Convective Asymmetries Associated with Tropical Cyclone Landfall.
Part I: f-Plane Simulations

JOHNNY C. L. CHAN
Laboratory for Atmospheric Research, Department of Physics and Materials Science, City University of Hong Kong, Hong Kong, China

XUDONG LIANG
Laboratory for Atmospheric Research, Department of Physics and Materials Science, City University of Hong Kong, and Shanghai Typhoon Institute, Shanghai, China

(Manuscript received 22 March 2002, in final form 6 January 2003)

ABSTRACT

This study investigates the physical processes associated with changes in the convective structure of a tropical cyclone (TC) during landfall using the fifth-generation Pennsylvania State University–National Center for Atmospheric Research Mesoscale Model, version 3 (MM5). The land surface is moved toward a spunup vortex at a constant zonal speed on an f plane. Four experiments are carried out with the following fluxes modified over land: turning off sensible heat flux, turning off moisture flux, setting a higher surface roughness, and combining the last two processes.

The results suggest that sensible heat flux appears to show no appreciable effect while moisture supply is the dominant factor in modifying the convective structure. Prior to landfall, maximum precipitation is found to the front and left quadrants of the TC but to the front and right quadrants after landfall when moisture is turned off and surface roughness increased.

To understand the physical processes involved, a conceptual experiment is carried out in which moisture supply only occurs over the ocean and at the lowest level of the atmosphere, and such supply is transported around by the averaged circulation of the TC. It is shown that the dry air over land is being advected up and around so that at some locations the stability of the atmosphere is reduced. Analyses of the data from the more realistic numerical experiments demonstrate that convective instability is indeed largest just upstream of where the maximum rainfall occurs. In other words, the effect of the change in moisture supply on the convection distribution during TC landfall is through the modification of the moist static stability of the atmosphere.

1. Introduction

The problem of tropical cyclone (TC) landfall has received more attention in recent years especially after it has been listed as one of the foci of the U.S. Weather Research Program (Marks et al. 1998). The concerns include track oscillations, wind and rainfall distributions, storm surges, etc. Most observational research has been based on individual case studies (e.g., Miller 1964; Parrish et al. 1982; Powell et al. 1982; Powell and Houston 1996, 1998; Geerts et al. 2000; Blackwell 2000; Ching et al. 2000; Ching 2002). While these studies point out peculiarities of meteorological parameters associated with particular TCs, or even common features associated with different TCs, the understanding of the underlying physical processes is quite difficult. Numerical modeling studies therefore become a useful tool in this respect.

Two of the earlier modeling efforts by Ooyama (1969) and Rosenthal (1971) investigated the sensitivity of TCs to surface boundary conditions using a simple model. Ooyama (1969) noted that the cyclone-scale circulation depends on the convection for the release of heat energy, while the continued and organized activity of convective clouds in the storm must depend on the large-scale circulation for efficient supply of water vapor. It is also unlikely that convective activity in a column would be supported mainly by the evaporative supply of water vapor from the part of the sea surface directly below the column. However, the evaporation from a large area under the TC circulation is extremely important for supporting the convective activity in the central region of the cyclone. Tuleya and Kurihara (1978, hereafter TK78) simulated TC landfall using a three-dimensional, primitive equation model and investigated the effects of land friction and evaporation through three experiments: friction on land without evaporation, smooth land with-
out evaporation, and increase in surface roughness with evaporation. Their study suggested that evaporation depletion over land is the single most important parameter in the decay sequence during landfall. Increased surface roughness tends to augment boundary layer inflow, which enhances kinetic energy generation so that the effect of surface dissipation is partially offset by this generation. They also found that during TC landfall, the forward-right quadrant of the TC is a prime location of maximum precipitation intensity. To have better resolution, Tuleya et al. (1984, hereafter TBK84) used a more sophisticated model with triple-nested movable domains. A moist land with a different surface temperature than the ocean, and a roughness parameter of 25 cm, were specified in that study. A constant steering flow and a variable Coriolis parameter were also included. They found that the evolution of the wind and pressure fields and the rainfall distribution were significantly changed by different land surface conditions. Tuleya (1994) used an improved version of the Geophysical Fluid Dynamics Laboratory TC model that includes diurnal radiation and a bulk subsurface layer with explicit prediction of land temperature to test the sensitivity of surface boundary conditions to tropical cyclone development and decay at landfall. He found that the thermal property of the subsurface had a dramatic influence on TC development.

