

1 **An investigation of how radiation may cause accelerated rates of**
2 **tropical cyclogenesis and diurnal cycles of convective activity**

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12 **Abstract**

13 Recent cloud-resolving numerical modeling results suggest that radiative forcing causes
14 accelerated rates of tropical cyclogenesis and early intensification. Furthermore, observational
15 studies of tropical cyclones have found that oscillations of the cloud canopy areal extent often
16 occur that are clearly related to the solar diurnal cycle. A theory is put forward to explain
17 these findings. The primary mechanism that seems responsible can be considered a
18 refinement of the mechanism proposed by Gray and Jacobson (1977) to explain diurnal
19 variations of oceanic tropical deep cumulus convection. It is hypothesized that differential
20 radiative cooling or heating between a relatively cloud-free environment and a developing
21 tropical disturbance generates circulations that can have very significant influences on
22 convective activity in the core of the system. It is further suggested that there are benefits to
23 understanding this mechanism by viewing it in terms of the lateral propagation of thermally
24 driven gravity wave circulations, also known as buoyancy bores. Numerical model
25 experiments indicate that mean environmental radiative cooling outside the cloud system is
26 playing an important role in causing a significant horizontal differential radiative forcing and
27 accelerating the rate of tropical cyclogenesis. As an expansive stratiform cloud layer forms
28 aloft within a developing system the mean low level radiative cooling is reduced while at mid
29 levels small warming occurs. During the daytime there is not a very large differential

1 radiative forcing between the environment and the cloud system, but at nighttime when there
2 is strong radiative clear sky cooling of the environment it becomes significant. Thermally
3 driven circulations develop, characterized by relatively weak subsidence in the environment
4 but much stronger upward motion in the cloud system. This upward motion leads to a cooling
5 tendency and increased relative humidity. The increased relative humidity at night appears to
6 be a major factor in enhancing convective activity thereby leading in the mean to an increased
7 rate of genesis. It is postulated that the increased upward motion and relative humidity that
8 occurs throughout a deep layer both aids in the triggering of convection, and in providing a
9 more favorable local environment at mid-levels for maintenance of buoyancy in convective
10 cells due to a reduction of the detrimental effects of dry air entrainment. Additionally, the
11 day/night modulations of the environmental radiative forcing appear to play a major role in
12 the diurnal cycles of convective activity in the cloud system. It is shown that the upward
13 velocity tendencies in the system core produced by the radiative forcing are extremely weak
14 when compared to those produced by latent heat release in convective towers, but
15 nevertheless over the course of a night they appear capable of significantly influencing
16 convective activity.

17

18 **1 Introduction**

19 Numerous studies utilizing IR satellite imagery have shown that there is a significant diurnal
20 cycle of cirrus cloud cover in tropical cyclones (e.g. Browner et al. 1977; Muramatsu 1983;
21 Lajoie and Butterworth 1984; Steranka et al. 1984; Kossin 2002). The maximum areal extent
22 of cloud canopies was found to occur in the early morning and the minimum in the early
23 evening. It has been generally thought that the cause is a diurnal oscillation in deep
24 convection near the storm center (Hobgood 1986). Browner et al. (1977) suggested that the
25 oscillation should also be associated with a diurnal cycle of rainfall. A recent study by Shu et
26 al. (2013) has confirmed this supposition showing that a significant diurnal variation of
27 rainfall occurs in Western North Pacific tropical cyclones.

28 Recent numerical modeling studies also suggest that radiation may increase the rate of
29 tropical cyclogenesis (Nicholls and Montgomery 2013, hereafter NM13; Melhauser and
30 Zhang 2014). Previous studies to these, although limited by relatively coarse grid resolution
31 and parameterization of moist convection, had already shown earlier intensification when a
32 simple longwave cooling scheme was included (Sundqvist, 1970; Hack and Schubert, 1980).

1 NM13 conducted idealized experiments of tropical cyclogenesis using the Regional
2 Atmospheric Modeling System (RAMS) developed at Colorado State University (Pielke et al.
3 1992; Cotton et al. 2003). The objective was to obtain a better understanding of two distinctly
4 different pathways to tropical cyclogenesis that occurred in the idealized numerical modeling
5 studies of Montgomery et al. (2006) and Nolan (2007). The latter two investigations
6 examined the transformation of a relatively weak initial vortex over a warm ocean surface
7 into a tropical cyclone using grid resolutions marginally capable of resolving convective
8 clouds. Interestingly, the results of the two studies were very different. Montgomery et al.
9 conducted simulations with RAMS and found a pathway occurring similar to that in a case
10 study of the formation of Hurricane Diana (1984) by Hendricks et al. (2004). Vortical Hot
11 Towers (VHT's) were the preferred coherent structures. The aggregate diabatic heating of the
12 VHTs provided an influx of low-level angular momentum causing low-level cyclonic winds
13 to increase. Eventually the low-level winds became stronger than those of the mid-level
14 vortex used to initialize the model and remained stronger or comparable to the winds aloft
15 until they reached tropical depression strength (considered by NM13 to be approximately 12
16 m s^{-1}). Development was characterized by a gradual decrease of the radius of maximum
17 winds. Vorticity gradually built in the center as the system-scale inflow produced increasing
18 cyclonic vorticity and as small-scale cyclonic vorticity anomalies were converged at low
19 levels and underwent aggregation. In contrast the simulations by Nolan (2007) with the
20 Weather Research and Forecasting (WRF) model showed a gradual strengthening of the
21 initial mid-level vortex followed by the sudden formation of a small surface concentrated
22 vortex near the center of the larger scale circulation, with a radius of maximum winds of only
23 a few kilometers. This vortex became the focus of a strengthening cyclonic circulation that
24 grew in size to form a small tropical cyclone. These differing results were perplexing since
25 the two pathways were distinctly different even though there were many similarities between
26 the models and the experimental designs.

27 Nicholls and Montgomery (2013), using a newer version of RAMS, were able to
28 demonstrate for the first time development along both pathways, depending on initial
29 conditions and model physics employed. There were some caveats to the development along
30 the second pathway. Instead of the initial mid-level vortex simply strengthening, the surface
31 cyclonic winds typically increased to begin with, similarly to the first pathway. It was only
32 when an extensive stratiform anvil had formed aloft that a second prominent mid-level vortex
33 developed. NM13 ran a set of experiments to investigate the two pathways described by

1 Montgomery et al. (2006) and Nolan (2007) in more detail, which were denoted pathway One
2 and pathway Two, respectively. They concluded that the ice phase was crucial for the
3 formation of a strong mid-level vortex and development along pathway Two. Environments
4 conducive to forming large quantities of ice aloft appeared to be more favorable for
5 development along pathway Two. Higher sea surface temperatures for instance appeared to
6 produce more intense and deeper convective cells, more ice aloft and therefore favored
7 evolution along the second pathway.

8 NM13 included simulations both with and without radiation and comparison of the
9 genesis rate for otherwise identical experiments reveal significant differences. Table 1 shows
10 results for four pairs of experiments that are exactly the same except for whether radiation is
11 included or not. The time that maximum azimuthally averaged tangential winds reach 12 m s^{-1}
12 is much quicker for the experiments with radiation included for each of the four pairs. The
13 subsequent time to go from winds of 12 m s^{-1} to tropical storm strength winds of 17.4 m s^{-1} is
14 also considerably faster for all cases. On the other hand, there is no systematic increase in the
15 later intensification rate from tropical storm strength to hurricane strength (33 m s^{-1}) when
16 radiation is included. NM13 also found a strong diurnal cycle of convective activity when
17 radiation was included.

18 Potential influences of radiation on oceanic tropical Mesoscale Convective Systems (MCSs)
19 and tropical cyclones include enhancing surface precipitation, causing diurnal cycles,
20 changing the rate of development and effecting structure and motion. There are three main
21 mechanisms that have been proposed for the role of radiation in these convective systems: (1)
22 Differential cooling between the weather system and its surrounding cloud-free region (Gray
23 and Jacobson 1977); (2) Large scale clear-sky environmental cooling (Dudhia 1989; Tao et al.
24 1996); (3) Changing thermal stratification due to cloud top and cloud base radiative forcing
25 (Webster and Stephens 1980; Hobgood 1986; Xu and Randall 1995).

26 The first mechanism was proposed by Gray and Jacobson (1977) who presented
27 observational evidence in support of the existence of a large diurnal cycle of oceanic, tropical
28 deep cumulus convection. They found that in many places, heavy rainfall is two to three times
29 greater in the morning than in the late afternoon and evening. They hypothesized that the clear
30 environment surrounding the weather system experiences significant deep radiative cooling
31 during the nighttime whereas the air within the cloud system experiences little radiative
32 forcing except near the cloud top where there is strong cooling. They propose that the

1 environmental cooling causes increased subsidence and low-level convergence into the cloud
2 cluster at night, which enhances convection during the morning. During the daytime the solar
3 heating of the environment offsets the longwave cooling to a large extent and environmental
4 subsidence and low-level convergence into the cloud cluster is reduced, thereby producing a
5 diurnal cycle of convective intensity. They emphasize that it is the larger low-level moisture
6 convergence that occurs in the morning that is responsible for the higher morning
7 precipitation rates. Several numerical modeling studies have investigated the potential of this
8 differential radiation forcing mechanism for causing diurnal cycles of convective intensity in
9 MCSs and conclude that it is probably of relatively minor importance (Dudhia 1989; Miller
10 and Frank 1993; Xu and Randall 1995; Tao et al. 1996). All of these studies were in two
11 dimensions; further details will be provided in the discussion of the other two mechanisms.
12 There have not been many numerical modeling studies that have investigated the importance
13 of the horizontal differential heating mechanism in tropical cyclones. Craig (1996), utilizing
14 an axisymmetric model with explicit convection, examined the effect of radiation for an initial
15 vortex with maximum winds of 15 m s^{-1} at a radius of 75 km. Development was quite rapid
16 with azimuthal winds reaching approximately 60 m s^{-1} by 45 h. There was little difference
17 between simulations with and without radiation until 60 h. The longwave cooling was found
18 to increase the maximum intensity by about 20%. Results of sensitivity experiments
19 suggested that differential cooling was the sole mechanism responsible for the enhanced
20 deepening. It was noted that the lack of difference in the early development could have been
21 due to the cloud pattern not being established until 40-50 h, implying little contrast between
22 cloud and cloud-free regions. Also, it was pointed out that initialization with an initial vortex
23 of 15 m s^{-1} without cloud is extreme, since in nature the cloud is likely to precede the
24 establishment of a balanced vortex. On the other hand it was noted that the cloud distribution
25 early in the development of a tropical cyclone is likely to be highly variable. Results of the
26 study were not considered conclusive and further work in this area was recommended.

27 Some modeling studies support the idea that large scale clear-sky environmental cooling
28 can increase precipitation rates in MCSs. Dudhia (1989) used a two-dimensional hydrostatic
29 model with parameterized convection to investigate the life cycle of an MCS in the South
30 China Sea that developed near the coast of Borneo. The MCS was a slow moving system with
31 convective cores that were embedded mainly on the upwind side of a broad area of stratiform
32 precipitating cloud. Sensitivity tests indicated that radiative clear-sky cooling aided the
33 convection by continually destabilizing the troposphere. Two-dimensional numerical

1 modeling studies by Miller and Frank (1993), and Fu et al (1995) of MCSs in an environment
2 typical of the East Atlantic Intertropical Convergence Zone also emphasized the importance
3 of large scale clear-sky cooling. Both of these studies simulated cloud lines with trailing cloud
4 anvils. Miller and Frank (1993) examined the sensitivity to removing the horizontal radiative
5 gradients, while retaining domain-wide radiative destabilization. They found that this resulted
6 in only a small difference in rainfall, leading them to conclude that large-scale radiative
7 destabilization was the main factor causing enhanced rainfall rates when radiation was
8 included. Tao et al. (1996) used a two-dimensional non-hydrostatic cloud-resolving model to
9 simulate the development of both a tropical oceanic squall line and a mid-latitude continental
10 squall line. Again this study found that large scale clear-sky radiative cooling played an
11 important role. However, their experiments indicated that it was not so much destabilization
12 that enhanced surface rainfall in their simulations when longwave radiation was included, but
13 increased relative humidity. They found that Convective Available Potential Energy (CAPE)
14 was not significantly increased by large scale clear-sky cooling. They emphasize that
15 increased relative humidity due to cooling allows condensation to occur more readily.
16 Furthermore, it reduces evaporation and the negative impact of dry air entrainment. These are
17 important additional insights into how large scale clear-sky radiative cooling works in this
18 context. Tao (1996) also found that solar heating reduced precipitation compared with runs
19 with longwave forcing only, and suggested that this was likely to be playing a significant role
20 in the diurnal precipitation cycle found over most oceans.

21 Tao et al. (1996) calculated time- and domain- averaged longwave radiative profiles over
22 the clear and cloudy regions for both squall lines that were simulated (Figure 13 of their
23 manuscript). For the tropical oceanic case below 11 km, the clear-sky longwave cooling was
24 approximately 1.5 K/day larger than for the cloudy regions. This is substantial, and it will be
25 shown in this paper that such a difference, when operative for a twelve-hour period, should
26 produce an unbalanced overturning circulation with significant consequences for convection.
27 As a sensitivity experiment they eliminated differential cooling between cloudy and cloud-
28 free regions by replacing the cloudy heating/cooling profiles with cloud-free radiative
29 cooling. They found that for both the tropical and mid-latitude cases surface rainfall was
30 actually increased, which led them to conclude tentatively that differential cooling was not the
31 mechanism responsible for enhancing the surface precipitation when longwave radiation was
32 activated in the model. An additional sensitivity test was run, in which longwave cooling was
33 allowed to act for six hours prior to triggering convection, and then the simulation was run

1 without any radiative processes. For the tropical oceanic case, which had a small saturation
2 deficit, there was a significant increase in surface rainfall similar to that occurring for the
3 simulation with full radiative interactions. These experiments led them to conclude that the
4 increase of humidity due to large scale radiative cooling was significantly more important
5 than the convergence / differential radiation mechanism¹ (Gray and Jacobson 1977).

6 There are several important caveats to this large scale cooling mechanism that are
7 important for understanding its potential influence on tropical cyclones and their simulation.
8 Typically model simulations are initialized without the initial presence of cloudy air and it
9 takes several hours, often 6-12 h, before deep convection develops. During this period, large
10 scale radiative cooling can occur, the amount of course being influenced by what time of day
11 the model is initialized, which results in increased relative humidity and possibly changes of
12 stability that can enhance convective activity. Once a thick stratiform anvil forms aloft, which
13 is the result of the merging of anvils from numerous deep convective cells, clear-sky cooling
14 becomes confined to the surrounding relatively cloud-free environment. At this point there are
15 two potentially important effects: The surrounding environment becomes more conducive for
16 the development of widespread convection as cloud-free cooling continues in this region, and
17 the low-level inflow of air into the system from the environment becomes more humid,
18 compared to a simulation without radiation. It has yet to be shown how important these
19 effects might be during the early stage of a tropical cyclone's lifetime. Moreover, the question
20 of whether day/night modulations of the radiative forcing in the surrounding environment can
21 by these processes cause the diurnal variations observed in tropical cyclones has yet to be
22 ascertained.

23 Another caveat to the large-scale radiative cooling mechanism is that it can take place in
24

25 ¹The distinction between this mechanism and the first, even though the first also involves clear-sky radiative
26 cooling, is that the resultant cooling of the air and increase of relative humidity occurs in situ, where the clear-
27 sky radiative cooling occurs, and is not a result of radiatively induced circulations. When radiative cooling is
28 horizontally homogeneous and there are no clouds, what happens is clear: the air will cool and there will be an
29 increase in relative humidity in situ. However, when there are clouds present, radiatively induced circulations
30 can occur causing subsidence in the clear-sky region and an associated adiabatic warming tendency that offsets
31 the radiative cooling in a subsiding air parcel. At a particular level drying may occur due to subsidence, since the
32 air aloft is usually drier, leading to reduced relative humidity. In this situation these effects need to be considered
33 as well.

1 between and above scattered cumulus and cumulus congestus. Also, large scale radiative
2 cooling above a low-level stratocumulus layer can lead to increases of humidity and stability
3 changes aloft. In this situation large scale cooling would be a more appropriate term than clear
4 sky cooling for causing the increased potential for convective activity.

5 The third mechanism mentioned is based on the finding that large values of radiative
6 heating during the day and cooling at night occur at the top of oceanic tropical cloud
7 canopies, as well as large longwave radiative heating at cloud base (Webster and Stephens
8 1980). The authors of this study concluded that there could be substantial destabilization of
9 the cloud layer due to this radiative forcing, particularly at night. Xu and Randall (1994) refer
10 to this mechanism as “direct radiation-convection interaction”, and results of their ensemble
11 numerical model simulations led them to conclude that it plays a dominant role in diurnal
12 cycles of deep convection over the tropical oceans. On the other hand, some of the studies that
13 have focused on individual MCSs suggest this mechanism is not as important as large scale
14 clear-sky cooling in enhancing surface precipitation (Miller and Frank 1992; Fu et al. 1995;
15 Tao et al. 1996).

16 This third mechanism has also been by examined in regard to tropical cyclones by
17 Hobgood (1986) who suggested that variations in thermal stratification aloft due to cloud top
18 cooling at night and warming during the daytime might possibly cause the observed diurnal
19 oscillations of the areal extent of cloud canopies. Numerical modeling results presented by
20 Hobgood (1986) indicated that the diurnal cycle of net radiation at the cloud tops was the
21 primary cause of the oscillations. Radiative cooling at night steepens the lapse rate and
22 increases convection. During daylight hours, the absorption of solar radiation reduces the
23 lapse rate, thus resulting in a minimum in convection. It was also suggested that this process
24 might be augmented by differential cooling of cloudy and clear areas, as proposed by Gray
25 and Jacobson (1977). While this early numerical model provided strong evidence of diurnal
26 oscillations in convective activity in tropical cyclones it was very basic compared to today’s
27 models and was unable to actually reproduce diurnal oscillations of the cirrus canopy.

28 A numerical modeling study of the effect of the diurnal radiation cycle on the pre-genesis
29 environment of Hurricane Karl (2010) using the Advanced Research WRF has been reported
30 recently by Melhauser and Zhang (2014). An observational analysis by Davis and Ahijevych
31 (2012) found an approximate diurnal cycle of convective fluctuations with a maximum in the
32 mid- to late-morning and a minimum in the late evening leading up to genesis of Karl. The

1 numerical modeling sensitivity tests showed a case where inclusion of both short and long
2 wave components of radiation led to genesis and intensification whereas a simulation without
3 radiation did not develop. Furthermore it was found that a simulation with nighttime only
4 radiation had a fast genesis and intensification, whereas a day-time only radiation case did not
5 develop. Therefore, these results indicate an important role of radiation in increasing the
6 genesis rate in agreement with NM13, and also showed significant day/night differences of
7 radiation on TC development. The effects of radiation in the Melhauser and Zhang (2014)
8 study were analyzed in terms of the “local environment” by horizontally averaging each
9 model level within a circle of radius 225 km from the vortex center, and the “large scale
10 environment” by averaging over an annulus from 300 to 450 km. The effects of radiation
11 were then independently assessed in each region. Their presented results did not explicitly
12 illustrate diurnal cycles of convective activity. The focus, rather, was on the simulated early
13 development. They noted that their results appeared consistent with the conclusions of
14 previous studies regarding destabilization due to large scale environmental cooling,
15 particularly at night, by Dudhia (1989), Miller and Frank (1993), and Tao et al. (1996).
16 During the daytime they conclude that local and large-scale reduction of relative humidity and
17 increased stability made the overall environment less conducive to deep moist convection.
18 Their study apparently did not examine any potential role of horizontal differential radiative
19 forcing in producing diurnal oscillations of convective activity.

20 Another recent numerical modeling study by Bu et al. (2014) investigated the influence of
21 cloud-radiative forcing on tropical cyclones. The primary model they used was Hurricane
22 WRF. The initialization that they employed resulted in winds of tropical storm strength being
23 reached by 24 h, so they did not focus on the early stages of development. The impact of
24 cloud-radiative forcing was examined by comparing the development of a tropical cyclone
25 having only clear-sky radiation with one having both clear-sky and cloud radiation. The main
26 impact of including cloud-radiative forcing at this stage was to significantly broaden the wind
27 field. A similarity was noted between their results for the vertical profiles of diurnally-
28 averaged net radiative forcing for clear and disturbed (cloudy) regions and those presented by
29 Gray and Jacobson (1977). Sensitivity tests showed that weak, primarily longwave, warming
30 within the cloud anvil was the major factor responsible for modifying the structure when
31 cloud-radiative forcing was included.

32 The present study examines the influence of radiative forcing on tropical cyclogenesis,

1 early intensification, and diurnal oscillations of convective activity. As already discussed
2 there appear to be limitations in the ability of the large scale environmental cooling
3 mechanism to explain continual diurnal cycles in a developing tropical cyclone once a cloud
4 shield has formed, and also to explain the increased genesis rate observed in recent numerical
5 simulations. Increased relative humidity of the inflow air during the nighttime could possibly
6 explain the cycles of precipitation rate. On the other hand, increasing environmental relative
7 humidity at night could cause more widespread convection possibly more conducive to
8 widening the low level circulation rather than strengthening it. This mechanism may indeed
9 be important, but it is not clear that it is the primary factor. The third mechanism also is
10 problematic for explaining an increased genesis rate. The very large oscillations of radiative
11 forcing at cloud top between the night and day could certainly influence ice growth and the
12 strength of convection aloft. It is not clear however that this could lead to an increased low
13 level inflow at night, or a mean increase in low-level convergence that could enhance the
14 genesis rate. The first mechanism discussed has been found to be of minor importance in
15 several numerical modeling studies of MCSs, nevertheless, simple idealized experiments that
16 will be presented in this paper suggest that it should be capable of raising relative humidity
17 values considerably during the nighttime in the cloud system throughout a deep layer.
18 Therefore, it is proposed here that this mechanism is likely to be playing a significant role in
19 both increasing the genesis rate and causing diurnal cycles of convective activity.

20 This is a preliminary study, but makes a strong case for the mechanism proposed by Gray
21 and Jacobson (1977) to explain diurnal cycles of precipitation rate in MCSs as being relevant
22 for tropical cyclones in their early stages of development. Furthermore, it is beneficial to
23 adopt a dynamical perspective recognizing that modulations of the horizontal differential
24 radiative heating generate thermally forced gravity waves, also known as buoyancy bores
25 (Nicholls et al 1991; Mapes 1993). Early work that examined the transient linear response of a
26 stratified atmosphere to prescribed heat sources as a simple model of moist convection, noted
27 that thermally generated gravity waves were playing a fundamental role in compensating
28 subsidence (Lin and Smith 1986; Raymond 1986; Bretherton and Smolarkiewicz 1989).
29 Following these works Nicholls et al. (1991) derived simple analytical two-dimensional
30 solutions to the linear hydrostatic Boussinesq equations for an atmosphere at rest with
31 prescribed heat sources and sinks. For a case with an idealized rigid lid and a deep heat
32 source, represented by a half-sine wave in the vertical, the thermally generated buoyancy
33 circulation was characterized by upward motion in the heated region with outflow aloft and

1 inflow at low levels. The outflow and inflow expanded rapidly on either side of the heat
 2 source. The leading edges of these expanding circulations were deep fast moving wave-like
 3 pulses of subsidence. So the subsidence compensating the central upward motion did not
 4 occur continually over broad regions on either side of the heat source, but had a distinct
 5 horizontally propagating character. The propagating subsidence regions caused adiabatic
 6 warming, adjusting the environmental potential temperature towards the perturbed values at
 7 the heated center for this two-dimensional framework with a rigid lid. Response to a thermal
 8 forcing profile more typical of an MCS having a stratiform region was also examined for a
 9 rigid lid. In this case a deep fast propagating circulation like the one previously discussed was
 10 superimposed on a slower propagating circulation characterized by a mid-level inflow and
 11 upper and lower level outflows. This second slower moving mode had a cool potential
 12 temperature anomaly at low levels and a warm potential temperature anomaly aloft. The
 13 leading pulses of vertical motion had upward motion at low levels and downward motion
 14 aloft. The speed of the modes is given by

$$15 \quad c = \frac{NH}{n\pi} \quad (1)$$

16 where N is the Brunt-Väisälä frequency, H the height of the rigid lid and n the vertical mode
 17 of the heating profile. This profile has the form $\sin(n\pi z/H)$, where z is height.

