ACPD-15-8479-2015
Responses to Reviewer 1 (Prof. R. Schumacher)
Date: 9 July 2015

Title: A numerical study of convection in rainbands of Typhoon Morakot (2009) with extreme rainfall: roles of pressure perturbations with low-level wind maxima
Authors: C.-C. Wang, H.-C. Kuo, R. H. Johnson, C.-Y. Lee, S.-Y. Huang, and Y.-H. Chen

1. General comments:

This study uses dynamical diagnosis of numerical model simulations to understand convective processes in the rainbands of Typhoon Morakot (2009). In particular, the authors focus on the behavior of updrafts in the strong low-level vertical wind shear in the rainbands. They conclude that rather than being forced by lift along a cold pool, the location of ascent associated with the updraft/shear configuration is responsible for the back-building of convection in the rainbands.

The topic of the manuscript and the methods used are generally sound, and the paper is clearly written. The study did raise several questions for me that I think should be more thoroughly addressed in the manuscript, but I believe these bigger-picture issues still only require minor revisions. Therefore, I recommend that the manuscript be accepted if these minor revisions are sufficiently considered by the authors. I don’t need to see the manuscript again, but would be willing to review a revised version of the editor prefers it.

Reply:

We appreciate the positive views and critical comments from all three reviewers, and have revised the paper accordingly. Among the changes, we have (1) added the diagnostic results at 0645 UTC (besides 0630 UTC) to show a dominant and persistent effect from the dynamical pressure perturbation in the mature cell, (2) employed 10-min radar CAPPI data at 3 km to show more clearly the merging and back-building behavior in the observation, and (3) estimated the contribution from convection versus stratiform clouds over the plains in the event. In addition, the figures are polished and font sizes enlarged, the method of diagnosis is validated, the scale of the low-level jet is clarified, the cold pool is examined, and the evolutions of model convective cells are discussed in more detail, as suggested.

The changes in the manuscript are marked in red, blue, and green for Reviewer 1, Reviewer 2, and Reviewer 3, respectively. The modifications made by ourselves during the revision are in orange (mostly to correct mistakes), and those made during the production stage of ACPD since our first submission (to meet the format requirements) are in pink. The point-by-point responses to each of the comments/suggestions from this reviewer are listed below.

2. "Big-picture" comments:

1) This study uses diagnostics originally developed for understanding supercell thunderstorms to look at convection in the outer rainbands of a tropical cyclone. In fact, some past studies have shown that this convection is indeed supercellular, e.g., Eastin and Link (2009); Morin and Parker (2011); see also the references therein. I think it would be useful to establish whether the convection in Morakot was consistent with supercell dynamics, or whether these convective cells were non-supercellular but potentially still explained by these diagnostic methods.

Reply:

It is clarified that although this diagnostics was initially developed to study supercell storms, it is also shown to be valid and used for quasi-linear rainbands by Parker and Johnson (2004). In addition, it is mentioned that several past studies have suggested that some convection in TC rainbands are indeed supercellular, and the works of Eastin and Link (2009) and Morin and Parker (2011) are both cited.

2) One thing that wasn't totally clear to me in the manuscript was whether the "low-level jet" (LLJ) being discussed here was something external to the tropical cyclone, or the flow associated with the TC itself (or perhaps a bit of both). Perhaps the findings here could be compared to the conceptual model of Hence and Houze (2008) for rainbands in the "secondary horizontal wind maximum"?

Reply:

In the revision, it is clarified that the LLJ was a part of the TC circulation, but was also most likely enhanced by the southwesterly monsoon. The study of Hence and Houze (2008) is also cited as suggested.

3) Although from my reading of the manuscript, the use of the pressure perturbation diagnostics is interesting and applicable to this case, the manuscript may be more convincing if at least one more time is shown, such as 0645 or 0650 UTC when we can see how the detailed diagnostics shown for 0630 UTC related to the subsequent back-building convection. I don't think it's necessary to show 4 or 5 complete figures for the later time, maybe just one figure illustrating how things evolved after this time.

Reply:

In the revision, it is clarified that the diagnostics initially developed by Rotunno and Klemp (1982) is also shown to be valid and used for quasi-linear rainbands by Parker and Johnson (2004). The diagnostic results associated with cell A1 at 0645 UTC are also shown in a new figure (Fig. 17) and compared with those at 0630 UTC (Figs. 15 and 16), as suggested. The results show similar patterns and the effect from p_d*'* continue to dominate over those from buoyancy and p_b*'* at the rear side of A1 (mature cell), so that they are persistent through the mature stage. The time evolution of the pair A1 and A2 after 0630 UTC is also shown in Fig. 8, and it is described in more detail in relation to the results of dynamical pressure diagnostics in the revision, as suggested. Consistent with the diagnostics, cell A1 continues to strengthen for about 15 min (through 0645 UTC) but travels at a slower speed after 0630 UTC (linked to the induced vertical PGF at its rear flank), and cell A2 shows significant development afterwards and eventually becomes the dominant cell of the pair (near 0700 UTC).

4) There is some interesting 3-dimensional structure to the low-level vertical velocity field to the west of the updraft, as seen in Figs. 11 and 12, but this isn't really addressed in the subsequent discussion. (I'm referring to the "line" of upward motion that extends to the south-southwest of the main updraft.) Is this ascent explained also by the pressure perturbation diagnostics, or is something else going on there? And is it important to the back-building convection?

Reply:

The vertical velocity (at 1058 m) surrounding A1 is largely forced by the low-level convergence (at 547 m), so the two fields have very similar patterns. This is pointed out and described more clearly during the discussion of Fig. 9. The pattern surrounding A1 extends south and west (toward the area of the initiation of A2) is largely due to the deceleration in *u*-wind and confluence in *v*-wind, respectively. Both the speed convergence in *u* and confluence in *v* across A1 are consistent with (and mainly in response to) the development of deep convection. Thus, even though the background westerly LLJ is consistent with the confluence in *v* and is stronger toward the east (cf. Fig. 7), the deep convection still exhibits significant modulation effects on the local wind field. In the revision, the above phenomena and their relationships are described more clearly and explained in more details.

References

Eastin, M.D., and M. C. Link, 2009: Miniature Supercells in an Offshore Outer Rainband of Hurricane Ivan (2004). Mon. Wea. Rev., 137, 2081–2104. doi: http://dx.doi.org/10.1175/2009MWR2753.1

Hence, D. A., and R. A. Houze Jr. (2008), Kinematic structure of convective-scale elements in the rainbands of Hurricanes Katrina and Rita (2005), J. Geophys. Res., 113, D15108, doi:10.1029/2007JD009429.
Morin, M. J., and M. D. Parker (2011), A numerical investigation of supercells in landfalling tropical cyclones, Geophys. Res. Lett., 38, L10801, doi:10.1029/2011GL047448.

ACPD-15-8479-2015
Responses to Reviewer 2 (anonymous)
Date: 9 July 2015

Title: A numerical study of convection in rainbands of Typhoon Morakot (2009) with extreme rainfall: roles of pressure perturbations with low-level wind maxima
Authors: C.-C. Wang, H.-C. Kuo, R. H. Johnson, C.-Y. Lee, S.-Y. Huang, and Y.-H. Chen

1. General comments:

This manuscript investigates the formation and evolution of deep convection inside Typhoon Morakot's rainband using CReSS model. The authors then discuss the back-building mechanisms and how the distributions of the dynamical pressure favored the new development of updraft on the west side (upstream) of a mature cell. The results appeared plausible and in general consistent with observations.

This paper should be accepted for publication after major revision. Specific comments are listed below.

Reply:

We appreciate the positive views and critical comments from all three reviewers, and have revised the paper accordingly. Among the changes, we have (1) added the diagnostic results at 0645 UTC (besides 0630 UTC) to show a dominant and persistent effect from the dynamical pressure perturbation in the mature cell, (2) employed 10-min radar CAPPI data at 3 km to show more clearly the merging and back-building behavior in the observation, and (3) estimated the contribution from convection versus stratiform clouds over the plains in the event. In addition, the figures are polished and font sizes enlarged, the method of diagnosis is validated, the scale of the low-level jet is clarified, the cold pool is examined, and the evolutions of model convective cells are discussed in more detail, as suggested.

The changes in the manuscript are marked in red, blue, and green for Reviewer 1, Reviewer 2, and Reviewer 3, respectively. The modifications made by ourselves during the revision are in orange (mostly to correct mistakes), and those made during the production stage of ACPD since our first submission (to meet the format requirements) are in pink. The point-by-point responses to each of the comments/suggestions from this reviewer are listed below.

2. Major comments:

1. The authors should be congratulated with this great simulation. What is the

potential for CReSS to perform real-time TC prediction?

Reply:

Thank you for the nice comment. In fact, the CReSS model has been used to perform real-time weather forecasts for several years and recent results of the first author (C.-C. Wang) demonstrate its superior capability particularly in quantitative precipitation forecasts (QPFs) for extreme rainfall events brought by the TCs. A few published works are also cited for reference (Wang et al., 2013b; Wang, 2014, 2015).

2. It is somewhat disappointing that the authors did not compare their results with ample radar observations on this particular rainband. Although radar observations were shown in Fig. 4, it would be helpful to show observed cells indeed went though this sequence. Some of the black arrows (indicating the sequence of back building) in Fig. 4 are not obvious. It is difficult to compare vertical velocity (Fig. 8) with reflectivity (Fig. 4). Perhaps the authors can pick one or two cells in the radar observations to demonstrate their life cycle.

Reply:

In the revision, series of CAPPI reflectivity observed by the Chigu radar (location marked in Fig. 3a) at 3 km every 10 min over two 30-min periods (0510-0540 and 0620-0650 UTC) on 8 August 2009 are used in Fig. 4 to replace the old figure, and these plots can better depict the back-building and merging behavior of the cells embedded inside the rainbands in the observation. Also, the faster moving speed of new cells (to the west) than the old cells (to the east) can be clearly seen prior to their merger. The description in text is also modified accordingly.

3. Please clarify the meaning of LLJ. Was this LLJ a synoptic scale feature that this rainband took advantages of growing on top of it or it was a mesoscale feature accompanied by this rainband. For example, did each rainband in the simulation accompanied by a distinct LLJ or the LLJ is a scale larger than the individual rainband. The formation and/or the source of the LLJ may be one of the key issues to characterize this type of rainband.

Reply:

In the revision, it is clarified that the LLJ was a part of the TC circulation but was also most likely enhanced by the southwesterly monsoon (cf. Fig. 3a), as suggested. Also, while the westerly LLJ forms in response to the convergence within the TC flow and with the monsoon (mainly the confluence in *v*-wind) in the background and thus is stronger toward the east (cf. Figs. 7 and 10a), the deep convection still exhibits significant modulation effects on the local wind field and give rise to the speed couplets (Figs. 9 and 10a). In the revision, the above phenomena and their relationships are described more clearly and explained in more details.

4. The figures are hard to interpret with distance represented in longitude. The authors should consider using km rather than lat and lon for the axes. Other than Fig. 9, there is no distance scale in other figures.

Reply:

The lengths of the vertical cross-sections in Fig. 10a and b and Fig. 14 (old Fig. 13) are given in the caption, and distance scales are also added in Figs. 12, 13, 15, and 16 (old Figs. 11, 12, 14, and 15) as well as the newly-added Fig. 17 in the revision, as suggested.

3. Minor comments:

1. Fig. 9 can include a vertical motion plot as panel (c) rather than having to refer back to Fig. 8a.

Reply:

We do appreciate the reviewer's suggestion to add a vertical motion plot in Fig. 9 as the third panel for easy reference. However, since the distribution of the simulated low-level *w* at 0630 UTC has already been provided in Fig. 8 (at 1058 m), Fig. 12 (at 547 m), and Fig. 13a-f (at both 550 and 1050 m) and is readily available for the readers to refer to, we feel that it is perhaps not necessary to add another similar plot in the manuscript.

2. Is there a reason Fig. 10 a and b showing two different cross-sections? It is confusing as the readers may compare the structures shown in 10 a and b then find out they are not suppose to do so.

Reply:

The E-W vertical cross-section in Fig. 10a is along 22.5°N and slices through (or near) three cells (C1, A1, and B1), while that in Fig. 10b is along 22.52°N and cuts through the center of A1 to provide a close-up view of this mature cell. These above reasons are stated in the text, and it is clarified that the cross-section in Fig. 10b is not the same as, and is slightly to the north (by about 2 km) of, that in Fig. 10a in the revision to avoid confusion.

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Responses to Reviewer 3 (Prof. C. Rozoff)
Date: 9 July 2015

Title: A numerical study of convection in rainbands of Typhoon Morakot (2009) with extreme rainfall: roles of pressure perturbations with low-level wind maxima
Authors: C.-C. Wang, H.-C. Kuo, R. H. Johnson, C.-Y. Lee, S.-Y. Huang, and Y.-H. Chen

1. General comments:

This manuscript documents a numerical simulation of Typhoon Morakot on 8 August 2009, during its multi-day historic interaction with Taiwan in which catastrophic flooding occurred, due in part from the repeated formation and west-to-east passage of intense convective cells that impacted a significant proportion of Taiwan, particularly the southern half of Taiwan. The numerical simulation of this study produces a realistic simulation of the rainband activity observed in Morakot on 8 August. The authors focus on the back-building behavior and merger of cells within within east-west-oriented rainbands that impinged upon the Central Mountain Range at a time in which such convective cells were particularly vigorous. A pressure perturbation analysis applied on a characteristic convective cell clearly shows the local shear vector (associated with a strong low-level jet) produced a favorable dynamic pressure perturbation force that favored upstream development of new updraft, a slowing of mature convection, and thereby a favored mechanism for convective updraft mergers. This is an excellent study that likely applies to many more tropical cyclone cases than this particular Morakot example. I therefore enthusiastically recommend that the manuscript should be published after some minor revisions. The minor revisions are listed below as specific comments and are only meant to enhance the current analysis, which appears to be sound overall.

Reply:

We appreciate the positive views and critical comments from all three reviewers, and have revised the paper accordingly. Among the changes, we have (1) added the diagnostic results at 0645 UTC (besides 0630 UTC) to show a dominant and persistent effect from the dynamical pressure perturbation in the mature cell, (2) employed 10-min radar CAPPI data at 3 km to show more clearly the merging and back-building behavior in the observation, and (3) estimated the contribution from convection versus stratiform clouds over the plains in the event. In addition, the figures are polished and font sizes enlarged, the method of diagnosis is validated, the scale of the low-level jet is clarified, the cold pool is examined, and the evolutions of model convective cells are discussed in more detail, as suggested.

The changes in the manuscript are marked in red, blue, and green for Reviewer 1, Reviewer 2, and Reviewer 3, respectively. The modifications made by ourselves during the revision are in orange (mostly to correct mistakes), and those made during the production stage of ACPD since our first submission (to meet the format requirements) are in pink. The point-by-point responses to each of the comments/suggestions from this reviewer are listed below.

2. Specific comments:

1. In several of the figures, axis, contour, and colorbar labeling is very difficult to read due to small font size (e.g., Figs. 2, 10-15 are very difficult, and Figs. 1b, 3a, 7, 9 are marginal). Please consider resizing the fonts to be more legible.

Reply:

All the figures in question are improved in font size to be more legible, as suggested.

2. p. 8489, l. 6-8: An interesting question that arises is what percentage of the accumulated rainfall is accomplished by the intense cells that are the focus of this study vs. the more widespread stratiform rain associated with the rainbands (seen in all panels of Fig. 7)? This is not an essential question to answer in revisions, but, as a suggestion, if the calculation is readily available, it may bolster the practical significance of this study.

Reply:

The rainfall on 8 August from deep convection versus stratiform over the plains is estimated using hourly rain-gauge data, and is described in the revision, as suggested. For sites over the southwestern plains with a 24-h total rainfall amount of ≥700 mm on 8 August, at least 84% (and up to 95%) came from convective rainfall with an intensity of 20 mm h^{-1} or more. Thus, the practical significance of the present paper can indeed be enhanced.

