Quantifying the Imprint of a Severe Hector Thunderstorm during ACTIVE/SCOUT-O3 onto the Water Content in the Upper Troposphere/Lower Stratosphere

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ABSTRACT

The development of a severe Hector thunderstorm that formed over the Tiwi Islands, north of Australia, during the Aerosol and Chemical Transport in Tropical Convection/Stratospheric-Climate Links with Emphasis on the Upper Troposphere and Lower Stratosphere (ACTIVE/SCOUT-O3) field campaign in late 2005, is simulated by the Advanced Research Weather Research and Forecasting (ARW) model and the Met Office Unified Model (UM). The general aim of this paper is to investigate the role of isolated deep convection over the tropics in regulating the water content in the upper troposphere/lower stratosphere (UT/LS). Using a horizontal resolution as fine as 1 km, the numerical simulations reproduce the timing, structure, and strength of Hector fairly well when compared with field campaign observations. The sensitivity of results from ARW to horizontal resolution is investigated by running the model in a large-eddy simulation mode with a horizontal resolution of 250 m. While refining the horizontal resolution to 250 m leads to a better representation of convection with respect to rainfall, the characteristics of the Hector thunderstorm are basically similar in space and time to those obtained in the 1-km-horizontal-resolution simulations. Several overshooting updrafts penetrating the tropopause are produced in the simulations during the mature stage of Hector. The penetration of rising towering cumulus clouds into the LS maintains the entrainment of air at the interface between the UT and the LS. Vertical exchanges resulting from this entrainment process have a significant impact on the redistribution of atmospheric constituents within the UT/LS region at the scale of the islands. In particular, a large amount of water is injected in the LS. The fate of the ice particles as Hector develops drives the water vapor mixing ratio to saturation by sublimation of the injected ice particles, moistening the air in the LS. The moistening was found to be fairly significant above 380 K and averaged about 0.06 ppmv in the range 380–420 K for ARW. As for UM, the moistening was found to be much larger (about 2.24 ppmv in the range of 380–420 K) than for ARW. This result confirms that convective transport can play an important role in regulating the water vapor mixing ratio in the LS.

1. Introduction

The characteristics of the upper-troposphere/lower-stratosphere (UT/LS) region are intrinsically determined by those of both spheres. The balance of processes that regulate its dynamical, radiative, and chemical characteristics is at the heart of the debate on UT/LS exchanges. In particular, the quantification of the respective

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role of convective overturning in the troposphere and radiative and/or diabatic overturning in the stratosphere in determining the entry of atmospheric constituents into the LS is still a polemical issue for air quality and climate applications (e.g., Forster and Shine 1999; Kirk-Davidoff et al. 1999). The troubling uncertainty regarding the balance of processes contributes to the uncertainty of trends in the future composition of the atmosphere.

Riehl and Malkus (1958) showed for the first time that towering cumulus clouds that reach the tropopause have a large impact on the energy balance of the tropical atmosphere. The height to which tropical convection reaches and affects the composition of the UT/LS region remains uncertain. The transition layer where both tropospheric and stratospheric processes interact is usually referred to as the tropical tropopause layer (TTL; e.g., Atticks and Robinson 1983; Holton et al. 1995; Highwood and Hoskins 1998). This transition zone is typically defined as the region extending from the main convective outflow (10–14 km; Folkins 2002) up to the cold point tropopause (16–17 km; Highwood and Hoskins 1998). The cold point is usually defined as the level of the minimum temperature in the vertical temperature profile. Note that the position of the cold point is thus defined very close to the position of the minimum water vapor saturation mixing ratio (e.g., Dessler 1998; Zhou et al. 2001). The TTL region is key for most air entering the stratosphere. Deep convective clouds occasionally reach the height of the TTL and even the deepest clouds often penetrate up to the cold point (e.g., Danielsen 1982, 1993). By destabilizing the TTL, convective updrafts affect the vertical transport of atmospheric constituents into the stratosphere. Indeed, several studies have clearly indicated a predominant role for such convective events in determining temperature (Sherwood et al. 2003; Kuang and Bretherton 2004; Robinson and Sherwood 2006), water vapor (Sherwood and Dessler 2000, 2001; Read et al. 2004), and other tracer profiles (Dessler 2002) near the cold point. Nonetheless, some studies have shown that most of the tropical convection does not extend beyond 14 km in altitude (e.g., Folkins et al. 1999). This altitude is often referred to as the mixing barrier, above which the effects of convection drop off rapidly. The altitude of the mixing barrier roughly corresponds to the lower boundary of the TTL. Hence, the cold point has been suggested as a stratospheric feature that is decoupled from convection (e.g., Thuburn and Craig 2002). In that case, cross-tropopause transport is a result of large-scale ascent (e.g., Reid and Gage 1996). This mechanism still implies air to cross the level of zero net radiative heating (LZH), above which air rises and below which air descends. Note that this picture does not require or exclude overshooting convection as an important mechanism for cross-tropopause transport. A thorough discussion, further focused on water vapor entry in the stratosphere, is presented in section 5.

As stated above, deep convection over the tropics is thought to play a significant role in determining UT/LS exchanges. Basically, there are three regions of intense deep convective activity over the tropics, namely South America, Central Africa, and the Maritime Continent, which includes the Indonesian archipelago and northern Australia. Tropical convection is especially intense over the Maritime Continent during the buildup of the monsoon (e.g., Keenan et al. 1994). During the Australian premonsoon season (typically in November–December), severe storms are usually tied to the islands on the southeast extremity of the Maritime Continent, which are strongly influenced by coastal and diurnal effects. Thunderstorms occur almost every afternoon over the islands during this period (Keenan et al. 1990). It is during the mature part of the buildup just prior to the monsoon onset when the storms reach their maximum intensity (May and Ballinger 2007). Note that during the monsoon season, convection usually peaks over the ocean in the early afternoon and over the land in the early morning (e.g., Keenan et al. 1989a; Chen and Houze 1997). The difference in the timing of convection is mainly due to day to night variations in radiative heating between cloudy and clear-sky regions (Gray and Jacobson 1977) and in radiative cooling of cloud tops (Kraus 1963), to the semidiurnal solar atmospheric tide (Brier and Simpson 1969) over the ocean, and to coastal and topographic effects over the land (Houze et al. 1981). Also, monsoon convection usually produces more rain, with a larger contribution of stratiform rain, over a larger area, but there are fewer vigorous convective cells (May and Ballinger 2007).

There have been numerous insightful observational studies devoted to deep tropical convection over the western equatorial Pacific (viz., in the range 100°E–180°). For instance, the Australia Monsoon Experiment (AMEX; Holland et al. 1986), the Equatorial Mesoscale Experiment (EMEX; Webster and Houze 1991), and the Stratosphere–Troposphere Exchange Project (STEP; Russell et al. 1993) programs were designed to investigate convective mesoscale systems in the monsoonal flow north of Australia. These three programs dealt mainly with mesoscale processes, and did not detail the isolated thunderstorms that develop over the islands during the premonsoon season. The Island Thunderstorm Experiment (ITEX; Keenan et al. 1989b) and the Maritime Continent Thunderstorm Experiment (MCTEX; Keenan et al. 2000) were the first programs to examine these storms in detail. These programs focused on the giant thunderstorms (known as Hector thunderstorms) that develop over the Tiwi Islands, around 80 km north of Darwin, Australia. Deep convection was found to be initiated mainly by cold air pool interactions and convergence of penetrating sea-breeze flows (e.g., Carbone et al. 2000). The Darwin Area Wave Experiment (DAWEX; Hamilton et al. 2004) program also investigated deep convection in the area of Darwin, and was further dedicated to the study of tropical internal gravity waves and their relation to tropospheric convection. In late 2005 and early 2006, the Darwin area hosted two intensive observation periods (IOPs), which involved three consortium programs, namely the Stratospheric-Climate Links with Emphasis on the Upper Troposphere and Lower Stratosphere (SCOUT-O3; Vaughan et al. 2008), the Tropical Warm Pool International Cloud Experiment (TWP-ICE; May et al. 2008), and the Aerosol and Chemical Transport in Tropical Convection (ACTIVE; Vaughan et al. 2008). SCOUT-O3 was held in November–December 2005 and was focused on deep convection and its relation to UT/LS exchanges and ozone chemistry. TWP-ICE was conducted in January–February 2006 and designed to relate cirrus properties to characteristics of both the environment and deep convection. As for ACTIVE, it was held concurrently with both SCOUT-O3 and TWP-ICE and thus served as an important complement to both programs. ACTIVE focused particularly on the measurement of a variety of chemical species, including aerosols (Allen et al. 2008). The aerosol component of ACTIVE was partly motivated by the outcome of the second phase of the Egrett Microphysics Experiment with Radiation Lidar and Dynamics (EMERALD; Whiteway et al. 2004) program, which provided a unique view of the anatomy of cirrus outflow from Hector thunderstorms. Aerosols were not targeted during this program while they strongly affect the dynamical and microphysical structure of cirrus clouds (e.g., Connolly et al. 2006).

ACTIVE was aimed at determining the impact of deep convection on the composition of the UT/LS region and its relative importance compared with large-scale ascent. During the first phase of ACTIVE, which was held concurrently with SCOUT-O3, the emphasis was put on studying isolated thunderstorms. The IOP was conducted from 7 November to 9 December 2005 over the Tiwi Islands during the buildup of the summer monsoon. Vaughan et al. (2008) described the science objectives and the design of the IOP. A range of observational equipment was deployed during this IOP including several research aircrafts, radiosondes, and precipitation radars, such as the Bureau of Meteorology Research Centre–National Center for Atmospheric Research (BMRC–NCAR) C-POL radar (Keenan et al. 1998). This dataset has provided significant insights into the onset and development of isolated thunderstorms over northern Australia.