18 The two-dimensional solution for a semi-infinite region, without a troposphere/stratosphere
 19 interface, shows considerable differences of the low level fields in some respects (Pandya et
 20 al. 1993). In particular, the magnitude of the subsidence is substantially reduced, and it occurs
 21 over a much broader region. Moreover, the axis of the peak vertical velocity in the low level
 22 subsidence region is no longer vertically aligned, but strongly tilted. Nevertheless, adiabatic
 23 warming behind the broader wave front still gradually approach the values at the heated
 24 center. Another factor to consider is that in the real atmosphere there is increased stability
 25 above the tropopause, which partially reflects waves and to some extent increases the
 26 similarity with the rigid lid solution. An early two-dimensional numerical simulation of a
 27 squall line showed a structure qualitatively similar to the first mode during the early stage of
 28 development (Nicholls 1987). The deep convective heating extending to the top of the
 29 troposphere produced a deep overturning circulation with surface mesolows growing laterally
 30 away from the center of the convection at a rapid pace. For the first deep convective mode
 31 that extends throughout the depth of the tropical troposphere, H is approximately 15 km and
 32 taking $N=0.01 \text{ s}^{-1}$ gives a horizontal propagation speed of 48 m s^{-1} . For the second mode the

1 speed is 24 m s^{-1} . So the first mode is very fast moving and while the second mode is
2 considerably slower its speed is still quite fast compared to typical atmospheric motions.

3 Mapes (1993) postulated that higher order modes of the heating profile in an MCS may
4 cause upward displacements at low levels in the nearby atmosphere, thus favoring the
5 development of additional convection nearby. He also emphasized that the wave-like
6 disturbances are not ordinary gravity waves, and pointing out their similarity to tidal bores in
7 water, referred to them as buoyancy bores. There has been considerable amount of research
8 since these earlier studies that has examined their role in convective systems (e.g. Pandya and
9 Durran 1996; McAnelly et al. 1997; Pandya et al. 2000; Nicholls and Pielke 2000; Shige and
10 Satomura 2001; Lane and Reeder 2001; Haertel and Johnson 2001; Fovell 2002; Liu and
11 Moncrieff 2004; Fovell et al. 2006; Tulich and Mapes 2008; Bryan and Parker 2010; Lane
12 and Zhang 2011; Adams-Selin and Johnson 2013). Inclusion of planetary rotation confines
13 the compensating subsidence and adiabatic warming caused by deep convection to a finite
14 distance, measured by the Rossby radius of deformation (Bretherton 1987, 1988; Liu and
15 Moncrieff 2004). Adjustment towards geostrophic balance when a spectrum of inertial-
16 gravity wave modes is generated by a heat source has been investigated in a two-dimensional
17 framework by Liu and Moncrieff (2004).

18 To the author's knowledge, the generation of thermally forced gravity waves, by radiative
19 forcing, rather than latent heating, has not been examined before. The modulations of
20 radiative forcing are generally quite slow, and the circulations generated very weak, so it
21 would be unlikely that these propagating circulations could be directly observed, especially
22 considering that they are typically superimposed on much stronger circulations. Nevertheless,
23 they are evident in the idealized numerical experiments that are discussed in this study and it
24 is useful to consider their properties in order to obtain a more complete understanding of the
25 circulations generated by radiative forcing.

26 Recently Dunkerton et al. (2009) developed the 'marsupial paradigm' that provides a
27 theoretical framework for understanding tropical cyclogenesis from easterly waves. The
28 Kelvin cat's eye within the critical layer, or 'wave pouch', was identified as a favorable
29 environment for tropical cyclogenesis. The theory is supported by both observations of a
30 developing Pacific easterly wave (Montgomery et al. 2010; Raymond and Lopez Carrillo
31 2011) and cloud-resolving numerical simulations (Zhang et al. 2011; Montgomery et al. 2010,
32 2012; Wang et al. 2010; Wang et al. 2012). A numerical model investigation of the

1 thermodynamic aspects near the center of the wave pouch found that it was characterized by a
2 high saturation fraction (Wang 2012). It was hypothesized that updrafts were more vigorous
3 in this region because of reduced dry air entrainment, and that this was favorable for tropical
4 cyclogenesis. This present study makes a similar argument for the effect of enhanced relative
5 humidity in the tropical disturbance due to radiative forcing. Support for the hypothesis is
6 provided by recent cloud model experiments that indicate that for non-supercell environments
7 a major effect of dry air aloft is to reduce the intensity of convection, including total
8 condensation and rainfall (James and Markowski 2009; Kilroy and Smith 2013).

9 The approach that will be used in this study will be to start off by examining the response
10 of an initially motionless atmosphere to a prescribed cooling rate, with a magnitude typical of
11 radiative forcing, applied in the large scale environment surrounding a non-forced core, which
12 represents the region occupied by a cloud cluster. For these simulations radiation,
13 microphysics and surface flux schemes are de-activated. Now it is obviously a considerable
14 idealization to set radiative forcing to zero in the region representing the cloud cluster. We
15 will show later in a full physics simulation of a system that evolves into a tropical cyclone,
16 that the radiation scheme produces in the cloud cluster reduced longwave cooling at low
17 levels, slight warming at midlevels, and very large magnitude forcing at cloud top and in the
18 outer region of the sloping stratiform cloud base. Nevertheless, this simplified framework
19 does give insight into the circulations expected to be induced by horizontal differential
20 radiative forcing at low- and mid-levels between a cloud cluster and its environment. It is also
21 worth noting that Gray and Jacobson (1977) presented estimated day and night radiative
22 forcing rates in the clear-sky and in the disturbance, which for the disturbance were very
23 small beneath 400 mb (Figure 13 of their manuscript). A set of sensitivity experiments using
24 this idealized approach is conducted, and then numerical experiments become progressively
25 more complex, culminating in an examination of the influence of radiation in cloud resolving
26 full physics simulations of tropical cyclogenesis. These latter full physics experiments do not
27 explicitly show the mechanisms causing radiation to influence tropical cyclogenesis, but
28 drawing on the results of the idealized simulations some strong inferences can be made. The
29 combined results of these numerical model experiments suggest that increased relative
30 humidity caused by large-scale environmental radiative cooling in situ does play a significant
31 role in accelerating the rate of tropical cyclogenesis, as other studies have similarly
32 demonstrated its importance for increasing the intensity of MCSs. However, there is probably
33 a more important role played by the circulations induced by horizontal differential radiative

1 forcing, both in accelerating the rate of genesis and in causing diurnal cycles of convective
2 activity. The larger magnitude forcing aloft appears to have less important impacts on these
3 two aspects, but this study does not focus on upper level forcing, so its potential importance
4 for tropical cyclones is left for future research.

5 An outline of the remaining paper is as follows: In Sect. 2 the numerical model is
6 described. In Sect. 3 the designs of the numerical experiments are discussed as well as the
7 particular motivations for them. Results are presented in Sect. 4 and conclusions in Sect. 5.

8

9 **2 Numerical model**

10 RAMS is a nonhydrostatic numerical modeling system comprising time-dependent equations
11 for velocity, non-dimensional pressure perturbation, ice-liquid water potential temperature
12 (Tripoli and Cotton 1981), total water mixing ratio and cloud microphysics. The microphysics
13 scheme has categories for cloud droplets, rain, pristine ice crystals, snow, aggregates and hail
14 (Walko et al. 1995). There have been several improvements to the model physics incorporated
15 since the NM13 study. A two-moment microphysical scheme is now used for all hydrometeor
16 species (Meyers et al. 1997). Additionally, an improved scheme for cloud-droplet riming that
17 uses a binned approach is employed (Saleeby and Cotton 2008). Another change is that the surface
18 parameterization of heat, vapor and momentum fluxes utilizes recent results of the Coupled
19 Boundary Layer Air-Sea Transfer Experiment (CBLAST) presented by Bell (2012).

20 The radiation scheme used in this study is two-stream, and treats the interaction of three solar
21 and five infrared bands with the model gases and cloud hydrometeors (Harrington 1997;
22 Harrington et al. 1999). This parameterization solves the radiative transfer equations for the
23 three gaseous constituents, H₂O, O₃ and CO₂, and for the optical effects of the hydrometeor
24 size spectra. The fast exponential sum-fitting of transmissions method of Ritter and Gelyn
25 (1992) is used for gaseous absorption. The optical properties of water drops are calculated
26 using Lorenz-Mie theory and for non-spherical ice crystals the theory of Mitchell et al. (1996)
27 is used. Since the parameterization includes interaction with liquid and ice hydrometeor size
28 spectra, this enables radiation to respond to variations in droplet size spectra.

29 A standard first-order sub-grid scale turbulence scheme developed by Smagorinsky (1963)
30 is used with modifications by Lilly (1962) and Hill (1974) that enhance diffusion in unstable
31 conditions and reduces diffusion in stable conditions. RAMS utilizes the two-way interactive

1 multiple nested grid scheme developed by Clark and Farley (1984). The radiation boundary
2 condition described by Klemp and Wilhelmson (1978) is used at the lateral boundary of the
3 coarse grid. A Rayleigh friction layer is included at upper levels.

4

5 **3 Description of experiments**

6 The experiments conducted in this study can be classified into three categories. First are
7 idealized experiments without radiation, cloud microphysics or surface fluxes. A list of the
8 ten main experiments in this category is given in Table 2. Cooling or heating rates that
9 represent idealized radiative forcing are prescribed. For many of the experiments a constant
10 cooling is prescribed at the initial time in an environment that surrounds a core region,
11 approximately the size of a cloud cluster, which is absent of any forcing. It is shown that
12 thermally generated gravity waves, or buoyancy bores, propagate quickly into the core and
13 induce upward motion resulting in significant changes of the core temperature and relative
14 humidity. Experiments are conducted for various vertical profiles of the cooling. Additional
15 sensitivity tests examine the response for cooling in annular regions of two different breadths
16 surrounding the core. This is to see how large the cloud free environment needs to be in order
17 to generate significant sustained upward motion in the core. The sensitivity to the width of the
18 core is also examined by increasing it from 200 km to 600 km. This has relevance for
19 contrasting how the response may be different for a small tropical cyclone versus a large one.
20 Then a simple idealized representation of diurnal environmental forcing is used to examine
21 how the core responds to changes in the environmental forcing. This is followed by an
22 experiment to examine the response to forcing within the core, which has a warming region
23 above a cooling region. The reason for this experiment, as previously remarked upon, is that
24 the full physics simulation with radiation included shows a warming at mid levels and a
25 cooling at low levels in the core. Also included at this stage are simulations with two vortices
26 of different strengths to examine how the presence of a vortex influences the induced upward
27 motion in the core. It was found that using a sharp transition between the forced and unforced
28 regions leads to some noise in the fields. To avoid this a linear ramp 50 km wide is used. For
29 instance, for a typical set up that might be designated “unforced for radius less than 200 km”,
30 the region 175 km is actually unforced and linearly changes to the fully forced value at a
31 radius of 225 km. The designation of “unforced for radius less than 200 km” is used for
32 brevity.

1 The second category of experiments has the radiation scheme turned on, but microphysics
2 and surface fluxes turned off. Sensitivity to whether or not radiation is included in the core is
3 examined when there is an initial mid-level vortex present identical to the one used to
4 initialize the full physics experiments. The two experiments in this category are listed in
5 Table 3. The first experiment, that has radiation activated in the core region as well as in the
6 environment, is relevant for the early development of the full physics simulations because it
7 takes some time before deep convection develops and for the cloud canopy to form. The
8 second experiment, which has no radiation in the core, is an idealization, but gives an idea of
9 what circulations and resultant effects caused by radiative forcing outside the core region
10 might be expected to occur in the lower and middle levels of the troposphere when a cloud
11 canopy forms. The simulation begins in the mid-morning and longwave cooling is negated to
12 a large degree by shortwave heating throughout the day. During the nighttime it will be shown
13 that a substantial increase of core humidity occurs for Experiment 12.

14 The third category, are the full physics simulations that include both microphysics and
15 surface fluxes. The main focus of these experiments is to provide further supporting evidence
16 that the differential radiative forcing mechanism is playing a primary role, once a thick cloud
17 canopy forms that is opaque in the infrared. A list of the twelve experiments is given in Table
18 4. The first five experiments are treated as a group. The first simulation, Experiment 13, with
19 radiation included everywhere in the domain is examined in some detail. It is shown how a
20 small tropical cyclone develops quickly along pathway Two (NM13). The following
21 sensitivity tests are carried out: (1) Radiation is turned off everywhere. This experiment
22 demonstrates the difference in the rate of genesis caused by radiative forcing; (2) Radiation is
23 turned off in the core (radius less than 200 km). This experiment is conducted to show that it
24 is likely that mean environmental radiative cooling is having a significant effect on the rate of
25 genesis. Prior to the canopy shield extending beyond a radius of 200 km the radiative forcing
26 can be expected to be similar to Experiment 12, and it can be seen whether the induced
27 upward motion in the core and enhanced humidity influences the intensity of convection; (3)
28 Radiation is turned off in the environment, but included in the core. Again this is to reinforce
29 the conclusion that environmental radiative cooling is important, since the result for this case
30 will be seen to be a slow genesis rate. Together experiments 15 and 16 will make a strong
31 case for differential forcing being important and the direct cloud-radiative interactions aloft as
32 being relatively unimportant in influencing the rate of development of the system; 4)
33 Radiation is turned off and a constant uniform cooling is applied outside the core, below a

1 height of 10 km. This is the same cooling function as used in Experiment 3. The objective is
2 to demonstrate that the changes in the core due to this constant applied differential thermal
3 forcing have a major impact on the genesis rate. The development of the low-level wind speed
4 is compared for the five cases to see which develop into tropical cyclones the quickest. Also
5 the time evolution of the total mass of hydrometeors is compared to see which cases develop
6 significant oscillations of convective activity. The following seven experiments shown in
7 Table 4 were conducted in order to clarify some issues brought up by results of the previous
8 experiments. The reasoning behind these experiments and relation to some of the idealized
9 experiments will be explained in the results section.

10 For the majority of the first category of experiments only one grid is used. The horizontal
11 grid increment is 12 km, with (x, y, z) dimensions of 170×170×48. The vertical grid
12 increment is 60 m and gradually stretched with height to the top of the domain at z=22.3 km.
13 The depth of the upper Rayleigh friction layer is 6 km. The horizontal dimensions of the grid
14 were increased in size for the large annulus experiment to 400 grid points. For the vortex
15 simulations a better horizontal resolution was necessary to resolve adequately the vortex and a
16 nested grid was added with a horizontal grid increment of 3 km and (x, y, z) dimensions of
17 202×202×48.

18 The second category of experiments also includes a nested grid with the same horizontal
19 grid increments and horizontal dimensions as for the previous vortex experiments. The
20 number of vertical grid points is increased to 56 to be consistent with the full physics
21 simulations. In this case the vertical grid increment is gradually stretched from 60 m but not
22 allowed to exceed 700 m, which occurs at approximately a height of 9 km, and thereafter held
23 constant to the top of the domain at z=22.9 km. The better resolution aloft was deemed
24 necessary for the full physics simulations, because the canopy top is near the tropopause and
25 plays a radiatively active role.

26 For the full physics simulations of the third category, three grids are used with horizontal
27 grid increments of 24, 6, and 2 km, and (x, y, z) dimensions of 150×150×56, 150×150×56,
28 and 203×203×56, respectively. Each grid is centered within the next coarsest grid.

29 The temperature and moisture profiles used to initialize the model are the mean Atlantic
30 hurricane season sounding of Jordan (1958), which is slightly moister at low levels than used
31 by NM13. The details of the procedure used for the initial vortex are discussed by
32 Montgomery et al. (2006). For the full physics simulations the initial mid-level vortex is

1 similar to that shown in Fig. 1 of NM13, with maximum winds of 8 m s^{-1} at a radius $r=75 \text{ km}$,
2 and a height of $z=4 \text{ km}$. The core of the vortex is moistened to 85% of saturation with respect
3 to water below 8 km. The model is configured for an f-plane at latitude of 15 degrees. The
4 center of the domain for these simulations is at a longitude of -40 degrees. The shortwave
5 radiation computation accounts for the longitudinal variation of solar zenith angle. All
6 simulations are begun at 1200 GMT. The sea surface temperature for the full physics
7 simulations is set to a constant value of $29 \text{ }^\circ\text{C}$.

8

9 **4 Results**

10 **4.1 Idealized experiments with prescribed forcing**

11 In the first experiment a cooling rate is specified for a radius greater than 200 km from the
12 center of the domain, and has maximum amplitude at a height of 8 km. The cooling extends
13 from the surface where the cooling rate is $-1.2 \times 10^{-5} \text{ K s}^{-1}$ up to a height of 16 km. The
14 maximum amplitude of the cooling at 8 km is $-2.8 \times 10^{-5} \text{ K s}^{-1}$. Figure 1 shows x/z vertical
15 sections through the center of the domain of potential temperature perturbation, pressure
16 perturbation, x-component of velocity u , and vertical velocity at 50 minutes. The cooling has
17 begun to produce a decrease in the temperature of the environment. A laterally propagating
18 wave-like circulation has formed at the boundary between the cooled and non-cooled regions.
19 A region of upward motion is propagating towards the center, whereas a region of downward
20 motion is propagating into the environment. The circulation has a similarity to the two
21 dimensional solutions obtained by Nicholls et al. (1991), but clearly the three dimensional
22 geometry of this simulation has some important implications. For instance, the ring of upward
23 motion near the center is stronger in magnitude than the ring of downward motion in the
24 environment. The circulation is associated with a high-pressure perturbation at the surface and
25 a low-pressure perturbation aloft which is propagating towards the center. The uplift is
26 causing the air to cool adiabatically creating a bulge of cold air in the core at mid-tropospheric
27 levels where the upward motion is strongest. The upward motion has a maximum of
28 approximately 6 mm s^{-1} and the horizontal motions while considerably larger with a
29 magnitude of approximately 0.1 m s^{-1} are also not particularly strong. The ring of downward
30 motion that propagates outward into the environment causes adiabatic warming that offset

31

1 s to some extent the applied cooling. However, the wave amplitude decreases rapidly with
2 distance travelled and, because the wave is propagating, its impact at any location is transient,
3 whereas the environmental cooling is persistent. Figure 2 shows fields at 12 h. A sustained
4 weak upward motion in the unforced region has led to cooling at the center, which above 3
5 km is very similar to that of the environment. A maximum perturbation of over 1 K occurs
6 just below a height of 8 km. The relative humidity has increased significantly as well,
7 particularly near the surface and further aloft within the region of maximum upward motion
8 with an increase of over 6%. Therefore, in spite of the vertical motion being relatively weak
9 over the course of a twelve-hour period, significant changes result in the core. Fig. 2d,
10 showing the vapor mixing ratio anomaly, indicates that the vertical motion has produced the
11 largest change in the lowest 2 km of approximately $.7 \text{ g kg}^{-1}$ where the vertical gradient of
12 moisture in the Jordan sounding is largest (not shown). There is a small increase in the low
13 level relative humidity near the surface in the environment, which must be due to the cooling
14 since there is no increase in the environmental vapor mixing ratio. Fig. 2e shows by far the
15 strongest horizontal motions occur near the surface at a radius of 200 km. The surface inflow
16 at 12 h is fairly substantial with a magnitude of approximately 0.8 m s^{-1} , which is
17 considerably stronger than the horizontal motions that occurred at 50 min associated with the
18 inward propagating wave-like feature shown in Fig. 1. Due to the Coriolis force there has
19 been a small but significant increase of the cyclonic winds at the same location of
20 approximately 1 m s^{-1} (Fig. 2f).

21 This simulation was also run in two dimensions to see if there was a significant three-
22 dimensional aspect to the relatively strong updraft in the unforced core. Interestingly, the
23 results were qualitatively similar in many aspects. The updraft was about 20% weaker at 12 h,
24 but it still resulted in significant changes to relative humidity and potential temperature in the
25 core. A major difference was that the deep pulses of subsidence propagating towards the
26 boundaries were much more evident since they didn't decrease in amplitude so quickly. The
27 leading edges of the pulses reached the open boundaries at approximately 4.5 h and then
28 propagated through them. The distance the pulses travelled to the lateral boundaries from
29 where they formed at 200 km from the center was about 800 km. This gives a propagation
30 speed of about 49 m s^{-1} . The pulses were about 14 km deep so the speed seems in reasonable
31 agreement with the speed of the first mode estimated from Eq. (1), discussed in the
32 introduction, to be about 48 m s^{-1} for a 15 km deep wave.

1 Figure 3 shows results for the second experiment at 12 h that has the same environmental
2 cooling as the first except there is enhanced cooling near the surface. This case produces
3 increased low-level upward motion in the core that increases the low-level relative humidity
4 with the largest perturbation occurring just below 2km (Fig. 3d). Also the increased
5 environmental cooling near the surface produces a notable increase of the low-level relative
6 humidity in the environment.

7 These two preliminary experiments suggest that in order to understand the effects of
8 radiation on a cloud cluster that has a canopy capable of significantly reducing longwave
9 cooling, account needs to be taken of the thermally driven circulations caused by differential
10 heating or cooling between the cloud cluster and the environment. These idealized
11 simulations resulted in significant changes to the relative humidity, lapse rates and surface
12 cyclonic circulation in the unforced core of the system. In addition to the changes caused by
13 the induced circulations there was a notable increase of low-level relative humidity in the
14 environment caused by the environmental cooling in experiment two, which could also
15 influence the development of convection. The latter is the second mechanism of radiative
16 influence on cloud clusters discussed in the introduction.

17 Experiment 3 examines the response to an environmental cooling that is uniform beneath
18 10 km with a value of -1.5 K day^{-1} . In Figure 4 this case is compared at 12 h with Experiment
19 4, which has the same uniform cooling rate prescribed beneath 10 km in an annulus between
20 200 to 400 km from the center. Although the air in the annulus became colder during the first
21 hour this did not persist, and by 12 h it can be seen there are only weak temperature changes
22 in the annulus. There is a weak cooling at the surface, but much of the air in the annulus is
23 actually very slightly warmer. The case with cooling throughout the environment is similar to
24 the first two experiments but is more uniform in the vertical, except at low levels in the
25 unforced core. There is weak downward motion of a few mm s^{-1} in the cooled annulus and
26 this results in a significant decrease of relative humidity. This is due to downward advection
27 of drier air, which is evident since the small potential temperature change in the annulus is not
28 enough to significantly influence the humidity except near the surface. The decrease in
29 potential temperature next to the surface is due to the applied cooling, which is not offset
30 much by adiabatic warming since the subsidence there is small. This results in an increase of
31 humidity in a very thin layer adjacent to the surface. In the core the upward motions are
32 considerably less at 12 h for the annulus case and there are regions of downward motion at

1 $z=2$ and 7 km. There is an increase in surface cyclonic flow but it is not as large as for the
2 case with cooling throughout the environment. There is also anticyclonic flow at the surface
3 at a radius of 400 km.

