Prediction and diagnosis of Tropical Cyclone formation in an NWP system. Part I: The critical role of vortex enhancement in deep convection.

K. J. Tory[*1], M. T. Montgomery[2], and N. E. Davidson[1]

[1]Bureau of Meteorology Research Center, GPO Box 1289K, Melbourne, VIC, 3001, Australia

[2]Department of Atmospheric Science, Colorado State University, Fort Collins, CO 80523-1371

Submitted: November 17, 2004

Resubmitted: August 28, 2005

[*] Corresponding author address: K.J. Tory, Bureau of Meteorology Research Centre, GPO Box 1289K, Melbourne, Victoria, 3001, Australia. E-mail address: k.tory@bom.gov.au.

Abstract

This is the first of a three-part investigation into Tropical Cyclone (TC) genesis in the Australian Bureau of Meteorology's Tropical Cyclone Limited Area Prediction System (TC-LAPS), an operational Numerical Weather Prediction (NWP) forecast model. The primary TC-LAPS vortex enhancement mechanism is presented in Part I, the entire genesis process is illustrated in Part II using a single TC-LAPS simulation, and in Part III a number of simulations are presented exploring the sensitivity and variability of genesis forecasts in TC-LAPS.

The primary vortex enhancement mechanism in TC-LAPS is found to be convergence/stretching and vertical advection of absolute vorticity in deep intense updrafts, which result in deep vortex cores of 60—100 km in diameter (the minimum resolvable scale is limited by the 0.15° horizontal grid spacing). On the basis of the results presented we hypothesize that updrafts of this scale adequately represent *mean* vertical motions in real TC genesis convective regions, and suggest that explicitly resolving the individual convective processes may not be necessary for qualitative TC genesis forecasts. Although observations of sufficient spatial and temporal resolution do not currently exist to support or refute our proposition, relatively large-scale (30 km and greater), lower to middle-level tropospheric convergent regions have been observed in tropical oceanic environments during GATE, EMEX and TOGA-COARE, and regions of extreme convection of the order of 50 km are often (remotely) observed in TC genesis environments. The vortex cores are fundamental for genesis in TC-LAPS. They interact to form larger cores, and provide net heating that drives the system-scale secondary circulation, which enhances vorticity on the system-scale akin to the classical Eliassen problem of a balanced vortex driven by heat sources. These secondary vortex enhancement mechanisms are documented in Part II.

Vortex enhancement in some recent TC genesis theories featured in the literature, largely ignore the deep convective mode. Instead, they focus on the stratiform mode. While it is recognized that vortex enhancement in the stratiform precipitation deck can greatly enhance mid-tropospheric cyclonic

vorticity, we suggest this mechanism only increases the potential for genesis, whereas vortex enhancement via the deep convective mode is necessary for genesis.

1. Introduction

As part of the ongoing development of the Australian Bureau of Meteorology's tropical cyclone (TC) version of the Limited Area Prediction System (LAPS), a detailed investigation of the TC life-cycle in TC-LAPS is in progress. Clearly one of the most important aspects of the TC forecast is genesis, i.e., will a TC develop in the forecast area or not. Before improvements can be made to the prediction system it is necessary to develop a basic understanding of genesis in the model, and an understanding of how realistic the model depiction of genesis is. As is frequently noted in the literature regarding genesis and intensification (e.g., Gray, 1998) there are many wrong ways to the right answer. Thus, if the model is performing well for the incorrect reasons, improvements in the realism of certain aspects of the modeling system could in fact reduce the performance of the system. With this in mind, a detailed examination of TC genesis in TC-LAPS is underway and the first round of results is presented in this study and two companion papers, Tory et al. (2005a,b, hereafter Part II and Part III). In this paper (Part I) the TC-LAPS genesis processes are introduced, and the primary vortex enhancement mechanism illustrated. Part II provides a detailed diagnostic analysis of the genesis processes and includes forecast verification for a simulation of TC Chris (February 2002, off the Western Australian coast). Sensitivity and variability of the genesis processes are examined in Part III, where a number of developing and non-developing TC genesis simulations are presented.

Tropical Cyclone genesis has been described as the process that leads to the development of a self-sustaining surface-concentrated vortex, in which the flux of energy from the sea-surface to the vortex, governed partly by the intensity of the vortex, is sufficient to maintain and amplify the vortex, i.e., the hurricane heat engine (e.g., Wind Induced Surface Heat Exchange, WISHE Emanuel, 1986; Rotunno and Emanuel, 1987). Alternatively, Saunders and Montgomery (2004), Hendricks et al. (2004, hereafter H04) and Montgomery et al. (2005, hereafter M05), each described the genesis process as the development of a warm core vortex that extends from the surface to at least the mid-troposphere. Both definitions require the generation of strong cyclonic surface vorticity.

Over the last decade or so the search for the process that generates such a finite amplitude surface vortex has focussed on the observation that TC formation is associated with Mesoscale Convective Systems (MCSs) and their accompanying Mesoscale Convective Vortices (MCVs). Although the dynamics and thermodynamics of mid-latitude terrestrial MCSs and their accompanying MCVs has been extensively studied and reasonably well documented (e.g., Fritsch et al., 1994; Raymond et al., 1998; Houze, 2004), comparatively little has been published about their tropical oceanic counterparts. Mapes and Houze (1995; hereafter MH) used airborne Doppler radar to construct vertical profiles of mean horizontal divergence in tropical oceanic MCSs to identify the dominant precipitation modes. These profiles can provide insight into the type of vorticity structure one might expect in such MCSs.

The well-documented mid-latitude terrestrial MCS is dominated by stratiform precipitation, and an associated mean divergence profile consisting of mid-level convergence with divergence above and below. Because convergence (divergence) increases (decreases) the absolute vorticity magnitude in a rotating environment, the MCV consists of a cyclonic vortex maximized in the middle troposphere. Evaporative cooling below the MCS is often responsible for a surface anticyclone. MH identified, on the other hand, both stratiform and convective mode divergence profiles in a tropical oceanic environment. Averaging over diameters of 30—60 km they identified examples of almost pure stratiform and pure deep convective convergence profiles, however most were comprised of combinations of both (see also Houze, 1997). The deep convective divergence profile consists of convergence in the low- to mid-troposphere and divergence above. It follows that in a deep convective rotating environment the absolute vorticity magnitude would be expected to increase throughout the low- to mid-troposphere and decrease above.

The majority of profiles presented in MH included non-trivial convergence from the surface to middle troposphere, which suggests the deep convective mode often dominated at low-levels. The mean divergence profiles in MCSs of this type within a TC genesis environment (non-trivial low- to mid-level cyclonic absolute vorticity) should, enhance cyclonic vorticity from the low- to mid-levels and lead to the generation of cyclonic vortex cores of equivalent depth.

The TC genesis theory proposed in Simpson et al. (1997) and Ritchie and Holland (1997), did not consider the convective mode in the genesis process. Although they acknowledged that it could exist, they believed it to act only as a TC enhancement mechanism after genesis was complete. The TC genesis theory of Bister and Emanuel (1997) considered the convective mode only in the final stages of genesis. Both theories were based on the mid-latitude terrestrial MCV conceptual model that includes a surface anticyclone (as mentioned above) and a stratiform mode considerably greater in scale than the convective mode. Thus these theories first required a mechanism to replace the surface anticyclone with cyclonic vorticity from above. The main focus of both theories was on mechanisms that bring mid-level cyclonic vorticity to the surface. Fritsch et al. (1994), when discussing a mid-latitude terrestrial MCV that propagated out to sea, commented that surface fluxes would likely erode the surface cold pool and associated anticyclone. It might be expected that the same surface fluxes in the tropical oceanic MCV would reduce the likelihood of a surface anticyclone. Indeed Simpson et al. (1997) commented that there was no surface anticyclone present during the genesis of TC Oliver (the event they used to illustrate their genesis theory). Using a mid-latitude terrestrial MCV conceptual model as a basis for their genesis theories may have been misleading, however, because they did not consider the potential for vorticity enhancement by the convective mode, and their theories presumed the existence of a mesoscale surface anticyclone that may not necessarily exist in tropical oceanic environments.

