Structure and Evolution of Rainfall in Numerically Simulated Landfalling Hurricanes

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ABSTRACT

Using an idealized landfalling model hurricane, the impact of different land surface characteristics on hurricane rainfall distribution before, during, and after landfall is investigated. Before landfall, maximum rainfall occurs on the right side of the storm track as a result of dry air intrusion from both the environmental flow behind the vortex and the land surface ahead of the vortex. These sources of dry air combine to destabilize the right side and stabilize the left side of the storm. Upon landfall, the rainfall maximum shifts to the left of the storm track over land, near the coast. Increased friction over land drives a region of convergence in the entire front half of the storm. While mean rainfall rates decrease, localized areas of large rainfall accumulations may occur as a result of this frictional forcing. Over land, the rainfall area broadens and mean rainfall rates decrease. No differences are detected in inner-core rainfall rates between cases, but outer-core rainfall rates and rainfall coverage increase with moister land surfaces. Hence, significant differences in rainfall accumulations occur depending on moisture availability of the land surface. Since reconnaissance planes cannot fly over land, forecasters are often forced to make extrapolations from reconnaissance data over water. They should take extreme caution in doing so, since rainfall distributions may change suddenly upon landfall as different forcing mechanisms take over.

1. Introduction

As a hurricane makes landfall, it moves from shallow coastal waters to solid land, leading to a rapid decrease in surface latent heat fluxes, the hurricane’s primary energy source. In addition, surface roughness increases substantially. These rapidly changing conditions affect the pattern, extent, and intensity of damaging winds and rainfall. The underlying physical processes are complex, occur over a very short time (less than a day), and depend on many coexisting factors including topography, coastline shape, storm angle of incidence, storm translation speed, and land surface properties.

Idealized numerical modeling studies of landfalling hurricanes (Tuleya and Kurihara 1978; Tuleya et al. 1984; Jones 1987; Tuleya 1994; Chan and Liang 2003; Shen et al. 2002) consider a hurricane making landfall on a flat land surface with a straight coastline. They confirm that reduction in surface evaporation near the storm core is the primary mechanism for hurricane decay over land. Tuleya (1994) explains how this might occur. As the cloud canopy moves in to block solar radiation, limited heat capacity and conductivity of the soil subsurface lowers the land surface temperature (LST). This causes a low-level cool pool to form beneath the storm, thus reducing latent heat fluxes over the land surface and depriving the storm of its fuel. Increased surface roughness and reduced relative wetness [or moisture availability (MA)] of the soil enhance decay but are not critical to it (Tuleya et al. 1984; Tuleya 1994).

The above modeling studies showed a decrease in tangential wind and an increase in radial wind as storms moved over land. A band of convergence was seen on the onshore flow side and a band of divergence on the offshore flow side. Tuleya and Kurihara (1978) and Tuleya et al. (1984) observed a rainfall maximum in the right-front quadrant of the landfalling storms and attributed this to the enhanced convergence in that area. Jones (1987) and Chan and Liang (2003) observed a rainfall maximum in the right-rear and left-front quadrants, respectively, and concluded that enhanced convergence on the onshore flow side was not a dominant factor in driving rainfall in landfalling storms. Jones
(1987) speculated the location of the rainfall maximum was caused by the interaction of the vortex and steering flow. Chan and Liang’s (2003) study did not include a steering flow, but they observed dry air from land being entrained into the vortex. The resulting instability explained the location of the precipitation maximum. Both Jones (1987) and Chan and Liang (2003) employed a fixed LST and an f plane in their models.

There seems to be considerable disagreement in both observational and modeling studies about the location of the rainfall maximum before and at landfall. Similarly, there is uncertainty about the forcing factors that drive rainfall distribution at landfall.

Chen and Yau (2003) simulated a landfalling hurricane using a straight coastline and flat land surface with marshland properties. Their model included an evolving LST and explicitly resolved cumulus convection with variable Coriolis parameter. The model fields of a previous simulation of Hurricane Andrew (Liu et al. 1997) were used as initial conditions, so a steering flow with vertical shear was included. The rainfall before landfall was concentrated on the right side of the vortex track; this asymmetry was attributed to vertical wind shear. Once the vortex made landfall, rainfall became equally distributed around the vortex track.

Powell (1982) observed a convergence maximum on the onshore flow (right) side and divergence on the offshore side in Hurricane Frederic (1979). Parrish et al. (1982) noted the formation of two intense convective features in Frederic’s eyewall, near the coast, to the right of the track. These features increased in intensity as they moved cyclonically around the northern eyewall and produced a rainfall maximum on the left of the track. In Hurricane Alicia (1983) low-level convergence was observed on both sides (Powell 1987). This was attributed to an environmental flow that was directed along the coast and perpendicular to the storm track.

As Hurricane Andrew (1992) crossed the Florida coastline, intense deep convection developed in the right-front (RF) quadrant of the eyewall and then moved into the left-front (LF) quadrant (Wakimoto and Black 1994). Chan et al. (2004) observed an increase in convection on the western side of two typhoons making landfall on the south China coast.

The current study uses state-of-the-art model physics, but a simplified landfall configuration, to investigate the structure and forcing factors of hurricane rainfall distribution before, during, and after landfall. The goal is to simulate important physical processes as realistically as possible (within the limitations of the current status of model development), while simplifying the analyses by considering only a limited number of factors that potentially affect rainfall distribution. The contributions from additional factors like topography and coastline shape will be introduced in follow-up work.

