The South-Central U.S. Flood of May 2010: Present and Future*

GARY M. LACKMANN

Department of Marine, Earth, and Atmospheric Sciences, North Carolina State University, Raleigh, North Carolina

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ABSTRACT

Previous studies have documented a feedback mechanism involving the cyclonic low-level jet (LLJ), poleward moisture flux and flux convergence, and condensational heating. Increased water vapor content and potentially heavier precipitation accompanying climate warming suggest the hypothesis that this feedback could strengthen with warming, contributing to amplification of precipitation extremes beyond what the thermodynamically controlled vapor increase would provide. Here, this hypothesis is tested with numerical simulations of a severe flooding event that took place in early May 2010 in the south-central United States.

Control simulations with a mesoscale model capture the main features of the May 2010 flooding event. A pseudo-global warming approach is used to modify the current initial, surface, and boundary conditions by applying thermodynamic changes projected by an ensemble of GCMs for the A2 emission scenario. The observed synoptic pattern of the flooding event is replicated but with modified future thermodynamics, allowing isolation of thermodynamic changes on the moisture feedback. This comparison does not indicate a strengthening of the LLJ in the future simulation. Analysis of the lower-tropospheric potential vorticity evolution reveals that the southern portion of the LLJ over the Gulf of Mexico in this event was strengthened through processes involving the terrain of the Mexican Plateau; this aspect is largely insensitive to climate change. Despite the lack of LLJ strengthening, precipitation in the future simulation increased at a super Clausius–Clapeyron rate because of strengthened convective updrafts.

1. Introduction

Anthropogenic climate change has the potential to increase the severity of both droughts and floods because of changes in the hydrologic cycle (e.g., Trenberth 1999; Allen and Ingram 2002; Held and Soden 2006). Increased atmospheric water vapor is a straightforward thermodynamic consequence of warming, and as expected, general circulation models (GCMs) consistently indicate a robust vapor increase in future projections of a warmer climate. This increase is consistent with the Clausius–Clapeyron equation, which yields roughly a ~7% specific humidity increase per degree Celsius of warming for typical lower-tropospheric temperatures

(e.g., Mitchell and Ingram 1992; Pall et al. 2007). Enhanced vapor suggests a likely precipitation increase for all weather systems (Trenberth 1999), although overall precipitation increases are expected to grow at a smaller rate than vapor increases via energy balance arguments (Allen and Ingram 2002; Wentz et al. 2007; Stephens and Ellis 2008; Rondanelli and Lindzen 2008; Stephens and Hu 2010; Posselt et al. 2012). An overall tendency for increased frequency of heavy rainfall has been documented in both numerical and observational studies (e.g., Karl and Knight 1998; Semenov and Bengtsson 2002; Groisman et al. 2005; Emori and Brown 2005; Lenderink and van Meijgaard 2010; Shaw et al. 2011).

The increase in lower-tropospheric water vapor that accompanies climate warming can lead to precipitation increases in many types of precipitating weather systems (e.g., Trenberth 1999). Larger water vapor content results in strengthened upward vapor flux even with the same upward vertical motion (the thermodynamic effect). Emori and Brown (2005) separate precipitation changes into a thermodynamic contribution due to vapor increase and dynamical contributions due to changes in atmospheric motion. Several studies have shown the dynamical

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Corresponding author address: Gary M. Lackmann, Department of Marine, Earth, and Atmospheric Sciences, North Carolina State University, 1125 Jordan Hall, Box 8208, Raleigh, NC 27695.
E-mail: gary@ncsu.edu

a.) Current

b.) Future

FIG. 1. Schematic diagram of hypothesized changes in an idealized cold-frontal zone in (a) current and (b) future conditions. (left) Plan projections showing region of cloud and precipitation (green shading), cold front, and lower-tropospheric wind vectors. (right) Corresponding cross section; location and strength of diabatically produced cyclonic PV features added.