3. Fig. 10b. It is easy to see that this convective cell does not produce an intense cold pool characteristic of some storms (such as midlatitude continental convection), but it is difficult to conclude whether there may be a weak cold pool or not. It seems conceivable even a 0.5 to 1-K magnitude cold pool (not uncommon in moist tropical cyclones) could produce some low-level lift, but such a cold pool would be difficult, if not impossible, to see in Fig. 10b. If possible, it might be nice to see a snapshot or two of the lowest model level's temperature field in the box shown in Fig. 7a at 0630

UTC with sufficiently small contouring intervals to discern the magnitudes of “cold” pools produced here.

Reply:

A new plot (Fig. 11) is produced for detailed examination of possible cold pool, as suggested. The results show only very weak cold pool (with surface temperature deficit within 0.5 K) and the weak outflow cannot reach the location of new cell development, and these results are described and discussed in the revision. The work of Yu and Chen (2011) is also cited for a comparison with the cold pool strength in the present case.

4. p. 8493, l.4: Equation (9) is not linearized.

Reply:

Corrected in the description, as suggested.

5. p.8493, l.6: This is the anelastic approximation, but not quite the Boussinesq approximation since the density is a function of height in the continuity equation. The fourth fluid extension term would disappear in eqn (13) in a Boussinesq fluid.

Reply:

Corrected in the description, as suggested.

6. This analysis may benefit from presentation (or verbal explanation) of the temporal evolution of vertical motion forcing mechanisms. For example, do the relative vertical motion forcing mechanisms (dynamic PGF, buoyant PGF, and buoyancy) maintain relative proportions of magnitude throughout the lifecycle of a given convective cell and/or birth of a new cell? This may help demonstrate also whether there are feedback loops. For example, higher buoyancy (even if transient) could induce stronger dynamic and buoyant pressure perturbations. Likewise, as I think is somewhat alluded to in this analysis (Fig. 14), the shearing terms in the dynamic pressure perturbation equation may induce a vertical motion pattern that reinforces the fluid extension term in a positive feedback loop. The temporal perspective may provide a deeper intuition into these complexities.

Reply:

In the revision, the diagnostic results associated with cell A1 at 0645 UTC are also shown in a new figure (Fig. 17) and compared with those at 0630 UTC (Figs. 15 and 16), as suggested. The results show similar patterns and the effect from p_d' continue to dominate over those from buoyancy and p_b' at the rear side of A1 (mature cell), so that they are persistent throughout the mature stage. The time evolution of the pair A1 and A2 after 0630 UTC is shown in Fig. 8, and it is described in more detail in relation to the results of dynamical pressure diagnostics in the revision, as suggested. It is noted that cell A1 maintains its strength through 0645 UTC, in agreement with the vertical PGF induced by p'_d. It is also noted that by strengthening the upward acceleration in the updraft, the shearing terms appear to also act to reinforce the fluid extension term (EX3) in Eq. (13), as suggested.

7. Fig. 15 is a very important figure that really brings together the manuscript as it clearly illustrates the impacts of the pressure perturbation forces vs. buoyancy on the vertical accelerations, particularly the importance of the dynamic PGF induced by the strong vertical shear structure. Still, in reference to the discussion on p. 8498, I recommend plotting the sum of the buoyancy and buoyant pressure perturbation gradient force alone (i.e., B - d pb / dz) as a separate panel, since it does appear that throughout a significant portion of the updraft, the buoyancy term B still dominates the buoyant pressure perturbation gradient force. Typically buoyancy dominates the PGF associated with buoyant pressure perturbations in mature updrafts in other idealized studies of convection.

Reply:

A new panel showing the sum of the buoyancy and buoyant pressure perturbation gradient force (in the vertical) has been added in Fig. 16 as Fig. 16d (old Fig. 15) as suggested, and the related description is also modified accordingly. In the newly-added Fig. 17, their sum (total buoyant effect) is also shown for 0645 UTC.

3. Technical corrections:

1. Eqn (13): The friction term from eqn. (12) mysteriously drops.

Reply:

Corrected in the description as suggested.

2. p. 8493, l.17: Simplify/spell check “are the Piosson equations of the laplacian of” to just “Poisson’s equations of” since the Laplacian operator is implicit to Poisson’s equation, by definition.

Reply:

Corrected and simplified as suggested.

A numerical study of convection in rainbands of Typhoon Morakot (2009) with extreme rainfall: roles of pressure perturbations with low-level wind maxima

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Abstract

This paper investigates the formation and evolution of deep convection inside the east-west oriented rainbands associated with a low-level jet (LLJ) in Typhoon Morakot (2009). With typhoon center to the northwest of Taiwan, the westerly LLJ was resulted from the interaction of typhoon circulation with the southwest monsoon flow, which supplied the water vapor for the extreme rainfall (of ~1000 mm) over southwestern Taiwan. The Cloud-Resolving Storm Simulator with 1-km grid spacing was used to simulate the event, and it successfully reproduced the slow-moving rainbands, the embedded cells, and the dynamics of merger and back-building (BB) on 8 August as observed. Our model results suggest that the intense convection interacted strongly with the westerly LLJ that provided reversed vertical wind shear below and above the jet core. Inside mature cells, significant dynamical pressure perturbations (p'_d) are induced with positive (negative) p'_d at the western (eastern) flank of the updraft near the surface and a reversed pattern aloft (> 2 km). This configuration produced an upward directed pressure gradient force (PGF) to the rear side and favors new development to the west, which further leads to cell merger as the mature cells slowdown in eastward propagation. The strong updrafts also acted to elevate the jet and enhance the local vertical wind shear at the rear flank. Additional analysis reveals that the upward PGF there is resulted mainly by the shearing effect but also by the extension of upward acceleration at low levels.

In the horizontal, the upstream-directed PGF induced by the rear-side positive p'_d near the surface is much smaller, but can provide additional convergence for BB development upstream. Finally, the cold-pool mechanism for BB appears to be not important in the Morakot case, as the conditions for strong evaporation in downdrafts do not exist.

1 Introduction

1.1 Literature review

Rainbands develop in response to linear forcing such as fronts, drylines, troughs, and convergence zone (e.g., Carbone, 1982; Browning, 1990; Doswell, 2001; Johnson and Mapes, 2001) or by self-organization in a sheared environment (e.g., Bluestein and Jain, 1985; Rotunno et al., 1988; Houze et al., 1990), and are a common type of precipitation systems around the world (e.g., Houze, 1977; Chen and Chou, 1993; Garstang et al., 1994; LeMone et al., 1998; Meng et al., 2013). These linear-shaped mesoscale convective systems (MCSs) are most well studied in mid-latitudes and classified by Parker and Johnson (2000, 2004) into three archetypes based on the location of stratiform region relative to the main line: trailing stratiform (TS), leading stratiform (LS), and parallel stratiform (PS), in response primarily to the different structure of environmental vertical wind shear. Most of these quasi-linear MCSs, especially the TS archetype, are squall lines and propagate relatively fast in direction more-or-less normal to the line (Houze et al., 1990).

The above motion of squall-line type systems, however, is not particularly conducive to high rainfall accumulation and the occurrence of flash floods, as Schumacher and Johnson (2005, 2006) found that the three archetypes together constitute only about 30% of extreme precipitation events caused by MCSs over the United State Great Plains. For the more hazardous linear systems that travel at small angles to their own alignment (also Dowsell et al., 1996; Brooks and Stensrud, 2000), the above authors further identified two common types of MCSs: the training line-adjoining stratiform (TL/AS) and back-building/quasi-stationary (BB), accounting for about 34% and 20% of the extreme rainfall events, respectively (Schumacher and Johnson, 2005, 2006). Inside the TL/AS type that often forms along or north of a pre-existing slow-moving boundary with an east-west (E-W) orientation, a series of embedded "training" convective cells move eastward (also Stevenson and Schumacher, 2014; Peters and Roebber, 2014). On the other hand, the BB systems are more dependent on mesoscale and storm-scale processes than synoptic boundaries, and while the embedded cells move downwind after initiation, new cells are repeatedly generated at nearly the same location at the upwind side, making the line as a whole "quasi-stationary" (also e.g., Chappell, 1986; Corfidi et al., 1996). Both the above configurations allow for multiple cells to pass through the same area successively, thus rainfall at high intensity to accumulate over a lengthy period (say, several hours) to cause extreme events and related hazards (Doswell et al., 1996). The common mechanism of repeated new cell generation at the end of BB MCSs are through the lifting at the leading edge (i.e., gust front) of the outflow of storm-generated cold pool (see e.g., Fig. 7 of Doswell et al., 1996), which forms on the upwind side of the system (e.g., Parker and Johnson, 2000; Schumacher and Johnson, 2005, 2009; Houston and Wilhelmson, 2007; Moore et al., 2012). Outside the North America, linear MCSs with embedded cells moving along the line are also often responsible for floods, such as the events in France, Australia, Hawaii, and East China (Sénési et al., 1996; Tryhorn et al., 2008; Murphy and Businger, 2011; Luo et al., 2014).

Another well-known theory through which the movement of convective cells, and thus the evolution of quasi-linear MCSs, can be modified is the dynamical pressure change induced by the shearing effect in environments with strong vertical wind shear, first put forward by Rotunno and Klemp (1982) to explain the propagation of isolated supercell storms (also e.g., Weisman and Klemp, 1986; Klemp, 1987). The related diagnostics is also shown to be valid for convection in quasi-linear MCSs by Parker and Johnson (2004). To be detailed in Sect. 3.3, the convective-scale dynamical pressure perturbation (p'_d) can be shown to be roughly proportional to the inner product of the vertical shear vector of horizontal wind ($\boldsymbol{S}$) and the horizontal gradient of vertical velocity ($\nabla_h w$). Thus, in an environment with westerly vertical shear ($\boldsymbol{S}$ pointing eastward), positive (negative) p'_d is induced to the west (east) of the updraft where $\nabla_h w$ points eastward (westward, see e.g., Fig. 7a of Klemp, 1987). This produces an eastward pressure gradient force (PGF) in the horizontal and favors new updraft development to the east (with $p'_d < 0$ aloft), and helps the storm to propagate forward. In Wang et al. (2009), multiple supercell storms near Taiwan are successfully simulated without the use of initial warm bubbles, and the perturbation pressure (p') couplets (rear-positive and front-negative with respect to $\boldsymbol{S}$) across the updraft are also reproduced. In the present study, the roles of pressure perturbations associated with convective cells inside the rainband of Typhoon (TY) Morakot in 2009 (e.g., Chien and Kuo, 2011; Wang et al., 2012) are investigated using the simulation results from a cloud-resolving model, and the background related to typhoon rainbands and this particular typhoon is introduced below.

1.2 Typhoon Morakot and back-building rainbands

Located over the western North Pacific (WNP), on average about 3-5 typhoons hit Taiwan annually and pose serious threats to the island. Some of them develop strong interaction with the monsoon that often further enhance the rainfall and worsen the damages. In the past 50 years, the most devastating case was TY Morakot in August 2009 (Lee et al., 2011; Chang et al., 2013), leading to 757 deaths and direct damages of roughly 3.8 billion U.S. Dollars (Wang et al., 2012). Based on the Joint Typhoon Warning Center (JTWC) best track, after its formation on 3 August, TY Morakot (2009) approached from the east since 4 August then impacted Taiwan during 6-9 August (Fig. 1a). Embedded inside a 4000-km monsoon gyre that enclosed two other TCs (e.g., Hong et al., 2010; Nguyen and Chen, 2011), Morakot was large in size (Fig. 1a and b) and moved very slowly near Taiwan under the influence of its background environment (e.g., Chien and Kuo, 2011; Wu et al., 2011). During the departure period on 8 August, its mean translation speed further dropped to below 2 m s^{-1} for about 24 h (cf. Fig. 1a), attributed to the effects of asymmetrical latent heating that concentrated at the rear side of the storm over the southern and eastern quadrants (Wang et al., 2012, 2013a; Hsu et al., 2013). At only category 2 on the Saffir-Simpson scale, Moarkot did its destruction almost entirely from the extreme rainfall (e.g., Hendricks et al., 2011) that reached 1624 mm in 24 h, 2361 mm in 48 h, and 2748 mm in 72 (Fig. 2a and b; Hsu et al., 2010) and approached the world record (Table 1). While the above studies clearly indicate that the event of Morakot was resulted from interactions across a wide range of scales, the interplay between TC motion and convection was especially important since the heaviest rainfall over southern Taiwan took place on 8 August when the TC moved the slowest (Figs. 1a and 2a and b, Wang et al., 2012).

During 8 August, two types of rainbands appeared persistently over or near Taiwan to cause the extreme rainfall. One was aligned north-south (N-S) near 120.7°E along the windward slopes of southern Central Mountain Range (Figs. 2a-d and 3a, Wang et al., 2012), produced through forced uplift of moisture-laden air by the steep topography at high precipitation efficiency (Yu and Cheng, 2013; Huang et al., 2014). The second type of rainbands, on the other hand, was nearly E-W oriented and parallel to the flow. On 8 August when Morakot's center was over northern Taiwan Strait, these E-W bands formed repeatedly over the southern Taiwan Strait, within the strong low-level convergence zone between the northerly to westerly TC circulation and the monsoonal flow from the southwest, as illustrated in Fig. 3a at 06:00 UTC as an example. Together, the two rainbands formed a "T-shaped" pattern and persisted into 9 August as the TC gradually moved away (Fig. 3b). Similar combinations of topographic (N-S) and TC/monsoon (E-W) rainbands were also observed in several past TCs (Kuo et al., 2010), such as Mindulle (2004), Talim (2005), Haitang (2005), and Jangmi (2008), so it is not unique to Morakot. While some of them have also been noted for evident interaction with the southwesterly monsoon (Chien et al., 2008; Yang et al., 2008; Ying and Zhang, 2012), all are among the 12 most-rainy typhoons in Taiwan, while Morakot (2009) sits at the very top of the list (Chang et al., 2013).

The E-W rainbands can be classified as secondary bands (Fig. 3a), as apposed to the three other types of TC rainbands outside the eyewall: principle, connecting, and distant bands (Willoughby et al., 1984; Houze, 2010). Within them, active and vigorous convective cells formed repeatedly off the coast of southwestern Taiwan and further upstream, then moved eastward over land, in a direction parallel to the band and the TC flow (Fig. 4, also Chen et al., 2010; Yu and Cheng, 2013). While several past studies have suggested that supercells may form in TC rainbands (e.g., Eastin and Link, 2009; Morin and Parker, 2011), the multi-cellular structure in Fig. 4 resembles those of both TL/AS and BB types of linear MCSs. An examination of hourly rainfall at gauges with a 24-h total ≥700 mm over the southwestern plains on 8 August reveals that at least 84% of the rain came from convective rainfall with an intensity ≥20 mm h^{-1}, and the percentage was higher (up to 95%) at sites with higher amount (>1000 mm). Thus, the embedded convection were clearly responsible for the heavy rainfall and serious flooding over much of the coastal plains in southwestern Taiwan (extending inland for about 50 km, roughly west of 120.5°E, cf. Figs. 2a-d and 3). In the examples shown in Fig. 4, back-building and merging and intensification of existing cells were both observed in the E-W bands, and prior to their merger, the new cell to the west often moves faster than the old one. Such behaviors are known to be largely controlled by processes at convective scale. Therefore, besides a favorable forcing of low-level convergence between the TC circulation and southwesterly monsoon at meso-α and β scale to trigger the convection and maintain the rainbands, whether processes at cloud and sub-cloud scale (meso-γ) also contributed in the detailed evolution of convection within the bands to cause the heavy rainfall over southwestern Taiwan? This is the focus of the present study.

