Dynamics and Thermodynamics of the Regional Response to the Indian Monsoon Onset

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ABSTRACT

The regional influence of the Indian monsoon onset is examined through observational analysis focusing on the Rodwell–Hoskins “monsoon-desert” hypothesis, which proposes that the strong diabatic heating associated with the monsoon produces a Gill-like Rossby wave response that thermodynamically interacts with the midlatitude westerly jet to produce subsidence and reduced rainfall to the west of the monsoon. Here, the authors analyze this proposed mechanism in terms of changes to the thermodynamic energy equation, regional circulation, and precipitation between the 10-day periods before and after the monsoon onset, for all onset dates in the 1958–2000 period. A Rossby-like response to the monsoon onset is clear in the observational data and is associated with horizontal temperature advection at midlevels as the westerlies intersect the warm temperature anomalies of the Rossby wave. Analysis of the thermodynamic equation verifies that the horizontal temperature advection is indeed balanced by subsidence over areas of North Africa, the Mediterranean, and the Middle East, and there is an associated decrease in precipitation over those regions. Despite the increased subsidence, diabatic heating changes are small in these regions so diabatic enhancement does not appear to be a primary factor in the response to the onset. This analysis also shows that the same processes that favor subsidence to the west of the monsoon also force rising motion over northern India and appear to be an important factor for the inland development of the monsoon. Comparison of strong and weak onsets further validates these relationships.

1. Introduction

The convection associated with the Indian monsoon comprises the largest single component of atmospheric heating (e.g., Yanai and Tomita 1998), with an upper-level circulation that extends around nearly half the globe (Hoskins and Rodwell 1995). The rainfall associated with the Indian monsoon exhibits large interannual variability that has a notable impact on agricultural productivity in one of the most heavily populated areas of the world (Webster et al. 1998; Abrol and Gadgil 1999). The monsoon appears to be influenced by climate change (Meehl and Washington 1993; Kripalani et al. 2003; Goswami et al. 2006; Meehl et al. 2007), further highlighting the need for understanding the regional influence.

The onset of the Indian summer monsoon is associated with large changes in convection and circulation not only over India but also over large areas of the Indian Ocean and surrounding environs (e.g., Ramage 1971; Rao 1976; Hastenrath 1991; Webster et al. 1998; Fasullo and Webster 2003; Wang et al. 2009). The monsoonal rainfall originates over Burma and Thailand around mid-May. Over India, the onset begins in the south near the beginning of June, then develops northwestward over the subcontinent over the course of several weeks. The large-scale convection pattern, however, is dominated by two oceanic maxima, one in the Arabian Sea and one in the Bay of Bengal.

The proximity of the vigorous monsoon circulation to the North African desert, the largest desert in the world, has naturally raised questions of association. Rodwell and Hoskins have proposed a “monsoon-desert hypothesis” (Hoskins 1996; Rodwell and Hoskins 1996; Rodwell and Hoskins 2001) whereby the direct atmospheric response to the monsoon diabatic heating thermodynamically interacts with the mean flow, resulting in downward
motion over North Africa and thus playing a role in desert conditions in that region. Their analysis of average summer conditions shows horizontal temperature advection over North Africa consistent with the hypothesis, and a linear primitive equation model is able to reproduce subsidence over the region in response to monsoon heating in the presence of mean wind. However, the model only reproduces the correct strength of subsidence when the local diabatic cooling over North Africa is included, and they further propose a “diabatic enhancement” process whereby the initial forcing of descent results in clearer skies, which increases radiative cooling, which then feeds back into increased descent.

Analysis of average summer conditions, however, is not straightforward to interpret, as multiple dynamical and radiative factors are important to the summer circulation in addition to the monsoon and it is therefore difficult to separate out the influence solely due to the monsoon. Here, we conduct an observational analysis keyed to the individual dates of the large-scale onset of the monsoon, so that we can analyze the changes directly linked to the monsoon heating and provide a stringent test of the hypothesis. The changes during the onset are considered for the terms of the thermodynamic equation and compared to changes in circulation and precipitation, to test the monsoon-desert hypothesis. The data and methodology are described in the next section.