Though these observational studies are crucial information sources to clarify the processes involved in UT/LS exchanges, measurements cannot provide a complete picture of the evolution of the 3D structure of deep convective events. In this context, high-resolution cloud-resolving models have been a helpful tool to refine and quantify these processes, and to confirm results from observations. In support of observational studies of isolated island thunderstorms, only a few numerical studies were performed. To our knowledge, the first ones were conducted by Golding (1993) using the Met Office Unified Model (UM; Golding 1992) at 3-km horizontal resolution to examine 2 days from the ITEX field campaign. Using a simplified initialization from soundings, the simulations were able to successfully reproduce the development of deep convection over the Tiwi Islands. Convection was found to be first initiated where the convergence of sea-breeze flows was maximum, due to the shape of the coastline. The timing, strength, and location of convection were predominantly determined by the convective available potential energy (CAPE), given sufficient heating. Saito et al. (2001) investigated an observed case from MCTEX using the Japanese Meteorological Research Institute Non Hydrostatic Model (MRI NHM; Saito and Kato 1999) at 1-km horizontal resolution. Good agreement was obtained between the simulated and observed developments of Hector. The sensitivity of the development of deep convection to microphysics, and the topography and size of the islands was also examined. Crook (2001) performed idealized simulations of Hector using both linear and nonlinear models. The effects of low-level wind, CAPE, and sensible and latent heat fluxes on the strength and location of convection were investigated. Consistent with observations, it was found that major convection occurs along the downwind side of the islands and that the intensity of Hector storms increases when the direction of the low-level flow turns toward the major (i.e., longer) axis of the islands.

The predictable and isolated nature of Hector thunderstorms makes them ideal case studies for evaluating numerical models. It is noteworthy that, apart from the simulation by Saito et al. (2001), there is no realistic simulation of Hector published in the literature. Saito et al. (2001) used a 1-km horizontal resolution, which is likely to be insufficient to resolve faithfully the basic thunderstorm structure (e.g., Bryan et al. 2003). Furthermore, the impact of deep convection on the composition of the TTL was not investigated.

In the present paper, we discuss results from the Advanced Research Weather Research and Forecasting (ARW) model (Skamarock et al. 2007) and the Met Office UM (Davies et al. 2005) for the golden day during the first phase of ACTIVE (i.e., 30 November 2005) for which a severe Hector thunderstorm developed. These two models are representative of the next generation of numerical weather prediction models and both of them have already been applied to tropical area. Using two models is expected to lead to more robust conclusions and to eventually provide deeper insights in the processes under investigation. High-resolution cloud-resolving simulations have been performed using a horizontal resolution as fine as 250 m. The setup of the models is presented in section 2. The strategy for evaluation of the models is discussed as well. The large-scale environmental flow over northern Australia during the Hector episode is described in section 3. Four main phases have been identified in the development of the Hector thunderstorm during this episode: onset, triggering, Hector, and dissipation. These phases are detailed in light of results of the ARW and UM models in section 4. The impact of convective overshooting on the redistribution of atmospheric constituents within the UT/LS is discussed in section 5. In particular, we focus on the understanding and quantification of the imprint of Hector onto the water content in the LS. A summary and concluding remarks are given in section 6.

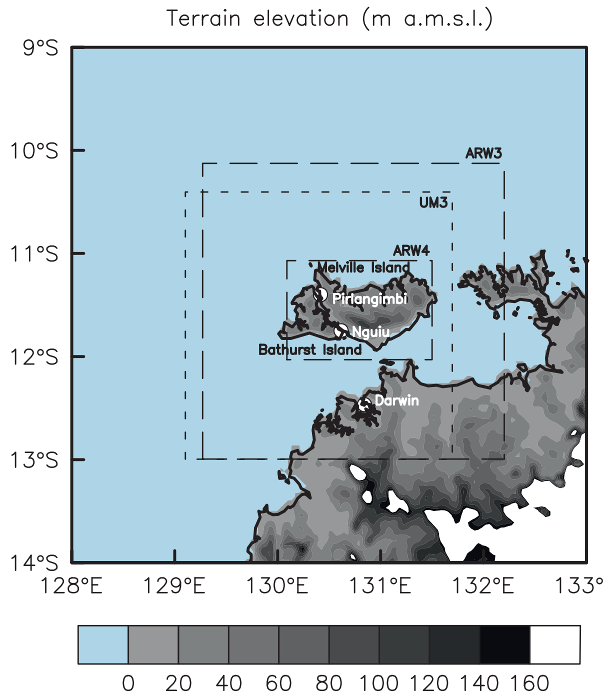

FIG. 1. Orography of the Tiwi Islands. The attached color scale indicates altitude (m MSL). The dashed polylines represent the areas of the domains ARW3, UM3, and ARW4 (see the text and Table 1).

2. Setup of the ARW and UM models

The ARW and UM models were used to investigate the dynamics of a severe Hector thunderstorm over the Tiwi Islands. The Tiwi Islands consist of a pair of closely spaced islands (Bathurst Island and Melville Island) separated from each other by a narrow tidal channel (Apsley Strait). This group of islands is about 100 km long and 75 km wide. The orography of the islands is quite gentle (see Fig. 1). The ridge to the south of Melville Island reaches only height up to about 120 m above mean sea level (MSL). The terrain slopes gently down to the sea with shallow tributary valleys. Vegetation is mainly composed of open forests and grasslands and is not dense because of the traditional indigenous land management practices. Indeed, vegetation is burnt during the dry season in order to prevent fire outbreaks from developing and to preserve grasslands for hunting.

a. Downscaling and grid setup

The two models were run on multiple grids using one-way nests. Table 1 gives the spatial coverage and the horizontal resolution of the grids used for the simulations. The domains using a 1-km horizontal resolution (ARW3 and UM3) are displayed in Fig. 1. The ARW model was run down to a horizontal resolution as fine as 250 m to investigate the sensitivity of the results to horizontal resolution. The motivation is that a 1-km horizontal resolution might be too coarse to faithfully represent physical processes of cloud turbulence that occur in deep convection (see Bryan et al. 2003, for a guide to resolution requirements for the simulation of deep moist convection). For that specific run at high resolution, 621°N–S $\times$ 621°E–W grid points were used to cover a domain over the islands (see Fig. 1). The computations with ARW were made on 115 vertical levels up to 3 hPa. The grid was stretched along the vertical axis to yield an averaged vertical grid spacing of 500 m and to accommodate a high resolution ($\sim$100 m on average) both close to the ground surface and within the UT/LS region. The spatial resolution within the UT/LS region needs to be fine enough to resolve a range of convective structures and to reduce artificial gravity wave damping. A vertical resolution of the order of 100 m is very likely to be sufficient to resolve the short vertical wavelengths of the gravity waves induced by deep convection in the Darwin area. Indeed, these wavelengths were observed to be typically in the order

TABLE 1. Resolution of the grids used for the simulations.

Domain	Typical extent	Grid points (N–S, E–W)	Grid size
UM0	Globe	324, 432	5°/9°, 5°/6°
ARW1	Northern Territory	141, 141	16 km
UM1		320, 320	0.11°
ARW2	Area of Darwin	221, 221	4 km
UM2		200, 310	0.036°
ARW3	Tiwi Islands (large)	341, 341	1 km
UM3		290, 290	0.009°
ARW4	Tiwi Islands (small)	621, 621	250 m

of a few kilometers (e.g., Tsuda et al. 2004). Two distributions of vertical grid spacings were employed for the UM runs. For the grids with a horizontal resolution larger than 1 km, we used 38 vertical levels, with a top at 1 hPa. The vertical grid spacing was in the order of 1000 m within the UT/LS region. As for the run with a 1-km horizontal resolution, 76 levels were used and the vertical grid spacings were reduced by half.

The topography of the Tiwi Islands is basically flat (see Fig. 1). Nonetheless the ridge to the south of Melville Island may have an impact on the timing, structure, and strength of deep convection. Saito et al. (2001) found that the development of Hector tends to be delayed and the maximum vertical velocity becomes smaller when the topography is flattened. Hence, a high-resolution topography would be required to accurately simulate Hector events. In our ARW runs we used digital elevation data from the Shuttle Radar Topography Mission (SRTM; Farr et al. 2007) at ~90-m resolution. As for the land cover, we used the Moderate Resolution Imaging Spectroradiometer (MODIS) MOD 12 land cover product at 1-km resolution (Friedl et al. 2002) for the domains ARW3 and ARW4 and remapped it to the U.S. Geological Survey (USGS) classification implemented in ARW. For the other domains and the other characteristics of the soil and the ground surface (such as soil type and monthly surface albedo), static data were derived from the default geographical data that are provided with the WRF preprocessing system. As for the UM, vegetation and soil types were extracted from the Advanced Very High Resolution Radiometer (AVHRR) International Geosphere-Biosphere Programme (IGBP) dataset (Loveland et al. 2000). Orography was processed from the Global Land One-Kilometer Base Elevation (GLOBE) dataset (Hastings and Dunbar 1998), except for the finer-resolved domain (UM3), for which the SRTM dataset was used (as for ARW). Both the AVHRR IGBP and GLOBE datasets are available at 1-km resolution. Other soil and ground surface data were specified from climatological values.

Initial and lateral boundary conditions of the coarser domain ARW1 were derived from the European Centre for Medium-Range Weather Forecasts (ECMWF) gridded analyses available every 6 h with a horizontal resolution of 0.5° on operational pressure levels up to 3 hPa for vertically distributed data, and surface and soil levels for surface and deep-soil data. As for the UM, the global run was initialized by a data-assimilated global dump provided by the Met Office. A grid nudging technique (e.g., Stauffer and Seaman 1990) was employed for the coarser domain ARW1 during the first 6 h in order to shorten the spinup time. No nudging was performed in the UM simulation. A relaxation zone covering a few grid cells around each domain (5 in ARW and 12 in UM) was employed to smooth gradients near the lateral boundaries. In our work, we focus on the Hector storm that developed in the afternoon on 30 November 2005. A tropopause fold passed over the simulated domains from 23 to 27 November 2005, irreversibly mixing stratospheric air into the troposphere. As a result, the troposphere was very dry in the range of 200–600 hPa. This persistent dry layer prevented convection from developing over land, reducing cloud cover (to about 10%) and precipitation (to about 50 mm a day) over the area of interest. Hector storms developed in the afternoon on 27–30 November 2005. We found that starting the ARW simulation on any of these days does not change much the timing, structure, and strength of Hector on 30 November 2005. However, it does have a significant impact on the water content in the UT/LS. This is shown in section 5c. Basically, we found that starting the simulation on 28 November 2005, gives the best set of results when compared with observations and that starting it later on in time may give misleading results. Therefore only the results from the ARW simulation starting on 28 November 2005 are presented hereinafter. As for UM, the simulation was started at 1200 UTC 29 November 2005, since starting it on 28 November 2005 would lower its forecasting skill considerably.

b. *Parameterization of physical processes*

For the simulations using a horizontal resolution down to 1 km, subgrid-scale (SGS) mixing was classically parameterized within the boundary layer physics. We used the Yonsei University (YSU; Hong et al. 2006) and operational (Lock et al. 2000) boundary layer parameterization schemes in ARW and UM, respectively. Both schemes are nonlocal and take explicitly into account entrainment at the top of the boundary layer. These schemes assume that there is a clear scale separation between the subgrid and resolved scales, so that they can be treated independently. This assumption is less clear as grid sizes approach a few hundred meters or less. In these cases, the schemes should be replaced by a fully 3D local SGS scheme, and such simulation is referred to as large-eddy simulation (LES).

