A highly configurable vortex initialization methodology has been constructed in order to permit manipulation of the initial vortex structure in numerical models of tropical cyclones. By using distinct specifications of the flow in the boundary layer and free atmosphere, an array of parameters is available to modify the structure. A nonlinear similarity model that solves the steady-state, height-dependent equations for a neutrally stratified, axisymmetric vortex is solved for the boundary layer flow. Above the boundary layer, a steady-state, moist-neutral, hydrostatic and gradient wind balanced model is used to generate the angular momentum distribution in the free atmosphere. In addition, an unbalanced mass-conserving secondary circulation is generated through the assumption of conservation of mass and angular momentum above the boundary layer. Numerical simulations are conducted using a full-physics mesoscale model to explore the sensitivity of the vortex evolution to different prescriptions of the initial vortex. Dynamical adjustment is found to be dominant in the early evolution of the simulations, thereby masking any sensitivity to initial changes in the secondary circulation and boundary layer structure. The adjustment time can be significantly reduced by arbitrarily enhancing the moisture in the eyewall region.

1. Introduction

One aspect of the numerical prediction of tropical cyclones (TCs) that receives significant attention, because of its considerable influence on forecast accuracy, is initialization. Since TCs spend a large percentage of their lifetime in data-sparse oceanic regions, highly accurate representation of the meteorological fields in both the inner core and environment is limited. As a result, synthetic data are often used to supplement coarse analyses or observations.

Kurihara et al. (1993) argued that the synthetic vortex should possess three properties to minimize dynamic adjustment and false spinup. The first is structural consistency, the second is resemblance to the “real storm,” and the third is that the new vortex should be compatible with the numerical model. To ensure these qualities are enforced in the generation of TC flows for model initialization, three general techniques, that work together or alone, have been developed: data assimilation, dynamical initialization, and vortex bogusing.

The most widely employed data assimilation techniques are variational methods and ensemble Kalman filters, and more recently hybrids of the two. Variational methods minimize a cost function in order to optimally combine observations with a background state, with knowledge of their respective errors. This is done either at a fixed time with a balance constraint imposed [three-dimensional variational data assimilation (3D-Var)], or over a short time interval using an adjoint model [four-dimensional variational data assimilation (4D-Var; Daley 1997)]. A disadvantage of 3D-Var is that the balance constraint may be inappropriate for tropical cyclones (e.g., geostrophic balance), resulting in a physically unrealistic vortex (Hendricks et al. 2011). In this regard, 4D-Var is superior in that it is able to utilize observations distributed in time, and it is able to produce primary and secondary circulations consistent with the numerical model through the forward integration (e.g., Zou and Xiao 2000). However, its disadvantage is that it is computationally expensive and not easily portable. Ensemble Kalman filters have shown promise for tropical cyclone initialization, but they also possess shortcomings...
such as the dominance of position errors in its covariance structure, sampling errors due to the low ensemble size, the treatment of model error, and the dynamical inconsistency between the analyzed vortex and the numerical model (Torn 2010; Aksoy et al. 2013).

Dynamic initialization, introduced by Kurihara et al. (1993), entails the use of an idealized model (in their case, axisymmetric) that is integrated to generate a vortex similar to that of the real storm. Nudging, or relaxation, of the wind or pressure field is done to provide a structural match to observations. Hendricks et al. (2011) performed vortex surgery to remove the analyzed vortex and blend in an idealized vortex that was nudged to the observed surface pressure in an idealized three-dimensional full physics model, an update to the technique developed by Kurihara et al. (1993). Ideally, the model used is the same as, or similar to, the forecast model to ensure compatibility (i.e., similar model equations, grid structure, and physical parameterizations) and a vortex that is well adapted to its environment. To this end, Nguyen and Chen (2011) introduced a dynamic initialization procedure in which the forward model was used in a series of 1-h cycles to intensify the analyzed vortex to one with a minimum surface pressure and maximum wind speed close to the observed values. To produce a more realistic TC boundary layer and an unbalanced secondary circulation in the initial condition, Cha and Wang (2013) utilized a 6-h cycle with a warm start model forecast initialization. Because of the warm start, spectral nudging is applied to the large scale (wavelengths greater than 1000 km) to minimize differences in the environment between the analysis and the dynamic initialization. However, like 4D-Var, dynamic initialization is complex and computationally expensive.

The method of vortex bogusing has been applied extensively (Leslie and Holland 1995; Low-Nam and Davis 2001; Pu and Braun 2001; Kwon et al. 2002; Kwon and Cheong 2010). By using mathematical constructs to produce both the radial and vertical structure of the tangential flow, a synthetic representation of the vortex may be implanted into the analysis. Not only does bogusing require removal of the analyzed vortex (as does dynamic initialization), but the inserted vortex will not be consistent with the model formulation. The advantages of bogusing are the ability to customize the vortex structure and its very low computational cost.

We will present a simple algorithm that produces a synthetic vortex with a realistic boundary layer and free atmosphere structure in addition to an unbalanced secondary circulation. The objective of the methodology presented herein is twofold: first, to provide a benchmark for more advanced methods that have been or will be developed, and second, to enable sensitivity studies through the customizable nature of the bogusing approach. For example, we will discuss the sensitivity of the time scale of the dynamical adjustment of mass and momentum fields to the presence of a prespecified secondary circulation. The paper is organized as follows. In section 2, the components of the tropical cyclone circulation and the methodology are presented. Section 3 details numerical simulations that test the technique and the sensitivity to several key parameters. Finally, in section 4, a concluding discussion is provided.