In this paper we advance the hypothesis that vortex enhancement through convergence associated with the convective mode is critical for TC genesis. Since, as of yet, there is insufficient observational data to prove or refute this proposition, it must remain a hypothesis until tested further. We acknowledge that the stratiform mode is likely to play a role in TC genesis by enhancing large areas of mid-level vorticty, particularly early in the genesis process but we suggest genesis will not proceed without vortex enhancement associated with the convective mode. Furthermore, in some instances there might be a system-scale transition from vortex enhancement dominated by the stratiform mode to that dominated by the convective mode. (Note, this transition is opposite to the typical individual MCS life-cycle, where the young MCS is dominated by the convective mode, before the stratiform region has time to grow.) If such a transition does exist it may vary from basin to basin. For example the dry Saharan air layer in the Atlantic basin is likely to support the stratiform mode by enhancing low-level evaporation, which both strengthens the stratiform circulation, and inhibits the growth of large areas of near downdraft free convection.

There are a number of modeling studies in support of the hypothesis that the deep convective mode is critical for genesis. Montgomery and Enagonio (1998) considered the convective mode when they looked at the interaction of an MCV with a low- to mid-level vortex core that would likely be generated by vorticity convergence into a convective hot tower. They found a single upright cyclonic vortex core resulted, and proposed such an interaction may provide the necessary low-level vortex enhancement for TC genesis. More recent studies by H04 and M05 found the convective mode to be critical for TC genesis in both realistic and idealized models. They found intense vortices were generated in the low- to mid- troposphere through vorticity convergence into convective hot towers. These vortices interacted to form larger low- to mid-level vortices. They also found the sum of diabatic heating from condensation and adiabatic cooling from expansion in the hot towers was slightly positive, and the net heating from all hot towers enhanced the system-scale secondary circulation in a process akin to the Eliassen balanced vortex model forced by heat sources (Eliassen 1951). They termed these two processes the vortex upscale cascade, and the System Scale Intensification (SSI) respectively.

More recently, an observational study by Reasor et al. (2005) identified low-level vortex enhancement in the vicinity of a convective hot tower (consistent with H04 and M05) in Doppler radar observations during the genesis of Hurricane Dolly. They also identified mid-level vortex enhancement in a stratiform precipitation region. The spatial resolution of their observational data was sufficient to identify both convective and stratiform vortex enhancement as separate entities within regions of similar scale to MH's averaging domains. Although they did not carry out a quantitative partitioning of the two modes, their results suggest that the two modes act together to enhance cyclonic vorticity in a TC genesis environment.

In Part I of this series we document the primary vortex enhancement mechanism in TC-LAPS, and show that the convergence associated with the convective mode is responsible for the construction of low- to mid-tropospheric PV/vortex cores. We show in Part II how these cores contribute to the construction of a PV/vortex monolith that forms the basis of the TC, through the vortex upscale cascade and SSI processes of H04 and M05. In Part III we begin to identify common TC genesis features, and necessary conditions for TC genesis in TC-LAPS, using simulations of a number of developing and non-developing TCs.

An outline of Part I is as follows. The modeling system used, TC-LAPS, is described in Section 2. The primary low-level vortex enhancement mechanism active in all TC-LAPS genesis simulations is illustrated in Sec. 3. Observational evidence in support of this mechanism and the ability of TC-LAPS to adequately represent the mechanism is presented in Sec. 4. The work is summarized in Section 6.

2. TC-LAPS: model description

LAPS is an operational Numerical Weather Prediction (NWP) forecast model. A number of domains of varying sizes and horizontal resolution are run twice daily, together with a global model, to make up the suite of NWP forecasts produced by the Australian Bureau of Meteorology. The largest LAPS domain is illustrated in Fig. 1. It encompasses the Australian continent and much of the surrounding oceans and seas, and neighboring islands (longitude 65.0° to 184.625°, Latitude –65.0° to 17.125°). The horizontal grid resolution of this domain is 0.375° (LAPS375). LAPS375 is nested in the Bureau of Meteorology’s global model GASP (Global Assimilation Prediction System). Three mesoscale versions of LAPS are nested (one-way) in LAPS375, as well as a TC version (TC-LAPS) for monitoring tropical systems of interest that develop in the LAPS375 domain[1]. By its nature the domain of TC-LAPS is not fixed. An example of the TC-LAPS domain embedded in LAPS375 is shown in Fig. 1 (large rectangle). It is the TC-LAPS domain used in the simulation of TC-Chris discussed in the next section, and in Part II. The operational resolution is 0.15° on 29 σ-levels and covers a 27° by 27° area centered on the system of interest.

A detailed description of LAPS is provided by Puri *et al.* (1998), and summary of the LAPS components, including a number of additional options now available and used in TC-LAPS, is outlined here. The Miller-Pearce time-stepping scheme (Miller and Pearce, 1974) is combined with a third-order upwinding advection scheme, and implemented on an Arakawa A-grid (non-staggered). The European Centres for Medium-range Weather Forecasts (ECMWF) boundary layer parameterization and land-surface schemes (ECMWF Research Department, 1995) are now available and are used in all operational LAPS models. The vertical diffusion parameterization employs a Monin-Obukhov surface layer with first order closure schemes for determining the exchange coefficients (eddy diffusivities) above the surface. Under stable conditions the exchange coefficients are determined from mixing length theory using Richardson number-dependent stability functions. For unstable conditions they are scaled by the boundary layer height, based on Troen and Mahrt (1986). Each version of LAPS now includes 29 levels in the vertical, including 9 in the lowest (approximately) 1000 m.

The Tiedke mass flux convection scheme (Tiedke, 1989) is used to parameterize sub-grid scale convection. The penetrative convection component is triggered where sub-cloud-layer moisture convergence is positive. The cloud base mass flux is determined from the sub-cloud-layer moisture convergence, which together with the mass convergence in the lower half of the cloud due to organized entrainment provides the basic closure for the paramaterization scheme. The scheme takes into account:

[1] An additional LAPS domain is centered on the equator (T-LAPS, longitude 70.0° to 189.625°, Latitude –45.0° to 44.625°). It provides nesting for TC-LAPS in tropical regions north of the LAPS375 domain.

turbulent entrainment and detrainment; cloud-top detrainment at and above the zero-buoyancy level; downdrafts, when sufficient dry environmental air mixed into the cloud evaporates cloud water and cools the air until it becomes negatively buoyant; precipitation; and evaporation of precipitation in the cloud and in cloud-air detrained into the environment. This convection parameterization influences only the grid scale temperature and humidity tendencies. For grid resolved updrafts, any moisture in excess of 100 % humidity is removed as rain.

The radiation scheme (Fels and Schwartzkopf, 1975) has not changed since Puri *et al.* (1998), although the radiation tendencies are now updated hourly instead of 3-hourly to better resolve the diurnal cycle. The analysis system (multivariate statistical interpolation, MVSI) is also largely unchanged. An option to use sea surface temperature (SST) analyzed daily has been added, but not used here (weekly averaged SST is used). Currently, there is no feedback between the ocean and atmosphere in any of the LAPS systems.

Dynamic Initialization

The initialization procedure follows Davidson and Puri (1992) with only a few changes. A summary of the current procedure follows. For genesis simulations no bogus vortices are included in the initialization. Initialization is performed in two steps. First the analysis step, where high-resolution analyses (called target analyses) are obtained at 6-hourly intervals during the initialization period. Second the dynamical nudging step. Dynamical nudging is used to "grow" a numerically consistent and balanced initial state that best represents the real atmosphere, by nudging a numerical forecast, during the simulation, towards an analyzed state of the atmosphere (the target analyses generated in step 1). In the current study and in the operational TC-LAPS the dynamical nudging is performed over the 24-hour period prior to the initial forecast time.

-- Analysis step

The target analyses are produced using the objective analysis system described in Puri *et al.* (1998), and an assimilation scheme involving four 6-hour model integrations. LAPS375 or T-LAPS analyses are used to initialize the first 6-hour model integration and for boundary conditions throughout the assimilation. Objective analyses are performed at the end of each 6-hour model integration, and the resulting fields provide both the initial conditions for the next 6-hour model integration, and the target analyses used in the dynamic nudging step. The observation base used by the objective analysis includes all standard observational data, together with the additional scatterometer and surface wind observations included in the research version of TC-LAPS used here.