Limitations of some previous modeling work have included a fixed LST, an f plane, exclusion of a steering flow, and the use of a cumulus parameterization. A fully evolving LST is important to realistically simulate the landfall process since it drives latent heat fluxes over land that were shown to be critical to storm decay (Tuleya 1994). Inclusion of a steering flow will allow the impacts of vortex interaction with the environmental flow to be included in the landfall process. The current study employs a uniform (in order to eliminate asymmetries forced by vertical wind shear) and relatively weak steering flow, explicitly resolved convection, variable Coriolis parameter, and finer horizontal and vertical resolutions than were used in previous model landfall studies. Possible forcing factors behind the distribution of rainfall before, during, and after landfall will be discussed. Additionally, the effects of different and realistic land-use categories on rainfall distribution will be investigated.

2. Model configuration and experiment design

The fifth-generation Pennsylvania State University–National Center for Atmospheric Research (NCAR) Mesoscale Model (MM5) is initialized with 8 m s\(^{-1}\) southerly flow. Embedded in this flow is a hurricane vortex with initial minimum surface pressure (PSMIN) of 970.6 hPa and 42 km radius of maximum winds (RMW). The intensity and size properties of this vortex are based on the averaged properties of hurricanes making landfall in the north-central Gulf of Mexico during 1988–2002. Construction of such a vortex follows the technique outlined in Kimball and Evans (2002). The sea surface temperatures (SSTs) in the model are kept constant at 28°C, and a straight, west–east-oriented coastline is located at 30°N, 387 km north of the initial position of the vortex. The vortex is asymmetric since it is embedded in relatively strong southerly flow; therefore, it is larger on the eastern side than it is on the western side. If we define size as the radius of 17 m s\(^{-1}\) winds, then in the LF (RF) quadrant the average size of the vortex is 115 (280) km. This means the core of the vortex does not interact with the coastline at the beginning of the simulation. The land surface is flat, has a height of 0.1 m above sea level, has an initial surface temperature of 28°C, and is covered by a single land-use category. A control simulation (noland) consisting of a water surface only is compared with simulations with different land-use categories. The experiment names, descriptions of the land-use categories and their physical properties are listed in Table 1.
A 42-h simulation is conducted using a coarse mesh of 9-km horizontal resolution, a nested grid of 3-km horizontal resolution, and 38 vertical levels. The coarse mesh is a 1350 by 1350 km square grid, while the nested grid measures 603 km in both dimensions. The center of the nest is located 150 km south of the landfall location of the vortex (which shrinks in size as it weakens and approaches the coastline). The nested domain is switched on 9 h after the start of the simulation. Convection is parameterized on the coarse mesh using the Kain–Fritsch scheme (Kain and Fritsch 1992) and is explicit on the fine mesh. Other parameterizations on both meshes include the Goddard Microphysics (including graupel) scheme (Lin et al. 1983; Tao et al. 1989, 1993), a long- and shortwave cloud radiation scheme (Dudhia 1989), and the Medium-Range Forecast model (MRF) boundary layer scheme (Hong and Pan 1996), which includes a 5-layer soil model. The surface flux parameterization makes use of a form of the bulk aerodynamic formulations (Blackadar 1979).

Henceforth, only nested domain results will be discussed and times will be given relative to the start of the nested domain ($t = 0$ h).

3. Results

a. Storm evolution

1) Storm track

When the nested domain is switched on, the storm center is located 195 km south of the coastline. Figure 1 shows storm tracks for all cases. No significant difference is seen in translation speed and tracks of the landfalling storms compared to the control (Figs. 1 and 2), in agreement with previous idealized modeling studies of landfalling hurricanes (Tuleya et al. 1984; Jones 1987; Tuleya 1994). A small track deviation occurs toward the simulation end, but it is never larger than about 20 km. All model fields are left to evolve freely over time. Because radiational and convective processes alter the heating and pressure distribution in the model domain, the background flow evolves. In this case, it changed from a southerly to southwesterly direction in the first 8–10 h of the simulation (i.e., before the nest is switched on). The southwesterly flow remains in place in the later hours of the simulation, but decreases in magnitude, explaining the direction (Fig. 1) and evolution in the storms’ speed (Fig. 2).
storms deviate from their steering flow by a very small poleward amount, which can be explained by the $\beta$ effect. Tuleya (1994) also found a deviation of the storm track from the initially imposed background flow. Fifteen hours after nested domain initiation, the storm centers cross the coastline about 222 km east of their initial location in the nested domain.

2) Storm intensity

Figure 3 shows the evolution of storm intensity measured in terms of PSMIN). The noland simulation weakens slowly until about $t = 18$ h when slow intensification begins. This storm remains a category 2 hurricane throughout the simulation. As expected, the

Fig. 2. Time series of translation speed (m s$^{-1}$) of all cases. A 3-point smoother has been applied to reduce noise. The vertical line indicates the time of landfall.

Fig. 3. Time series of PSMIN (hPa) for all cases. The vertical line indicates the time of landfall.
other storms continue to weaken after their centers cross the coastline at $t = 15$ h. Taking the case with the lowest roughness length (RL) and highest MA (irrigated; 15 cm and 50%, respectively) as a reference point, it becomes clear that increasing RL (broadleaf; 50 cm; 50%) has a bigger impact on enhancing storm decay than decreasing MA (savanna; 15 cm; 15%). This was also found by Tuleya (1994), but with MA values of 100% and 30%, and RL values of 1 cm and 25 cm. Furthermore, all large RL cases (50 cm: urban, broadleaf, needle), including the case with the highest MA (broadleaf) and potentially the largest surface evaporation, weaken more than low RL cases. Cases with identical MA but larger RL always evolve into a weaker storm. A possible reason for the extra weakening of a high RL case may be that increased RL reduces the surface wind speed. This reduces surface latent and sensible heat fluxes further than in a low RL case with identical MA. These same conclusions hold when viewing a time series of the maximum low-level wind speed (not shown). However, while the pressure decrease becomes only slightly larger once the storm center crosses the coastline, the low-level wind speed drops substantially in a period of about 2 h as the center crosses the coastline. For cases with large roughness lengths this drop is significantly larger. A more detailed analysis of the evolution of the storm wind speed will be presented in a future publication.

