The Effect of Spaceborne Microwave and Ground-Based Continuous Lightning Measurements on Forecasts of the 1998 Groundhog Day Storm

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ABSTRACT

This study seeks to evaluate the impact of several newly available sources of meteorological data on mesoscale model forecasts of the extratropical cyclone that struck Florida on 2 February 1998. Intermittent measurements of precipitation and integrated water vapor (IWV) distributions were obtained from Special Sensor Microwave/Imager (SSM/I) and Tropical Rainfall Measuring Mission (TRMM) Microwave Imager (TMI) observations. The TMI also provided sea surface temperatures (SSTs) with structural detail of the Loop Current and Gulf Stream. Continuous lightning distributions were measured with a network of very low frequency radio receivers. Lightning data were tuned with intermittent spaceborne microwave radiometer data through a probability matching technique to continuously estimate convective rainfall rates.

A series of experiments were undertaken to evaluate the effect of those data on mesoscale model forecasts produced after assimilating processed rainfall and IWV for 6 h. Assimilating processed rainfall, IWV, and SSTs from TMI measurements in the model yielded improved forecasts of precipitation distributions and vertical motion fields. Assimilating those data also produced an improved 9-h forecast of the radar reflectivity cross section that was validated with a coincident observation from the TRMM spaceborne precipitation radar.

Sensitivity experiments showed that processed rainfall information had greater impact on the rainfall forecast than IWV and SST information. Assimilating latent heating in the correct location of the forecast model was found to be more important than an accurate determination of the rainfall intensity.

1. Introduction

Outbreaks of winter squalls over the Gulf of Mexico have produced widespread damage along the Gulf coast and Florida. Such damage includes flash flooding, electrical power outages, tornadoes, wind shear that adversely affects aviation safety, and rough seas that impact marine activities and adjacent coastal areas. Improved 6–9-h forecasts could be beneficial to mitigate those hazards.

The lack of sufficient data to correctly specify divergence and moisture in the initial conditions of numerical forecast models has contributed to the often-encountered spinup problem described by Davidson and Puri (1992) and others. Because the distribution of latent heating is a major factor affecting the development of tropical and extratropical cyclones, Uccellini (1991), Manobianco et al. (1994), Petty and Miller (1995), and Jones and Macpherson (1997a) were able to demonstrate that assimilating rainfall rates during early stages of cyclogenesis could improve model forecasts. Using lightning as a continuous proxy for latent heating in mesoscale models, Jones and Macpherson (1997b) and Alexander et al. (1999) further improved forecasts of extratropical cyclones.

Unfortunately, it has been difficult in the past to acquire necessary data continuously over oceanic regions during periods of cyclogenesis. The present study takes...
advantage of new measurements to improve numerical weather predictions of a cyclone that developed over the Gulf of Mexico where conventional data were not available. These include additional spaceborne microwave radiometers to measure distributions of precipitation, integrated water vapor (IWV) over oceans, and sea surface temperatures (SSTs), and a spaceborne weather radar that was also recently placed into orbit. A long-range lightning observing network that furnished continuous observations of convective activity had also been established.

Electrification of precipitating clouds is coupled to vertical air motions and microphysical conditions that depend on the development stage of convection within a storm cloud. MacGorman and Rust (1998) describe several studies that suggest that charge separation in thunderstorms is caused by noninductive ice–ice interactions that occur at temperatures \(< -10\,^\circ\text{C}\) in the presence of supercooled liquid water. Those charging mechanisms require strong updrafts to generate supercooled water contents between 0.1 and 5 g m\(^{-3}\). Because of the difference in terminal velocity between the denser negatively charged graupel or hail particles and the positively charged small ice crystals and snow particles, strong updrafts are also needed to separate the charged hydrometeors. MacGorman and Rust (1998) cited studies that related lightning flash rates to radar-echo-top heights, the radar reflectivity above the freezing level that exceeded 40 dBZ, and the vertically integrated liquid water derived by converting radar reflectivity to its liquid water equivalent, and then summing the water content in each column. Correlations thus may be expected between lightning and rainfall generated by convective ice processes.

MacGorman and Rust (1998) cite several rainfall–lightning relationships in which the factors of proportionality between rainfall and lightning varied by several orders of magnitude, depending on the storm regime, that is, oceanic, tropical, arid, frontal, etc. Many of those studies only measured cloud-to-ground lightning, and no distinction was made between contributions from convective rain and warm or stratiform rain. The present study derives convective rainfall from microwave radiometric measurements because those dense ice particles that efficiently scatter microwave radiation also contribute to charge separation.

Alexander et al. (1999) showed that assimilating latent heating derived from lightning had a positive impact on their forecast of the 1993 Superstorm. The 1993 Superstorm evolved close to the coast of the Gulf of Mexico within observing range of the U.S. National Lightning Detection Network (NLDN) described by Cummins et al. (1998). The NLDN consists of ~100 direction finding receivers that permit cloud-to-ground strokes to be measured within the continental United States and out to ~400 km beyond its borders. The present study differs from that of Alexander et al. (1999) in the following respects:

1) The genesis of the cyclone considered in this study began over the Gulf of Mexico and it led to the 2 February 1998 Groundhog Day storm in Florida. Cyclogenesis occurred beyond the range of operational weather radars, NLDN, and other conventional weather observation systems. However a recently developed Sferics Timing and Ranging Network (STARNET-1) permitted the very low frequency (VLF) radio noise emitted by lightning, known as sferics, to be continuously monitored at ranges of several thousands of kilometers from the continental United States. (Sferics can be generated both by cloud-to-ground and by intracloud lightning. The STARNET-1 receivers were not designed to separate those events. Any lightning that triggered a sferics event will be designated simply as “lightning”.) Lightning observations, suitably tuned with intermittent spaceborne microwave radiometer data, could thus be used to continuously estimate convective rainfall distributions. The algorithm relating lightning events to convective rainfall used in the present study was modified from that used by Alexander et al. (1999).

2) Two Defense Meteorological Satellite Program (DMSP) satellites carrying Special Sensor Microwave/Imager (SSMI) microwave radiometers, described by Hollinger et al. (1990), were available. This should have provided better sampling of the Groundhog Day storm thereby improving the forecasts: in fact it did not because the DMSP F-13 and F-14 satellite orbits left critical data gaps. The additional measurements were however used to implement the algorithm that retrieved convective rainfall from lightning data.

3) A new and unique set of sensors on the Tropical Rainfall Measuring Mission (TRMM) satellite, described by Kummerow et al. (1998), had also become available. These included a 13.8-GHz Precipitation Radar (PR), the first weather radar in space, a TRMM Microwave Imager (TMI) and a Lightning Imaging Sensor (LIS). The PR provided a unique opportunity to measure vertical hydrometeor distributions over oceanic regions. The spatial resolution of the TMI was twice as great as that of the SSM/I, although the observing swath was proportionally reduced. An additional 10-GHz channel on TMI enabled SSTs to be determined within only a few days. The LIS, described by Christian (1999), yielded optical measurements of the location, duration, and intensity of lightning strokes that were coincident with the PR and TMI observations. The LIS data could thus be used to calibrate the ground-based STARNET-1 sferics measurements.