4 Experiment 5 has a much larger annulus that extends to a radius of 1000 km. Figure 5
5 shows results at 5 h. At this stage the air is colder in the cooled annulus but a very deep region
6 of subsidence is evident travelling from the edge of the outer annulus boundary inwards
7 towards the center that is tending to warm adiabatically the air in the annulus, counteracting
8 the diabatic cooling below 10 km, as well as warming the air above 10 km. At the same time a
9 region of upward motion that is weaker in magnitude can be seen propagating into the
10 environment. At the inner annulus boundary a ring of upward motion propagated into the core
11 and caused a strong sustained upward motion as seen in Fig. 5b, similarly to the other cases.
12 A ring of downward motion also propagated outwards from the inner annulus boundary and is
13 superimposed with the inward propagating downward motion. It is smaller in magnitude
14 however, so the inward propagating downward motion is more evident. It is clear that this
15 inward propagating downward motion is going to significantly influence the vertical
16 velocities in the center when reaching it. This situation also occurred for the 200 km annulus
17 previously discussed but is more easily seen for this larger annulus case. At this time there has
18 been a significant increase in relative humidity in the center and a small decrease in the
19 annulus.

20 Figure 6 shows results at 12 h. There is colder air at low levels in the annulus and a cool
21 region in the core at a height of about 1.5 km. There is also a cold anomaly at the top of the
22 annulus. The middle levels have not cooled significantly however either in the annulus or
23 core. The core does not show significant upward motion as it did at 5 h except close to the
24 surface. There has been a significant increase in the relative humidity in the core, but not as
25 large as for the case shown in Fig, 4e with the cooling throughout the environment. There is
26 some increase of relative humidity at low levels in the annulus where apparently the
27 downward motion is too weak to counteract the increase due to cooling. The anticyclonic
28 surface winds are considerably stronger at the outer boundary of the annulus than at the inner
29 boundary.

30 The annulus experiments show that the response in the core becomes more significant as
31 the size of the region that is cooled is increased as might be expected. They also show that the
32 response can be quite complicated and that there can be significant drying within the cooled

1 region due to subsidence. This case is extremely idealized and would not occur in nature, but
2 is relevant to a full physics simulation of tropical cyclogenesis that will be discussed later. It
3 further illustrates the fast propagation of these thermally generated circulations that travel in a
4 wave-like manner and how the magnitude of the vertical velocity is increased or diminished
5 depending on whether the direction of propagation is towards the center or away from it,
6 respectively.

7 In light of the results of the annulus experiments two other simulations were run without
8 an annulus, which will be briefly mentioned. It would appear that it is important to have a
9 very large surrounding cloud-free environment for the region representing the cloud cluster to
10 be significantly influenced by circulations induced by environmental cooling. A simulation
11 was run similar to Experiment 3, but with a much reduced domain size with a width of only
12 600 km. Additionally, a second simulation was conducted for this reduced grid size with the
13 open boundary conditions replaced by cyclic boundary conditions. In spite of the reduced size
14 of the domain, results for the first simulation were quite similar to Experiment 3, showing
15 significant modification of the core due to induced upward motion. The open boundary
16 conditions allowed the outward propagating pulse of downward motion to propagate without
17 reflection when it reached the boundary and persistent low-level inflow and upper-level
18 outflow at the boundary was established. The strong horizontal pressure gradients responsible
19 for accelerating the inflow at low levels and outflow at high levels were for the most part still
20 represented in the smaller domain simulation. So it was not necessary to have a large domain
21 size to simulate sustained upward motion in the core when radiative boundary conditions
22 were used. One might say, however, that in the real atmosphere, if horizontal gradients in
23 radiative forcing in the far environment surrounding a cloud cluster were at some point in
24 time to occur, possibly due to clouds, they would generate circulations that would propagate
25 towards the cluster, and depending on their strength and duration, could eventually have some
26 influence on it. The effect of environmental cooling at night is therefore going to be more
27 effective when there is a large region of clear sky surrounding the cloud cluster. The second
28 simulation showed a significant difference with Experiment 3 since the pulse of outward
29 propagating downward motion propagated back into the domain because of the cyclic
30 boundary conditions. This resulted in less modification of the unforced core's relative
31 humidity and potential temperature. This latter experiment could be considered to qualitatively
32 mimic what might happen when there are nearby cloud clusters, which would apparently

1 reduce the sustained upward motion in the core produced by nighttime clear-sky radiative
2 cooling.

3 Experiment 6 examines the sensitivity to increasing the size of the unforced core region to
4 a radius of 600 km for a uniform cooling below 10 km in the environment. Figure 7 shows
5 that for this case the colder air at 12 h is less horizontally homogeneous than for a radius of
6 200 km (Fig. 4a). A minimum occurs at middle levels in the core. There is still a fairly
7 significant increase in relative humidity in the core although this is noticeably a maximum
8 near its periphery. This result suggests that the center of a large tropical cyclone will be less
9 impacted by the differential radiative forcing mechanism than a small one.

10 Experiment 7 investigates the response to an idealized diurnal oscillation in the cooling
11 function in the environment. Figure 8 shows time series of the environmental forcing and
12 potential changes it produces both in the environment and the core. A uniform cooling is
13 applied below 10 km and for a radius greater than 200 km, which is constant for 12 h and then
14 decreases in magnitude to zero by 18 h and increases in magnitude back to the original
15 cooling rate by 24 h using a half-sine wave. This approximately represents a cooling at night
16 due to infrared cooling and then a reduced net cooling rate in the daytime due to shortwave
17 heating, which is a maximum at noon. The simulation was run for 72 h. Fig. 8b compares the
18 potential temperature change at a height of 5 km in the environment with the change in the
19 core at the same height. The first 12 h is identical to Experiment 3 (Fig. 4) that was compared
20 to the annulus experiment. The temperature decreases first in the environment, but two hours
21 later it has decreased by as much in the center of the core. Later in the simulation the core
22 temperature anomaly is actually slightly larger in magnitude than in the environment,
23 although the reason for this behavior is unclear. The core temperature at this level is closely
24 matching what happens in the environment with little time lag. During the nighttime there is a
25 steady decrease in temperature and then a leveling off in the daytime.

26 Figure 9a-h shows vertical sections at various times for the idealized diurnal cycle
27 simulation. Fig 9a shows the vertical velocity at 18 h, when the environmental cooling rate is
28 zero, representing the middle of the day. The upward motion that existed at 12 h (Fig. 4c) has
29 been replaced by very weak downward motion through most of the core below 10 km. Fig. 9b
30 shows the vertical velocity at 30 h in the middle of the night. Significant upward motion now
31 extends below 10 km through the core except near the surface. Fig. 9c shows the vertical
32 velocity at 72 h, which is at the end of the third day. There is downward motion near the

1 surface in the core and a weak shallow outflow at the surface (Fig. 9g). The temperature of
2 the air in the core is similar to that in the environment except at low levels (Fig. 9d). There
3 has been a significant increase in the relative humidity both in the environment and in the
4 core. In the environment it is due to the diabatic cooling, whereas in the core it is due to the
5 forced upward motion, which was strong in the three nighttime periods. The core relative
6 humidity is larger than in the environment except at the surface. Similarly to Experiment 2 the
7 perturbation of relative humidity from the initial state is largest just below 2 km. Although the
8 flow at 72 h is slightly divergent at the surface, there was mean low level convergence during
9 the simulation, which resulted in a notable increase of the cyclonic flow of approximately 3 m
10 s^{-1} (Fig. 9f).

11 Results of this diurnal experiment suggest that if a cloud canopy forms aloft that reduces
12 longwave cooling beneath it, then the environmental radiative forcing will induce a
13 temperature tendency at mid-levels in the core that will be similar to what occurs in the
14 environment with only a small time lag. The temperature tendencies will follow the day/night
15 modulations of the environment and except near the surface would also have the same vertical
16 profile. The small time lag appears to be a consequence of the fast propagation speed of the
17 thermally forced gravity wave modes. The temperature tendencies would presumably be
18 superimposed on those occurring in the cloud system caused by latent heating or cooling
19 although this has not been demonstrated here. The induced mean upward motion tendency
20 would increase the humidity in the core most likely favoring the development of convection.

21 Experiment 8 examines the response to an upper-level heating above a low-level cooling
22 for radius less than 200 km. This forced core experiment has relevance because the full
23 physics simulations show that when an extensive canopy forms aloft there is weak mid to
24 upper level radiative heating and weak low level cooling. One of the full physics simulations
25 examines what happens when radiative forcing is only activated for a radius less than 200 km
26 from the center, which results in a similar scenario. There has been no attempt to match the
27 magnitude and spatial distribution of the heating aloft and cooling below with the full physics
28 simulations. Instead a uniform heating rate of 1.5 K day^{-1} is specified between 5 to 10 km and
29 a uniform cooling rate of -1.5 K day^{-1} between the surface and 5 km, for a more straight
30 forward comparison with some of the previous simulations. Figure 10a-f compares fields at
31 1h and 12h. Fig. 10a shows that at 1 h there is air that is slightly warmer above air that is
32 slightly colder as might be expected from the spatial distribution of the heating/cooling

1 function. The magnitude of heating/cooling matches what is expected from the applied
2 forcing ~ 0.06 K/h. Figure 10b shows, however, that at 12 h this pattern has changed
3 considerably. The only notable warm and cold regions at 12 h are at the vertical boundaries of
4 the heating/cooling regions, although they are still very small in magnitude. Fig. 10c shows
5 that significant vertical motions originate at the lateral boundaries of the heating/cooling
6 regions and propagate both towards the center and into the environment. This behavior is
7 because the heating/cooling functions are horizontally homogeneous except at the lateral
8 boundaries. If the heating/cooling functions were strongly peaked at the center then the largest
9 magnitudes of vertical velocity would have started at the center. This situation is similar in
10 some ways to a sea breeze circulation that begins at the land-sea boundary. Since the depth of
11 the thermally driven circulations are shallower than for previous simulations, their lateral
12 propagation is slower (Nicholls et al. 1991, and Eq. 1 of this paper). At 12 h there is broad
13 upward motion in the heated region and downward motion in the cooling region although
14 there are considerable inhomogeneities. This has led to a notable increase of relative humidity
15 in the heating region and decrease in the cooling region. One aspect that this experiment
16 illustrates is that care has to be taken when inferring changes in static stability that might arise
17 from radiative heating profiles that vary in the vertical, because in this case, where the forcing
18 has limited horizontal extent, these changes become negligible in a short time.

19 Experiment 9 considers the response when there is a relatively weak vortex present. The
20 radial distribution of uniform cooling $z < 10$ km is the same as in Experiment 3. The vortex has
21 a depth of 12 km and is strongest at the surface with a wind speed of 12 m s^{-1} at a radius of 75
22 km. Figure 11a shows the initial relative humidity, which is slightly perturbed in the vortex
23 because of the temperature and pressure anomalies associated with it, while Fig. 11b-e show
24 vertical sections at 12 h of y-component of velocity, potential temperature perturbation,
25 vertical velocity and relative humidity, respectively. At 12h there has been only a slight
26 weakening of the vortex strength. The temperature has cooled significantly in the
27 environment and throughout most of the vortex except near the surface central region where
28 the vertical velocity is weak. The vertical velocity peaks just outside the radius of maximum
29 winds, which leads to the most significant increase in relative humidity at this location. This
30 result shows that the vortex is having a considerable influence on the induced motion in the
31 unforced core.

1 Experiment 10 is similar to the previous experiment except that the vortex strength is
2 increased to 30 m s^{-1} . Figure 12 shows a more pronounced minimum in vertical velocity at the
3 center of the vortex. Just outside the radius of maximum winds there is still a significant
4 increase in the relative humidity. This result suggests that environmental radiative forcing at
5 lower and middle levels could still have a significant influence on the convection within a
6 strong tropical cyclone, but that it is more likely to impact the outer region of the system.

7 Several recent theoretical studies have investigated the response to heating in a tropical
8 cyclone-like vortex under the assumption of gradient wind balance in the radial momentum
9 equation (Wirth and Dunkerton 2006, 2009; Pendergrass and Willoughby 2009; Vigh and
10 Schubert 2009). These analyses lead to consideration of the “transverse circulation equation”
11 first derived by Eliassen (1951). A similar analysis that includes representation of
12 environmental radiative forcing could potentially provide a fuller interpretation of the results
13 found for the vortex experiments shown in this study. The neglect of the time derivative of
14 radial velocity in the radial momentum equation means that the thermally generated pulses of
15 vertical motion that eventually propagate into the far environment would not be simulated,
16 nevertheless the induced sustained upward motion in the vortex should qualitatively be the
17 same. This could lend insight into the reason for the radial variation of vertical velocity
18 induced in the vortex. A reasonable speculation is that it is a consequence of the inertial
19 stability of the vortex.

21 **4.2 Idealized experiments with radiation scheme included**

22 Experiment 11 proceeds to examine the response to the Harrington radiation scheme, instead
23 of specified forcing, for the same initial vortex that will be used for the full physics
24 simulations. There are no clouds or cloud-radiative feedbacks. Figure 13 shows vertical
25 sections of the initial y-component of velocity and relative humidity. The core of the vortex
26 has been moistened to 85 % of saturation similarly to many of the experiments in NM13,
27 which encourages the development of a tropical cyclone in the full physics simulations. A
28 difference with the previous simulations is that instead of the vapor mixing ratio abruptly
29 being set to zero above 11 km there is a more gradual decrease with height. The infrared
30 cooling between 7-9 km is quite sensitive to the existence of small amounts of water vapor
31 aloft. A more gradual reduction with height decreases the infrared cooling rate in this layer
32 (Norman Wood, personal communication), which without this modification is quite large.

1 Figure 14 shows the radiative flux convergence and vertical velocity at 4 h into the
2 simulation, which is during the middle of the day. At this time the solar heating is stronger
3 than the infrared cooling in the upper troposphere and between 1-6 km. The strongest cooling
4 occurs at the top of the moistened core and the strongest heating between 5-6 km, producing
5 downward and upward motion, respectively. An east-west asymmetry can be seen in the
6 radiative forcing due to longitudinal variation of the solar radiation.

7 Figure 15 shows fields at 10 h, during the early nighttime. A layer of strong cooling of
8 approximately -3K day^{-1} occurs outside the moistened core between $z=7-8$ km. An even
9 stronger cooling in this layer occurs in the moistened core. There is a maximum near 4 km
10 and moderate cooling below in the environment. There is a maximum at low levels in the
11 moistened core. At this time the only significant vertical velocity is downward at the top of
12 the moistened core where the cooling is strongest. At this time there hasn't been a significant
13 change to the relative humidity from the initial values. The strong environmental cooling
14 produced by the Harrington radiation scheme at around the 8 km level is similar to the result
15 found for infrared cooling during moist conditions in the equatorial Western Pacific by Zhang
16 and Chou (1999) who calculated a maximum cooling at around 300 mb (Fig. 4 of their
17 manuscript).

18 Figure 16 shows fields at 24 h, which is mid-morning. The potential temperature
19 perturbation is negative in the environment throughout the troposphere except at the very top.
20 Beneath 3.5 km there has been a cooling of approximately -1.5 K. Between 4-5 km it is about
21 half as large, whereas it is more than -2 K day^{-1} between 7-8 km. The overall cooling has led
22 to a significant increase in relative humidity. Note that the contour intervals in Fig. 16c have
23 been changed from previous figures for better comparison with the next experiment.
24 Subsidence is still noticeable at the top of the vortex at 24 h and weak upward motion above.
25 The vortex winds have not changed a great deal in the 24 h period showing only a slight
26 weakening, probably due to model diffusion.

27 Experiment 12 is similar to the previous one, except the radiation is turned off for radius
28 less than 200 km. Figure 17 shows fields at 4 h. The radiative flux convergence in Fig. 17a is
29 identical to that in Fig. 14a for radius greater than 200 km. A significant upward motion is
30 induced in the core within the 7-12 km layer, where there is cooling in the environment.
31 Conversely downward motion is induced in the core within the 3-6 km layer where there is
32 warming in the environment.

1 Figure 18 shows that at 10 h during the nighttime there is significant upward motion at
2 upper levels associated with the strong environmental cooling aloft. The sustained upward
3 motion both during the day and night has led to increased humidity in the core at upper levels.
4 The perturbed relative humidity field also shows that even though a nested grid was used to
5 increase horizontal resolution that there is still some noticeable lateral diffusion due to the
6 subgrid scale turbulence scheme where the velocity gradients are large.

7 Figure 19 shows fields at 24 h. There has been a more significant reduction in potential
8 temperature in the warm core aloft compared to Experiment 10 (Fig. 16a). Also the air at the
9 surface in the core is significantly warmer than the environmental air. At this time the
10 strongest upward motion is near the surface at a radius of 200 km. There has been a large
11 increase of relative humidity in the core to values far in excess of saturation at a height of 8
12 km. The larger increase of relative humidity aloft compared to low levels for this simulation
13 that uses the radiation scheme is in contrast to the results of Experiments 2 and 7 that showed
14 largest increases at low levels. Note that since the model does not have microphysics
15 activated the accumulation of moisture to values well in excess of saturation is able to occur.
16 Another difference with the previous experiment is that wind speeds increase at the surface by
17 approximately 1.5 m s^{-1} .

18 The results of Experiment 11 during the first 12 h are similar to what occurs in the full
19 physics simulation with radiation activated, which will be discussed shortly. There are some
20 differences since the full physics simulation has surface fluxes, and also, shallow clouds
21 develop by 12 h. There is a small but still significant increase in low-level relative humidity
22 both in the core and in the environment in Experiment 11, which in the full physics simulation
23 tends to promote the development of low-level clouds. In the full physics simulation with
24 radiation included in the whole domain, deep convection develops after 12 h, which shortly
25 thereafter causes an upper level canopy to form that modifies the radiative fluxes
26 considerably. It will be shown that this results in a reduction of longwave cooling at low
27 levels in the core and a slight warming at middle levels. Experiment 12 is idealized since the
28 radiative forcing is set to zero in the core beneath 10 km, but does illustrate that weak but still
29 significant circulations could be expected to develop in the more complex full physics
30 simulations associated with the differential radiative forcing between the environment and the
31 core when a cloud canopy forms.

32 **4.3 Full physics experiments**

1 The next set of experiments to be discussed have surface fluxes and cloud microphysics.
2 Experiment 13 that has radiation included will be described in some detail. Figure 20 shows a
3 horizontal section of total hydrometeor mixing ratio at a height of 11.7 km at 15 h, and
4 vertical sections through the center of the domain of total hydrometeor mixing ratio and
5 radiative flux convergence at 21 h. Figure 20a shows that several deep moist convective cells
6 have developed by 15 h and a canopy aloft is starting to form as the anvils from the individual
7 cells begin to merge. Figure 20b shows that six hours later the canopy aloft is more
8 widespread. A strong cell is evident at $x=90$ km. Figure 20c shows that at this time, which is
9 during the night just before daybreak, there is strong cooling at the top of the canopy. There is
10 also strong warming at the base of the canopy in the outer region of the core.

11 Figure 21 shows vertical sections at 30 h, which is in the late afternoon. The cloud canopy
12 has continued to develop. There is significant shortwave warming of the cloud canopy aloft,
13 but longwave cooling at the top of the cloud canopy is also starting to occur by this time. It is
14 notable that the shortwave warming of the cloud canopy aloft occurs just beneath the region
15 of longwave cooling. The x-component of velocity shown in Fig. 21c has strong velocities
16 associated with the intense cell present at $x=50$ km. There is also a sloping inflow evident at
17 the base of the cloud layer, which is very persistent and which leads to the spin up of the mid-
18 level circulation seen to be occurring in Fig. 21d, between $z=4-7$ km. The mid-level
19 circulation continued to intensify and at 38 h a small vortex suddenly formed, which was
20 concentrated at the surface and close to the center of the domain.

21 Figure 22 shows a horizontal section of the vertical relative vorticity near the surface at 40
22 h and a vertical section of the y-component of velocity at 48 h. There can be seen a small
23 region of intense low-level positive vertical vorticity at the center of the domain at 40 h. This
24 small vortex became the focus of a rapidly intensifying circulation and cyclonic winds
25 exceeded 30 m s^{-1} by 48 h. The development of this tropical cyclone was along pathway Two
26 as discussed in NM13, and appears to be similar to the results of Nolan (2007). The evolution
27 was similar to other cases discussed in NM13 that evolved along pathway Two, even though
28 there are some differences in the model parameterizations and in the initial moisture profile,
29 as discussed in section 2. The tropical cyclone that formed is small and compact, which is an
30 advantage for these experiments since it does not require a large fine-scale grid.

31 Figure 23 shows time series of the maximum near surface azimuthally averaged
32 tangential wind speeds and total hydrometeor mass in the domain for the first five

1 experiments shown in Table 4. Also shown is the shortwave cloud-free radiation reaching the
2 surface for a simulation without microphysics, initial vortex or surface fluxes. Experiment 13
3 with radiation throughout the whole domain intensifies very similarly to Experiment 15 with
4 radiation only in the environment (radius greater than 200 km). Near surface wind speeds
5 reached 12 m s^{-1} around 36 h followed by a rapid rate of intensification until 48 h and then a
6 somewhat slower rate until the end of the simulation at 84 h. The fastest development
7 occurred for Experiment 17 with environmental cooling beneath a height of 10 km at a
8 constant rate of -1.5 K day^{-1} . The case without radiation or any prescribed forcing did not start
9 to intensify until the very end of the simulation. The case with radiation only in the core also
10 did not develop until late in the simulation, but did show a very rapid intensification
11 beginning at 72 h. These results suggest that the radiation in the environment is an important
12 factor in causing the accelerated rate of development when radiation is activated in the model.
13 Including radiation only in the core dramatically slowed the development compared with
14 including radiation everywhere. Therefore, it appears unlikely that the very strong upper level
15 radiative heating and cooling rates in the core of the system are having a large effect on the
16 rate of development. It seems more likely that the circulation induced by differential heating
17 and cooling between the environment and the cloudy region at low and middle levels is the
18 main cause of the accelerated development when radiation is included. The previous idealized
19 experiments suggest that an increase in relative humidity due to weak upward motion in the
20 core, particularly significant during the nighttime when environmental cooling is strongest, is
21 the main factor responsible.

22 The total hydrometeor mass in the fine grid domain in Fig. 23b shows a very significant
23 oscillation between 36-48 h occurring for both the case with radiation included everywhere
24 and for the case with radiation only included in the environment (Experiments 13 and 15).
25 Fig. 23c shows that this coincides with the second night and that the rapid increase in total
26 hydrometeor mass came to an end during the early daylight hours. The case with radiation
27 included only in the environment also had a significant oscillation that began during the first
28 night. In NM13, diurnal oscillations of the total hydrometeor mass also occurred for
29 simulations with radiation included, a particularly prominent example being shown in Fig. 11
30 of that paper. It can also be seen that Experiment 17 with thermal forcing constant in time also
31 had two moderately large oscillations early on.