The LES capability of ARW was not applicable to real cases in its current release (version 2.2). As a matter of fact, the SGS scheme was not coupled with the surface fluxes, which are calculated by the soil–vegetation model. For the finer-resolved domain (ARW4) using a horizontal resolution of 250 m, we recoupled these fluxes and included them in the buoyancy and shear production terms of the turbulent kinetic energy equation (E. D. Grell 2007, personal communication). Note that we used the level-1.5 subgrid-scale scheme by Deardorff (1980).

The classical Monin–Obukhov surface layer scheme was used to provide surface forcing in terms of momentum, heat, and moisture fluxes. The land surface energy budget was calculated by the Noah soil–vegetation model (Ek et al. 2003) in ARW and by the Met Office Surface Exchange Scheme (MOSES; Essery et al. 2003) in UM. The radiation schemes provide surface downward longwave (LW) and shortwave (SW) radiation for the land surface energy budget as well as atmospheric heating due to radiative flux divergence. In ARW, we used the Community Atmosphere Model, version 3 (CAM3), radiation package (Collins et al. 2006). As for the UM, the Edwards–Slingo radiation scheme (Edwards and Slingo 1996) was employed. Both schemes include a multiband calculation of atmospheric heating for both the LW and SW streams.

For the coarser grids with a horizontal resolution larger than 4 km, SGS effects of convective and shallow clouds were parameterized by a mass flux–based cumulus scheme. We used the ensemble cumulus scheme introduced by Grell and Dévényi (2002) in ARW and a scheme based on the Gregory and Rowntree (1990) cumulus scheme in UM. Note that the latter scheme had numerous enhancements to the original scheme since (in particular, separate updraft and downdraft parameterization, convective momentum transport, and CAPE closure). For the finer-resolved grids with a horizontal resolution of 1 km and less, convection was explicitly resolved and the cumulus scheme was switched off. We used the microphysical schemes by Thompson et al. (2004, 2008) and Wilson and Ballard (1999) in ARW and UM, respectively. The Thompson scheme has six classes of hydrometeors: water vapor, cloud liquid water, rain, cloud ice, snow, and graupel. In addition, the number concentration of cloud ice is predicted as is the case in double-moment schemes. As for the version of the Wilson and Ballard scheme that we used, it has prognostic water vapor, cloud liquid water, rain, total ice (ice and snow), and graupel.

c. *Strategy for evaluation of the models*

The models were evaluated using observational data from the IOP. Results of the simulations at 1-km horizontal resolution were compared with both ground surface and vertically distributed measurements. In the present study, the strategy for evaluation of the models and further analysis has to select the most appropriate variables to optimally characterize the key flow features and physical processes. Hereinafter, we decided to use the following variables: vertical velocity, sensible heat flux, precipitation rate, and water mixing ratios.

The nominal boundaries of the TTL region can be defined in several different ways based either on the thermal or dynamical structure of the region. A dynamical definition of the TTL using particular values of potential vorticity (PV) might not be appropriate over the tropics where PV becomes ill conditioned (e.g., Gettelman and Sobel 2000). The definition adopted in our work is based on the thermal structure of the UT/LS region. Convenient boundaries for the TTL are the potential temperature lapse rate minimum at about 11 km in altitude (~345 K in isentropic coordinates) and the cold point tropopause at about 17 km (~370 K). The main focus of our study is whether and how deep convection affects the water content in the UT/LS. Hence, an important boundary that occurs within the TTL is the LZH at about 15 km (~355 K). Indeed, only convection that reaches this level facilitates the ascent of air into the LS. Therefore, the LZH is used as the lower boundary of the TTL (as in Sherwood and Dessler 2000) and termed the TTL height hereinafter.

The minimum time period at which the results need to be output to capture the nonstationary structure of deep convection should be smaller than the typical overturning time scale τ of strong updrafts. Let us assume that the typical vertical extension h of strong updrafts is in the order of the TTL height (15 km) and that their maximum vertical velocity w_* is about 25 m s^{-1} (see section 4b). This yields $\tau \equiv h/w_* = 10$ min. Hence we fixed the frequency of the outputs to 10 min.

3. Large-scale environmental flow

The synoptic condition over the Maritime Continent, and in particular over northern Australia, is mainly driven by the monsoon trough (e.g., McBride 1987). When the wind regimes from the Northern and Southern Hemispheres are well distant from the equator, the intertropical convergence zone (ITCZ) forms a monsoon trough. Basically, the trough lies north of Australia during the Australian summer monsoon. The air converges over the Maritime Continent (Ramage 1968), which is dominated by low-level westerly flow and upper-level easterly flow. During the IOP, the onset of the monsoon started on 26 December 2005. The preceding period may be considered as a buildup period. During the buildup of the monsoon, the environmental flow is usually dominated by easterly flow (e.g., Keenan and Carbone 1992; Kullgren and Kim 2006). Low-level winds with a westerly zonal component were not often observed during the IOP. A low-level northwesterly flow was observed on 30 November 2005, the flow above being westerly up to about 12 km and easterly aloft (see Fig. 2). As suggested for instance by Crook (2001), this synoptic pattern favors the triggering of Hector thunderstorms. Indeed, deep convection is initiated as a result of convergence of sea-breeze flows and its intensity is maximum when the low-level flow is along the major axis of the islands. The timing and structure of the sea-breeze flows is detailed in section 4.

A severe Hector thunderstorm started to develop on 30 November 2005 at about 1400 LT (local time = UTC + 9.5 h). The initiation of deep convection requires conditional instability and a trigger mechanism, which is discussed in section 4. During MCTEX, Keenan et al. (1990) found that thunderstorms develop in an environment with low shear, moderate CAPE [O(1500 J kg^{-1})], and high moisture. In our case, at 1430 LT the convective inhibition was rather weak and the CAPE was very large (~3464 J kg^{-1}; see Fig. 2). The precipitable water was found to be 5.0 cm, which value is close to that observed during MCTEX (4.7 cm).

Figure 3 shows Multi-Functional Transport Satellite (MTSAT-1R) satellite images over northern Australia every 6 h from 1000 to 2200 LT. Prior to deep convection this region did not experience high clouds (see Fig. 3a at 1000 LT). By early afternoon deep convective clouds extended over the area of Darwin and the southeast coast of Melville Island. As from about 1500 to 1600 LT, deep convective storms intensified (see Fig. 3b at 1600 LT) and produced heavy precipitation. Thunderstorms lasted for about 2 h. At the end of the afternoon high clouds spread all over the islands. By early evening the convective system dissipated. Finally, in the evening remaining clouds had moved off the islands (see Fig. 3c at 2200 LT).

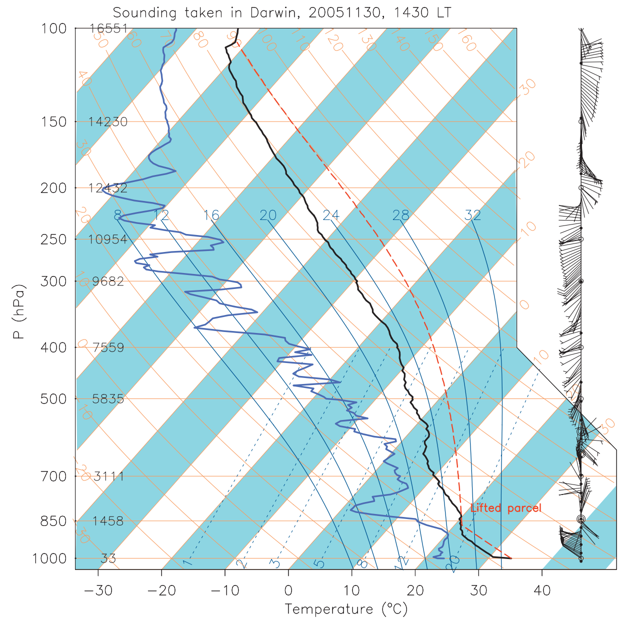

FIG. 2. Skew T–log$p$ diagram plotted from the sounding taken in Darwin at 1430 LT 30 Nov 2005.

4. Structure of the Hector thunderstorm

a. Horizontal structure

During the morning hours, the boundary layer grows rapidly as a result of ground surface heating, reaching a height z_i of about 2000 m by early afternoon. The air originating from the westerly flow above the low-level northwesterly flow is entrained into the mixed layer once the low-level flow encroaches on the layer of westerly flow, which is located at about 1000 m (see Fig. 2). Later on in the afternoon, the height of the boundary layer is more difficult to define. As Hector develops, deep convective cells strongly modify the structure of the boundary layer. Basically, the depth of the mixed layer may be either shallow or deep depending on the location and strength of the convective cells. In this case, the boundary layer height is a quantity that is not really appropriate to characterize the deep convective cells.