Cases with identical RL and lower MA do not always become weaker. This is illustrated by needle (RL = 50 cm; MA = 30%) and broadleaf (RL = 50 cm; MA = 50%); shortly after landfall broadleaf is stronger, but toward the end of the simulation it is weaker despite more available surface moisture. The same applies to dryland and irrigated cases. However, differences are small. The largest difference in intensity between any two cases (urban and dryland) at the simulation end is 6 hPa, about half the pressure range of Saffir–Simpson category 2.

b. Rainfall accumulation

Figures 4a–c show accumulated rainfall for the simulation period for 3 cases. Noland (Fig. 4c) displays a mostly symmetric pattern of accumulated rainfall around the storm center. As the storm intensifies around $t = 18$ h, rainfall amount and areal coverage increases. The landfalling storms display pronounced asymmetries with rainfall maxima to the right of the track before landfall in all cases, as can be seen for urban and broadleaf in Figs. 4a and 4b, respectively. As the storms approach the coastline, rainfall totals drop off. After landfall, rainfall accumulations remain less than amounts recorded to the right of the track while the storms are over water, but rainfall becomes more symmetrically distributed around the storm center. However, local areas of larger rainfall accumulations (≥140 mm) are seen just onshore to the left of the storm track and farther inland on either side of the storm track.

Figure 5a shows a time series of hourly rainfall accumulations totaled over the entire nested domain (603 × 603 km) for all simulations. In all cases, hourly rainfall accumulations drop off steeply during model spinup ($t = 0–6$ h). In the landfalling cases, a gradual decline follows as land is approached. Rainfall totals increase once storm centers have crossed the coastline (after $t = 15$ h) until $t = 27–28$ h, coincident with a minimum in storm propagation speed (Fig. 2) which enhances rainfall accumulation. Rainfall decreases during the last 5 or 6 h of the simulation as the storms dissipate. In noeland, total rainfall decreases during the first 10 h of the simulation, followed by a steady increase as the storm intensifies and reduces its propagation speed (Fig. 2). Differences in total rainfall accumulations between the various cases are subtle over water, but clearly visible after landfall (see also Fig. 4). Rainfall totals rank according to MA: storms with 50% MA produce the most rainfall, while those with 10%–15% produce the least. Cases with identical MA but larger RL produce slightly more rainfall most of the time.

No large differences in translation speed and track shape [section 3a(1)] were seen between cases, therefore, these are not the cause of the differences in rainfall accumulations seen in Figs. 4 and 5a. Figure 5b shows a time series of hourly rainfall rates averaged over a circle with a 55-km radius from the storm center, where most of the rainfall occurs. After the model adjustment period, rainfall rates hold steady at around 11 mm h$^{-1}$, followed by a slight increase just before the storm centers cross the coast. After landfall, rainfall rates drop off slowly. No big differences are seen between cases, but these values agree well with observed rainfall rates in hurricanes (Marks 1985; Burpee and Black 1989; Lonfat et al. 2004; Rodgers et al. 1994; Miller 1958) as summarized in Table 2. The decrease in hourly rainfall accumulations from $t = 6$ to 16 h (Fig. 5a) coincides with a slight increase in mean rainfall rates in the inner core of the storm (Fig. 5b). Figure 5c shows mean rainfall rates in the outer core (radius of 55–110 km), which decrease from the end of the spinup period until about $t = 19$ h. This explains the drop in hourly rainfall accumulations, but does not explain differences in hourly accumulations between landfalling cases (Fig. 5a).

Figure 5d is a time series of areal coverage of rainfall rates greater than 0 mm h$^{-1}$. Between $t = 6$ and 9 h,
rainfall coverage decreases. As the storms approach the coastline, coverage steadily increases, until it drops off at around $t = 29$ h. A big difference exists between the various cases in areal coverage. Coverage increases with MA and explains the differences in hourly rainfall accumulations in Fig. 5a. Between $t = 23$ h and the simulation end, these differences are enhanced because of differences in outer-core rainfall rates, which are also correlated with MA (Fig. 5c). These rainfall rate differences are caused by a rainband east and southeast of the storm center (e.g., Fig. 4b) that is more pronounced in moister cases. Rainbands are discussed further in section 3c(3).

Figure 6 summarizes contrasts in rainfall coverage and rates seen between two different MA cases with the same RL. Hourly rainfall totals are shown at 6-h intervals for savanna and irrigated. Before landfall ($t = 9$ h), rainfall coverage does not show a large difference (see also Fig. 5c). Rainfall rates are larger in the drier case and are concentrated to the right of the storm track (Fig. 1) in both cases. At landfall ($t = 15$ h), both cases display a rainfall maximum to the left of the storm track.
FIG. 5. Time series for all cases of (a) hourly rainfall accumulations (mm) totaled over the entire domain, (b) rainfall rates (mm h$^{-1}$) averaged azimuthally and over a radius of 0 to 55 km, (c) rainfall rates (mm h$^{-1}$) averaged azimuthally and over a radius of 55–110 km, and (d) areal coverage ($\text{km}^2$) of rainfall rates greater than 0 mm h$^{-1}$. The vertical line indicates the time of landfall.