-- Dynamical nudging step

The T = -24 hour target analysis is used to initialize TC-LAPS during the 24-hour period of dynamical nudging. Linear interpolation in time between the neighboring target analyses provides a target analysis at every nudging time-step during the procedure. Throughout the 24-hour period the numerical solution is nudged towards the analyzed vorticity, while allowing the model to develop its own divergence. This leads to the preservation of the observationally reliable rotational wind component in the target analyses, while allowing the model-generated divergent wind-field to dominate over the less reliable analyzed divergent winds. The surface pressure and temperature fields are also nudged towards the target analyses to preserve the mass field.

As mentioned in Davidson and Puri (1992), this form of dynamical nudging does not guarantee the convection will develop in the right place or time. To overcome this problem Cloud Top Temperature nudging is applied, where artificial heat sources are used to force model convection in regions of deep convection identified by satellite observations. This forcing is applied during the assimilation process as well as the dynamical nudging. The method is well described in Davidson and Puri (1992) and will not be elaborated on here except where changes to the procedure have been made. These changes include the reduction of the triggering temperature (the maximum cloud top temperature at which the heat source is applied) from 273 to 233 K, to ensure forced convection is only applied where the observed convection is deep. Another change, although minor, involves nudging of the numerically predicted atmosphere towards the imposed heating profile, rather than the direct replacement of the convective parameterization with the heating profile, as mentioned in Davidson and Puri (1992).

3. Primary TC-LAPS vortex intensification mechanism

Here we document the primary vortex intensification mechanism present in all TC-LAPS genesis simulations: horizontal convergence (or stretching) of absolute vorticity in deep convective cores. The mechanism in itself is not responsible for vortex amplification to TC-intensity. Instead the mechanism builds vortex cores that interact with one another (H04 and M05's vortex merger), and provide heating that fuels the SSI process (H04, M05). (Hereafter, we label the vortex merger and SSI processes as secondary vortex enhancement mechanisms.) In this way the primary mechanism indirectly contributes to the construction of a monolithic vortex structure of sufficient intensity and scale for the self-sustaining and self-amplification process (e.g., WISHE amplification) to set in. The construction of the vortex monolith in a simulation of TC Chris is analyzed and described in detail in Part II.

The horizontal convergence/stretching of absolute vorticity in deep convective cores was described by H04 and M05. They labeled the resulting vortices Vortical Hot Towers (VHTs). The main difference between the TC-LAPS convective vortex cores and VHTs is the scale of the updrafts and resulting vortices. As mentioned in Section 2, the TC-LAPS horizontal grid-spacing is 0.15˚ (about 15 km), giving a minimum resolvable convective scale of about 60 km. Typically the resolved updrafts are of the order of 60—100 km in diameter[2], about 4—5 times greater than those documented in H04 and M05, and extend to depths greater than 14 km. The convection in these updrafts is both explicit and parameterized. At 0.15˚ grid spacing the individual convective cells are not resolved and thus a parameterization scheme is required. The parameterization scheme, in theory, adjusts the temperature and moisture fields to represent the effect of the unresolved convection. It does not transport mass;

[2] Horizontal smoothing has been applied to the vertical motion plots featured in this paper, which tend to give the appearance of greater horizontal scale to the updrafts.

instead the explicit, or resolved, updrafts represent the mean vertical motion within the larger convective region. It is not clear how realistic this combination is, but it is clear that some form of convective parameterization is required and that the parameterization alone can not represent the relatively large deep-convective mode divergence profiles observed by MH in tropical oceanic MCSs. Below the mature updrafts boundary layer subsidence (up to 400 m deep) is often present with associated cooling of up to 2 K (not shown). This subsidence is likely to have developed in response to the temperature and moisture field adjustment associated with the parameterization of downdrafts and evaporative cooling.

Examples of the TC-LAPS updrafts are shown in Fig. 2. The images in this figure come from the simulation of TC Chris featured in Part II, and show a cycle of updraft development and decay between 4, 6 and 8 hours into the simulation. Vertical motion on the σ = 0.25 level (approximately 10 km) is combined with the low-level horizontal winds to identify regions of deep intense updrafts embedded in low-level cyclonic flow. In the left panel there are two updraft cores labeled A and B. Core A is an old updraft undergoing decay, while Core B is relatively young. The right panel shows Core B in its mature stage, with only a small remnant of Core A remaining. The lower panel shows the emergence of a third updraft labeled C just prior to the decay of Core B. Vertical cross sections of vertical velocity, absolute vorticity, and the contributions to absolute vorticity tendency from vertical advection and stretching, are shown in Fig. 3, for Core B during the developing, mature and decaying stages (same times as Fig. 2). The white lines in Fig. 2 indicate the locations of the cross-sections. There appears to be a pattern present in a number of TC-LAPS updrafts in which the maximum updraft intensity occurs at relatively low levels early in the life of the updraft and at progressively higher levels as the updraft matures, then decays. This is evident for updraft B in Fig. 3. Note also that updraft A, which is in a state of decay, contains a maximum intensity at an elevated level (near 9 km). The second row of Fig. 3 shows the intensification of absolute vorticity at low- mid-levels in the vicinity of Core B. During this 4-hour period the maximum cyclonic vortex intensity increased from about -3×10^{-4} s^{-1} to about -8×10^{-4} s^{-1} (southern hemisphere), with the maximum value located in the lowest 2 km.

The tendency equation for the vertical component of absolute vorticity (ς_a) can be expressed as,

$$\frac{\partial \varsigma_a}{\partial t} = -\overline{u}_h.(\nabla_h \varsigma_a) - \omega \frac{\partial \varsigma_a}{\partial p} - (\nabla.\overline{u}_h)\varsigma_a - \left(\frac{\partial \omega}{\partial x}\frac{\partial v}{\partial p} - \frac{\partial \omega}{\partial y}\frac{\partial u}{\partial p} \right),$$

where the subscript *h* refers to the horizontal components of the wind and the other variables have there usual meaning. The terms on the right hand side represent, horizontal advection, vertical advection, stretching/convergence and tilting of absolute vorticity. The contributions to absolute vorticity tendency from these terms were calculated and the dominant vortex enhancement terms, vertical advection and stretching/convergence, are illustrated in Fig. 3.

The contributions from the horizontal advection and tilting terms do not contribute to cyclonic vortex enhancement in the vicinity of the updraft core. Tilting opposed the contribution from vertical advection, but was considerably weaker in magnitude. The contribution from horizontal advection is more difficult to assess because it is highly dependent on the frame of reference in which the term is calculated. It represents both the advection of the vortex core within the background monsoon circulation, and the advection of vorticity into and out of the vortex core. The former, which is of no interest to understanding the development of the vortex core, can be removed by subtracting the background flow. However, because it is not uniform it is impractical to subtract the background flow at every point of interest. Instead we investigated the effects of horizontal advection into and out of the vortex cores at only a few points (not shown). We found anticyclonic (cyclonic) tendencies in convergent (divergent) regions due to the inward (outward) flow in an environment of increasing cyclonic vorticity towards the center of the vortex core, i.e., they opposed the tendencies from the convergence/stretching term. At low-levels where the convergence was maximized this opposition was up to 25% of the convergence/stretching term.

Figure 3 shows the vortex is enhanced at low- to mid-levels is by convergence/stretching of absolute vorticity in the deep intense updraft, and at mid- to upper-levels by vertical advection of absolute vorticity. Thus vorticity is being concentrated and stretched from below and advected upwards by the updraft. Figure 3 also shows how the tendency terms evolve with the updraft life cycle. During the developing phase, when the updraft maximum is located at relatively low-levels, the contribution from the stretching/convergence term is confined to low-levels, and it grows deeper with time as the location of the updraft maximum moves upward with time (associated with the evacuation of mass from the horizontal plane where the upward flow is accelerating)[3]. Also evident in Fig. 3 is anticyclonic growth on the left edge of the updraft above about 7000 m (compare the center and right panels, second row). This is due to tilting and appears to indicate that the net change in absolute vorticity is nearly zero at this level. This would be consistent with Haynes and McIntyre's (1987) comment that there can be no net change in absolute vorticity on a pressure surface. The PV associated with the same convective burst is presented in Fig. 7 of Part II, where we show in another cross section two hours later that the cyclonic anomaly does become considerably more intense at these levels. With time vortex advection effects lead to the ejection of the anticyclonic anomaly from the cyclonic core region.

