavy rains, and flash floods that come with the summer monsoon. Atmospheric scientists are also familiar with the events, but a complete understanding of the monsoon is still not in hand (Douglas et al., 1993). Much of the past research on the Southwest monsoon has focussed on either, case studies of singular events, or mesoscale aspects of the phenomenon (Douglas, 1992b, 1992c). This thesis is a climate diagnostic study of the summer monsoon. The results may be of interest to climate modellers, since the monsoons are an important part of the global circulation. The primary goal is to provide other researchers and forecasters with a large scale view of the

\(^{1}\)The character of the seasonal distribution of rainfall at any place (AMS, 1959)
phenomenon, with a view to its better prediction in the future. In order to accomplish this goal the following specific objectives are undertaken:

a) to document the changes that occur in summer precipitation (spatial changes are well known but temporal changes require further investigation);

b) to examine changes in atmospheric circulation that coincide with the changes in surface precipitation (circulation from the surface to the top of the troposphere will be investigated);

c) relate the changes that occur in atmospheric circulation to changes in surface precipitation;

d) examine independent evidence (outgoing long-wave radiation which is a surrogate for convective precipitation) to corroborate the relationship between changes in atmospheric circulation and changes in surface precipitation;

e) re-examine the spatial scale of the monsoon; and

f) examine the interannual variability of the monsoon.

2.3 OVERVIEW OF THE THESIS

The thesis starts with a review of past literature related to the phenomenon. An examination of monthly precipitation in North America identifies different precipitation regimes, including the summer monsoon. A simple index is used to explore the rapid changes in precipitation associated with the monsoon. Standard meteorological fields are then analyzed, in an attempt to dynamically interpret the summer monsoon. Divergence calculations show that upper-level anticyclones are related to areas of convective activity. Outgoing Long-wave Radiation (OLR) is used to corroborate the relationship. The year-
to-year variability of the monsoon is explored with OLR difference fields. Finally, conclusions and implications of the research are presented.
CHAPTER 3. DATA SOURCES

3.1 PRECIPITATION

Monthly precipitation data used in this study are from the World Monthly Surface Station Climatology (WMSSC) and the United States Climate Division data set, both available from the National Climatic Data Center (NCDC), in Asheville, North Carolina.

The WMSSC contains monthly precipitation time series for over 3000 stations worldwide. NCDC provide quality control by scanning and correcting for several hundred gross errors related to incorrect data entry. Extreme values beyond 5 standard deviations from the long period monthly mean are also inspected. Those that NCDC believe to be obviously the result of publication error are set to "missing".

In addition to the NCDC quality control the following procedures were implemented for the data used in this study. Monthly precipitation data were first filtered to find missing values. An objective check was conducted to compare average annual precipitation to individual anomalously high monthly precipitation. Finally each year, for every station used in this study, was inspected visually by means of a time series of monthly precipitation to highlight possible incorrect values. No additional suspect data were found using these additional techniques.

The United States Climate Division data set was developed by the National Climate Data Center (NCDC). Each individual state was separated into 5 to 10 homogeneous climate divisions, and all available stations within each area were weighted equally and averaged, to produce the mean monthly precipitation for each division. The size of the division depends on the spatial homogeneity of the climate within a state. The borders of the climate divisions are based on station availability, topography, and
differences in monthly station data. The borders changed slightly in 1965 from their original definition in 1951. Obviously, this creates a problem when attempting to evaluate trends in the time series. The changes also create problems in the early part of the record, when some divisions had only one or two stations. However, it is one of the most comprehensive monthly precipitation data sets for the United States and is essentially continuous from 1895 to 1988. Only climate divisions in Arizona and New Mexico were examined in this study.

Daily precipitation data for Albuquerque, Logan, Tempe, and Winnemucca are from the Daily Climate Summary, available from NCDC.

The accuracy of rainfall measurements is hard to determine. The treatment of trace values as zero in the monthly totals can lead to an underestimate of precipitation, as can evaporation from rain gauges. On the other hand, the possibility of rain splashing into the gauge can lead to an overestimate of rainfall. Groisman and Legates (1994) estimate a rain gauge measurement error of 5% during summer over the United States.

3.2 UPPER AIR

Upper air data used in this study were taken from the National Meteorological Center’s (NMC) Northern Hemisphere octagonal grid data set. Gridded fields of monthly mean sea level pressure, 500 mb height, 200 mb height, and daily 250 mb zonal and meridional wind components were used. Mean sea level pressure and 500 mb height fields have the most extensive past coverage starting in January 1946. Obviously, over the decades there have been considerable changes both in the availability of observations, and the objective analysis schemes used to assimilate those observations.
A lack of surface observations in the early part of the record, and a serious lack of upper air observations led to the practice of "bogusing" data. Skilled meteorological analysts constructed "subjective analyses" based on the available data to estimate the pressure or height in data sparse regions. As the number of observations increased over the years, and as remotely sensed data became available in the late 1960s, the practice of "bogusing" was reduced. Hence, the early part of the pressure and height data sets are of questionable accuracy. As time progressed the data sets improved in quality. As to exactly when the data sets improved in quality enough to justify accepting them without question is not known, and some may argue that the time has yet to come.

Objective analysis schemes changed considerably over the period of the data sets. One of the first schemes used at the National Weather Analysis Center (the forerunner to the NMC) employed a least squares fitting method to infer grid point values from surrounding observations. In practice, the method employed a minimum of ten observations (Dey, 1989), which meant it performed poorly in data sparse areas. The technique also failed to preserve temporal continuity in analyses, since each analysis was constructed without prior knowledge of preceding data. This often led to substantial errors in the analysis in data sparse regions. In 1958, NMC changed their objective analysis scheme to one developed by George Cressman (Cressman, 1959). It corrected for temporal continuity by using NMC's previous 12 hour forecast field, valid for the time of the analysis, as the first guess field for the analysis. Grid point values were determined by weighting observations around the grid point. Errors in observations could be readily detected when they varied considerably from the first guess field. Data sparse areas could be analyzed much more accurately using objective rather than subjective means. The Cressman scheme lasted until 1974 when it was supplanted for a short period by the
Hough analysis. This new scheme incorporated vertical consistency by using spectral objective techniques. In 1978 the analysis scheme was changed to a technique employing multivariate statistical methods. This technique remained in place for the duration of the available data set, however, there were continual improvements to the basic scheme. In fact, Trenberth and Olson (1988) have documented over 40 changes to the NMC objective analysis scheme from 1978 to 1987.

The NMC gridded data fields are not without error, and the objective schemes have changed over the period of record, but they remain one of the most error-free analyses available for use in climate and meteorological studies. The larger concern with the NMC analyses is that the archive is incomplete. Of particular concern to this study is 13.4% of the data fields are missing during the summers of 1982 and 1983 (Trenberth and Olson, 1988).

NMC archives the analyses in global, hemispheric, and octagonal formats. The octagonal grid is a specific grid designed by NMC so that a 47 by 51 array of evenly spaced grids can be overlaid on a Northern Hemisphere polar stereographic secant projection. The four corners of the array were cut off to reduce the size of the data set, and yet maintain as much Northern Hemisphere coverage as possible. Figure 2 shows the octagonal grid overlaid on a polar stereographic secant projection. Each dot represents a grid point. It should be noted that the octagonal u and v wind components are not in relation to compass directions, but to the x and y coordinate system of the polar stereographic projection. Figure 3 is an expanded view of North America. Note the lack of data in the southwest corner of the picture. Overlaid on Figure 3 are the locations of the upper air sites on which the grid data is based. The weather ships are no longer in operation.
Figure 2 NMC octagonal grid
Figure 3 NMC octagonal grid for North American sector with radiosonde stations overplotted
3.3 OUTGOING LONG-WAVE RADIATION

Measurements of Outgoing Long-wave Radiation (OLR) from polar-orbiting NOAA satellites started in June 1974 (Gruber and Krueger, 1984). Each time a satellite failed, another satellite was sent up to continue the data collection. In spite of the different radiometer characteristics a twice-daily global OLR data set has been derived. The satellite infrared radiometer measures in a narrow-band window, typically from 10.5 to 12.5 μm. This radiance is converted to an equivalent blackbody window temperature, $T_b$, by means of a regression equation that relates $T_b$ to 99 different atmospheres with varying vertical moisture and temperature profiles. Because several different satellites have collected the data in the past, another regression equation is used to convert all the different satellite $T_b$ data to the same flux temperature, $T_f$. Outgoing long-wave radiation can then be estimated using the $T_f$ data via $\sigma T_f^4$, where $\sigma$ is the Stefan-Boltzmann constant. Given that the actual infrared window has changed over the years from 10.5 - 12.5 μm, to 10.5 - 11.5 μm, and then to 11.5 - 12.5 μm, it is somewhat surprising to find the data are fairly consistent, with expected errors of only 11 W m$^{-2}$ (Gruber and Krueger, 1984). Another problem exists with the OLR data set. Each time a different satellite was launched, it had a slightly different Equator crossing time, which introduces a diurnal bias since the satellite was observing the same areas at different times of the day. The simplest way to reduce the error associated with different Equator crossing times is to average the twice-daily data. This reduces the error to at most 7 W m$^{-2}$ (Gruber and Krueger, 1984). A correction has already been applied to the OLR data to account for the different equator cross times (Gruber and Krueger, 1984).
Outgoing Long-wave Radiation (OLR) flux data are available from the National Oceanic and Atmospheric Administration (NOAA). It is a twice-daily global data set with resolution of 2.5° latitude by 2.5° longitude.
CHAPTER 4. PRECIPITATION

4.1 PRECIPITATION REGIMES

In "The Climates of the Continents", Kendrew (1922) was one of the first to describe the different precipitation regimes in North America (excluding Mexico). Examination of monthly and daily precipitation over the United States shows a peak in rainfall during summer over much of the country (Epstein and Barnston, 1988). This summer rainfall is extremely important for agricultural purposes (Fritsch et al., 1986). In studying daily precipitation events in summer that were greater than 12.7 mm (0.5 inch) in depth, Heideman and Fritsch (1988) found that 80% of all warm season precipitation is convective in nature, being indirectly or directly associated with thunderstorms. This contrasts with winter precipitation, when maxima correspond to preferred storm tracks of synoptic cyclones. Topography can play a role in enhancing precipitation in both seasons.