The presentation of this study is ordered as follows. Section 2 describes the setting of the Groundhog Day storm, spaceborne radar and microwave radiometer measurements, and algorithms to retrieve convective
rainfall. Lightning measurements and the algorithm that relates lightning rates to convective rainfall rates are then presented. The procedure for estimating IWV and new measurements of SSTs are also discussed in section 2. The mesoscale numerical model, the IWV nudging technique, and the rainfall assimilation technique are described in section 3. Section 4 presents results of the model output using those assimilation procedures. Impacts of the assimilated measurements on rainfall forecasts are then evaluated from the results of experiments that systematically omit each of those data types. Section 4 also presents 9-h forecasted cross sections of vertical motion and radar reflectivity. The latter is compared to the coincident radar reflectivity measured with the TRMM PR. Implications of these findings are then discussed in section 5.

2. Measurement and data retrieval methods

a. Case description

The recent El Niño was at its peak around early February of 1998, a time when Goodman et al. (2000) noted that lightning over the Gulf of Mexico was more frequent than it had been during the previous year. The Groundhog Day storm of 1998 was one of the more intense storms during that period. It evolved from a low pressure system that originated over the southwestern Gulf of Mexico on 1 February and then moved northeastward. As a strong jet streak moved across the Gulf on 2 February, dynamically forced vertical motion over the eastern Gulf intensified the storm system. A mid-tropospheric cutoff low was then formed and it moved in phase with the surface low. That low created an intense squall line that moved across Florida on 3 February. Rippey (1998) noted that the Groundhog Day storm damaged 235 aircraft, produced seven tornadoes, and developed wind gusts of 45 m s⁻¹ around Miami. Much of the Groundhog Day storm evolved beyond the range of operational weather radars and other conventional weather warning systems. Winter cyclogenesis in the Gulf of Mexico poses a special challenge because of strong surface temperature gradients; that is, the North American land surface is colder than the warm and variable Loop Current in the Gulf of Mexico and the Gulf Stream in the Atlantic. The Groundhog Day storm thus provided an opportunity to test a data assimilation technique for winter storms that develop in a data-sparse region. Another motivation for studying this case was that most of the precipitation produced along the squall line was in the form of convective rain that could be inferred from lightning observations and be assimilated accordingly.

Figure 1 shows six infrared (IR) images observed by the Geostationary Operational Environmental Satellite-8 (GOES-8) that record the development of clouds around the Groundhog Day storm. The convective cells appear in somewhat isolated clusters at 0600 UTC on 2 February 1998 and cirrus anvils cover much of the Gulf coast. A squall line developed by 1200 UTC. A secondary convective region that started to develop northwest of the western tip of Cuba is evident at 1800 UTC on 2 February. The cyclone was fully mature by 0000 UTC on 3 February and a dry air intrusion can be seen behind the squall line. The secondary development appears to have become more intense than the original squall line with the passage of time. Unfortunately it was difficult to infer the rainfall and IWV quantitatively by means of IR imagery because of obscuration by cirrus anvils.

b. Rainfall–lightning relationships

1) MICROWAVE AND RADAR MEASUREMENTS OF CONVECTIVE RAINFALL

Rainfall rates can be estimated from operational radar observations, but the maximum range of the radars restricted those data within ~200 km of the continental United States. Rainfall measurements beyond that range limit were provided by SSM/I radiometers that were carried aboard the DMSP F-13 and F-14 satellites and the TMI on the TRMM satellite. The TMI is similar to the SSM/I except that it has an added 10-GHz channel and its spatial resolution was improved twofold. For example, the 85.5-GHz channel footprint size is 4 km × 6 km. This higher resolution improved the accuracy of the retrieved precipitation by reducing beam filling errors. Convective and stratiform rainfall rates were determined from the microwave radiometric measurements using the Goddard Profiling (GPROF) algorithm described in Kummerow et al. (1996). Convective–stratiform separations were obtained from those measurements using the procedure described by Hong et al. (1999). Latent heating profiles were then derived by means of the method described by Olson et al. (1999). Figure 2 shows total rainfall distributions derived from SSM/I and TMI microwave radiometer data during the experiment period. Note the limited swath widths and the irregular temporal coverage. It is especially vexing that the orbits of the DMSP satellites had gaps in their coverage that ran eastward of Florida along the U.S. east coast on 2 and 3 February 1998. Unfortunately, the TRMM satellite orbit overpasses only provided limited sampling improvements.

The PR, operating at 13.8 GHz, measures profiles of precipitation with 250-m vertical resolution. The horizontal resolution is 4.3 km at nadir, but the cross-track resolution is degraded off nadir as the PR scans across the satellite track. The PR reflectivity measurements are limited to values greater than 17 dBZ. Nonetheless the PR provides uniquely detailed information on hydrometeor distributions, especially over oceanic regions previously unobserved from land-based radars. The PR data were therefore used to validate the forecasts made in this study.
2) Lightning Measurements

Lightning data were used to fill sampling gaps because lightning could be measured continuously over large distances, including oceans, from the STARNET-1 ground-based radio receivers. Lightning was monitored by means of the VLF radio noise that it emits between 5 and 15 kHz, known as VLF sferics (hereafter, designated only as sferics), described by Pierce (1977). Because sferics propagate several thousands of kilometer in the earth–ionosphere waveguide, sferics receivers can be widely separated so that a sparse network can observe oceanic regions at a reasonable cost. Sferics are mainly produced by vertical discharges that may occur within clouds or as cloud-to-ground discharges. (Horizontal discharges may also be observed, but their detection efficiency is more sensitive to the location of the discharge with respect to the receivers.) For example, Lee (1986) developed an operational sferics observing network in Europe that provides lightning data over much of Europe and as far as Central America.

STARNET-1, based on Lee's design, benefited from three recent technological developments: 1) the availability of cheap, yet fast, personal computers and signal processing boards operating at speeds greater than 100
Fig. 2. Rainfall rate distributions inferred from the GPROF algorithm of Kummerow et al. (1996) from TMI and SSM/I microwave data obtained between 1210 UTC on 2 Feb and 0235 UTC on 3 Feb 1998. The solid lines represent the boundaries of the observed swaths.
LIS is a ground-based system. The stroke location accuracy of NLDN. Isolated convection at 2300 UTC on 2 February and 0600 UTC on 3 February because a power outage in the Miami area was caused by lightning associated with passage of the squall line. However, data were available from the NLDN and those were normalized to the NLDN data when both were available. Those normalized NLDN data were employed during the outage period. NLDN was more vulnerable to disruption because it relied only on five receivers; future systems will have some redundancy and operate with more flexible signal processing software.

The performance of NLDN was evaluated with respect to that of the LIS and the NLDN. The LIS carried aboard the TRMM satellite is an optical sensor that observes not only the vertically propagating strokes seen by the ground-based receivers, but it can measure all intracold and cloud-to-ground strokes. LIS may therefore measure an order of magnitude more events than the NLDN and NLDN-1 ground-based systems. The stroke location accuracy of LIS is $\sim 4$ km within a $600 \times 600$ km scene. Unfortunately LIS only observes a given location on the earth for 90 s. That however was sufficient to provide an estimate of the location accuracy of strokes over oceanic regions observed by NLDN-1 that were beyond the range of NLDN.