component to be smaller than the thermodynamic contribution (e.g., Emori and Brown 2005; Chou et al. 2012). However, there are several distinct dynamical mechanisms that could operate in warming environments. For convective storms, enhanced CAPE that is expected to accompany climate warming (e.g., Trapp et al. 2007) should lead to strengthened convective updrafts. Synoptic-dynamic research has established the importance of condensational heating to the dynamics of extratropical cyclones and frontal systems; using potential vorticity inversion, this role has been quantified (e.g., Davis and Emanuel 1991; Davis 1992; Davis et al. 1993; Stoelinga 1996). The pre-cold-frontal low-level jet (LLJ), often located at the western boundary of the cyclonic warm sector, accompanies a warm, moist airstream referred to as the warm conveyor belt (e.g., Harrold 1973; Browning and Pardoe 1973; Browning 1990). More recently, these features have been dubbed atmospheric rivers (e.g., Newell et al. 1992; Zhu and Newell 1994, 1998; Ralph et al. 2004), and their role in heavy precipitation has been studied extensively (e.g., Eckhardt et al. 2004; Field and Wood, 2007; Knippertz and Martin 2007; Knippertz and Wernli 2010; Berry et al. 2011; Boutle et al. 2011). The LLJ can be strengthened by condensational heating and lower-tropospheric cyclonic potential vorticity (PV)

generation in cold-frontal precipitation to the west (e.g., Lackmann and Gyakum 1999; Lackmann 2002; Brennan et al. 2008), suggesting a positive feedback between condensational heating and enhanced poleward moisture transport in the LLJ. Given the role of condensation in this scenario, increased future water vapor content associated with climate warming could strengthen this feedback mechanism (Fig. 1). The strengthening of the LLJ in response to warming would lead to precipitation increases beyond the thermodynamic effect, due to increased horizontal moisture flux and flux convergence. A stronger LLJ could also potentially increase the threat of severe convective storms due to increased lower-tropospheric wind shear. From a climate dynamics standpoint, such a change could alter transient meridional transports in the midlatitude storm tracks (Held and Soden 2006). Changes to upper-tropospheric jets could also result from increased heating, but here the focus is restricted to the lower troposphere.

The question of how enhanced vapor and condensation might alter mesoscale dynamics during heavy precipitation has received somewhat limited attention in the literature and is the focus of the current paper. The hypothesis summarized in Fig. 1 is tested here using the “pseudo-global warming” approach introduced by Schär

FIG. 2. Summary information from May 2010 flooding event: (a) schematic summary from Moore et al. (2012, their Fig. 17b); (b) Storm Prediction Center storm reports for 1 May 2010; and (c) 72-h gauge-adjusted radar quantitative precipitation estimate for period 0000 UTC 30 Apr through 0000 UTC 2 May 2010 (graphic obtained online from <http://nmq.ou.edu/>); and (d) annotated GFS analysis showing 850-hPa wind barbs and water vapor mixing ratio (g kg^{-1} ; gray shading) and 850–700-hPa potential vorticity (PVU, $1 \text{ PVU} = 10^{-6} \text{ m}^2 \text{ K kg}^{-1} \text{ s}^{-1}$; blue shading every 0.25 PVU above 0.5 PVU) for 0000 UTC 3 May 2010.

et al. (1996) and Frei et al. (1998) to compare present-day simulations of an extreme flooding event to a future replication of the same event, but with modified thermodynamics consistent with climate warming. Simulations of much higher resolution than GCMs can provide (6-km grid length or less) are necessary to allow analysis of changes due to moist mesoscale processes. It is recognized that many events must be examined in order to adequately test this hypothesis, but here we begin by analyzing an exceptional flooding event that took place in May 2010.

Severe flooding occurred in early May 2010 in parts of the south-central United States; central and western portions of the state of Tennessee were particularly hard hit. Storm-total damage estimates exceeded \$2 billion (NWS 2011) and, in addition to flooding, there were numerous reports of severe weather accompanying a series of convective storms during the event (Figs. 2a,b). A slowly evolving synoptic pattern during this event

featured an upper-level ridge (trough) over eastern (western) North America and a persistent lower-tropospheric southerly flow of anomalously warm and humid air from the Yucatan Peninsula and Gulf of Mexico into the vicinity of a quasi-stationary frontal boundary over the flooding region (Fig. 2a; Moore et al. 2012). The synoptic environment was conducive to slow mesoscale convective system (MCS) motion, with back building and echo training leading to extensive flooding in the Nashville, Tennessee, metropolitan area (Moore et al. 2012). The 3-day gauge-adjusted radar precipitation estimates for the period from 0000 UTC 1 May through 0000 UTC 4 May 2010 exceeded 250 mm over a large area of western Tennessee and parts of Kentucky, with maximum values greater than 500 mm (Fig. 2c). Two quasi-stationary MCSs were responsible for these extraordinary precipitation totals (Moore et al. 2012; Durkee et al. 2012); the characteristic back building

and echo training observed during this event is consistent with the findings of previous studies of flash-flooding events by Maddox et al. (1979), Doswell et al. (1996), and Schumacher and Johnson (2006).