In the typhoon environment, the maximum wind speed typically occurs near the top of the planetary boundary layer (PBL) due to thermal wind relationship and the influence of surface friction (e.g., Hawkins and Imbembo, 1976; Anthes, 1982, Sect. 2.3). Thus, the vertical wind shear is strongly cyclonic below the level of maximum wind and reverses in direction above it. At 00:00 UTC 8 August, the areal-averaged environmental wind upstream and near southern Taiwan indeed exhibited a distinct west-northwesterly low-level jet (LLJ) reaching 20 m s^{-1} at 850 hPa, and the vertical shear below it reversed in direction at 700-200 hPa (Fig. 5a), reminiscent to the "hairpin" shape noted by Schumacher and Johnson (2009). At 06:00 and 12:00 UTC (Fig. 5b and c), the LLJ and the associated shear through deep troposphere turned slightly to the left (into more E-W directions), in response to the northwestward movement of Morakot (cf. Fig. 1a). While the LLJ was clearly a part of the TC circulation (e.g., Hence and Houze, 2008), it was most likely enhanced by the southwesterly monsoon (cf. Fig. 3a). Based on Rotunno and Klemp (1982) and Klemp (1987), the interaction between convective updraft and its environmental flow (with LLJ and a reversed wind-shear in profile) would produce an anomalous high to the rear and an anomalous low ahead of the updraft near the surface (below the jet core) but a reversed pattern farther aloft (above the jet core). As illustrated by the schematic in Fig. 6, the effect from such dynamical pressure perturbations would be to favor updraft intensification to the west of mature cells and new development there, i.e., to the back side of the E-W rainbands in our case. Thus, in the present study, we examine such a mechanism and the possible roles played by the pressure perturbations surrounding convective cells in storm evolution (such as back-building and cell merger) mainly through cloud-resolving numerical simulation. In the Morakot case, such meso-γ scale processes and their potential roles in rainfall accumulation have not been studied previously.

2 Data and methodology

2.1 Observational data

The observational data used in this study include standard weather maps, the best-track data from the JTWC, infrared cloud imageries from the geostationary Multifunctional Transport Satellite (MTSAT) of Japan, and data from the rain-gauge network (Hsu, 1998) and radars operated by the Central Weather Bureau (CWB) of Taiwan during our case period. The radar data also include the vertical maximum indicator (VMI) and constant-altitude point position indicator (CAPPI) of reflectivity. Many of the above data have been used in the figures discussed so far.

2.2 Model and experiment

In Wang et al. (2012), the evolution of TY Morakot on 8 August 2009 was simulated in close agreement with the observations using the Cloud-Resolving Storm Simulator (CReSS; Tsuboki and Sakakibara, 2007), with a horizontal grid spacing of 3 km, a dimension of 480 × 480 × 50 (vertically stretched grid with spacing Δz = 100-745 m), and a model top of 25 km. Using European Center for Medium-range Weather Forecasts (ECMWF) Year of Tropical Convection (YOTC) analyses (0.25° × 0.25° and 20 levels, every 6 h, e.g., Waliser and Moncrieff, 2007; Moncrieff, 2010) as the initial and boundary conditions (IC/BCs), this particular experiment (their R01 run) served as the control to compare with sensitivity tests designed to examine the effects of asymmetric latent heating on the slow-down of Morakot over the northern Taiwan Strait during its departure. Besides research, the model has also been used to perform real-time forecasts including those for TCs (Wang et al., 2013b; Wang, 2014, 2015).

For this study, a run similar to R01 was performed with identical setup, except that the model top is increased from 25 to 36 km with slightly reduced vertical resolution. This experiment is referred to as the 3 km run (Table 2) and provides the IC/BCs to a second experiment using a horizontal grid spacing of 1 km and 55 levels, such that detailed structure and evolution of the convective cells embedded inside the rainbands can be reproduced and studied. The integration length of the 1 km run is 24 h starting from 00:00 UTC 8 August 2009, and model output intervals are 7.5 min. The detailed domain configuration and physics of the 3 km and 1 km runs are shown in Table 2 (also cf. Figs. 1b and 3a). For later analysis and discussion, only outputs from the 1 km model are used.

2.3 Analysis of pressure perturbations

In this study, to obtain the perturbation pressure p' associated with the convection in the 1 km model results for analysis, two different methods are employed. The first method is to define a background field that varies both with space and time, then separate p' by subtracting the background from the total field. In our case, since the TC center is located to the north and gradually moving away from the analysis area, set to 22-23°N, 119.2-120.2°E (1° × 1°, cf. Fig. 7a), the accompanying changes, including those with time, need to be partitioned into the background field. For pressure $p\,(x, y, z, t)$, its spatial mean over a fixed area is $\bar{p}_A(z,t)$ and its time average over a period from t_1 to t_2 is $\bar{p}_t(x,y,z)$. The time average of the spatial mean of p is thus $\bar{p}_{At}(z)$, which varies only with z. Here, we define Δp as the deviation of $\bar{p}_A(z,t)$ from its time average $\bar{p}_{At}(z)$ as

$$\Delta p(z,t) = \bar{p}_A(z,t) - \bar{p}_{At}(z)\,. \quad (1)$$

Thus, Δp can account for the gradual increase of the areal-mean pressure with time as the TC moves northward. Containing the averaged spatial pattern plus the change in its mean value with time, the background pressure (p_0) is defined and computed as

$$p_0(x,y,z,t) = \bar{p}_t(x,y,z) + \Delta p(z,t)\,, \quad (2)$$

and p' is obtained subsequently as

$$p'(x,y,z,t) = p(x,y,z,t) - p_0(x,y,z,t)\,. \quad (3)$$

Thus, the background pressure p_0 is not a function of z only, but also varies with location and time. Here, the time period for analysis is selected to be 03:00-12:00 UTC 8 August. When needed, the above method is also applied to other variables to obtain their perturbations, including horizontal wind (u, v) and virtual potential temperature (θ_v).

To further examine the detailed roles of pressure perturbation on the development and evolution of convection, the vertical momentum equation is analyzed and the two components of p', the dynamical (p'_d) and buoyancy (p'_b) pressure perturbations (thus $p' = p'_d + p'_b$), are evaluated following Rotunno and Klemp (1982), Klemp (1987), and Parker and Johnson (2004). In this second method, p'_d and p'_b and any of their contributing terms can be solved numerically through the relaxation method by iteration. For better clarity, the relevant formulation and procedure will be described later in Sects. 3.2 and 3.3, immediately followed by the results obtained from the 1 km simulation for selected convective cells with the presence of a westerly wind speed maximum near the top of the PBL in the Morakot case.

3 Results of model simulation

3.1 Model result validation

The CReSS model-simulated column maximum mixing ratio of total precipitating hydrometeors (rain + snow + graupel) in the 1 km run over the period of 06:00-08:00 UTC 8 August 2009 is shown in Fig. 7, which can be compared with the radar reflectivity composites in Fig. 3 and Wang et al. (2012, their Figs. 6e-g and 7). Comparison of these figures suggests that the model successfully reproduces the rainbands associated with TY Morakot near southwestern Taiwan over this period. On many occasions, more than one roughly E-W aligned bands (as observed) are simulated along a relatively wide zone of low-level convergence between the northerly to westerly TC flow and the west-southwesterly monsoon flow (Fig. 7). These modeled rainbands are in general agreement with earlier observational studies using radars (e.g., Chen et al., 2010; Yu and Cheng, 2013; Wei et al., 2014) while there are often slight displacements in their exact locations, typically by no more than 50 km. Nevertheless, the simulated accumulated rainfall distribution (Fig. 2e-g) compares favorably with the gauge observations on 8 August, including both 00:00-12:00 and 12:00-24:00 UTC (cf. Fig. 2b-d).

In the model, convective cells embedded inside the rainbands are repeatedly generated and move eastward after initiation, as in the observation (cf. Fig. 4), and the phenomena of back-building and cell mergers are successfully captured. For example, Fig. 8 shows the development and evolution of several convective cells near 22.5°N using model outputs at 1058 m every 7.5 min over 06:30-07:00 UTC. Already mature at the beginning of this 30 min period, cell "A1" moves eastward at an estimated speed of 26.1 m s^{-1}, while cells "B1" and "C1" also travel slightly faster at their mature stage. Meanwhile, new cells, labeled as "A2", "B2", and "C2", respectively, are initiated just upstream (to the west) of each of the three mature cells (Fig. 8), corresponding to back-building behavior as observed (cf. Fig. 4). At early development stage, the new cells also tend to travel faster than the adjacent old cells, most evidently for A2 that reached 31.1 m s^{-1} and eventually catches up and merges with cell A1 shortly after 07:00 UTC. Thus, the merging behavior of convective cells is also reproduced and linked to the slowing-down of mature cells, with a cycle of roughly 30-40 min in agreement with the observation (cf. Fig. 4). Although Figs. 4 and 8 only show a few selected cells as examples, similar back-building and merging behaviors are quite common both in observations and the 1-km run throughout 8 August (detailed figures not shown).

Since cells A1 and A2 exhibit typical evolution in the model with clear merging and back-building behavior, this particular pair of mature-new cells is selected for detailed study. Figure 9 presents the model-simulated horizontal winds and convergence/divergence associated with cell A1 at 547 and 2013 m at 06:30 UTC. While the rainband develops within the low-level convergence zone and the background westerly flow (i.e., the LLJ) increased in speed toward the east (about 35 m s^{-1} at 1058 m near A1; cf. Fig. 7b), the airflow surrounding the mature cell is very different, indicating significant local modulation on the jet by deep convection. Near the surface (Fig. 9a), a wind speed maximum-minimum couplet exists across the updraft (higher wind speed upstream) with a west-southwest-east-northeast (WSW-ENE) orientation and strong deceleration and convergence. At 547 m, the pattern of convergence closely resembles that of vertical motion at 1058 m (cf. Fig. 8, top panel) and extends south and west toward the area of new cell initiation (of A2), mainly due to the deceleration in u-wind and confluence in v-wind, respectively. From 39 to 22 m s^{-1}, the speed convergence in u across A1 at this level is about 5×10^{-3} s^{-1} and twice the magnitude of the confluence in v, consistent with the deep convection (Fig. 9a). On the other hand, the divergence at 547 m is generally located east and southeast (SE) of the updraft. At 2013 m (Fig. 9b), the updraft core appears slightly to the east and thus is tilted downstream, while the wind speed couplet turns slightly to a SW-NE alignment with similar convergence (from ~45 to 29 m s^{-1}).

In the E-W vertical cross-section along 22.5°N, which slices through (or near) several cells including C1, A1, and B1 at 06:30 UTC (near 119.5, 119.75, and 120.05°E, respectively, cf. Figs. 7b and 8), the local deceleration of westerly winds and convergence across convective cells at low levels are evident, while the oncoming environmental flow clearly has the structure of a LLJ with increased speed downstream and a core near 1 km in altitude (Fig. 10a). Away from the jet core level, the wind speed decreases much more rapidly below than above, suggesting strong westerly vertical wind shear near the surface (>10^{-2} s^{-1} in vorticity) but weak easterly shear above the LLJ, in agreement with Fig. 5. Across the u-wind couplets, the maximum speed typically occurs near 2 km, indicating an upward transport of momentum of the jet by the updraft, and the minimum speed is toward the surface. Another cross-section (about 2 km north) that cuts through the center of A1 (cf. Fig. 8) is shown in Fig. 10b and provides a close-up view of this mature cell, whose updraft indeed tilts eastward (downwind) with height (also Wei et al., 2014). Inside the updraft, air parcels accelerate upward to reach near 20 m s^{-1} at mid-levels and this clearly contributes to the strong low-level convergence, and thus the wind speed couplet, through continuity. Due to the eastward tilt of the updraft, maximum precipitation and near-surface downdraft (below 2 km) both occur slightly downwind (Fig. 10b; cf. Fig. 8). This is also depicted in Fig. 11, which in addition confirms only small surface temperature variations, within 0.5 K, over the rainfall area (centered near 119.77°E). Thus, the cold pool is very weak (cf. Yu and Chen, 2011) and the induced outflow cannot reach the vicinity of A2 (Fig. 11). These figures suggests that locally, the evaporative cooling in the downdraft (reflected by a downward decrease in hydrometeors) is barely enough to overcome the adiabatic warming effect, since the near-surface air is very moist (close to saturation) and a mid-level drier layer is also lacking in typhoon environment. Thus, the cold-pool mechanism commonly seen in mid-latitudes to initiate new cells (Doswell et al., 1996) does not seem important in our case here. Nevertheless, from Figs. 8-10, we see that the model cells, especially the mature ones, are moving at speeds slower than their low-level background flow, and we focus on the possible roles played by the pressure perturbations at sub-cloud scale in merging and back-building behaviors. Below, the vertical momentum equation is analyzed.

3.2 Analysis of vertical momentum equation

Following Rotunno and Klemp (1982) and Klemp (1987), the three-dimensional momentum equation can be expressed as

$$\frac{d\mathbf{v}}{dt} + \frac{1}{\rho}\nabla p = -g\hat{\mathbf{k}} - f\,\hat{\mathbf{k}} \times \vec{\mathbf{v}} + \mathbf{F}^*, \qquad (4)$$

where $\mathbf{v}(u, v, w)$ is the velocity vector, ρ is air density, g is gravitational acceleration, f is the Coriolis parameter, $\hat{\mathbf{k}}$ the unit vector in z direction, and $\boldsymbol{F}^*$ (F_x^*, F_y^*, F_z^*) the frictional term. Both p and ρ can be separated into the background and perturbation (i.e., $p = p_0 + p'$ and $\rho = \rho_0 + \rho'$, note that p_0 here is not the same as the one given in Eq. 2) and the former is assumed to be in geostrophic and hydrostatic equilibrium. At convective scale, the Coriolis force is neglected and friction is replaced by a turbulent mixing term $\boldsymbol{F}(F_x, F_y, F_z)$. Thus, the horizontal acceleration is caused by the perturbation PGF and turbulent mixing as

$$\frac{d\mathbf{v}_h}{dt} = -\frac{1}{\rho}\nabla_h p' + F_x + F_y \,, \tag{5}$$

where the subscript h denotes the horizontal components, while the vertical acceleration (dw/dt) can be approximated as

$$\frac{dw}{dt} = -\frac{1}{\rho}\frac{\partial p'}{\partial z} - \frac{\rho'}{\rho} g + F_z \approx -\frac{1}{\rho_0}\frac{\partial p'}{\partial z} - \frac{\rho'}{\rho_0} g + F_z = -\frac{1}{\rho_0}\frac{\partial p'}{\partial z} + B + F_z \,, \tag{6}$$

where $B = -g\,(\rho'/\rho_0)$ is the buoyancy acceleration. Thus, the vertical acceleration is driven by an imbalance among the perturbation PGF, buoyancy, and turbulent mixing.

The buoyancy B is composed of effects from gaseous phase and condensates, and the former can be accounted for by the virtual potential temperature perturbation (θ_v', where $\theta_v = \theta_{v0} + \theta_v'$) and the latter is the drag by cloud particles and precipitation, such that

$$B = -\frac{\rho'}{\rho_0} g = g\,\frac{\theta_v'}{\theta_{v0}} - g\,(q_c + q_i + q_r + q_s + q_g)\,, \tag{7}$$

where q_c, q_i, q_r, q_s, and q_g are mixing ratios of cloud water, cloud ice, rain, snow, and graupel, respectively, and available from model outputs. The separation of θ_{v0} and θ_v' is performed using the same method described in Sect. 2.3. Using $p' = p'_d + p'_b$, Eq. (6) can be rewritten to divide the perturbation PGF into two separate terms as

$$\frac{dw}{dt} = \frac{\partial w}{\partial t} + u\frac{\partial w}{\partial x} + v\frac{\partial w}{\partial y} + w\frac{\partial w}{\partial z} = -\frac{1}{\rho_0}\frac{\partial p'_d}{\partial z} - \left(\frac{1}{\rho_0}\frac{\partial p'_b}{\partial z} - B\right) + F_z \,. \tag{8}$$

Here, the equation of total derivative is used, and the vertical acceleration is driven by the dynamical perturbation PGF, the buoyancy effect (which contains both the buoyancy perturbation PGF and buoyancy acceleration), and turbulent mixing.