Figure 3 shows the 15-min accumulation of sferics at 2-h intervals between 1400 UTC on 2 February and 1200 UTC on 3 February. Because the angles between azimuth vectors from neighboring NLDN receivers got smaller with increasing distance from the continental United States, NLDN location errors appear to have radially spread the lightning distribution somewhat more than the NLDN-1 measurements. Nonetheless those data provided needed continuity. The absence of events south of Cuba between 0200 and 0600 UTC on 2 February was probably caused by the range limits of NLDN. Isolated convection at 23°N, 86°W, northwest of the western tip of Cuba, begins to be evident at 1600 UTC. It is seen to grow and become more intense than the original squall line by 2200 UTC. Comparison of this sequence with Fig. 1 illustrates the benefit of continuous complementary lightning observations.

Because graupel and hail needed to produce charge separation also scatters 85.5-GHz microwave radiation, the 85.5-GHz brightness temperature observed by TMI diminishes in proportion to the columnar graupel and hail density. The 85.5-GHz brightness temperatures are also affected by the emissivities of underlying land or water surfaces, which depend on the polarization of the emitted radiation. That background effect can be removed by linearly combining horizontally and vertically polarized brightness temperatures to produce an 85.5-GHz polarization corrected temperature (PCT) defined by Spencer et al. (1989). Figure 4 compares the 85.5-GHz PCT and LIS flashes at 2010 UTC with data from NLDN and TMI at 2000 UTC on 2 February. All four images reveal similar features, although the limitations of each sensor system are also evident.

3) CONVECTIVE RAINFALL–LIGHTNING REGRESSION ALGORITHM

The earlier approach of Alexander et al. (1999) derived a ratio between total convective rain flux from SSM/I over the entire domain and total number of lightning flashes. (The horizontal distribution was not considered.) That information was then used to scale lightning rates at intermediate times between SSM/I overpasses. In this study, a more realistic relationship between convective rainfall rate and sferics rate is introduced and applied to estimate continuous convective rainfall distributions.

MacGorman and Rust (1998) show that the region and time of particularly intense rainfall and that of lightning do not exactly coincide. However that effect may be mitigated, but not completely eliminated, if the lightning and convective rainfall are sampled in an area that adequately encloses individual cells, that is, tens of kilometers on a side, and time intervals that sample a significant fraction of the development time of individual cells, that is, $\sim 15$ min.

A procedure for relating convective rainfall to sferics rates that did not rely on exact collocation between grid cells containing lightning and convective rainfall was employed in this study. It was based on the logic presented in the algorithm of Calheiros and Zawadski (1987). That algorithm also avoided giving undue weight to frequently observed small values at the expense of the less frequent, but more significant, large convective rainfall and sferics rates. The algorithm is summarized below:

Apply GPROF to TMI data from 11 orbits over the Gulf obtained during February 1998 to derive convective rainfall rates.

Compute the average convective rainfall rates over $0.5^\circ \times 0.5^\circ$ boxes.
Fig. 3. Sferics distributions measured with STARNET-1 accumulated in 15-min segments displayed at 2-h intervals between 1400 UTC on 2 Feb and 1200 UTC on 3 Feb 1998. NLDN data are displayed between 0200 and 0600 UTC on 2 Feb followed by a restoration of STARNET-1 data at 0800 UTC on 3 Feb 1998. Note the emergence of a second squall line to the northeast of Cuba that becomes more intense than the original squall line after 1400 UTC on 2 Feb.
Count sferics events over $0.5^\circ \times 0.5^\circ$ boxes that occur within $\pm 7.5$ min of the TRMM satellite overpasses. Determine cumulative probability distribution for both convective rainfall rates and sferics rates. For the particular convective rainfall rate, find a sferics rate that has the same cumulative probability to that of the convective rainfall rate.

Figure 5 shows the relationship that was empirically derived from convective systems in the Gulf during early February 1998. Solomon and Baker (1998) showed that relationships between rainfall and flash rates depend on several environmental factors including, among others, the strength of updrafts within the convective clouds. The convective rainfall rate corresponding to zero flash rate may therefore vary from zero to tens of millimeters per hour. A particular rainfall–flash rate relationship should therefore only be expected to be valid under limited conditions. The points in Fig. 5 are results of the closest matching of cumulative probability between convective rainfall rates and sferics rates. The least squares fitted function (curve in Fig. 5) to those data is

$$R = \begin{cases} \frac{1}{30} \times S^{0.46} - 1.1 & (S \geq 1) \\ 0 & (S < 1) \end{cases} \tag{1}$$

where $R$ is the convective rainfall rate in (mm h$^{-1}$) and $S$ is the number of sferics accumulated in a 15-min interval surrounding the TRMM overpass time.

Convective rainfall derived from the GPROF algorithm applied to the TMI measurements is shown in Fig. 6a, and that from the coincident sferics measurements using Eq. (1) is shown in Fig. 6b. The TMI swath is
the same as that shown in Fig. 4a. The location of the squall line up to Yucatan is quite well depicted by the sferics observations even though the rainfall amount is slightly underestimated. A pixel-by-pixel comparison between TMI and sferics rain is presented in the scatter diagram of Fig. 6c. For this particular TMI swath, the threat score for a 1 mm h\(^{-1}\) threshold value is 0.72, the root mean square error is 5 mm h\(^{-1}\), and the mean bias is less than 1 mm h\(^{-1}\). It is also worth noting that the regions with no convective rain in Fig. 6a, which also provide important information for assimilation, are depicted well by this estimation procedure.

c. Integrated water vapor

Distributions of IWV assimilated within the model domain between 1200 and 1800 UTC on 2 February were derived from brightness temperatures measured over sea surfaces by the SSM/I and TMI. Various statistical and physical algorithms described by Sohn et al. (1998) and by Wentz (1997) have been used for that purpose. The IWV distributions were obtained from SSM/I and TMI data in the Remote Sensing Systems Inc.’s archive using algorithms based on the work of Wentz. The maritime IWV distribution from the archived National Centers for Environmental Prediction (NCEP) moisture analysis at 1200 UTC on 2 February is shown in Fig. 7a and the corresponding sferics observation is shown in Fig. 7b. It is evident that the dominant SW-NE tilt of the squall line is not captured in the archived data. These figures can be compared to IWV distributions from SSM/I at 1210 UTC (Fig. 7c) and 1456 UTC (Fig. 7d). Note that the IWV retrieval algorithms screened regions that were obscured by rain so that IWV gaps appear along the squall line. Data from coastal areas, likely to produce dubious IWV retrievals, were also rejected. Comparing the IWV distribution from the archived NCEP analysis with that from SSM/I at around 1200 UTC shows that shapes of those distributions differed significantly along the moist region surrounding the squall line. Moreover, the IWV from the NCEP analysis was larger along the squall line than that derived from the SSM/I measurement. We will show that the erroneous IWV distribution from the NCEP archive produced a poor forecast of the evolution of the squall line. Keeping the IWV distribution consistent with that of the precipitation revealed in the sferics observations overcomes that difficulty.

The temporal coverage for IWV was as incomplete as that shown in Fig. 2. The IWV distributions at 1210 and 1456 UTC, shown in Figs. 7c and 7d, respectively, were the only measurements available during the model assimilation interval. Those distributions were therefore used to morph the IWV between those times using the technique described by Alexander et al. (1999). Morphing requires a subjective evaluation of tie points that identify features in two microwave images that are assumed to have been transformed from the first scene to the second. A least squares coordinate transformation that distorts the first image into the shape of the second is then sought. The coefficients of that transformation are assumed to vary linearly between their value at the first time step and the second. The amplitude of the IWV is also assumed to vary linearly with time on that transformed coordinate system.