Kendrew (1922) categorized precipitation in the southwest United States as the Arizona type; a winter maximum due to cyclonic activity, and a summer maximum due to local heating causing convective rainfall. Preceding the summer rains is an almost rainless June. The southeastern United States was labelled the Gulf type distinguished by a summer precipitation maximum later than the Arizona type, and overall greater rainfall throughout the year. Markham (1970) combined both the Arizona and Florida areas, and called it the August tropical regime. Markham assumed it was the strength of the trade winds, and height of the trade inversion, that was the main factor for the tropical rains. The higher the inversion, the greater the strength of the convection. Texas was not included in the tropical regime since its maximum in precipitation is delayed until
September. Markham (1970) reasoned that occasional torrential rainfall from hurricanes tended to influence the Texas statistics.

Harmonic analysis of rainfall (Horn et al., 1957 and Horn and Bryson, 1960) was used to determine the annual march of precipitation in the United States. These results show that the region of summer rainfall in Arizona and New Mexico also extends into southern Nevada and Utah. They also indicated another summer maximum over the panhandle of Florida. A more elegant way of displaying the harmonics was used by Hsu and Wallace (1976) employing vectors. The length of the vector indicated the amplitude of the harmonic, and the direction indicated the time of the maximum amplitude in precipitation. Essentially, the same summer maxima as those of Horn and Bryson (1960) were found.

Walsh et al. (1982) used a rotated factor analysis to determine areas of spatially coherent monthly precipitation. They found large areas of coherence to be related to cyclone tracks during winter, but during summer the spatial coherence broke down due to the predominance of convective precipitation. Englehart and Douglas (1985) found that precipitation frequency has more spatial coherence than precipitation amount. Using precipitation frequency they found a spatially coherent precipitation regime during summer over Arizona and western New Mexico. Horn et al. (1957) used a simpler approach of intermonthly precipitation changes to delineate different precipitation regimes. Positive and negative changes in precipitation amounts for successive months were analyzed. The southwest monsoon was easily identified by a sharp increase in rainfall from June to July, accompanied by drying in adjacent areas. Horn et al. (1957) hinted at the existence of some synoptic circulation patterns that affect all the areas simultaneously, but offered no suggestions.
Even the Baja Peninsula of Mexico exhibits unusual precipitation patterns. Hastings and Turner (1965) found that the precipitation patterns of the northern and southern ends of the Baja are completely different. Over the extreme northwest corner of the peninsula, close to San Diego, rainfall is a maximum during winter and rare in the summer. On the other hand, the southern tip has at least twice the rainfall of the northwest area, but it comes during summer and fall. There is a secondary maximum in winter, but it is considerably weaker, and there is no rain in the south during the spring.

Research by Ives (1949) and Wallen (1955) led to a description of the rainfall distribution in Mexico. In general annual rainfall is a maximum in southern Mexico, with the northwest or Sonoran state being the driest year round. Another maximum exists along the western slopes of the Sierra Madres. Portig (1965) described annual rainfall in Central America ranging from 1 m at inland sheltered areas to 6.5 m in exposed coastal areas. All stations possess rainfall maxima in the summer season, but as in Mexico many stations exhibit a bimodal distribution, with rainfall peaks occurring in June and September with relatively drier conditions prevailing in July and August. In Mexico this phenomenon is referred to as the "la canícula" or August drought, while many in Central America refer to it as "el veranillo". Portig (1965) contends that in Central America this minimum in July and August may be related to the movement of the intertropical convergence zone, and in Mexico by changes in the intensity of the Bermuda anticyclone. He later acknowledges (Portig, 1976) that the strength of the anticyclone and the associated increase in subsidence is not in phase with the precipitation, so the reasoning is not completely satisfactory.
Mosino and Garcia (1974) suggest other mechanisms. Common hypotheses for the dry spell invariably relate it to the variation of the Sun's declination, and its effect on insolation and heating. The maximum in June occurs in southern Mexico when the Sun's overhead position moves northward to the Tropic of Cancer, and the second maximum in September occurs as the Sun's overhead position moves southward. However, this would not account for the dry spell near the Tropic of Cancer over northeastern Mexico, since the Sun is directly overhead only once. More importantly, as Mosino and Garcia (1974) point out, the rainfall peaks in June and September do not occur on the same dates every year as would be the case if it was dependent on the Sun's zenith angle. They propose that the dry spell is related to changes in the atmospheric circulation. In midsummer an upper air trough develops over the Atlantic and extends southwestward over Florida and into Central America. Thus trade wind disturbances, such as easterly waves and tropical cyclones tend to get caught by the upper trough and recurve northeastwards along the eastern edge of the trough. This has the effect of cutting moisture off from the Gulf of Mexico, Mexico and Central America. However, this does not explain why the dry spell is restricted to the eastern side of Mexico, since easterly waves would cross Mexico to the Pacific ocean. Cavazos and Hasting (1990) postulate that the rainfall maximum in September in northeastern Mexico is due to tropical storm activity.

The annual cycle of precipitation (mm) based on monthly data for various stations over the southern United States, Mexico, and Cuba (for locations see Figure 4) is shown in Figures 5a to 5l. Table 1 gives the mean monthly precipitation (mm) and the record length (years) for each station. Monthly data for Logan and Tempe were derived from daily precipitation data. A total of twelve sites were chosen. Site selection was based on the availability of the data in the World Monthly Surface Station Climatology (WMSSC),
Figure 4 Location map
with the requirement of a minimum record length of 25 years (except for Caibarien which only possessed 10 years of data). The first six stations represent locations where the precipitation increases from June to July. The remaining six stations have contrasting summer precipitation regimes.

<table>
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<tr>
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<td>128</td>
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</table>

Table 1. Mean monthly precipitation (mm) and record length (years).

Before discussing the precipitation climatologies in Figure 5 an index is introduced to help evaluate the changes in rainfall between successive months. A simple index of the change in rainfall from June to July and from August to September, normalized by the two month total precipitation amount (x100) is calculated to show the precipitation changes in adjacent months. The indices given in Table 2 are as follows:

\[(\text{JULY - JUNE})/(\text{JULY} + \text{JUNE}) \times 100\]

\[(\text{SEPTEMBER - AUGUST})/(\text{SEPTEMBER} + \text{AUGUST}) \times 100\]
The index ranges from -100 to +100. If June has zero precipitation then the index will be +100. If July is zero then the index will be -100. Therefore, the index represents the relative change from month to month.

<table>
<thead>
<tr>
<th>Station</th>
<th>Jul-Jun (index)</th>
<th>Sep-Aug (index)</th>
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</thead>
<tbody>
<tr>
<td>Grand Canyon</td>
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<td>-17</td>
</tr>
<tr>
<td>Tempe</td>
<td>74</td>
<td>-22</td>
</tr>
<tr>
<td>Guaymas</td>
<td>96</td>
<td>-43</td>
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<td>Albuquerque</td>
<td>31</td>
<td>-23</td>
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<tr>
<td>El Paso</td>
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<td>-8</td>
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<tr>
<td>Chihuahua</td>
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<td>-8</td>
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<tr>
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<td>8</td>
</tr>
<tr>
<td>Mobile</td>
<td>13</td>
<td>-13</td>
</tr>
</tbody>
</table>

Table 2. Precipitation index changes

The Grand Canyon, Tempe², and Guaymas (Figures 5a, 5b, and 5c respectively), all exhibit southwest monsoon signatures, characterized by fairly dry Junes followed by very wet Julys. The July-June indices range from a low of 53 at the northern location to a high of 96 at the southern site of Guaymas. Note that the above three stations have a major peak in precipitation during summer, and a secondary peak of precipitation during winter.

Albuquerque, El Paso, and Chihuahua (Figures 5d, 5e, and 5f respectively) exhibit weaker southwest monsoon signatures, with indices ranging from 31 to 42. These stations do not exhibit a secondary winter precipitation peak.

²Tempe, AZ is a suburb of Phoenix, AZ and was chosen instead of Phoenix because its precipitation record remained unbroken by station relocations.
Figure 5 a) Monthly precipitation (mm) for Grand Canyon, Arizona (top).
  b) Monthly precipitation (mm) for Tempe, Arizona (bottom)
Figure 5 c) Monthly precipitation (mm) for Guaymas, Mexico (top),
d) Monthly precipitation (mm) for Albuquerque, New Mexico (bottom)
Figure 5 e) Monthly precipitation (mm) for El Paso, Texas (top), f) Monthly precipitation (mm) for Chihuahua, Mexico (bottom)
Figure 5 g) Monthly precipitation (mm) for Winnemucca, Nevada (top),
h) Monthly precipitation (mm) for Logan, Utah (bottom)
Figure 5 i) Monthly precipitation (mm) for Brownsville, Texas (top), j) Monthly precipitation (mm) for Caibarien, Cuba (bottom)
Figure 5 k) Monthly precipitation (mm) for Miami, Florida (top),
l) Monthly precipitation (mm) for Mobile, Alabama (bottom)
Winnemucca and Logan (Figures 5g and 5h, respectively) are examples of the drying that takes place to the north of the southwest monsoon in July. Both sites experience a considerable drop in precipitation from June to July, while their counterparts to the south simultaneously experience a large increase in precipitation.