3.3 Analysis of dynamical and buoyancy pressure perturbations

Through the use of nearly incompressible Poisson equation, Rotunno and Klemp (1982) and Klemp (1987) can obtain p'_d and p'_b as the following. First, Eqs. (5) and (6) in Sect. 3.2 can be combined as

$$\frac{\partial \mathbf{v}}{\partial t} + \mathbf{v}\cdot\nabla\,\mathbf{v} + \frac{1}{\rho_0}\nabla p' = B\hat{\mathbf{k}} + \mathbf{F} \,, \tag{9}$$

and the anelastic continuity equation is

$$\nabla \cdot \rho_0 \mathbf{v} = 0 . \tag{10}$$

When Eq. (9) is multiplied through by ρ_0 then applied the three-dimensional gradient operator, the first lhs term vanishes using Eq. (10), and the dynamical and buoyancy terms can be separated as

$$\nabla^2 p'_b = \frac{\partial}{\partial z}(\rho_0 B) \text{ and} \tag{11}$$

$$\nabla^2 p'_d = -\nabla \cdot (\rho_0 \mathbf{v} \cdot \nabla \mathbf{v}) + \nabla \cdot (\rho_0 \mathbf{F}) . \tag{12}$$

After expansion and cancellation of terms with friction omitted, Eq. (12) can be rewritten as

$$\nabla^2 p'_d = -\rho_0 \left[\underbrace{\left(\frac{\partial u}{\partial x}\right)^2}_{\text{EX1}} + \underbrace{\left(\frac{\partial v}{\partial y}\right)^2}_{\text{EX2}} + \underbrace{\left(\frac{\partial w}{\partial z}\right)^2}_{\text{EX3}} - \underbrace{w^2 \frac{\partial^2}{\partial z^2}(\ln \rho_0)}_{\text{EX4}} \right] - 2\rho_0 \left(\underbrace{\frac{\partial v}{\partial x}\frac{\partial u}{\partial y}}_{\text{SH1}} + \underbrace{\frac{\partial u}{\partial z}\frac{\partial w}{\partial x}}_{\text{SH2}} + \underbrace{\frac{\partial v}{\partial z}\frac{\partial w}{\partial y}}_{\text{SH3}} \right) \tag{13}$$

where the first rhs term inside the brackets is the fluid extension term and the second is the shearing term. Together, Eqs. (11) and (13) are the Poisson equations of p'_d and p'_b, and a maximum (minimum) in laplacian corresponds to a minimum (maximum) in pressure perturbation. From Eq. (11), it can be seen that p'_b is related to the vertical gradient of buoyancy B. While a variety of processes in Eq. (13) can lead to the change in p'_d, the fluid extension effect includes four terms: three terms from divergence/convergence and the fourth term linked to w and the vertical gradient of ρ_0 (and will be referred to as EX1, EX2, EX3, and EX4, respectively). The fluid shearing effect consists of three terms related to horizontal wind shear and vertical shear of u and v, respectively (referred to as SH1, SH2, and SH3), and SH2 and SH3 contain the shearing effect (of $\boldsymbol{S} \cdot \nabla_h w$) mentioned in Sect. 1.1. After $\nabla^2 p'_b$, $\nabla^2 p'_d$, or any of its rhs terms is obtained using Eq. (11) or (13), the relaxation method is used to solve the associated pressure perturbation through iteration.

The results of $\nabla^2 p'$ obtained by the two different methods are compared in Fig. 12 at 547 m at 06:30 UTC 8 August as an example. The patterns are generally very similar, with positive $\nabla^2 p'$ (implying $p' < 0$) to the east and negative $\nabla^2 p'$ (implying $p' > 0$) to the west of mature cells (e.g., A1 and B1) or positive $\nabla^2 p'$ at and to the south-southeast (SSE) of rising motion of developing cells and negative $\nabla^2 p'$ to the north-northwest (NNW, e.g., A2, B2, and C1). This suggests that the separation method described in Sect. 2.3 also gives reasonable results. However, the contrast between positive and negative $\nabla^2 p'$ values obtained from Eqs. (11) and

(13) tends to be slightly larger. A comparison between Fig. 12b with patterns of $\nabla^2 p'_b$ and $\nabla^2 p'_d$ indicates that p' is dominated by p'_b (i.e., p'_d is minimal) in most area (where w is small) except near strong updrafts and downdrafts (not shown).

In Fig. 13, a closer view of the model-simulated vertical velocity w, the total pressure perturbation p' obtained through background separation and relaxation method, as well as p'_d through relaxation method at (or near) three different heights of 550, 1050, and 2050 m near cell A1 for the time at 06:30 UTC are shown, together with horizontal winds and vertical wind shear vector $\boldsymbol{S}$. Again, from the separation method, areas of $p' < 0$ at these three levels are typically found to the southern quadrants of the cell where ascending motion often also appears, while $p' > 0$ is generally to the north and weaker (Fig. 13a, d and g). The peak value of $p' < 0$ near the updraft is roughly -1 hPa, and located to its SE at 550 m but to the southwest at 2050 m. The patterns of total p' solved by the relaxation method at the three levels are generally similar, with $p' < 0$ to the SE and south of the updraft and $p' > 0$ to the north and northwest (NW, Fig. 13b, e and h). However, the N-S difference in p' surrounding the cell is considerably larger (near 3 hPa inside the plotting domain) to give a lower value in minimum p' (~ -1.5 hPa). The reason for this is most likely two fold: (1) the variation in background pressure p_0 on the xy-plane at 06:30 UTC is larger than the time mean used, so that p' centers are under-estimated for this time using the separation method, and (2) the frictional effect that tends to reduce the contrast in p' is not taken into account in Eqs. (11)-(13), causing some over-estimation in p' from the relaxation method. In agreement with earlier discussion (cf. Fig. 10a), the vertical shear across the updraft at 550 m is northwesterly to westerly and quite strong (roughly 1-2 $\times 10^{-2}$ s^{-1}, Fig. 13c) and a clear couplet in p'_d can be found with $p'_d > 0$ to the NW and $p'_d < 0$ to the SE of the updraft, which can already reach about 5 m s^{-1} at this level (cf. Fig. 10b). Consistent with the gradual veering of environmental wind with height in the lower troposphere (cf. Figs. 5b and 13a, d and g) the vertical shear turns clockwise at 1050 and 2050 m, and the alignment of the high-low couplet in p'_d also gradually changes into NNW to SSE and even north-northeast to south-southwest (Fig. 13f and i). Although the vertical shear near 1-2 km is considerably weaker (mostly $< 1 \times 10^{-2}$ s^{-1}) than at 500 m, the larger w and its horizontal gradient allow for a comparable magnitude in p'_d (cf. Fig. 10b). Thus, the patterns of p'_d up to 2 km are consistent with the SH2 and SH3 terms in Eq. (13) and our hypothesis, and with a difference of about 0.6-1 hPa across the high-low couplet, the dynamical pressure perturbations can account for a large part of the difference in total p' near the updraft (Fig. 13).

In Fig. 10a, it is seen that deep convection can locally modify the vertical wind profile and change the structure of the LLJ, so here we examine such changes and the resultant effect on the shearing term in Eq. (13) in greater detail before further discussion on the pressure perturbations due to different terms. In Fig. 14, the E-W vertical cross sections as in Fig. 10b (along 22.52°N) but for kinematic variables, vertical wind shear, and $\nabla^2 p'_d$ from SH2 are presented. In Fig. 14a, it can be clearly seen that the low-level convergence induced by the mature cell A1 also causes the LLJ to accelerate toward the updraft and decelerate beneath the updraft core. The upward transport of momentum inside the updraft (~6 km in width), consistent with its eastward tilt with height, also elevates the jet to 2-4 km. Due to the formation of the tilted maximum-minimum wind couplet (Fig. 14a), the vertical wind shear directly below the updraft core is enhanced (near 119.71-119.75°E and below 1 km, Fig. 14b). Associated with the rise of the LLJ, the northwesterly shear above the jet core at the western flank of the updraft (near 119.72°, 2-4 km) is also strengthened to some extent. In response to this profile of vertical wind shear modified by convection, the pattern of $\nabla^2 p'_d$ from SH2 (multiplied by -1 to have the same sign as p'_d) exhibits positive p'_d below 1-1.5 km to the west and negative p'_d below ~3 km to the east of the updraft center, and a reversed pattern above to at least 5 km (Fig. 14b), again consistent with our hypothesis.

Figure 15 presents the total p' from background separation and p'_b and p'_d and the major contributing terms of p'_d from the relaxation method along the same E-W vertical cross-section as Fig. 14, in addition to w. The total p' associated with cell A1 obtained from the two different methods have similar patterns on the vertical plain (Fig. 15a and b), with largest negative p' of about -1 hPa near 3-4 km and higher p' at the surface immediately to the west of the updraft core. Directly underneath the updraft and to its east, on the other hand, p' obtained through the relaxation method is smaller (more negative) near the surface, which is mainly due to the effects of p'_b rather than p'_d (Fig. 15c and d). To the west of the updraft, however, p'_d is the main reason for the total p' to become positive near the surface and negative further aloft as expected (Fig. 15a-c), and the major contributing terms to this pattern is the shearing and divergent effects of the updraft (SH2 + EX3, Fig. 15e and f), in agreement with Fig. 14. The region of $p'_b < 0$ below and east of the updraft at low levels is resulted from an increase in buoyancy B with height (i.e., $\nabla^2 p'_b > 0$, cf. Eqs. 7 and 11). On this section plane, rhs terms in Eq. (13) other than SH2 and EX3 are much smaller, especially EX4 which is about two orders of magnitude smaller. Thus, these other terms are not shown.

4 Discussion

In the previous section, the dynamical pressure perturbation p'_d near the updraft of the mature cell in the rainband is found to exhibit a pattern consistent with our hypothesis, i.e., with positive (negative) perturbation below (above) the LLJ upwind from the updraft, and this pattern is attributed mainly to the shearing effect of the updraft on the vertical wind shear associated with the jet (SH2) but also to the extension term from vertical acceleration inside the tilted updraft (EX3). The induced PGF by the total p' and its components (p'_b and p'_d) in the vertical can be computed using Eq. (8), and this is shown in Fig. 16 on the same cross section through cell A1 (along 22.52°N). With its pattern shown in Fig. 15c, the high-low couplet of p'_d west of the updraft induces an upward-directed PGF there below about 3 km, with a peak value of roughly 7×10^{-2} m s^{-2} at 1.5 km (Fig. 16a). Such acceleration can produce an upward motion of 5 m s^{-1} under 75 s, or from 5 to 15 m s^{-1} in 150 s across a distance of about 1.5 km, very comparable to the acceleration below the core of the main updraft. Consistent with this result, cell A1 maintains its strength but travels at a slower speed after 0630 UTC (cf. Figs. 8 and 17). Thus, the distribution of p'_d to the rear side of the updraft can certainly affect the evolution of cell A1 and cause it to slow down in moving speed. In other words, the cell merging behavior in the rainbands consisting of multiple cells, when they develop in an environment with an intense LLJ as in the present case, can be explained by the mechanism of dynamical pressure perturbations induced through the shearing (and extension) effect. In Figs. 15 and 16, by strengthening the upward acceleration in the updraft, the shearing effect appears to also act to reinforce the fluid extension term (EX3) in Eq. (13). In addition to its role in cell merger, the reduced propagation speed of mature cells implies an enhancement in low-level convergence upstream. Using Fig. 8 (A1 travelling at 26.1 m s^{-1}) and assuming a LLJ of 35 m s^{-1} about 40 km upstream (cf. Figs. 10a), the speed convergence implied is about 2.2×10^{-4} s^{-1}, or 3.2×10^{-4} s^{-1} larger than the background with speed divergence of $\sim 1 \times 10^{-4}$ s^{-1}.

Because of the surface-based negative p'_b below and east of the updraft, the induced vertical PGF by p'_b is also negative (directed downward) below 3 km (Figs. 15d and 16b), suggesting that the downwind side of the updraft is less favorable for its maintenance and further development. The buoyancy B is mostly positive inside the updraft (Fig. 16c) and this can only come from an increased θ_v' due to latent heat release (cf. Eq. 7). Although B is also positive below the updraft core and even to the west (below 1 km), its values are smaller than the upward acceleration induced by the p'_d pattern and largely cancelled by the effect of p'_b (Fig. 16d). Thus, when all three terms in Fig. 16a-c are added together in Fig. 16e, their total effect on vertical acceleration (cf. Eq. 8) resembles that from the effect of p'_d alone in both the pattern and magnitude (cf. Fig. 16a).

The patterns of p', p'_d, and p'_b, their induced perturbation PGF in the vertical, and B in association with the mature cell, as shown in Figs. 15 and 16, are also quite persistent through time. For example, at 06:45 UTC when A2 grows significantly stronger (cf. Fig. 8), the distributions of p'_d (Fig. 17a) and vertical accelerations from the dynamical (Fig. 17b), total buoyant (Fig. 17c), and their combined effects (Fig. 17d) along the E-W cross section (through A1) all remain similar to those 15 min earlier (cf. Figs. 15c and 16a, d, and e, respectively), although some of them have weakened somewhat. The effects of p'_d, nevertheless, still dominate over those from p'_b and B at the rear side of cell A1. Similar results are also found in mature cells in other rainbands (not shown).

From Fig. 15c and e, the positive p'_d near the surface can be seen to also produce horizontal PGF apart from the vertical PGF, and the westward-directed PGF upstream from A1 (west of 119.7°E) can be estimated using Eq. (5) to be about 3-7 × 10^{-3} m s^{-2}, which is about one order of magnitude smaller than the PGF in the vertical from p'_d. Nevertheless, a value of 5 × 10^{-3} m s^{-2} is enough to decelerate the oncoming westerly flow by 1 m s^{-1} in 200 s and induce a speed convergence of roughly 1.4 × 10^{-4} s^{-1} (again assuming a background flow of 35 m s^{-1}). Even though this value is one order of magnitude smaller than the convergence associated with cell A2 at 06:30 UTC during its early stage of development (~1.5 × 10^{-3} s^{-1}, cf. Fig. 9a) the combined convergence with that implied by a slower moving speed of A1 (as discussed earlier) would be about 4.6 × 10^{-4} s^{-1} larger compared to its surrounding. This is certainly not negligible and can provide additional forcing to favor new cell development upstream from the old cell, consistent with the time evolution of A2 (Figs. 8 and 17). Therefore, the role played by the dynamical pressure perturbation in producing an anomalous high near the surface and additional uplift at the rear flank of the updraft of mature cells can favor both merging and new cell initiation further upstream, i.e., the behavior commonly found in back-building MCSs, in the rainbands of TY Morakot (2009) when a strong LLJ is present in the background. Since the LLJ is a common feature in the TC environment, our results are likely also applicable to cell evolution in the rainbands of other TC cases.

In the Mei-yu season (May-June), quasi-linear MCSs also often develop near the Mei-yu front in an environment with a LLJ (e.g., Chen, 1992; Chen and Chou, 1993; Chen et al., 2005; Wang et al., 2014), and they may exhibit characteristics reminiscent to the TL/AS or BB systems described by Schumacher and Johnson (2005, 2006) and cause heavy rainfall and flash floods (e.g., Lin et al., 1992; Wang et al., 2005; Jou et al., 2011). Thus, although typhoon rainbands are studied here and the LLJs are typically not as strong in the Mei-yu season, a similar interaction between the updrafts and LLJ may promote cell merging and initiation of new cells upstream, and thus contribute to heavy rainfall and related weather hazards.

5 Conclusion and summary

Typhoon Morakot in August 2009 was the most devastating TC to hit Taiwan over the past 50 years, with extreme rainfall that came close to the 24 h and 48 h world records. During the period of heaviest rainfall on 8 August, when the TC center was over the northern Taiwan Strait, the E-W oriented, persistent, and slow-moving rainbands and the embedded deep convection that propagated eastward (parallel to the bands) were responsible for the serious and wide-spread flooding over the southwestern plains of Taiwan. Developing inside the low-level convergence zone between the TC vortex (from the N/NW/W) and the monsoon flow (from the WSW/SW) over southern strait, as also observed in several other past TCs, these rainbands were collocated with a westerly LLJ and exhibited frequent cell merging and back-building behavior that contributed to the heavy rainfall. Thus, the possible roles of pressure perturbations associated with deep convection on rainband behavior of TY Morakot (2009) are investigated in this study, mainly through the use of simulation results from the CReSS model at a horizontal grid-spacing of 1 km every 7.5 min.