To illustrate that this primary vortex enhancement mechanism is a very common feature of TC-LAPS simulations a similar analysis has been performed on another three updrafts from three additional TC forecasts. These are: the Elcho Island storm (Arafura Sea, January 2003, it reached TC intensity just prior to land-fall and as a consequence was not named), TC Fiona (Arafura Sea, February 2003) and TC Erica (Coral Sea, SW Pacific, March 2003). Part III contains a more detailed analysis of these events. The vertical velocity, absolute vorticity and two absolute vorticity tendency terms (vertical advection and convergence/stretching) for the three updrafts are presented in Fig. 4. This figure provides an indication of the variability of the process from case to case.

The size and intensity of the updrafts from the Elcho Island (left column, Fig. 4) and TC Erica simulations (right column) are examples of strong TC-LAPS updrafts, whereas the two updrafts evident

[3] The negative convergence tendency between 4000 and 8000 m in the decaying Core A (Fig. 3, lower, left panel) appears to contradict the combination of absolute vorticity and implied divergence from the panels above. Horizontal smoothing has merged two side-by-side negative tendency anomalies. On the left the absolute vorticity is anti-cyclonic and the flow is horizontally convergent. On the right the flow is divergent and absolute vorticity cyclonic.

in the TC Fiona simulation (center column) are particularly weak. The Elcho Island "snap-shot" was taken during a time of rapid development on both the updraft and system-scales (5 hours into the forecast). The rapid development on the updraft scale is evident in the two tendency terms in Fig. 4, which are of similar intensity to the mature updraft from TC Chris mentioned above (center column, Fig. 3). One obvious difference is the negative contribution to intensification from the vertical advection term below 2800 m, which results from the positive vertical gradient of cyclonic absolute vorticity in this layer (i.e., the absolute vorticity maximum is located at 2800 m). This low-level negative contribution is more than compensated for by the strong positive contribution from the stretching/convergence term, which leads, with time, to a downward migration of the maximum in cyclonic absolute vorticity. It is interesting to note that such a downward migration would give the appearance of the vortex growing down from mid-levels, when clearly the dynamics describe intensification from below. This phenomenon has been observed in a number of TC-LAPS simulations, and is also evident in Chen and Frank (1993).

In Chen and Frank a numerical model was used to simulate the growth of an MCS and associated MCV. Between 4 and 8 hours into the simulation a TC-LAPS-like updraft core developed and the subsequent vortex development was almost identical to that shown in Fig. 4 for the Elcho Island storm, except the vortex maximum was deeper, and as a consequence the tendency terms extended over a deeper layer of the lower troposphere. This higher initial mid-level vorticity maximum in Chen and Frank led to a more pronounced downward moving vorticity maximum with time. It would appear that they assumed all model resolved flow in the MCS to be associated with stratiform precipitation dynamics. As a result they concluded that the vortex enhancement they observed to grow downwards resulted from the stratiform mode, when clearly the divergence profile was consistent with the deep convective mode. This led them to conclude that stratiform dynamics can lead to downward growth of a MCV to the surface. It is possible that this conclusion may have influenced the search for genesis mechanisms that bring mid-level vorticity down to the surface, while focusing almost entirely on the stratiform mode.

The TC Fiona simulation was characterized by very slow development on the system scale. This was due in part to relatively weak and short-lived updrafts, and relatively long periods between convective outbreaks that fuelled the genesis process. The weak contributions to vortex intensification on the updraft scale are evident in Fig. 4 (compare with the other two examples). Clearly, as Eq. 1 would suggest, the stronger the updraft and the greater the absolute vorticity the updraft is embedded in, the greater the vortex intensification.

This point is very apparent in the case of TC Erica, where the two tendency terms are about double the magnitude of the Elcho Island and TC Chris terms. Despite this very significant vortex enhancement on the updraft scale, the TC Erica simulation failed to “spin-up” a TC (see Part III). This particular updraft formed about 100 km outside of a monsoon gyre, and about 300 km from the center of the gyre. It was also subjected to low-level shear. The shear in itself may have been sufficient to inhibit development, however it is also possible that the cyclonic environment in the vicinity of the convection was not sufficient large or intense to sustain the development (see Part III).

4. Discussion

In the previous section we have described a vortex intensification mechanism active in the TC-LAPS updrafts, and we note that it is the primary mechanism responsible for TC genesis in that it provides seed vortices and net heating that drive the secondary mechanisms of vortex upscale cascade and the SSI process (Parts II and III). Clearly, a number of questions arise regarding the realism of the simulations.

1. Is the primary vortex enhancement mechanism active in the real atmosphere?

2. If so, is the primary vortex enhancement mechanism critical for TC genesis in the real atmosphere?

3. If so, are we adequately representing and resolving the mechanism?

Given that two prominent genesis theories (Simpson et al., 1997; Holland and Ritchie, 1997; and Bister and Emanuel, 1997) focus on the stratiform mode, and that the stratiform mode does not appear to play an obvious role in the TC-LAPS simulations, further questions arise.

4. Does the stratiform mode play an important role in TC genesis?

5. Does TC-LAPS adequately incorporate the stratiform mode?

Unfortunately, observational evidence of sufficient temporal or spatial resolution does not exist to answer these questions with any certainty. However, we believe the observations that do exist provide strong evidence in favor of the primary vortex enhancement mechanism as a critical genesis mechanism. This is backed up by contemporary modeling studies (H04 and M05). In presenting the argument for the primary vortex enhancement mechanism being active and important for genesis in the real world we will address the five questions posed above.

a) Is the primary vortex enhancement mechanism active in the real atmosphere?

Forecasters for decades have known that vortex intensification often follows periods of sustained deep convection in tropical systems ranging from depressions through to full blown hurricanes. Zehr (1992) documented such observations focusing mostly on satellite imagery from two seasons (1983, 1984) of tropical storms and typhoons in the northwest Pacific (50 events in total). Low-level U.S. Air Force reconnaissance flight-level data (at approximately 500 m above sea level) were available for many of the systems, which provided confirmation of the role such convection played in intensifying low-level vorticity. He found that after periods of greatly enhanced deep cumulonimbus convection the low-level vorticity was significantly enhanced. We note that such observations do not rule out the genesis theories based on vortex enhancement by the stratiform mode, since it is conceivable that decaying convection contributed to the construction of a stratiform precipitation deck. However, we feel it is unlikely that this mode was responsible for the low-level vortex enhancement reported by Zehr, given the response time required to build the stratiform precipitation deck (e.g., Houze, 2004) and then somehow transfer the vorticity down from mid-levels. Zehr suggests the intensification is evident in a matter of hours, which is consistent with direct low-level vortex intensification by convergence associated with the convective mode. Furthermore, in the TC genesis environment large regions of very cold cloud tops (significant overshooting) are often observed. Zehr showed one satellite image (Zehr's Fig. 7.5) of cold cloud top temperatures covering nearly 3°×3° square region of clouds less than -65°C with an almost 1°diameter region less than -80°C embedded, suggesting a large contiguous region of deep intense convection was active.

MH provide evidence in support of vortex intensification by the convective mode in tropical oceanic regions in non-genesis environments. They produced vertical profiles of mean horizontal divergence within ten MCSs observed during TOGA-COARE, and found that both the startiform and convective modes were active. On average both modes were significant, suggesting that vortex enhancement by the convergence associated with these modes would intensify vorticity from low- to mid-levels. They noted that some systems sampled were dominated by the stratiform mode and others by the convective mode. Data collected from two of the systems dominated by the convective mode, included a TC rainband (TC Oliver), and a young vigorous, near-downdraft free, convective region.

Further evidence of mesoscale low- to mid-level convergence in large convective regions has been found in the Global Atmospheric Research Program Atlantic Tropical Experiment (GATE) and the jointly conducted Equatorial Mesoscale Experiment (EMEX) and the Australian Monsoon Experiment (AMEX). Zipser and Gautier (1978) documented the intensification of a Tropical Depression (TD) during GATE near Dakar on 15 July 1974. They noted, among other things, mesoscale cyclogenesis was preceded by mesoscale organization of deep convection, accompanied by strong mesoscale convergence at low-levels; strong convective echoes were “within and downwind of a broadzone of 10^{-4} s^{-1} convergence at 990 mb that covers an entire 1° square”. Mapes and Houze (1992) commented that MCSs that developed in weak cyclonic depressions were characterized by very broad, deep bands of convective clouds, with large apparent upward mass flux. They illustrated this with an example from EMEX where an updraft exceeded 1 ms^{-1} continuously along a 40 km flight leg. Although the flight passed along a convective line, they noted that it was not a particularly narrow one.