A different summer precipitation regime occurs around the Gulf of Mexico. In particular, Brownsville and Caibarien (Figures 5i and 5j, respectively) have two rainy seasons interrupted by a relatively drier mid-summer period. Caibarien shows this best, with rainfall maxima in May and June, as well as September and October, and a relatively drier July and August. The drying in August at Brownsville is not as pronounced as in July so its pattern is not as clear as that at Caibarien.

The southeastern United States exhibits a mixture of precipitation regimes. Miami (Figure 5k) resembles Caibarien with a drier mid-summer period, while just to the northwest, Mobile (Figure 5l) experiences a single summer precipitation peak, that is more reminiscent of a weak monsoon signature.

Winnemucca and Logan (Figures 5g and 5h, respectively) begin to dry at the same time as Tempe and Albuquerque (Figures 5b and 5d, respectively) begin their monsoon rains. In the next section this is investigated further using daily data.

4.2 WET-DRY INDEX

Researchers normally study precipitation as total amounts of precipitable water whether in daily, monthly, or yearly formats. The use of total amounts has two important limitations. One is that trace amounts of precipitation are neglected. This is inconsequential if one is interested solely in the total amount of precipitation. However, if
one is concerned with the percentage of days on which precipitation occurs, then trace amounts have to be accounted for, in some way. Another limitation is that totals are often biased towards large events. For example, heavy convective rains that occur during summer severe weather have been known to produce total rainfall amounts in one day that match or exceed the long-term monthly average for the location.

To deal with the biases introduced by both large precipitation events and trace measurements a new method of studying daily precipitation is developed here. The first step is to recode the total daily precipitation amounts into a more simplistic measure. All precipitation events, whether large or trace, are recoded as "1"s. Then all days without precipitation are recoded as "-1"s. All missing data are recoded as zeros. Thus a running sum of the recoded data, the wet-dry index, increases during wet periods and decreases during dry periods. Extended periods of missing data are easily visible graphically since the "wet-dry index" remains constant.

Figure 6 shows an example of two such running sums for 1971. Figure 6a shows a relatively wet period during winter and spring in Logan, and a relatively dry period during summer and fall. Tempe on the other hand (Figure 6b) is generally dry throughout this year except during the summer when some indication of the monsoon appears. Note that the rains did not fall every day during the summer of 1971 but were interspersed with dry days.

The final index number at the end of the year is a measure of the year's dryness or wetness. For instance, an index of -365 would be a year of 365 continuous dry days. Conversely, +365 would indicate a year of 365 continuous wet days. The index can also be directly converted into the number of dry or wet days in the year as follows:
Figure 6 a) Wet-Dry index example for Logan, 1971 (top),
b) Wet-Dry index example for Tempe, 1971 (bottom)
WET = (365 + INDEX)/2  
DRY = (365 - INDEX)/2

Of course, the total number of days (365 in the above example) must be adjusted for Leap Years as well as missing data.

Figure 7a shows the time series of the annual wet-dry index for Logan from 1928 to 1986 and Figure 7b is a time series of annual total precipitation for the same time period. Note that even though the period beyond the mid 1960s shows a relatively wetter period than earlier, this is only reflected in the total annual precipitation in the 1980s. Hence, the number of wet days increased in the latter part of the record but this was not necessarily accompanied by increases in rainfall amounts. Figure 8a is the time series of the annual wet-dry index for Tempe from 1926 to 1984, and Figure 8b is a time series of annual total precipitation for the same time period. Tempe does not show the discrepancy in the latter part of the record shown for Logan. Figure 9 gives an idea of how well the wet-dry index is correlated with precipitation. A correlation coefficient of 0.74 (95% confidence limit from 0.59 to 0.84) for Logan means that approximately 54% of the variance is explained. Tempe has a slightly larger correlation of 0.79 (95% confidence limit from 0.67 to 0.87), and 62% of the variance is explained.

A 59-year average of the wet-dry index for each day of the year for Logan is shown in Figure 10a. (Note that almost all days have negative wet-dry index values. This indicates that on average there are more dry than wet records for any given calendar day). What is notable is a break in the wet-dry index as one goes from a very wet to a very dry period in the year, in a relatively short period of time. Thus the average break occurs sometime in the middle of June (year day, YD, 160-170). A gradual return to wetter conditions occurs in the later part of the year. It is interesting to note that the wet-dry
Figure 7 a) Annual wet-dry index for Logan, 1928-1986 (top), b) Annual precipitation (mm) for Logan, 1928-1986 (bottom)
Figure 8 a) Annual wet-dry index for Tempe, 1926-1984 (top).
b) Annual precipitation (mm) for Tempe, 1926-1984 (bottom)
Figure 9 a) Scatter plot of wet-dry index versus precipitation for Logan, 1928-1986 (top),
b) Scatter plot of wet-dry index versus precipitation for Tempe, 1926-1984 (bottom)
index does not reach the same level in the last two months of the year as in the first two months of the year, even though the monthly precipitation chart (Figure 5h) shows that the precipitation amounts to be almost equal. This indicates that there are fewer days in November and December which have precipitation, but they must have on average a higher daily precipitation total to produce monthly totals on par with January and February. Thus frequency of precipitation could be used as another technique to distinguish between precipitation regimes. Figure 10b is the 59-year average of the wet-dry index for each day for Tempe. Here one goes from a very dry period of the year which lasts to around YD 175 to a very wet period by YD 195.

In order to remove some of the noise in the signal a running sum was applied to the wet-dry index. Figure 11a is the running sum of the average daily wet-dry index for Logan, and Figure 11b for Tempe. This display makes it easier to find the break in slope for Logan, but the change at Tempe is now harder to see. The robustness of the break of slope for both Logan and Tempe were tested using two standard techniques. The first compares the odd years to the even years, and the second compares the first half of the record with the latter half. Since the breaks persist (not shown), it seems that both Logan and Tempe exhibit different temporal summer precipitation regimes. Logan goes from wet to dry in the summer and Tempe goes from dry to wet.

To better emphasize the break point the time series of the indices were adjusted so as to have a mean of zero. To accomplish this the final running sum index value at the end of the year was divided by 365. This average value is then subtracted from the index given for each day (i.e., before a running sum is done). Now a running sum of this adjusted data should force the index to end at zero. Figure 12a shows the transformed index for Logan with a more pronounced break point near YD 165. The transformed index
Figure 10 a) Average daily wet-dry index for Logan, 1928-1986 (top),
b) Average daily wet-dry index for Tempe, 1926-1984 (bottom)
Figure 11 a) Running sum of average daily wet-dry index for Logan, 1928-1986 (top), b) Running sum of average daily wet-dry index for Tempe, 1926-1984 (bottom)
Figure 12 a) Transformed wet-dry index for Logan, 1928-1986 (top),
b) Transformed wet-dry index for Tempe, 1926-1984 (bottom)
also identifies the more gradual return to a wetter climate that begins after YD 300. Tempe (Figure 12b) shows the onset of the monsoon beginning around YD 190. This happens to correspond to the first week of July, which is why the monsoon signal shows up so strongly in the monthly data set. The change in Logan occurs in the middle of June, which gives the appearance in the monthly data that the drying occurs more gradually.

A similar procedure is undertaken for precipitation totals, and the final transformed precipitation is shown in Figure 13. Both transformed precipitation curves show the same breaks as the transformed wet-dry index curves. This indicates that neither Logan or Tempe was unduly biased by trace precipitation records or large precipitation events.

Two more stations are examined to show that Logan and Tempe are not unique to the region. The transformed wet-dry index curves are shown for Winnemucca and Albuquerque (Figures 14a and 14b, respectively). Winnemucca is almost identical to Logan with the break occurring near YD 165. Albuquerque has a much earlier break than Tempe and occurs near YD 170.

In the next Chapter the changes in atmospheric circulation that accompany the rapid onset of the monsoon rains are examined.
Figure 13 a) Transformed precipitation for Logan, 1928-1986 (top),
b) Transformed precipitation for Tempe, 1926-1984 (bottom)
Figure 14 a) Transformed wet-dry index for Winnemucca, 1928-1986 (top),
b) Transformed wet-dry index for Albuquerque, 1931-1986 (bottom)
CHAPTER 5. NORTH AMERICAN ATMOSPHERIC CIRCULATION

5.1 MONTHLY COMPOSITES

General circulation studies of the Northern Hemisphere summer have been relatively rare. Almost all of the work has concentrated on winter circulation patterns when gradients are at their strongest and more exciting changes in weather and climate occur. Summer circulation patterns are fairly flat with little gradient in pressure, height, or temperature, hence the lack of research interest in the season.

An exception is a paper by White (1982), who produced an observational study of Northern Hemisphere summers (June to September inclusive composited from 1966 to 1977). Composited 200 mb height fields showed a zonal flow pattern over North America extending from the arctic to 30°N latitude. South of 30°N the gradient was weak and the only feature was an upper-level high centred just south of the Baja peninsula. Such upper-level highs are characteristic of all monsoon circulations around the globe.

Because of the lack of summertime observational studies the first step in the present study was to examine the standard monthly meteorological fields. Monthly fields were readily available for Mean Sea Level Pressure (MSLP), and 500 millibar height (500 mb) data from January 1946 to June 1989. Monthly composites for both were calculated from 1946 to 1988, inclusive.