2. Vortex generation

Construction of the synthetic vortex is a five-step procedure. First, a radial profile of the tangential wind is specified along the top of the boundary layer. Second, the tangential, radial, and vertical flow in the boundary layer are determined by the similarity model of Foster (2009, hereafter F09). Third, the tangential wind profile is determined above the boundary layer through the Emanuel (1986, hereafter E86) steady-state structure model. Fourth, the radial and vertical flow above the boundary layer is determined. Finally, the azimuthal-mean temperature and pressure distributions are determined through hydrostatic and gradient wind balance.

a. Radial structure

Currently, two possible parameterizations of the radial profile of tangential wind are employed. First, a modified Rankine (MR) vortex has been used as it permits an adjustable scaling exponent. The tangential wind of the MR vortex is given by

\[
V(r) = V_{\text{max}} \left( \frac{r}{\text{RMW}} \right)^{\alpha} \quad (r < \text{RMW}),
\]

\[
V(r) = \frac{V_{\text{max}} \text{RMW}^\alpha}{r} \quad (r > \text{RMW}),
\]

where \(\alpha\) is the inverse power-law scaling exponent and RMW and \(V_{\text{max}}\) are the radius of maximum wind (RMW) and tangential velocity at the RMW, respectively. The MR profile is used when a profile is to be fit to observed values of wind at particular radii, such as those given by real-time Automated Tropical Cyclone Forecast (ATCF) data. From Eq. (1b), \(\alpha\) can be determined by knowledge of the wind at two radii, such as the RMW and the radius of gale-force winds, so that

\[
\alpha = \frac{\log \left( \frac{V_{\text{RMW}}}{V_r} \right)}{\log \left( \frac{r_2}{r_1} \right)},
\]
Thus, the decay rate of the tangential wind outside of the RMW may be taken from observations or manipulated manually. Observations have shown that, through the variation of \( \alpha \), the MR vortex is a good approximation for real tropical cyclone tangential wind profiles of all intensities (Mallen et al. 2005).

As an alternative to the MR vortex, a piecewise continuous radial profile of tangential wind based on a statistical fit of nearly 500 observed profiles provided by Willoughby et al. (2006, hereafter WDR06) is provided as an option in the algorithm but is not utilized in this study. The WDR06 profile is given by

\[
V(r) = V_i V_{\text{max}} \left( \frac{r}{R_{\text{MW}}} \right)^n (r \leq R_1),
\]

\[
V(r) = V_i (1 - w) + V_o w (R_1 \leq r \leq R_2),
\]

\[
V(r) = V_o V_{\text{max}} \left[ (1 - A) \exp \left( -\frac{r - R_{\text{MW}}}{X_1} \right) + A \exp \left( -\frac{r - R_{\text{MW}}}{X_2} \right) \right] (R_2 \leq r),
\]

where \( w \) is a ninth-order polynomial ramping function; \( R_1 \) and \( R_2 \) define a transition zone straddling the RMW; \( X_1 \) and \( X_2 \) are the decay lengths of the near and far wind profiles outside of the transition zone, respectively (see WDR06’s Figs. 1 and 2 for additional detail); and \( A \) is a weighting parameter for the significance of the two decay profiles. The parameter \( n \) is the scaling exponent for the power law. With specification of the RMW, \( V_{\text{max}} \), and the latitude, these parameters are determined through a statistical fit as noted above.

The profiles of both vortices are shown in Fig. 1 with a \( V_{\text{max}} \) of 40 m s\(^{-1} \), RMW of 40 km, and \( \alpha \) of 0.5 for the MR profile. In the inner core (Fig. 1a), the MR vortex has a discontinuity in the radial derivative at the RMW while the WDR06 vortex, which utilizes a transition zone, displays a smooth transition about the RMW. At the outer radii (Fig. 1b), the WDR06 vortex is seen to decay more rapidly with radius, though the decay rate of both vortices can be controlled through specification of either \( \alpha \) for the MR vortex or the latitude, through its control of the parameter \( A \) in Eq. (3c), for the Willoughby vortex. An additional shaping tool has been included to modify the outer wind decay rate:

\[
V_{\text{final}}(r) = V(r) \exp \left( -\frac{r}{r_{\text{cut}}} \right)^\gamma,
\]

where \( r_{\text{cut}} \) is the radius where the cutoff function transitions from 1 to 0 and \( \gamma \) is the decay exponent. The thin lines in Fig. 1b shows the same vortex profile as given by

![Fig. 1. Tangential wind radial profiles with a \( V_{\text{max}} \) of 40 m s\(^{-1} \) and RMW of 40 km: (a) near-eyewall Willoughby profile (solid) and modified Rankine profile (dash) with an \( \alpha \) of 0.5; (b) full Willoughby profile (solid), Willoughby profile with forced decay (thin solid), modified Rankine profile (dash), and modified Rankine profile with forced decay (thin dashed); (c) inertial stability (s\(^{-1} \)) profile for the vortices in (a).](image-url)
the thicker line but with an $r_{\text{cut}}$ value of 600 km and $\gamma = 6$. Use of $r_{\text{cut}}$ proves particularly useful when the tangential wind falls off slowly with radius as in the MR vortex.

b. Vertical structure

Many previous studies—Wang (1998), Jones (2004), and Kwon and Cheong (2010) just to name a few—have created various mathematical constructs to represent the vertical structure of idealized vortices. One option for vertical structure is to use Gaussian decay functions both above and below the boundary layer top (Nolan 2007). Such a vertical structure takes the following form:

$$V(r, z) = V(r) \exp\left(-\frac{|z - z_{\text{max}}|^\alpha}{aL^\alpha}\right), \quad (5)$$

where $z_{\text{max}}$ is the height of the boundary layer top and $\alpha$ and $L$ are the length scale of the vertical Gaussian profile and decay rate, respectively. The variables $\alpha$ and $L$ may be specified as different values above and below the boundary layer to further customize the vertical structure. The Gaussian decay option is highly simplistic; it would be preferable to define a flow consistent with the governing system of equations for both the boundary layer and the free atmosphere of a tropical cyclone.