<table>
<thead>
<tr>
<th>Study</th>
<th>Storm(s)</th>
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Fig. 6. Hourly rainfall accumulations (mm) and 10-m wind arrows for (left) savanna and (right) irrigated at $t = 9, 15, 24, \text{ and } 27 \text{ h}$. The coastline is indicated by the black horizontal line. Land is to the north of this line, water to the south. Labels on $x$ and $y$ axes indicate distance in kilometers.
track, just onshore. Six hours later, the eyewall has broadened in both cases and a rainband is located to the storm’s southeast, both of which account for the increases in areal coverage and outer-core rainfall rates shown in Figs. 5c,d. Rainfall rates are now higher in the moister case, probably because air parcels that have traveled over a wetter landmass remain moister and not because there is more moisture available directly underneath. This is reinforced by the much lower low-level specific humidity values surrounding the storm core in the dry cases (not shown). At $t = 27$ h, the eyewall has broadened further, especially in the moister case, which displays both larger areal coverage and larger rainfall rates (the latter associated primarily with the rainband). Similar differences are seen for the other pair of storms with identical RL and different MA (urban and broadleaf).

c. Rainfall distribution

1) Prelandfall

Figures 4 and 6 indicated that a left–right (L–R) asymmetry exists in the rainfall distribution of the landfalling storms. Figure 7 displays hourly rainfall rates averaged over each storm half (left and right of the track), as a function of time and distance from the storm center, for irrigated and noland. The asymmetry is strongest in the eyewall during the period $t = 5–10$ h. The evolution of the eyewall rainfall field during this period is examined further in this section using time-
averaged radius–azimuth plots. Since the eyewall is not usually a perfectly round circle and changes in shape (from circular to oval and anything in between) and size (i.e., its diameter contracts and expands) over time, using a circular ring with fixed inner and outer radius would miss (portions of) the eyewall at some azimuths. To capture the entire eyewall at all azimuths, an annulus is used where the inner and outer radii of the annulus vary with azimuth, but the distance between the two radii (i.e., the width of the annulus) remains constant (48 km). In other words, the annulus maintains its width but adjusts its shape (becomes more or less oval or contracts and expands) in order always to envelope the evolving eyewall and/or to follow the outward-tilting eyewall with height. The annulus method is also ideally suited to constructing time composites, which allows smoothing of short-term fluctuations in the fields. Such a time average is presented in Fig. 8 for rainfall rates and surface fluxes in the eyewall over the $t = 5$–10-h period, before the storm center crosses the coastline. Radius is shown along the abscissa and azimuth (in 30° increments) along the ordinate. The azi-
muths are oriented such that 0° is always parallel to the direction of motion of the storm; 0°–90° is the (LF) quadrant; 90°–180° the left-rear (LR), and so forth. Figures 8a,c,e compare eyewall rainfall rates for noland, broadleaf, and urban (the latter two are both high RL cases). There is weak left–right asymmetry in noland with rainfall concentrated in the RF and LF quadrants. When land is present, the asymmetry increases with most rain occurring in the right-rear (RR) quadrant. Observations of rainfall in hurricanes over water (e.g., Marks 1985; Burpee and Black 1989; Lonfat et al. 2004; Rodgers et al. 1994) show a preference for maximum rainfall to occur in the LF and/or RF quadrants in response to low-level convergence as a result of storm motion (Shapiro 1983). This is consistent with case noland. However, Marks (1985) observed a shift of the rainfall maximum to the RR quadrant in Hurricane Allen (1980) as the storm approached the Texas coast, just like the current landfalling storms. Therefore, the low-level convergence forcing mechanism may be overruled by other forcing mechanisms when a hurricane approaches land.

In dry case urban (Fig. 8c) more rain falls on the right side of the storm and less on the left than in moister case broadleaf (Fig. 8c), so drier cases display a more pronounced L–R asymmetry in the rainfall field (this also applies to the drier cases with low RL not shown here). This difference in asymmetry could be explained...
by a difference in L–R asymmetry in surface fluxes and energy supply to the storm. Figures 8b,d,f show the latent plus sensible heat fluxes (shaded) for the same three cases. Again, the land cases display a more pronounced L–R asymmetry in surface fluxes than noland, but this time the degree of the asymmetry does not increase with decreasing MA. In dry case urban the fluxes are higher at all azimuths than they are in moister case broadleaf. A similar trend is seen for moist and dry cases in the low RL group (not shown). This could be caused by stronger low-level winds and/or larger vertical moisture and temperature gradients in drier cases.

Figures 9a–d show a time-averaged radius–azimuth plot (like Fig. 8) of the 10-m wind speed (shaded) for noland, broadleaf, irrigated, and urban. The wind field shows an L–R asymmetry in all cases, which is partly caused by storm propagation speed since the forward motion vector is aligned with storm cyclonic winds to the right of the track and opposes storm cyclonic flow on the left side of the track. The wind speed fields of the landfalling cases differ significantly from noland, which is interesting because all cases move at the same propagation speed. Therefore, something else besides propagation speed contributes to the wind field asymmetry. It is also seen that the wind fields of the landfalling cases do not differ significantly from one another in spite of different MA and RL values. Since the rainfall fields do
differ among landfalling cases, it is concluded that the wind field is not the driving force behind the rainfall asymmetry. Instead, it could be the reverse: the asymmetric rainfall field forces the asymmetric wind fields. The latter is supported by Figs. 10a and 10b, which show the 40-m tangential and radial winds, respectively, as a function of radius and azimuth for broadleaf. Shown in contours are speed divergence of the tangential wind \((\times 1000, \text{Fig. } 10a\) and 10, 20, 30, and 40 mm h\(^{-1}\) rainfall rates (Fig. 10b). All fields [rainfall, upward motion (not shown), radial winds, tangential winds] are collocated. In other words, a rainfall maximum accompanied by strong updrafts, forces a low-level inflow (i.e., radial wind) maximum. This causes inward advection of angular momentum and, via angular momentum conservation, stronger tangential winds at the location of the rainfall. The speed divergence field of the tangential wind in Fig. 10a shows maximum convergence, as expected, located just downstream from the tangential wind maximum. If anything, the rainfall maximum occurs in an area of low-level speed divergence of the tangential wind in all cases. The con-