Finally, Reasor et al. (2005) provide evidence of low-level vortex enhancement associated with deep convection during the genesis of Hurricane Dolly.

b) Is the primary vortex enhancement mechanism critical for TC genesis in the real atmosphere?

While we cannot answer this question with any certainty yet, we feel there is sufficient evidence to suggest it is quite likely. As noted above: (i) bursts of intense convection have long been associated with vortex intensification, and TC genesis; (ii) we suggested the vortex enhancement response time is likely to be more consistent with Zehr’s observations for the convective mode mechanism than the indirect stratiform mode; and (iii) we believe the relatively large areas of very cold cloud tops represent overshooting in deep convective regions where the mean divergence profile is likely to be consistent with MH’s convective mode. For these reasons we associate the relatively short-lived (order 6 hours) but intense convective bursts that tend to occur sporadically throughout the genesis environment (e.g., Part II; Ritchie and Holland, 1997; Ritchie et al, 2003; Harr et al., 1996), with vorticity enhancement via the deep convective mode.

c) Are we adequately representing and resolving the primary vortex enhancement mechanism?

To answer the question regarding the adequate representation of the mechanism we need to be sure that the scale and intensity of the TC-LAPS updrafts realistically represent the vertical profile of the mean horizontal divergence. The observations of Zipser and Gautier (1978) from GATE, the Mapes and

Houze (1992) EMEX updraft observations, the MH observational study of convergence profiles taken during TOAGA-COARE, the large areas of very cold cloud top temperatures observed by Zehr (1992), and finally the comments by Gray (1998) that pockets of extreme convection on the scale of 50 km "sometimes act as the focus from which the centers of tropical cyclones develop", all suggest that relatively large areas (30 km or greater) of enhanced low- to mid-level convergence do exist in the oceanic tropical environment. A comment in Houze (1997, p2186) would suggest that deep convective convergence profiles observed in young vigorous convection during TOGA COARE *typically* reached 140 km in diameter. This latter scale is of the same order as the largest TC-LAPS updraft cores. The mean convergence in core B (Fig. 2 and 3) averaged over a diameter of 30 km yields values of $-4\times10^{-4}s^{-1}$ above the boundary layer decreasing in magnitude to about $-2\times10^{-4}s^{-1}$ near 3 km (not shown). This is about twice the intensity of the deep convective profiles measured by MH, although it should be noted that they did not sample deep convective regions in an intensifying TC genesis environment. On the other hand, when averaged over a $1°\times1°$ square it is of similar intensity to that measured by Zipser and Gautier (1978) in an intensifying Tropical Depression. Essentially this means the vertical profiles of horizontal mean divergence observed in TC-LAPS are of similar type and spatial scale as those frequently observed in tropical oceanic environments, and of similar magnitude to one measurement in a TC genesis-like environment.

To answer the question regarding the adequate resolution we need to understand the importance of the finer details of the convection for TC genesis. Currently the TC-LAPS updrafts are consistent with a MH divergence profile dominated by the convective mode, associated with a large convective region. The associated vortex enhancement fuels the secondary intensification mechanisms that lead to TC genesis. It would appear from the many simulations we have performed using TC-LAPS (some reported in Parts II and III) that the TC-LAPS updrafts responsible for the primary vortex enhancement mechanism are qualitatively accurate to drive TC formation and non-formation. However, the finer details may be quantitatively important when it comes to accurately forecasting the location, timing and intensity of the developing TC.

d) Does the stratiform mode play an important role in TC genesis?

MH have shown the stratiform mode is active in most non-genesis oceanic tropical environments, and that it tends to dominate in long-lived MCSs. Vortex structures, consistent with the well-documented mid-latitude terrestrial MCV that form in the stratiform precipitation deck, have been documented in the TC genesis environment (e.g., Bister and Emanuel, 1997; Reasor et al., 2005). Clearly, vortex enhancement by the stratiform mode mid-level convergence is favorable for TC genesis, in that it enhances the mid-level pool of vorticity. However, the associated low-level divergence weakens the absolute vorticity magnitude at low-levels, which is counter to the genesis process. Montgomery and Enagonio (1998) showed that the interaction of two vortices, one consistent with construction by the stratiform mode (traditional MCV) and the other consistent with construction by the convective mode (vortex core maximized at low-levels), leads to the generation of a single upright vortex core. Thus the stratiform mode is likely to be important for TC genesis when it interacts with a vortex enhanced by the convective mode. Thus, we believe the stratiform mode may play a role in improving the likelihood of genesis, but ultimately genesis will not proceed without vortex enhancement associated with the convective mode (i.e., the primary vortex enhancement mechanism).

e) Does TC-LAPS adequately incorporate the stratiform mode?

The MH vertical profiles of mean horizontal divergence have shown that MCSs can often be equally dominated by the stratiform and convective modes. It can be assumed in these cases the MCS being sampled consists of distinct convective and stratiform regions, with their typical respective divergence profiles (convergence maximized at low-levels decreasing to zero at mid-levels with increasing divergence above, and convergence maximized at mid-levels with divergence above and below, respectively). However, averaged over the whole sampling region significant convergence extends from low- to mid-levels. This is evident in the TC-LAPS updrafts, particularly the mature updrafts (e.g., see the lower panels of Figs. 3 and 4). Whether this is just the nature of convective updrafts in the TC-LAPS TC genesis environment (i.e., maximum diabatic heating at 7 or 8 km, rather than 5 or 6 km), or whether TC-LAPS is in some way incorporating the stratiform mode is yet to be determined. Whatever the reason it is evidence that a convergence profile consistent with a combination of stratiform and convective modes, dominated by the latter, is being incorporated in the TC-LAPS updrafts.

5. Summary and conclusions

The search for an understanding of TC genesis over the last ten years or so has focused on the observation that TCs often develop in the vicinity of an MCS. It was believed that the transition from MCS to a TC-like vortex required the generation of low-level vorticity below the MCS, and the search for a TC genesis mechanism became the search for a mechanism that provided this sub MCS low-level vorticity. Two of the earlier genesis theories focused almost entirely on the vortex enhancement in the stratiform region of the MCS (Simpson et al, 1997, Bister and Emanuel, 1997), and thus required mechanisms for bringing mid-level vorticity down to the surface. Another genesis theory suggested the necessary low-level vorticity could be provided by vortex enhancement by low-level convergence into deep convective regions (Montgomery and Enagonio, 1998). More recent modeling studies have highlighted the importance of vortex enhancement by the convective mode in realistic (H04) and idealized environments (M05). The convective mode has been identified as the dominant vortex enhancement mechanism in the TC-LAPS genesis simulations.

The vorticity tendency analysis of the TC-LAPS updrafts presented in Section 3 showed that vortex cores were generated by convergence and vertical advection of vorticity in convective updrafts. We have labeled this process the primary vortex enhancement mechanism. We show in Part II that the net heating generated in these cores drives the system-scale intensification process (H04, M05) and the vorticity generated in the cores contributes to the upscale vortex cascade (H04, M05), both secondary vortex enhancement mechanisms that lead to the construction of a monolithic vortex core that forms the basis of the TC.

The generation of deep vortex cores is fundamental to genesis in TC-LAPS, but it is not yet clear whether this mechanism is of equal importance in the real atmosphere. Evidence of deep convective regions of TC-LAPS scales do exist in tropical oceanic environments (Zipser and Gautier, 1978; Mapes and Houze, 1992; MH), and more specifically are very commonly observed in genesis environments (Zehr, 1992; Gray, 1998). These latter observations are consistent with the hypothesis we propose that deep convective regions are critical for genesis in the real atmosphere. The TC-LAPS resolved updrafts ideally represent the mean flow in real convective regions. It may be that the interaction of these updrafts with the larger-scale environment is of sufficient accuracy to forecast formation with qualitative success, whereas greater resolution may be required to adequately represent smaller scale features and interactions (and possibly separate convective and stratiform modes) before improvements in formation timing, rate and intensity can be made.