The large-scale environmental flow affects the timing, structure, and strength of the atmospheric circulation over the islands as discussed for instance by Crook (2001). Nonetheless, thunderstorms develop almost every afternoon during the premonsoon season. The life cycle of the observed Hector storm is described in the following. Sea-breeze fronts form all along the coastline as a result of the temperature gradient between the ocean and the land. This is the onset stage. The inflow depth was about 800 m and lay between the mixed layer depths over the ocean and over the land. This result is consistent with the findings of Oliphant et al. (2001) and Schafer et al. (2001) during MCTEX though being shallower by a few hundred meters (depths averaged 1200 m). The northwesterly low-level flow led to conditions that were particularly favorable for convective initiation. Indeed, the low-level flow was almost along the major axis of the islands. Hence, the convergence of

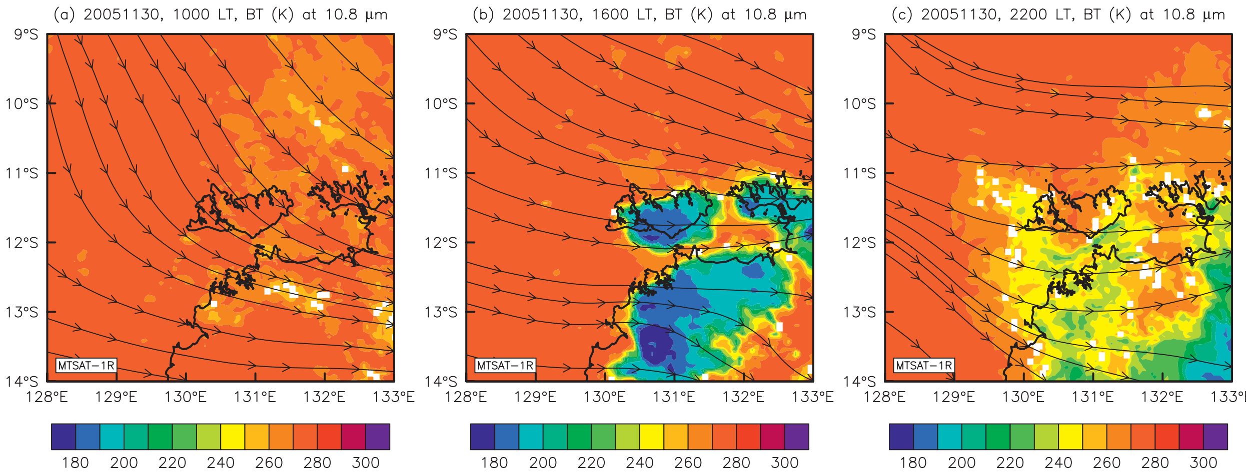

FIG. 3. Brightness temperature (BT) at 10.8 μm from MTSAT-1R satellite imagery over the area of Darwin at (a) 1000, (b) 1600, and (c) 2200 LT 30 Nov 2005. Typically, high clouds are characterized by the lowest BT at 10.8 μm. Streamlines at 850 hPa derived from ECMWF analyses are superimposed.

the sea-breeze flow is intensified (see also Liu and Moncrieff 1996; Carbone et al. 2000; Crook 2001; Wilson et al. 2001). The air is thus more likely to be lifted to the level of free convection. Note that enhanced convergence may also occur as a result of either gust or cold or dry fronts, or be induced by topographic features (Kingsmill 1995).

Sea-breeze flows first develop around 1300 LT along the windward coast north of the islands and propagate more rapidly than along the leeward coast. Indeed, along the south coast the sea-breeze front opposes the low-level flow. Thus, the progression of the sea-breeze front is slowed down and the air is lifted further. At that stage (hereinafter referred to as triggering), the convective activity over the interior of the islands is still typical of that of the convective boundary layer. Shallow clouds form at the top of the boundary layer and spread over the islands. Maximum temperature and minimum pressure take place over the center of the islands defining a heat low that causes island-scale convergence (see also Skinner and Tapper 1994).

As from about 1400 LT, sea-breeze flows penetrate inland rather rapidly (within one hour or so) and converge over the center of the islands (see Figs. 4a,b). Thunderstorms develop along the convergence line north of the ridge to the south of Melville Island. Though the ridge reaches only height up to about 120 m, the topography has a significant impact on the atmospheric circulations and consequently on the characteristics of the storm. The elevation associated with the escarpment makes the ridge a region of maximum heating along the convergence line and thus a preferred region of storm genesis. By 1500 LT, the sea-breeze flows have penetrated to the center of the island and reach their maximum strength. At that time, vertical velocities close to the ground surface are maximum and exceed 1 m s^{-1} (see Figs. 4c,d). This is the Hector stage.

By 1530 LT, the air temperature in the center of the islands has dropped by about 3°–5°C as a result of the development of cold pools from thunderstorm outflows. At the same time the surface water vapor mixing ratio has increased because of the heavy rainfall. Radially spreading cold pools force air to rise, generating new convective cells, which in turn generate cold pools. At that stage, convection is self-sustained and free from the sea-breeze fronts. As Hector develops, the strength of the sea-breeze flows is dampened rapidly by the thunderstorm outflows.

Sea-breeze flows all along the coastlines have significantly decreased in intensity by 1600 LT and an offshore flow develops as cold air induced by the storms spread out (see Figs. 4e,f). As the storms spread out horizontally at the surface over the center of the islands, gust fronts traveling offshore develop occasionally. The gust fronts can force warm air to rise and locally lead to the development of clouds along the coastlines (as illustrated in Fig. 8b). By 1700 LT, the storm is moving downstream westward, similarly to convection bands during the monsoon (e.g., Keenan and Brody 1988) and dissipates (see Figs. 4g,h). The dissipation stage lasts for about 2 h. Later on, the oceanic boundary layer air is advected over the islands and resets the timing and structure of the Hector convective system.

It is noteworthy that the values of sensible heat flux (see Fig. 5) obtained with ARW and UM can differ by up to 20%–30%. This can be explained by different

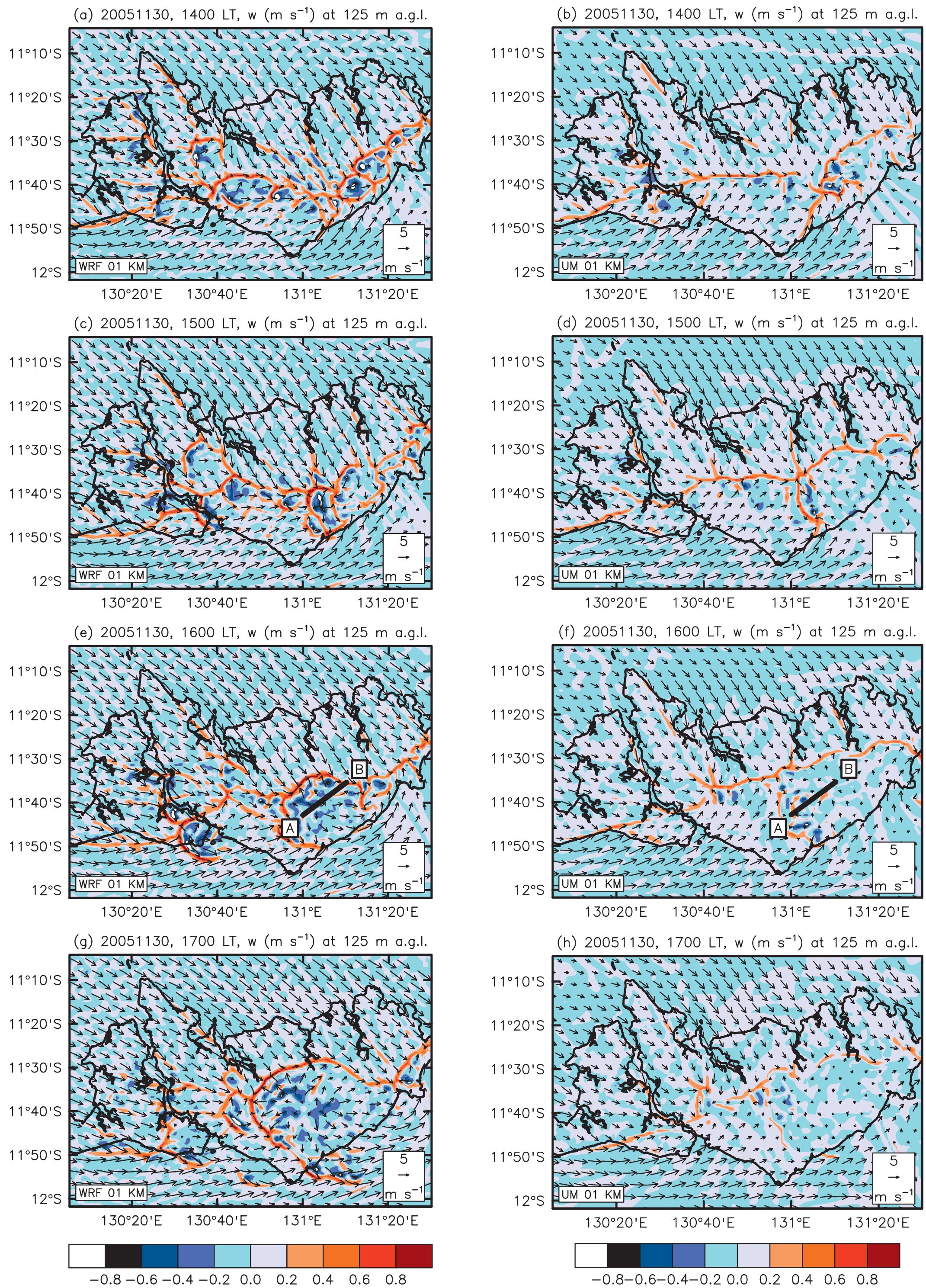

FIG. 4. Vertical velocity field w at 125 AGL simulated by (left) ARW and (right) UM in domains ARW3 and UM3, respectively, at (a),(b) 1400; (c),(d) 1500; (e),(f) 1600; and (g),(h) 1700 LT. Horizontal wind vectors at 10 m AGL are superimposed. The black line (A–B) in (e) and (f) gives the position of the vertical cross section in Figs. 7 and 8c.