No microwave radiometric data were available after 1456 UTC on 2 February for the remainder of the assimilation period; the morphing procedure of Alexander et al. (1999) could not therefore be used at those times. Because continuous measurement of the distribution of lightning permitted the squall line to be tracked, the IWV derived from the SSM/I swath at 1456 UTC (Fig. 7d) was displaced eastward in phase with the squall line movement revealed by the sferics measurements. Processed IWV distributions were thus generated for time steps between 1456 and 1800 UTC on 2 February.

In order to illustrate the validity of displacing the IWV distributions up to 1800 UTC, the 1456 UTC IWV distribution, shown in Fig. 7d, was displaced beyond the end of the assimilation period to 2100 UTC. The processed IWV distribution at 2100 UTC, shown in Fig. 8a, was derived from the IWV at 1456 UTC rotated cyclonically in phase with the sferics distribution displayed in Fig. 3. The processed IWV distribution in Fig. 8a can be compared to the IWV derived from two TMI swaths (Fig. 8b) that passed within ±50 min of 2100 UTC. Because the IWV data in Fig. 7d were only available over the Gulf at 1456 UTC, a data fissure evolved to the east of Texas and Mexico as the processed IWV distribution was moved offshore.
d. Sea surface temperatures

The addition of a 10-GHz channel to the TMI permitted SSTs to be measured even in the presence of clouds by a technique based on the study by Liu (1983). SST maps were produced by Remote Sensing Systems Inc. from 4-day composites of TMI observations. The SST distribution from the NCEP archived $2.5^\circ \times 2.5^\circ$ analysis, shown in Fig. 9a, is compared with that produced by TMI in Fig. 9b. Strong SST gradients around the Loop Current, the Gulf Stream, and near shorelines that were blurred in the NCEP archived analysis are evident in the TMI data. The area around the Loop Current with SST $> 25^\circ$C turned out to be much larger than that in the NCEP archived analysis. The SST in the Yucatan channel shown in Fig. 9b was also greater than that shown in Fig. 9a. Results of sensitivity experiments using both SSTs throughout the entire 24-h model runs will be presented to demonstrate their impact on the forecasts.

3. Simulations

a. Mesoscale numerical model

Version 2 of the Pennsylvania State University–National Center for Atmospheric Research fifth-generation nonhydrostatic Mesoscale Model (MM5) described by Grell et al. (1994) was used in this study. The model physics selected for this study included the Kuo–Anthes cumulus parameterization of Anthes (1977); the explicit treatment of rainwater, cloud water, snow, and cloud ice of Dudhia (1993); a prognostic surface energy budget equation for ground temperature; and the boundary layer parameterization of Hong and Pan (1996). The atmospheric radiation scheme of Dudhia (1989) accounted for the interaction of longwave and shortwave radiation with the surface and with the atmosphere including clouds and precipitation. Davies and Turner (1977) relaxation boundary conditions were used at the lateral boundaries and the radiative boundary conditions of
Klemp and Durran (1983) were used at the top of the model.

The hydrometeors were assumed to be either rain or snow with size distributions as presented in Grell et al. (1994):

\[ N(D) = N_0 \exp(-\Lambda D), \] (2a)

where values of \( N_0 \) are assumed to be \( 8 \times 10^6 \, \text{m}^{-4} \) for rain. The expression for \( N_0 \) of snow is based on the size distribution of Sekhon and Srivastava (1970) and a power law terminal velocity, cited by Sauvageot (1992),

\[ N_0 = 0.108[(\rho_{\text{snow}}/\rho_{\text{air}})/(\pi \rho_{\text{snow}}/\rho_{\text{air}})]^{0.10} \] (2b)

and the \( \Lambda \) (m\(^{-1}\)) in Eq. (2a) is

\[ \Lambda = \left( \frac{\pi N_0 \rho_x}{q_x \rho_{\text{air}}} \right)^{0.25}, \] (2c)

where \( x \) can be liquid or snow as the case may be, \( q_x \) is the mixing ratio of precipitation at each level, \( \rho_x \) are the densities of the hydrometeors (1000 and 100 kg m\(^{-3}\) for \( x = \) rain and snow respectively), and \( \rho_{\text{air}} \) is the density of air at each level.

The model employed 23 vertical layers on a terrain-following sigma vertical coordinate system, where sigma is defined as \( \sigma = (p - p_s)/(p_t - p_s) \), with \( p \) being the pressure, \( p_t \) is the pressure at the top of the model domain, and \( p_s \) is the surface pressure. The resolution of the vertical levels is highest in the planetary boundary layer and decreases above there, with \( \sigma = 0.995, 0.985, 0.97, 0.945, 0.91, 0.87, 0.825, 0.775, 0.725, 0.675, 0.625, 0.575, 0.525, 0.475, 0.425, 0.375, 0.325, 0.275, 0.225, 0.175, 0.125, 0.075, \) and 0.025. The model domain was centered at 30\(^\circ\)N, 85\(^\circ\)W; it consisted of 100 \( \times \) 120 grid points with 40-km separation.

The initial and boundary conditions for the experiments were obtained by interpolating the NCEP 2.5\(^\circ\) \( \times \) 2.5\(^\circ\)
2.5° analysis to the model grid points and then enhancing those with rawinsonde soundings using the objective analysis scheme of Benjamin and Seaman (1985).

b. Rain assimilation technique

Continuous rainfall rate distributions inferred from combined lightning and microwave radiometry in section 2 were assimilated into the forecast model. The rainfall assimilation scheme was based on those of Manobianco et al. (1994) and Jones and Macpherson (1997b), which scaled the model predicted latent heating profiles to the observed rainfall as

$$\Delta T_{\text{m}} = (1 + \alpha) \Delta T_{\text{ml}},$$  \hspace{1cm} (3)

where $\Delta T_{\text{ml}}$ is the temperature tendency calculated from the model latent heating, $\Delta T_{\text{m}}$ is the adjusted temperature tendency, and $i$ is the index identifying the vertical levels of the model. The scaling factor, $\alpha$, is derived from the processed convective rainfall rates, $R_o$, inferred from combined lightning and microwave radiometric measurements and the convective portion of the model predicted rainfall, $R_m$. It is explicitly defined as

$$\alpha = \frac{(R_o - R_m)}{R_m}.$$  \hspace{1cm} (4)

Three possibilities for introducing latent heating profiles were considered. The first regime includes areas in which both $R_o$ and $R_m$ were greater than zero (i.e., $R_o > 0$ and $R_m > 0$). The second regime includes areas where $R_o = 0$ and $R_m > 0$. The third regime includes areas where $R_o > 0$ and $R_m = 0$.

In regime 1, the model-predicted latent heating is scaled by $\alpha$. However scaling is limited to $\alpha \leq 2$ to prevent large temperature changes at any given time step. If the rainfall rate predicted by the model was too low, that is, if $R_o > 3 R_m$, the predicted latent heating profile was assumed to be unreliable. Nearby model grid points were then searched for a more closely matching rainfall rate, which should have been associated with a more realistic latent heating profile. If one of the nearby model-predicted rainfall rates closely matched the processed rainfall, the rainfall rate and latent heating profile at that grid point were used to replace the original grid point. Searches extended to a range of 240 km in this study and more than 90% were successful for this experiment. If the search failed, a maximum scale factor of $\alpha = 2$ was used. If on the other hand, the model-predicted rainfall rate was too high, that is, if $R_o > 3 R_m$, the scaling factor was limited to a minimum value of $\alpha = -2/3$ to ensure that too much heating was not removed from the model. This lower bound was assumed because the actual rainfall may have been from clouds that produced no lightning.