While these numerical studies have shown the effects of land conditions on the TC structure, the underlying physical processes are still not clearly understood. In particular, changes in one condition often affect other conditions. For example, the primary effect of an increase in roughness is to reduce the wind speed. However, such a reduction leads to a concomitant decrease in heat and moisture fluxes so that it is difficult to isolate the physical processes involved.

This study therefore represents a numerical modeling effort to investigate the physical processes that lead to changes in the convective distribution in a TC prior to, during, and after landfall under idealized conditions. The mesoscale model used is briefly described in section 2. Results of a control run are also presented, not only to demonstrate the validity of the model, but also to serve as a basis for comparison with other experiments. The basic effects of sensible heat, moisture (latent heat), and momentum fluxes on modifying the convective distributions are described in section 3. While the results appear to be reasonable and consistent with a number of observations, they also deviate from the conclusions of some modeling studies. To investigate this apparent discrepancy, a conceptual experiment is introduced in section 4 to understand the fundamental physical processes involved. Based on the result from this experiment, the numerical data generated in section 3 are analyzed to demonstrate that these processes can be used to explain the present simulation results. Section 5 gives a summary of the study together with future investigation directions.

2. The model and control run

Version 3 of the fifth-generation Pennsylvania State University–Mesoscale National Center for Atmospheric Research Model 5 (MM5; Dudhia 1993; Grell et al. 1994) is employed. There are $301 \times 301$ grid points in the x and y directions with a grid spacing of 15 km. The vertical coordinate consists of 16 $\sigma (=\rho/\rho_0)$ levels, 1.0, 0.99, 0.97, 0.94, 0.90, 0.85, 0.79, 0.71, 0.61, 0.49, 0.37, 0.27, 0.18, 0.10, 0.04, and 0.00. The value of the Coriolis parameter f is fixed at 20°N. The physics options include the Betts–Miller cumulus parameterization (Betts and Miller 1993; Janjic 1994), simple ice scheme (Dudhia 1989), and high-resolution Blackadar planetary boundary layer (PBL) scheme (Zhang and Anthes 1982; Blackadar 1979).

The initial conditions of all experiments are obtained by integrating for 36 h a prespecified vortex in a quiescent atmospheric environment, in the same way as described in Chan et al. (2001). The prespecified vortex has an initial minimum sea level pressure (MSLP) of 980 hPa and a radius of 15 m s$^{-1}$ winds of 250 km. The vertical temperature structure of the environment is obtained from the reanalysis data of the European Centre for Medium-Range Weather Forecasts for a particular day over the western North Pacific and the humidity values are specified. For further details, the reader is referred to Chan et al. (2001).

Since the TC cannot move on an f plane, the “coastline” is moved in the model domain, similar to the strategy used in TK78. The land and sea parts of the domain (separated by the coastline) are characterized by different values of various PBL parameters depending on the experiment to be performed. The coastline is oriented north–south with a movement of 5 m s$^{-1}$ (i.e., towards the TC from west to east). As pointed out by TK78, moving the coastline toward the vortex avoids any complication involving a steering current. Peng et al. (1999) and Chan et al. (2001) indeed found that a uniform steering flow could change the vortex intensity in their respective models. However, by eliminating the steering current, the effect of asymmetries arising from the superposition of a steering current onto a vortex, which is present in the real atmosphere, cannot be simulated. Nevertheless, it is felt that since the objective here is to examine the effects of the changes in the various fluxes over land on the convective structure of the vortex, the lack of such asymmetries should still be acceptable, and in fact desirable. Inclusion of these asymmetries will be left to future studies.

The control experiment (to be labeled as experiment 1 for future reference—see Table 1) is run with ocean PBL conditions throughout the domain. Chan et al. (2001) have demonstrated that the model is capable of

\footnote{It should be pointed out that directions such as east and west are for reference only and do not have much physical meaning since all experiments are performed on an f plane.}
developing and sustaining the vortex. An important analysis in the control run is the temporal variation of rainfall distribution near the TC center (Fig. 1), which is obtained by summing the hourly rainfall within each 15-km, 1° azimuth box (centered on the TC center) out to 300 km. The radius of 300 km is chosen because almost all of the rainfall occurs within this radius (the 10-cm isohyet after 36 h of integration in most of the experiments has roughly a radius of this value). It is obvious from Fig. 1 that the maximum rainfall areas rotate cyclonically around the TC center.