Since pressure and height gradients are relatively weak during the summer over North America, non-standard contour intervals were used. Figure 15 shows the MSLP composites for June, July, August, and September (hereafter referred to as JJAS) using a 1 mb contour interval. The thermal trough line is shown in bold. The Pacific and Atlantic surface highs show approximately the same position and strength throughout the summer.
Figure 15a June composite of mean sea level pressure (mb) with a 1 mb contour interval (bold line indicates the thermal trough axis)
Figure 15b July composite of mean sea level pressure (mb) with a 1 mb contour interval (bold line indicates the thermal trough axis)
Figure 15c August composite of mean sea level pressure (mb) with a 1 mb contour interval (bold line indicates the thermal trough axis)
Figure 15d September composite of mean sea level pressure (mb) with a 1 mb contour interval (bold line indicates the thermal trough axis)
months. In between the highs, a thermal trough is evident in Mexico and the southwestern United States. The thermal trough weakens by 1 mb from June to July and moves slightly westward through the summer. Pressures increase over eastern Mexico, New Mexico, and western Texas by over 2 mb in July. An increase in pressure is opposite to what one would expect with the arrival of the summer rains in July. The preferred location of the thermal trough and change in strength is corroborated by Rowson and Colucci (1992).

Figure 16 shows the 500 mb height composites for JJAS using a 30 metre interval. The 500 mb ridge line is shown in bold. By focussing on the ridge line that lies parallel to the latitude circle, we can see that the ridge moves northward over North America rather quickly, moving from 25°N latitude to 35°N from June to July. The position remains relatively stable from July to August, but retreats to 25°N in September.

Evidence of this upper-level anticyclone dates as far back as 1921 when balloon launches from San Francisco revealed a southerly wind aloft and not the easterlies that would be expected in a land-sea breeze circulation (Reed, 1933). Even then forecasters believed that the tropical moisture affecting the southwest was being advected northwestwards from the Gulf of Mexico around the anticyclone that they discovered at 4000 m (Reed, 1937). Bryson and Lowry (1955a, 1955b) came to the same conclusion some time later.

Since the JJAS sequence of 500 mb height composites changed much more than MSLP did, the next step was to investigate higher levels in the atmosphere. Monthly fields of 200 mb heights were not readily available, but were constructed from twice daily 200 mb height data available from 1962 to 1989. All available data were averaged to produce monthly composites for JJAS. Figure 17 shows the 200 mb height JJAS
Figure 16a June composite of 500 mb height (m) with a 30 m contour interval (bold line indicates the 500 mb ridge axis)
Figure 16b July composite of 500 mb height (m) with a 30 m contour interval (bold line indicates the 500 mb ridge axis)
Figure 16c August composite of 500 mb height (m) with a 30 m contour interval (bold line indicates the 500 mb ridge axis)
Figure 16d September composite of 500 mb height (m) with a 30 m contour interval (bold line indicates the 500 mb ridge axis)
sequence using a 40 metre contour interval. In June there is an indication of an upper ridge building over North America. By July a closed upper high of considerable strength forms over northwestern Mexico. The high remains quasi-stationary during August, and by September it has weakened and retreated southward. Thus an upper-level anticyclone similar to that found in the Asian, Australian, and African monsoons is found over North America as well.

5.2 UPPER WIND FIELD

The height field gradient revealed by the summer composites is quite weak over the southern portions of North America. Therefore, to better emphasize the upper-level circulation, the 250 mb wind fields were examined. Twice-daily 250 mb level u and v wind component data were available from May 1965 to June 1989. The JJAS sequence of monthly composites based on daily data is shown in Figure 18. A 25.7 m s⁻¹ (50 knot) wind is equivalent to a wind vector length of 22 mm.

In June anticyclonic flow is centred near 20°N over western Mexico. Magana and Yanai (1991) refer to the upper-level anticyclone over Central America as the Mexican anticyclone. They believe it is "maintained by divergence associated with convection over Central America". The mean position of the polar jetstream cuts across the continent near 50°N. A weaker subtropical jetstream tracks over the northern sections of the Baja peninsula, and gradually merges with the polar jetstream over the continent. By July the upper-level anticyclone is much farther north, centred around the Mexico-Arizona-New Mexico border. There is also clear evidence of an eastward extension of the anticyclonic circulation toward northern Florida. The polar jetstream is still in the same relative position as in June, but now the subtropical jetstream has moved farther
Figure 17a June composite of 200 mb height (m) with a 40 m contour interval
Figure 17b July composite of 200 mb height (m) with a 40 m contour interval
Figure 17c August composite of 200 mb height (m) with a 40 m contour interval
Figure 17d September composite of 200 mb height (m) with a 40 m contour interval
Figure 18a June composite of 250 mb wind
(26 m s$^{-1}$ wind vector = 22 mm in length)
Figure 18b July composite of 250 mb wind
(26 m s\(^{-1}\) wind vector = 22 mm in length)
Figure 18c August composite of 250 mb wind
(26 m s\(^{-1}\) wind vector = 22 mm in length)
Figure 18d September composite of 250 mb wind
(26 m s\(^{-1}\) wind vector = 22 mm in length)
north and merges with the polar jetstream over the northwestern United States. The main circulation cell drifts slightly southward in August, and by September it has retreated southward to 25°N along western Mexico. The extension towards the east is now harder to discern. The subtropical jetstream has returned to its mean position over the northern Baja peninsula.

South of the upper-level ridge there is often an area of cyclonic shear, referred to as the Tropical Upper Tropospheric Trough (TUTT) (Whitfield and Lyons, 1992). A TUTT is a relatively narrow cyclonic shear zone, elongated southwest to northeast, that appears in the climatological mean summer circulation pattern of the upper troposphere. An example is given for the mean 250 mb-level wind field for July (Figure 19). Vorticity maxima at the 200 mb-level propagate along the TUTT axis. The TUTT or vorticity maximum is not usually reflected in any feature at the surface. The TUTT is characterized by anomalous sinking motion near the centre due to upper-level convergence, and weak rising motion along the eastern flank which can produce some rainfall (Whitfield and Lyons, 1992). The rainfall associated with TUTTs is overshadowed by the monsoon rains, but TUTTs can be very important in day-to-day convective events.

To better examine the motion of the upper-level anticyclone, weekly 250 mb-level winds were composited out of twice daily data. The individual sequence of maps are not shown but the seasonal evolution of the estimated central position of the main Mexican high is shown in Figure 20. The number beside each point is the week that was composited. The northward motion of the upper-level high is extremely fast during the first 7 weeks. The high then tends to meander during the height of summer (weeks 7-10). When the high begins to retreat it appears to stall (weeks 12-14) for some time before
heading farther south at a somewhat slower pace than it had originally advanced. Since
the northward advance is more rapid than its retreat, the motion is asymmetric.
Figure 19 Example of a Tropical Upper Tropospheric Trough (TUTT)
Figure 20 Weekly averaged positions of the 250 mb upper-level anticyclone
CHAPTER 6. NORTH AMERICAN MONSOON

6.1 UPPER-LEVEL DIVERGENCE

Given that a thermal surface trough occurs throughout the summer months over Mexico and the southwestern United States, there should also be low-level convergence and upward vertical motion. If enough moisture is available, the ascending motion would likely lead to the development of convective clouds. In addition, upper-level divergence is associated with mid-tropospheric ascent.

Krishnamurti (1971) found east-west upper-level circulation patterns across the tropics while examining upper-level data for the summer of 1967. The strongest upper-level divergence centres are located over western Mexico and southeast Asia. The corresponding areas of upper-level convergence of the thermally-direct circulation are over the Atlantic and eastern Pacific Oceans around 30°N. He discovered that the intensity of the east-west upper-level circulation is comparable to the Hadley-type circulation, and that it is distinct from the Walker circulation which is known to be farther south along the Equator.

The relationship between upper-level anticyclones and convective precipitation has been noted in a number of studies. Leary (1979) and Leary and Houze (1979) showed how the development of a single tropical cloud cluster (100 km diameter) was related to an upper-level (200 mb) anticyclone in the large-scale flow. A cloud cluster formed in an area of weak upper-level divergence. Low-level convergence in the cluster eventually increased until a closed surface cyclonic circulation formed. Finally, upper-level divergence increased as the cloud cluster reached maturity. McBride and Gray (1980)
found that upper-level divergence was evident in 87 cloud clusters they examined in both the Pacific and Atlantic Oceans.

In the case study of a severe thunderstorm by Hales (1975) there was a definite 300 mb anticyclone located directly over the southwestern United States, but he only examined the low-level convergence zones. Two squall lines in a particular Arizona monsoon season were examined by Smith and Gall (1989) and upper-level anticyclonic flow was clearly found in both cases. Unfortunately, upper-level divergence calculations were only available for one case, thus it was only suggested that upper-level divergence could have played a role in the development of the squall line.

While studying mesoscale convective complexes (MCC) Fritsch and Maddox (1981a, 1981b) found that all ten cases exhibited a strong anticyclonic flow perturbation at 200 mb. Unsure as to whether the anticyclonic flow was there prior to the development of the MCC or whether the MCC developed its own upper-level anticyclone, they ran a numerical experiment which showed that an upper-level anticyclone developed after the complex formed. However, they still regard it as a "chicken and egg" question.

6.2 DIVERGENCE AND PRECIPITATION

In May 1986 NMC changed the way they conducted their analyses, resulting in improved tropical upper-troposphere divergence fields (Mo and Rasmusson, 1993; Dey, 1989). Hence pre-1986 upper-level divergence calculations are not as good as those in later years. However, the interest here is in the changes that take place in the upper circulation, and this should be less sensitive to alterations in the analysis scheme. The changes are most abrupt between June and July, and less so between August and
September. Hence, difference fields between adjacent summer months were calculated for divergence fields. Hopefully, this differencing will reduce the biases introduced by the new analysis scheme. Figure 21 shows the smoothed July minus June upper-level divergence field. The upper-level flow becomes increasingly divergent aloft over northwestern Mexico, while smaller increases occur just to the west of Florida. There is also an increasingly divergent flow aloft over Lake Superior, which may be anomalous since it is based on a single grid point. However, no glaring error is evident in the data set. The flow becomes increasingly convergent aloft over the northwestern United States, and from the Great Plains States to the Gulf of Mexico, as well as over Cuba and Jamaica.