Low-resolution analyses of the lower-tropospheric moisture, PV, and wind field suggest that diabatically generated PV maxima lie to the west of a southerly LLJ extending from the Gulf of Mexico into the region of heavy precipitation (Fig. 2d). Additional calculations are needed in order to determine if this elongated series of PV maxima are due to condensational heating, but the synoptic pattern appears consistent with the type of moisture-transport feedback schematically outlined in Fig. 1.

The objective of this study is to answer the question: If the synoptic pattern accompanying the May 2010 flooding event were to repeat 100 years in the future, would large-scale thermodynamic changes result in a dynamically driven precipitation increase beyond what the thermodynamic change alone would provide? Changes in the strength of the LLJ and convective updrafts between two high-resolution numerical simulations will be examined to address this question.

2. Methods

a. Numerical model experimental design and present-day simulation

The Advanced Research Weather Research and Forecasting (WRF) model (ARW; Skamarock et al. 2008) version 3.2.1 is used for mesoscale simulations in this study. To resolve the mesoscale evolution of this event, the model domain features a larger outer domain with 54-km grid spacing, along with 18- and 6-km nested domains (Fig. 3a). Additional simulations with a 2-km nest were performed (not shown) but, due to computational constraints, the size of this domain was limited and results will be therefore primarily shown for the 6-km domain. The model was run with 28 vertical levels and a model top of 50 hPa.

The Betts–Miller–Janjic convective parameterization (CP) scheme was utilized on 54- and 18-km domains, but CP was omitted on the 6- and 2-km domains. One could argue that 6 km is too coarse a grid length at which to omit CP (e.g., Weisman et al. 1997; Bryan et al. 2003), but comparison to 2-km simulations and observed radar (Fig. 4) demonstrate that the organized convective storms that characterized this event were represented adequately for our purposes on the 6-km grid. For the simulations presented here, all domains employed the WRF single-moment 6-class (WSM6) microphysics scheme, the Yonsei University (YSU) PBL scheme, and the National Oceanic and Atmospheric Administration (NOAA) land

surface model. Longwave radiation was handled with the Rapid Radiative Transfer Model (RRTM) scheme, and shortwave radiation was handled by the Dudhia scheme.

Initial and lateral boundary condition data were based on Global Forecast System (GFS; Sela 1980; Kleist et al. 2008) final analyses on a 1° latitude–longitude grid; simulations were initialized at 0000 UTC 30 April 2010 and run for 96 h to 0000 UTC 4 May 2010. For nested domains, lateral boundary conditions are provided by the parent domain and one-way nesting is used.

b. Future simulations

There is no simple way to anticipate how changes in future synoptic patterns might influence the frequency or severity of extreme precipitation events. The experimental design employed here addresses this issue by replicating a recently observed synoptic pattern but with future large-scale thermodynamic changes imposed. While it is acknowledged that an identical synoptic pattern would not occur in the future, it is reasonable to assume that a similar pattern could occur. Because the model simulation is allowed to evolve dynamically for the duration of the synoptic event, the resulting changes, brought about by imposing thermodynamic change, can include dynamical changes as well. Two types of possible dynamical change to be examined here are the hypothesized future strengthening of the LLJ and stronger future convective updrafts. Several previous studies have utilized similar methods; the current experimental design is similar to the pseudo–global warming approach of Schär et al. (1996) and Frei et al. (1998) and more recently by Sato et al. (2007), Hara et al. (2008), and Kawase et al. (2009), among others. This is a conservative approach, in that the multiple averaging periods result in a thermodynamic change signal that is smoothly varying and robust across the GCMs. Future extreme events could exhibit thermodynamic changes that deviate strongly from the average change. Nevertheless, to test the hypothesis that a diabatically driven moisture-transport feedback would amplify with climate warming, this experimental design serves our purposes well.