2. Data and methodology

The reanalysis data from the National Centers for Environmental Prediction (NCEP)—National Center for Atmospheric Research (NCAR) (Kalnay et al. 1996) are used as the primary source of circulation data and as the basis for calculating the terms of the thermodynamic equation. Daily-averaged (0000 and 1200 UTC) data with a horizontal resolution of 2.5° on a regular latitude–longitude grid for the period 1958–2000 are used. The thermodynamic equation is calculated on the following pressure levels (Holton 2004):

$$\frac{dT}{dt} = -\mathbf{V} \cdot \nabla T + \frac{S_p}{c_p} \omega + \frac{J}{c_p},$$

where $T$ is the temperature, $\mathbf{V}$ is the horizontal wind vector, $S_p$ is the static stability parameter, $c_p$ is the specific heat of dry air, and $J$ is the diabatic heating. Static stability is proportional to the vertical gradient of potential temperature, so $S_p$ represents the vertical advection of temperature, accounting for adiabatic changes. The diabatic heating term is calculated as a residual from the thermodynamic equation as in Barlow et al. (2005). We have also repeated the thermodynamic analysis using the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40) 4-times-daily data (Uppala et al. 2005), to verify that the results are not sensitive to dataset or spatial and temporal resolution.

The hydrologic onset and withdrawal index (HOWI) is used to determine the monsoon onset (Fasullo and Webster 2003). It is a diagnostic index of monsoon onset and withdrawal based on changes in the vertically integrated moisture transport over southern India. The advantage of using HOWI-derived monsoon onset dates is that they agree well with other criteria of onset and are objectively derived from reanalysis datasets. Critically, the HOWI focuses on the large-scale onset, which determines the regional atmospheric response, rather than the arrival of precipitation at a particular region, which may not be associated with significant changes at the large scale. We have also repeated our analysis using the monsoon onset dates for Kerala reported by the India Meteorological Department (IMD); the results are similar but generally weaker with less spatial extent, consistent with the use of a local rather than a large-scale metric. The onset changes are considered here as the difference between the 10-day average after the onset (day +1 to day +10) and the 10-day average before the onset (day −10 to day −1). This 21-day period centered on the onset day is long enough for Rossby waves to develop but short enough to isolate the onset signal as distinct from the more general seasonal evolution. The results are not sensitive to the exact length of the averaging period.

To compare how the dynamical links change between strong and weak onsets, we have also developed a simple measure of the strength of the onset. The interannual variability of the Indian monsoon has traditionally been measured in terms of precipitation but it is known to be inhomogeneous and subject to local effects (Webster and Yang 1992). There is a strong relationship between Indian monsoon rainfall and circulation that shows that the interannual variability of the Indian monsoon is largely driven by fluctuations of the monsoon convective heat source over Bay of Bengal and Arabian Sea (Goswami et al. 1999). Therefore, we developed a dynamic index defined using diabatic heating (derived from NCEP–NCAR reanalysis data as described above) during onset at 400 hPa over the monsoon region (10°–25°N and 60°–95°E) as a criterion to differentiate between strong and weak monsoon onset years and then chose 12 strongest and 12 weakest years for our analysis.

Obtaining consistent precipitation estimates for the whole domain of interest is difficult because of the sparse observational coverage in many regions. For precipitation over the eastern Mediterranean and North African regions during the monsoon onset we use Global Precipitation
Climatology Project (GPCP) Pentad data from National Oceanic and Atmospheric Administration (NOAA)’s National Climatic Data Center (NCDC) for a period of 22 yr from 1979 to 2000. For northern India, the Asian Precipitation–Highly-Resolved Observational Data Integration Toward Evaluation of the Water Resources (APHRODITE’s Water Resources) daily precipitation data with a horizontal resolution of 0.25° on a regular latitude–longitude grid for the period 1961–2000 (Yatagai et al. 2009) are used.