Similar difference calculations were done for adjacent monthly precipitation data from the WMSSC data set. Figure 22 shows whether a particular station increases in precipitation from June to July (indicated by a "+" symbol), or decreases in precipitation from June to July (indicated by a "-" symbol). Generally, there is an increase in precipitation over western Mexico, and northward into Arizona, New Mexico, Colorado, and southern Utah and Nevada. The eastern United States also experiences an increase in precipitation from June to July, while precipitation decreases in the rest of the United States, eastern Mexico, Cuba, and Jamaica.

The relationship between changes in upper-level divergence and changes in surface precipitation becomes clearer when the two fields are superimposed (Figure 23). Over the southern United States and Mexico, wherever there is an increasingly divergent flow aloft from June to July, we observe a corresponding increase in precipitation. Conversely, wherever there is an increasingly convergent flow aloft from June to July, we observe a corresponding decrease in precipitation. For example, there is an increase in
Figure 21 July minus June divergence ($10^{-8} \text{ s}^{-1}$) difference field with a contour interval of $2.0 \times 10^{-7} \text{ s}^{-1}$
Figure 22 July minus June composite monthly precipitation (locations of stations with monthly precipitation available for at least 10 years from the WMSSC data set. A "+" denotes an increase in precipitation from June to July and locations with a "-" denotes a decrease in precipitation from June to July)
Figure 23 July minus June composite monthly precipitation overlaid on July minus June divergence \((10^{-8} \text{ s}^{-1})\) difference field with a contour interval of \(2.0 \times 10^{-7} \text{ s}^{-1}\)
precipitation through the southeastern United States down into Florida, until we reach
the southern tip, where precipitation actually decreases in July. This decrease extends
southeastward into Cuba and Jamaica and corresponds well with an increasingly
convergent flow aloft. A decrease in July precipitation is evident throughout the Great
Plains which corresponds to an area of increasingly convergent flow aloft. Through much
of western Mexico and northward into Arizona and New Mexico, there is an increase in
precipitation, which corresponds to an area of increasingly divergent flow aloft. The
relationship does not hold as well through Texas and eastern Mexico where the area of
increasingly convergent flow aloft is to the east over the Gulf of Mexico. An increase in
precipitation over the northeastern United States does not correspond to the increasingly
convergent flow aloft, which may be related to precipitation mechanisms other than
convection.

Figure 24 shows September minus August divergence superimposed on
September minus August precipitation changes. The dominant features in this pattern are
similar to those in the July minus June pattern, but of reversed polarity. The upper-level
flow becomes increasingly divergent over the extreme northwestern United States and
from the Great Plains States to the Gulf of Mexico, as well as Cuba and Jamaica.
Conversely, the flow becomes increasingly convergent aloft over Arizona, New Mexico,
western Mexico, and the eastern United States. Northern California does show
increasingly convergent flow aloft, which is inconsistent with the increase in precipitation
over that region. However, it should be noted that by September convective precipitation
is no longer the dominant type of precipitation at the latitude of northern California.
Synoptic weather systems coming from the Pacific become more dominant. Increases in
precipitation are apparent in the northwestern United States, Texas and eastern Mexico,
Figure 24 September minus August composite monthly precipitation overlaid on September minus August divergence ($10^{-8}$ s$^{-1}$) difference field with a contour interval of $2.0 \times 10^{-7}$ s$^{-1}$.
as well as Cuba and Jamaica. Decreases in precipitation are found in northwestern Mexico, Arizona, New Mexico, sections of the Great Plains, and the eastern United States.

Once again, the relationship between changes in upper-level divergence and changes in the surface precipitation is again clear (Figure 24). Over the southern United States and Mexico wherever the upper-level flow becomes increasingly divergent (convergent) aloft from August to September we observe a corresponding increase (decrease) in September precipitation. For example, there is a decrease in precipitation through the southeastern United States down into Florida, until we reach the southern tip, where precipitation increases in September. The increase extends southeastward into Cuba and Jamaica and corresponds well with the upper atmosphere becoming increasingly divergent aloft. An increase in September precipitation is evident in Texas, and eastern Mexico, which corresponds to an area of increasingly divergent flow aloft. Through much of western Mexico and northward into Arizona and New Mexico, there is a decrease in precipitation, which corresponds to an area of increasingly convergent flow aloft.

The relationship is no longer valid in some sections of the northern United States. The weakness of the relationship can probably be attributed to the effect of synoptic scale weather systems that begin to affect the northern United States in September.

A simple linear regression analysis is employed to show the scatter of July minus June divergence, to July minus June precipitation. Upper-level divergence values are interpolated from the raw grid point data at every precipitation site in Jamaica, Cuba, Mexico, and the United States up to a latitude of 42°N, which coincides with the border of northern California, Nevada, and Utah. Figure 25a shows the July-June scatterplot of
153 data points with the best fit line as shown. The correlation coefficient ($r$) is 0.57 (95% confidence limits from 0.45 to 0.67). The coefficient of determination ($r^2$) is 0.33, indicating that 33% of the variance can be explained, which shows that there is a fairly strong relationship between changes in divergence aloft and changes in surface precipitation. Similarly, September minus August data (Figure 25b) gives a correlation coefficient of 0.45 (95% confidence limits from 0.32 to 0.57) and a coefficient of determination of 0.20. The lower value was expected because the retreat of the monsoon does not regularly occur at the end of August but sometimes occurs as late as the end of September.

Horel et al. (1989) used Outgoing Long-wave Radiation (OLR) to study the annual cycle of convection during 1980-87 over the Amazon basin, and found that the position of the Bolivian high at 200 mb correlated well with inferred convection during the rainy season. Their results gives credence to similar correlations found here over Mexico, the southern United States, and the northern Caribbean.

6.3 PRECIPITATION REGIMES REVISITED

Building on this notion of the relationship between changes of surface precipitation and those in upper-level circulation, it is appropriate to re-examine the summer precipitation regimes given in Section 4.1.

Figure 26 shows the onset index (from Table 2) overlaid on the July-June change in upper-level divergence. Figure 27 shows the demise index (from Table 2) overlaid on the September-August change in upper-level divergence.

The monsoon is surprisingly apparent in the southeastern United States, although Hsu and Wallace (1976) did suggest that "Small areas of late summer maximum
Figure 25 a) Scatter plot of July minus June monthly precipitation (mm) versus July minus June divergence ($10^{-8}$ s$^{-1}$) for stations in Mexico, Cuba, Jamaica, and the United States up to 42°N latitude (top)
b) Scatter plot of September minus August monthly precipitation (mm) versus September minus August divergence ($10^{-8}$ s$^{-1}$) for stations in Mexico, Cuba, Jamaica, and the United States up to 42°N latitude (bottom)
Figure 26 July-June precipitation index for stations in Table 2 overlaid on July minus June divergence ($10^{-8} \text{ s}^{-1}$) difference field with a contour interval of $2.0 \times 10^{-7} \text{ s}^{-1}$
Figure 27 September-August precipitation index for stations in Table 2 overlaid on September minus August divergence ($10^{-8}$ s$^{-1}$) difference field with a contour interval of $2.0 \times 10^{-7}$ s$^{-1}$.
[precipitation] over Florida.....are probably monsoonal in character". Only small changes in precipitation are evident, but they are consistent with the changes in divergence aloft. Most researchers agree that the convective rainfall in Florida is associated with the low-level convergence zones associated with sea breezes. However, Burpee (1979) shows that the strength of convergence of the sea breezes has surprisingly little correlation with the amount of rainfall. Mobile has one of the lowest July-June precipitation indices (see Table 2) because the upper-level ridging to the east is considerably weaker over the Gulf coast states.

Much of the U.S. Midwest shows precipitation peaks in June and September, with slightly lower precipitation amounts in July and August. Keables (1989) concluded that the month of June is wet in the Midwest because the 700 mb flow was more often southwesterly, providing moisture advection from the Gulf of Mexico. As summer progresses a 700 mb ridge builds over British Columbia turning the flow more northwesterly, hence cutting off the supply of moisture to the region. Drier Julys and Augsts result. As the end of summer approaches, the 700 mb ridge to the west diminishes somewhat allowing a more southwesterly flow pattern again, which leads to wetter Septembers. Our analysis of the 700 mb flow patterns (not shown) indicate a direct westerly flow over the US Midwest in June and September, while the flow turns slightly northwestwards (approximately 300°) during July and August. The mean 700 mb flow is never southwesterly. During the months of July and August the upper-level flow becomes more convergent aloft than in June and September, which coincides nicely with the bimodal precipitation pattern Keables (1989) was examining. Similarly, the relative "dry" spell in July and August through Texas and eastern Mexico, is consistent with the upper-
level flow becoming increasingly convergent aloft. In southern Mexico, the relationship is not as clear.

To the north of the monsoon, considerable drying takes place as the monsoon rains begin, in early July. The effect is strongest in northern Nevada and Utah but stretches northward to the Pacific northwest. Lowry (1956) first suggested that the atmospheric circulation links Oregon's drying trend in early July to the monsoon to the south.