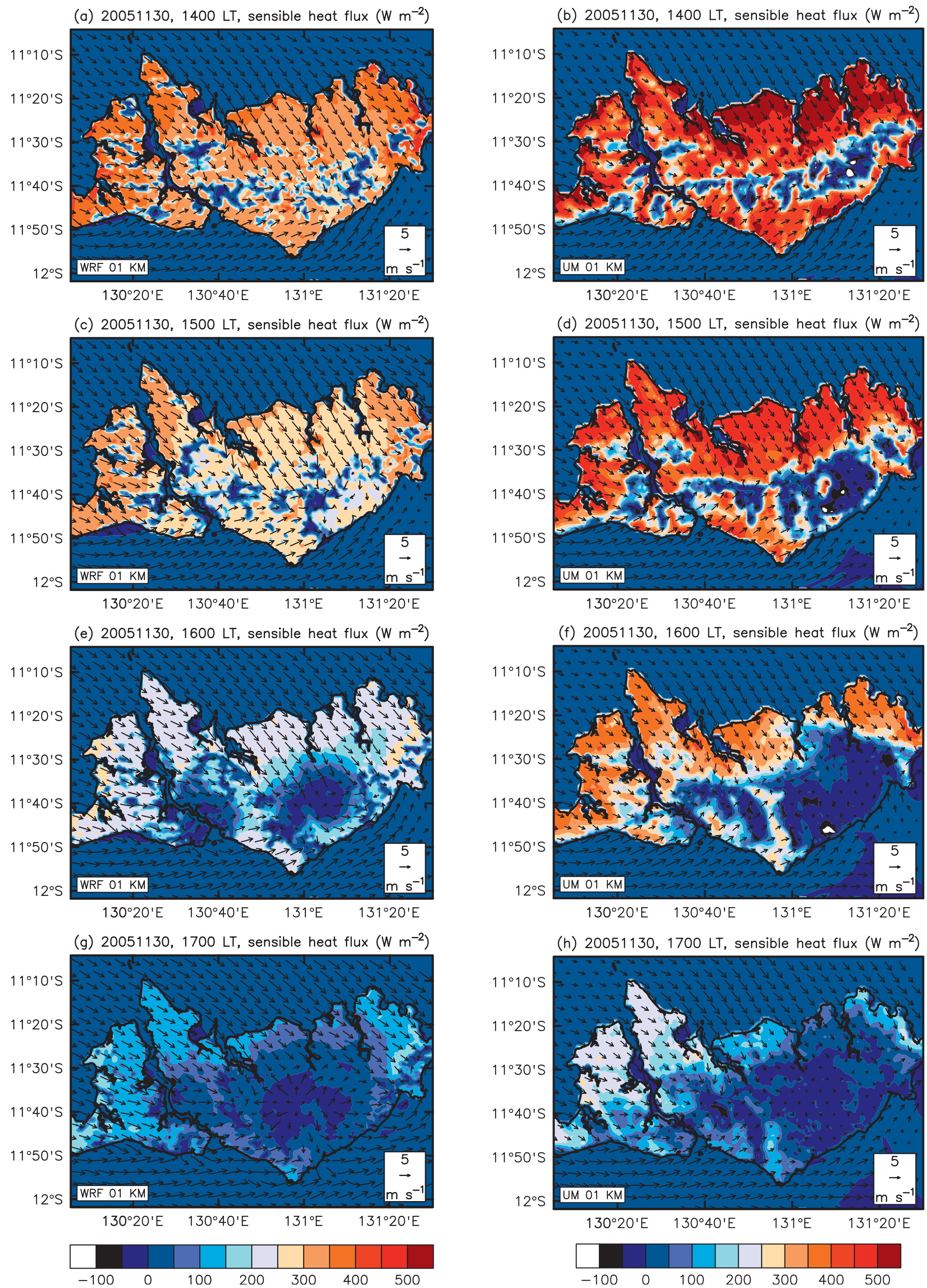

FIG. 5. As in Fig. 4, but for sensible heat flux.

soil and ground surface characteristics (as mentioned in section 2a) as well as a more extensive cloud cover simulated by UM. One might expect the structure and strength of Hector to be different in both runs. Indeed, the development of deep convection depends on the surface heat fluxes since they warm, mix, and moisten the boundary layer, which needs to contain sufficient moisture to support deep convection as well as precipitation.

Interestingly, both models reproduce the development of Hector fairly well. Consistent with this result, Simpson et al. (1993) found that the islands still provided sufficient surface heat fluxes to initiate convection on a day when the islands were under a dense cirrus overcast. This would mean that the intensity of Hector storms could not be controlled only by the surface heat fluxes. Robinson et al. (2008) suggested that the spatial extent of ground surface heating can control the intensity of Hector. The authors argued that deep convection is stronger when the vertical to horizontal aspect ratio of the thermally driven boundary layer is equal to the ratio of the vertical wavelength to horizontal wavelength of propagating internal gravity waves.

The early stages of the development of Hector (onset and triggering) show very little convective precipitation. Initially, the storm has characteristics of nonsquall multicellular convection (i.e., convective cells separated by stratiform regions). The convective regions are characterized by large vertical velocities greater than the typical fall speed of hydrometeors (i.e., about 2–3 m s^{-1}). The convective cells propagate in a discrete way and are persistent in a sense that at least one cell is active. As for the stratiform regions, they are characterized by vertical velocities that are much less than the typical fall speed of hydrometeors. In that case, ice particles within the upper layers of clouds fall while growing (e.g., Steiner et al. 1995).

Thunderstorms are first initiated to the east of Melville Island and then move westward (see also Keenan et al. 1994; Carbone et al. 2000). During the mature stage of Hector (see Fig. 6), the convective activity over the islands increases rapidly. The merging of active convective cells leads to a rapid development of the Hector thunderstorm. The convective complex organizes along an east–west line over the center of the islands. The most intense precipitation is associated with this period of explosive convection, with precipitation rates locally as large as 100 mm h^{-1}. Later on, when the developing cold pool generated by thunderstorm outflows encompasses most of the islands, a reorientation of the convective system to a north–south line perpendicular to the environmental shear is observed. The storm then moves westward with a squall-like structure (i.e., convective front with a trailing stratiform region). In the dissipation stage, stratiform precipitation becomes prevalent and moves off the islands westward. To complement the qualitative comparison of the spatial distribution of rainfall rate in Fig. 6, the performance of the models has been further quantified by a statistical comparison of the rainfall rate over the domain ARW4. Hourly-accumulated precipitation generated by ARW and UM every hour from 1230 to 2130 LT has been compared with that from the BMRC/NCAR C-POL radar measurements. The normalized mean bias is −21.15% for ARW and −51.62% for UM.

b. Vertical structure

The initial convection during the triggering stage is rather shallow with cloud tops in the range of 8–10 km. Upward motion in some clouds reaches 8–10 m s^{-1} and precipitation starts. After that the development of Hector has a stepwise character as storm cells merge to form tall hot towers. The duration of this intense convection is in the order of 1 h. A convective complex with an extensive anvil cloud develops in the mature phase, producing heavy rainfall. Note that mergers represent only about 10% of the convective systems while producing about 90% of the rainfall over the islands (Simpson et al. 1993). In Fig. 7, the vertical distribution of microphysical variables simulated by the models at 1600 LT is compared with the classification of hydrometeors from the BMRC/NCAR C-POL radar measurements at 1430 LT. The measurements were processed by the Bureau of Meteorology, Australia, to create horizontal cross sections at different altitudes. Note that the time for comparison is not identical. Indeed, for the comparison to be qualitatively fair, we had to select a time for which the radar scanned an active convective cell. Cloud-top height reaches 17 km and an anvil develops at the top of the convective complex. The maximum vertical velocity is located around 10–12 km and exceeds 22 m s^{-1}. Note that the maximum simulated values over the inner domains ARW3 and UM3 reach 30–40 m s^{-1} (not shown). This is consistent with the continuous measurements at Darwin discussed by May and Rajopadhyaya (1999), for which the most intense updrafts occur above the melting level (at about 5 km altitude). Hydrometeors are produced and uplifted by deep convection. Within updrafts exceeding several meters per second, upper-level hydrometeors grow rapidly by riming (liquid water freezing onto ice crystals). Graupel (ice crystals that undergo extensive riming) is found between the melting level and the LZH (at about 15 km altitude), and either falls out or provides nuclei for hail. Note that in the simulations performed with ARW, hail is not discriminated from graupel. The melting level is easy to identify in Fig. 7a as it is known to give a bright band of reflectivity in the stratiform cloud layers and hence a discontinuity in the class of hydrometeors. While falling out, hydrometeors grow by vapor diffusion and precipitate out below the melting level. Rain eventually occurs and a downdraft forms all around the updraft while creating a cold pool at the ground surface.

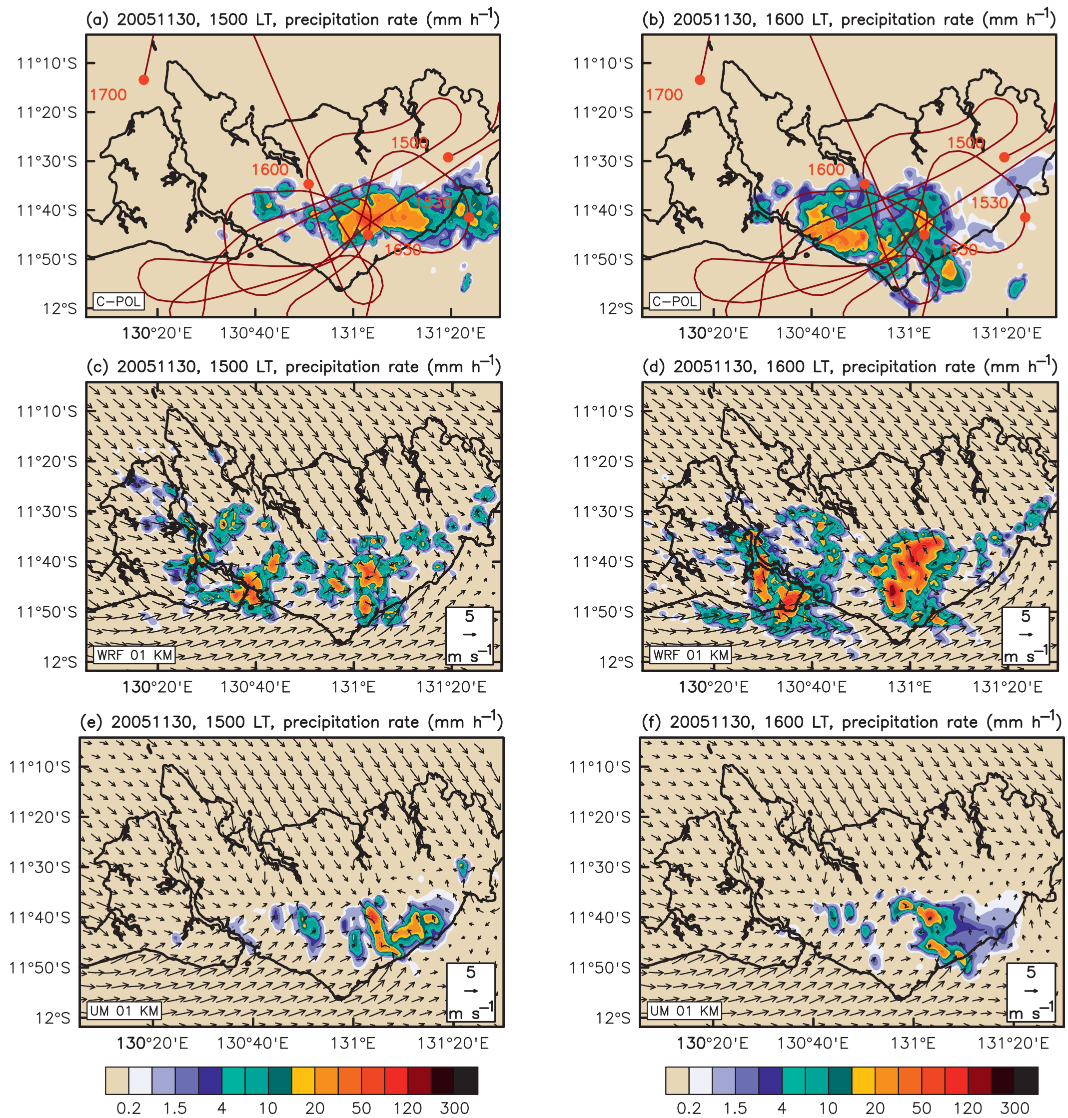

FIG. 6. Hourly-averaged precipitation rate (top to bottom) derived from the BMRC/NCAR C-POL radar measurements and simulated by ARW and UM in domains ARW3 and UM3, respectively, at (a),(c),(e) 1500 and (b),(d),(f) 1600 LT. The flight track of the M55-Geophysica research aircraft from 1500 to 1700 LT 30 Nov 2005 is superimposed on (a) and (b) and is discussed in section 5b. Horizontal wind vectors at 10 m AGL are superimposed on (c)–(f).