### 3. Basic experiments on the effects of PBL fluxes

Three properties can be exchanged between the atmosphere and its underlying surface through their respective fluxes: sensible heat, moisture (latent heat), and momentum. Therefore, to examine the changes in the characteristics of the TC that is about to make landfall, the following experiments are performed (see also Table 1): experiment 2—turning off the sensible heat flux over land; experiment 3—turning off the moisture flux over land; experiment 4—setting the roughness length to be 0.25 m over land; and experiment 5—combining the effects in experiments 3 and 4. As in TK78, landfall is defined as the instant when the TC center encounters the coastline.

In the high-resolution Blackadar PBL scheme (Grell et al. 1994), the friction velocity $u_*$ is computed from

$$u_* = \max \left( \frac{\kappa V}{\ln \left( \frac{z}{z_0} \right)} - u_{*0} \right),$$

where $\kappa$ is the von Kármán constant, $z$ is the height above the surface, $z_0$ is the roughness length, and $V$ is the wind speed at a reference height. This equation accounts for the variation of wind speed with height and the influence of surface roughness on the friction velocity.

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**TABLE 1. Location of the TC center in the model domain in experiments 2, 3, and 4. The coordinates are from the bottom-left corner of the domain.**

<table>
<thead>
<tr>
<th>Experiment</th>
<th>36 h</th>
<th>72 h</th>
</tr>
</thead>
<tbody>
<tr>
<td>Expt 2</td>
<td>(151, 151)</td>
<td>(151, 150)</td>
</tr>
<tr>
<td>Expt 3</td>
<td>(151, 151)</td>
<td>(151, 148)</td>
</tr>
<tr>
<td>Expt 4</td>
<td>(151, 151)</td>
<td>(151, 149)</td>
</tr>
</tbody>
</table>

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**FIG. 1.** Total hourly rainfall within 300 km at each azimuth from 1 to 36 h after the vortex has been spun up for 36 h (i.e., 1 h means 37 h of integrating time from the initial conditions) for the control run (experiment 1).
where \( k \) is the von Kármán constant, \( \psi \) a nondimensional stability parameter that is a function of the bulk Richardson number, \( u_0 \) a background value (0.1 m s\(^{-1}\) over land and zero over water), and \( V \) is given by
\[
V = (V_0^2 + V_0^2)^{1/2},
\]
with \( V_0 \) being the wind speed at the lowest model layer, and \( V_0 \), a convective velocity defined under unstable and neutral conditions as
\[
V_0 = 2(\theta_g - \theta_a)^{1/2},
\]
which is equal to zero under stable conditions, \( \theta_g \) the potential temperature at the lowest model layer and \( \theta_a \) that at the ground, \( z_0 \) the height of the lowest \( \sigma \) level, and \( z_0 \) the roughness parameter. Over land, \( z_0 \) is specified as a function of land-use category. Over water, \( z_0 \) is calculated as a function of friction velocity such that
\[
z_0 = 0.032 u_0^2/g + z_0, \tag{3.4}
\]
where \( z_0 \) is a background value of 10\(^{-4}\) m and \( g \) the acceleration due to gravity.

In all the experiments, land use at each grid point is set to be water body, and the surface temperature is inviable with time during the integration. Therefore, the surface temperature will not be affected by the changes in the heat flux in the PBL. In experiment 2 (turning off the sensible heat flux over land) only the sensible heat flux is set to zero over “land” and nothing else is changed. Similarly, only the latent heat flux is set to zero in experiment 3 (turning off the moisture flux over land). The land in each experiment is therefore not real land but only an area over which one or more parameters is/are changed, because the land use is still the water body. In other words, when the sensible heat flux is set to zero in experiment 2, the roughness length \( z_0 \) is also calculated according to (3.4). In experiment 4 (friction experiment), when calculating the momentum flux, the roughness length \( z_0 \) over the land is set to 0.25 m [that over the sea still calculated according to (3.4)]; however, when calculating the sensible and latent heat fluxes, \( z_0 \) is calculated according to (3.4) for the entire domain (i.e., both over land and sea). This way, the effect of friction on the momentum transfer can be isolated from the effect due to heat exchanges.

\( a. \) Effect of sensible heat flux

In experiment 2, the sensible heat flux is turned off over land while that over the sea surface is the same as in the control run. The largest departure of total rainfall after 36 h in this experiment from the control one is only 40 mm (not shown), which is very small relative to the total rainfall of >1000 mm in the inner area. The effect of sensible heat flux change due to landfall is therefore apparently not evident.