Quantification of projected thermodynamic change due to increased anthropogenic greenhouse gases is accomplished using an ensemble of Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4) GCM simulations. Experiments are conducted using both the A1B and A2 emission scenarios (Solomon 2007), with only the A2 simulations presented here because of overall similarity of the A1B and A2 simulations. Monthly averaged temperature and mixing ratio values are available on isobaric levels and at the 2-m level; monthly averaged sea surface temperature data are

FIG. 3. (a) WRF model domains used in current and future simulations; (b) north–south cross section of applied temperature change ($^{\circ}\text{C}$) derived from IPCC AR4 GCM ensemble for decadal May averages using (2090–99) – (1990–99). (c),(d) As in (b), but for 700 hPa in (c) and 200 hPa in (d).

also available. The May monthly fields were averaged over a subset of five GCMs for which reliable temperature data were available at all vertical levels [Bjerknes Centre for Climate Research Bergen Climate Model version 2 (BCCR BCM2), Centre National de Recherches Météorologiques Coupled Global Climate Model version 3 (CNRM-CM3), Institute of Numerical Mathematics Coupled Model version 3 (INM-CM3.0), Max Planck Institute (MPI) ECHAM5, and third climate configuration of the Met Office Unified Model (HadCM3)]. A decadal average temperature change was then computed from the May monthly averages from the GCM ensemble for the periods 1990–99 and 2090–99. Temperatures at the 2-m level, sea surface, and all available isobaric levels were averaged. These spatially varying average fields were interpolated to the same 1.0° GFS grid as for the initial and

lateral boundary conditions used in the control simulation. Differences between the 2090s and 1990s were computed for each grid cell, and these changes were subsequently added to the original GFS analyses. Given that the flooding event in question took place in May 2010 and that we are adding a change that was calculated from a 100-yr difference, the future simulation can be interpreted as loosely representing thermodynamic conditions around the year 2110 that would arise from anthropogenic greenhouse gas forcing alone in the A2 scenario. The same synoptic weather pattern from May 2010 also characterizes the future simulation, and this fictitious date does not correspond to a specific date in future GCM projections. The triple averaging process (decadal and ensemble averages of May monthly means) provides a smoothly varying large-scale thermodynamic

FIG. 4. NEXRAD observed composite radar reflectivity (dBZ; shaded) valid (a) 0000 UTC 1 May, (c) 0000 UTC 2 May, (e) 1200 UTC 2 May, and (g) 0000 UTC 3 May. Model-simulated radar at corresponding times: (b) 24, (d) 48, (f) 60, and (h) 72 h. Radar reflectivities <20 dBZ are omitted because of excessive coverage and ground clutter at times.

FIG. 6. Summary of 900–800-hPa potential vorticity (PVU; shaded as in legend at bottom of panel) and sea level pressure (hPa): (a) 6, (b) 18, (c) 30, (d) 42, (e) 54, and (f) 66 h for 6-km domain of current simulation.

FIG. 7. Summary of lower-tropospheric water vapor and LLJ, including 900–800-hPa potential vorticity (PVU; red contours every 0.5 PVU starting at 0.5 PVU), 850-hPa water vapor mixing ratio (g kg^{-1} ; shaded as in legend at bottom of panel), and 850-hPa wind barbs (barbs omitted $< 15 \text{ m s}^{-1}$): (a) 6, (b) 18, (c) 30, (d) 42, (e) 54, and (f) 66 h for 6-km domain of current simulation.

FIG. 8. Geostationary Operational Environmental Satellite-13 (GOES-13) enhanced infrared satellite imagery for (a) 0615 and (b) 1815 UTC 1 May 2010. Infrared brightness temperature shaded as in legends at left of panels.

diabatic PV tendency, which is related to the projection of the heating gradient onto the absolute vorticity vector. This term also indicates that the strength of the diabatic PV generation is proportional to the magnitude of the absolute vorticity and to the PV itself; this means that diabatic PV changes are potentially stronger for a given heating gradient in regions of larger background vorticity or PV.