6.4 OUTGOING LONG-WAVE RADIATION (OLR)

Outgoing long-wave radiation has been used as a proxy measurement for convection over the tropics for some time (e.g. Bess et al., 1989 and Maddox et al., 1992b). Any OLR values less than or equal to 240 W m$^{-2}$ are regarded as being associated with cold convective cloud tops (Arkin et al, 1989), and therefore, convective rainfall. Assuming blackbody conditions, 240 W m$^{-2}$ is equivalent to 255 K (-18°C). Lifting a saturated air parcel with a temperature of 20°C, from the 1000 mb level, the cloud top must ascend to over 6 km (20000 feet) to cool to -18°C in a standard atmosphere. Therefore, if cloud tops are over 6 km, the clouds are considered to be convective. Of course, this calculation is for a single measurement of a cloud top. In the case of an individual OLR grid point, which is averaged over a 2.5 degree latitude by 2.5 degree longitude box, cold convective cloud tops would be averaged with warmer surroundings and produce significantly warmer average temperatures than -18°C. Hence, the actual convective cloud top heights are more likely considerably colder than -18°C and much higher than 6 km. A considerable amount of research has been done to calculate exactly what average OLR grid value corresponds to cold convective cloud tops.
that are presumably raining underneath. The current consensus is that this threshold corresponds to 240 W m\(^{-2}\). There are still errors associated with using OLR as a proxy for convective precipitation. The largest such error is the influence of cirrus clouds, which can have very low OLR values yet not necessarily be associated with convective precipitation.

Figure 28 shows composite OLR for JJAS using a contour interval of 10 W m\(^{-2}\). Here the solid lines indicate areas of convective activity. It must be emphasized here that over the extreme northern United States the ground temperatures are cool enough to bias the OLR values. Hence, it is not appropriate to use OLR as a proxy for convective precipitation under such circumstances. Since we are concerned only with the summer months the distinction between cool ground and cold cloud tops is not that difficult to make. In other seasons, it would be more difficult and some kind of parameter for cloud cover would have to be incorporated. The OLR values are also monthly averages, hence areas which do not have daily convective clouds will not reach the threshold value of 240 W m\(^{-2}\).

Convection in June is evident over northwestern South America, Central America, and westward along the intertropical convergence zone. By June convection has also started its northward march into southern Mexico. The high (warm) OLR values over northwestern Mexico are an indication of both clear skies and the intense surface thermal heat low already evident in the area. By July rapid change has occurred and convection reaches well into northwestern Mexico, close to the Arizona and New Mexico borders. Note that the 240 W m\(^{-2}\) threshold does not reach into the southeastern United States, which indicates that convective cloud tops do not occur daily over the region. There is also a slight northward extension of the intertropical convergence zone over the eastern
Figure 28a June composite of OLR (W m$^{-2}$) with a contour interval of 10 W m$^{-2}$
(contours less than or equal to 240 W m$^{-2}$ are solid)
Figure 28b July composite of OLR (W m$^{-2}$) with a contour interval of 10 W m$^{-2}$ (contours less than or equal to 240 W m$^{-2}$ are solid)
Figure 28c August composite of OLR (W m\(^{-2}\)) with a contour interval of 10 W m\(^{-2}\) (contours less than or equal to 240 W m\(^{-2}\) are solid)
Figure 28c September composite of OLR (W m\(^{-2}\)) with a contour interval of 10 W m\(^{-2}\) (contours less than or equal to 240 W m\(^{-2}\) are solid)
Pacific. The convective rainfall is at full intensity during the month of July. The pattern is similar in August. However, by September the area of convection is retreating southward. Note again that the onset and demise of the convective rainfall is not symmetrical, that is, the progression is quicker than the retreat. Negri et al (1993) found similar OLR distribution patterns while examining the rainfall climatology over Mexico.

6.5 OLR AND PRECIPITATION

OLR data are used to corroborate the relationship between upper-level divergence and precipitation. Figure 29 is July minus June OLR with a contour interval of 5 W m\(^{-2}\). Negative values are indicative of cooler cloud tops, hence the more negative the values, the more convective July is compared with June. Conversely, positive areas indicate less convective activity in July than June. Obviously, there is a considerable decrease in OLR, and hence increase in convective activity, over western Mexico and into Arizona and New Mexico. There is also a small decrease in OLR, and hence increase in convection, along the northern Gulf of Mexico as far east as Florida. Areas showing a decrease (increase) in convection (OLR) include Central America eastward to Cuba. Other areas showing strong decreases (increases) in convection (OLR) are centred over the U.S. Midwest southward into Texas, as well as from Nevada northeastwards into Montana. When compared with the divergence and precipitation map for the same period (Figure 23) the general pattern is almost identical. The increase in convection inferred from the decrease in OLR over northwestern Mexico matches well with an increasingly divergent flow aloft and an increase in precipitation, and the position of the upper-level anticyclone. The extension of the upper-level anticyclone eastward into Florida corresponds to an increasingly divergent flow aloft, increased (decreased) convection (OLR) and
Figure 29 July minus June difference field for OLR (W m\textsuperscript{-2}) with a contour interval of 5 W m\textsuperscript{-2} (negative contours are dashed)
precipitation. Similarly, areas of decreased (increased) convection (OLR) and precipitation compare well with regions of increasingly convergent flow aloft.

Figure 30 is September minus August OLR and is essentially a reversal, albeit a weaker change than July minus June. Now convection (OLR) decreases (increases) over western Mexico, Arizona, New Mexico, and the northern Gulf of Mexico region. While convection (OLR) increases (decreases) slightly over Texas, Central America, and areas south of Florida. There are at least four reasons for the weaker September minus August changes. Firstly, convection is indirectly being inferred from the OLR, and the relationship is not perfect. For example in September the cooler ground temperatures in the northern United States can begin to bias the OLR data. Secondly, the onset and demise of the monsoon is asymmetric. Recall that the monsoon onset is more rapid than its demise, hence changes in precipitation are less dramatic. Thirdly, the demise does not fall nicely at the transition periods between the two months of August and September, whereas the onset does start near the beginning of July. Finally, the increased effect of synoptic weather systems over the northern sections of the United States during September will tend to mask out convective precipitation.

Mo and Rasmusson (1993) looked at OLR, divergence, and precipitation in a different manner. Instead of comparing precipitation with divergence, they compared the OLR and divergence fields for the summer of 1987. They found that OLR departures from the mean had a spatial correlation coefficient of 0.8 with 200 mb divergence. They also found that OLR is a poor proxy for divergence in areas of weak convection or areas of subsidence. That is, areas of 200 mb convergence are poorly related to OLR departures. This means that our comparison of precipitation and divergence would also fail in areas of upper-level convergence.
Figure 30 September minus August difference field for OLR (W m\(^{-2}\)) with a contour interval of 5 W m\(^{-2}\) (negative contours are dashed)
6.6 INTERANNUAL VARIABILITY

The largest increases in OLR from June to July occur over northwestern Mexico. Monthly OLR means are calculated from the twice daily OLR data, for each year of available data. Fifteen years of data from 1974 to 1990 were used (1978 and 1988 were missing).

The onset of the monsoon brings the most noticeable change in convective activity. The 15 years of July minus June OLR difference fields are shown in Appendix A. Many years closely resemble the average July-June field shown in Figure 29. Others show the monsoon did not reach as far northward as Arizona and New Mexico. Some show an increase (decrease) in convection (OLR) across the northern Gulf of Mexico, while others do not. One year in particular (1979) showed a weak Mexican monsoon with a stronger inferred increase in convection over the Gulf of Mexico.

Because the spatial variability of the monsoon makes it difficult to estimate its strength each year, a simple objective index is developed. The largest change in OLR between June and July almost always occurs over northwestern Mexico, as in the average field. The OLR grid is based on 2.5 by 2.5 degree latitude-longitude boxes, and 12 out of the 15 year maximum OLR differences occurred at the same grid point (one was just to the left of that grid point and two were to the right). Thus, the maximum value of those three particular grid points (points A,B,C see Figure 4) is used to create an index to examine the temporal variability of the monsoon. That is, the maximum July-June OLR difference normalized by the July+June OLR for either point A,B, or C is the simple objective OLR index.
The mean OLR difference over the 15 year period was -48.0 W m\(^{-2}\), with a standard deviation of 19.0 W m\(^{-2}\). This indicates considerable variability, and shows that the onset of the monsoon is not always a strong event. Gadgil \textit{et al.} (1992) found that the OLR values from NOAA-SR (1974-78) were consistently higher than those from NOAA-7 (1982 and onward) and could not be accounted for by an increase in convection since rainfall data did not appear to corroborate the lower values. Thus, any long term trend analysis of OLR data is subject to criticism. However, we are looking at difference amounts between adjacent months, which should be less sensitive to the effects of systematic error.

Figure 31a shows the OLR index for increased convective activity versus year. The more negative the number the stronger the onset of the monsoon. The 1970s still appear to have had stronger monsoons than the 1980s. This is consistent with Gadgil \textit{et al.}'s (1992) systematic error hypothesis, even though OLR difference values were employed.

In order to verify that the OLR index is an appropriate measure of the monsoon, comparison is made to monthly precipitation. Unfortunately, the area of the index grid points has no precipitation station. The closest station from the WMSSC data set is Guaymas, Mexico which is 2° south of the points A, B, and C. To the north is Arizona and New Mexico, whose closest stations are approximately 2° to the north of points A, B, and C. Therefore, to give an estimate of the precipitation in the area of points A, B, and C an average was taken of Guaymas, and the southeastern Arizona and southwestern New Mexico climate data divisions (each climate data division is an average of all available stations within each division). Figure 31b is a plot of the combined average precipitation index (July - June)/(July + June).
Figure 31 a) OLR index from 1974-1990 (top),
b) Precipitation index from 1974-1988 (bottom)
The scatter plot (Figure 32) gives a correlation coefficient (r) of -0.70 (95% confidence limits range widely from a low of -0.24 to a high of -0.90 because of the limited sample size). There are only 13 years of overlapping data, so any conclusions are tenuous at best. Nevertheless, agreement between the two indices gives support to the suggestion that the OLR index is an appropriate measure of the monsoon.