In the model, the rainbands, multiple cells embedded, their eastward movement, and merging and back-building behavior are all successfully captured in close agreement with the observations, although slight positional errors are often unavoidable. In its mature stage, a particular cell at 06:00 UTC 8 August is selected for detailed study. As hypothesized (cf. Fig. 6) following Rotunno and Klemp (1982) and Klemp (1987), in an environment of a LLJ (~35 m s^{-1}) with reversed vertical wind shear below and above, the interaction between convective updraft and such a vertical shear profile produces a positive dynamical pressure perturbation p'_d to the west (rear) and a negative p'_d to the east of the updraft near the surface (~500 m), but $p'_d < 0$ to the west and $p'_d > 0$ to the east farther aloft above the jet core (> 2 km). At the rear side, the positive-negative couplet of p'_d in the vertical has a difference of about 1 hPa in p, and induces an upward directed PGF whose magnitude is comparable to the acceleration inside the updraft core. This configuration favors updraft development at the rear flank, and subsequently causes the mature cells to slow down and merge with approaching new cells, which remain faster without a strong updraft. The intense updraft of this mature cell, reaching 10 m s^{-1} at 1.3 km and about 20 m s^{-1} at mid-level, is also found to elevate the jet and act to enhance the local vertical wind shear both above and below the jet core at its rear flank, aided by the induced near-surface convergence at cloud scale.

Through the use of relaxation method to solve for each contributing terms of p'_d (and the buoyancy perturbation pressure p'_b, where $p' = p'_d + p'_b$), our results indicate that the above vertical couplet of p'_d to the rear flank is mainly caused by the shearing effect (SH2 in Eq. 13, or $-2\rho_0(\partial u/\partial z)(\partial w/\partial x)$), but also contributed by the vertical extension term (EX3 in Eq. 13, or $-\rho_0(\partial w/\partial z)^2$), i.e., by the upward acceleration in the updraft, at low levels, while the effect of p'_b is nearly counteracted by the buoyancy B in the mature cell examined. Near the surface, the westward-directed horizontal PGF induced by the positive p'_d at the rear side, when combined with the effect from the slow-down of mature cells, can produce an estimated additional convergence (in speed) roughly 1/4-1/3 of the value associated with developing new cell further upstream. Thus, a positive p'_d near the surface in the couplet is also helpful to new cell initiation some distance upstream, i.e., the back-building process, compared to the conditions without a mature cell. Finally, the updraft in the mature cell in our case tilts eastward (downwind) with height due to the presence of the LLJ, and the maximum precipitation and near-surface downdraft occur at the eastern side. However, only a weak cold pool is found since the low-level air is very moist and a dry layer does not exist at mid-levels in the TC environment. Thus, the cold-pool mechanism typical in mid-latitudes to initiate new cells in back-building systems does not appear to be important in our case here.

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Table 1. Comparison of maximum accumulated rainfall (mm) observed in TY Morakot (2009) in Taiwan (Hsu et al., 2010) and the World's record rainfall, including location and date (source: World Meteorological Organization World Archive of Weather and Climate Extremes, available at http://wmo.asu.edu/#global; Holland, 1993; Guhathakurta, 2007; Quetelard et al., 2009).

Duration	Morakot	World record		
		Amount	Location	Date
24 h	1624	1825	Foc-Foc, La Réunion	7-8 Jan 1966
48 h	2361	2493	Cherrapunji, India	15-16 Jun 1995
72 h	2748	3930	Cratère Commerson, La Réunion	24-26 Feb 2007
96 h	2855	4936	Cratère Commerson, La Réunion	24-27 Feb 2007

Table 2. Domain configuration, physics, and experiment design used in this study. In the vertical, the grid spacing (Δz) of the CReSS model is stretched (smallest at the bottom), and the averaged spacing is given in the parentheses.

	3 km	1 km
Projection	Lambert Conformal, center at 120°E, secant at 10°N and 40°N	
Grid dimension (x, y, z)	$480 \times 480 \times 50$	$450 \times 500 \times 55$
Grid spacing (km)	$3.0 \times 3.0 \times$ 0.1-0.98 (0.72)	$1.0 \times 1.0 \times$ 0.1-0.718 (0.5)
Topography and SST	Real at (1/120)°, and weekly mean at 1° resolution	
Initial/boundary conditions (IC/BCs)	ECMWF YOTC analyses (0.25° × 0.25°, 20 levels, 6 h)	Outputs from 3 km run (3 km, 55 levels, 15 min)
Initial time	0000 UTC 8 Aug 2009	
Integration length	48 h	24 h
Output frequency	15 min	7.5 min
Cloud microphysics	Bulk cold rain scheme (mixed phase with 6 species)	
PBL parameterization	1.5-order closure with TKE prediction	
Surface processes	Energy/momentum fluxes, shortwave and longwave radiation	
Soil model	41 levels, every 5 cm to 2 m deep	

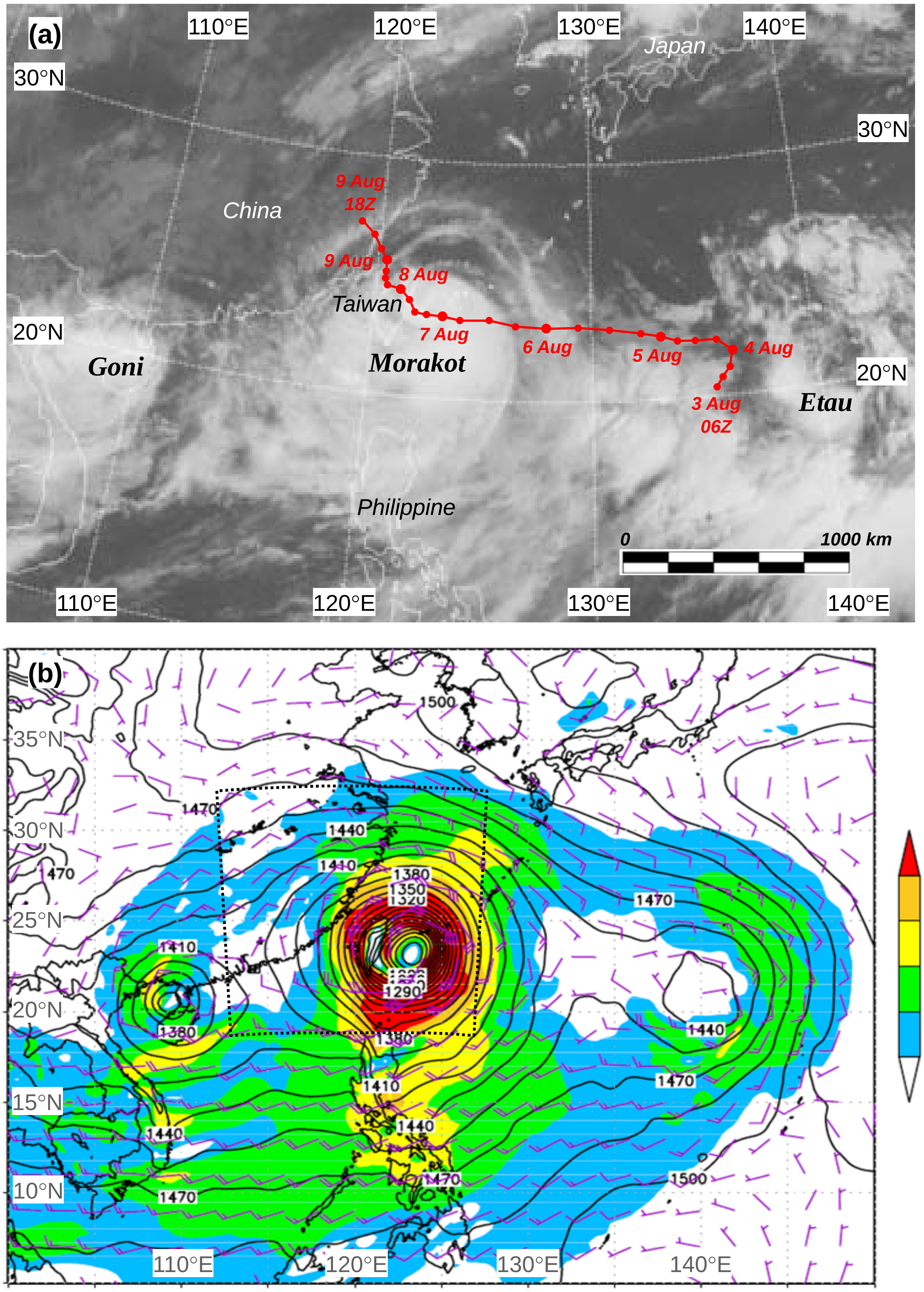


Figure 1. (a) The JTWC best-track of Typhoon Morakot (TY0908) overlaid with MTSAT IR cloud imagery at 00:30 UTC 7 August 2009. The TC positions are given every 6 h. (b) The ECMWF-YOTC analysis of geopotential height (gpm, contours) and horizontal winds [m s^{-1}, full (half) barb = 10 (5) m s^{-1}, wind speed shaded] at 850 hPa at 00:00 UTC 7 August 2009. The domain of 3 km experiment is also plotted (dotted region).

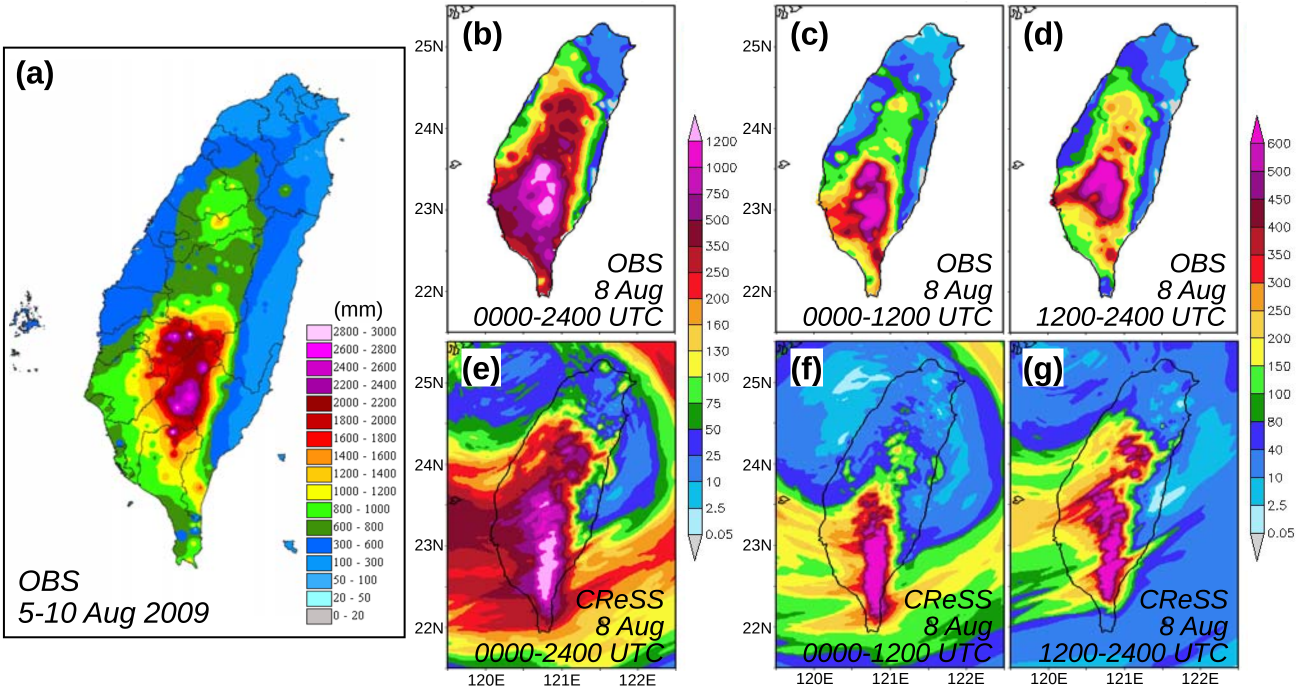


Figure 2. Observed (a) total rainfall distribution (mm) in Taiwan over 5-10 August, (b) daily (00:00-24:00 UTC) rainfall distribution on 8 August, and 12 h rainfall over (c) 00:00-12:00 UTC and (d) 12:00-24:00 UTC on 8 August, during Morakot (2009). (e-g) Same as (b-d) except from 1 km CReSS model simulation.

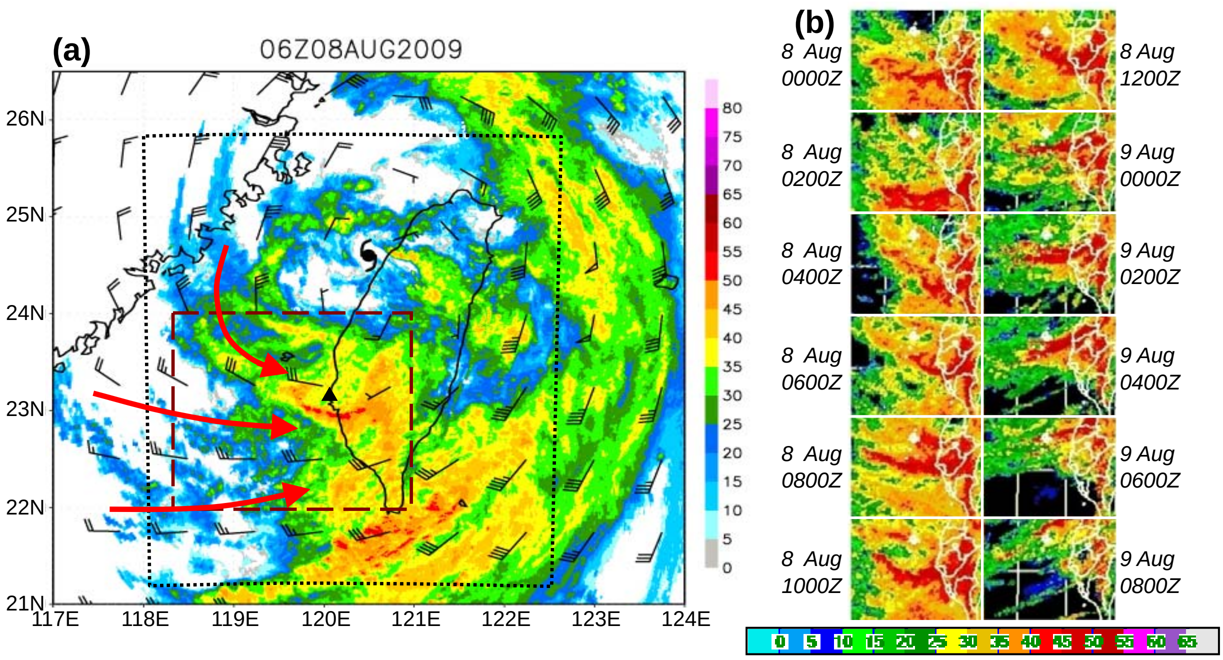
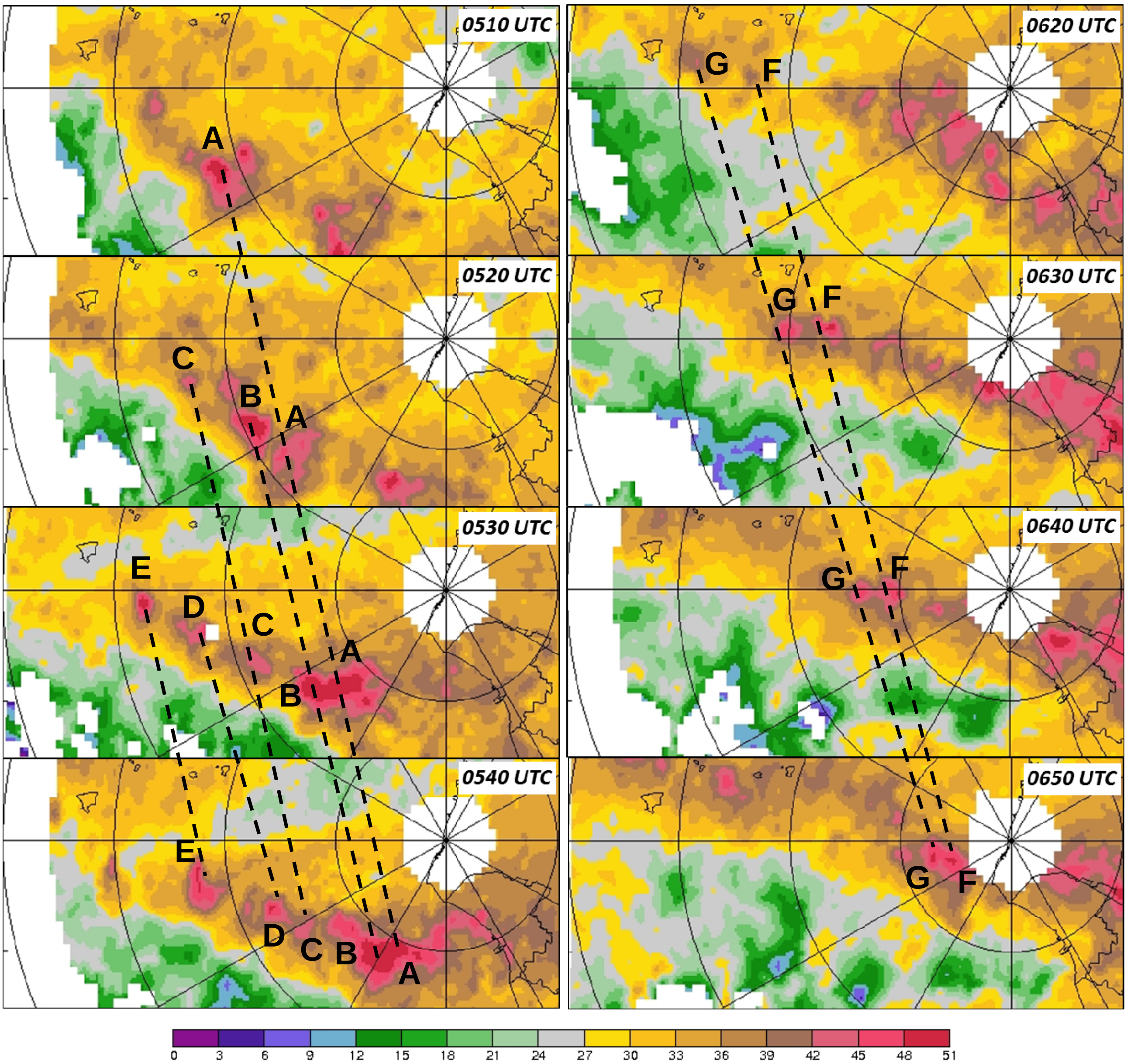