In regime 2, that is, if $R_o = 0$ and $R_m > 0$, the scaling factor $\alpha$ was set to $-2/3$, because of the possibility that electrically inactive convective or stratiform precipitation had occurred. Because latent heating profiles were not generated by the model in regime 3, that is, $R_o > 0$ and $R_m = 0$, the search algorithm in regime 1 was first applied. If the search failed, a normalized parabolic convective heating profile with a midlevel heating maximum at about 500 mb was introduced as in Karyampudi et al. (1998). This specified heating profile was then used to vertically distribute the total latent heating from the observed rain through the temperature tendency equation:

$$\frac{\partial p*T}{\partial t} = p* \frac{L}{C_p} g p_m R_i N_i,$$  \hspace{1cm} (5)
where $N_h$ is the normalized parabolic heating function, $\rho_{\text{liq}}$ is the density of water, $T$ is the model-predicted temperature from latent heat forcing only, and $p^*$ is defined as $p_s - p_r$, where $p_s$ is surface pressure and $p_r$ is the pressure at the top of the model. For regime 3, levels at which latent heating occurs are forced to be saturated as they were in Manobianco et al. (1994). In this study about 15% of the scaled cases fell within regime 3.

c. IWV nudging

Water vapor is an important variable for severe weather forecasting because the moisture distribution directly affects the formation of clouds and precipitation. The processed IWV field described in section 2c was also assimilated into the model using the technique described by Kuo et al. (1993), which was based on the Newtonian nudging technique of Hoke and Anthes (1976) and Stauffer and Seaman (1990). Following the analysis nudging procedure, for the specific humidity variable $q$, the model’s prognostic equation is written as

$$\frac{\partial p^* q(t)}{\partial t} = F + Gp^*[q_{\text{obs}}(t) - q(t)].$$

where $F$ is the normal model forcing terms and $G$ is the nudging coefficient. In this study, we chose $G$ to be $3 \times 10^{-2}$ based on the work of Kuo et al. (1993). Since the IWV is not a three-dimensional variable, they proposed an iterative method to estimate $q_{\text{obs}}$. The basic idea was to nudge the model’s IWV toward the observed value while retaining the vertical structure of the model’s humidity field. Details of this procedure are described by Kuo et al. (1993).

d. Experimental design

The impact of combinations of microwave radiometric and lightning data as proxies for convective rainfall, IWV, and SST on the simulation of the Groundhog Day storm were evaluated in a set of numerical experiments. All experiments were initiated at 1200 UTC on 2 February and integrated for 24 h using 120-s time steps.

Table 1 summarizes the various experiments conducted in this study. The control experiment (CTL) started from initial conditions of the NCEP analysis using only conventionally available data. Experiment ASM replaced SSTs from the NCEP analysis with those described in section 2d throughout the entire integration period. Processed rainfall rates and IWV were assimilated during the first 6 h of the ASM model run. Sensitivity of the forecasts to the new datasets was evaluated by selectively removing one set of data from ASM. NORAIN, NOIWV, and NOSST excluded the data identified by the experiment’s designation.

Because retrieving latent heating from convective rainfall employing sferics and microwave data is a relatively recent development, experiments were also conducted to determine the effect of errors in latent heating derived from convective rainfall based on that technique. For the experiments designated as SCL0.5, SCL0.7, and SCL2.0, the rainfall rates in Eq. (1) were scaled by factors of 0.5, 0.7, and 2.0, respectively, to evaluate the effect of biases in GPROF. In order to determine the possible effect of misregistration of grid boxes containing lightning and rainfall, random errors of $\pm 5$ mm h$^{-1}$ were superimposed on the rainfall rates retrieved from Eq. (1) for experiment RDM5.0. Rainfall rates were set to zero if adding a negative error made them less than zero.
Fig. 10. Surface rainfall rate (mm h\(^{-1}\)) distributions at 1800 UTC on 2 Feb 1998 for (a) CTL, (b) ASM, (c) NORAIN, and (d) NOIWV.

Table 1. Summary of experiments.

<table>
<thead>
<tr>
<th>Experiment designation</th>
<th>Additional assimilated data</th>
<th>Remarks (applied to the processed rainfall rates)</th>
</tr>
</thead>
<tbody>
<tr>
<td>CTL</td>
<td>None (control simulation)</td>
<td></td>
</tr>
<tr>
<td>ASM</td>
<td>Rain, IWV, SST</td>
<td></td>
</tr>
<tr>
<td>NORAIN</td>
<td>IWV, SST</td>
<td></td>
</tr>
<tr>
<td>NOIWV</td>
<td>Rain, SST</td>
<td></td>
</tr>
<tr>
<td>NOSST</td>
<td>Rain, IWV</td>
<td></td>
</tr>
<tr>
<td>SCL0.5</td>
<td>Rain, IWV, SST</td>
<td>Scale by factor of 0.5</td>
</tr>
<tr>
<td>SCL0.7</td>
<td>Rain, IWV, SST</td>
<td>Scale by factor of 0.7</td>
</tr>
<tr>
<td>SCL2.0</td>
<td>Rain, IWV, SST</td>
<td>Scale by factor of 2.0</td>
</tr>
<tr>
<td>RDM5.0</td>
<td>Rain, IWV, SST</td>
<td>Added ±5 mm h(^{-1}) random errors</td>
</tr>
</tbody>
</table>
4. Results

a. Assimilation results

All experiments were initiated at 1200 UTC on 2 February and the 6-h assimilation duration was sufficient to overcome spinup problems. Because the assimilation results can be regarded as new initial conditions for the subsequent forecasts, we now examine the characteristics of the assimilation results for each experiment.

1) Rainfall

Figure 10a shows the surface rainfall at 1800 UTC on 2 February produced by the CTL experiment that used the NCEP analysis to initialize a 6-h forecast. That figure can be compared to Fig. 10b, which shows the rainfall distribution computed at the end of the assimilation period from the ASM experiment that used the newly available SSTs, processed IWV, and convective rainfall. The effect of selectively omitting the convective rainfall in experiment NORAIN is shown in Fig. 10c and Fig. 10d shows the effect of omitting the processed IWV in experiment NOIWV.

Those figures can also be compared with the sferics distribution at 1800 UTC on 2 February shown in Fig. 3, which revealed a well-developed squall line west of the Florida coast that extended to the northern tip of the Yucatan Peninsula with a branch just off the west coast of Cuba. The surface rainfall distribution shown in Fig. 10b is also similar to that derived from the TMI at 2010 UTC, shown in Fig. 6a, if the 2-h time difference is taken into account. The ASM results, in Fig. 10b, are somewhat more diffuse than the actual rainfall because of the 40 km × 40 km resolution of the MM5 model. By contrast CTL, in Fig. 10a, shows a disjointed distribution of precipitation extending to the north of the Yucatan Peninsula, with a spurious cell around Havana, Cuba.
The relative importance of assimilating processed convective rainfall and IWV can be seen when each of these is omitted separately from the assimilation. NORAIN, in Fig. 10c, demonstrates that failing to assimilate processed convective rainfall created a rainfall distribution nearly as disjointed as in CTL. The lack of rainfall southwest of 25°N, 85°W could probably be attributed to the dry air introduced in that region by the processed IWV derived from SSM/I. The rainfall distribution in NOIWV, shown in Fig. 10d, was better organized than that of NORAIN because the assimilated convective rain retained its influence in this experiment. However, NOIWV produced light rain northwest of Yucatan, which may have been caused by a wetter IWV distribution behind the squall line.