**Fig. 9.** Wind speed (m s\(^{-1}\); shaded) at 10 m averaged over \(t = 5-10\) h as a function of radius (km) and azimuth (degrees) for (a) noland, (b) urban, (c) broadleaf, and (d) irrigated. Arrows are horizontal wind vectors at 10 m with the radial component exaggerated by a factor of 4.
vergence in the radial wind is almost certainly a result and not a cause of the convection.

Another possible forcing mechanism is local vertical shear. A baroclinic vortex in an environment with variable Coriolis parameter, develops $\beta$ gyres whose magnitude changes with height because the magnitude of the vortex’s circulation changes with height (e.g., Bender 1997). This leads to local vertical wind shear across the center of the vortex. In Bender’s (1997) idealized simulations this shear forced a rainfall maximum in the southeast quadrant of the vortex, regardless of the direction of the environmental flow field. The southeast quadrant in the current simulations corresponds to the RR quadrant. The local shear magnitude and direction were calculated (not shown) and correspond to those found by Bender (1997). However, if the local shear played a dominant role, then each of the seven cases should display a rainfall maximum in that quadrant. In the landfalling cases, (e.g., Figs. 8c,e) a rainfall maximum is indeed observed in the RR quadrant. However, Fig. 8a shows that the no land case does not display a rainfall maximum in that area. Evidently, whatever dominates the forcing of the rainfall asymmetry happens to cause the rainfall maximum in the landfalling cases to be located in the same quadrant as local vertical shear forcing would have been.
Since a change in the surface flux and rainfall asymmetries was observed as MA changes, dry air intrusion is considered as a dominant forcing factor. Drier cases display larger surface fluxes on both sides of the storm despite wind speeds on each side being almost identical to the moist cases. Therefore, differences in low-level moisture and/or temperature fields must play a role in driving surface flux differences in the land cases. As the storm cloud canopy’s outer edges cross land, the LST and, hence, surface latent and sensible heat fluxes over land reduce (not shown). Dry and cool land air may flow toward the left half of the vortex while it is still over water and entrain into the eyewall. This dry and cool air intrusion may increase the vertical moisture and temperature gradients at the surface, thereby enhancing surface fluxes.

Detailed examination of the three-dimensional equivalent potential temperature ($\theta_e$) and wind fields in the eyewall annulus reveals all cases experience dry air intrusion from the drier (by definition, since a hurri-
cane core is moist) environment. Figure 11 shows the $\theta_e$ (upper panel) and hydrometeors (lower panel) averaged over $t = 5–10$ h across an inner slice (2–9 km) of the eyewall annulus for cases noland, broadleaf, and urban. In noland (Fig. 11a), a patch of low-$\theta_e$ air can clearly be seen in the right-front quadrant between about 850 and 650 hPa and in the left rear between the surface and about 800 hPa. In broadleaf (Fig. 11c), the right side dry air pocket occupies a larger azimuthal and vertical space, while the left pocket covers a wider azimuthal range but is less dry. In urban (Fig. 11e), both areas are larger and drier compared to broadleaf. The low-$\theta_e$ air pockets in noland force two slantwise statically unstable regions in the LF and RR quadrants, with high-$\theta_e$ air present at the surface (Fig. 11a) and cumulus towers forming downstream (Fig. 11b). The RR area is more unstable because of the lower values of $\theta_e$ upward and downstream. Hence, more and stronger

Fig. 11. (top) Equivalent potential temperature (K; shaded), radial wind speed ($\text{m s}^{-1}$; white contours), and reflectivity (dBZ; black contours) and (bottom) snow plus graupel plus ice mixing ratios (g kg$^{-1}$; shaded) and rainwater mixing ratio (g kg$^{-1}$; colored contours) averaged over $t = 5–10$ h and over a 24-km annulus in the eyewall as a function of height (hPa) and azimuth (°) for (a), (b) noland; (c), (d) broadleaf; and (e), (f) urban. Arrows are tangential–vertical wind vectors with the vertical component enlarged by a factor of 90.
cumulus towers form on that side. The cloud and rain shafts lean upstream with height because the tangential wind speed in a hurricane decreases with height, advecting the hydrometeors at lower levels at a faster pace. Some of the raindrops are advected in to the left half of the storm. In broadleaf, surface air on the left side of the storm has lower $\theta_e$ than noland while overlying air has higher $\theta_e$, making the air more stable on that side. Hence, air rises on the right side of the storm only. The larger azimuthal extent of the dry air on the left causes evaporation of raindrops that are advected in to that half. These features are further accentuated in even drier case urban (Figs. 11e,f).