Figure 3. (a) The composite of radar VMI reflectivity (dBZ, scale on the right) near Taiwan overlaid with the ECMWF-YOTC 850 hPa horizontal winds [m s^{-1}, full (half) barb = 10 (5) m s^{-1}] at 06:00 UTC 8 August 2009. The TC center is marked by the typhoon symbol, and the 1 km domain used in this study is also plotted (black dotted region). The triangle marks the location of Chigu radar. (b) Radar VMI reflectivity composites (dBZ, scale at bottom) over the brown dashed box in (a) every 2 h over 00:00-12:00 UTC 8 August and 00:00-08:00 UTC 9 August 2009.



1 Figure 4. Series of CAPPI reflectivity (dBZ, scale at bottom) at 3 km observed by the Chigu radar (cf. Fig. 3 for location) every 10 min during 05:10-05:40 and 06:20-06:50 UTC 8 August 2009, for regions of 100 × 45 km^2 covering the E-W oriented rainbands (courtesy of Prof. C.-H. Liu). The concentric rings are 20 km apart. Selected back-building cells are labeled as A-G and cell motions are marked (dashed lines).

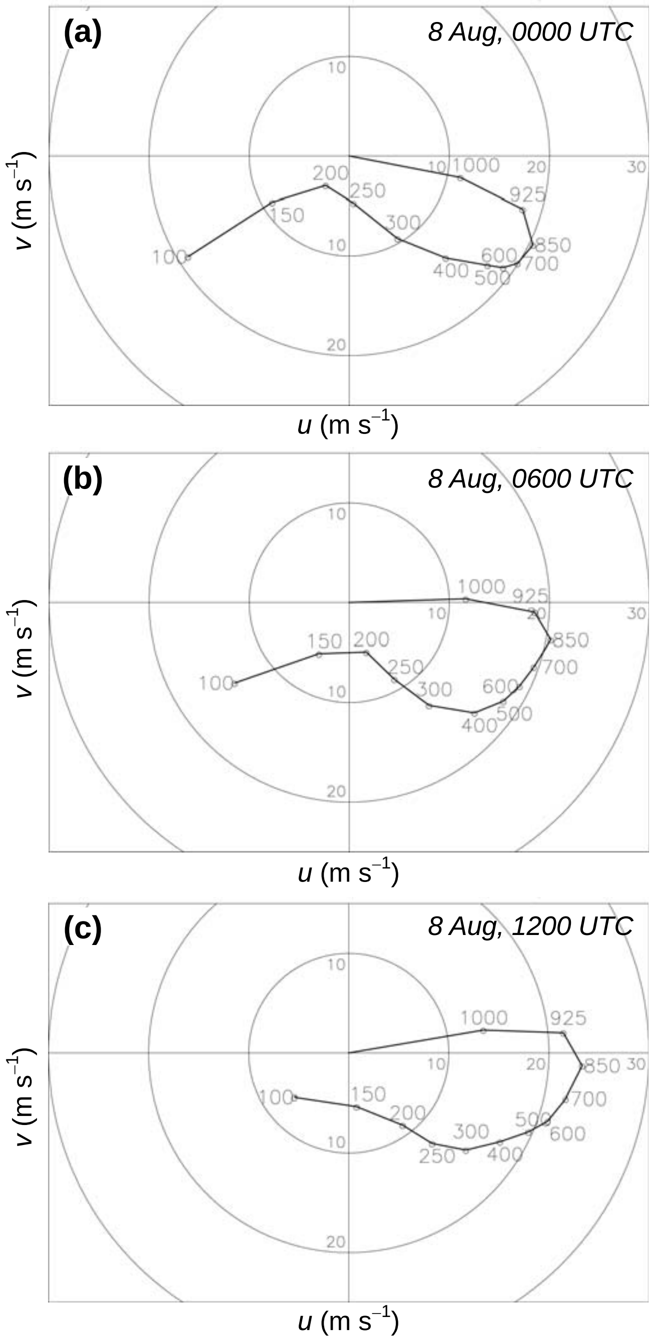


Figure 5. Hodograph of the mean wind (m s^{-1}) inside 22-24°N, 118.3-121°E (brown dashed box in Fig. 3a) computed from the ECMWF-YOTC analyses at (a) 00:00, (b) 06:00, and (c) 12:00 UTC 8 August, 2009. The numbers along the curve indicate pressure (hPa) at nearby dots.

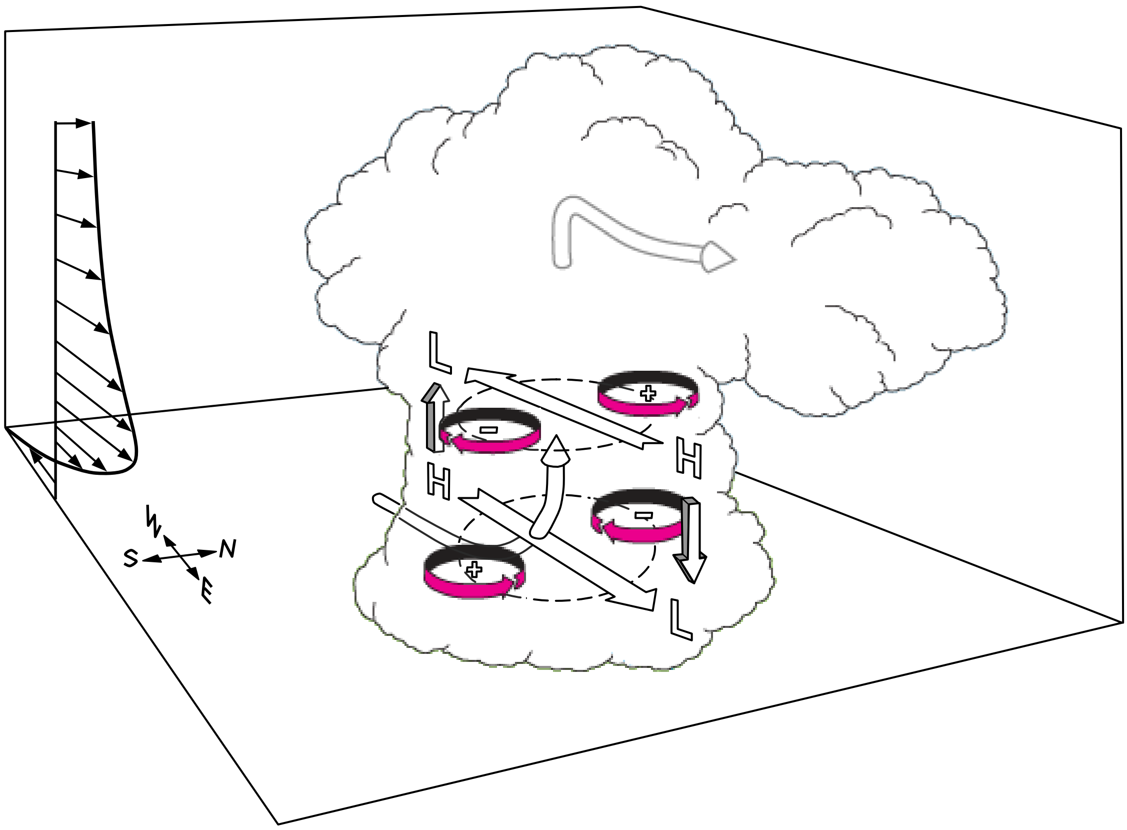


Figure 6. A schematic showing the distribution of dynamical pressure perturbations (marked by “H” for anomalous high and “L” for anomalous low) relative to a mature storm cell in an environment with the presence of a westerly low-level jet (LLJ) and westerly (easterly) vertical shear below (above) the jet core as in our case. The configuration is in favor of new development at the rear side and a slower moving speed of the cell.

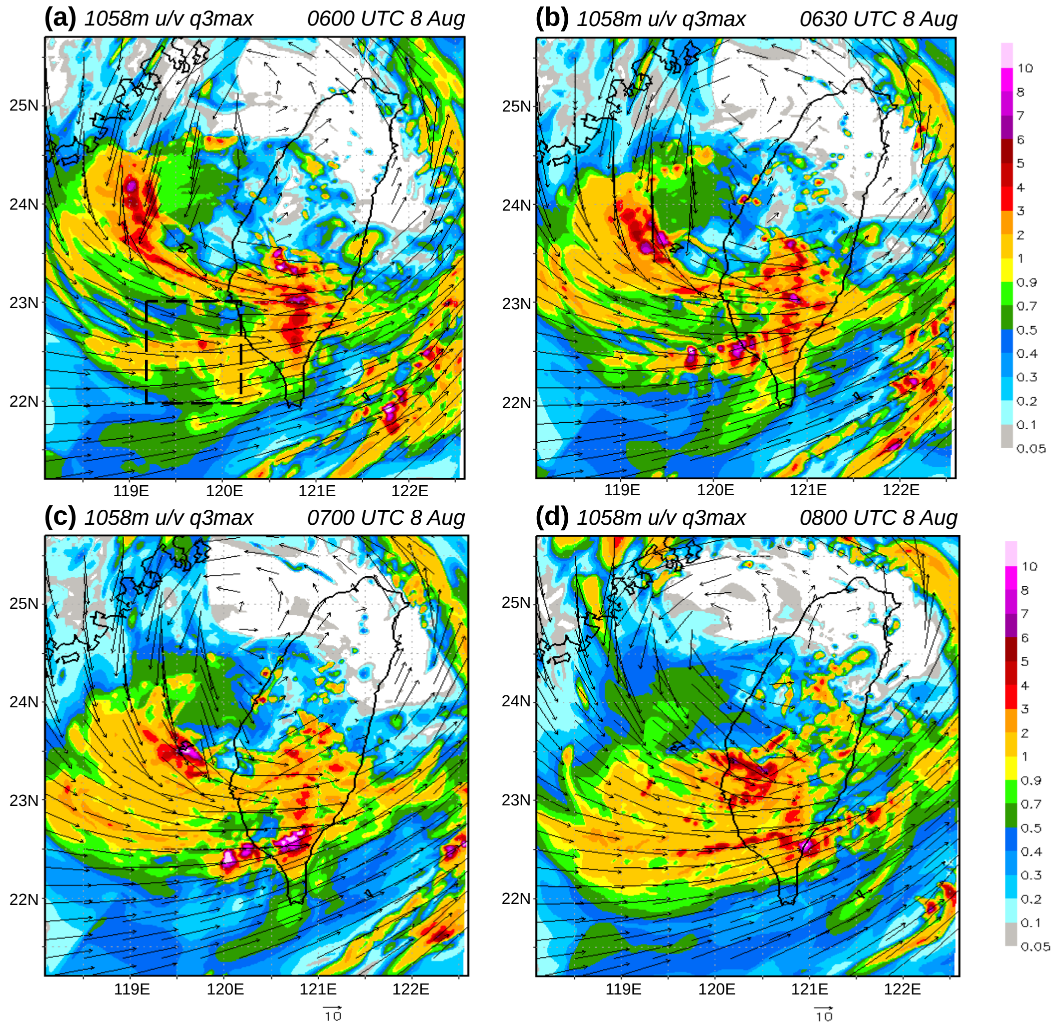


Figure 7. Model-simulated column maximum mixing ratio of total precipitating hydrometeors (g kg^{-1}, rain + snow + graupel, colors, scales on the right) and horizontal winds (m s^{-1}, reference vector at bottom) at 1058 m at (a) 06:00 UTC, (b) 06:30 UTC, (c) 07:00 UTC and (d) 08:00 UTC 8 August 2009 in the 1 km run. The dashed box in (a) depicts the area of pressure perturbation calculation (22-23°N, 119.2-120.2°E).

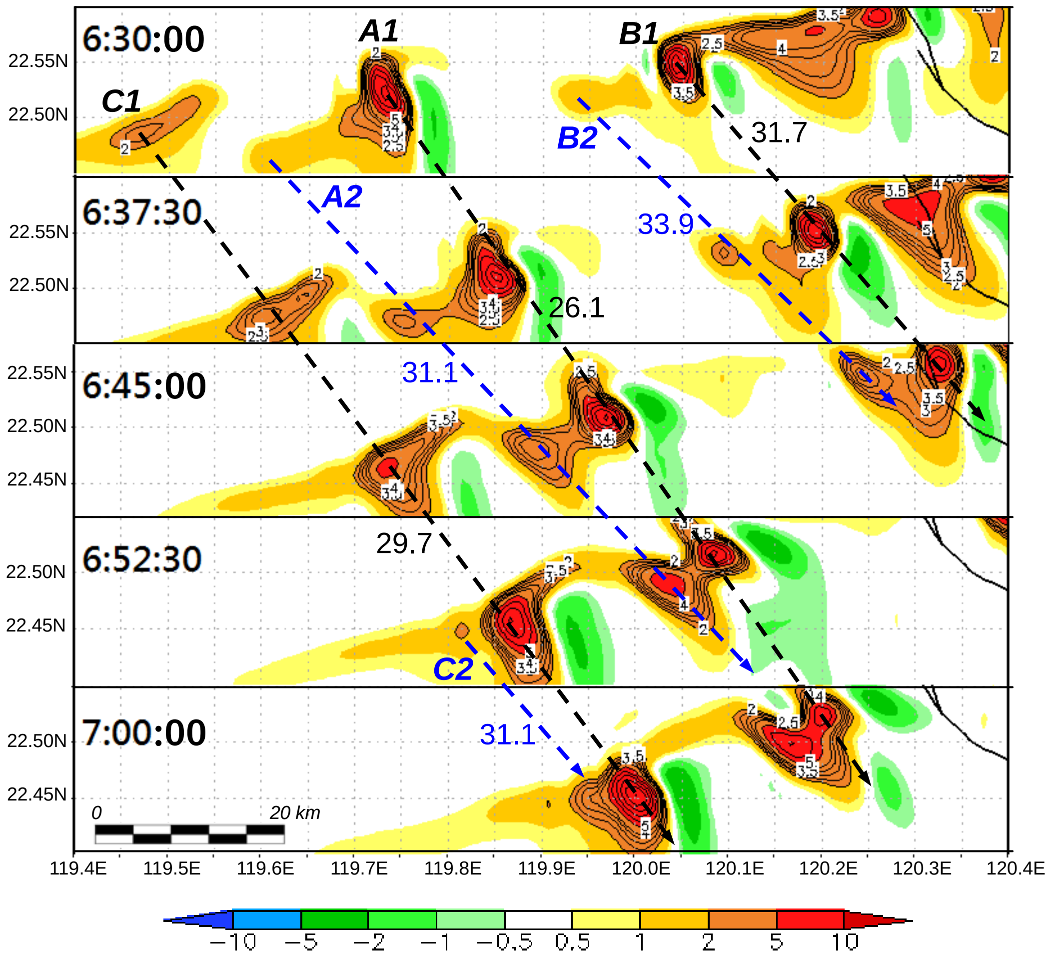


Figure 8. Similar to Fig. 4, but showing model-simulated vertical velocity (m s^{-1}, color shades and contours) at the height of 1058 m (sixth output level) over the period of 06:30-07:00 UTC (every 7.5 min) 8 August 2009 from the 1 km experiment. Contours start from 2 m s^{-1}, at intervals of 0.5 (1.0) m s^{-1} up to (above) 4 m s^{-1}. Old cells (A1, B1, and C1) and nearby new cells (A2, B2, and C2) and their estimated propagation speeds (m s^{-1}) are labeled.