2) EQUIVALENT POTENTIAL TEMPERATURE

We review the equivalent potential temperature, $\theta_e$, in the lower troposphere generated by the various experiments. Figure 11 shows $\theta_e$ generated by the CTL, ASM, NORAIN, and NOIWV experiments for $\sigma = 0.995$ at 1800 UTC on 2 February. Compared to the $\theta_e$ distribution of CTL (Fig. 11a), the ASM (Fig. 11b) produced a higher $\theta_e$ band along the two main rainbands in Fig. 10b and lower $\theta_e$ around it. As a result, the horizontal $\theta_e$ gradient for ASM was increased and that probably promoted convective activity around the squall line.

It is evident from the comparison of $\theta_e$ between NORAIN (Fig. 11c) and CTL (Fig. 11a) that most of the reduced values of $\theta_e$ over the Gulf for NORAIN were caused by the assimilation of IWV. Since the initial moisture amount in CTL over the Gulf of Mexico was larger than the IWV observed from SSM/I (Fig. 7), the IWV nudging procedure during the first 6 h for NORAIN and ASM reduced the moisture amount. The distribution of $\theta_e$ for NOIWV (Fig. 11d) shows that the increase of $\theta_e$ compared to CTL mainly resulted from the heating and moistening introduced by rainfall assimilated along the squall line. Consequently, $\theta_e$ changes in ASM appear to have been caused by the combined effect of increased $\theta_e$ from the rainfall assimilation and reduced $\theta_e$ from the IWV assimilation.

3) RELATIVE VORTICITY

The assimilation of processed rainfall rates also impacts the distributions of dynamic fields. Figure 12 compares the 500-mb relative vorticity at 1800 UTC on 2 February for experiments CTL, shown in Fig. 12a, and ASM, shown in Fig. 12b. The CTL experiment produced strong negative vorticity over the Straits of Florida, where intense lightning activity occurred (Fig. 3). By contrast the ASM experiment produced a strong positive vorticity maximum at 24°N, 84°W, which is consistent with the observed lightning location. It is worthwhile to remember that this new developing convection intensified and finally passed over Florida (Fig. 3), suggesting that the latent heating scaling technique efficiently generated a dynamic field distribution along the squall line.

4) LATENT HEAT FLUX

The availability of improved SSTs from the TMI changed the latent heat flux distributions at the surface. Both the CTL and the NOSST experiments used the archived NCEP analysis of SSTs, shown in Fig. 9a, while the ASM used the TMI SST distribution shown in Fig. 9b. The surface latent heat fluxes for CTL, ASM, and NOSST at 1800 UTC on 2 February are shown in Fig. 13. The latent heat fluxes for ASM (Fig. 13b) exhibit a significant moisture flux increase in ASM over
the Straits of Florida, reaching northeastward along Florida’s east coast and following the Gulf Stream, while CTL (Fig. 13a) produced much less intense latent heat flux over the Straits of Florida. This feature is consistent with the squall line track for the subsequent forecast mode. Intense changes of latent heat flux around the Straits of Florida are not only due to increased SST, but they are also caused by the redistribution of moisture and intensified surface wind speed brought about by the assimilation of rainfall and IWV. However, the latent heat flux for NOSST (Fig. 13c) turned out to be much less intense than that of ASM, which suggests that the intensified surface flux in Fig. 13b was mainly due to the improved SST distribution.

The differing SST distributions had a negligible effect on rainfall distribution in ASM and in NOSST at 1800 UTC (not shown) because the same precipitation had been assimilated into both cases up to that time. The effect of differing SSTs on rainfall forecasts at later times will be considered in the next section.

b. Impact on forecasts

1) Sea level pressure and surface wind

The impact of rainfall assimilation on the storm intensity and location was examined for the 18-h forecasts. Sea level pressures and winds for the lowest model layer \((\sigma = 0.995)\) at 1200 UTC on 3 February are shown from the NCEP analysis in Fig. 14a. The corresponding simulation results from the CTL experiment are shown in Fig. 14b, and those from the ASM experiment are shown in Fig. 14c. The analysis showed that the surface pressure center was located more than 180 km south of the Florida panhandle with a 992.5-mb minimum sea level pressure. The minimum sea level pressure for CTL was \(\sim 4\) mb greater than that of the analysis. By contrast
Fig. 14. Sea level pressure (mb) and wind distribution at $\sigma = 0.995$ at 1200 UTC on 3 Feb 1998 for (a) NCEP analysis, (b) CTL, and (c) ASM.

the minimum sea level pressure for ASM was ~2 mb greater than that of the analysis. The low pressure center of the CTL experiment was about 470 km to the northeast of the low pressure center of the analysis. The location of the surface pressure center found by the ASM was slightly closer to that of the analysis with a 350-km error. Also, the maximum wind speed around the low pressure center of the ASM was more intense than the CTL. Assimilated latent heating using processed convective rainfall data had a weaker impact on the surface pressure field in this case than in the earlier studies of Alexander et al. (1999) and Manobianco et al. (1994). That insensitivity might possibly be attributed to the fact that the Groundhog Day storm was not as intense during the 24-h integration period as those cyclones whose minimum sea level pressures ultimately plunged to 960 and 936 mb, respectively.

2) RAINFALL DISTRIBUTIONS

Forecasts at 0300 UTC on 3 February, that is, 9 h past the end of the assimilation period, yielded rainfall distributions shown in Fig. 15. Those forecasted rainfall distributions can be compared to those observed by the National Weather Service (NWS) radar network shown in Fig. 15a and the distribution obtained from the TMI microwave radiometer at 0235 UTC in Fig. 2. The limited range of the radars prevented observation of some of the rain that occurred over the Straits of Florida and over Cuba suggested by the lightning shown in Fig. 2. The maximum rainfall is evident along a band extending from Fort Lauderdale, Florida, to Havana, Cuba.

The control forecast, CTL, shown in Fig. 15b, reveals a rainfall pattern lacking any coherent structure. The forecast from the ASM experiment is shown in Fig. 15c. ASM was able to produce the observed rainband along the Florida Atlantic coast, and considering the half-hour time difference between the TMI observation, the rain over Cuba is predicted reasonably well. The main discrepancy is evident over the Gulf south of Tallahassee, Florida. Although precipitation was evident there at 2331 UTC on 2 February in the SSM/I data shown in Fig. 2, none of the model experiments were able to shut
Fig. 15. Surface rainfall rate (mm h⁻¹) distributions at 0300 UTC on 3 Feb 1998 for (a) NWS radar observation, and forecast of (b) CTL, (c) ASM, (d) NOIWV, (e) NORAIN, and (f) NOSST. The rainfall rates from radar were assumed to be related to the reflectivity, $Z$ (mm² m⁻³), as $Z = 300 R^{1.4}$. The maximum range of the radars confined those observations within ~200 km of the continental United States.
down that precipitation in the subsequent 3 h to match the radar rainfall distribution observed in Fig. 15a.

Figure 15d shows that removing the IWV information in NOIWV produced the rainbands along the Florida east coast as in ASM, albeit with lighter rainfall, and two erroneous rainbands over the Gulf west of Florida. The displaced IWV distribution used in the ASM provided drier air behind the squall line that appears to have suppressed those rainbands.