The remaining hydrometeors that do not fall out are mainly composed of snow and ice crystals above the freezing level. These hydrometeors are subjected to weaker vertical motions and remain aloft to form an anvil, which can be either advected or left decaying. The strong updraft displayed in Fig. 7 does overshoot its level of neutral buoyancy. In that case, rather small ice particles are lifted higher up and spread out. The persistence of the anvil as well as the injection of small ice particles above the tropopause can lead to changes in the radiative balance in the atmosphere and either a dehydration or a moistening of the LS. Elements of quantification of such impacts are proposed in section 5.

c. *Sensitivity to horizontal resolution*

To assess the impact of refining horizontal resolution on the development of Hector we compared results of the simulations performed with ARW using horizontal resolutions of 1 km (domain ARW3) and 250 m (domain ARW4). Figure 8 shows the Hector stage at 1600 LT as simulated over the domain ARW4. Basically the features of the storm do not change much and the flow structure is much more detailed when

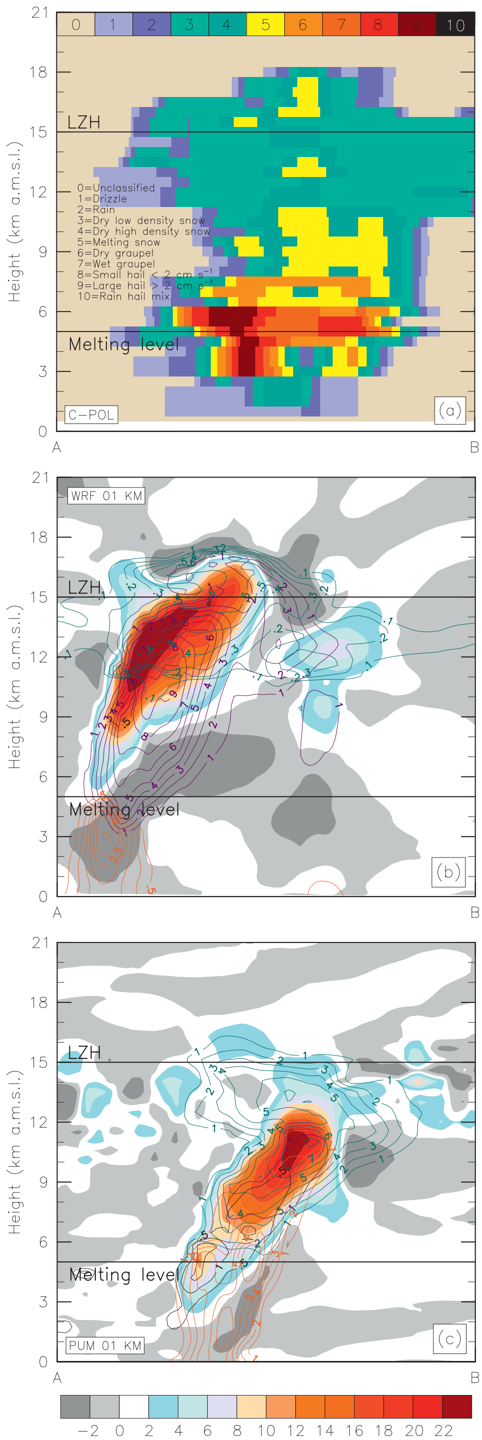

FIG. 7. Vertical cross section along the line A–B in Figs. 4e,f: (a) microphysical classification from the BMRC/NCAR C-POL radar measurements at 1430 LT; (b) rain (orange), cloud liquid water (black), graupel (purple), and ice (green) mixing ratios simulated by ARW at 1600 LT in the domain ARW3, with 0.5, 0.5, 1, and 0.1-g kg^{-1} interval contours, respectively; (c) as in (b), but for UM in the domain UM3, except with 1 g kg^{-1} interval contours for the cloud liquid water and total ice mixing ratios. The attached color scale for (b) and (c) indicates vertical velocity (m s^{-1}).

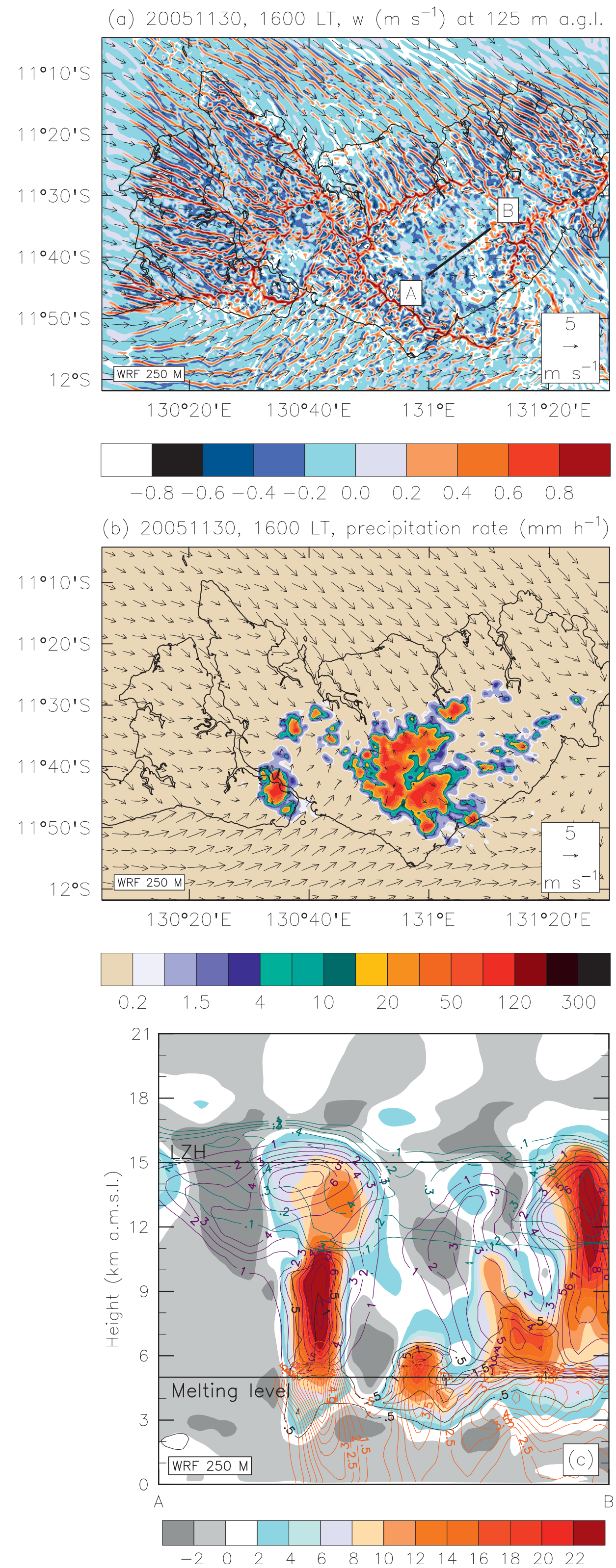

FIG. 8. As in Figs. 4e, 6d, and 7b, but for ARW in the domain ARW4.

increasing horizontal resolution. As expected, the higher-resolution simulation better resolves the convection. The normalized mean bias for hourly-accumulated precipitation is 15%. This represents an improvement on the coarser-resolution simulation, for which the normalized mean bias was −21.15%.

The low-level northwesterly flow imposes an onshore flow along the windward coast north of the islands and an offshore flow along the south coast. The onshore flow facilitates the inland penetration of the sea-breeze front. Conversely, the offshore flow results in a sharp front close to the coastline. This explains that Hector first developed along the leeward coast since the front is likely to produce greater lifting. Horizontal convective rolls [see Etling and Brown (1993) for a review] perpendicular to the sea-breeze fronts develop over the clear-sky areas. This feature is illustrated in Fig. 8a and was not so clear in the 1-km horizontal-resolution simulations. Note that horizontal convective rolls were also suggested by Saito et al. (2001) during MCTEX. The wavelength that is associated with the horizontal convective rolls is in the order of 3–4 km. This value is in good agreement with the theoretical wavelength $\ell = 2\sqrt{2}z_i$ determined by Kuettner (1971), yielding a roll aspect ratio of about 2.8. As discussed in section 4a, the boundary layer height z_i is a quantity that is not really appropriate during the triggering and Hector stages, but prior to convection z_i was typically in the order of 2 km, giving $\ell \sim 5$ km. These rolls develop in an environment with sufficient speed shear (e.g., Etling and Brown 1993) and tend to be aligned parallel to the vertical wind shear vector. A low-speed shear environment usually produces cellular convection close to that observed in the simulations at 1-km horizontal resolution. The fact that horizontal convective rolls are not recognizable using a horizontal resolution of 1 km is certainly due to a too coarse horizontal resolution. Indeed, a high horizontal resolution being 500 m or less is necessary to produce horizontal convective rolls (e.g., Balaji and Clark 1988).

As discussed in great detail by Dailey and Fovell (1999), the interaction of the sea-breeze front with these rolls produces spatial variation in the sea-breeze front and can provide sufficient lifting to initiate deep convection. Indeed, lifting is enhanced at frontal locations where roll updrafts intersect. The enhanced convection induces subsidence above the roll downdraft intersections that generates a positive convective feedback helping to maintain convection. The same mechanism do exist along the convergence line in Fig. 8a where along-line vertical velocity maxima enhance deep convection. A similar mechanism was observed by Wilson et al. (1992) at the intersection of a convergence line with horizontal convective rolls near Denver, Colorado. The results of the simulation using a 250-m horizontal resolution suggest that this mechanism might play an important role in modulating convection in the frontal environment over the islands. However, one need to bear in mind that the situation can be much more complicated when the sea-breeze front and the environmental low-level flow are not aligned (e.g., Fovell and Dailey 2001).