1 In order to obtain a better understanding of the role of radiative forcing, time and
2 azimuthal averages of the radiative flux convergence were made for Experiment 13 and are
3 shown in Figure 24. Fig. 24a shows an average between 24-48h for the whole troposphere,
4 whereas Figs. 24b shows the same cross section magnified for the lower troposphere. Fig. 24c
5 shows a 6 h average for the previous night between 15-21 h, and Fig. 24d show a 6 h average
6 in the daytime between 24-30 h. Fig. 24a is dominated by the strong radiative forcing at upper
7 levels. The warming at the base of the stratiform ice layer is quite smeared due to its variation
8 over this 24 h period. Fig. 24b, which provides more detail of the lower and middle
9 troposphere, shows significant horizontal gradients of the 24 h averaged radiative forcing.
10 There is a slight cooling at low levels and a slight warming at middle levels in the core. The
11 horizontal difference between the core and the environment is about -1 K day^{-1} at low levels
12 and slightly larger in magnitude at mid levels. During this period a shallow cloud layer
13 developed, which extended into the environment and there is quite strong longwave cooling
14 that can be seen at the top of this layer. Figure 24c shows a pronounced horizontal gradient
15 during the previous nighttime that is particularly large aloft. In the daytime Figure 24d shows
16 that the gradients are a lot weaker and at low levels reversed. Based on the idealized
17 simulations discussed previously, the magnitude of the averaged horizontal gradients at low
18 and mid levels appear large enough to drive circulations resulting in significant changes in the
19 core capable of promoting convective activity. Since the gradients of radiative forcing are
20 stronger aloft the largest increase of relative humidity in the core is also likely to be aloft.
21 This also applies to cooling, which would tend to make convective updrafts stronger at upper
22 levels. For this environment and these particular RAMS simulations a tentative conclusion is
23 that the radiative forcing is making the mid level local environment more favorable for
24 enhancing convection that has been triggered at low levels, and this could be more important
25 than causing an increase of the low level convective triggering, although both are likely
26 occurring.

27 Experiments 18-24 were conducted in order to obtain a clearer picture of the role of
28 radiative forcing suggested by the foregoing results. For instance, Experiment 16 with
29 radiation included only for radius less than 200 km and which developed slowly seems to
30 suggest that mean cooling of the environment is a major factor responsible for the accelerated
31 rate of genesis. Presumably if the radius within which radiation is included were increased say
32 to 400 km, this would not be large enough to encompass much of the environment
33 surrounding the cloud disturbance and therefore genesis would remain slow. Therefore

1 Experiment 18 includes radiation for radius less than 400 km. It also might be expected that if
2 radiation were included only in the environment but for a larger radius than 200 km, which
3 was the value used in Experiment 15, then there would be an accelerated rate of genesis, but
4 probably not as rapid as for Experiment 15. Therefore Experiment 19 includes radiation for
5 radius greater than 500 km. An interesting issue is how important radiative forcing at low
6 levels is for increasing the rate of genesis when radiation is included. To examine this issue
7 Experiment 20 has radiation included except in the lowest 1.5 km. Equally interesting is the
8 importance of radiation at upper levels for increasing the genesis rate. Therefore Experiment
9 21 has radiation turned off above 8 km to see how this affects the rate of development. It has
10 been noted in this and other studies that clear sky radiative cooling of the atmosphere will
11 produce increased relative humidity, which might be expected to promote convective
12 development. Experiment 22 examines the effect of persistent horizontally uniform cooling to
13 examine how the resultant increase in relative humidity affects convective development. For
14 this case the radiation scheme is turned off but a uniform cooling rate of -1.5 K day^{-1} is
15 prescribed beneath a height of 10 km, throughout the domain. Finally, two more experiments
16 are conducted, one with only radiative cooling (Experiment 23), and one with only radiative
17 warming (Experiment 24). Combined with the results of other experiments these give a better
18 indication of the relative importance of environmental radiative cooling versus mid-level
19 radiative warming in the core, in enhancing the development of the tropical disturbance.

20 Figure 25 shows the maximum near surface azimuthally averaged tangential winds for
21 Experiments 18-24. As expected Experiment 18, which has a 400 km core region where
22 radiation is included, has a slow genesis rate. The development was however quite complex.
23 As a large stratiform anvil formed, extending out to approximately 200 km radius, there was
24 still a considerable cloud-free area beyond it where the radiative transfer scheme likewise was
25 activated. In this clear region surrounding the cloudy core, subsidence predominantly due to
26 nighttime cooling produced a region of less humid air surrounding the developing cloud
27 system. This scenario has similarities to Experiment 4 that examined the response to a small
28 annulus of cooling between 200-400 km. The cyclonic low-level wind speeds increased fairly
29 significantly early on, more so than for Experiment 16 that had a 200 km core where radiation
30 was included. This could have been partly due to upward motion induced by the surrounding
31 subsidence, although Experiment 4 suggests that this might not be very large. Another factor
32 is that the system was notably more compact, apparently because the subsiding ring of air was
33 less humid, thereby reducing convective activity on the periphery of the cloud system. The

1 more centrally focused convection may possibly have played a role in the early strengthening
2 of the low-level circulation. In the long term the dry surrounding ring of air is likely to have
3 been a factor inhibiting convection until the stratiform anvil grew large enough to reduce the
4 cloud-free radiative forcing. The lack of mean cooling in the environment beyond a radius of
5 400 km also appears to have been a major factor in the slow genesis rate, as expected.

6 Experiment 19, with radiation included beyond a radius of 500 km, developed a tropical
7 cyclone more slowly than Experiment 15, with radiation beyond a radius of 200 km (Fig.
8 23a). Development was still however considerably faster than the case without radiation. This
9 supports the conclusion that mean cooling in the large-scale surrounding environment is a
10 major factor responsible for increasing the rate of tropical cyclogenesis in the model when
11 radiation is included.

12 The case with no radiation beneath 1.5 km developed slightly slower than the case with
13 radiation everywhere (Fig. 23a). This result indicates that the effects of radiative forcing
14 above 1.5 km are having the most influence on the genesis rate.

15 Experiment 21 with no radiative forcing above 8 km developed the fastest among these
16 five experiments. The rate of development was similar to Experiment 13 with radiation at all
17 levels. This result suggests that radiative forcing's at lower and middle levels have the most
18 impact on the genesis rate, in spite of there being much stronger heating and cooling rates
19 aloft.

20 Experiment 22 with a prescribed horizontally uniform cooling beneath 10 km (including
21 the core inside 200 km) developed considerably slower than most of the other cases, but still
22 significantly faster than the case without radiation. This supports the idea that increases of
23 relative humidity due to large scale radiative cooling without horizontal gradients can
24 promote convective activity (e.g. Dudhia 1989; Tao 1996). However it is apparently not as
25 effective in promoting convective activity as differential horizontal radiative forcing, which
26 generates circulations in the core. Moreover, the experiment is very idealized since once a
27 cloud canopy develops it will likely cause differential horizontal forcing. As discussed in the
28 introduction the question becomes how influential is the increase in relative humidity in the
29 cloud free environment where cooling occurs? Another factor worth mentioning is that
30 because there continues to be some weak longwave cooling at low levels in the simulated
31 cloud system, which would tend to increase the humidity in situ, this could enhance
32 convective activity to some extent.

1 Experiment 23 with negative radiative forcing only underwent genesis fairly quickly, but
2 still significantly slower than with the full radiative forcing. Experiment 24 with positive
3 radiative forcing only showed the slowest development in Fig. 25, but still faster than the no-
4 radiation case (Experiment 19, Fig. 23a). Therefore, it appears that warming, probably at mid
5 levels in light of Experiment 21, contributes to increasing the rate of genesis, but it is not as
6 important as the large scale environmental cooling in inducing a circulation in the core.

7

8 **5 Conclusions**

9 The results of this numerical modeling study suggest that circulations generated by
10 differential radiative forcing may increase significantly the rate of tropical cyclogenesis and
11 also play a major role in diurnal oscillations of convective activity. Experiments show that the
12 influences of the radiatively induced circulations on convective activity in the system include:
13 (1) increased relative humidity (2) changes in static stability, and (3) increased low level
14 cyclonic circulation. The first and second of these differential radiative forcing effects are
15 caused by upward motion within the core, which is strongest during the night. The model
16 results indicate relatively weak vertical motions are induced in the core during the day, so that
17 in the mean there is a net upward motion tendency caused by differential radiative forcing.
18 The third influence is due to convergence of absolute vorticity by the low level inflow, which
19 results in a small increase in the low level cyclonic winds. This is also likely to increase the
20 rate of genesis.

21 It has not been shown explicitly in this study that vertical motions induced by differential
22 radiative forcing are predominantly the cause of the accelerated rates of genesis in the full
23 physics simulations when radiation is included, but nevertheless it is a strong inference that
24 can be made. The radiatively induced vertical motions are very small compared to those
25 produced by latent heating in convective towers, so it is very hard to discern them explicitly
26 in a full physics simulation. It is assumed here that the small vertical velocity tendencies that
27 would be expected due to the differential radiative forcing are superimposed on the stronger
28 vertical motion produced in convective towers and in the regions between them where often
29 there is also relatively strong vertical motions, for instance in stratiform cloud layers, shallow
30 cumulus clouds, and in turbulence and gravity waves. This has not been demonstrated
31 explicitly, but the fact that idealized simulations with prescribed differential forcing show
32 significant impacts in the core, and that the same differential forcing applied in full physics

1 simulations show increased convective activity and faster rates of genesis, supports this
2 contention.

3 This study interprets this problem in terms of the propagation of thermally generated
4 gravity waves, or buoyancy bores. It was noted that a deep prescribed radiative cooling in a
5 finite radial range produces deep rapidly propagating wave-like pulses of vertical motion in
6 line with previous studies that have focused on their generation by latent heat release in
7 convective systems. Observational studies by Bryan and Parker (2010) and Adams-Selin and
8 Johnson (2010, 2013) appear to have detected the passage of this kind of gravity wave mode
9 generated by convection in intense mesoscale systems. Modulations of radiative forcing in the
10 real atmosphere are unlikely however to be large enough in magnitude, or rapid enough, to
11 ever produce observable gravity wave modes of this type. The numerical results however,
12 strongly suggest that they are present. The result for the idealized diurnal forcing experiment,
13 that the potential temperature changes at mid levels in the core closely parallel those in the
14 environment with hardly any noticeable delay, are evidence of the fast propagation speed of
15 these waves. Examination of the propagation of these waves also helped to explain the results
16 of the annulus and forced core experiments. Although the inward propagating thermally
17 induced circulation that occurs in the idealized environmental cooling experiments shows a
18 wave-like character, once it reaches the center a relatively steady sustained deep upward
19 motion ensues. The low-level inflow associated with this circulation becomes substantial,
20 reaching magnitudes of approximately 0.8 m s^{-1} in these experiments.

21 These experiments indicate that the large scale clear sky environmental cooling
22 mechanism (Dudhia 1989; Tao et al. 1996) is also having a noticeable effect. During the early
23 stages of the simulation, before an extensive cloud shield develops aloft, it leads to an
24 increase in humidity that favors the development of moist convection. It is possible that for
25 the case of a prolonged genesis period where convection is sporadic and there are large
26 regions of cloud free air this mechanism could be the main driver of diurnal cycles of
27 convective activity in a tropical disturbance. After the formation of a cloud shield the large-
28 scale surrounding environment continues to cool significantly during the nighttime. It is not
29 completely clear from the experiments how influential this aspect is on the cloud system.
30 What stands out from the idealized experiments is the very large impact of differential forcing
31 on relative humidity and potential temperature in the core of the system, so it seems unlikely
32 that the large scale clear sky environmental cooling mechanism is as important a factor in

1 influencing the development of a tropical disturbance once an extensive cloud canopy has
2 formed. The simulation with uniform cooling applied throughout the domain supports this
3 conclusion (Experiment 22), since the system did not develop as fast as for the differential
4 forcing case (Experiment 15).

5 The very large heating and cooling rates that occur at the canopy top and the large heating
6 rate at the stratiform ice base did not seem to have much influence on the genesis rate.
7 Experiment 21 with radiative forcing aloft turned off still underwent genesis almost as
8 quickly as Experiment 13 with radiative forcing aloft. Experiments 15 and 18, with radiative
9 forcing aloft but no environmental forcing, did not develop quickly, again supporting this
10 conclusion. It is however possible that the large variations of heating and cooling at the
11 canopy top could influence the areal extent of cloud cover aloft. This study has not looked at
12 this issue.

13 The weak mid-level longwave warming that occurs in the core of the system when
14 radiation is included appears to have a fairly small effect on the genesis rate, but not an
15 insignificant one. Experiments 15 and 18 with radiation only in the core both underwent
16 genesis quicker than the no-radiation case, probably because of the mid-level warming. For
17 the simulation with radiative cooling only (Experiment 23), the system underwent genesis
18 considerably slower than for the case with positive and negative radiative forcings
19 (Experiment 13), which is consistent with the view that mid-level warming enhances the
20 genesis rate. The idealized simulation with mid-level warming in the core (Experiment 8)
21 showed increased relative humidity aloft, and mid-level radiation warming also clearly
22 contributes to increasing the horizontal differential radiative forcing at night shown in Fig. 24.
23 However it is not the dominant radiative influence on the genesis rate in these simulations. It
24 is predominantly the nighttime cooling in the environment that is responsible for creating the
25 strong gradient of differential forcing that leads to an increased genesis rate.

26 The idealized experiments with a prescribed cooling outside a wide core (Experiment 6)
27 and the strong vortex simulation (Experiment 10), both suggest that for a large intense
28 tropical cyclone the effects of differential radiative forcing caused by mean environmental
29 cooling at low and mid levels might enhance convection on the periphery of the system. It
30 would be interesting to examine this aspect in future work.

31 As discussed in the introduction several previous numerical modeling studies of MCSs
32 concluded that the differential radiative forcing mechanism proposed by Gray and Jacobson

1 (1977) played only a minor role. These results are at variance with the present study, and it is
2 not clear why these previous investigations did not find a significant effect. Gray and
3 Jacobson (1977) stated that the more intense the convection and the more associated it is with
4 an organized weather system the more evident the diurnal cycle. The system simulated in this
5 study with radiation activated probably has more intense convection than those simulated in
6 the previous studies and it is certainly an organized and persistent system surrounded by a
7 relatively cloud free environment, so this is likely to be contributing factor responsible for the
8 different result. Also, the majority of the previous MCS studies were of convective lines with
9 a leading convective region and a predominantly trailing stratiform region, so that the location
10 of the strongest differential radiative forcing may not have coincided with the location of the
11 strongest convection.

12 This study may have ramifications for why tropical cyclogenesis often occurs in weak
13 vertical wind shear (e.g. Gray 1968; McBride and Zehr 1981). Vertically sheared
14 environments with winds aloft significantly different than those at low levels may not be
15 favorable for the formation of symmetrical optically thick stratiform canopies that remain
16 above the center of low level convergence. This could be a contributing factor for why large
17 vertical wind shear is unfavorable for tropical cyclogenesis.

18 The validity of this study depends crucially on whether significant differential radiative
19 forcing actually exists between tropical disturbances and their surrounding environment. The
20 physical processes are complex involving interaction between radiation and microphysics.
21 Moreover, accurate numerical modeling of these processes requires that the grid resolution be
22 good enough to realistically simulate deep convective cells since they are primarily
23 responsible for the formation of an extensive cloud canopy aloft. Further studies are
24 necessary, particularly ones that simulate real systems, to see if this radiative mechanism is
25 indeed playing a major role.

26

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Table 1. General statistics comparing four cases without and with radiation from NM13. Moist, dry refers to the core having an initial moisture anomaly or not. All cases are for a weak initial vortex having a maximum tangential wind speed of 8 m s^{-1} at $z=4 \text{ km}$. Shown are the time the maximum averaged tangential wind speeds at the surface reach 12 m s^{-1} ; the near-surface radius of maximum winds (RMW) at this time; the pathway taken to genesis; the time at which the system becomes a tropical storm (TS); the RMW at this time; the time the system becomes a hurricane (H); the RMW at this time.

Exp.	Description	$T_{12}(\text{h})$	RMW ₁₂ (km)	Path	$T_{\text{TS}}(\text{h})$	RMW _{TS} (km)	$T_{\text{H}}(\text{h})$	RMW _H (km)
1	No radiation, moist, small, weak, SST29	82	9	2	92	13	103	13
2	Radiation, moist, small, weak, SST29	60	5	2	65	11	78	13
3	No radiation, dry, small, weak, SST 29	151	5	2	174	11	189	13
4	Radiation, dry, small, weak, SST29	101	11	2	105	13	119	15
5	No radiation, moist, large, weak, SST29	89	5	2	103	23	114	23
6	Radiation, moist, large, weak, SST29	55	5	2	61	19	69	15
7	No radiation, moist, large, weak, SST28	112	41	1	118	37	127	23
8	Radiation, moist, large, weak, SST28	49	5	2	52	9	68	15

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Table 2. Experiments with prescribed thermal forcing

Experiment	Description
1	Maximum cooling aloft: Environmental cooling for $r > 200$ km, $z < 16$ km, with maximum amplitude at 8 km.
2	Maximum cooling aloft and at low-levels: Environmental cooling for $r > 200$ km, $z < 16$ km, with maximum amplitude at 8 km, and a secondary maximum at the surface.
3	Uniform cooling: Uniform environmental cooling for $r > 200$ km, and $z < 10$ km.
4	Small annulus: Uniform cooling between $r = 200$ km to $r = 400$ km, and $z < 10$ km.
5	Large annulus: Uniform cooling between $r = 200$ km to $r = 1000$ km, and $z < 10$ km.
6	Wide unforced region: Uniform environmental cooling for $r > 600$ km, and $z < 10$ km.
7	Diurnal forcing: Idealized diurnal oscillation of uniform environmental forcing for $r > 200$ km, and $z < 10$ km.
8	Core forcing: Uniform warming between $z = 5$ km to 10 km, uniform cooling below 5 km, for $r < 200$ km.
9	Weak vortex: Uniform environmental cooling for $r > 200$ km, $z < 10$ km, and a vortex with surface winds of 12 m s^{-1} .
10	Strong vortex: Uniform environmental cooling for $r > 200$ km, $z < 10$ km, and a vortex with surface winds of 30 m s^{-1} .

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Table 3. Radiative transfer scheme activated with a mid-level vortex and a moistened core

Experiment	Description
11	Radiative transfer scheme activated in the whole domain, with a weak mid-level vortex and a moistened core.
12	Radiative scheme activated in the environment for $r > 200$ km, with a weak mid-level vortex and a moistened core.

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Table 4. Experiments with full physics.

Experiment	Description
13	Radiation in the whole domain.
14	No radiation.
15	Radiation only in the environment, $r > 200$ km.
16	Radiation only in the core, $r < 200$ km.
17	Prescribed uniform environmental cooling, for $r > 200$ km, $z < 10$ km.
18	Radiation only in a large core, $r < 400$ km.
19	Radiation only in the environment outside a large unforced region, $r > 500$ km.
20	No radiative forcing below 1.5 km.
21	No radiative forcing aloft, above 8 km.
22	Prescribed horizontally homogeneous cooling throughout the domain, uniform below 10 km.
23	Radiative cooling only.
24	Radiative warming only.

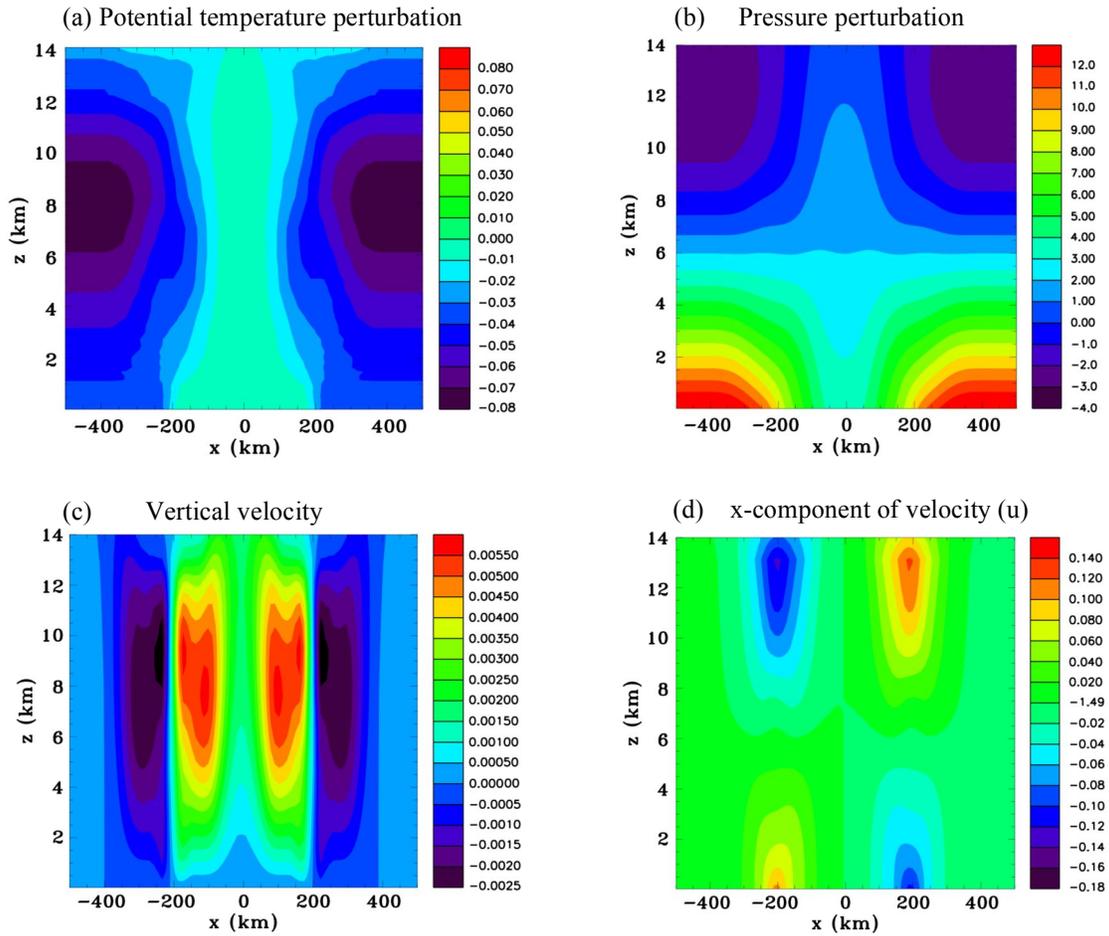


Figure 1. Vertical sections for Experiment 1: the maximum cooling aloft case, at $t=50$ minutes. **(a)** Potential temperature perturbation (K), **(b)** pressure perturbation (mb), **(c)** vertical velocity (m s^{-1}), and **(d)** x-component of velocity, u (m s^{-1}).