3. Present-day simulation

A comparison of simulated composite reflectivity from the current-day WRF simulation to the actual composite

radar mosaic demonstrates that the simulation captures the general character of the precipitation event in the region of interest (Fig. 4). At simulation hours 24 (Figs. 4a,b) and 60 (Figs. 4e,f), the simulated reflectivity elements exhibit strong similarity to the observed radar across the northern portion of the model domain. This result is perhaps fortuitous considering that the lateral boundary of the outermost 54-km domain is the only analysis information constraining the simulation during the 96-h integration period. At times, the simulated precipitation over the southern portions of Louisiana and Mississippi is heavier than observed (Figs. 4e–h). Details of the

FIG. 9. As in Fig. 6d, but for no-terrain simulation.

model-simulated MCSs differ from observations, but the general location and orientation of the simulated convection matches the observations reasonably well throughout the model integration. The primary goal is not to capture the MCS details but to compare current and future versions of the event.

a. Precipitation

The gauge-adjusted radar-derived quantitative precipitation estimate (QPE) for the 72-h period ending 0000 UTC 2 May reveals a large area of greater than 250 mm accumulation, with peak values in excess of 500 mm over western Tennessee (Fig. 2c). Although the corresponding WRF-simulated precipitation exhibited a similar maximum value, the axis of heaviest precipitation is offset southward from that observed (Fig. 5). Model precipitation was heavier than observed across much of Louisiana and parts of southern Mississippi and was too light across a portion of Kentucky. Despite these differences, the magnitude of maximum precipitation and the areal coverage of the 250-mm contour are sufficiently similar between observation and simulation to warrant further comparisons with a thermodynamically modified simulation representing a future thermodynamic regime.

b. Low-level jet and lower-tropospheric PV

The lower-tropospheric PV field exhibits the characteristic signature of cyclonic maxima to the west of the

southerly LLJ. Animations of the lower-tropospheric PV field (see supplemental material) reveal the spontaneous appearance of cyclonic PV in the lower troposphere over the eastern slopes of the Mexican Plateau during the overnight hours on each of the first 3 days of the simulation (Figs. 6a,c,e); during the afternoon hours, these PV maxima subsequently drift eastward (Figs. 6b,d,f). As the cyclonic PV maxima generated near the elevated terrain (referred to as topographic PV) drift north and east, they appear to merge with condensationally generated PV, consistent with the development of strong lower-tropospheric southerly and southwesterly winds over the south-central United States (Fig. 7). The strongest portions of the LLJ are found immediately to the east and southeast of the condensationally strengthened cyclonic PV maxima over the southern United States.

Little or no condensational heating is taking place in or near the Mexican Plateau cyclonic PV generation region; satellite imagery (Fig. 8) indicates clear skies in this area throughout the event. This suggests that the cyclonic PV maxima developing there are due to radiational or frictional processes associated with the elevated terrain. The generation of PV in the vicinity of topography has been examined in numerous studies, many of which implicate frictional processes (e.g., Aebischer and Schär 1998). Eastward-propagating orographically generated PV maxima have been analyzed by Li and Smith (2010). They document midlevel cyclonic PV maxima and propose that daytime heating leads to the

FIG. 10. Comparison of control (current) and future A2 simulation of simulated composite reflectivity (shaded) and sea level pressure (contour interval 2 hPa) for a subset of the 6-km WRF domain: present-day simulation hours (a) 24, valid 0000 UTC 1 May; (c) 48, valid 0000 UTC 2 May; (e) 60, valid 1200 UTC 2 May; and (g) 72, valid 0000 UTC 2 May. (b),(d),(f),(h) As at left, but corresponding to the simulation hours of the present-day simulation but modified with thermodynamic changes from the A2 climate change scenario.

FIG. 11. As in Fig. 7, but for future A2 simulation.

PV generation, which differs from the present case that features nocturnal PV growth. Detailed investigation of the exact mechanism of the topographic PV generation is tangential to the current objective. However, it is

useful to clarify the role of terrain in this event because the topographic PV appears to contribute directly to the strength of southerly flow and moisture transport in the lower troposphere over the Gulf of Mexico. By increasing

FIG. 12. Difference field (future – current; shaded) for 24-h time-averaged 850-hPa wind speed (m s^{-1}) for hours 48–72. Vector subtraction (m s^{-1}) is shown as black arrows.

the background PV, the first right-hand term of (3) may be rendered more effective for a given heating gradient.