\( b. \) Effect of moisture (latent heat) flux

In a TC, although most of the latent heat is released in the areas near the eyewall and the moisture can be transported into the TC through inflow in the lower troposphere, the bulk of the water vapor comes from the PBL through moisture flux. Therefore, cutting off the moisture flux over land should have a profound effect on the TC characteristics.

Significant differences between this experiment (experiment 3) and the control run begin to occur at \( \sim 12 \) h (216 km) before landfall when the maximum rainfall occurs to the west (i.e., ahead) of the TC (Fig. 2). Then, another maximum to the north of the TC begins to rotate westward and by 6 h (108 km) before landfall, the maximum rainfall is located in the west to southwest quadrant (i.e., in the forward and left quadrants relative to the TC motion). This maximum remains stationary in this quadrant up to 9 h after landfall, with the highest value occurring due west at around the landfall time.

This result appears to be consistent with time-lapse animations of radar reflectivities associated with Hurricane Andrew (1992) that show a convective cycle with cells forming on the north side of the eyewall, maturing on the west and south sides, and decaying on the east side (Dodge and Burpee 1993; Willoughby and Black 1996). A recent study of radar reflectivities associated with four typhoons making landfall near Hong Kong also showed a maximum in convection to the west (left) of the typhoons at and after landfall (Ching 2002).

It is interesting to note that even when the moisture flux is turned off over land, the rainfall in some areas is still increased (not shown). TK78’s experiment also suggested a small increase in the average precipitation rate in the model storm near landfall.

\( c. \) Effect of surface friction

In this experiment (experiment 4), the roughness length over land is set at 25 cm (same as that in TK78) while the value over the sea is calculated according to (3.4) for the rest of the domain. The temporal variations of rainfall indicate that when the TC reaches about 108 km (6 h) from the coast, the rainfall in the sector from the northwest to west and south is enhanced (Fig. 3). After landfall, the maximum rainfall is located in the west and southwest (i.e., again, forward and left) quadrants. The difference in the total rainfall between this experiment and the control run shows a steady increase from the time when the TC is still nearly 300 km from the coast, reaching a maximum about 3 h after landfall (Fig. 4). The difference then begins to decrease although the rainfall in experiment 4 is still larger than that in the control run even at 8 h (144 km) after landfall. The distributions of surface wind speeds (decrease over land), divergence and vorticity [both parameters being positive (negative) in the offshore (onshore) sector] have similar patterns as those in TK78 (and hence not shown).
d. Changes in track and intensity

Since the experiments are all on an $f$ plane, any displacement of the vortex must be due to the specific changes in the various fluxes. However, it is observed that the track changes are minimal (Table 1), with experiment 3 giving the largest southward displacement of 45 km in 36 h. This suggests that in the absence of other factors, changes in the track due to landfall are generally not significant, which is consistent with the result of TBK84.

On the other hand, since the amount of latent heat release varies among the experiments, the MSLP also varies (Table 2). While the vortex in the control experiment (experiment 1) continues to deepen slightly, that in experiment 2 maintains its intensity. If the moisture is cut off (experiment 3) or the roughness length increased (experiment 4), the vortex weakens by $\sim 10$ hPa in 36 h. The result from experiment 3 agrees with that of TK78, but that in experiment 4 does not. One possibility for this discrepancy is in TK78’s statement that “evaporation was not suppressed” (section 5b in TK78’s paper). This is interpreted to mean that even when the surface roughness was increased, the moisture flux was maintained. As a result, an increase in surface
friction would enhance the amount of surface convergence through Ekman pumping so that more moisture became available, which led to an increase in the intensity of the modeled TC. In experiment 4 of the current study, the moisture flux is reduced because of the decrease in surface wind, which leads to a decrease in intensity.

e. Combining experiments 3 and 4

In TK78’s basic experiment, the total rainfall before TC landfall tends to increase to the right of the storm track where winds are onshore. However, a careful examination of the locations of maximum rainfall (their Fig. 10) suggests that these locations actually tend to occur more often to the left of the TC. During and after landfall, the location of maximum precipitation is biased toward the forward-right quadrant of the storm. In the current experiments 3 and 4, the maximum rainfall generally exists ahead and in the forward-left quadrant of the TC near, during, and after landfall. In other words, an apparent discrepancy between the current study and TK78 exists for the period near and after landfall, which may be due to the fact that in experiments 3 or 4, only one of the fluxes is specified to be different over land.