The weakest OLR index occurs in 1987, which corresponds to the lowest precipitation index. The second weakest OLR index is in 1984 which corresponds to the second lowest precipitation index. The next two weakest OLR indices are 1979 and 1986. Precipitation indices for these two years are fairly low but do not correspond to the same ranking as the OLR index. The four weakest OLR indices in order of increasing strength are 1987, 1984, 1986, and 1979.

The monsoon season in 1984 is a special case in that the monsoon started a month earlier than normal, with the upper-level anticyclone already into northwestern Mexico in June. Thus the July-June indices were artificially low since they were the differences between two wet months rather than a dry to wet transition. Hence, 1984 was not a weak monsoon year, but started a month earlier instead.

Examination of the remaining years (1979, 1986, 1987) reveals a common feature. Each year shows the upper-level anticyclone had not yet reached northwestern Mexico by July, and in fact was still weak and centred over the southern Baja peninsula. In both 1986 and 1987 the upper-level high managed to move farther northward by August, and hence the monsoon was delayed somewhat. However in 1979, the upper-level high never reached farther north than the middle of the Baja peninsula so the monsoon did not reach into the southwestern United States. Reyes and Cadet (1986,1988)
Figure 32 Scatterplot of OLR index versus Precipitation index
investigated water vapour fluxes during the summer of 1979, in an attempt to determine the moisture sources for the southwest monsoon. Unfortunately, this turned out to be one of the weakest monsoons of the period which is probably why their findings were not as substantial as they had hoped.
CHAPTER 7. SUMMARY AND CONCLUSIONS

Monsoon circulations are found throughout much of the extratropics. Using a scheme based on surface wind criteria, Ramage (1971) classified Africa, southern Asia, and northern Australia as having a monsoon climate. An atmospheric circulation pattern that coincides with all monsoons is a large upper tropospheric anticyclone (Das, 1986). A westerly jetstream is commonly observed to the north of the anticyclone, while an easterly jetstream is to the south (Chenetat, 1989; Soman and Kumar, 1993). During winter a strong surface anticyclone establishes itself over continental regions. This creates an "offshore" wind component and dry conditions generally prevail throughout the region. In summer the flow reverses as the surface anticyclone is replaced by a strong thermally induced low pressure system. The "onshore" wind component often coincides with the onset of heavy summer rainfall in the region.

Ramage's (1971) classification scheme excludes any monsoon climates in North America because there is no distinct seasonal shift in surface winds. However, Kendrew (1961) categorizes portions of Mexico and Central America as having a monsoon climate based on a summer rainfall maximum followed by a dry winter. Arizona is not included as a monsoon climate because it does not have a dry winter. However, the term "Arizona or Mexican monsoon" has become entrenched in the research literature to describe the summer precipitation in the southwestern United States and northwestern Mexico.

The Arizona or Mexican monsoon has been extensively studied in the past. Early research efforts focussed on moisture sources for the heavy summer rains. However, there are still many questions that remain to be resolved with the current hypothesis of moisture transport up the Gulf of California (Hales, 1974). Case studies of singular events
or mesoscale aspects of the phenomena comprise the bulk of the more recent research efforts. Thus much of the research has essentially been done in isolation from the larger scale atmospheric flow. As a result there are considerable gaps in our understanding of the relationship between the monsoon and the larger scale atmospheric circulation. This thesis attempts to address those gaps in understanding.

In Chapter 4 an investigation of monthly precipitation data showed several summer precipitation regimes throughout the southern United States and Mexico. Over the southwestern United States and northwestern Mexico rainfall increases sharply from June to July. A less abrupt increase takes place over the southeastern United States. The reverse is true over the northern United States Rocky Mountains, Texas, eastern Mexico, and the northern Caribbean, where rainfall decreases from June to July. These simultaneous but reverse changes in precipitation hint at the possibility of a larger scale connection between these summer precipitation regimes.

To highlight the temporal change in precipitation a few stations in the western United States were analyzed in more depth using daily precipitation data. Mexican daily precipitation data were not available for use in this study. A simple wet-dry index was developed and applied to graphically show the abrupt changes that occur in summer precipitation. The heavy summer rains begin in earnest over the southwestern United States near the end of June or beginning of July. To the north over the northern United States Rocky Mountains strong drying takes place near the middle of June.

The question of where the moisture for the heavy summer rains initially comes from is difficult to answer. Preliminary analysis of water vapour from satellite imagery suggests there are multiple potential moisture sources. One such possible moisture source could be the northern edge of the ITCZ, which moves northward during summer. Waliser
and Gautier's (1993) study of a satellite derived climatology of the ITCZ for a 17 year period shows that the large-scale deep convection along the ITCZ widens during July, with the northern edge reaching southern Mexico. Satellite imagery shows tropical water vapour plumes (well-defined boundaries of middle to upper-level moisture) extending from the northern edge of the ITCZ through Mexico and into the United States on a regular basis during the summer monsoon (Thiao et al., 1993). This large scale source of moisture hints at the possibility that local scale sources of moisture in Arizona or surrounding areas may be less important than previously thought.

In Chapter 5 the changes in large scale atmospheric circulation that accompany the onset and demise of the monsoon are examined. The summer rainfall so characteristic of other monsoons, does not coincide with the presence of the deepest surface thermal trough. In fact, the deepest thermal low occurs in June over Arizona a month in advance of the monsoon rains. The arrival of the monsoon rains is accompanied by a weak rise in mean sea level pressure (MSLP) over the affected area. This is contrary to what is normally expected with convective rainfall. It also suggests the possibility that other mechanisms, other than local convection, might be involved.

An examination of available National Meteorological Center upper level data shows an upper-level anticyclone which corresponds to surface precipitation as in other monsoons around the world. Weekly composites of upper-level wind data show the rapid progression of the upper-level anticyclone from southern Mexico during the beginning of June to the southwestern United States-northwestern Mexico border by the middle of July. The anticyclone meanders near the border for the remainder of July. In the beginning of August the anticyclone starts its southward track only to stall in the later half of August just southeast of Guaymas. The retreat begins again in the beginning of
September but at a slower pace than its initial advance through Mexico. Since the northward advance is more rapid than its retreat, the motion is asymmetric.

Analyses of upper-level divergence in Chapter 6 show that changes in precipitation are consistent with changes in divergence aloft. Over northwestern Mexico and the southwestern United States upper-level divergence corresponds well with the summer monsoon signature. Increased precipitation from June to July corresponds to an increasingly divergent flow aloft. As precipitation decreases from August to September an increasingly convergent flow aloft is evident. Similarly over the northern United States Rocky Mountains decreased precipitation from June to July corresponds to increasingly convergent flow aloft. The reverse is true from August to September, as precipitation increases over the Rockies correspond to increasingly divergent flow aloft. The bimodal peak in precipitation in the summer half of the year, around the Gulf of Mexico, also matches the increasingly divergent/convergent patterns aloft.

Using OLR as an independent data set, it is confirmed that convective rainfall correlates well with upper-level divergence. Precipitation, OLR, and upper level divergence information demonstrate a remarkable spatial correlation across the southern United States, Mexico, and the northern Caribbean.

These analyses show a great deal of consistency between the large scale atmospheric circulation and the different summer precipitation regimes in the area. A conceptual model of the monsoon at its height in July shows this in Figure 33. An upper-level anticyclone is well established over northwestern Mexico, with a ridge extending eastward to a weaker anticyclone over the southeastern United States. The subtropical jet has moved farther north where it converges with the polar jet over the northern
Figure 33 Conceptual model of the North American monsoon (Thick lines are jetstream axes. The strong upper-level anticyclone position is designated by a large H. The weak anticyclone is designated by a small H. Areas of increasingly divergent flow aloft (DIV) and increasingly convergent flow aloft (CON) are also noted)
United States. From June to July areas of increasingly divergent/convergent flow aloft match changes in precipitation (OLR) at the surface.

This thesis was originally undertaken to better understand the summer precipitation regimes that exist in North America, in particular, the Arizona or Mexican monsoon. Its primary goal was to provide other researchers and forecasters with a large scale view of the phenomenon, in hopes that this broadened context would provide a basis for improved prediction of summertime rainfall anomalies over North America. It is now clear that the onset of the monsoon is part of a much larger pattern of circulation changes that occur rather abruptly in late June to early July. Precipitation, OLR and upper air data present a mutually consistent picture of these changes, which extend over most of the southern United States, Mexico, and the northern Caribbean. The rapid northward development of the upper level anticyclone coincides with the abrupt changes found in the precipitation and OLR data. The upper-level anticyclone provides a much better indicator for the onset and demise of the monsoon than the previously used 500 mb ridge lines, or the strength and position of the thermal low (Carleton, 1985,1986, 1987). Forecasters would be well advised to monitor this atmospheric feature in order to help forecast the monsoon onset.

The monsoon can no longer be viewed as a meso- to synoptic scale phenomena: it is an integral part of the summertime circulation pattern over North America. Hence it would be better to rename it the North American monsoon. A re-evaluation of the presently accepted precipitation climatology of the area should also be undertaken. The summer precipitation regimes that heretofore have been treated as separate, should be treated as connected entities. The view that summer monsoons exist only in Asia, Africa, and Australia also needs re-examination. Figure 34 shows Ramage's (1971) monsoon
Figure 34 Global monsoon region as defined by Ramage (after Ramage, 1971) with approximate area of North American monsoon added
region as defined by his wind criteria with the approximate extent of the North American monsoon as defined by precipitation and OLR data. The northern and western boundaries coincide with the approximate limits of the monsoon rains over the southwestern United States. The eastern boundary is more artificial since precipitation data are not analyzed over the Atlantic Ocean. The southern boundary is undefined because the differential heating between land and ocean masses decreases as the land area diminishes.