1) BOUNDARY LAYER

To yield a realistic axisymmetric flow in the boundary layer, we utilize the model developed by F09. The F09 similarity model solves the steady-state, height-dependent equations for a neutrally stratified, axisymmetric flow in the boundary layer. The stress on the lower boundary is computed from a bulk drag law formulation. With the assumption of a vortex in gradient wind balance and a constant pressure gradient through the depth of the boundary layer, pressure is removed as a variable. For simplicity, we use a constant vertical eddy diffusivity $K$ in the diffusion term of the momentum equations.

When shaping a tangential wind profile to observations for a customizable tangential velocity profile, use of the MR vortex is necessary. However, the MR vortex provides a complication for the F09 model in the form of a discontinuity in the radial derivative of the tangential wind at the RMW. The boundary layer model equations are nondimensionalized using a depth scale inversely proportional to the inertial stability (Eliassen 1971). Thus, with a discontinuity in the inertial stability (Fig. 1c, dashed line), the boundary layer solution is at best unrealistic, and at worst, the F09 solution fails to converge. To alleviate the discontinuity issue, attempts were made to patch the MR vortex with fifth- or seventh-order Hermitian interpolating polynomials with first and second derivative matching conditions at the endpoints. Fluctuations in the boundary layer flow persisted. Such fluctuations are also observed in the Willoughby vortex, which uses a ninth-order polynomial straddling the RMW, due to an inertial stability profile that is non-monotonic (Fig. 1c, thick solid line). The most consistent solution found was to apply 10 iterations of a moving average with a 5 point window to the entire profile (Fig. 1c, thin solid line). If we chose to apply the smoothing to a subset of the profile centered on the RMW, an inertial stability similar to that of the Willoughby profile would result.

An example of the axisymmetric flow for a constant eddy diffusivity is shown in Fig. 2. A large RMW is used to aid in visualization. The boundary layer flow displayed is similar to that seen in observations (Bell and Montgomery 2008; Zhang et al. 2011) and numerical simulations (Nolan et al. 2009). The supergradient jet in the tangential velocity field is located inside the peak inflow in the region of strong vertical shear of the radial flow. Near the surface, the radial inflow overshoots the RMW, rises and flares outward where it forms the eye-wall updraft near the top of the boundary layer. Vertical motion is determined by the vertical integration of the continuity equation.

To gauge how the flow characteristics change with the magnitude of the eddy diffusivity, vertical profiles of the wind fields taken at the RMW at the boundary layer top, set at 2 km throughout this study, are shown in Fig. 3. The values of $K$ chosen span the range of typical eddy diffusivities and are similar to the values chosen in F09. The near-surface vertical shear is largest for the smallest values of the eddy diffusion coefficient $K$. In association with the strongest supergradient flow is the strongest low-level inflow that carries angular momentum inward past the radial position where the boundary layer wind speed matches the gradient wind speed, where the radial flow transitions from accelerating to decelerating. As $K$ increases, the degree of vertical smoothing of the radial and tangential winds increases. Such smoothing is evident in the weakening and raising of the supergradient jet. Furthermore, the weaker but deeper radial inflow for increasing $K$ yields a larger vertical velocity near the top of the boundary layer.

As the F09 boundary layer model requires the specification of the wind profile at the boundary layer top and operational estimates provide information at the surface, it is necessary to develop a relationship between the surface and boundary layer top wind profiles. More specifically, we need to relate the parameters that define the wind profile. For the MR vortex, for example, the boundary layer model was run, at a specified constant
FIG. 2. Axisymmetric wind field produced by the boundary layer model (F09) using an eddy diffusivity of 55 m$^2$ s$^{-1}$: (a) tangential wind (m s$^{-1}$), (b) radial wind (m s$^{-1}$), and (c) vertical wind (cm s$^{-1}$).

FIG. 3. Vertical wind profiles at the radius of maximum wind produced by the boundary layer model (F09) for different values of eddy diffusivity: (a) tangential wind (m s$^{-1}$), (b) radial wind (m s$^{-1}$), and (c) vertical wind (cm s$^{-1}$).
eddy diffusivity, for a range of surface values of tangential velocity \( V \), radius \( r \), and decay exponent \( \alpha \) [Eq. (2)]. Multiple linear regression was used to determine each of the MR parameters along the boundary layer top as a function of all three parameters at the surface. That is,

\[
V_{\text{max}_\text{btl}} = F(V_{\text{max}_\text{sfc}}, \text{RMW}_{\text{sfc}}, \alpha_{\text{sfc}}),
\]

\[
\text{RMW}_{\text{btl}} = G(V_{\text{max}_\text{sfc}}, \text{RMW}_{\text{sfc}}, \alpha_{\text{sfc}}),
\]

\[
\alpha_{\text{btl}} = H(V_{\text{max}_\text{sfc}}, \text{RMW}_{\text{sfc}}, \alpha_{\text{sfc}}),
\]