To isolate the contribution of terrain to the LLJ and precipitation in this event, an additional simulation was undertaken with the terrain removed. The results feature a surprisingly prominent role for terrain in this case; without terrain, the lower-tropospheric cyclonic PV maximum is largely absent, and much weaker southerly geostrophic flow is implied by sea level isobars over the Gulf of Mexico (Figs. 6d, 9). Precipitation in the no-terrain simulation is drastically reduced, with maximum values less than one-third of that in the control simulation for the outermost domain (not shown). Additional analysis of this aspect is beyond the scope of this paper. However, if the nocturnal PV generation in the vicinity of the terrain was the result of radiational processes, this would be only weakly sensitive to climate change (due to

slightly reduced radiational cooling commensurate with greenhouse gas increases). Because of the important role of terrain in upstream PV generation in this case, the climate response of the LLJ in this event may differ from cases with a more purely condensational enhancement of the LLJ.

4. Comparison of current and future simulations

Thermodynamic changes computed from a small ensemble of IPCC GCM output for the A2 emission scenario (Solomon 2007) were applied to the initial and lateral boundary conditions for this event, as discussed in section 2 (Fig. 3b). The resulting simulation is a replication of a highly similar synoptic pattern to that in the control simulation, but with a projected future thermodynamic environment. Because of increased water vapor

FIG. 13. Averaging areas used for (a) 54-km grid length domain (domain 1) and (b) 6-km grid length domain (domain 3).

content, strengthened condensational heating is hypothesized to lead to a stronger future LLJ and enhanced water vapor transport, allowing a nonlinear increase in accumulated precipitation beyond what the background vapor increase alone would provide.

Inspection of simulated radar for the two model runs (Fig. 10) indicates a modest southward shift in the axis of heavy precipitation in the future. Simulated radar reflectivity values in convective cells also appear to be stronger in the future, consistent with increased CAPE. The southward shift is also evident in the location of the LLJ axis and lower-tropospheric cyclonic PV maxima, as well as in the storm-total precipitation fields. Examination of upper-tropospheric fields (not shown) indicates that the upper-level trough to the west is slightly more progressive in the future simulation, consistent with an increase in the upper westerly flow mentioned in section 2b and with the southward precipitation shift.

The increased lower-tropospheric mixing ratio evident in the future simulation was essentially an imposed change (Figs. 7, 11). For the range of near-surface temperatures in question, vapor content increases roughly 7% per degree of warming under conditions of constant RH, so the lower-tropospheric warming of $\sim 3^{\circ}\text{C}$ suggests a background vapor increase of approximately 20%. Because these changes are applied only at initialization, at the lateral boundaries, and indirectly through warmed sea surface and soil temperature, the humidity in the interior of the model domain may not maintain the initial increase over the duration of the model simulation. However, the averaged lower-tropospheric specific humidity evolution reflects a sustained vapor increase (not shown).

Comparison of the LLJ strength from Figs. 7 and 11 is facilitated by time averaging and difference-field computations, which were performed for various

lower-tropospheric pressure levels and times. The change in time-averaged 850-hPa wind speed reveals that an increase in the strength of the LLJ is not evident (Fig. 12). In fact, the LLJ is, on average, weaker in the future simulation. Additional spatial averaging was performed for the regions shown in Fig. 13 to provide information about change profiles with height over these areas. Averages were computed for each of the domains, over the northern averaging area (bounded by 30° – 37°N and 95° – 82°W), and for the 24-h period from simulation hours 48 to 72 for the current and future simulations. These confirm that there is no systematic increase in the meridional wind component in the lower troposphere, contrary to our hypothesis (Fig. 14a). This result is not sensitive to shifts in averaging area, averaging time interval, or coarse- versus high-resolution model domains.

There is a considerable increase in the averaged latent heating profile in the future simulation (Fig. 14b), and this corresponds to an increase in nonadvective PV tendency in the troposphere, with enhanced cyclonic (anticyclonic) tendency below (above) the level of maximum heating (Fig. 14c). The change in lower-tropospheric PV, however, is small. Over the depth of the troposphere, the averaged PV profile exhibits a future increase that maximizes in the upper troposphere, with reduced PV above the 200-hPa level in the future simulation and an implied increase in the height of the dynamic tropopause (Fig. 14d). The future PV increase in the upper troposphere may seem surprising, given that an enhanced anticyclonic nonadvective PV tendency in the future is evident there in Fig. 14c. However, recall that the imposed GCM-derived temperature change profile features strongest warming aloft and thus a more stable lapse rate. The averaged temperature change over the northern region, taken as the difference between the averages from