5. Imprint of Hector onto stratospheric water vapor

It is widely accepted that overshooting convection does occasionally penetrate the TTL up to the cold point (e.g., Liu and Zipser 2005) as is the case in Fig. 7 for instance. There has been considerable discussion over the years as to whether a significant amount of tropospheric air is injected into the stratosphere by deep convection at the global scale. Understanding whether and how overshooting convection affects the water content in the UT/LS region is particularly important since water vapor does have a significant impact on the radiative balance of the atmosphere as a greenhouse gas, and hence on climate.

a. *Review of our current understanding*

The extreme dryness of the stratosphere prevailed on Brewer (1949) to suggest that air enters the stratosphere primarily in the tropics, where it is dried by condensation. Newell and Gould-Stewart (1981) narrowed down this tropical region where the tropical tropopause is cold enough to dehydrate the air entering the stratosphere to account for the low water vapor mixing ratio. The authors referred to this temporally and spatially limited region as the stratospheric fountain, which included the western tropical Pacific, northern Australia, Indonesia, and Malaysia during the November–March period and the Bay of Bengal and India during the monsoon season. However, it turned out that the stratospheric fountain hypothesis is not necessary since Mote et al. (1996) and Dessler (1998) showed that the temporal and spatial restrictions, respectively, do not hold. Hence, air can enter the stratosphere at any time and any longitude in the tropics rather than at preferred locations and times. Zhou et al. (2001) examined an extended version of the radiosonde dataset used by Dessler (1998) and found a cooling trend in the mean tropical tropopause temperature and a long-term increase in stratospheric water vapor. This long-term increase is significantly larger than the one expected from oxidation of increased methane (Gettelman et al. 2000), indicating that the stratospheric fountain might still be necessary. Fueglistaler and Haynes (2005) found no cutting-edge evidence in 40 yr ECMWF Re-Analyses (ERA-40) that explains the long-term increase in stratospheric water vapor. This suggests that dynamical pathways and the tropopause temperature might not be the only important factors controlling water vapor in the LS. Hence, it is difficult to account for the extreme dryness of the stratosphere unless tropical regions determine the dehydration of the air entering the stratosphere. The role of tropical convection in UT/LS exchanges is not well understood but deep convection eventually provides an alternative pathway for regulating the water content in the UT/LS region.

The regulation of the water content in the UT/LS by deep convection results from a combination of several processes. These processes are briefly summarized in the following. A large-scale ascent from the UT to the LS is expected above the LZH, where the clear-sky mass flux is upward (e.g., Highwood and Hoskins 1998). This process does not require convection to overshoot the tropopause. Horizontal transport through the cold trap regions then causes air reaching the tropopause elsewhere to be dehydrated to the low saturation mixing ratios characteristic of the tropical tropopause (Holton and Gettelman 2001). Overshooting deep cumulus clouds do occasionally penetrate the tropopause up to the cold point (Liu and Zipser 2005). These events do not support a permanent vertical transport, but a temporally and spatially localized one. Since deep convective structures overshooting their level of neutral buoyancy are colder than their environment, air can be dehydrated rapidly (e.g., Danielsen 1982; Kley et al. 1982; Danielsen 1993; Sherwood and Dessler 2000, 2001). As the air rises in deep convective turrets it is cooled, and water vapor is frozen out. Most of the ice particles are removed by sedimentation, and thus dehydrate the air. However, if ice particles do not settle out and remain in the LS the air is moistened as ice particles evaporate (Vömel et al. 1995; Grosvenor et al. 2007). There are other processes such as optically thin cirrus (Jensen et al. 2001) and internal gravity waves (Potter and Holton 1995; Teitelbaum et al. 2000), which may contribute to the dehydration of the LS. The relative importance of these processes in regulating the water content in the UT/LS region is still uncertain (e.g., Holton et al. 1995). Gettelman and Birner (2007) argued, based on the results of global-scale simulations, that large-scale processes, and the large-scale effects of small-scale processes, are the ones that are globally controlling the variability of the TTL. While the details of small-scale processes in these simulations may not be important for the climatology of the TTL, the authors indicated that further investigation is needed to understand the role of cloud processes and whether models are providing a correct picture of the TTL for the right reasons.

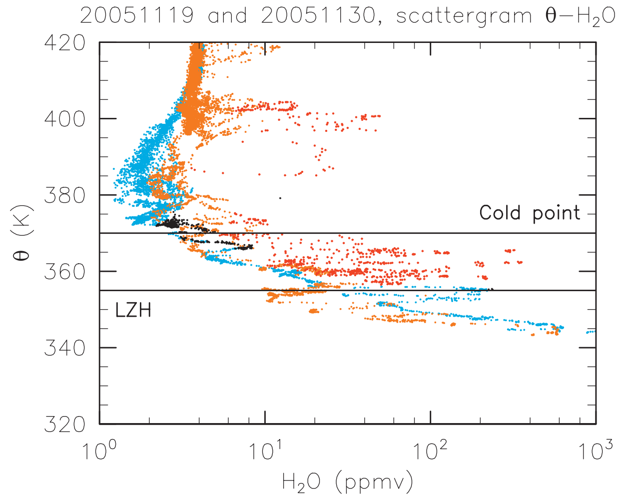

FIG. 9. Scattergram θ–H_2O using observational data from the FISH hygrometer on board the M55-Geophysica research aircraft on 19 (blue and black) and 30 (orange and red) Nov 2005 (see text for flight times and Figs. 6a,b for the flight track from 1500 to 1700 LT 30 Nov 2005). Black and red colors are used to emphasize points for which total water RH_{ice} is greater than 150%, clearly indicating the presence of ice particles.

b. *Total water observations above Hector*

Total water (water vapor and ice) was measured by a fast in situ stratospheric hygrometer (FISH; Zöger et al. 1999) on board of the M55-Geophysica research aircraft during the ACTIVE/SCOUT-O3 field campaign. The FISH measurements provided invaluable information on the vertical distribution of total water above and around Hector. Unfortunately, the aircraft cannot fly into the updrafts and convective overshoots. Hence, the measurements have to be considered as representative of an overall signature of Hector, for which the most extreme values are discarded. Figure 9 shows two vertical distributions of total water obtained from flight tracks from 1255 to 1712 LT and from 1300 to 1721 LT 19 and 30 November 2005, respectively. The flight track from 1500 to 1700 LT 30 November 2005 is displayed in Figs. 6a,b. Note that the virtual potential temperature θ is used as the vertical coordinate.

Hector did not develop on 19 November, so that the corresponding vertical distribution may be regarded as a typical background profile. The mean water vapor mixing ratio at 380 K is about 2.5 ppmv, which value is close to the one found by Kelly et al. (1993) over Darwin during the STEP tropical mission in 1987 (values averaged 2.4 ppmv at 375 K). The imprint of the Hector event on 30 November onto the water content in the UT/LS region appears clearly when superimposed on the background profile. Within the TTL, total water mixing ratios as high as 100–200 ppmv were observed. The region of minimum water vapor (often referred to as hygropause) lies well above the TTL at around 380–390 K. The elevated hygropause may be regarded as a manifestation of the seasonal cycle in tropopause temperature (Mote et al. 1996). Teitelbaum et al. (2000) provided an alternative explanation based on an uplift of the tropopause during an intense convection event and a reestablishment of the temperature minimum at a lower altitude after the event while the hygropause remains elevated. Considerable amounts of ice are present above the hygropause, especially around 400 and 420 K with values as large as 10–40 ppmv (Corti et al. 2008), for which total water RH_{ice} is greater than 150% (see Fig. 9). This ice must have been dropped off over there as a result of overshooting convection as we show in section 5c. The presence of ice may lead to a dehydration or moistening of the LS via sedimentation or sublimation, respectively. There is observational evidence that the water vapor was subsaturated with respect to ice in the LS on that day (Corti et al. 2008) and therefore the presence of ice would lead to a moistening of the LS. The size distribution of ice particles and the water vapor mixing ratio are key to determining the potential for either dehydration or moistening. Smaller particles fall out more slowly than larger particles, allowing more ice evaporation while mixing with the surrounding air and eventually a moistening (or rehydration) of the LS. Larger particles sediment out more rapidly, allowing ice to be removed and the air to become dehydrated. A major question that we cannot answer at that time (from the measurements presented herein) is the relationship between the overshooting air and the amount of small ice particles.

c. Simulated total water above Hector

There have been few high-resolution numerical studies on the role of overshooting tropical convection in regulating the water content in the UT/LS region. Küpper et al. (2004) investigated a case using a cloud-resolving model run to equilibrium. The results showed no dehydration and that the nonconvective fluxes of mass and water were predominant when compared with the corresponding convective fluxes. However, the model did not produce overshoots penetrating above the cold point, so that the effect of the most vigorous overshooting events was not represented. In a real case numerical study, Chaboureau et al. (2007) evidenced overshooting convection transporting mass and water vapor across the tropopause during the second Tropical Convection, Cirrus and Nitrogen Oxides (TROCCINOX; Schumann 2005) field campaign, which took place in Brazil in early 2005. The simulated deep convective event was found to induce an upward transport of water vapor of a few tons per second across the tropopause. Consistent with this study, Grosvenor et al. (2007) found a significant moistening (up to 0.26 ppmv) of the LS by deep convection using idealized cloud-resolving simulations over the same area in Brazil during the Impact of Tropical Convection on the Upper Troposphere and Lower Stratosphere at Global Scale experiment (HIBISCUS; Pommereau et al. 2007) IOP. In contrast to the cases studied by Chaboureau et al. (2007) and Grosvenor et al. (2007), Marécal et al. (2006) and Rivière et al. (2006) found a negligible impact of deep convection on the composition of the LS for an extreme event observed in February 2001 during the preparation campaign of the HIBISCUS IOP. Jensen et al. (2007) simulated overshooting convection in an idealized framework using a cloud-resolving model. The authors found no evidence that overshooting convection can dehydrate the TTL when it is initially ice subsaturated. Conversely, deep convection can dehydrate the LS if the TTL is initially supersaturated. Hence, the entry of water vapor in the LS depends on the TTL relative humidity (with respect to ice).