From Fig. 11 it can be concluded that dry air enters the storms from two sides. After detailed examination of $\theta_e$ maps and cross sections (not shown), the dry air's path was identified and is depicted in Fig. 12. In each case, including noland, the dry air enters the vortex in the southeast because the southerly environmental flow is slightly stronger than the forward motion of the vortex. The dry air gets entrained into the outer edges of the vortex circulation and wraps cyclonically from the southeast around the outer edge of the front half of the eyewall annulus (Fig. 12a). The upper portion (above about 850 hPa) of the encroaching dry air experiences outflow in the front half of the vortex [Fig. 11a; possibly because the environmental flow or a vortex relative flow at that level is faster than the propagation speed of
the vortex and, hence, pushes its way through the vortex’s circulation as was also found in Bender (1997)] and, therefore, the dry air continues to encircle the vortex until it reaches the rear. Here, the vortex experiences inflow up to about 600 hPa at the outer edge of eyewall annulus (Fig. 11a). The dry air enters the outer edge of the eyewall annulus in the RR quadrant and moves cyclonically and inward to reach the inner edge of the eyewall in the RF quadrant (bold black arrow in Fig. 12). The lower portion (below 850 hPa) of the dry environmental flow entering the vortex in the southeast, experiences inflow when it reaches the front of the vortex and, hence, enters the eyewall annulus in the LF eyewall quadrant, and establishes itself as the dry tongue seen in Fig. 12a (dark gray arrow) extending into the left half of the vortex. The lower portion of the RR eyewall quadrant remains relatively unaffected by dry air intrusion. With the pocket of low $\theta_e$ aloft in the RF quadrant a slantwise unstable situation exists (Fig. 11a). As a result, large amounts of cloud and rainwater (Fig. 11b) form downstream of the RR quadrant. In the LF quadrant a similar slantwise instability is seen but it is capped by high-eyewall-$\theta_e$ air and less convective activity occurs than in the RF quadrant. The convection in the left half of the noland storm causes low-level inflow on that side, explaining why more dry
air is seen there (Fig. 11a) than in the land cases (Figs. 11c,e).

In the land cases, the situation is more complex (Fig. 12b). In addition to the environmental source, another source of dry air enters the vortex from the landmass to the north (light gray arrow). Reduced LST and surface fluxes lower moisture and temperature over land and reduce $\theta_e$ of the inflowing air. Additional reasons for cooler air over land are 1) enhanced friction forcing air to flow toward the lower pressure center of the storm and expanding and cooling as it does so and 2) evaporation of falling rain (Li et al. 1997). One dry air channel flows off land to the left of the vortex and entrains into the outer edges of the vortex in the LR quadrant (Fig. 12b). Here the air rises and merges with the environmental dry air channel (black arrow). This reinforces the higher-level dry air—and associated instability—on the right side of the vortex (Figs. 11c,e). Again slantwise instability allows the formation of convection but this time the maximum lies between the RR and RF quadrants. Additionally, dry air from the land surface enters the outer edges of the front of the eyewall (Fig. 12b); any raindrops advected into this area by the tangential winds evaporate in this dry air. This dry air also explains the larger surface fluxes on the left of the storm for drier cases. As the land surface decreases in MA, more and lower-$\theta_e$ air is entrained into the vortex. This can be seen in Fig. 11e for urban.

**2) DURING LANDFALL**

The prelandfall right side rainfall maximum flips over to the left-hand side as the storm center crosses the coastline between $t = 13$ and 16 h (Figs. 4, 6, and 7). The larger maxima occur in the large RL cases; large MA is a secondary driving force (Table 3) with one exception: irrigated’s high MA compensates for its low RL, causing it to rank above higher RL case urban. The fact that rainfall increases with RL suggests that low-level convergence—resulting from differential surface friction between land and water—may be a forcing fac-

---

**TABLE 3.** Landfalling cases ranked according to the accumulated rainfall maximum observed just onshore to the left of the storm track.

<table>
<thead>
<tr>
<th>Case</th>
<th>Accumulated rainfall (mm)</th>
<th>RL (cm)</th>
<th>MA (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Broadleaf</td>
<td>200</td>
<td>50</td>
<td>50</td>
</tr>
<tr>
<td>Needle</td>
<td>180</td>
<td>50</td>
<td>30</td>
</tr>
<tr>
<td>Irrigated</td>
<td>160</td>
<td>15</td>
<td>50</td>
</tr>
<tr>
<td>Urban</td>
<td>150</td>
<td>50</td>
<td>10</td>
</tr>
<tr>
<td>Dryland</td>
<td>140</td>
<td>15</td>
<td>30</td>
</tr>
<tr>
<td>Savanna</td>
<td>140</td>
<td>15</td>
<td>15</td>
</tr>
</tbody>
</table>

---

**FIG. 12.** Schematic of dry air intrusion for the (a) no land case and (b) land cases. The thin black circles represent the storm eyewall. The thin solid black arrow indicates the direction of motion of the storm; motion relative quadrants are shown using abbreviations LF, RF, LR, and RR. In (a) the thin curved arrows show where storm-relative in- and outflow occurs; the solid arrows represent inflow below 650 hPa, the dashed black arrows outflow above 850 hPa, and the dashed gray arrows inflow below 850 hPa. For clarity these arrows have been left off in (b). The thick black arrows show dry air intrusion above 850 hPa, the dark gray arrows show dry air intrusion below 850 hPa but above the surface, and the light gray arrows indicate dry air originating at the surface over land.
tor. In other words, RL is needed to drive convergence and MA is needed to fuel convection. According to previous modeling studies (e.g., Tuleya and Kurihara 1978; Tuleya et al. 1984), however, a rainfall maximum should occur on the right side where land–sea roughness contrasts force low-level convergence. This apparent discrepancy is further investigated below.