FIG. 14. Temporally and spatially averaged domain-1 fields for northern area shown in Fig. 13a: (a) meridional wind component (m s^{-1}) for model simulation hours 48–72 for current (blue dashed) and future (red) simulations; (b) condensational heating (units 10^{-5} K s^{-1}); (c) nonadvective PV tendency ($10^{-5} \text{ PVU s}^{-1}$); (d) PV (PVU); and (e) temperature change (K; future – current).

current and future simulations, demonstrates that the PV increase (decrease) corresponds to regions of decreased (increased) temperature lapse rate (Figs. 14d,e). Computation of the background PV increase due to the decreased tropospheric lapse rate is sufficient to explain the PV changes seen in Fig. 14d. An increased tropopause height over this region is consistent with deeper convective clouds and also with changes to the upper-trough

dynamics and surface cyclogenesis; investigation of these aspects are beyond the scope of the current paper.

Despite the lack of strengthening of the LLJ, we would still expect precipitation to increase in the future simulation due to the enhanced water vapor content, as seen in a comparison of Figs. 11 and 7. The precipitation pattern is shifted southward in the future simulation, hampering direct comparison of rainfall totals (Figs. 15a,b).

FIG. 15. Difference field of 72-h total precipitation ending 0000 UTC 3 May (mm; shaded, future – current).

However, it is evident from Fig. 15c that the difference field couplet is not symmetric: the positive difference values are considerably larger than the negative values. The domain-3 maximum 72-h simulated precipitation in the current simulation is 511 mm, comparable to the radar-derived QPE maximum. The maximum in the future simulation is 688 mm, a 35% increase. Comparison of local maxima is prone to uncertainty due to small-scale convective activity, prompting computation of area-averaged 72-h precipitation for the averaging region shown in Fig. 13b. Average domain-3 precipitation over this region is 52.6 mm (75.4 mm) in the current (future) simulation, a 43% future increase.

Given that the strength of the LLJ did not increase, it is important to understand why the precipitation exhibited a larger percentage increase than did the water

vapor content, which exhibits an approximate 20% increase in the lower troposphere (Fig. 16a). One hypothesis is that greater instability owing to increased lower-tropospheric water vapor content (and larger CAPE) led to stronger convective updrafts, increasing the moisture flux convergence. Examination of temporally and spatially averaged CAPE from current and future simulations over the entire duration of the simulation indicates a pronounced increase from approximately 300 J kg^{-1} in the current simulation to over 560 J kg^{-1} in the future. Averaged upward pressure-coordinate vertical velocity during the 48–72-h time interval increases by up to 30% for the 6-km domain 3 (Fig. 16b). This analysis indicates that the increase in precipitation observed in the future simulation was due to a combination of more vigorous convective motions

FIG. 16. Domain-3 spatial and temporal averages for area bounded by 30° – 37°N and 95° – 82°W (Fig. 16b) for hours 48–72: (a) specific humidity (g kg^{-1}) and (b) upward omega ($10^{-1} \text{ Pa s}^{-1}$). Blue dashed (red solid) contour corresponds to current (future) simulation.

(a dynamical effect) in conjunction with increased water vapor content but was evidently not due to an increase in the strength of the horizontal moisture-transport feedback.

The model-simulated composite reflectivity can be used as a proxy for rainfall rate. To ascertain changes in this quantity, reflectivity histograms at 1-dBZ intervals were computed over the entire 96-h model simulations. Differences in the shape of the current and future reflectivity histograms indicate a tendency for higher frequency of heavier instantaneous rainfall in the future simulation (Figs. 17a,b). To better quantify this, a difference histogram reveals that the frequency of composite reflectivity values above ~ 20 dBZ increases in the future, with a sharp decrease in the frequency of lower reflectivity values (Fig. 17c). Additional difference histograms of pressure-coordinate vertical motion (Fig. 18a) and hourly precipitation (Fig. 18b) are consistent with Fig. 17 and indicate stronger future ascent in simulated convective grid cells. All histogram bins above $\sim 5 \text{ mm h}^{-1}$ exhibited an increase in the future simulation, with very strong frequency decreases in the 0 – 4 mm h^{-1} range. These results are consistent with earlier studies and hold societal implications in terms of flash-flooding potential.