To explore this further, an experiment (experiment 5) similar to the basic landfall experiment of TK78 is designed, with the moisture flux being cut off and the roughness length set to 0.25 m over land. Prior to landfall, the location of maximum precipitation still occurs
in the forward and forward-left quadrants (Fig. 5), similar to the results of experiments 3 and 4. However, after landfall, the location of maximum rainfall shifted to the forward-right quadrant of the TC. Thus, to a large extent, the current result is consistent with that of TK78. In other words, a cutoff in moisture together with an increase in roughness tends to shift the maximum precipitation at landfall to the forward-right quadrant. The physical processes involved are to be discussed in the next section.

4. Effect of moisture supply

Two main results from the last section are: 1) the maximum rainfall tends to rotate cyclonically, which suggests that the rainfall variations should depend on the three-dimensional wind and moisture distributions of the TC; and 2) prior to and during landfall, the maximum rainfall primarily occurs in the forward and forward-left quadrants. The results from experiment 3 (no moisture flux over land) further suggest that the convergence (divergence) in the onshore (offshore) area does not appear to play a key role in determining the precipitation distribution, since no change in the wind field exists in this experiment and only the moisture flux is cut off over land. Furthermore, the rainfall distributions in experiments 3, 4, and 5 before landfall are very similar, which suggests that the moisture flux is very important in determining the rainfall pattern before and during TC landfall since in all three experiments, the common element that is modified is the moisture flux. In a case study of Hurricane Donna of 1960, Miller (1964) also concluded that the removal of oceanic heat source was the primary reason for the weakening of Donna.

However, the effect of a change in moisture flux in the convective structure is obviously not local since the rainfall pattern is not uniform over land. In other words, some other physical processes must be operating to produce the simulated rainfall distribution when the moisture flux is cut off or decreased. To address this issue, a conceptual experiment is performed to understand the

<table>
<thead>
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<th>Experiment</th>
<th>36 h (hPa)</th>
<th>72 h (hPa)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Expt 1</td>
<td>942</td>
<td>939</td>
</tr>
<tr>
<td>Expt 2</td>
<td>942</td>
<td>941</td>
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<tr>
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<td>952</td>
</tr>
<tr>
<td>Expt 4</td>
<td>942</td>
<td>954</td>
</tr>
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</table>
effect of moisture supply on the convective structure associated with a landfalling TC.

a. Conceptual experiments

To study the effect of moisture flux, a conceptual experiment is designed such that the variation of water vapor (represented by the specific humidity $q$) at each grid point is determined by the three-dimensional advection and surface moisture flux:

$$\frac{dq}{dt} = \left[ \frac{4.67 \times 10^{-4}}{\exp\left(\left(\frac{i - ic}{22}\right)^2 + \left(\frac{j - jc}{22}\right)^2\right)} \right]_{i, j}$$

where $(ic, jc)$ is the grid point of the TC center. Since $dq/dt$ exists only at the lowest level ($\sigma = 1$), the qualifier $\mid_{\sigma=1}$ for $dq/dt$ will be omitted for simplicity. The value of $dq/dt$ calculated by (4.2) is very close to the moisture flux in experiment 1 at the initial time (Fig. 6). Note also from (4.2) that the value of $dq/dt$ is independent of time.

The discrete form of (4.1) is

$$q_i = q_{i-1} - \left[ \left( \frac{\Delta q}{\Delta x} - \nu \frac{\Delta q}{\Delta y} - w \frac{\Delta q}{\Delta z} \right)_{i-1} + \frac{dq}{dt} \right]_{i-1} \Delta t,$$

where $\Delta t$ is the length of the time step. The domain used to solve (4.3) is similar to that in experiment 1 (control run) except that the domain size is smaller. The horizontal grid spacing is 15 km with $113 \times 113$ grid points, and 15 vertical levels. The TC center is fixed at the middle of the domain ($ic = 57, jc = 57$).

The initial value of $q$ at each grid point and level is equal to the averaged specific humidity between 1 and 36 h in experiment 1. The time step is 90 s, and at each time step the “moisture flux” at the lowest level is calculated from (4.2). Because no prognostic equation is available to predict the temporal evolution of the wind fields, the wind fields of the control run averaged between 1 and 36 h are used to calculate the advective quantities. In other words, in solving (4.3), the three-dimensional wind fields do not change with time.