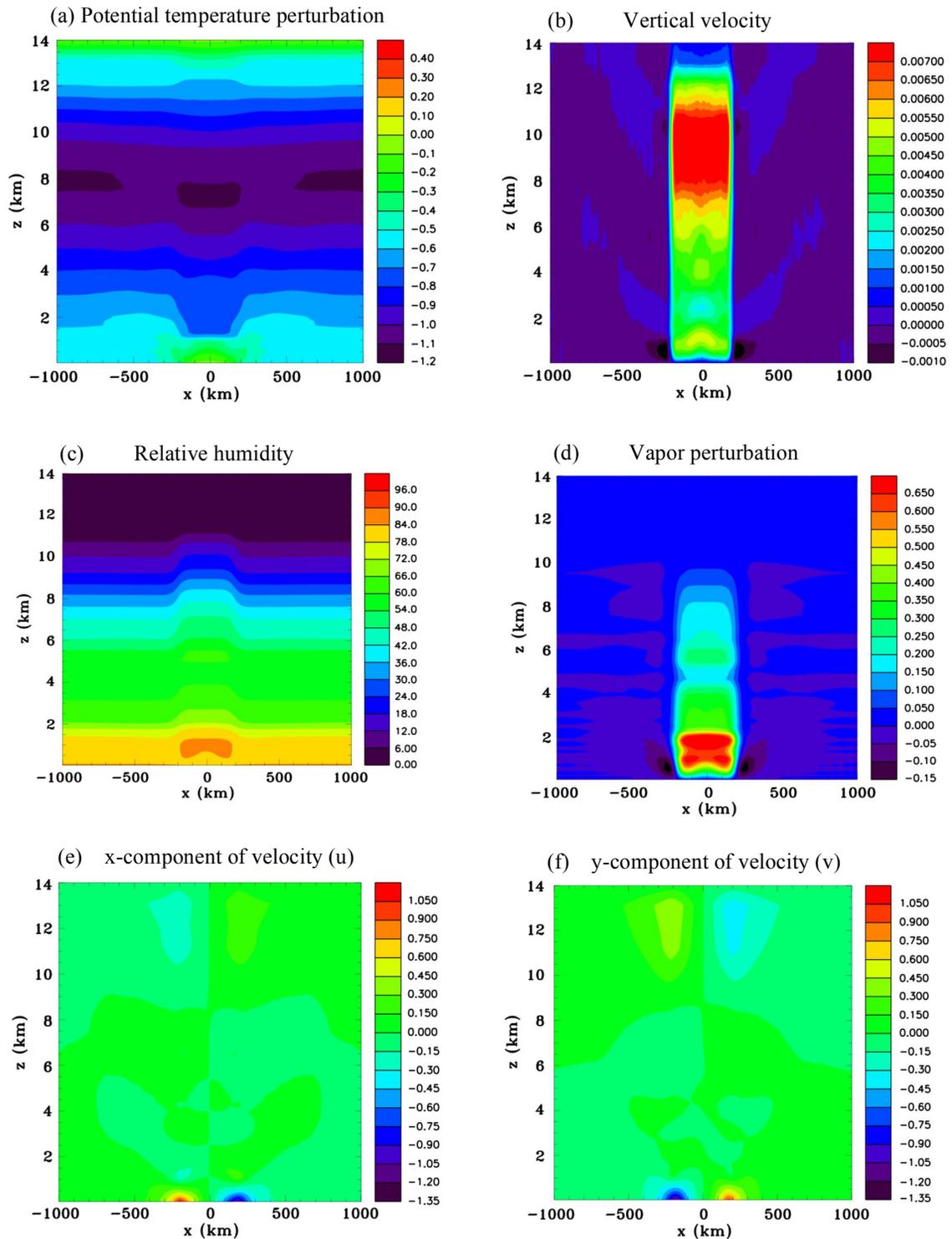


Figure 2. Vertical sections for Experiment 1: the maximum cooling aloft case, at $t=12$ h. **(a)** Potential temperature perturbation (K), **(b)** vertical velocity (m s^{-1}), **(c)** Relative humidity **(d)** Vapor perturbation (g kg^{-1}), **(e)** x-component of velocity, u (m s^{-1}), and **(f)** y-component of velocity, v (m s^{-1}).

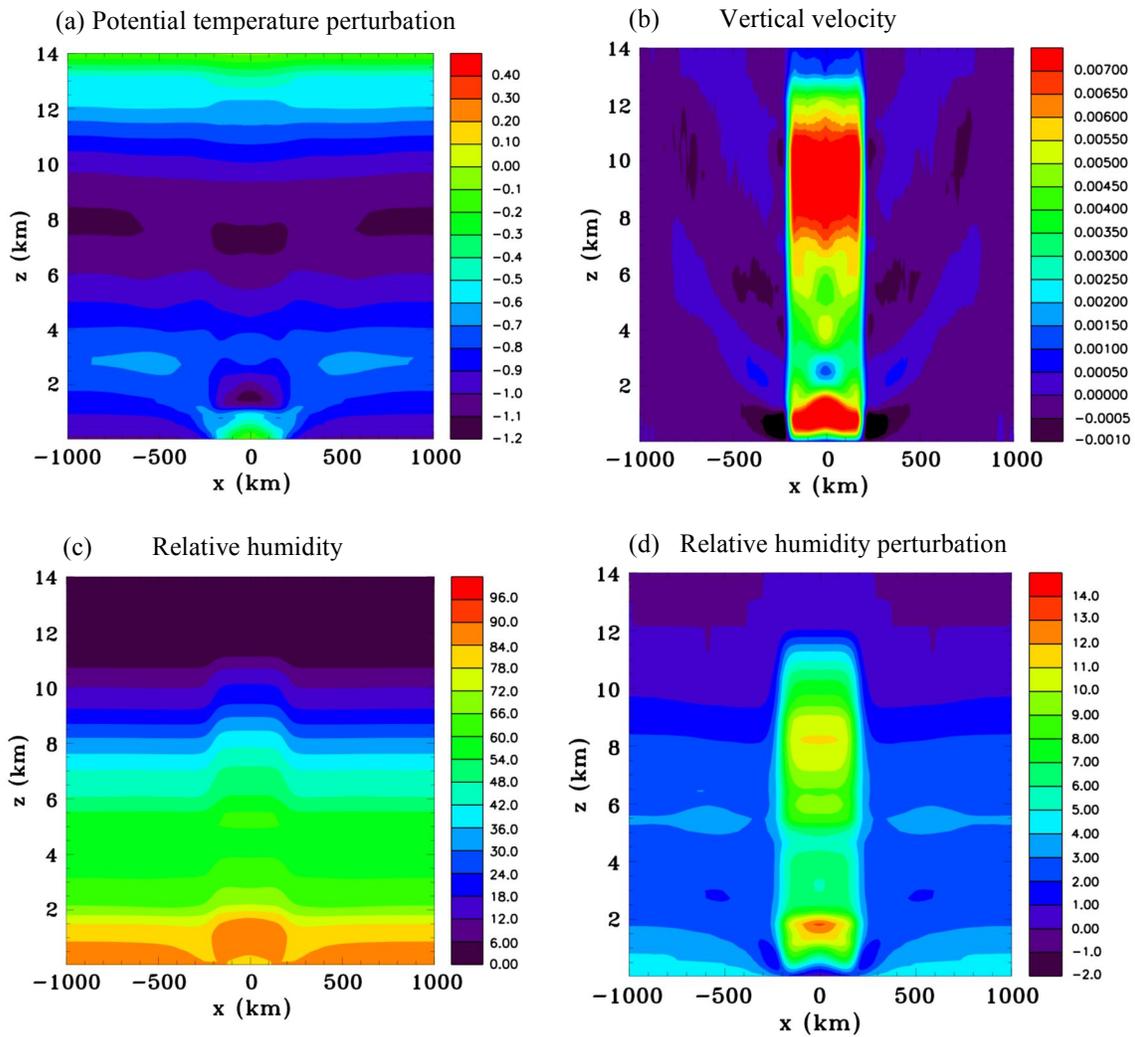


Figure 3. Vertical sections for Experiment 2: the maximum cooling aloft and weaker maximum at low levels case, at $t=12$ h. **(a)** Potential temperature perturbation (K), **(b)** vertical velocity (m s^{-1}), **(c)** relative humidity, and **(d)** relative humidity perturbation.

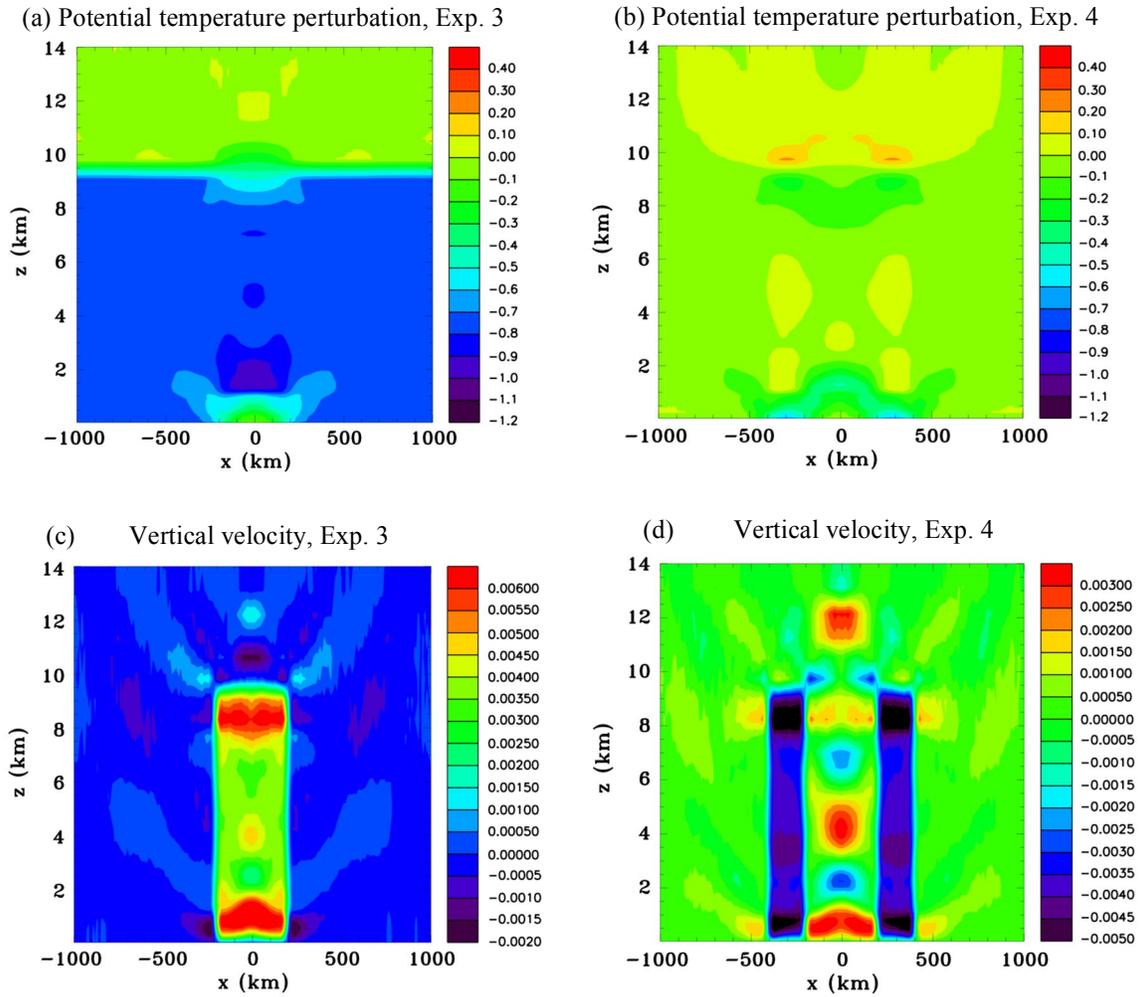


Figure 4. Vertical sections comparing Experiment 3: the uniform cooling case with Experiment 4, the small annulus case, at $t=12$ h. (a) and (b) Potential temperature perturbation (K), (c) and (d) vertical velocity (m s^{-1}), (e) and (f) Relative humidity, and (g) and (h) y-component of velocity, v (m s^{-1}), for Experiments 3 and 4, respectively.

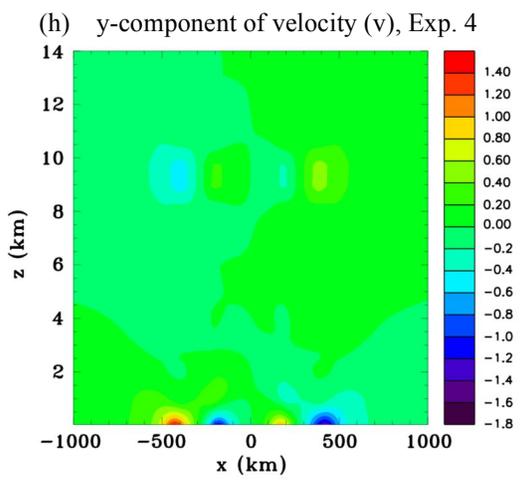
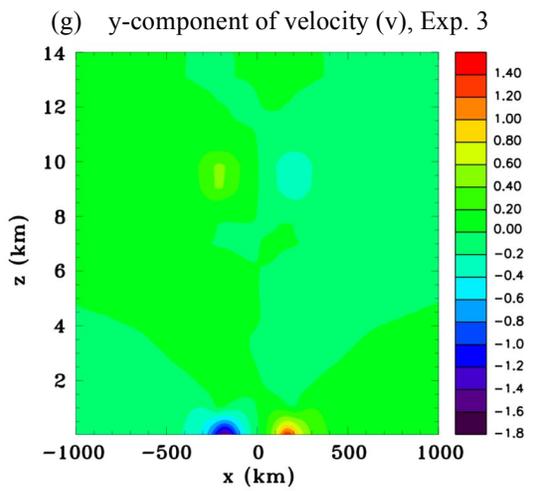
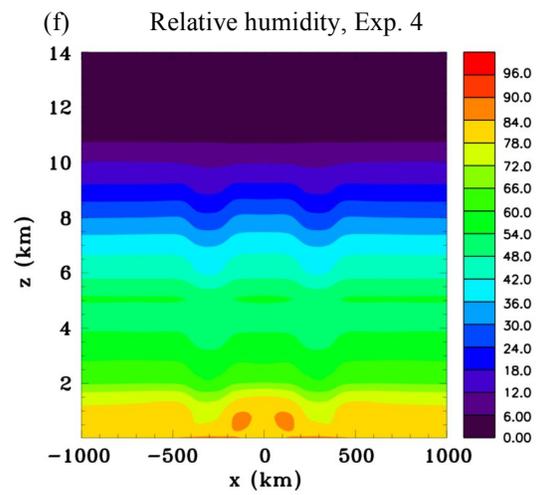
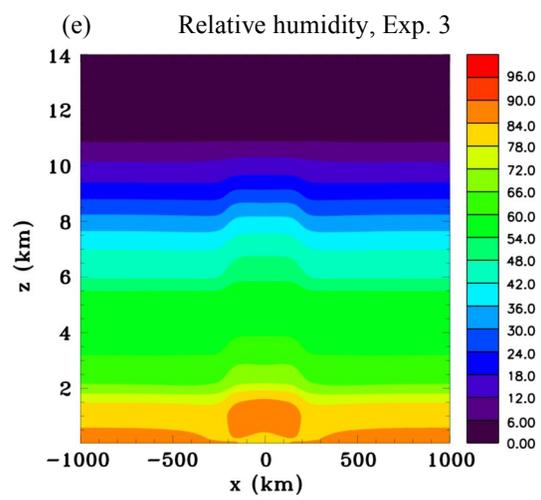


Figure 4 continued.

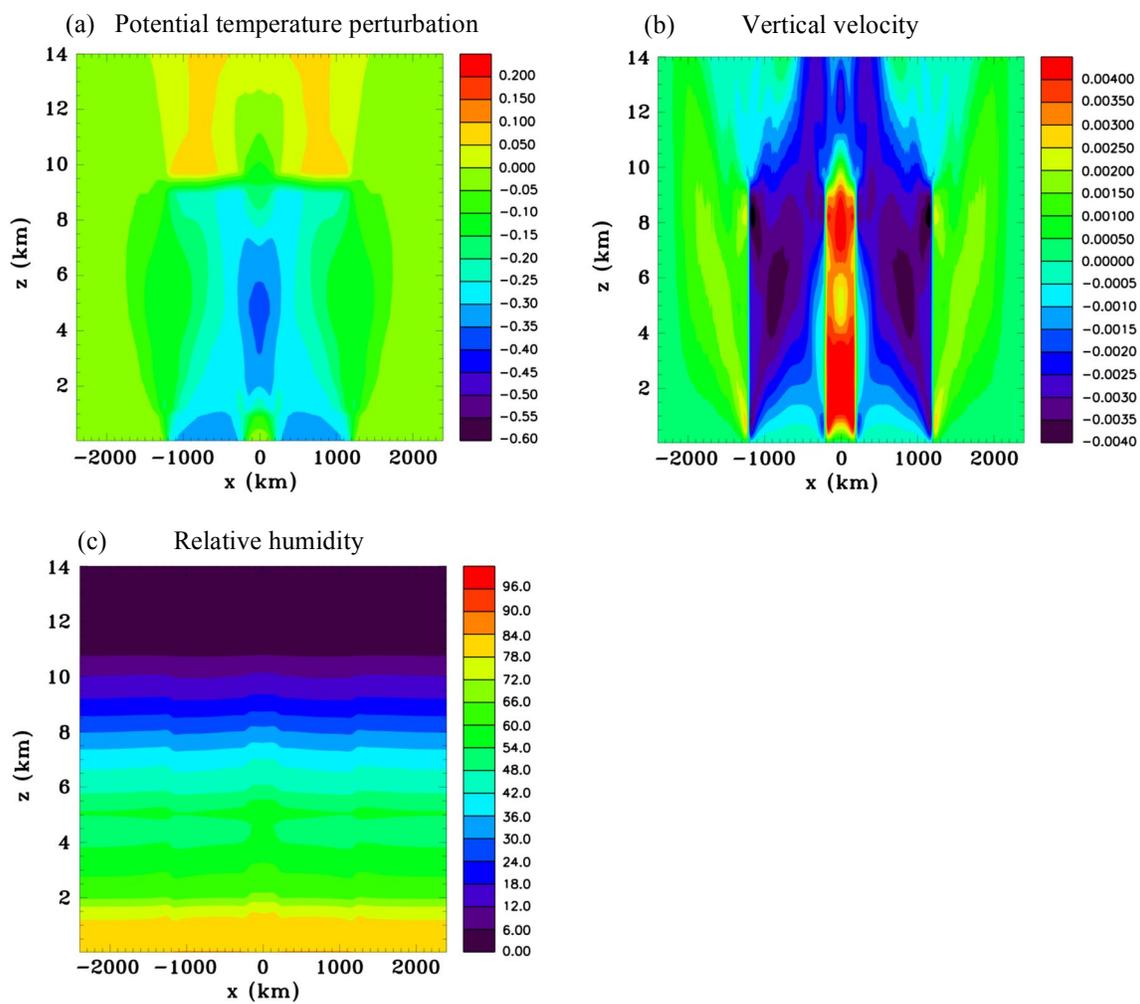


Figure 5. Vertical sections for Experiment 5: the large annulus case, at $t=5$ h. **(a)** Potential temperature perturbation (K), **(b)** vertical velocity (m s^{-1}), and **(c)** relative humidity.

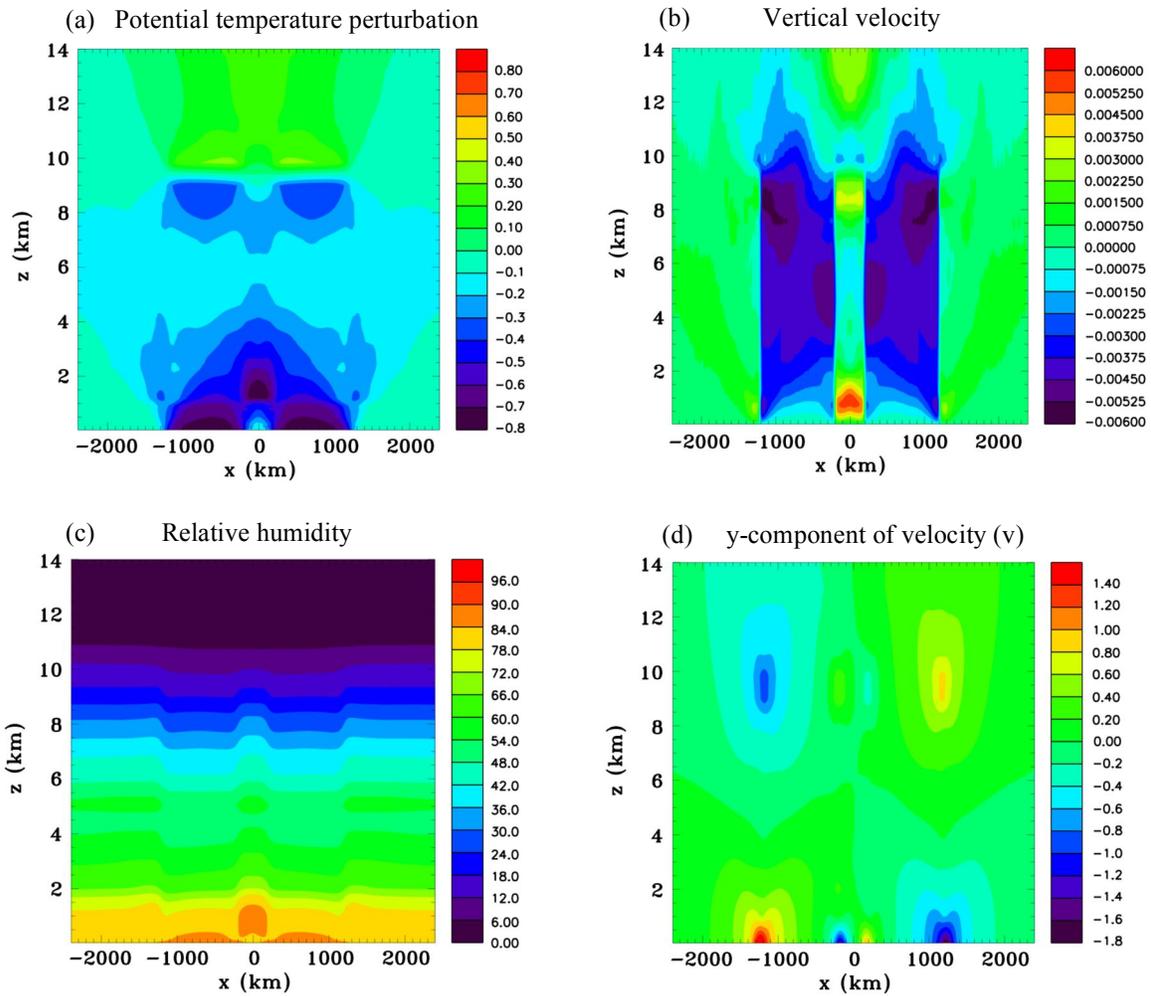


Figure 6. Vertical sections for Experiment 5: the large annulus case, at $t=12$ h. **(a)** Potential temperature perturbation (K), **(b)** vertical velocity (m s^{-1}), and **(c)** relative humidity.