The result of the NORAIN experiment that removed the processed convective rainfall while retaining all other improvements in ASM is shown in Fig. 15e. The disorganized rainfall distribution in this figure clearly demonstrates the importance of assimilating convective rain during the spinup of the model.

The effect of the improved SST distribution is illustrated in the experiment that reverted back to the SST provided by the NCEP analysis, designated NOSST, whose resulting rainfall distribution is shown in Fig. 15f. The NOSST experiment erroneously placed an 18 mm h \(^{-1}\) maximum rainfall near Fort Pierce, Florida, compared to the maximum rainfall of 25 mm h \(^{-1}\) placed over the Keys by the ASM. By contrast, the rainfall forecasted over the Keys by NOSST was less than 7 mm h \(^{-1}\). Although the increased latent heat flux along the Gulf Stream, produced by the improved SST, may not have changed the rainfall distribution during the assimilation period, it played a positive role in the improvement of the forecast at 0300 UTC on 3 February.

The effect of errors in the processed convective rainfall amount and location on 9-h forecasts was also considered. C. A. Morales (1999, unpublished manuscript) showed that systematic STARNET-1 sferics location biases were less than 50 km over the Texas Gulf coast and less than 20 km approaching Florida. Two experiments arbitrarily displaced the sferics distributions by 80 km. The retarded derived rain distribution produced rainfall forecasts that moved eastward too slowly and the experiment that moved the processed rainfall 80 km too far to the east produced a squall line that moved eastward too rapidly (not shown). These results suggest that sferics location errors less than 80 km do not affect squall-line structure, but they do impact the forecasted positions of the squall line.

Because retrieving latent heating from processed convective rainfall data is a relatively recent development, an experiment was conducted to determine the effect of biases and errors in latent heating based on that technique. Figure 16 shows the sensitivity of rainfall forecasts to biases that might have been introduced by biases in the GPROF retrieval that was used to derive Eq. (1). That potential problem was simulated by scaling the rainfall rate in Eq. (1) by 0.5, 0.7, and 2.0. The result of scaling the processed convective rainfall rate by 0.5, shown in Fig. 16a, demonstrates that assimilating only half of the latent heating produces a rainfall distribution that is as disorganized as the CTL experiment. The assimilated latent heating appears to become irrelevant at that level. Increasing the scaling to 0.7, in Fig. 16b, produces a significant improvement in the ASM-forecasted rainfall distribution. Scaling the ASM rainfall by a factor of 2.0 produced the rainfall distribution shown in Fig. 16c, which was relatively unchanged from Fig. 15c. Some spurious rainfall behind the squall line was reduced, but the maximum rainfall amount over Florida Key was almost unchanged.

Errors in the placement of lightning were simulated by superimposing ±5 mm h \(^{-1}\) of random rainfall on the processed rainfall rates from Eq. (1), keeping all values positive. Figure 16d shows that those errors did not appreciably change the forecasted rainfall.

### 3) CROSS SECTION OF VERTICAL MOTION AND PRECIPITATION

Forecasts of the convective outlook are distributed to the aviation community to warn of potential hazards. The availability of reflectivity cross sections from the TRMM PR radar at 0235 UTC permitted a unique validation of the MM5 forecasts of the vertical extent of precipitation at 0300 UTC. The plan view of reflectivities, dBZ, of frozen hydrometeors directly above the melting layer at 5 km is shown in Fig. 17a. The swath observed by the PR is identified by the two parallel boundaries. A cross section presented in Fig. 17b shows the vertical distribution of dBZ measured by the PR along the line designated A–B shown at the center of that swath.

A comparison between effective reflectivity, \(Z_{\text{eff}}\) (mm\(^6\) m\(^{-3}\)), at the 13.8-GHz operating frequency of the PR and those forecasted by the MM5 can only be qualitative. Computing the \(Z_{\text{eff}}\) of frozen hydrometeors is difficult because of the large number of undetermined parameters needed to model the electromagnetic scattering. Because sferics were observed along the squall line, the frozen hydrometeors probably consisted of small hail or graupel rather than the snow that was assumed to be the only frozen hydrometeor in the MM5 model. The relationship between the effective reflectivity, and the mixing ratio, \(q_{\text{ice}}\) (g kg\(^{-1}\)), of graupel and hail was represented by

\[
Z_{\text{eff}} = 720 \beta k \left( \frac{\rho_{\text{ice}}}{\rho_{\text{air}}} \right)^2 \left( \frac{\rho_{\text{ice}} q_{\text{ice}}}{\pi \rho_{\text{air}}} \right)^{1.75} N_e^{-0.75},
\]

where \(\beta = 0.224\) is the ratio of components of the refractive indices of ice and water; \(\rho_{\text{ice}}\), \(\rho_{\text{air}}\), and \(\rho_{\text{air}}\) are the densities of graupel (400 kg m\(^{-3}\)), water, and air, respectively, as in Eq. (2b); \(q_{\text{ice}}\) is the mixing ratio of graupel at each level, assumed to be the same as \(q_{\text{snow}}\); the snow mixing ratio predicted by the models; and \(N_e\) is \(4 \times 10^6\) m\(^{-4}\) for the graupel size distribution, from Grell et al. (1994). The constant, \(k = 10^4\), converts units of m\(^3\) to mm\(^6\) m\(^{-3}\).

While Eq. (7) has been used in the mesoscale modeling community, its validity in this situation is limited.
Because of the nonlinearity of the \( \text{dBZ}_{\text{eff}} \), the grid-averaged \( q_{\text{ice}} \) is not proportional to \( \text{dBZ}_{\text{eff}} \) averaged over the 40 km \( \times \) 40 km model grid. The grid-averaged mixing ratios, \( q_{\text{ice}} \), were also derived from forecast models that assumed that the frozen hydrometeors were snow rather than graupel. The Rayleigh approximation, used in the derivation of Eq. (7), is only valid for ice particles with \( D < 0.35 \) cm for scattering by 13.8-GHz waves (Sauvageot 1992). Ferrier (1994) suggests that graupel and hail particles can be larger than this; hence, Mie theory should be more appropriate to derive \( \text{dBZ}_{\text{eff}} \). The effect of neglecting to do so overestimates the effective reflectivity for a given mixing ratio. The graupel may not have been dry either; Ulbrich and Atlas (1982) showed that shells of water around an ice core could enhance the effective reflectivity by more than 10 \( \text{dBZ}_{\text{eff}} \) compared to dry ice particles of equal volume. That could offset the effect of the Rayleigh approximation. The PR measurements are limited to values greater than \(-17 \text{ dBZ} \) and the PR data have higher spatial resolution than the forecast models.

In view of these ambiguities, the distribution of \( \text{dBZ}_{\text{eff}} \) shown in Figs. 17c and 17d should be used only for qualitative guidance. Comparing Fig. 17b with Fig. 17c shows that the location of strongest reflectivity in the CTL experiment does not agree well with observations, whereas the ASM experiment results, shown in Fig. 17d, are in better qualitative agreement with the measured radar reflectivity distribution.