The solution of (4.3) for the control run (experiment 1) after 36 h (to be labeled as experiment 6) gives the expected result that $q$ is advected from the outer area to the eyewall of the TC in the lower levels and subsequently moved up to higher levels (Fig. 7). To simulate the effect of moisture flux change due to the encounter with land, another experiment (experiment 7) is run with $dq/dt$ set to zero over land and the coastline moved from west to east at a speed of 5 m s$^{-1}$ (i.e., similar to experiment 3). Because the coastline moves towards the TC, $dq/dt$ is also changed when the sea surface becomes the land surface. Therefore, $q$ at the lowest level should be reduced over land. Subtracting the value of $q$ in experiment 7 from that in experiment 6 will then give a negative value, which will be termed negative anomaly. These negative anomalies will be advected around by the three-dimensional flow.

After 36 h, the moisture anomalies show that moisture flux is reduced in the sector of the TC over land (Fig. 8b). Further, these anomalies are advected from the outer into the inner area through convergence at the lower levels and then vertical advection (Fig. 8a). Note that only negative anomalies exist since no mechanism exists to increase the moisture.

The time variations of $q$ suggest that prior to landfall, negative anomalies of $q$ due to the cutoff in moisture flux over land occur in the southwest at the surface (9 h; Fig. 9) and in the southeast at the lower levels (9 h, 3; Fig. 10). The maximum negative anomalies are found to the east and northeast at the midtroposphere ($\sigma = 5$ and 7; Fig. 10). In other words, the maximum negative anomaly is rotated cyclonically upward from south to north when the TC moves from sea to land. These anomaly distributions have a profound effect on the stability of the atmosphere. For example, to the north and west, water vapor is reduced in the midtroposphere but remains at about the same value as in experiment 6 at the low levels (because of moisture advection from the ocean). As a result, the atmosphere in the northwest to southwestern quadrants becomes more unstable, which then leads to stronger convection, and hence more rainfall. After landfall, these areas should still remain unstable but since the reduction of water vapor now also occurs at the lower levels in the north and east, the rainfall should be reduced compared to the situation...
before landfall. When the TC moves farther inland, the rainfall should become less and less, and the TC should weaken.

Because of horizontal and vertical advection, the three-dimensional anomalies shown in Figs. 9 and 10 rotate cyclonically as the TC moves from sea to land. However, this is not exactly the case in experiment 3 (see section 4b below). The reason for this difference is because in experiment 7, the negative anomalies, once generated, will always be present (since no mechanism exists to cause their removal or destruction) and are advected around by the cyclonic flow and vertical motion of the TC. As a result, these anomalies are accumulated and rotated cyclonically. In the more realistic case (experiment 3), the water vapor anomalies would be vertically advected by upward motion, and removed by rainfall and other mechanisms such as diffusion etc. Nevertheless, the results in these conceptual experiments can be used to provide an understanding of the rainfall distributions associated with TC landfall, as will be seen in the next section.

b. Analyses of the data in the numerical experiments

In interpreting the diagnoses of the data from the numerical experiments, the following conceptual framework needs to be established. First, it is assumed that a reduction of moisture flux from the land surface will likely decrease the amount of water vapor in the air, and this drier air is subsequently advected into the TC center. Second, the drier air may not always decrease the precipitation, since it can be advected to higher levels,
which then increases the instability of the whole column of air, and enhances the upward motion of the moist air advected from the sea surface. Finally, rain occurs only when the air is raised up to a certain height (way above the lifting condensation level) and not at the very low levels.

The potential pseudoequivalent temperature $\theta_w$ is given by (Bolton 1980):

$$
\theta_w = T \left( \frac{1000}{p} \right)^{0.2857(1 - 0.28 \times 10^{-3}r)} \times \exp \left( \frac{3.376}{T_L} - 0.00254 \right) \times r(1 + 0.81 \times 10^{-3}r), \tag{4.4}
$$

where $p$ is the pressure (hPa), $r$ the mixing ratio (g kg$^{-1}$), $T$ the temperature (K), and $T_L$ the temperature at lifting condensation level computed from

$$
T_L = \frac{2840}{3.5 \ln T - \ln e - 4.805} + 55, \tag{4.5}
$$

where $e$ is the water vapor pressure (hPa) given by

$$
e = \frac{pr}{622 + r}. \tag{4.6}
$$

Since the configuration in the conceptual experiment in section 4a is closest to that in experiment 3 (no moisture flux over land), the values of $\partial \theta_w / \partial z$ (used as indicator of instability) are first calculated at 1-h intervals using the data from experiment 3. The asymmetric horizontal distribution of $\partial \theta_w / \partial z$ and $q$ (specific humidity) are then computed by subtracting the wavenumber 0 (i.e., the axially symmetric component) from the total value. Since the heaviest rainfall and the strongest vertical motion occur within ~100 km from the TC center (not shown), the values of these two quantities within 150 km are examined.