Figure 13 shows maps of the 40-m tangential and radial winds for \( t = 9 \) and 15 h for savanna. Hourly rainfall totals for the same time periods can be found in Figs. 6a,c. At \( t = 9 \) h the rainfall is concentrated in the right half of the storm (which is tracking toward the northeast) relative to the storm motion (Fig. 6a). Strong inflow can be seen in that same storm half but slightly upwind from the rainfall maximum (Fig. 13a). The tangential wind also displays a maximum in the right half of the storm (Fig. 13c), partially as a result of the north–eastward motion of the storm, partly as a result of the convective maximum on that side [see section 3c(1)]. Six hours later (Fig. 6c), the rainfall has reduced and is concentrated in the left half of the storm (relative to the track) over land. However, the radial wind is now better collocated with the rainfall maximum and is stronger than 6 h earlier despite the reduced rainfall (Fig. 13b). This means the increased radial wind over land is mostly frictionally driven instead of driven by the convection in the eyewall. Because of surface friction, tangential winds are weaker over land (Fig. 13d). Both are in agreement with previous landfall modeling studies (Tuleya and Kurihara 1978; Tuleya et al. 1984; Jones
These wind configurations cause low-level divergence patterns that may drive the observed rainfall maximum. The radial wind pattern certainly complies, with low-level convergence occurring on the left side of the storm. However, the tangential winds cause an area of low-level divergence in that same location. The horizontal divergence in cylindrical coordinates [Eq. (1)] will aid in further investigating the relative contribution of each wind component to the total divergence:

\[
\nabla \cdot \mathbf{u} = \frac{1}{r} \frac{\partial (ru_r)}{\partial r} + \frac{1}{r} \frac{\partial v_t}{\partial \phi} + \frac{v_r}{r} \frac{\partial u_t}{\partial r} + \frac{1}{r} \frac{\partial u_t}{\partial \phi} = A + B + C .
\]

Here \(v_r\) and \(v_t\) are the radial and tangential winds, respectively, \(r\) is the radial direction, and \(\phi\) is the azimuth. Term A is the directional divergence of the radial wind, term B is the speed divergence of the radial wind, and term C is the speed divergence of the tangential wind. Each of these terms has been plotted in Fig. 14, as well as their sum, at 40 m and \(t = 9\) and 15 h for savanna.

At \(t = 9\) h, term A results from a combination of frictional convergence and eyewall convection (Fig. 14a). The storm is over water at the time and no external forcing other than surface friction exists to force winds to converge toward the hurricane core. The winds turn inward near the convective maximum, and therefore term A is a little stronger just downwind from the rainfall maximum. At \(t = 15\) h (Fig. 14b), a larger maximum in term A is seen over land. This is a result of increased frictional turning of the winds toward low surface pressure, which in turn increases the radial component of the wind. This enlarges term A in that area.

Term B, the radial variation of \(v_r\), displays alternating bands of convergence and divergence associated with in- and outflow of the eyewall and rainbands. This term is largest near the eyewall where radial winds approach rapidly and then turn upward to ascend in the eyewall. This sudden decrease in the radial wind component forces a speed convergence maximum (Figs. 14c,d). The magnitude of this term is similar over water and land, and in both situations its maximum is collocated with the rainfall maximum. This suggests term B is driven by eyewall convection and not surface characteristics.

The speed divergence pattern of the tangential wind (term C; Fig. 14e) in the eyewall at \(t = 9\) h is consistent with the tangential wind dipole shown in Fig. 13d. Along the coastline, a region of divergence exists on the offshore flow side of the storm center and convergence on the onshore flow side. These fields are enhanced when the storm crosses the coastline (Fig. 14f) and are forced by the difference in surface roughness between land and water.

The aggregate effect of the three divergence terms is shown in Figs. 14g,h. At \(t = 9\) h, an almost uniform annulus of convergence is seen in the eyewall, while 6 h later convergence is concentrated in the northern half of the storm, over land. Even without term B (convergence driven by the convection and not vice versa), this remains true. On the west side, the convergence is a result of surface-driven inward turning of the wind (term A); the directional convergence of the radial wind is larger than the speed divergence of the tangential wind, even in a low surface roughness case like savanna. East of the storm center, differential surface roughness forces speed convergence of the tangential wind (term C).

Figure 15 shows an azimuthal–vertical cross section of rainwater (contoured) and graupel, ice, and snow (shaded) averaged over a 36-km annulus in the eyewall at \(t = 15\) h for savanna and irrigated. In the dry case (savanna) one cluster of cumulus towers is observed in the LF quadrant, while in the moist case (irrigated) two clusters of towers are seen, one in the LF and one between the LR and RR quadrants. In savanna, the eastern low-level convergence maximum forces the formation of cumulus towers which are advected slightly downstream (i.e., to the LF quadrant of the eyewall) by the tangential winds. The rainfall maximum occurs even farther downstream as stronger lower-level tangential winds advect the raindrops downstream from the cumulus tower (Fig. 15a). In moist cases, the rainfall maximum extends all the way to the southern part of the eyewall (Fig. 5d). This is because convergence on the vortex’s west (left) side has enough surface moisture available to force a second cluster of cumulus towers (Fig. 15b). This cluster of towers is subsequently advected downstream from the surface convergence maximum to the west-southwest and raindrops are advected even farther downstream to the south (or RR quadrant) of the storm center. Two groups of towers are also observed in cases broadleaf and needle. This explains why more rain falls to the left of the vortex track at landfall in cases with 1) high RL combined with medium-to-high MA, or 2) low RL and high MA (Table 3). On the west side of the vortex, the RL drives the low-level convergence and in moist cases sufficient MA provides fuel for convection. On the east side of the vortex, moisture is supplied, in all cases, by moist onshore flow. This agrees with observations. As Hurricane Andrew (1992) crossed the Florida coastline, in-
tense deep convection developed in the RF eyewall quadrant and moved to the LF quadrant (Wakimoto and Black 1994). Parrish et al. (1982) observed two intense convective features that formed in the eyewall, near the coast, and to the right of Hurricane Frederic’s (1979) track during landfall on the Alabama coast. These features increased in intensity as they moved cyclonically around the northern eyewall and produced a rainfall maximum on the left of the track.