3. Results and discussion

The changes in midlevel diabatic heating and upper-level winds associated with the monsoon onset for 1958–2000 are shown in Fig. 1, where the changes are the difference between the 10-day average before the onset and the 10-day average after the onset. [Calculating the changes from the first half (1958–79) and second half (1980–2000) of the data individually confirms that the pattern is stable.] Here, 400 hPa is used as representative of midlevels—changes are similar at 500 hPa. Note that while the monsoon onset is typically defined in terms of land precipitation due to societal relevance, the greatest precipitation, heating, and thus atmospheric forcing occurs primarily over the adjacent ocean. Associated with the diabatic heating in Fig. 1 is a circulation very similar to the Gill–Matsuno Rossby wave circulation response (Matsuno 1966; Gill 1980) that is centered over Southwest Asia and extends over much of North Africa. As is well known, the circulation is anticyclonic at upper levels and cyclonic at lower levels (e.g., Rodwell and Hoskins 2001; Gill 1980; Kershaw 1985; Cadet 1979; Hsu et al. 1999). This change in circulation represents a strengthening and westward extension of the Tibetan high.

The changes in the terms of the pressure-level thermodynamic equation during the onset are shown in Fig. 2 for the NCEP–NCAR (left panels) and ERA-40 reanalysis (right panels). The temperature tendency term (not shown) is negligible with respect to the other terms. The static stability is approximately constant, so that the changes in the vertical velocity term (Fig. 2b) are balanced by changes in temperature advection (Fig. 2a) and diabatic heating (Fig. 2c). As expected, the tropical balance is largely with diabatic heating and the extratropical balance is largely with horizontal temperature advection with some contribution from diabatic heating (Mohanty et al. 1983; Hoskins and Karoly 1981).

The intense upward motion in the Indian Ocean and southeastern Arabian Sea (Fig. 2b) is in balance with the diabatic heating in those areas (Fig. 2c). The ascent over northern India (Fig. 2b), in contrast, is primarily balanced by temperature advection (Fig. 2a) with only a small contribution from diabatic heating (Fig. 2c). Using ERA-40 data, we have verified that, while there are some minor differences (the vertical velocity is somewhat more vigorous in ERA-40 and the monsoon is stronger over the Bay of Bengal) the overall patterns and relationships are quite similar.

To further test the link between convection, temperature advection, and vertical motion, the thermodynamic terms are compared between strong onsets (Figs. 3a–c) and weak onsets (Figs. 3d–f). In the strong cases where the diabatic heating is very large, the cold advection over the Middle East and North Africa is stronger as well, as is the subsidence. In the weak cases, there are still areas of cold advection and subsidence but they are much weaker. The strength and extent of the Rossby-like circulation in wind is also directly related to the strength of the onset (not shown). Thus temperature advection and vertical motion are not only linked to the average onset, but they vary in strength according to the strength of the onset as well.

To further explore the changes in temperature advection, they are divided in Fig. 4 into the contribution from advection of the temperature anomalies by the mean wind, as in the Rodwell–Hoskins hypothesis, and the other component—advection of the mean temperature by the wind anomalies. The midlevel warm temperature anomalies, shown in Fig. 4a, are spatially coherent with the upper-level anticyclone in Fig. 1, as expected from the thermal wind balance. In the middle troposphere, the temperature anomalies lie within the westerlies, thus resulting in cold advection to its west and warm advection to
Fig. 2. Comparison of NCEP–NCAR reanalysis and ERA-40 for the changes in the terms of the thermodynamic equation at 400 hPa associated with the large-scale onset: (a),(d) the temperature advection term (red represents warm advection and blue represents cold advection), (b),(e) the vertical velocity term (blue represents rising motion and red represents subsidence), and (c),(f) the diabatic heating term (red represents positive values and blue represents negative values). The contour interval is 0.7 K day$^{-1}$ throughout and the onset changes are determined as in Fig. 1.

From an isentropic perspective, as noted by Rodwell and Hoskins (1996, 2001), the warmer temperatures pull isentropic surfaces downward and the jet winds will thus be deflected down (isentropic downgliding) on the western side of the anomaly, resulting in descent over North...
Africa and deflected back up (isentropic upgliding) on the eastern side of the anomaly, resulting in rising motion over northern India as evident from our results (Figs. 2 and 4).