where the subscripts btl and sfc represent the boundary layer top and surface, respectively; and \( F \), \( G \), and \( H \) are the functional relationships derived from the regression. Note that the same moving mean applied to the general algorithm is also applied in running of the boundary layer model for the regression analysis. As a result, the boundary layer top vortex will be larger in amplitude and significantly smaller in size (before it is smoothed). These new boundary layer top values are then used to run the algorithm to produce the synthetic vortex. The \( R^2 \) values for the regression are 0.99, 0.98, and 0.97 for the RMW, \( V_{\text{max}} \), and \( \alpha \), respectively. Suppose two observations, one of which corresponds to the RMW, are found to produce a MR profile with \( V_{\text{max}} = 40 \text{ m s}^{-1}, \text{RMW} = 50 \text{ km}, \) and \( \alpha = 0.4 \) (Fig. 4, solid line). Equations (6a)–(6c) yield boundary layer top parameters of \( V_{\text{max}} = 57 \text{ m s}^{-1}, \text{RMW} = 14 \text{ km}, \) and \( \alpha = 0.47 \) (Fig. 4, solid line with circles). As discussed above, the wind profile needs to be smoothed in order to eliminate the discontinuity in the inertial stability profile. The smoothed boundary layer top profile, which is used to drive the F09 model that has been rescaled to the unsmoothed maximum wind, is given by the solid line with pluses. Finally, the resultant surface wind from the boundary layer model, designed to match the observational profile, is given by the dashed line. Inspection of Fig. 2 reveals that the F09 boundary layer model produces different wind profiles at the surface and at the boundary layer top. These different wind structures are apparent in Fig. 4 as given by the dashed and solid plus lines. The surface wind, not surprisingly, is 25% weaker. In addition, the surface wind appears to be slightly more peaked and decays at a slower rate with radius than the boundary layer top profile.

2) FREE ATMOSPHERE

Above the boundary layer, the vertical structure of the axisymmetric tangential wind is given by the steady-state, moist-neutral, hydrostatic, and gradient wind balanced model of E86. The advantage of the Emanuel structure is the incorporation of a radial variation of the RMW with height. Given the radial distribution of angular momentum at the top of the boundary layer, the imposition of moist neutrality allows one to find the altitude of angular momentum surfaces through conservation of saturated moist static energy. The tangential wind profile can then be obtained by inversion of the angular momentum relationship. Stern and Nolan (2011) found, in observations, theory, and numerical simulations, that the normalized decay of the maximum tangential wind with height was independent of the size and intensity of the tropical cyclone. Thus, the authors concluded that a single level would be sufficient to determine the vertical profile of maximum wind speed above the boundary layer. Here, we use a simplified version of the E86 wind field as implemented by Moon and Nolan (2010). They used the E86 theory to define only the radius of maximum winds at each height, \( \text{RMW} = \text{RMW}(z) \), and the maximum tangential wind \( V_{\text{max}} = V_{\text{max}}(z) \). Then at each height the same tangential wind profile as prescribed at the top of the boundary layer was applied using the derived values of RMW and \( V_{\text{max}} \). This constructs a complete wind field that has a very similar outward slope and vertical decay as the full E86 solution. The mass field is diagnosed from gradient wind and hydrostatic balance using the iterative scheme of Nolan.
et al. (2001). The outflow temperature, which controls the depth of the circulation in the free atmosphere (the E86 model uses temperature as a vertical coordinate), can be user specified or taken from an analysis as a quasi-horizontal mean at the level representing the tropopause, which will be discussed in the following section.

3) SECONDARY CIRCULATION AND MOISTURE ENHANCEMENT

To match the flow between the boundary layer and the free troposphere, conservation of angular momentum $M$ above the boundary layer is exploited. A mass-conserving scalar streamfunction $\psi(r)$ can be used to define the free troposphere secondary circulation. First, the radial-height angular momentum distribution is calculated from the axisymmetric tangential wind field. Second, the streamfunction along the boundary layer top is determined by the inward integration of the vertical motion field at the upper boundary of the F09 model, thus providing $\psi(M)$. Third, the radial-height distribution of $\psi$ is determined by maintaining a constant $\psi$ along $M$ surfaces. Finally, the secondary circulation is derived from $\psi$.

Figure 5 displays the total tropospheric flow that matches with the boundary layer flow shown in Fig. 2. Note the sloping RMW with height in Fig. 5a.

Once the mass and momentum fields are determined on the axisymmetric grid, they are expanded to a three-dimensional grid with Arakawa C staggering. Asymmetry in the full three-dimensional fields is recovered when the axisymmetric fields are added to the analysis or background state. Only the vertical motion remains axisymmetric. Such symmetry is not long lived because of the grid geometry and vortex tilt from the influence of large scale vertical shear during dynamic adjustment.

Finally, the option is available to use either the original moisture field from the analysis or to provide a moisture enhancement to accelerate development, both of which are examined in this study. Currently, the moisture enhancement is axisymmetric with a radial profile that matches that of the tangential wind (i.e., an annulus)

$$q = q' + q' \left[ \frac{V(r, z)}{V_{\text{max}}} \right],$$

where $q'$ is a user-defined enhancement parameter. As such, the moisture perturbation is large where the winds are large and deep convection develops from frictional convergence. Other techniques can easily be incorporated into the methodology. For example, suppose an idealized distribution of equivalent potential temperature $\theta_e$ is desired. Then an iterative scheme for the pressure, temperature, and specific humidity can be implemented. Pressure would be derived from gradient

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Fig. 5. Full axisymmetric wind field produced by matching the boundary layer model (F09) with the free-atmosphere model (E86) as described in the text: (a) tangential wind (m s$^{-1}$), (b) radial wind (m s$^{-1}$), and (c) vertical wind (cm s$^{-1}$).
wind balance, temperature from hydrostatic balance, and specific humidity from the definition of $\theta_e$.