As might be expected from the results of the conceptual experiment, prior to landfall, dry air from the land is advected into the inner area of the TC in the south and east at the low levels and then raised up in the east and north (18 h; Fig. 11). At the same time, moist air from the sea is advected into the inner areas in the north and west at the low levels and raised up in the west and south. As the distance between the TC and the coastline decreases (9 and 18 h; Fig. 11), the negative asymmetries (dry air) at the low levels extend counterclockwise from south to east and north. However, the negative asymmetries in the midtroposphere reach farther downstream due to upward motion and horizontal advection. For example, at 18 h, the negative asymmetries cover from the south to east and north at $\sigma = 3$, while they are over the east and north at $\sigma = 5$, and even reach to the west at $\sigma = 7$. At around landfall time (27 h; Fig. 11), moisture asymmetries at the low levels are positive (negative) in the north and west (south and east) but are nearly opposite in the midtroposphere. After landfall (36 h; Fig. 11), the distributions of moisture asymmetries remain generally the same as at 27 h except in the inner areas (within 50 km) where the moist (dry) air reaches the southern (northern) part of the TC at the low levels, but dry (moist) air is still present in the northwest (southeast) at $\sigma = 7$. 

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**Fig. 6.** Values of the moisture flux (solid; unit: $10^{-4}$ kg kg$^{-1}$) along the line across the TC center in the zonal direction defined by Eq. (4.2). The dashed line is the value (unit: kg m$^{-2}$) in the initial conditions of the control run (experiment 1). The abscissa is the grid number starting from the west side of the domain.
The vertical distributions of dry and moist air will decrease or increase the instability, which can be evaluated by examining the asymmetric distributions of $\partial \theta_w / \partial z$. The asymmetric distributions of $q$ in Fig. 11 suggest that the distributions of instability in each level will be changed. In experiment 3, before landfall (18 h; Fig. 12), the atmosphere is more unstable to the east at $\sigma = 3$. However, the area of negative asymmetric $\partial \theta_w / \partial z$ covers from east to north and west at $\sigma = 5$ and north to west and south at $\sigma = 7$. It is obvious that this “band of instability” is rotated from the outer area at the low levels to the inner areas in the midtroposphere through horizontal and vertical advection. At around landfall (27 h; Fig. 12), instability is mostly confined to the outer areas to the north, east, and south at $\sigma = 3$, but covers most of the northeast to northwest at $\sigma = 5$ and almost the entire western half at $\sigma = 7$. The patterns are generally similar after landfall (36 h; Fig. 12). The increased (decreased) instability will lead to increase (decrease) of upward motion, so that enhanced upward motion begins to the east and north of TC at the lower levels and
Fig. 9. Distribution of negative anomaly of $q$ (unit: kg kg$^{-1}$) in experiment 7 on the lowest $\sigma$ level at (a) 9, (b) 18, (c) 27, and (d) 36 h. The abscissa and ordinate are the zonal and meridional distances (km) relative to the TC center, which is represented by the typhoon symbol. The straight long-dashed line indicates the coastline.

migrates to the west and south in the midtroposphere. Since rainfall occurs only after the air is raised to certain heights and not at the very low levels, enhanced rainfall is found to the northwest and southwest of the TC.

An examination of the corresponding distributions of moisture and instability for experiments 4 and 5 gives essentially the same patterns. The major difference in experiment 4 (increased friction over land) is that because the moisture flux over land is only reduced rather than cut off, the moisture and instability asymmetries tend to be smaller in magnitude (not shown). In experiment 5 (moisture cut off over land plus increased friction), the coupling of these two effects created very similar asymmetric distributions to those in experiment 3 (cf. Figs. 12 and 13) except for some details. At around and after landfall, the maximum negative asymmetries in experiment 3 (27 and 36 h; Fig. 12) at $\sigma = 7$ extend more southwestward than those in experiment 5 (27 and 36 h; Fig. 13). As a result, after landfall, the maximum rainfall shifts towards the northwest in experiment 5 (see Fig. 5) compared with that in experiment 3 (see Fig. 2). These results also suggest that, before and during TC landfall, the coastal convergence, which exists in experiments 4 and 5 to the north of the TC, is not very important in determining the rainfall distributions. Of course, after landfall, friction over land has more distinct effects because a large part of the TC circulation is over land and affected through stronger friction. Such a decay process after landfall is more complicated and will be considered in future studies.