Many of the remaining questions concerning the North American monsoon relate to the following questions:

1) Is there another unexplored source of moisture for the monsoon?
2) Why is the onset of the monsoon so rapid?
3) How does the monsoon affect surrounding areas of North and Central America?
4) What controls the interannual variability of the monsoon?

A cursory examination of hourly satellite imagery during several North American summer monsoons has led the author to believe that some of the water vapour that condenses over the southwestern United States and northwestern Mexico comes from the southwestern Gulf of Mexico through the Rio Balsas river basin (south of Mexico city) then northwestward through the Rio Grande river basin to the west coast of Mexico. From there the convection remains close to the western slopes of the Sierra Madre Occidental mountain range all the way north to the Mexico-United States border near Arizona and New Mexico. This southerly flow may be related to the gulf surges or low-level southerly jetstream found in the Gulf of California and surrounding areas during the summer monsoon (Douglas, 1992a).
This hypothesis could be investigated using 850 mb gridded wind data to examine the air flow around the region. However, the scale of the river basins is small enough that both the observation network and the objective analysis scheme that converted the data to gridded values might not be able to resolve it. It might be possible to use satellite derived winds from cloud motion to see if the flow across this basin at the top of the planetary boundary layer is strong enough to account for the moisture flux that supplies the monsoon and if it tends to be particularly strong during periods of heavy monsoon rainfall. If there were not enough trade wind cumulus clouds to support estimates of the wind field in this region, (subject to the availability of resources) constant density balloons could be used to track the air flow from the pathway to northwestern Mexico.

Another question deals with the rapid onset of the North American monsoon. More work is needed to better document the onset date of the monsoon as a function of geographical location. Weekly rather than monthly composites of infrared satellite imagery would better define the onset. Higher resolution OLR data are now available to better define the mesoscale structure of the monsoon. Eventually high quality daily precipitation data will be available for Mexico to help provide a better temporal sequence for the monsoon.

Since the solar forcing changes only slightly from June to July it is surprising that northwestern Mexico and the southwestern United States go from their driest to their wettest seasons in less than a month. How does the atmosphere change so quickly?

One possible avenue to explore is the atmosphere's ability to move from one steady state equilibrium to another in a short period of time. For example, the upper flow over western North America during wintertime can be meridional for weeks at a time and then, within a matter of a few days, change to zonal and remain that way for weeks,
before switching back to a more meridional flow. Perhaps the monsoon circulation and non-monsoon circulation are another example of such a phenomena. A diagnostic study of gridded upper air data could determine whether such abrupt changes in circulation are common. However, a more theoretically-oriented general circulation modelling study would be necessary in order to begin to understand why such rapid changes occur in the atmosphere.

Another interesting possibility is to investigate the effect of boundary conditions on the climate (Meehl, 1994): Extreme heating of the surface can occur only after the soil moisture has been largely depleted. Thus the differential heating between land and ocean masses can increase quickly, once the solar energy that was used to melt mountain snows and evaporate soil moisture during the spring becomes available to heat the parched surface during early summer. Reduced water availability will also cause changes in vegetation which, in turn, affect the albedo of the area. An analysis of observational data could be undertaken to see if relationships exist between certain anomalous boundary conditions and the monsoon. If any relationships are indicated then a climate model could be used to determine how much of a role the absence or presence of a particular boundary condition plays in the rapid onset of the monsoon.

When the monsoon begins over northwestern Mexico and the southwestern United States an almost simultaneous drying occurs both to the north and to the east. A more comprehensive analysis of daily precipitation data can better define the strength and extent of this negative correlation. The simultaneous occurrence of flooding in the Mississippi river basin and a relatively weak monsoon in the summer of 1993 is a reflection of such a negative correlation. A more comprehensive diagnostic study could
verify whether strong (weak) monsoons are consistently associated with suppressed (enhanced) summer convection in surrounding areas.

Several processes could contribute to the observed negative correlation described in the previous paragraph. It is conceivable that the monsoon takes most of the available moisture in the region, leaving less for the surrounding areas, or that it induces subsidence over neighbouring areas thus creating an inversion and reducing convection. General circulation modelling studies would be best to explore whether these hypotheses bear merit.

A preliminary examination of the OLR data shows that there is considerable interannual variability of the North American monsoon. Both the timing and intensity of the onset of the monsoon vary. The causes underlying the interannual variability of the North American monsoon need to be investigated. Since the monsoon is believed to occur in response to the differential heating of the land and adjacent oceans it seems reasonable that the variability of either land and ocean thermal characteristics could give rise to monsoon variability (Joseph et al, 1991, 1994).

Much like the heating of the Tibetan plateau is thought to play a major role in the Asian monsoon, the Great Basin of the United States may play a role in the North American summer monsoon (Tang and Reiter, 1984). There are several factors to examine in determining the thermal characteristics of the land. Since the amount of snow coverage varies from year to year, the eventual melting of all the snow takes a different amount of time each year. Once the energy is no longer needed to melt the snow it can go towards evaporating soil moisture. Once the soil is dry the energy can go towards heating the surface and overlying air, enabling the heating to occur at a faster rate. Hence, both snow cover and soil moisture may be important factors in determining whether a weak or
strong monsoon will occur. Hence, the preceding seasons may be very important to monsoon variability. A colder than normal winter with more snow could conceivably lead to a weaker monsoon, or an anomalously warm, dry spring may lead to a strong monsoon. Even vegetation and albedo changes could affect differential heating in the summer.

Changes in sea-surface temperature (SST) in either the eastern Pacific and the Gulf of Mexico could give rise to changes in differential heating. On a local scale warmer SST in the Gulf of California would increase evaporation, thereby increasing the available moisture for the monsoon, while colder SST could lead to a stronger sea breeze circulation which could enhance the vertical motion along the western flanks of the Sierra Madre Occidental mountain range during the day.

Large scale changes to SST in the tropical Pacific associated with ENSO have been shown to be related to the Asian monsoon (Joseph et al, 1994). Hence, there exists the possibility that the North American monsoon may feel the effects of SST anomalies in the eastern equatorial Pacific.

During El Nino years the ITCZ tends to be south of its climatological position. Previously, the ITCZ was proposed as a source of moisture for the North American monsoon, since tropical water vapour plumes regularly extend from the northern edge of the ITCZ through Mexico and into the United States during the summer (Thiao et al, 1993). Hence, the strength and position of the ITCZ could also play a role in the North American monsoon.

Diagnostic studies could be performed to see if relationships exist between any of the above mentioned factors and the summer monsoon. If significant relationships are
found, climate modelling experiments would have to be employed in order to try to understand the underlying linkages.

It is conceivable that both local and remote factors could contribute to the variability of the North American monsoon. Since the monsoon is such an important component of the summer climate for much of North America, progress in understanding the factors that influence it could yield major benefits in terms of improved summertime climate prediction.
REFERENCES


APPENDIX

OLR VARIABILITY 1974-1990

The following maps of OLR have undergone "smoothing" hence the central
values are not as high as the OLR index, which were obtained from the raw grid point
data.
Figure A1 July minus June OLR (W m\(^{-2}\)) difference field for 1974 with a contour interval of 5 W m\(^{-2}\) (negative contours are dashed)
Figure A2 July minus June OLR (W m$^{-2}$) difference field for 1975 with a contour interval of 5 W m$^{-2}$ (negative contours are dashed)
Figure A3 July minus June OLR (W m\(^{-2}\)) difference field for 1976 with a contour interval of 5 W m\(^{-2}\) (negative contours are dashed)
Figure A4 July minus June OLR (W m\(^{-2}\)) difference field for 1977 with a contour interval of 5 W m\(^{-2}\) (negative contours are dashed)
Figure A5 July minus June OLR (W m$^{-2}$) difference field for 1979 with a contour interval of 5 W m$^{-2}$ (negative contours are dashed)
Figure A6 July minus June OLR (W m\(^{-2}\)) difference field for 1980 with a contour interval of 5 W m\(^{-2}\) (negative contours are dashed)
Figure A7 July minus June OLR (W m⁻²) difference field for 1981 with a contour interval of 5 W m⁻² (negative contours are dashed)
Figure A8 July minus June OLR (W m\(^{-2}\)) difference field for 1982 with a contour interval of 5 W m\(^{-2}\) (negative contours are dashed)
Figure A9: July minus June OLR (W m$^{-2}$) difference field for 1983 with a contour interval of 5 W m$^{-2}$ (negative contours are dashed)
Figure A10 July minus June OLR (W m$^{-2}$) difference field for 1984 with a contour interval of 5 W m$^{-2}$ (negative contours are dashed)
Figure A11 July minus June OLR (W m$^{-2}$) difference field for 1985 with a contour interval of 5 W m$^{-2}$ (negative contours are dashed)
Figure A12 July minus June OLR (W m$^{-2}$) difference field for 1986 with a contour interval of 5 W m$^{-2}$ (negative contours are dashed)
Figure A13 July minus June OLR (W m\textsuperscript{-2}) difference field for 1987 with a contour interval of 5 W m\textsuperscript{-2} (negative contours are dashed)
Figure A14 July minus June OLR (W m$^{-2}$) difference field for 1989 with a contour interval of 5 W m$^{-2}$ (negative contours are dashed)
Figure A15 July minus June OLR (W m\(^{-2}\)) difference field for 1990 with a contour interval of 5 W m\(^{-2}\) (negative contours are dashed)