The stratiform vortex enhancement mode does not appear to play a significant role in the TC-LAPS simulations. The mid-level vortex enhancement mechanism associated with the stratiform mode increases the likelihood of genesis by enhancing mid-level vorticity, but as mentioned above genesis cannot proceed without some other mechanism to enhance low-level vorticity (the stratiform mode weakens low-level vortictiy). Of the two dominant horizontal divergence modes identified by MH, only the convective mode in a cyclonic environment can directly enhance the low-level vorticity. Thus the stratiform mode would not appear to be essential for genesis, but it may combine well with the convective mode to provide deep net convergent regions in MCSs, provided the deep convective mode dominates at low-levels. The vertical velocity profiles in the TC-LAPS updrafts are consistent with such deep net convergence, which suggests TC-LAPS may be adequately incorporating the stratiform mode at least during periods of strong convective activity.

Acknowledgements

This work was supported in part by NICOP-ONR Grant N00014-02-1-0431 as part of a collaboration with the Naval Research Laboratory. Additional support came from Grants ONR N00014-02-1-0474, NSF ATM-0101781 and NSF ATM-0132006. Recent support from an ONRIFO VSP Grant N00014-04-1-4062 funded a visit by the lead author to the USA, during which significant advancements in this work were made. We would like to thank Jeff Kepert, Peter May, John McBride, Andrea Saunders, Laurie Trenary and John Persing for helpful comments and suggestions, and we would like to acknowledge Dr. Greg Holland for some interesting and illuminating discussions on the topic of TC genesis.

References

Bister, M., and K. A. Emanuel, 1997: The genesis of Hurricane Guillermo: TEXMEX analyses and a modeling study. *Mon. Wea. Rev.*, **125**, 2662—2682.

Chen, S. S., and W. M. Frank, 1993: A numerical study of the genesis of extratropical convective mesovortices. Part I: Evolution and dynamics. *J. Atmos. Sci.*, **50**, 2401—2426.

Davidson, N. E. and K. Puri, 1992: Tropical prediction using dynamical nudging, satellite defined convective heating and a cyclone bogus. *Mon. Wea. Rev.*, **120**, 2501—2522.

ECMWF Research Department, 1995: ECMWF forecast model: physical parametrization. Fourth Edition, ECMWF, Shinfield Park, UK.

Eliassen, A., 1951: Slow thermally or frictionally controlled meridional circulation in a circular vortex. *Astrophysica Norvegica*, **5**, 18—60.

Emanuel, K. A., 1986: An air-sea interaction theory for tropical cyclones. Part I: Steady-state maintenance. *J. Atmos. Sci.*, **43**, 585—604.

Fels, S. B. and M. D. Schwartzkopf, 1975: The simplified exchange approximation: a new method for radiative transfer calculations. *J. Atmos. Sci.*, **32**, 1475—1488.

Fritsch, J. M., J. D. Murphy and J. S. Kain, 1994: Warm core vortex amplification over land. *J. Atmos. Sci.*, **51**, 1780—1807.

Gray, W. M., 1998: The formation of Tropical Cyclones. *Meteor. Atmos. Phys.* **67**, 37—69.

Harr, P. A., R. L. Elsberry and J. C. L. Chan, 1996: Transformation of a large monsoon depression to a tropical storm during TCM-93. *Mon. Wea. Rev.* **124**, 2625—2643.

Haynes, P. H. and M. E. McIntyre, 1987: On the evolution of vorticity and potential vorticity in the presence of diabatic heating and frictional or other forces. *J. Atmos. Sci.*, **44**, 828—841.

Hendricks, E. A., M. T. Montgomery and C. A. Davis, 2004: On the role of "vortical" hot towers in the formation of tropical cyclone Diana (1984). *J. Atmos. Sci.*, **61**, 1209—1232

Houze, R. A., 1997: Stratiform precipitation in regions of convection: A meteorological paradox? *Bull. Am. Meteorol. Soc.*, **78**, 2179—2196.

Houze, R. A., 2004: Mesoscale Convective Systems. *Reviews of Geophysics*, **42**, RG4003, 1—43.

Mapes, B. E. and R. A. Houze, 1992: An integrated view of the 1987 Australian monsoon and its mesoscale convective systems. I: Horizontal structure. *Quart. J. Roy. Meteor. Soc.*, **118**, 927—963.

Mapes, B. E. and R. A. Houze, 1995: Diabatic divergent profiles in Western Pacific Mesoscale Convective Systems. *J. Atmos. Sci.*, **52**, 1807—1828.

Miller, M.J. and R. P. Pearce, 1974: A three-dimensional primitive equation model of cumulonimbus convection. *Quart. J. Roy. Meteor. Soc.*, **100**, 133—154.

Montgomery, M. T., M. E. Nicholls, T. A. Cram and A. Saunders, 2005: A "vortical" hot tower route to tropical cyclogenesis. *J. Atmos. Sci.*, in press.

Montgomery, M. T. and J. Enagonio, 1998: Tropical cyclogenesis via convectively forced vortex Rossby waves in a three-dimensional quasigeostrophic model. *J. Atmos. Sci.*, **55**, 3176—3207.

Puri, K., G. S. Dietachmayer, G. A. Mills, N. E. Davidson, R. A. Bowen and L. W. Logan, 1998: The new BMRC Limited Area Prediction System, LAPS. *Aust. Meteor. Mag.*, **47**, 203-223.

Raymond, D. J., and H. Jiang, 1990: A theory for long-lived Mesoscale Convective Systems. *J. Atmos. Sci.*, **47**, 3067—3077.

Raymond, D. J., C. L′opez-Carillo, and L. L′opez-Cavazos, 1998: Case-studies of developing east Pacific easterly waves. *Quart. J. Roy. Meteor. Soc.*, **124**, 2005—2034.

Reasor, P. D., M. T. Montgomery and L. F. Bosart, 2005: Mesoscale observations of the genesis of Hurricane Dolly (1996). *J. Atmos. Sci.*, in press.

Ritchie, E. A. and G. J. Holland, 1997: Scale interactions during the formation of Typhoon Irving. *Mon. Wea. Rev.*, **125**, 1377—1396.

Ritchie, E. A., J. Simpson, W. T. Liu, J. Halverson, C. S. Velden, K. F. Brueske and H. Pierce, 2003: Present day satellite Technology for hurricane research. Chapter 12, *Hurricane: Coping with disaster*. Ed. R. Simpson, American Geophysical Union, 360 pp.

Rotunno, R. and K. A. Emanuel, 1987: An air-sea interaction theory for tropical cyclones. Part II: Evolutionary study using a nonhydrostatic axisymmetric numerical model. *J. Atmos. Sci.*, **44**, 542—561.

Saunders, A. and M. T. Montgomery, 2004: A closer look at vortical hot towers within a tropical cyclogenesis environment. Dept. of Atmospheric Science bluebook, Colorado State University.

Simpson, J., E. A. Ritchie, G. J. Holland, J. Halverson and S. Stewart, 1997: Mesoscale interactions in Tropical Cyclone Genesis. *Mon. Wea. Rev.*, **125**, 2643—2661.

Tiedke, M., 1989: A comprehensive mass flux scheme for cumulus parameterization large-scale models. *Mon. Wea. Rev.*, **117**, 1779—1800.

Tory, K. J., M. T. Montgomery, N. E. Davidson and J. D. Kepert, 2005a: Prediction and diagnosis of Tropical Cyclone formation in an NWP system. Part II: A detailed diagnosis of tropical cyclone Chris formation. Submitted to *J. Atmos. Sci.*

Tory, K. J., M. T. Montgomery and N. E. Davidson, 2005b: Prediction and diagnosis of Tropical Cyclone formation in an NWP system. Part III: Developing and non-developing storms. Submitted to *J. Atmos. Sci.*

Troen, I. B. and L. Mahrt, 1986: A simple model of the atmospheric boundary layer: sensitivity to surface evaporation. *Bound. Lay. Meteor.*, **37**, 129—148.

Zehr, R., 1992: Tropical cyclogenesis in the western North Pacific. NOAA Tech. Rep. NESDIS 61, 181 pp.

Zipser, E. J. and C. Gautier, 1978: Mesoscale events within a GATE tropical depression. *Mon. Wea. Rev.*, **106**, 789—805.

List of Figures

Figure 1: LAPS375 domain, with the TC Chris TC-LAPS domain embedded. The innermost rectangle represents the sub-domain featured in Fig. 2.