3. Results

a. Vortex removal and asymmetries

All simulations performed use the Weather Research and Forecasting Model (WRF) mesoscale numerical weather prediction system version 3.1.1 (Skamarock et al. 2008) with nested grids of 27-, 9-, and 3-km resolution and $240 \times 180$, $180 \times 180$, and $240 \times 240$ grid points, respectively. Physical parameterizations include the Yonsei University (YSU) boundary layer (Noh et al. 2003) with modifications to the drag formulation following Donelan et al. (2004). The WRF single-moment 6-class microphysics scheme (Hong and Lim 2006) is used for microphysical processes. Radiation physics is handled by the Rapid Radiative Transfer Model (RRTM) for longwave radiation (Mlawer et al. 1997; Iacono et al. 2000) and the Goddard scheme (Chou et al. 1998) for shortwave radiation. Finally, the Kain–Fritsch convective parameterization (Kain 2004) is activated on the outer grid.

The objective of this study is the presentation of a new synthetic vortex that may be used for numerical modeling or forecasting of TCs. Thus, vortex removal is not of primary concern, but a necessary step to conduct the synthetic vortex experiments. Vortex removal follows the method used in the GFDL Hurricane Prediction System and, therefore, will not be discussed at length. Interested readers should refer to Kurihara et al. (1993, 1995) for details. After the global analysis has been subsetted to the domain of interest, the initial field is separated into basic and disturbance fields. The separation occurs through filtering, either using the smoothing function described in Kwon et al. (2002), an updated version of the Kurihara et al. (1993, 1995) smoothing functions, or through a low-pass transform with a specified wavenumber cutoff (not utilized in the current study). Fields that are filtered include momentum, geopotential, potential temperature, and the dry weight of the grid column, termed $\mu$ in WRF. The nonhurricane component of the disturbance field is generated by removal of the hurricane circulation by applying optimal interpolation inward from a calculated radius, determined by values of the tangential wind and its derivative at each azimuth, as is done in the Geophysical Fluid Dynamics Laboratory (GFDL) model. Finally, the nonhurricane component of the disturbance is added back to the basic state. Vortex removal is executed at all levels in the vertical in which the level-mean potential vorticity (PV) is smaller than $1.5 \times 10^{-9}$ km$^2$ kg$^{-1}$ s$^{-1}$, a value taken to represent the dynamic tropopause. In intense storms, this value should be increased to ensure complete vortex removal. Results of the technique are shown in Fig. 6, which shows the perturbation hydrostatic pressure of the column in colorfill and boundary layer top wind vectors for Hurricane Lili on 0000 UTC 3 October 2002. The basic state, in which the analyzed vortex has been removed, is shown in Fig. 6b. Away from the cyclone, it is evident that there is little change in the environmental mass and momentum fields between the original Global Forecast System (GFS) analysis (Fig. 6a) and that of the initial condition produced by the technique described in this study (Fig. 6c).

To investigate the loss of asymmetry in the cyclonic momentum field, consider the insertion of an idealized balanced vortex for the case of Hurricane Lili at 0000 UTC 3 October 2002. To assess the degree of asymmetry, The National Oceanic and Atmospheric Administration (NOAA) Hurricane Research Division H*wind product (Powell et al. 1998) is displayed in Fig. 7a. With storm motion to the west-northwest, a translational right-of-motion asymmetry is observed. According to the Statistical Hurricane Intensity Prediction Scheme (SHIPS) dataset (DeMaria and Kaplan 1994), westerly shear increased from roughly 4.5 to 7.5 m s$^{-1}$ in the 6 h preceding 0000 UTC 3 October 2002. Thus, gustiness from downshear and downshear-left convection may also contribute to the observed asymmetry. Figure 7b shows the coarse GFS analysis depiction of Lili with a similar asymmetric structure. As can be seen in Figs. 7c and 7d, utilizing a 27- and 9-km resolution grid, respectively, the addition of the axisymmetric vortex to the base-state field also yields a reasonable low-level structure.

b. Numerical simulations

The nature simulation utilized for this study is that of Atlantic Hurricane Bill (2009) from 0000 UTC 16 August to 0000 UTC 23 August 2009, using GFS initial and boundary conditions from GFS analyses. It is important to emphasize that the nature simulation is just a proxy for the real atmosphere and that all experiments will be compared to the nature case, not the actual evolution of Hurricane Bill. Bill provided an excellent case as it did not experience a significant land interaction. As can be seen in the pressure and maximum 10-m wind traces (Fig. 8a), the simulation captures many aspects of the tropical cyclone life cycle. Figure 8b displays the track for the nature simulation (pluses). Only small deviations in track from the nature simulation are observed in all synthetic vortex simulations, an example of which is given by the dots in Fig. 8b. Therefore, further consideration of track deviations will not be considered.
To test the ability of the synthetic vortex to reproduce the intensity evolution of the nature simulation, the azimuthally averaged tangential wind field was calculated at hour 78 of the nature simulation, in the middle of the intensification phase (Fig. 8). The center of circulation was determined by calculating the PV centroid (Jones 2004) at 1-km altitude in a $1620 \times 1620$ km$^2$ box centered on the minimum in surface pressure. The 10-m azimuthally averaged tangential wind at the RMW and the radius of gale-force winds were then determined so that the RMW, $V_{max}$, and $a$ could be determined at the boundary layer top by multiple linear regression [Eq. (6)]. Use of the azimuthally averaged maximum wind as a peak wind is a subjective choice, but is a more reasonable approximation than the three-dimensional peak wind.