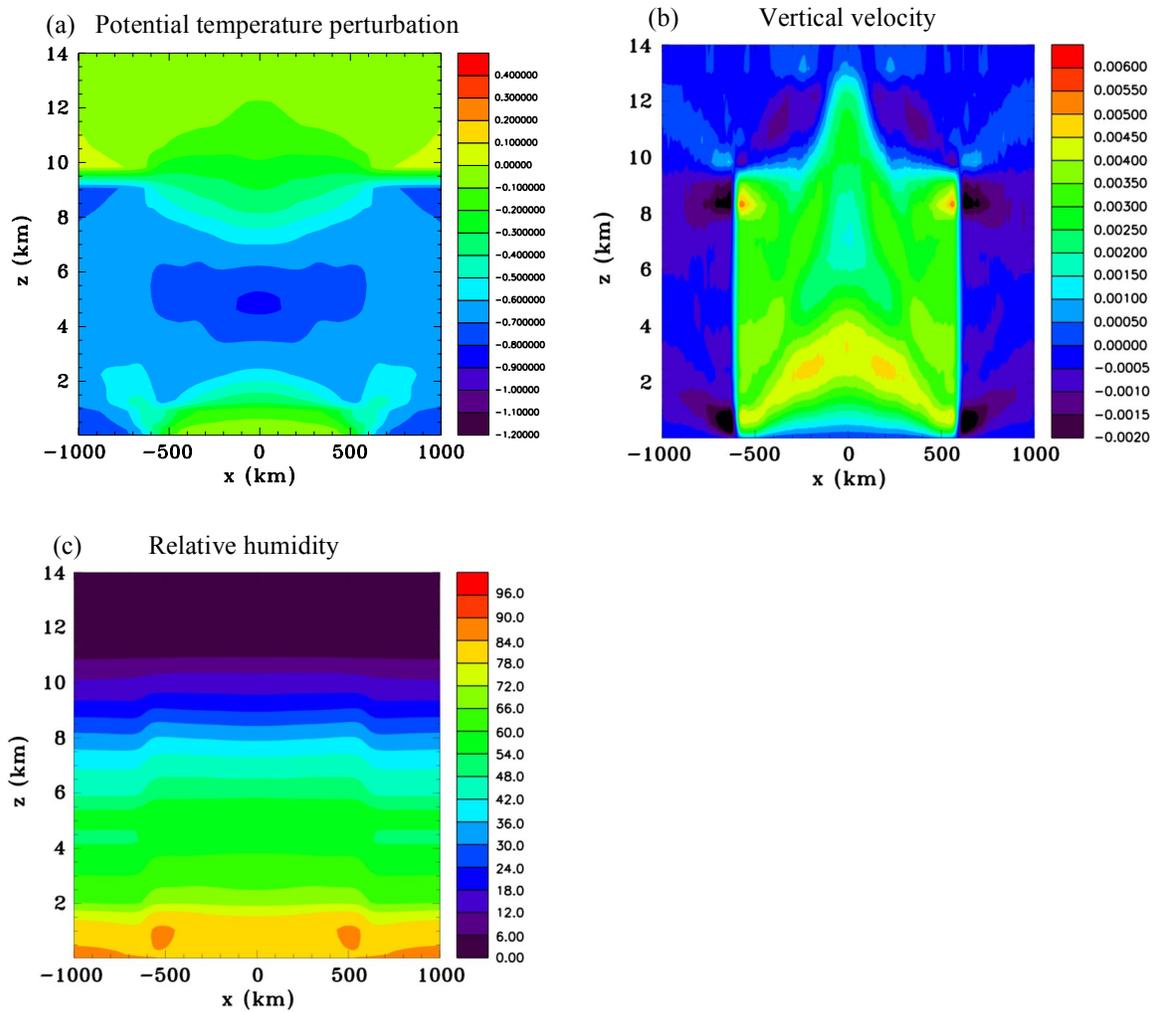


Figure 7. Vertical sections for Experiment 6: the wide unforced region case, at t=12 h. **(a)** Potential temperature perturbation (K), **(b)** vertical velocity (m s^{-1}), and **(c)** relative humidity.

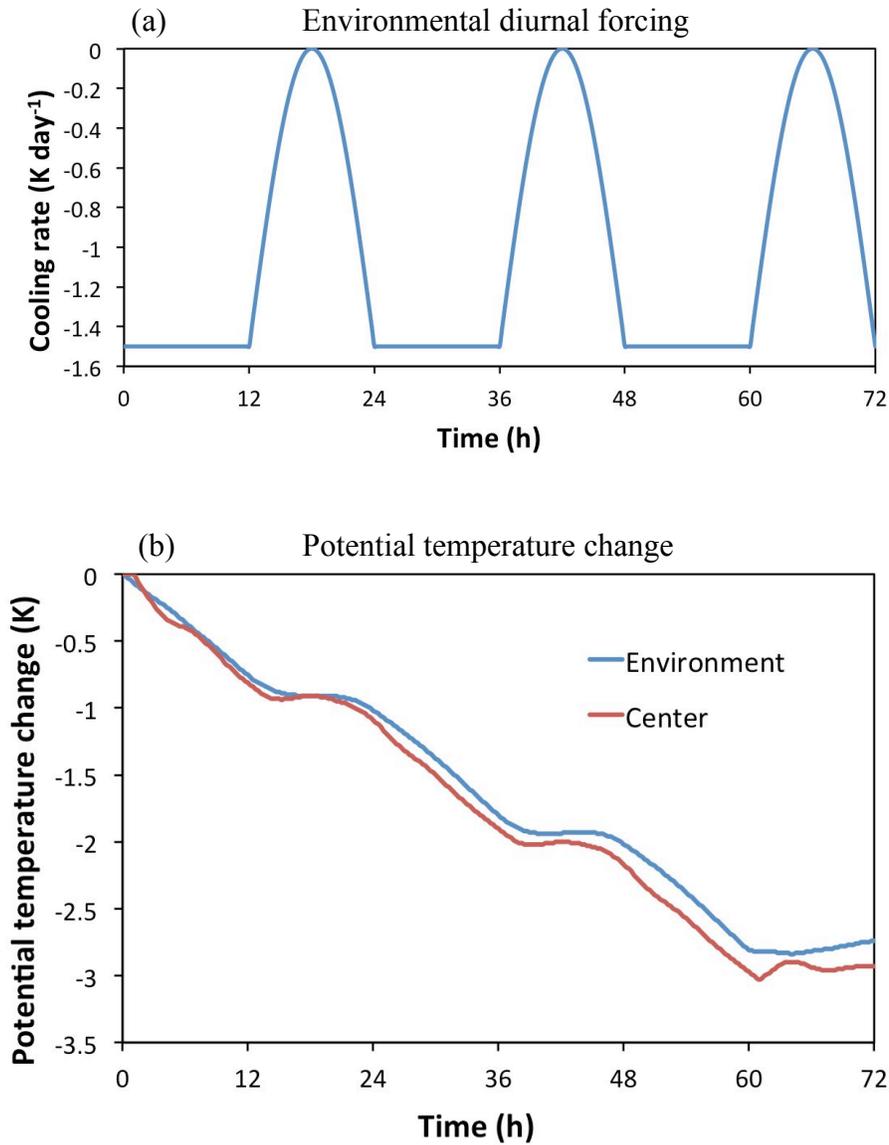


Figure 8. Time series for Experiment 7: the idealized diurnal forcing case. **(a)** Environmental forcing for $r > 200$ km, $z < 10$ km, and **(b)** potential temperature change in the environment at $r = 660$ km and $z = 5$ km, and at the center of the unforced region at $z = 5$ km.

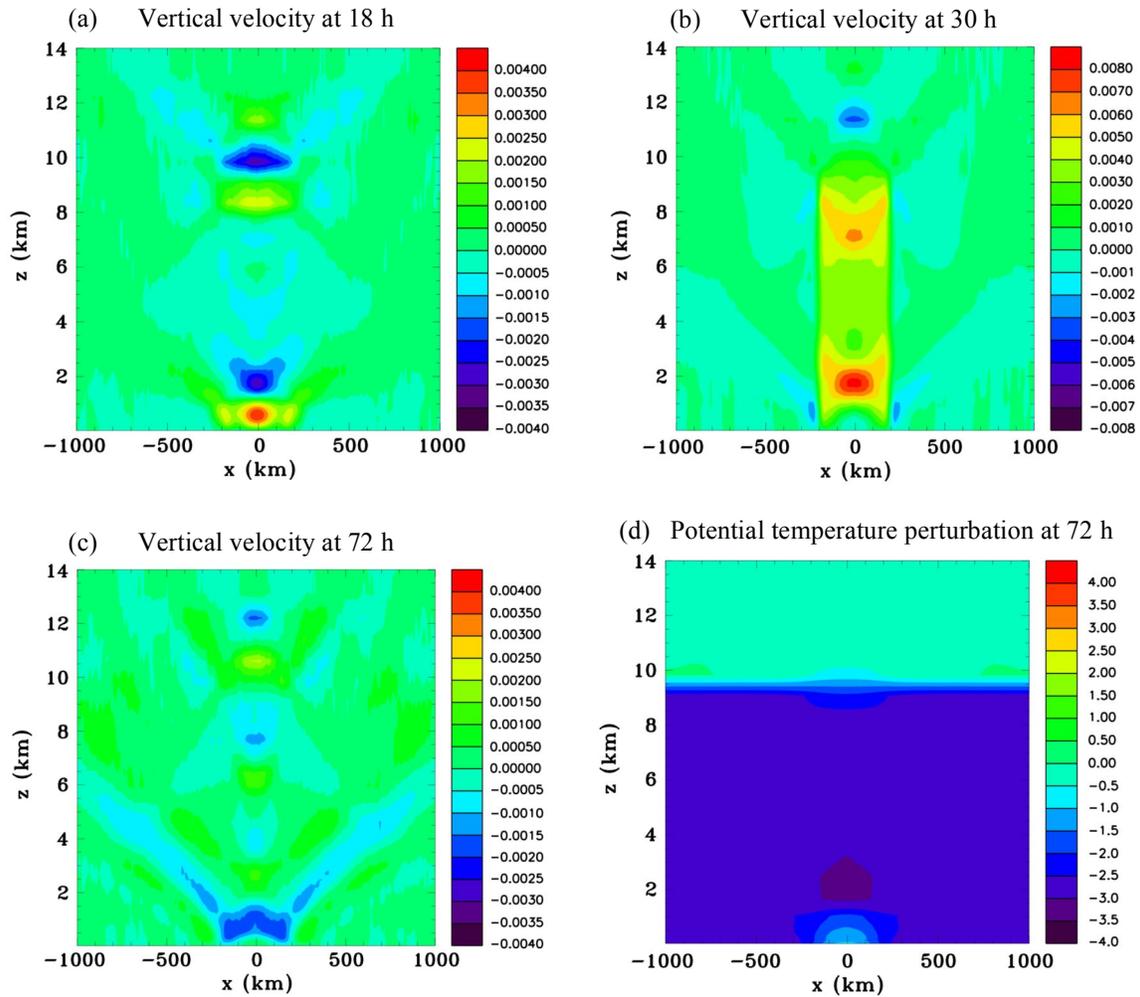


Figure 9. Vertical sections for Experiment 7: the idealized diurnal forcing case. **(a)** Vertical velocity at $t=18$ h (m s^{-1}), **(b)** vertical velocity at $t=30$ h (m s^{-1}), **(c)** vertical velocity at $t=72$ h (m s^{-1}), **(d)** potential temperature perturbation at $t=72$ h (K), **(e)** relative humidity at $t=72$ h, **(f)** y-component of velocity, v at $t=72$ h (m s^{-1}), **(g)** x-component of velocity, u at $t=72$ h (m s^{-1}), and **(h)** relative humidity perturbation at $t=72$ h.

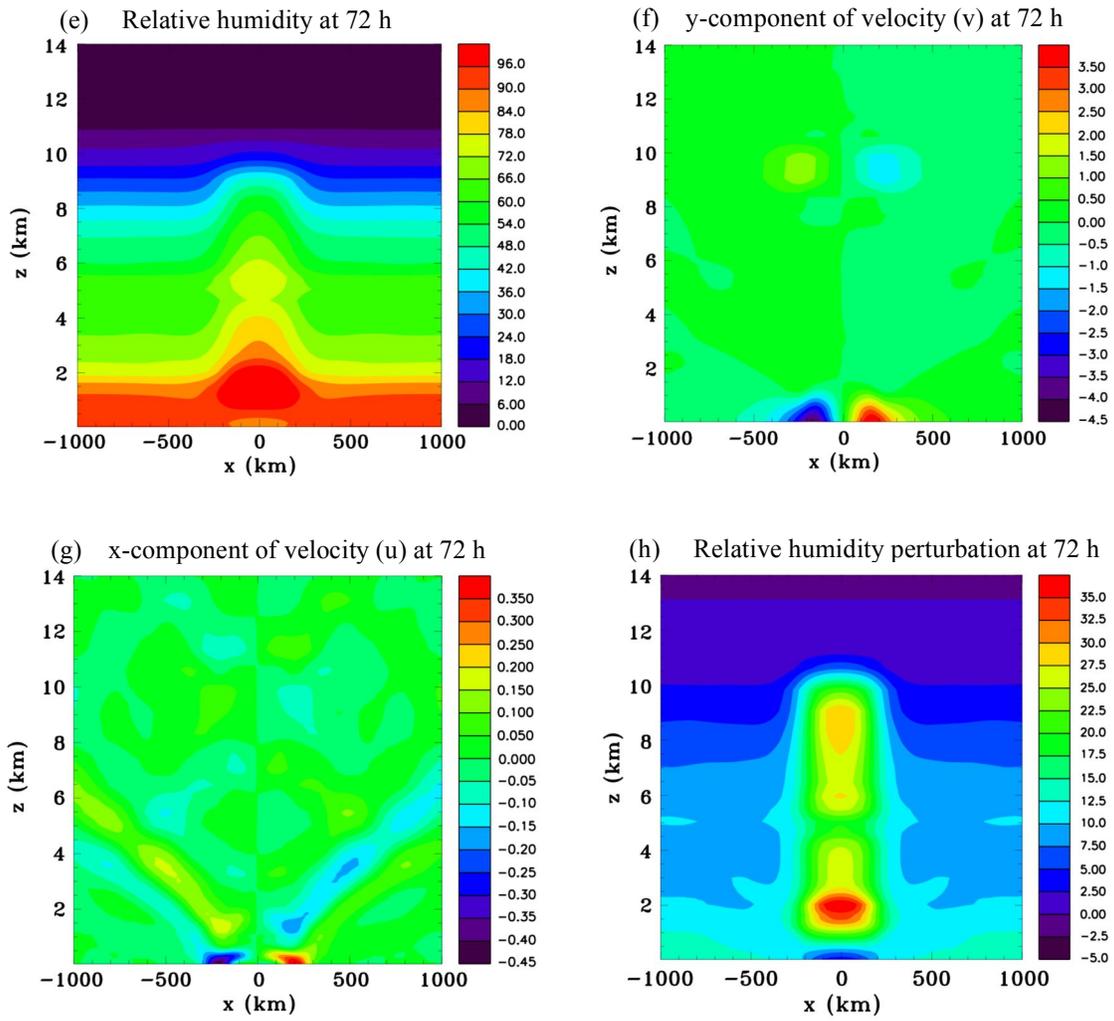


Figure 9 continued.

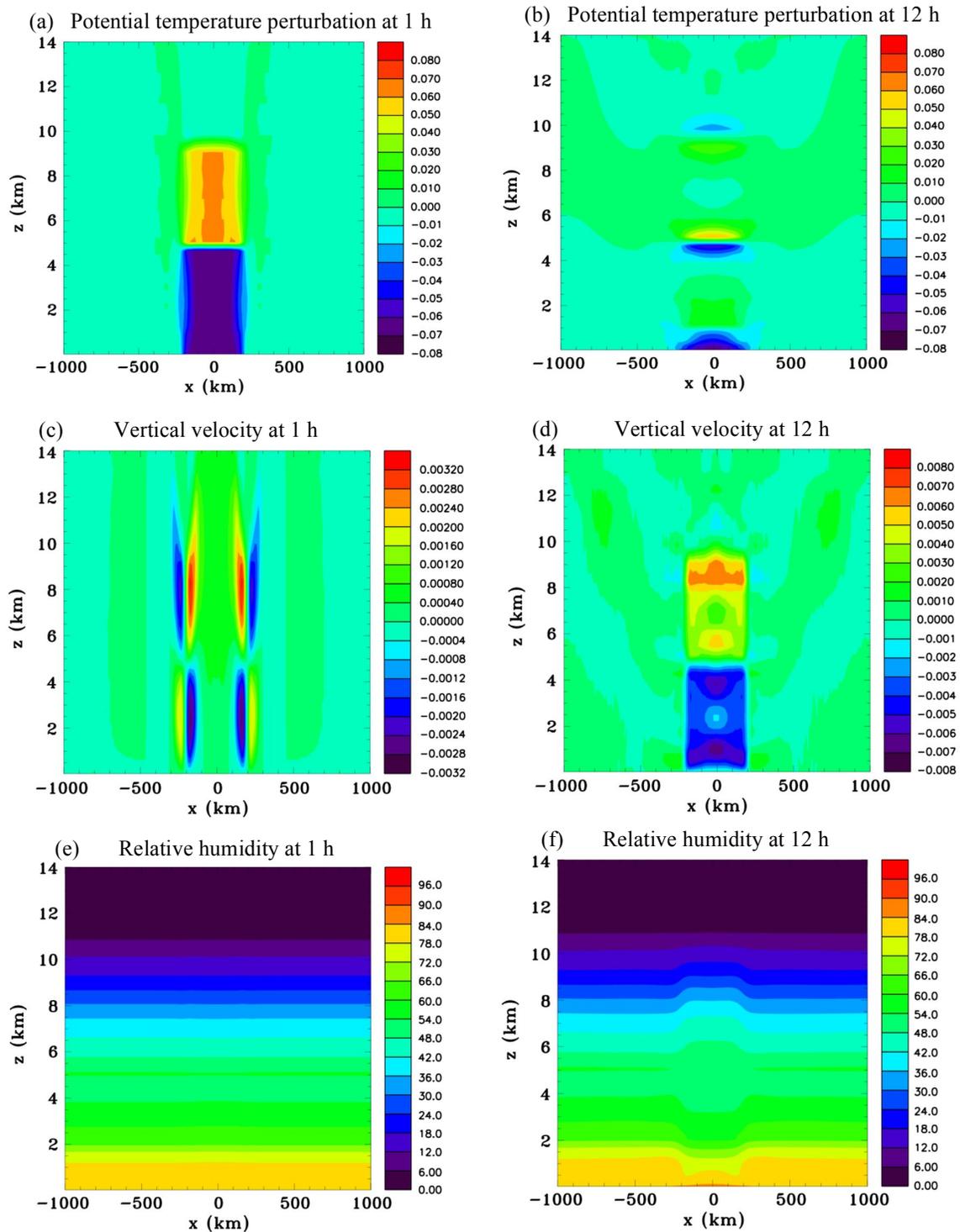


Figure 10. Vertical sections for Experiment 8: the forced core case at $t=1\text{h}$ and 12h . (a) and (b) Potential temperature perturbation (K), (c) and (d) vertical velocity (m s^{-1}), and (e) and (f) relative humidity, at $t=1$ and 12h , respectively.

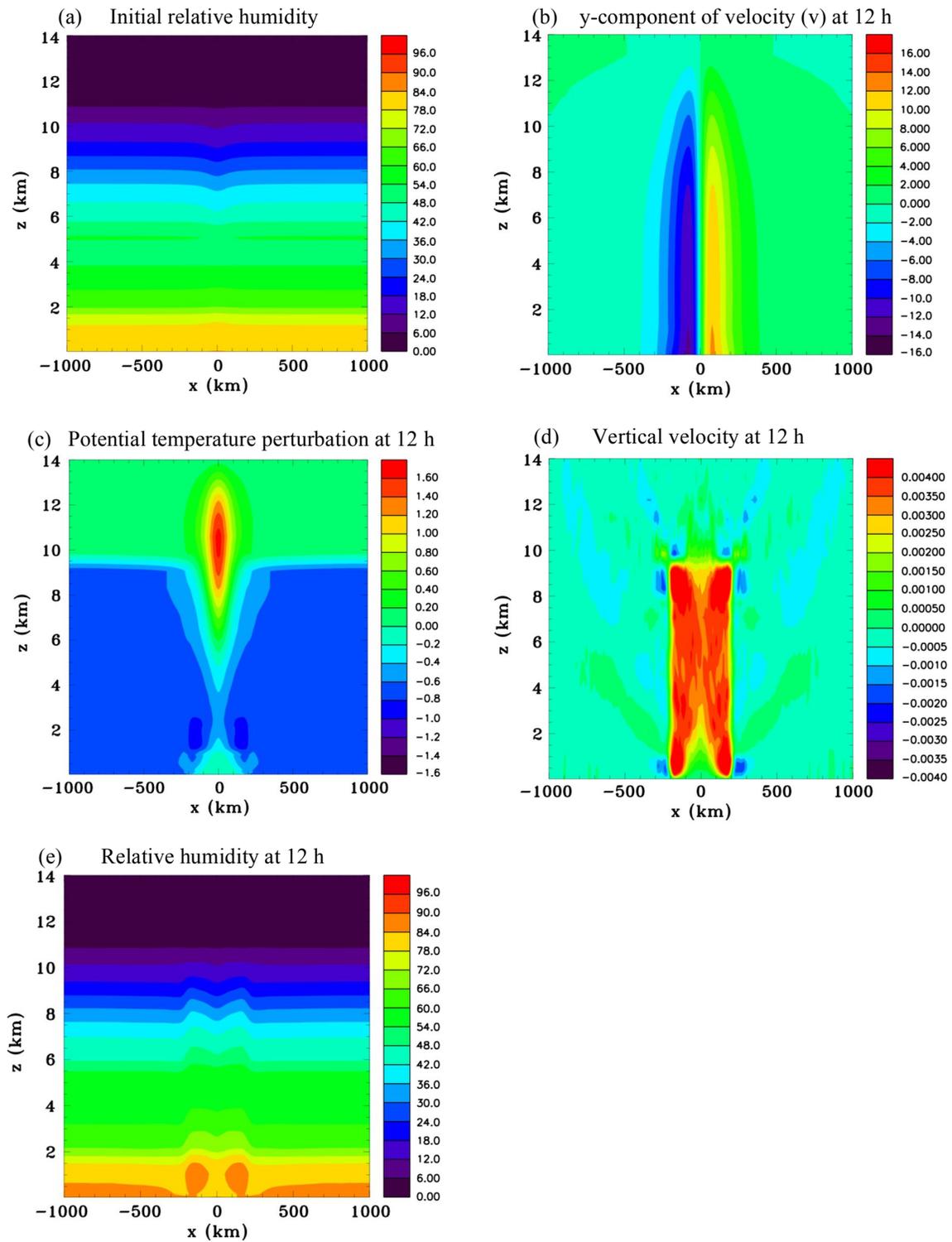


Figure 11. Vertical sections for Experiment 9: the weak vortex case. **(a)** Relative humidity at $t=0$ h, **(b)** y-component of velocity, v at $t=12$ h (m s^{-1}), **(c)** potential temperature perturbation at $t=12$ h (K), **(d)** vertical velocity at $t=12$ h (m s^{-1}), and **(e)** relative humidity at $t=12$ h.