3) POSTLANDFALL

After the storm center crosses the coastline, rainfall accumulation over land varies considerably from case to case (Fig. 4). Dry cases urban and savanna display smaller rainfall accumulations and less areal coverage than their moister counterparts. As MA increases, so do rainfall accumulations and areal coverage. Rainfall becomes more symmetrically distributed around the
storm center except for dry cases savanna and, especially, urban, which display higher rainfall accumulations on the right, or water, side. Also noticeable is a rainband least prominent in dry cases savanna and urban. A cross-shaped feature is seen just offshore to the right of the storm track in the accumulated rainfall totals in all cases except urban; this is associated with the rainband being oriented more perpendicular to the coastline around $t = 21$ h and at a more acute angle around $t = 27$ h (Figs. 6c–f). Between those time peri-
ods the areal extent and rainfall rates of the rainband reduce.

All cases display rainbands on the right side of the storm track early during the simulation (Figs. 6a,b and 7). A few hours before landfall (around $t = 12$ h), the rainfall in the bands dissipates (Figs. 6c,d, 4, and 7) but the confluen
t region in the wind field remains (Figs. 6c,d). Dry air intrusion may explain the demise of rainband precipitation with dry air flowing toward the confluence region from the northwest where land is located. It can indeed be seen (Figs. 4 and 7) that rainband precipitation dissipates earlier in drier cases and reestablishes (around $t = 21$ h) to a lesser degree. Once the confluence region reaches the coastline, enhanced low-level convergence on the right side of the storm (Figs. 6e,f) may be responsible for retriggering convection in the rainbands, in all except very dry case urban. A portion of the band remains offshore, but part of it moves onshore (Figs. 6g,h). Onshore rainband rainfall accumulations increase with increasing MA, but are smaller than rainfall accumulations under the path of the eyewall (Fig. 4b).

After the storm centers have crossed the coastline, rainfall coverage increases (Fig. 5d) and rainfall rates in the core (Fig. 5b) decrease. The system broadens (Fig. 6) once the core is located over land, in agreement with previous modeling studies (Tuleya and Kurihara 1978; Tuleya et al. 1984; Jones 1987; Tuleya 1994; Chan and Liang 2003). A possible explanation is that reduction in inertial stability, as the rotational winds reduce, decreases the resistance to horizontal displacement. Figure 16 shows vertical–azimuthal cross sections of ice, snow, and graupel mixing ratios (shaded) and rainwater (colored contours) averaged over $t = 25–30$ h and a 36-km annulus in the eyewalls of urban and broadleaf. Figure 11d shows the same quantities as Fig. 16b for broadleaf but for a time period before landfall. Cloud structures have become mostly stratiform after landfall with decreased rainfall rates. A similar comparison can be made for urban (Figs. 11f and 16a). Li et al. (1997) also saw a transition from convective to stratiform precipitation in their modeled landfalling typhoons.

4. Discussion and conclusions

In this modeling study of hurricane landfall with different surface characteristics, increasing RL has a bigger impact on enhancing storm decay than decreasing MA. However, MA has a greater impact on rainfall amount and distribution. Prior to landfall, approaching storms display a rainfall maximum on the right side of the track. This asymmetry is enhanced for drier land surfaces. At landfall, friction plays a temporary role in forcing a rainfall maximum to the left of the storm track over land. More rain falls in that location for cases with 1) high RL combined with medium-to-high MA or 2) low RL and high MA. After landfall, mean inner-core rainfall rates decrease but do not show much difference between cases. Outer-core mean rainfall rates and rainfall coverage increase and show distinct differences between cases. Both increase with increasing MA, leading to larger and more widespread rainfall accumulations for moister land surfaces. This can be mostly attributed to convection in a rainband flaring up as it interacts with the coastline. Moister cases have more active rainbands. Mean rainfall rates can differ by as much as 4.5 mm h$^{-1}$ which can lead to substantial differences in rainfall accumulations as the storm moves through the area.

The prelandfall asymmetry is forced by dry air intrusion from both the environmental flow behind the vortex and the dry, cool land surface in front of the vortex. Both sources of low $\theta_c$ air combine to destabilize the right side of the vortex and to stabilize the left side. Upon landfall, frictionally driven directional convergence of the radial wind exists on the offshore flow (left) side and speed convergence of the tangential wind on the onshore flow (right) side of the storm center. Cumulus towers form over both convergent regions, but rainfall is advected downstream by strong low-level tangential winds and falls on the left side of the storm track. In all cases the right side convergent zone has access to moisture from the ocean. In moist land surface cases, the convergent region on the left side has surface moisture to work with, leading to a larger rainfall maximum than the drier cases. Even though mean inner-core rainfall rates decrease during landfall, this frictional forcing can lead to large localized rainfall accumulations over land and near the coast of up to 200 mm.

After landfall, all cases experience broadening of the rainfall region and decrease in mean rainfall rates. Except for driest case urban, the rainfall accumulation is distributed evenly on both sides of the storm track. Urban displays more rainfall on the ocean side of the track.

The above rainfall structures are in agreement with some observed hurricanes. Other observations may differ because of the presence of vertical shear, varying SST, and topography—features that were not included in this study. Their inclusion will be the objective of follow-up work.

The left–right rainfall asymmetry has a maximum to the right of the track before landfall, which flips across the left of the track during landfall. This suggests that rainfall structures observed by reconnaissance planes...
over water do not necessarily persist in the same configuration when the storm makes landfall. Therefore, operational forecasters should take extreme caution when making extrapolations from prelandfall reconnaissance data.

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