Six synthetic vortex simulations were conducted to examine the role of varying the eddy diffusivity constant in the boundary layer model, the influence of an initial unbalanced secondary circulation, and the role of the initial moisture enhancement. The benchmark test simulation, hereafter CONTROL, uses $K = 55$ m$^2$ s$^{-1}$, while experiments K25 and K100, use $K = 25$ and 100 m$^2$ s$^{-1}$, respectively. Simulation NOSEC removes the secondary circulation for the initial condition and therefore represents a true “cold start.” For the above simulations, the moisture field is taken directly from the nature simulation at the time of initialization. Simulation MOISTPLUS adds a moisture enhancement to the initial water vapor field. The moisture enhancement has a maximum value of 10% (at the RMW). It is important to note that the initializations performed here remove all condensate from the nature simulation as WRF cannot cold start with condensate. All experiments use the same specification of boundary layer top and outflow level temperatures used for the E86 model representation of the free atmosphere vortex. The boundary layer top temperature was taken from the base-state analysis in the same fashion as the outflow temperature except the model level corresponds to that of the boundary layer top. Recall that the altitude of the angular momentum surfaces and, through moist neutrality, the moist isentropes, are determined through the constancy of moist static energy along these surfaces. Therefore, temperature along the boundary layer top is needed to derive the boundary layer top moist static energy distribution. In addition, all simulations utilize Eq. (4), with $r_{cut} = 600$ km, to force the wind profile toward zero at large radius. Figure 9 displays the azimuthally averaged tangential wind profiles at the top of the boundary layer and at the lowest model level (near $z = 100$ m) for the nature run (NATURE) and simulations CONTROL, K25, and K100 taken from the WRF initialization files.

Fig. 6. Perturbation hydrostatic pressure (colorfill; hPa) and boundary layer top wind vectors for Hurricane Lili at 0000 UTC 3 Oct 2002: (a) analysis, (b) analysis without vortex, and (c) analysis with synthetic vortex.
The 10-m wind is a diagnostic variable in WRF and, therefore, not available at the initial time. As the eddy diffusion coefficient increases and the vertical profiles are smoothed, a smaller magnitude boundary layer top wind is needed to produce a given surface wind.

The results of all simulations, using the maximum azimuthally averaged tangential wind at the lowest model level as a measure of intensity, as shown in Fig. 10. It is immediately evident that all test simulations, except MOISTPLUS, experience a 12-h dynamic adjustment. After this period, the rapid intensification of the test simulations are larger than that of the nature simulation so that roughly one day into the integration, all simulations have a similar intensity. After this time, the test simulations maintain a slightly higher equilibrium intensity.

The initial adjustment period is one of vortex spin-down, the result of surface friction. Just as in an intensifying TC where the tangential flow strengthens to balance the developing warm core, the primary circulation responds to changes in the secondary circulation during adjustment. Even though an unbalanced secondary circulation has been introduced in the initialization, Ekman pumping dominates and the deep secondary circulation decays at the expense of radial outflow at the top of the boundary layer, as seen in Fig. 11a, 3 h into the integration. Not surprisingly, the most significant changes in the primary circulation are centered around the boundary layer top where the radial outflow is weakening the tangential circulation (Fig. 11b; note the change in height scale between the two figures). At 10 h (Figs. 11c,d), the synthetic upper-level outflow and the boundary

![Figure 7](https://example.com/fig7.png)

**Fig. 7.** Cyclonic wind (m s\(^{-1}\))—Hurricane Lili: (a) HRD surface H*wind at 0130 UTC 3 Oct 2002; (b),(c),(d) first model level (approximately 100 m) at 0000 UTC 3 Oct 2002 for 1° FNL analysis interpolated to 27-km grid, bogus 27-km resolution, and bogus 9-km resolution, respectively.
layer tangential flow are at their weakest. However, deep frictionally driven convection signals the onset of intensification. By 20 h (Figs. 11e,f), the initial synthetic circulation has been replaced by a strong, model-consistent circulation that is undergoing intensification.

One possible explanation for the observed adjustment is the loss of information with respect to convection. While mass continuity is satisfied in the initialization, the strength of the secondary circulation may be significantly weaker than that of the nature simulation at the time of initialization. Nascimento and Droegemeier (2006), in a three-dimensional study of a bow echo, explored dynamic adjustment in convective simulations through resetting of either the horizontal or vertical wind components several hours into the model integration. It was found that the horizontal convergence is of central importance for forcing the correct vertical motion. Resetting of the vertical motion field produced

![Figure 8](image1.png)

**FIG. 8.** (a) Minimum surface pressure (hPa; solid) and maximum 10-m tangential wind (m s\(^{-1}\); dashed) of the nature simulation. Open dots display the times at which the bogus algorithm is applied. (b) Track of TC from nature simulation (pluses) and simulation CONTROL (circles).

![Figure 9](image2.png)

**FIG. 9.** Azimuthally averaged tangential wind (m s\(^{-1}\)) profile taken from the initialization of simulations discussed in text at (a) boundary layer top and (b) lowest model level (approximately 100 m).
little difference from the control simulation. Figure 12 shows the azimuthally averaged radial and vertical motion for CONTROL subtracted from NATURE at the initial time. With a peak radial flow just under 15 m s$^{-1}$ for the nature simulation, the secondary circulation for CONTROL is roughly 40% that of the nature case. To some extent, smoothing of the tangential wind profile for the boundary layer model is responsible for this deficit. The vertical motion field in Fig. 12b is largely a reproduction of the azimuthally averaged vertical motion of the nature simulation as it is an order of magnitude larger than that produced in the synthetic vortex initialization. The convective evolution and adjustment dynamics of an intensifying TC are different from that of a mature bow echo, so underrepresented vertical motion and the absence of condensate may be of significance. Testing such influences remains for future work.