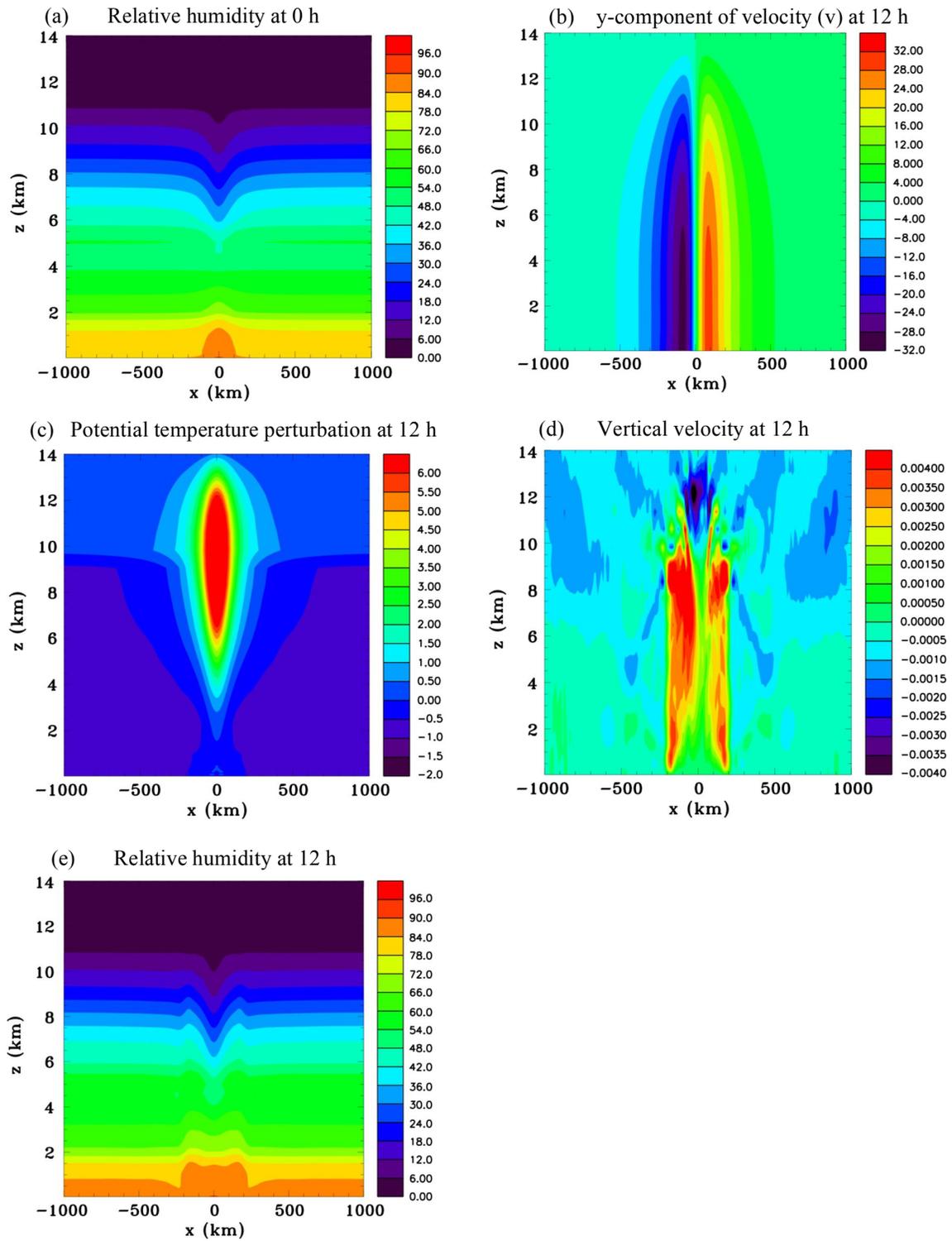


Figure 12. Vertical sections for Experiment 10, the strong vortex case. **(a)** Relative humidity at $t=0$ h, **(b)** y-component of velocity, v at $t=12$ h (m s^{-1}), **(c)** potential temperature perturbation at $t=12$ h (K), **(d)** vertical velocity at $t=12$ h (m s^{-1}), and **(e)** relative humidity at $t=12$ h.

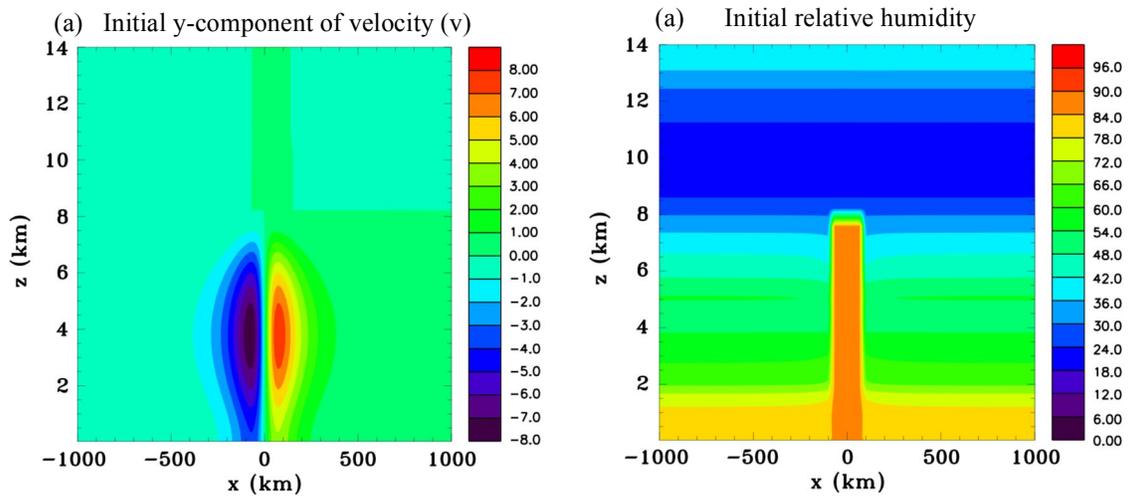


Figure 13. Vertical sections for Experiment 11: the radiation scheme activated in the whole domain with a mid-level vortex case. (a) y-component of velocity, v (m s^{-1}), and (b) relative humidity, at $t=0$ h.

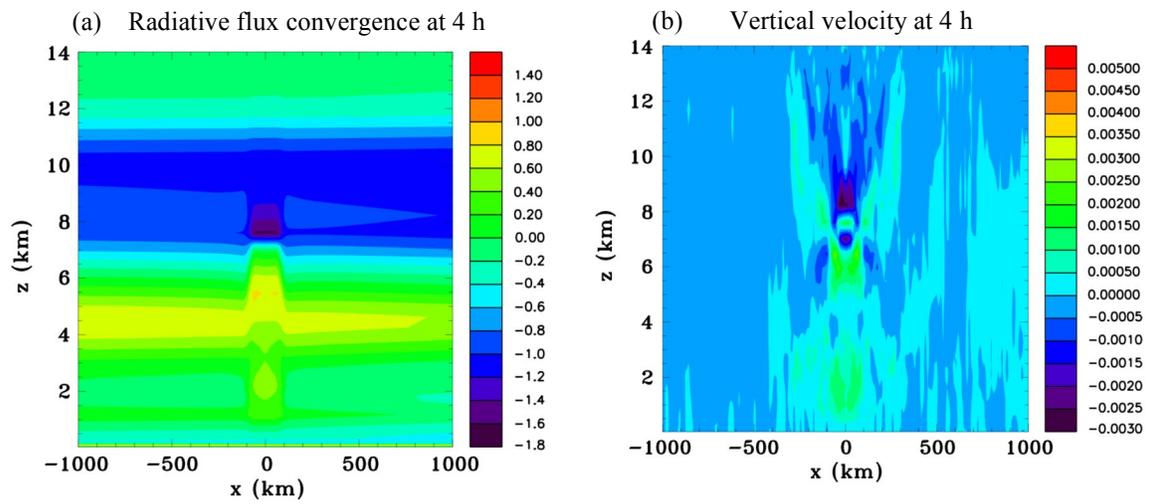


Figure 14. Vertical sections for Experiment 11: the radiation scheme activated in the whole domain with a mid-level vortex case. **(a)** Radiative flux convergence ($\text{K s}^{-1} \times 10^{-5}$), and **(b)** vertical velocity (m s^{-1}), at $t=4$ h, during the day.

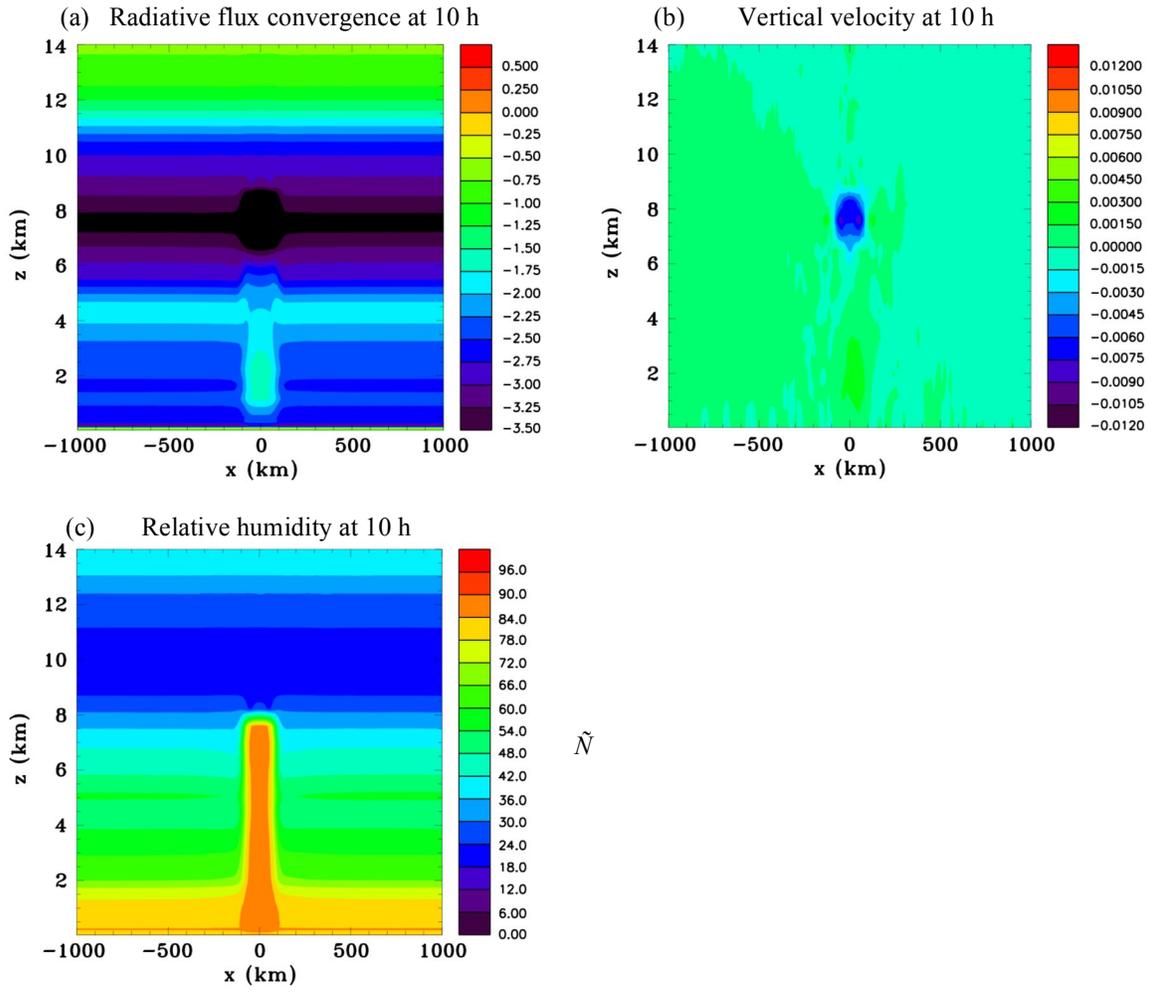


Figure 15. Vertical sections for Experiment 11: the radiation scheme activated in the whole domain with a mid-level vortex case. (a) Radiative flux convergence ($\text{K s}^{-1} \times 10^{-5}$), (b) vertical velocity (m s^{-1}), (c) relative humidity, at $t=10$ h, during the night.

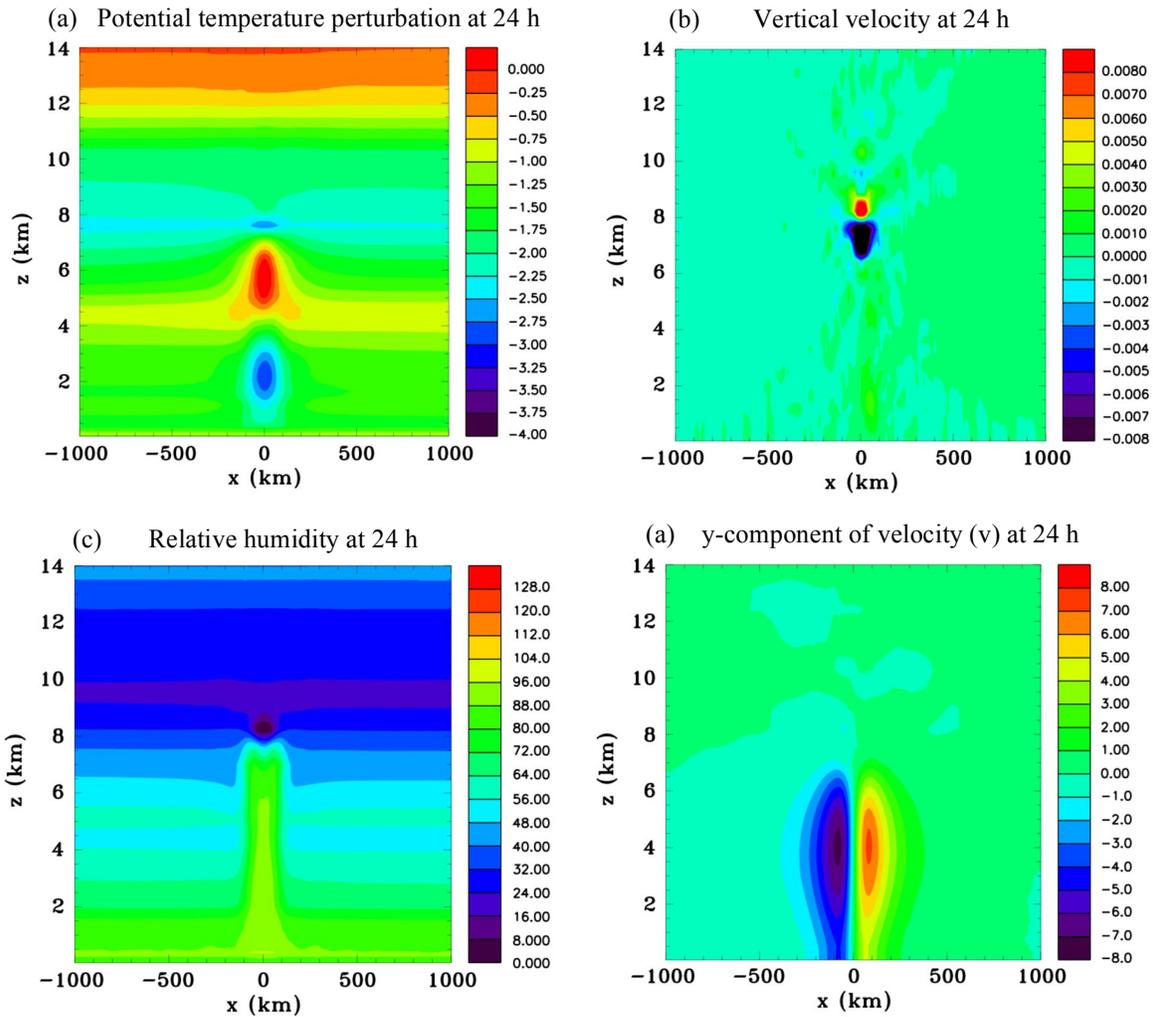


Figure 16. Vertical sections for Experiment 11: the radiation scheme activated in the whole domain with a mid-level vortex case. (a) Potential temperature perturbation (K), (b) vertical velocity (m s^{-1}) (c) relative humidity, and (d) y-component of velocity, v (m s^{-1}), at $t=24$ h, in mid-morning.

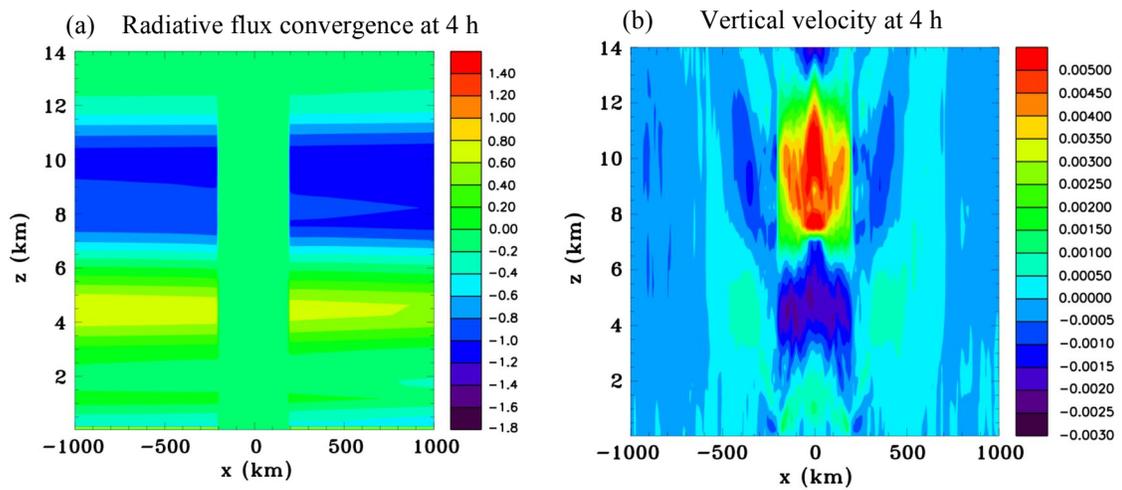


Figure 17. Vertical sections for Experiment 12: the radiation scheme activated in the environment, $r > 200$ km. (a) Radiative flux convergence ($\text{K s}^{-1} \times 10^{-5}$), and (b) vertical velocity (m s^{-1}), at $t=4$ h.

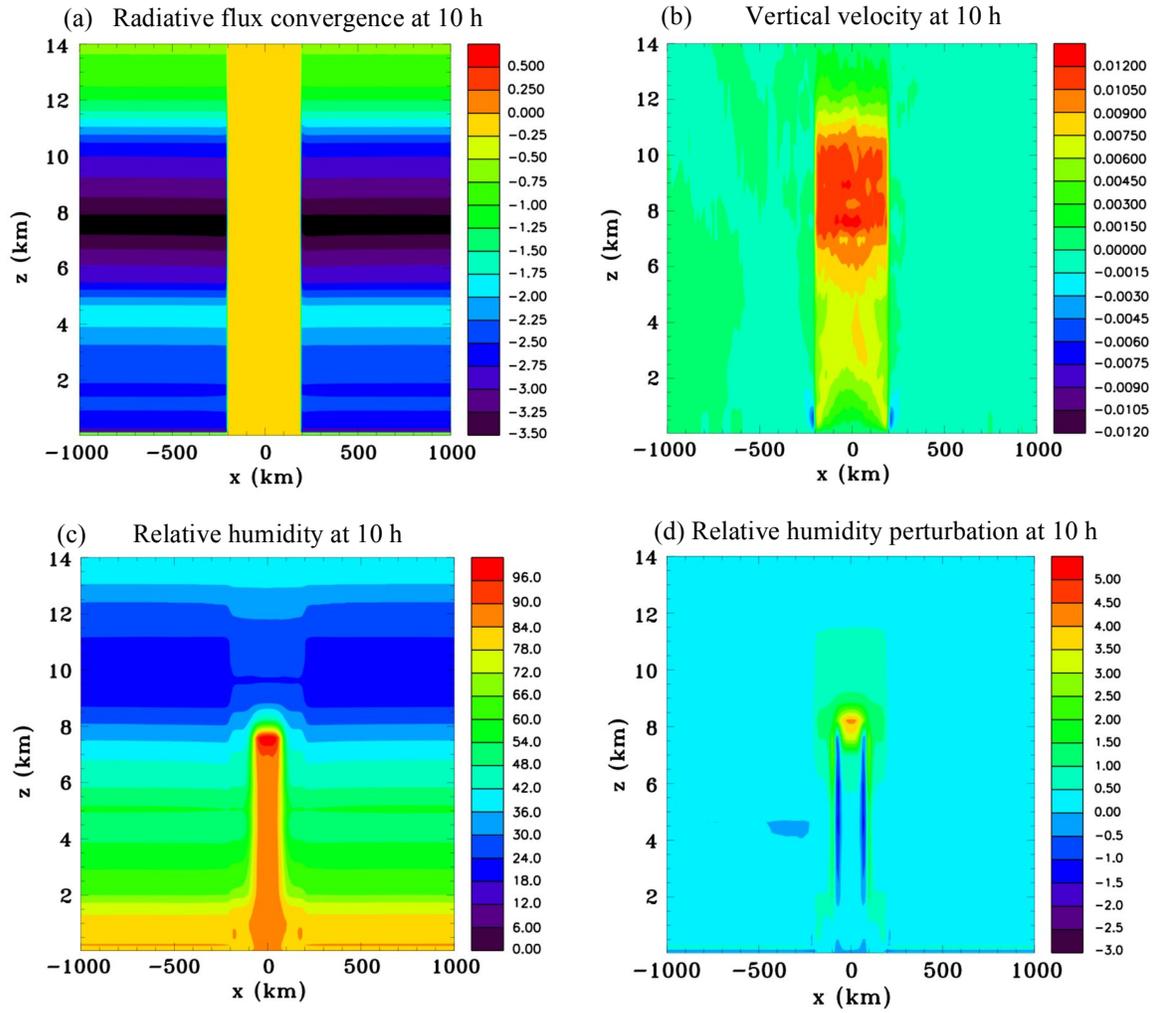


Figure 18. Vertical sections for Experiment 12: the radiation scheme activated in the environment, $r > 200$ km. (a) Radiative flux convergence ($\text{K s}^{-1} \times 10^{-5}$), (b) vertical velocity (m s^{-1}), (c) relative humidity, and (d) relative humidity perturbation, at $t=10$ h.