Figure 17c also shows cross sections of the forecasted
effective reflectivity superimposed on the vertical motion field, $\omega = \frac{dp}{dt}$ (m bar s$^{-1}$), predicted by the 15-h CTL forecast. These forecasted quantities are inconsistent because the $\omega$ is directed downward 1000 km southeast of A where the precipitation is predicted. By contrast, the upward-directed $\omega$ and the reflectivity maxima are nearly coincident at 1100 km southeast of A in Fig. 17d. That result is more consistent with the observed reflectivity cross section shown in Fig. 17b. A second, weaker updraft 350 km southeast of A in Fig. 17b is also reproduced in the ASM forecast at around 500 km.

The ASM experiment predicted the squall-line location better than the CTL experiment and it also produced narrower and deeper convective motion in the correct locations than the CTL experiment. The vertical motions in the model, smoothed over 40 km $\times$ 40 km grid boxes, are probably two orders of magnitude smaller than the actual updrafts inside the convective cores observed by the PR that were needed to suspend graupel and hail particles. This exercise suggests that operational convective outlooks over remote regions could benefit from the assimilation of the cited new measurements in models.

5. Discussion and conclusions

This study sought to evaluate the impact of several newly available sources of meteorological data on mesoscale model forecasts of the 1998 Groundhog Day storm. Measurements of precipitation and IWV distributions were provided by two SSM/I microwave radiometers and the improved TMI microwave radiometer on the TRMM satellite. The GPROF algorithm, to retrieve precipitation distributions from those measurements, was also improved compared to the algorithm used by Alexander et al. (1999). SSTs measured from the TMI yielded a map of SSTs that had a warmer Loop Current and Gulf Stream with sharper horizontal SST gradients than the NCEP-archived SST distribution. The TRMM also carried a PR that provided high-resolution
measurements of radar reflectivity that could be used to validate the model forecasts.

Although lightning is not as prevalent over oceans as over land, there were a sufficient number of sferics to permit a quantitative analysis of convection over the Gulf of Mexico and over the Atlantic Ocean. This was consistent with documented observations of maritime lightning by Maury (1963) and Orville (1990). The STARNET-1 long-range sferics monitoring network provided coverage of convective activity along the squall line over all of the Gulf of Mexico that enabled the evolution of the Groundhog Day storm to be monitored almost continuously. The NLDN provided complementary data over some of the Gulf of Mexico during a brief STARNET-1 data outage. STARNET-1 was able to measure sferics originating within a few thousand kilometers from the receivers, but the sensitivity and ranging accuracy of those receivers diminished with range. Because the sensitivity and stroke location accuracy of LIS are uniform, coincident LIS and STARNET-1 observations were used to evaluate the performance of STARNET-1.

Convective rainfall estimates, obtained from TMI microwave measurements, were compared to nearly coincident sferics rates to derive an empirical relationship between sferics and convective rainfall rates. The correlation between the rainfall rates inferred from sferics and those retrieved from the TMI microwave measurements had a 0.72 threat score and a 5 mm h\(^{-1}\) root-mean-square error with no evidence of bias for 1 mm h\(^{-1}\) threshold value. However, the validity of that empirical relationship was only demonstrated over the Gulf of Mexico for the Groundhog Day storm and it should be regarded as provisional until more comprehensive studies determine the variability of such an empirical relationship.

Experiments were undertaken to evaluate the impact of assimilating latent heating from processed rainfall, displaced IWV, and revised SST distributions on the forecasts. The 6-h assimilation results produced reasonable distributions of temperature and moisture as well as the organization of the vorticity field, which were necessary to initiate the Groundhog Day storm. The assimilation technique that scaled the model-produced latent heating profiles in proportion to processed rainfall rates successfully reproduced observed rainfall distributions. The state of the model atmosphere following the assimilation period initiated conditions that improved the forecast of the evolution of the Groundhog Day storm. Sensitivity tests showed that assimilating the correct processed rainfall was more important for reproducing observed rainfall distributions than either IWV or SST. However, the IWV and SST measurements also appeared to have contributed somewhat to conditions that favored the correct evolution of the squall line.

The 6-h assimilation of processed rainfall and IWV along with the enhanced SST significantly improved the 9-h rainfall forecast and it slightly improved the 18-h storm track and intensity forecast. Sensitivity experiments showed that processed rainfall information had the greatest effect on the successful 9-h rainfall forecast. The IWV assimilation and enhanced SST made positive, albeit smaller, contributions to the 9-h rainfall forecast. IWV assimilation eliminated some of the erroneous rainbands behind the squall line. Although the improved SSTs did not change the rainfall distribution during the assimilation period, increased latent heat flux along the Loop Current and the Gulf Stream, produced by the enhanced SSTs, was probably able to improve the 9-h rainfall forecast.

Because the CTL simulation produced too little rainfall over the observed squall-line region and spurious rainfall over the west coast of Florida during the assimilation period, initiation of the squall line in the ASM experiments depended mainly on the processed rainfall over those regions. Processed rainfall reduced by a factor of 0.5 was unable to initiate squall-line evolution during the following forecast period, but increasing the processed rainfall to a factor of 0.7 was sufficient to initiate the evolution of the squall line. Although a doubled ASM latent heating was assimilated in one experiment, the model tended to adjust to environmental conditions and diffusion processes in the subsequent forecast that seemed to balance the excessive latent heating. This suggests that the forecast is relatively insensitive to overestimated processed rainfall while underestimation of processed rainfall only produces forecasts that revert back to the original CTL results. It is encouraging that the processed convective rainfall estimation technique proposed in this study can operate within a somewhat broad range of biases in GPROF and in geographic and temporal variations of the sferics detection efficiency. Forecast results were also insensitive to \(\pm 5\) mm h\(^{-1}\) random errors.

These results are consistent with those found by Manobianco et al. (1994) who pointed out that errors of a factor of 2 in assimilated rainfall rates had little effect on forecasts, but that the raining regions needed to be correctly located. Accurate measurements of rainfall from spaceborne sensors are desirable for many applications. However, this study showed that accurate location of intense convection is more important for assimilation into a forecast model. Those findings may relax accuracy requirements of spaceborne microwave brightness temperature measurements, rainfall retrieval algorithms, and relationships between sferics rates and rainfall.

The results of this study showed that the ASM experiment employing recently available measurements predicted the squall-line location within 100 km in advance. It also predicted a hydrometeor distribution that was qualitatively similar to that observed from a spaceborne precipitation radar. The ASM also forecasted convective motion that was narrower and deeper than the CTL forecast that used conventionally available
data. These limited results suggest that the newly available observations, especially of precipitation, may ultimately contribute to improved aviation weather hazard warnings.

More studies to extend the relationship between lightning and convective rainfall to other precipitation regimes will be necessary. The intense rainfall in this case was convective so that lightning measurements were able to track the most significant latent heating. That will not be generally true, so that the total rainfall for other cases may have to be derived from other continuously available data, such as IR imagery from geostationary satellites.

STARNET-1 was shut down in March 1998. However, long-range sferics observing networks, whose design is based on experience gained with STARNET-1, are being put into operation in Europe and the western Pacific basin. A long range lightning detection network, described by Cramer and Cummins (1999), has also been assembled from receivers comprising the NLDN and the Canadian Lightning Detection Network. Data from that system are being disseminated by the NWS Aviation Weather Center.

A geostationary optical Lightning Mapping Sensor (LMS) has been proposed by Christian et al. (1989) to provide continuous, uniform lightning measurements over the Western Hemisphere in the next decade. The present analysis suggests how continuous lightning data from LMS and other long-range lightning observing systems might be used to improve weather forecasts.

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