Simulation MOISTPLUS has a significantly shorter adjustment phase as noted above. The addition of the moisture enhancement results in saturation of the mesoscale core, leading to stronger, more sustained convection. In addition, the near-zero stratification in a saturated convecting environment results in a deeper secondary circulation, reducing outflow at the boundary layer top.

As Ekman pumping dominates the early evolution in simulations without the moisture enhancement, the variations in eddy diffusivity and the presence of an unbalanced flow are less significant. Therefore, the experiments conducted above are repeated with a 15% moisture enhancement added in each case. While the 10% moisture enhancement saturated the inner core, the areal extent of saturation is increased in the 15% moisture enhancement simulations. Tangential wind traces similar to that of Fig. 10 are reproduced for the moisture enhancement simulations in Fig. 13. In all cases, the addition of the moisture enhancement has significantly decreased the adjustment time. In fact, the rapid development of model convection mitigates the need to initialize with a divergent flow. During the first few hours of the simulation, the presence of the initialized secondary circulation produces a noticeable increase in convective gustiness over simulation NOSEC. By 6 h, the onset of rapid intensification, the model-developed secondary circulation is evident (not shown), which reflects the problems associated with using simplified (nonmodel) equations to specify the initial vortex flow and secondary circulation. For the simulations in which the eddy diffusivity is varied, only the smallest value of K used, 25 m$^2$s$^{-1}$, produces a variation from the other cases. It is reasonable that a larger ensemble of simulations with the same K could yield members with adjustment times similar to that of the others. The large vertical shear of the smallest K case (Fig. 3) yields a turbulent boundary layer that mixes larger tangential wind downward, possibly producing the spike in the wind traces of Figs. 10 and 13.

It is interesting to compare the results found utilizing the F09 model to those of a vortex whose boundary layer was computed without the radial advection, residual curvature, and eddy diffusivity terms (i.e., gradient wind balance). As a result of the structure of the F09 vortex, we have two choices for the gradient wind profile that is constant with height: a profile that matches the wind speed at the surface or one that matches the wind speed at the top of the boundary layer. We have tested both, and the vertical profiles of the azimuthally averaged wind at the RMW, along with the F09 profile (NOSEC) at the RMW, are shown in Fig. 14a. Recall that the 10-m wind is a diagnostic and not included in the initial condition. It is included here to aid in visualization. The resultant intensity evolution using the lowest model level azimuthally averaged tangential wind is shown in Fig. 14b. Using the tangential wind profile at the boundary layer top produces an evolution similar to the F09 model simulation as the wind profile matches that of the F09 model above the boundary layer and overestimates the magnitude of the wind near the surface, leading to a slightly stronger intensity early in the evolution. The simulation using the surface-based wind profile intensifies at the same rate as the other simulations but takes longer to reach steady state as the winds are significantly weaker above the surface than the other.

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**Fig. 10.** Azimuthally averaged maximum tangential wind (m s$^{-1}$) at lowest model level (approximately 100 m) of simulations discussed in text. The time axis represents hours of the nature simulation. All synthetic vortex initializations were conducted at hour 78.
FIG. 11. (a),(c),(e) Azimuthally averaged, full tropospheric radial (grayscale; m s$^{-1}$) and vertical motion at 3, 10, and 20 h of simulation CONTROL. Vertical motion contours are 0.1, 0.25, and 0.45 m s$^{-1}$ at 3 h and 0.3, 0.6, and 0.9 m s$^{-1}$ at 10 and 20 h. (b),(d),(f) Azimuthally averaged tangential wind (m s$^{-1}$) in the boundary layer at 3, 10, and 20 h of simulation CONTROL.
cases. In both gradient wind simulations there is a precipitous dropoff in the strength of the vortex despite the moisture enhancement. While the behavior is not observed in the global maximum wind because of convective gustiness, the vortex weakening represents the adjustment process as described above. One possible explanation for the significant vortex weakening compared to simulation NOSEC, as noted above, is the absence of downward vertical mixing of strong winds from the supergradient jet in NOSEC.

Two additional times were selected to test the skill of the vortex initialization scheme: at hour 24 during the slow intensification or preintensification phase and at 108 h at the steady state (Fig. 8). In the first case, 24 h was chosen so that the nature run had undergone a complete adjustment from its cold start. For the early stage case, sensitivity to the secondary circulation was not considered, and sensitivity to eddy diffusion was not considered for either case. CONTROL and MOISTPLUS \([q^0 = 0.1\) in Eq. (7)] experiments with similar initializations to that employed in the intensification phase experiments are conducted for simulations initialized at \(t = 25\) h and \(t = 108\) h. In addition, a nondivergent initial condition is examined for the steady-state simulations.