5. Conclusions

Previous studies have indicated a positive feedback mechanism involving condensational heating and moisture transport in low-level jets. Climate warming and the attendant increase in lower-tropospheric water vapor are expected to result in greater precipitation, especially during heavy rainfall events (e.g., Trenberth et al. 2003; Chou et al. 2012). Future warming and moistening thus have the potential, due to increased condensational heating, to strengthen this moisture-transport

feedback. This possibility is investigated here using the south-central U.S. flooding event of May 2010 as a case study. High-resolution present-day and future simulations of this event were undertaken with the WRF model. Century-scale temperature and moisture changes were derived from decadal averages of May monthly means between the present (1990s) and future (2090s) from a five-member GCM ensemble, based on the A2 emissions scenario. These large-scale changes were applied to analyzed initial and lateral boundary conditions for the May 2010 flooding event, and an additional future simulation of the event was produced. This approach is similar to the pseudo-global warming method of Schär et al. (1996) and others.

The detailed synoptic and mesoscale analyses of this flooding event provided by Moore et al. (2012) and Durkee et al. (2012) eliminate the need for such analysis here, although the importance of Mexican terrain in this event is emphasized in the current analysis. Examination of the lower-tropospheric PV evolution reveals a nocturnal cyclonic PV generation mechanism in the vicinity of the eastern slopes of the Mexican Plateau. This PV generation took place in clear-sky conditions and appears to be radiatively driven, although additional analysis of this phenomenon is needed. These cyclonic PV maxima subsequently advect northeastward, helping to fuel the southerly LLJ and northward moisture transport over the Gulf of Mexico. An experimental simulation with terrain removed produced drastically reduced precipitation amounts and much weaker northward moisture transport over the Gulf of Mexico. Moore et al. (2012) also emphasize the geographically fixed nature of the LLJ, consistent with a topographic process. Because the topographic PV generation mechanism is likely to exhibit only weak dependence on climate change, the

FIG. 17. Histograms of domain-3 model-simulated composite reflectivity for entire 96-h simulation: (a) current simulation; (b) future simulation; and (c) difference histogram, future – current. Reflectivity bins are taken every 1 dBZ, and ordinate is multiplied by 10^4 .

LLJ in this event may not be optimal for testing the moisture-transport feedback mechanism proposed here. If a topographic contribution of this type is common during heavy rainfall events in this geographical location, it may imply less climate sensitivity for events of this type in this region, although additional research would be required in order to explore this speculation.

The present-day WRF simulation at 6-km grid length performed adequately, producing 3-day precipitation totals similar to those analyzed, albeit with a southward shift relative to analyses. The simulation of the May 2010 event with modified future thermodynamics exhibited a considerable increase in precipitation, with the maximum 72-h total increased by 35% and the

area-averaged value by 43%. These increases are greater than the 20% increase in water vapor content, qualifying as “super Clausius–Clapeyron” in the nomenclature of Lenderink and van Meijgaard (2010). The mechanism responsible for the increased precipitation is not a strengthened horizontal moisture-transport feedback because, although the horizontal moisture flux increases in the future simulation, the strength of the LLJ does not. The increased horizontal transport is entirely attributable to increased vapor content in these simulations. The amplified precipitation is due to stronger convective updrafts, consistent with an increase in CAPE in the future simulation. Difference histograms (Figs. 17, 18) demonstrate an increased future frequency of higher

FIG. 18. As in Fig. 17c, but for (a) pressure-coordinate vertical motion (mb s^{-1}) and (b) hourly precipitation rate (mm h^{-1}).

composite reflectivity, strong upward vertical motions, and heavier hourly rain rates in this event. Enhanced upward vertical motion leading to precipitation changes that exceed vapor increases is consistent with the findings of Emori and Brown (2005) and Sugiyama et al. (2010). These studies were based on coarser GCM simulations that utilize convective parameterization, whereas the current results were obtained with explicit convection with grid-scale convective updrafts.

Despite the failure of the hypothesis test concerning the LLJ, additional cases must be examined before the hypothesis can be rejected. Cases in which the LLJ was more condensationally driven, as opposed to the topographically driven LLJ in this case, may better serve to test this hypothesis. Even without the positive LLJ feedback, updraft-driven precipitation and rain rate increases in the future simulation should be cause for concern as a potential consequence of climate warming.

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