Figure 2: Vertical motion on the $\sigma = 0.25$ level (Pa/s, shaded) with horizontal wind on the $\sigma = 0.9943$ level overlaid (vectors), 4 (left), 6 (right) and 8 (lower) hours into the simulation of TC Chris initialized at 1100 UTC 1 February, 2002. The white lines indicate the positions of cross-sections presented in Fig. 3, and convective updrafts/cores referenced in the text are labeled A, B and C.

Figure 3: Vertical cross-sections of vertical motion (top, contour interval 1 Pa/s), absolute vorticity (second row, contour interval 1E-04 s^{-1}), contributions to the absolute vorticity tendency from vertical advection and stretching (third and bottom rows respectively, contour interval 2.5E-08 s^{-2}), at 4 (left), 6 (center) and 8 (right) hours into the simulation of TC Chris initialized at 1100 UTC 1 February, 2002. All zero contours have been omitted. Dashed lines represent downward flow (top), anticyclonic vorticity (second row), and anticyclonic vorticity tendency (third and bottom rows). White lines in Fig. 2 indicate the cross-section locations, and convective updrafts/cores referenced in the text are labeled A and B. Shading has been added to the lower three rows to provide an indication of the updraft location (updraft intensity shaded, $\omega > 2.5$ Pa/s). Horizontal and vertical axis scales are km and m respectively.

Figure 4: As in Fig. 3 except for three additional TC simulations: Elcho Island Storm (left), TC Fiona (center) and TC Erica (right). These simulations are presented in Part III.

Figures

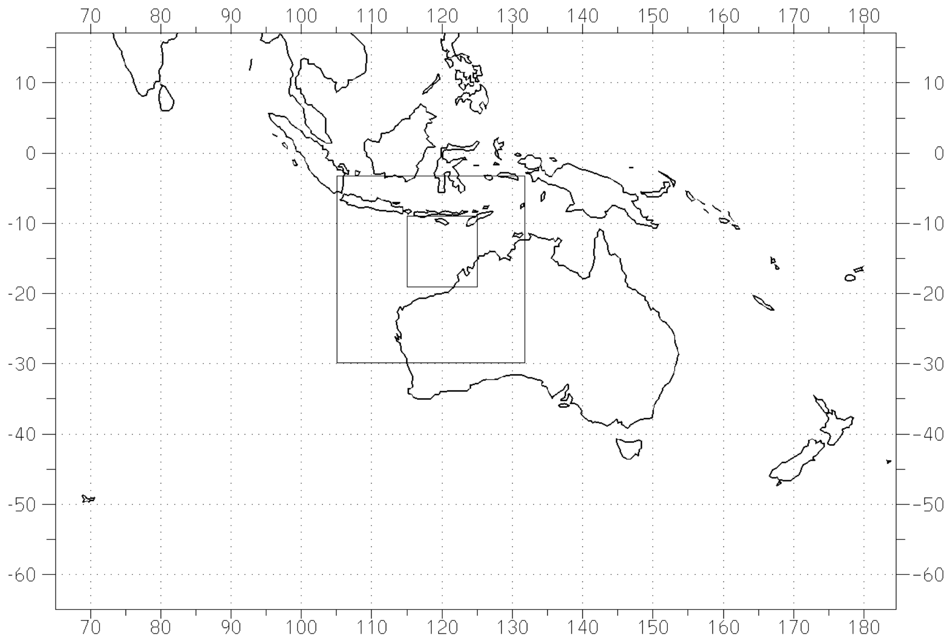

Figure 1: LAPS375 domain, with the TC Chris TC-LAPS domain embedded. The innermost rectangle represents the sub-domain featured in Fig. 2.

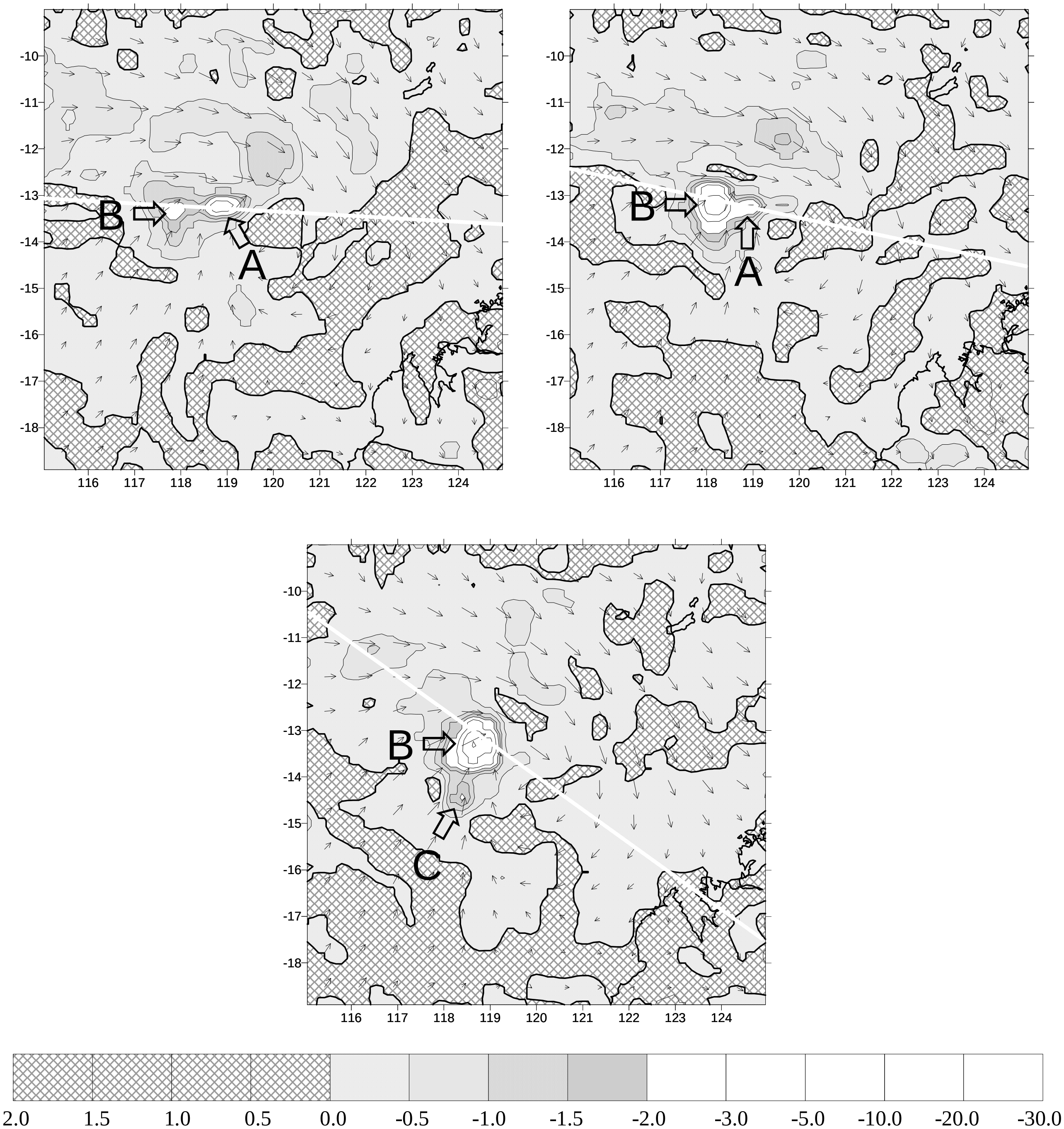

Figure 2: Vertical motion on the σ = 0.25 level (Pa/s, shaded) with horizontal wind on the σ = 0.9943 level overlaid (vectors), 4 (left), 6 (right) and 8 (lower) hours into the simulation of TC Chris initialized at 1100 UTC 1 February, 2002. The white lines indicate the positions of cross-sections presented in Fig. 3, and convective updrafts/cores referenced in the text are labeled A, B and C.