5. Summary and discussion

a. Summary

This study investigates the physical processes associated with changes in the convective structure of a tropical cyclone (TC) during landfall using the fifth-generation Pennsylvania State University–National Center for Atmospheric Research Mesoscale Model,
version 3 (MM5). The land surface with properties different from the ocean is moved toward a spunup vortex at a constant zonal speed of 5 m s$^{-1}$ on an $f$ plane. Four experiments are carried out with the following modifications over land: turning off sensible heat flux, turning off moisture flux, setting a higher surface roughness, and combining the last two processes.

The results suggest that sensible heat flux appears to show no appreciable effect while moisture supply is the dominant factor in modifying the convective structure. In particular, prior to landfall, maximum precipitation is found to the front and left quadrants of the TC in all the experiments. If only the moisture is cut off or the surface roughness is increased, the location of such an asymmetry is generally maintained during and after landfall. If both of these processes are included, the maximum rainfall shifts toward the front and right quadrants, in agreement with the studies of TK78 and...
TBK84. Such asymmetries in the rainfall distribution suggest that the effect of moisture supply is not local. Furthermore, the convergence (divergence) along the coastline is not very important for producing the rainfall anomalies in the inner area of TC because the maximum precipitation does not necessarily occur on the converging (onshore) side before landfall, and of the fact that in the moisture flux experiment (experiment 3), no extra convergence exists along the onshore side.

To understand the physical processes involved, a conceptual experiment is carried out in which moisture supply only occurs over the ocean and at the lowest level of the atmosphere, and such supply is transported around by the averaged circulation of the TC. It is shown that the dry air over land (where moisture supply is set to zero) is being advected up and around so that at some locations, the stability of the atmosphere is reduced (dry air overlying moist air).
Based on the result of this experiment, data from the more realistic numerical experiments are examined. Using the three-dimensional distributions of moisture, the vertical variations of saturated equivalent potential temperature are calculated. The results show that instability is indeed the largest around and just upstream of where the maximum rainfall occurs. In other words, the change in moisture supply modifies the moist static stability of the atmosphere and hence the convection distribution during TC landfall.

To summarize, as a TC makes landfall, the convection associated with the TC is modified by the changes in the fluxes over land. The change in sensible heat flux appears to have very little effect. The most important effect is from a reduction in moisture flux, which occurs not only through a decrease in moisture supply from the surface, but also through a decrease in wind speed due to an increase in surface roughness. However, the convergence associated with the decrease in surface wind speed on the onshore side of the TC apparently does

Fig. 12. As in Fig. 10 except for the values of $\partial \theta_e/\partial z$ in experiment 3. Unit: K km^{-1}; contour interval: 1.5 K km^{-1}. 
not produce an increase in rainfall there. Rather, the changes in the rainfall distribution results from the three-dimensional advection of the dry air around the TC, which then changes the moist static stability of the atmosphere. Areas where dry air is present over moist air will likely have enhanced convection, and hence maximum rainfall occurs at and downstream of those locations.

**b. Discussion**

Although the results from this study are from idealized situations, they are largely consistent with those from previous observational and modeling studies, and the convective asymmetries can be explained physically. Of course, these situations are far from realistic. Crude assumptions include the use of an $f$ plane and the coast-
line is moved toward the TC. The former (latter) ignores the beta (steering) effect, both of which can contribute toward wind asymmetries, and even convective asymmetries, in the TC. How these asymmetries will modify the present results needs to be determined through additional modeling work. Further, an implication from this result is that the distribution of rainfall associated with TC landfall is largely independent of the direction of approach of the TC. Although Ching (2002) did show that for four typhoons that hit Hong Kong in 1999 from different directions, the distributions of convection were quite similar, the extent to which this statement is generally true needs to be ascertained through both modeling work with more realistic situations and further observational studies. Finally, the effect of topography is ignored in the present study. The effects of mountains, water bodies, and even urban environments on the modifications of the convective distributions also need to be investigated.

Acknowledgments. The authors would like to thank Mr. Simon Ching for helpful discussions. This research is sponsored by the City University of Hong Kong Strategic Research Grant 7001042 and the National Natural Science Foundation of China Grant 49975014.

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