The time evolution of the vertical distribution of total water as simulated by ARW and UM averaged over the domains ARW3 and UM3, respectively, is shown in Fig. 10. Prior to Hector (see Figs. 10a,b), the vertical profile of total water is close to the background profile presented in Fig. 9. The altitude of the hygropause is lower than observed and lies at about 370 K nearby the cold point. As Hector develops, the total water profile becomes multivalued, which indicates spatial variability and nonuniform ascent. During the mature stage of the storm (see Figs. 10c,d), the total water mixing ratio exceeds a few tens of ppmv above the cold point. Total water RH_{ice} values greater than 150% above the cold point confirm that overshoots contain more than enough ice particles, which can either dehydrate or moisten the LS. As the preconvection water vapor in the LS was subsaturated with respect to ice (Corti et al. 2008), sublimation of the injected ice particles provides a source of water vapor. This condition is favorable for the air to moisten over the next few hours. The structure of the postconvection TTL after the dissipation stage is dramatically different in the ARW and UM simulations (see Figs. 10e–h). The TTL structure is not so much different during and after convection in the UM simulation. In the ARW simulation, the vertical distribution of total water reverts toward a distribution close to the preconvection one. The difference between the postconvection vertical distribution at 1830 LT (a couple of hours after Hector had formed) and that of the mature stage of Hector at

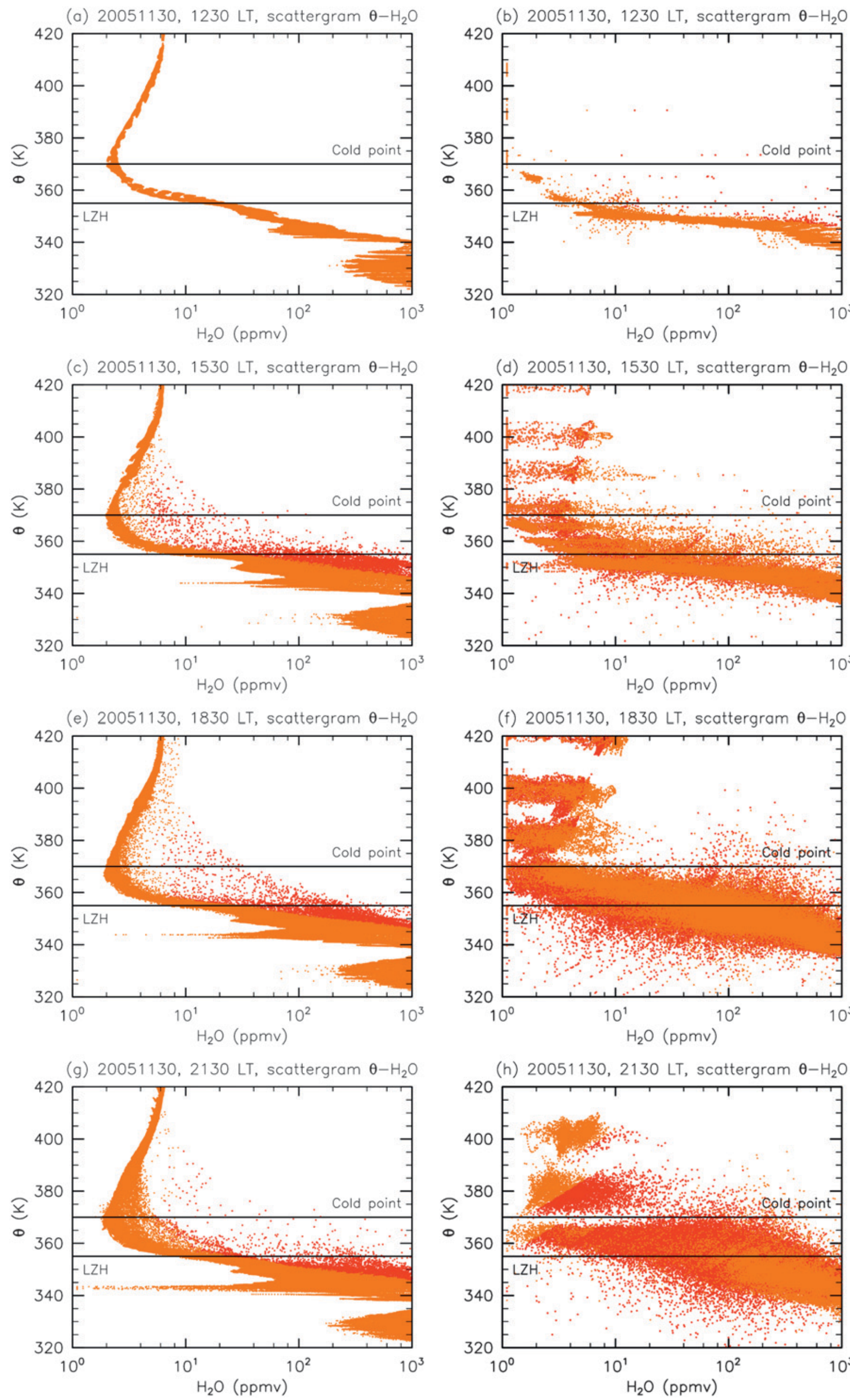

FIG. 10. Scattergrams θ–H_2O using results from the (left) ARW and (right) UM simulations in domains ARW3 and UM3, respectively, at (a),(b) 1230, (c),(d) 1530, (e),(f) 1830, and (g),(h) 2130 LT. The red color is used to emphasize points for which total water RH_{ice} is greater than 150%, clearly indicating the presence of ice particles.

1530 LT (see Fig. 11) suggests a moistening of the LS (above 380 K) up to 0.12 ppmv in the ARW simulation and up to 5.91 ppmv in the UM simulation. The moistening averages about 0.06 and 2.24 ppmv in the range 380–420 K for ARW and UM, respectively. The moistening derived from the UM simulation is much larger than that derived from the ARW simulation. The reason for this difference can be attributed to UM transporting more water vapor and ice in the LS than ARW (see Figs. 10c–f). Further research is necessary to assess the primary causes of this difference between the two models. Maximum moistening is predicted below the LZH. However, it requires that this moist air is lofted across the tropopause via large-scale upwelling to have an impact in the LS. If we start the ARW simulation on 29 November 2005 rather than on 28 November 2005, we found more moistening above 380 K (see Fig. 11). This is consistent with the idea of the water vapor mixing ratio being driven to saturation more rapidly in the LS as Hector puts water one day after another. This buildup is likely to be an important factor when quantifying the water budget in the LS.

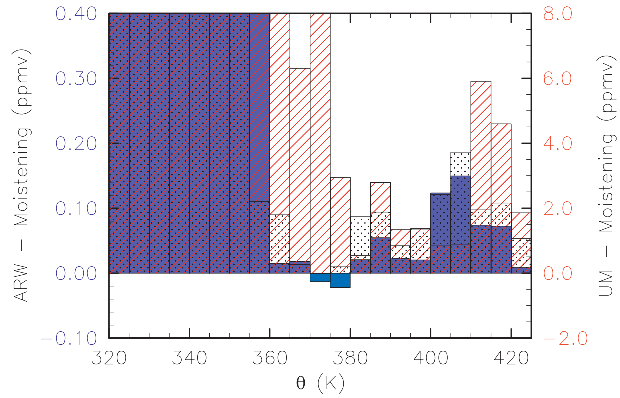

FIG. 11. Moistening computed using results from ARW and UM over the domains ARW3 and UM3, respectively, as the difference between the mean water vapor mixing ratios within 5-K virtual potential temperature bins at 1830 LT (a couple of hours after Hector had formed) and at 1530 LT (during the mature stage of Hector). The blue shaded bars correspond to the ARW simulation starting at 0930 LT 28 Nov 2005, while the stipple-filled bars correspond to the one starting at 0930 LT 29 Nov 2005. The red hatched bars correspond to the UM simulation. Note that the two models have their own separate "*y* axis."

6. Summary and conclusions

The development of Hector on 30 November 2005 was simulated using the Advanced Research Weather Research and Forecasting and the Met Office Unified Model cloud-resolving models. The simulations were performed with nests down to a horizontal resolution of 1 km. At this resolution, the timing, structure, and strength of deep convection were reproduced fairly well when compared with field campaign observations. The timing of this episode can be broken down into four main phases: onset (~1100–1300 LT), triggering (~1300–1500 LT), Hector (~1500–1700 LT), and dissipation (~1700–1900 LT). These stages were described in great detail in section 4. We found that refining the horizontal resolution to 250 m with ARW leads to a better representation of convection with respect to rainfall. Horizontal convective rolls are produced at this horizontal resolution, while being not recognizable using a horizontal resolution of 1 km. The lifetime of Hector and its maximum height did not change much from the 1-km to 250-m horizontal resolution simulations, so that we might argue that a 1-km horizontal resolution is fine enough to simulate faithfully this Hector thunderstorm.

From our results, we clearly evidenced the intermittent penetration of overshooting convection through the isentropic barrier (~355 K), which lead to troposphere–stratosphere exchanges. In particular, overshooting convection does affect the entry of water vapor in the LS. From a mechanistic point of view, tropospheric air crosses the tropopause by convective overshooting and then slowly rises as a result of radiative heating. This view is consistent with the parcel mixing model proposed by Sherwood and Dessler (2000, 2001). As stressed by Folkins et al. (2002), the potential for dehydration or moistening of the LS do exist provided that layers below overshooting convection detrain air within the TTL. In addition, only air detraining above the LZH is expected to affect significantly the water content in the UT/LS region. Convective ice lofting was found to be crucial in explaining the water content in the LS. Note that it was also found in previous studies to be essential in explaining the isotopic composition of the stratosphere (e.g., Keith 2000; Dessler et al. 2007). As the preconvection water vapor was ice subsaturated in the LS, we found a moistening rather than a dehydration of the LS. The moistening was found fairly significant above 380 K and averaged about 0.06 ppmv in the range of 380–420 K for ARW. As for UM, the moistening was found to be much larger (about 2.24 ppmv in the range of 380–420 K) than for ARW.

The question as to whether the rather infrequent overshoots within and above the TTL are sufficient to significantly affect the composition of the LS has not been addressed in the present study. This challenge requires a multiscale approach from the global scale down to the cloud scale. Convective transport into the LS is certainly small compared with transport induced by larger-scale systems. Nevertheless, its efficiency in quickly transporting boundary layer air makes it key to injecting short-lived species of tropospheric source (e.g., Dickerson et al. 1987) such as radon (e.g., Kritz et al. 1993) and some halogen compounds. More work needs to be done to quantify the relative importance of gradual versus episodic transport through the tropopause. Performing high-resolution simulations over a large domain of a few tens of thousands of square kilometers would be especially valuable for that purpose. We have left this point for further research.

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