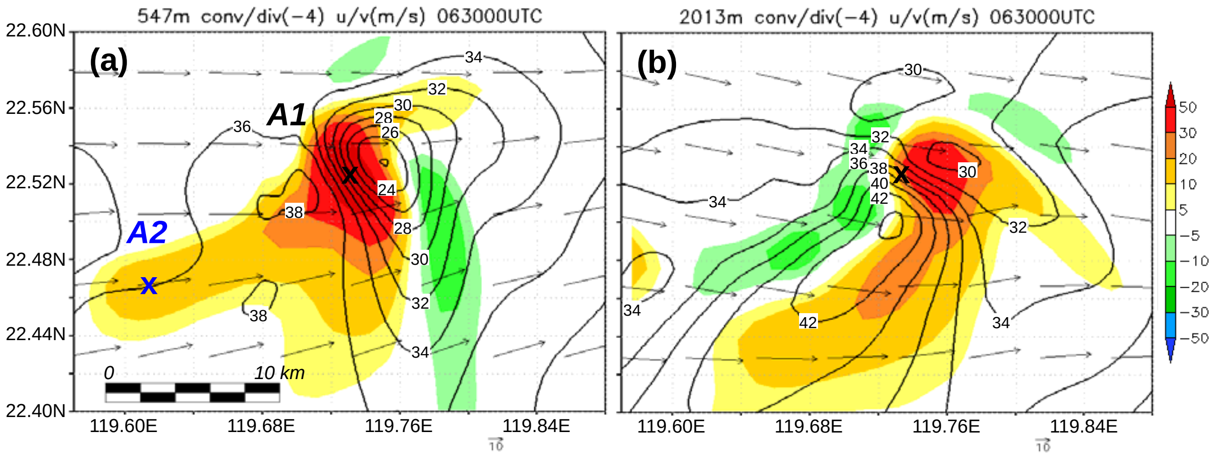


Figure 9. Model-simulated horizontal winds (m s^{-1}, vectors, reference vector at bottom), wind speed (contours, intervals: 2 m s^{-1}), and convergence/divergence (10^{-4} s^{-1}, color, positive for convergence, scales on the right) at (a) 547 m (fourth output level) and (b) 2013 m (ninth output level) associated with the convective cell "A1" off the southwestern coast of Taiwan at 06:30 UTC 8 August 2009. The "x" marks the updraft center at 1058 m (cf. Fig. 8).

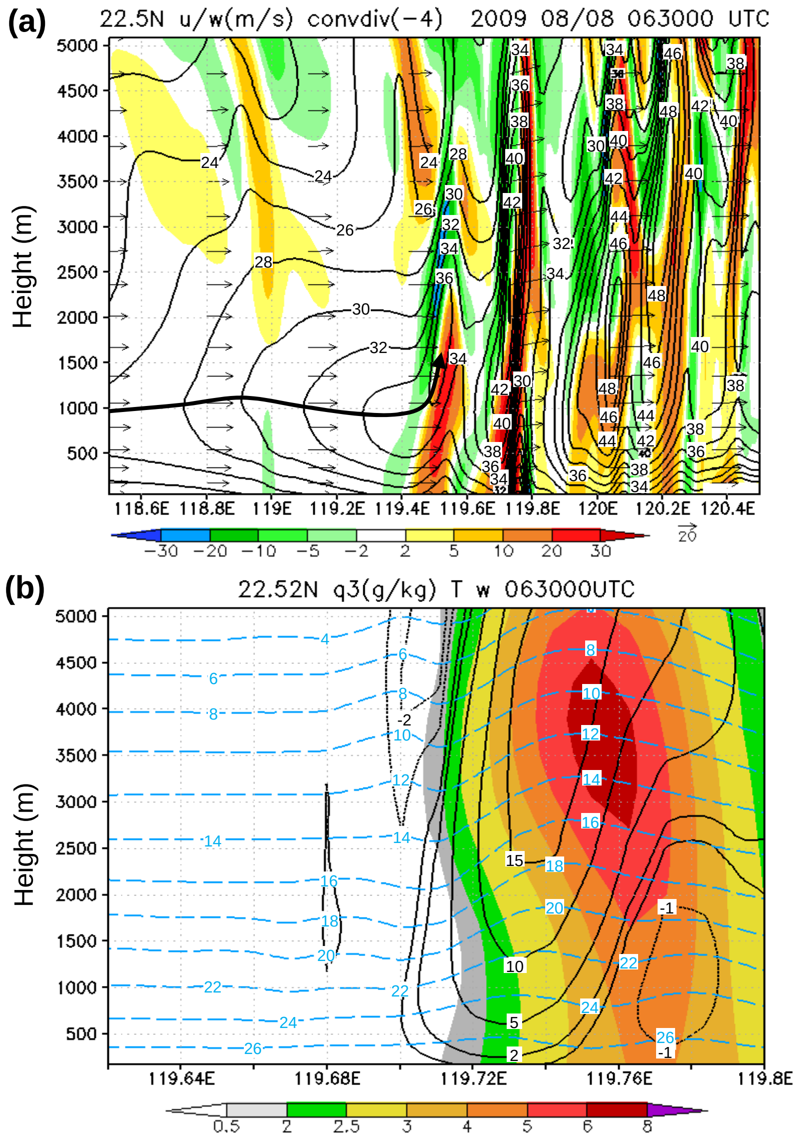


Figure 10. E-W vertical cross sections, through cell A1, of model-simulated (a) wind vectors on the section plane (m s^{-1}, reference vector at bottom), and horizontal wind speed (m s^{-1}, isotachs, every 2 m s^{-1}) and convergence/divergence (10^{-4} s^{-1}, color, scale at bottom, positive for convergence) along 22.5°N from 118.5 to 120.5°E (about 200 km in length), and (b) vertical velocity (m s^{-1}, contours, at ±1, ±2, ±5, ±10, and ±15 m s^{-1}, dotted for downward motion), temperature (°C, dashed cyan isotherms, every 2°C), and mixing ratio of total precipitation (g kg^{-1}, color, scale at bottom), along 22.52°N from 119.62 to 119.8°E (about 18 km in length) at 06:30 UTC 8 August 2009. The thick arrow-line in (a) marks the axis of LLJ in the background flow.

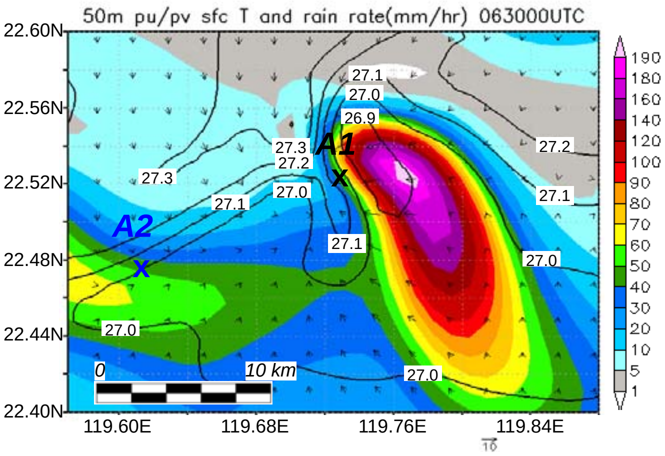


Figure 11. As in Fig. 9a, but showing model-simulated surface air temperature (°C, contours, intervals: 0.1°C), horizontal wind perturbation (u', v') at 50 m (m s^{-1}, vectors, reference vector at bottom), and (instantaneous) rainrate (mm h^{-1}, color, scales on the right) at 06:30 UTC 8 August 2009. The “x” marks the updraft/ascending centers of A1 and A2 at 50 m.

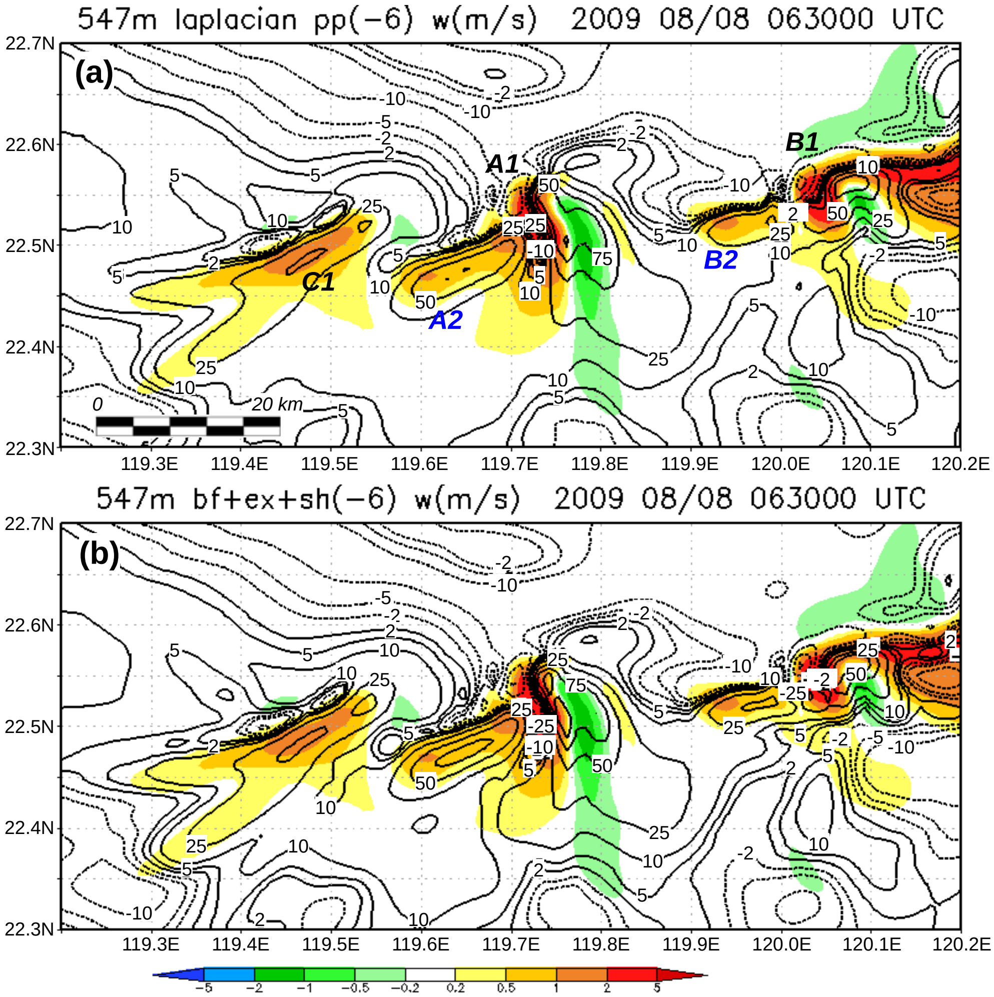


Figure 12. Model-simulated w (m s^{-1}, color, scales at bottom) and (a) $\nabla^2 p'$ (10^{-6} Pa m^{-2}, contours, dashed for negative values) computed from p' using the separation method and (b) $\nabla^2 p'_b + \nabla^2 p'_d$ obtained by adding all the rhs terms in Eqs. (11) and (13) together, at 547 m at 06:30 UTC 8 August 2009. Contour levels are at ±2, ±5, ±10, ±25, ±50, ±75, and ±100 × 10^{-6} Pa m^{-2}, respectively, and are the same in (a, b). The cells are labeled as in Fig. 8.

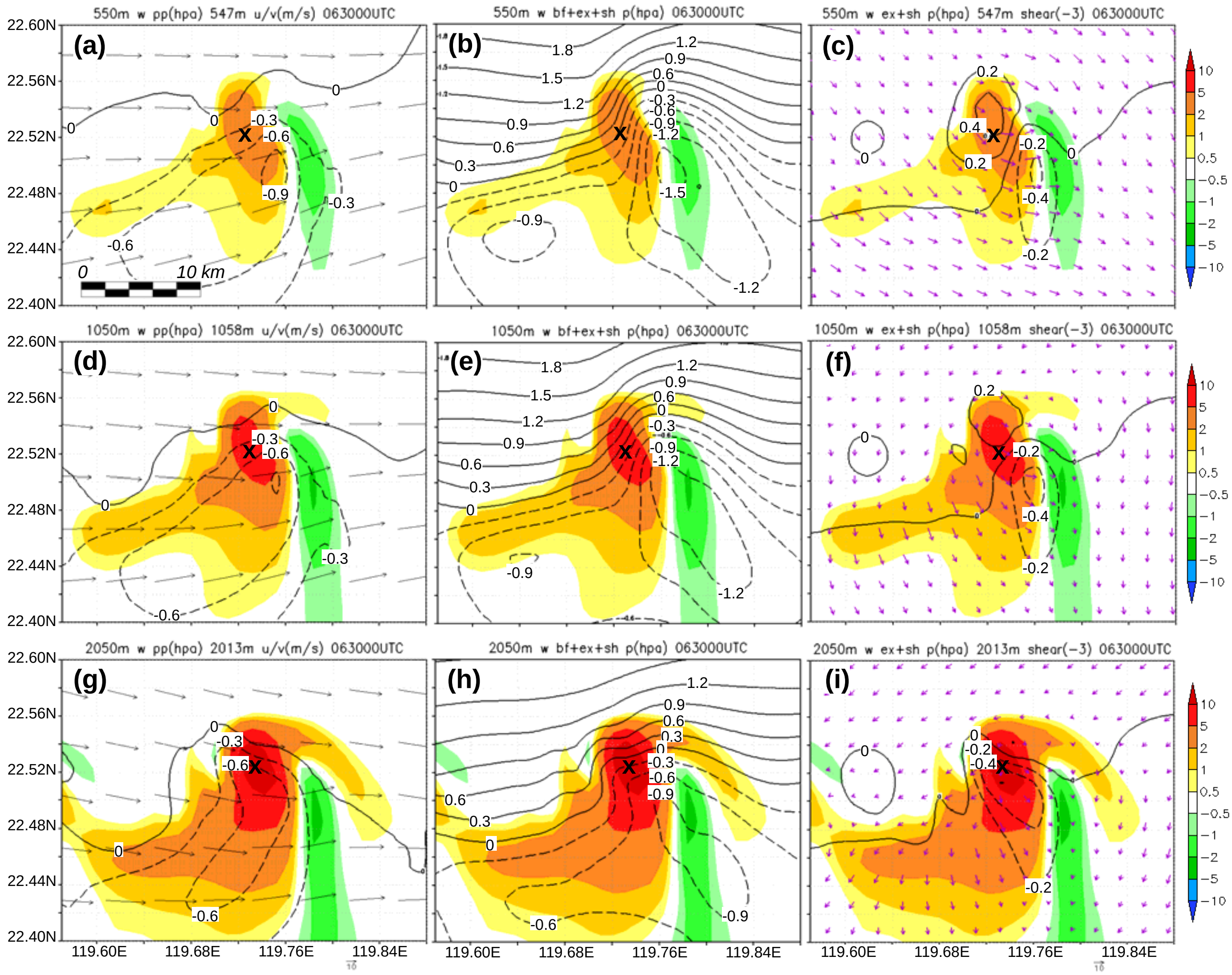


Figure 13. Model-simulated w (m s^{-1}, color, scales on the right) and (a) total p' (hPa, contours, every 0.3 hPa, dashed for negative values) and horizontal winds (m s^{-1}, vectors, reference vector at bottom), (b) $p'_b + p'_d$ (hPa, contours) solved by the relaxation method, and (c) p'_d (hPa, contours, every 0.2 hPa) solved by the relaxation method and vertical wind shear (10^{-3} s^{-1}, purple vectors, reference vector at bottom) at 550 m at 06:30 UTC 8 August 2009. (d-f) and (g-i) as in (a-c), except at 1050 and 2050 m, respectively. For horizontal wind and vertical wind shear, vectors at the closest model output level (547, 1058, and 2013 m, respectively) are plotted, and the "x" marks the updraft center in each panel.