Given that the circulation and temperature anomalies are consistent with the Rossby wave response, are directly tied to the monsoon heating in this onset analysis, and are in close accordance with theoretical and modeling expectations for the response to subtropical heating, we interpret the warm temperature advection over North Africa and the Middle East (Fig. 2a) and the
resultant necessary balance in descending motion (Fig. 2b) as forced by the monsoon diabatic heating. This provides direct observational confirmation of the basic thermodynamic mechanism of the monsoon-desert hypothesis. There are, however, the following two caveats—one, the effect during the onset is largely limited to eastern North Africa and the Middle East; and two, changes to diabatic heating in those regions are small during the onset (Fig. 2c)—so the role of a local diabatic feedback does not appear to be a primary factor here. The onset changes considered here are the difference between the 10-day averages before and after the onset, so there should be ample time for radiative effects to develop. Although there is significant diabatic cooling over North Africa and the eastern Mediterranean during this period, it does not appear to change significantly in association with the onset, probably because of the small total values of precipitation.

The percentage change in precipitation rate at monsoon onset is plotted for the descent region in Fig. 5. While North Africa and the Middle East receive relatively modest total amounts of precipitation, it is of considerable local societal significance; therefore, using percentage change instead of total precipitation for these regions is more useful for the present study. Figure 5 shows that precipitation is suppressed in North African and eastern Mediterranean regions during monsoon onset. This remote effect observed at the time of Indian monsoon onset enhances drier conditions in North African and eastern Mediterranean deserts, which agrees well with theoretical and modeling studies. However, precipitation in these regions has considerably decreased even before the monsoon onset in most years, so the net effect on seasonal precipitation may be modest.

Additionally, we would like to emphasize the converse of the hypothesis considered so far, which is the link between warm advection (Fig. 2a) and associated rising motion (Fig. 2b) over northern India. From the diabatic heating in Fig. 2d we can see that, during the large-scale onset, the monsoon has not yet developed strongly into northern India. However, strong increases in rising motion are already beginning to develop in that region (Fig. 2a) as a result of temperature advection, and therefore...
the thermodynamic mechanism considered here may also be an important part of the northward development of the monsoon. To analyze this further, we calculate the long-term composite of northern Indian precipitation and the northern Indian horizontal temperature advection averaged over latitudes 20°–30°N and longitudes 70°–90°E relative to the HOWI monsoon onset dates varying from 15 days before onset to 60 days after onset (Fig. 6). Before the monsoon onset, the horizontal temperature advection is negative, consistent with subsidence and the inhibition of local precipitation. After the monsoon onset, the horizontal temperature advection rapidly increases as the warm anomalies of the Rossby circulation response develop, and becomes slightly positive (mean rising motion), allowing development of the northern arm of the monsoon. That is, the warm temperature advection associated with the Rossby response to the large-scale onset counteracts the average subsidence that is present before the local onset; thus removing a large inhibiting factor for the development of precipitation over northern Indian. There is some suggestion of a second phase of precipitation increase about 3 weeks after the initial local onset; this feature, if robust, does not appear to be linked to temperature advection.

4. Summary

Analysis of the changes in circulation, thermodynamics, and precipitation relative to individual monsoon onset dates allows for a direct observational evaluation of the Rodwell–Hoskins ‘‘monsoon-desert’’ hypothesis. We confirm that the thermodynamic interaction of the Gill-like response to the monsoon heating with the jet stream does result in significant cold advection, descent, and suppressed precipitation over eastern North Africa and parts of the Middle East. However, we find little evidence of local diabatic enhancement in those regions, at least during the onset phase. We have also compared these interactions during weak and strong onsets, which show that stronger convection during the onset results in stronger cold advection and subsidence, further verifying the hypothesized mechanism. Preliminary thermodynamic analysis of the monsoon retreat suggests the same processes may also be important at the end of the summer dry season in those regions, and merits further investigation.

Additionally, the other implication of the monsoon-desert hypothesis—warm advection and rising motion over northern India—appears to be important in the northward continental development of the monsoon after the initial large-scale onset. Our results show that precipitation over northern India is inhibited by subsidence until the onset of the large-scale monsoon and the associated warm advection, which raises a new dynamic mechanism to be considered in understanding the northward progression of the monsoon with time.

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REFERENCES