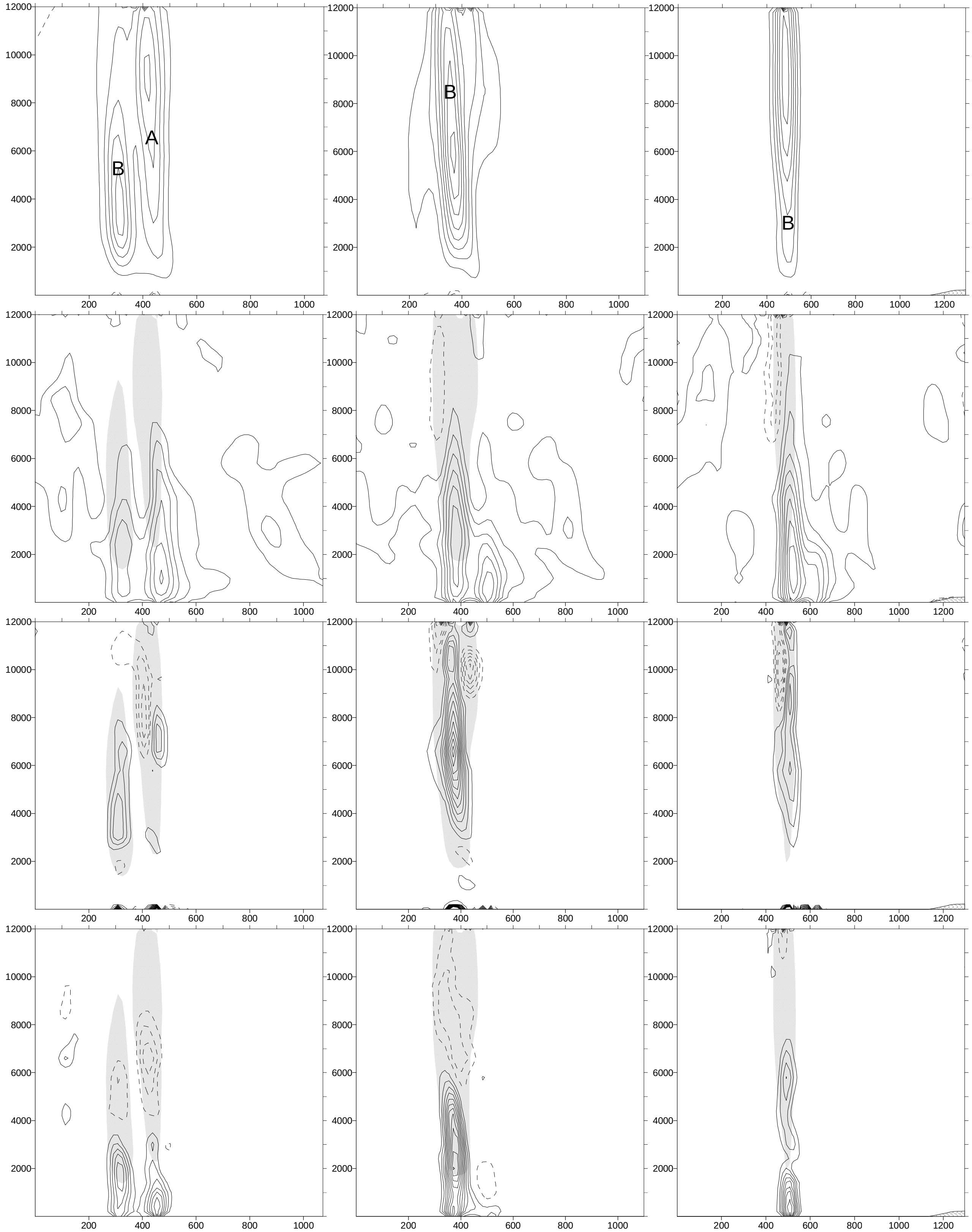

Figure 3: Vertical cross-sections of vertical motion (top, contour interval 1 Pa/s), absolute vorticity (second row, contour interval 1E-04 s^{-1}), contributions to the absolute vorticity tendency from vertical advection and stretching (third and bottom rows respectively, contour interval 2.5E-08 s^{-2}), at 4 (left), 6 (center) and 8 (right) hours into the simulation of TC Chris

initialized at 1100 UTC 1 February, 2002. All zero contours have been omitted. Dashed lines represent downward flow (top), anticyclonic vorticity (second row), and anticyclonic vorticity tendency (third and bottom rows). White lines in Fig. 2 indicate the cross-section locations, and convective updrafts/cores referenced in the text are labeled A and B. Shading has been added to the lower three rows to provide an indication of the updraft location (updraft intensity shaded, $\omega > 2.5$ Pa/s). Horizontal and vertical axis scales are km and m respectively.

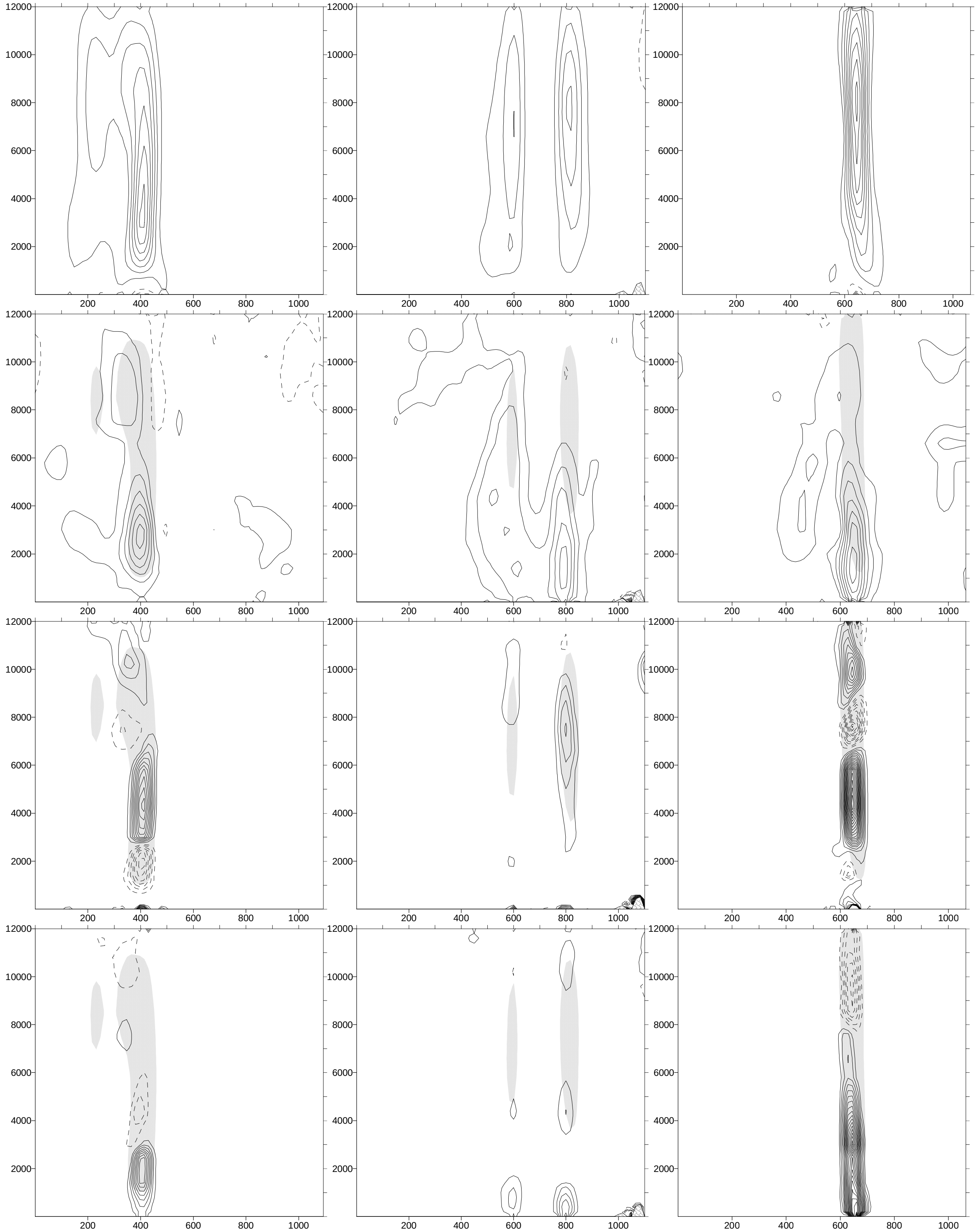

Figure 4: As in Fig. 3 except for three additional TC simulations: Elcho Island Storm (left), TC Fiona (center) and TC Erica (right). These simulations are presented in Part III.