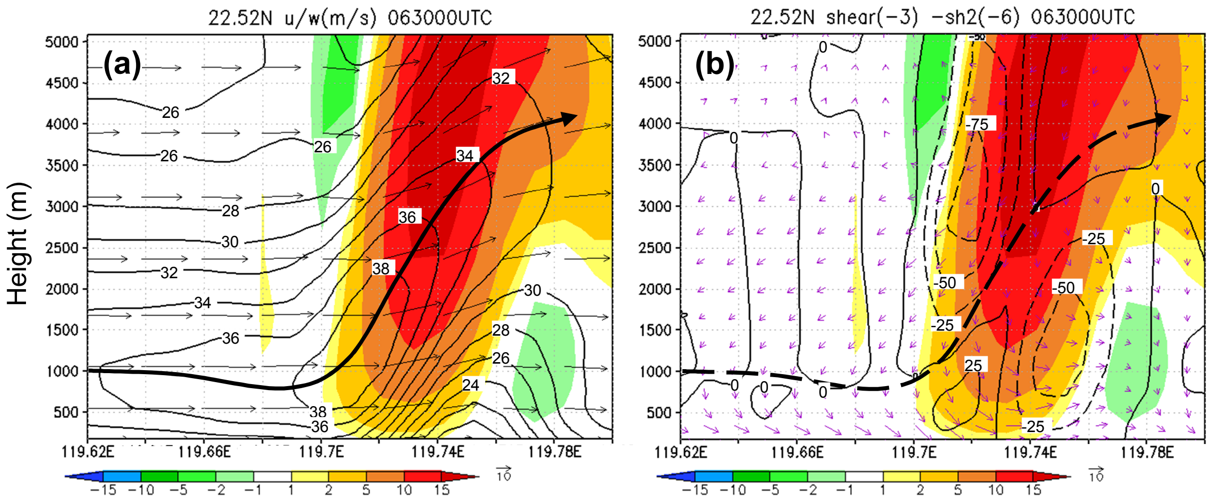


Figure 14. E-W vertical cross sections, through cell A1, of model-simulated w (m s^{-1}, color, scale at bottom) and (a) wind vectors on the section plane (m s^{-1}, reference vector at bottom), and u-wind speed (m s^{-1}, isotachs, every 2 m s^{-1}) and (b) vertical wind shear (10^{-3} s^{-1}, purple vectors, reference vector at bottom) and −1 times the x-component of shearing term in Eq. (13) (10^{-6} Pa m^{-2}, every 25 × 10^{-6} Pa m^{-2}, dashed for negative values) along 22.52°N from 119.62 to 119.8°E (about 18 km in length) at 06:30 UTC 8 August 2009. The thick (dashed) arrow lines in (a) and (b) mark the axis of LLJ.

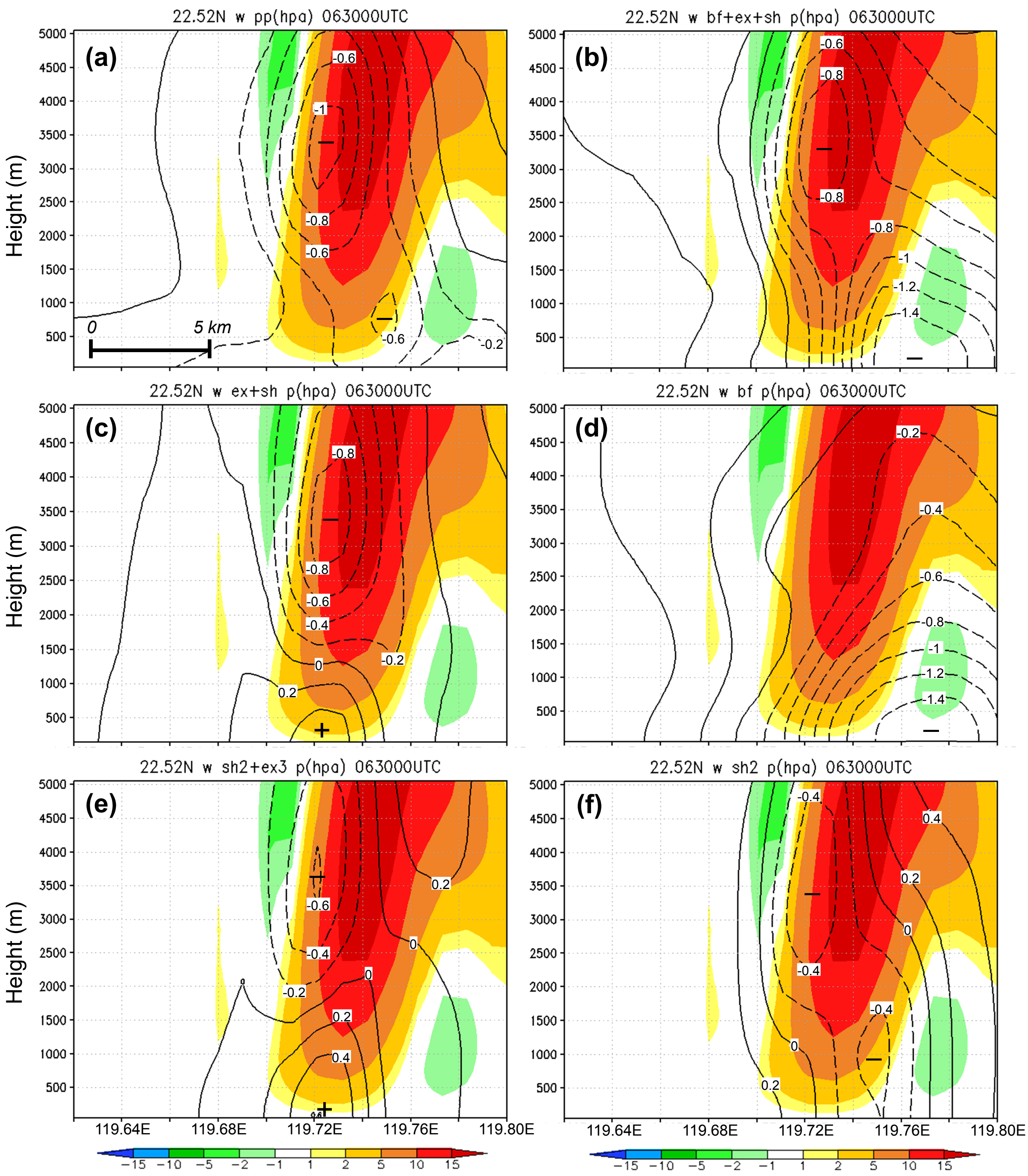


Figure 15. As in Fig. 14, except for w (m s^{-1}, color, scale at bottom) and (a) total p' (hPa, contours, every 0.2 hPa, dashed for negative values) separated from the background, and (b) $p'_b + p'_d$, (c) p'_d, (d) p'_b, (e) the portion of p'_d from SH2 and EX3, and (f) the portion of p'_d from SH2 alone (all in hPa, contours) in Eq. (13) solved by the relaxation method. The positive and negative centers are labeled by plus and minus signs, respectively.

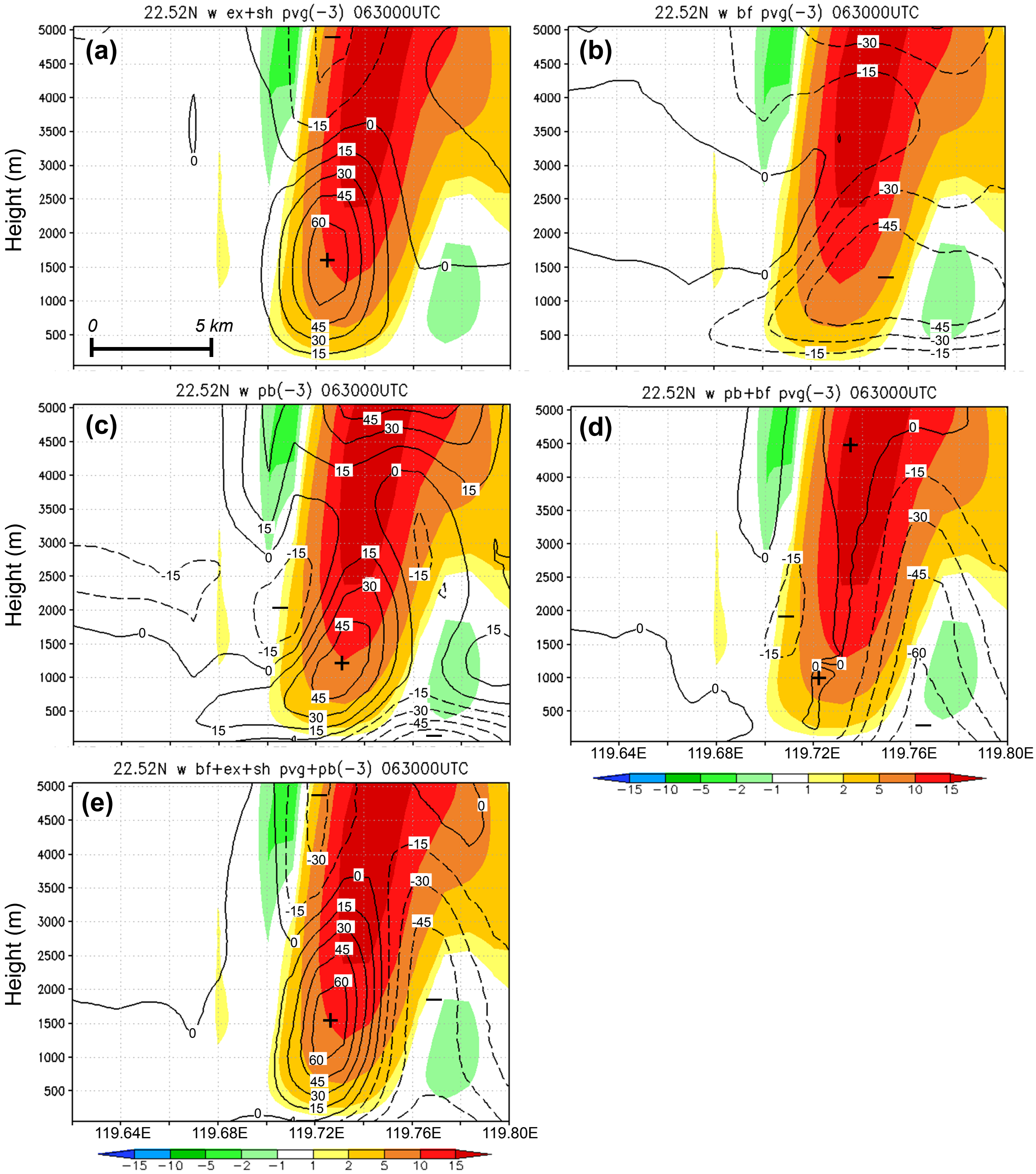


Figure 16. As in Fig. 14, except for w (m s^{-1}, color, scale at bottom) and the vertical (z) component of perturbation pressure gradient force (10^{-3} m s^{-2}, contours, every 15×10^{-3} m s^{-2}, dashed for negative values) from (a) p'_d, (b) p'_b, (c) buoyancy (B), (d) $p'_b + B$, and (e) $p'_d + p'_b + B$, respectively. The positive and negative centers are labeled by plus and minus signs, respectively.

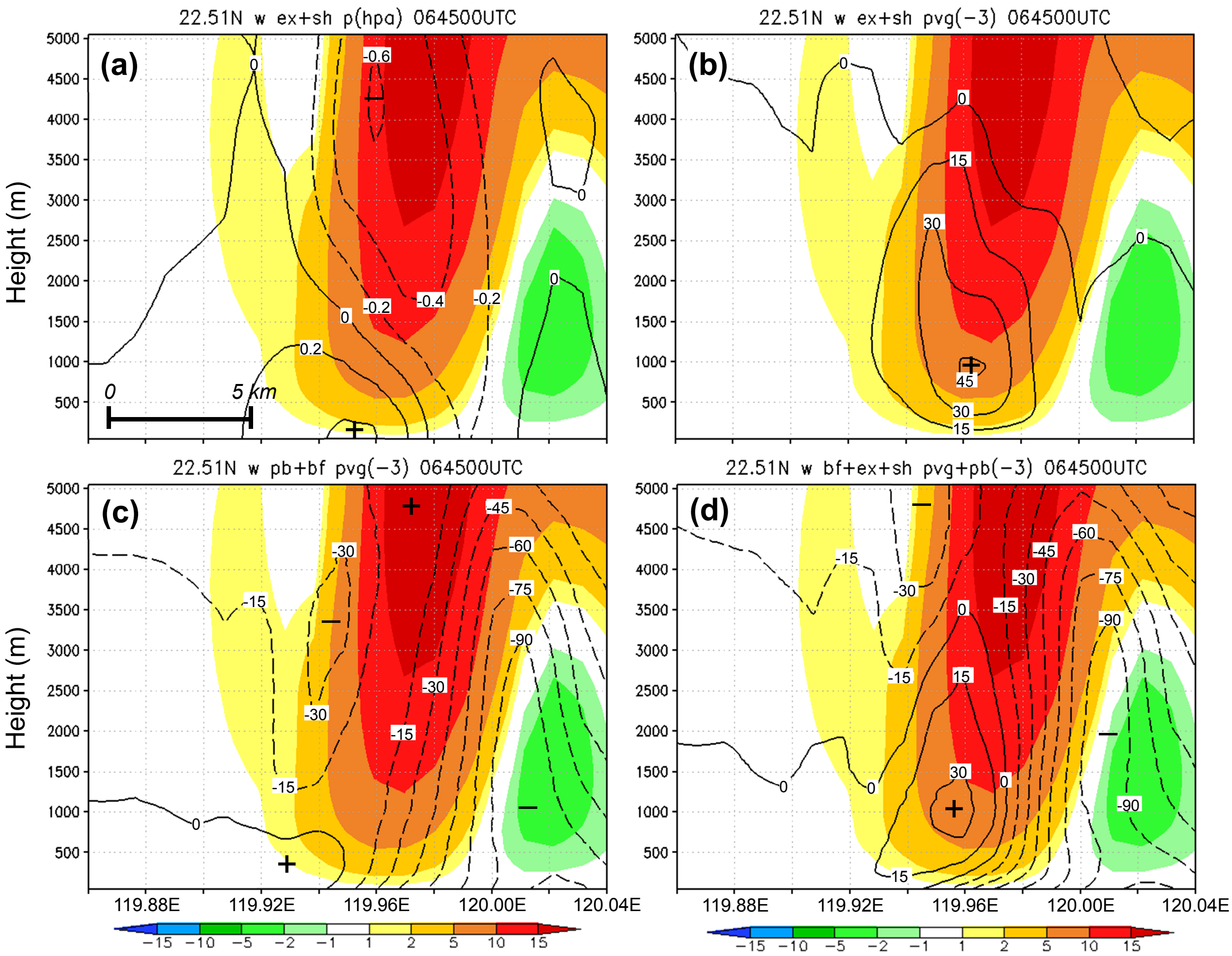


Figure 17. (a) As in Fig. 15c, except for w (m s^{-1}, color) and p'_d (hPa, contours, every 0.2 hPa) along 22.51°N from 119.86 to 120.04°E (about 18 km in length) at 06:45 UTC 8 August 2009. (b)-(d) As in Fig. 16, except for w and the z-component of perturbation pressure gradient force (10^{-3} m s^{-2}, contours, every 15 × 10^{-3} m s^{-2}) from (b) p'_d, (c) $p'_b + B$, and (d) $p'_d + p'_b + B$, respectively, at 06:45 UTC along the same cross-section as (a).