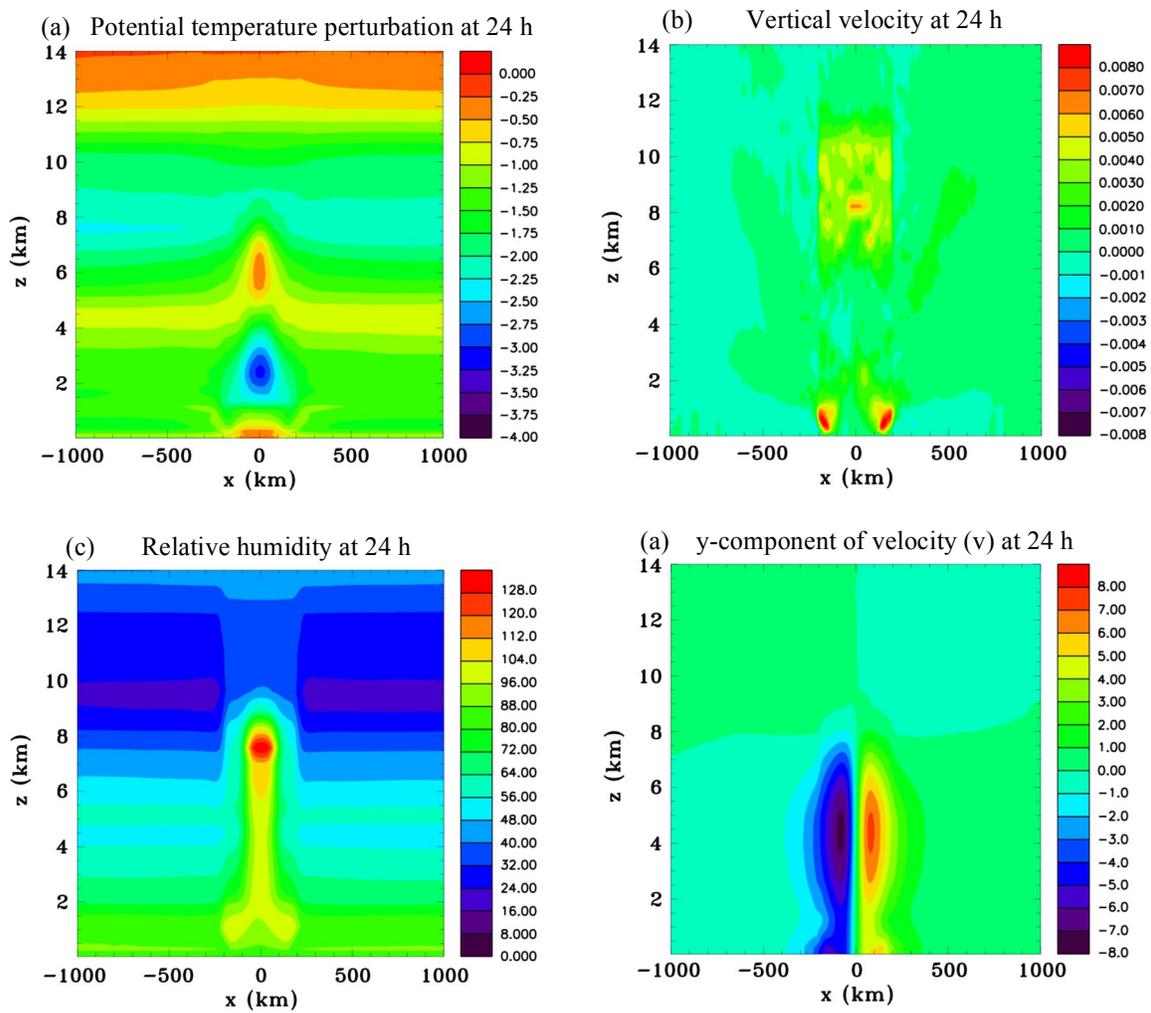


Figure 19. Vertical sections for Experiment 12: the radiation scheme activated in the environment, $r > 200$ km. (a) Potential temperature perturbation (K), (b) vertical velocity (m s^{-1}) (c) relative humidity, and (d) y-component of velocity, v (m s^{-1}), at $t = 24$ h.

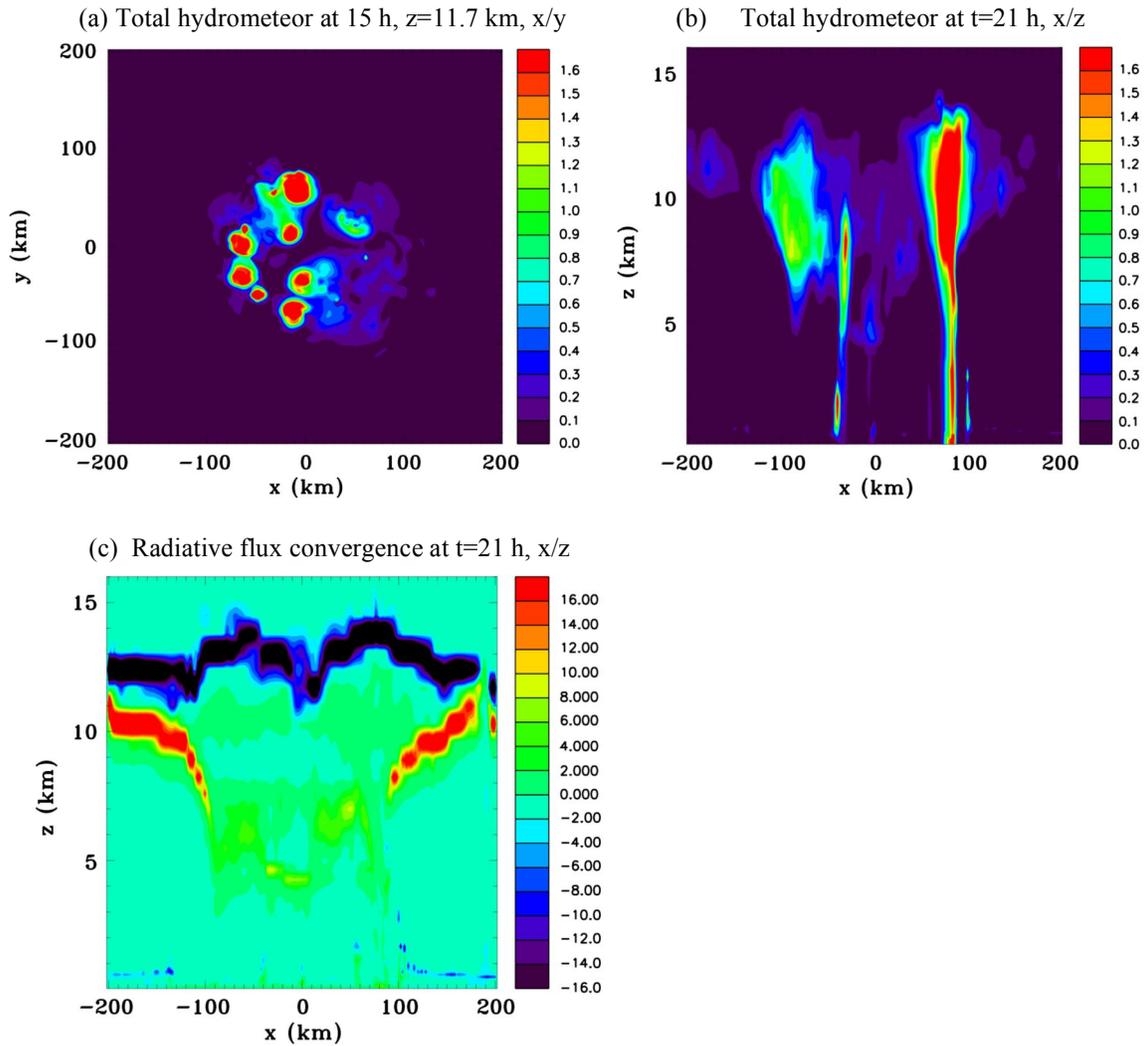


Figure 20. Experiment 13: Full physics with radiation in the whole domain. (a) Horizontal section of total hydrometeor mixing ratio at $z=11.7$ km, $t=15$ h ($g\ kg^{-1}$), (b) vertical section of total hydrometeor mixing ratio at $t=21$ h ($g\ kg^{-1}$), and (c) vertical section of radiative flux convergence at $t=21$ h ($\times 10^{-5}\ K\ s^{-1}$).

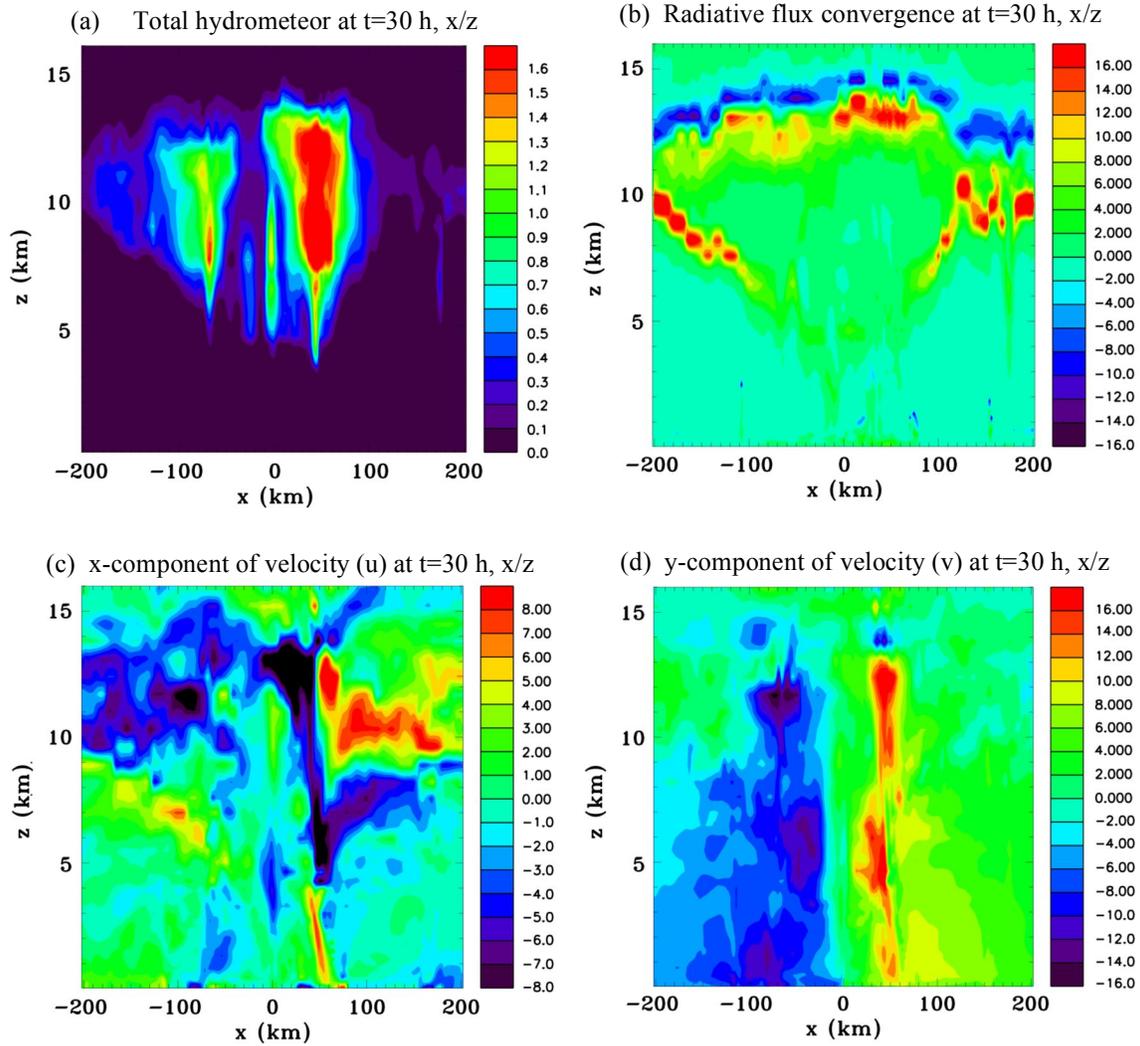


Figure 21. Experiment 13: Full physics with radiation in the whole domain. Vertical sections at t=30 h of (a) total hydrometeor mixing ratio (g kg^{-1}), (b) radiative flux convergence at t=21 h ($\times 10^{-5} \text{ K s}^{-1}$), (c) x-component of velocity, u (m s^{-1}), and (d) y-component of velocity, v (m s^{-1}).

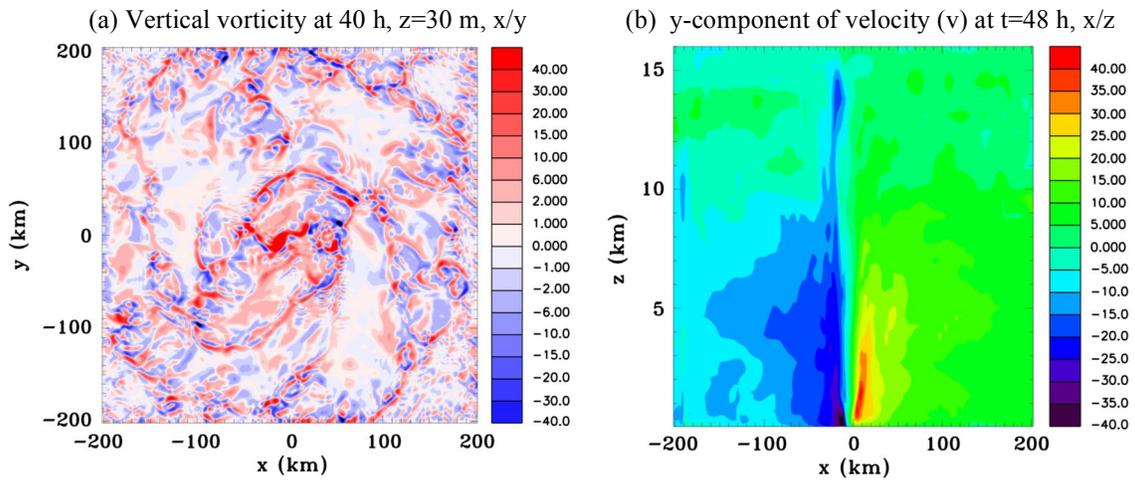


Figure 22. Experiment 13: Full physics with radiation in the whole domain. **(a)** Horizontal section of surface vertical vorticity, at $t=40$ h ($\times 10^{-4}$ rad s^{-1}), and **(b)** vertical section of y-component of velocity, v , at 48 h ($m s^{-1}$).

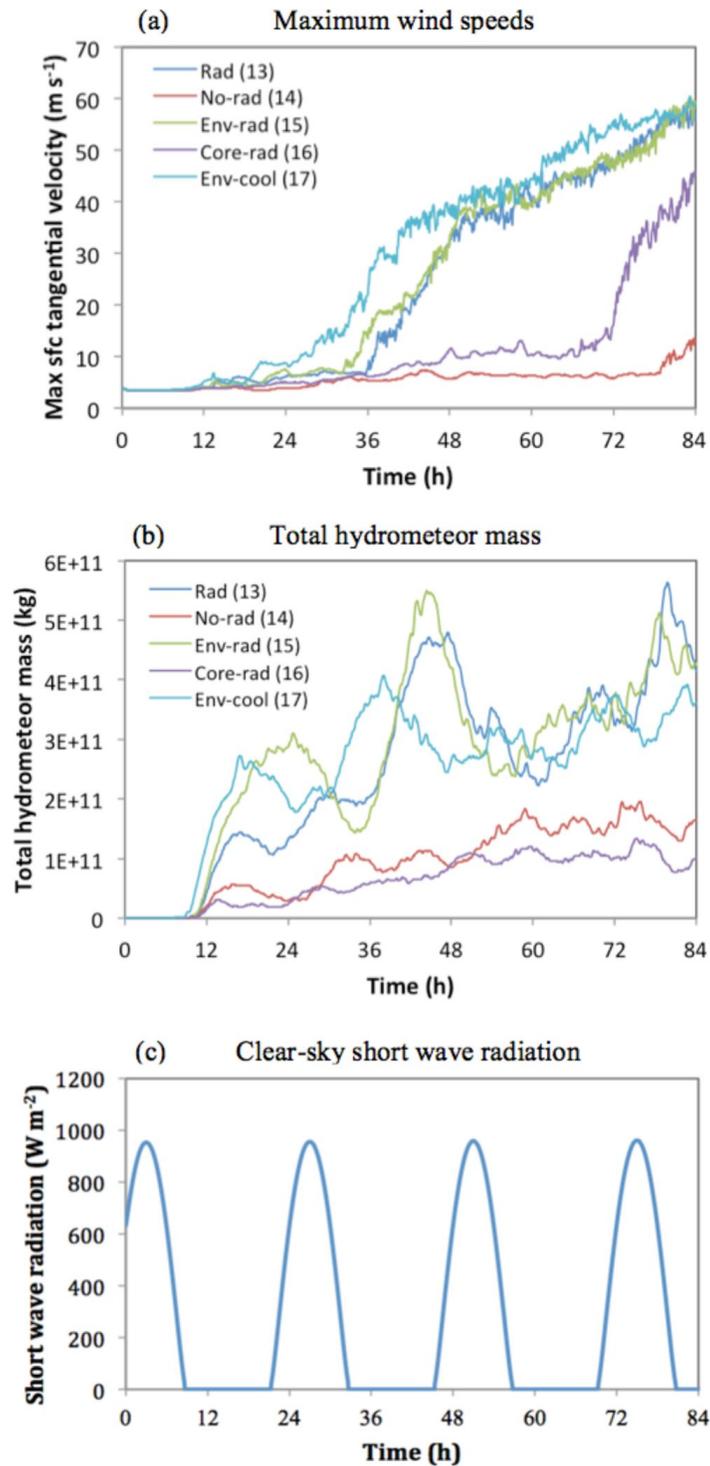


Figure 23. Time series for Experiments 13-17: (a) Maximum azimuthally averaged tangential wind speeds at $z = 29.5$ m (m s^{-1}), (b) Total hydrometeor mass in the domain (kg), and (c) clear-sky short wave radiation at the surface (W m^{-2}).

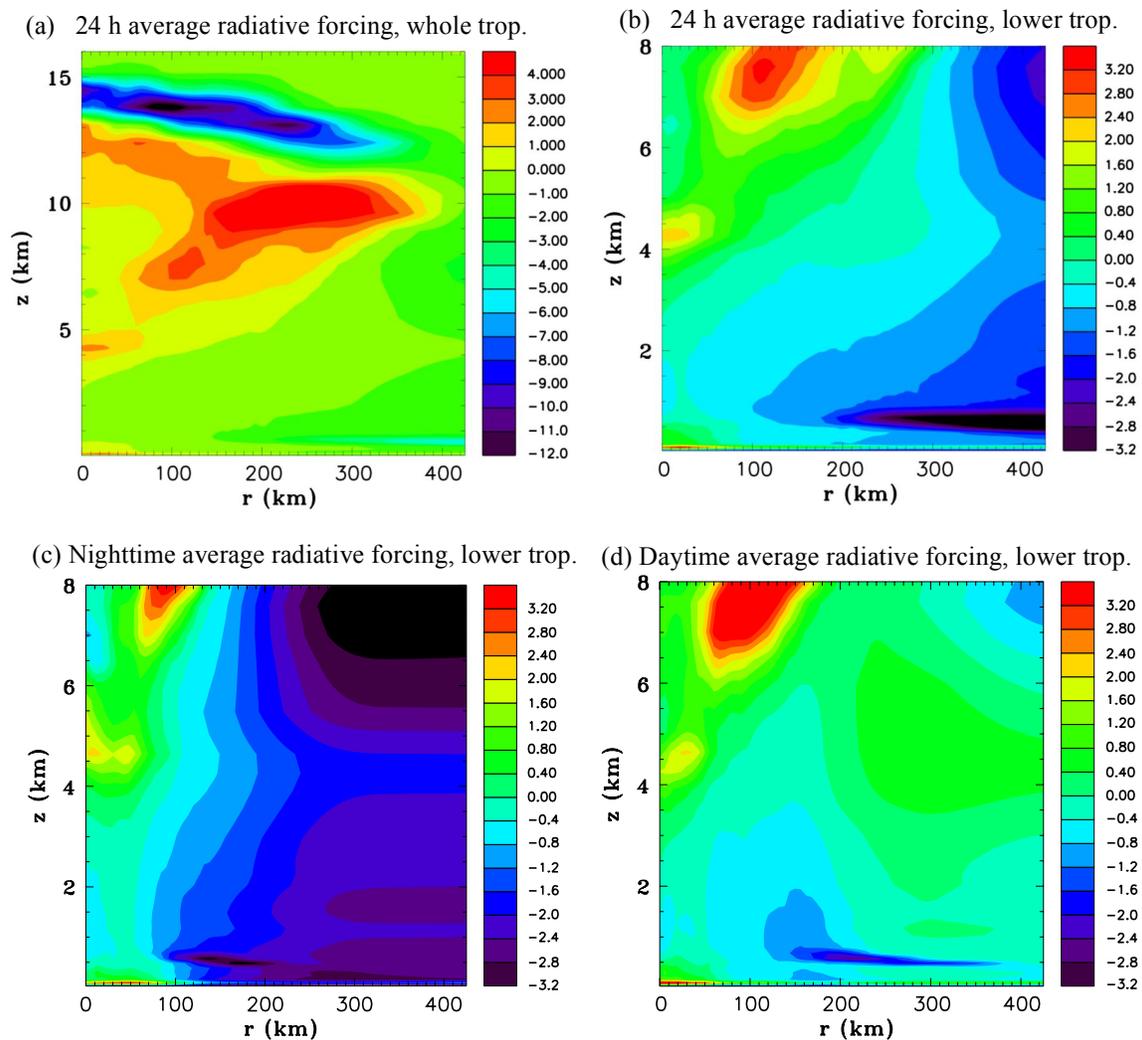


Figure 24. Time and azimuthally averaged radiative flux convergence for Experiment 13: the full physics simulation with radiation activated in the whole domain case. (a) 24 h average from 24 to 48 h shown for the whole troposphere, (b) 24 h average from 24 to 48 h shown for the lower troposphere, (c) 6 h nighttime average from 15 to 21 h shown for the lower troposphere, and (d) 6 h daytime average from 24 to 30 h shown for the lower troposphere, ($\times 10^{-5} \text{ K s}^{-1}$).

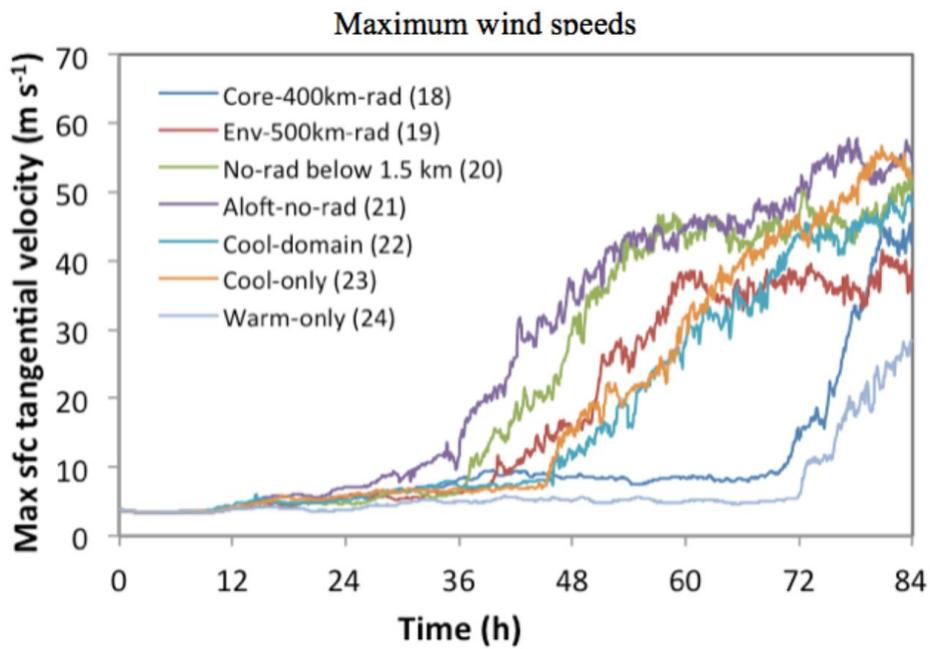


Figure 25. Time series of maximum azimuthally averaged tangential wind speeds at $z=29.5$ m (m s^{-1}), for Experiments 18-24.