The lowest model level tangential wind trace for the preintensification stage simulations is shown in Fig. 15. Because of the weak intensity, the radius of half the value of gale-force winds (245 km) is used to define the vortex structure of the nature simulation. As in the intensification stage simulations, the synthetic vortices display an initial weakening period due to dynamic adjustment. The adjustment phase is partially obscured in the intensification rate by the convective gustiness at this early stage of evolution. It is difficult to discern the influence, if any, of the moisture enhancement due to the aforementioned convective gustiness. However, the enhanced moisture simulation does undergo a slightly
more rapid intensification than the control simulation for the first 20 h after adjustment. As just a single realization of each initial condition is evaluated, it is difficult to attach meaning to such rate differences in comparison to natural variability. In addition, with a longer integration period prior to the steady state compared to the intensification stage simulations, the equilibrium intensities of the preintensification stage simulations are more consistent with the nature simulation than those of the intensification stage simulations.

Results for the steady-state simulations are shown in Fig. 16. As in the intensification phase simulations, the specification of a divergent initial condition does not influence the intensity evolution relative to the other...
synthetic vortex simulations. The enhanced moisture simulation intensifies to the steady state more rapidly than the other test cases for this phase. However, the prior discussion related to variability is equally valid here. Simulation CONTROL is already saturated on the mesoscale (not shown) and, therefore, the moisture enhancement acts to increase the geometrical size of the saturated region, which has little influence at this stage of evolution. The weakening that occurs during the adjustment phase is most evident for this set of simulations. Intensification of the synthetic vortex cases, where at the steady state none should be observed, is a result of the weakening during adjustment and grid interpolation differences. All intensity measures utilize data from the high-resolution grid while initialization occurs on the coarse grid. Although the nature and CONTROL vortices are similar on the coarse grid (Fig. 17), interpolation of the bogus to the finest grid yields a vortex that is slightly weaker than nature, at this vertical level, at this time.

While the temporal evolution of the intensity metric shows little variation in the synthetic vortex simulations, the structural evolution does display a sensitivity to the initial condition. Such sensitivity has recently been explored in the numerical simulations of Xu and Wang (2010) and Stern and Nolan (2011). Recall that the synthetic wind profile is generated from just two data points and as a result may significantly depart from “truth.” For example, if the observed wind deviates from the MR power law formulation inside the RMW, as is the case for the preintensification simulation (Figs. 18a,b), the synthetic vortex may bear little resemblance to the actual vortex. Another possible source of departure from the observed structure is the outer wind profile, as was the case for the intensification state synthetic vortex simulations, where, despite the use of a cutoff radius [Eq. (4)] larger angular momentum at large radius produced a larger vortex during the course of the integration (not shown).

Figure 18 shows radius–height plots of the azimuthally averaged tangential wind for both the nature simulation and the preintensification stage moisture enhanced simulation after the adjustment phase. Note that the hours given in the figures denote time after initialization of the synthetic vortex, which in this case is 24 h into the nature run. During the early hours of the integration of MOISTPLUS the winds in the boundary layer weaken as a result of adjustment as was discussed previously with regard to Figs. 12a and 12b. At 24 h into simulation MOISTPLUS (Fig. 18d), intensification is well under way as the diffuse boundary layer wind maximum has increased in magnitude and decreased in radius. The RMW in the nature simulation is nearly twice that of MOISTPLUS, which has surpassed the peak azimuthally averaged wind of the nature simulation. The vortex in simulation NATURE has contracted significantly by 48 h, but the peak azimuthally averaged wind remains weaker than that of the MOISTPLUS simulation. At 72 h, both simulations have a similar RMW and nearly equal intensities. At this time, the nature simulation has developed a vertical gradient of the lower and midtropospheric azimuthally averaged tangential wind outside of the RMW (Fig. 18g). Despite some subtle differences, there is good similarity between the vortices at this time.
FIG. 18. (a),(c),(e),(g) Radius–height plots of the azimuthally averaged tangential wind for the NATURE simulation at 0, 24, 48, and 72 h. (b),(d),(f),(h) Radius–height plots of the azimuthally averaged tangential wind for the MOISTPLUS simulation at 0, 24, 48, and 72 h.
4. Conclusions

Tropical cyclone initialization continues to be a challenge in numerical weather prediction, since the predicted storm evolution is often sensitive to the initial vortex structure. A new method for initializing a synthetic or “bogus” vortex has been introduced. The vortex initialization produces realistic tropical cyclone fields that are physically consistent without the need to utilize data assimilation.

The methodology is separated into two components, each having a number of tunable parameters to customize the vortex structure. First, the radial structure is specified by choosing a radial profile of tangential winds at the top of the boundary layer. Currently, the modified Rankine and the WDR06 statistical fit profiles are options provided in the algorithm. Additional profiles may be easily incorporated. The second component of the algorithm is specification of the vertical structure of the synthetic vortex. Vertical structure is decomposed into boundary layer and free atmospheric components, both of which utilize equation sets consistent with the physics of the region. Implementation of the boundary layer flow uses the axisymmetric similarity model of F09, which is a generalization of the Ekman layer solution with the inclusion of the nonlinear advection, curvature, and eddy diffusion terms to the equations of motion. The free-atmosphere flow was derived from the steady-state, moist-neutral, hydrostatic, and gradient wind balanced model of E86. The resultant vortex provides a more realistic structure than previously applied bogus vortices.

Despite a significantly more accurate vortex than other synthetic vortices used in other studies, the algorithm does contain a number of shortcomings compared to more involved initialization techniques such as dynamic initialization and data assimilation. These shortcomings were first discussed by Kurihara et al. (1993) and include differences in the physics that define the vortex and the forward model. Furthermore, the unbalanced secondary circulation is underrepresented as it constructed from conservation of momentum in a dynamic model, without consideration of diabatic heating or environmental stratification and inertial stability. While these issues are more aptly handled in the more advanced methods, the algorithm is simple and computationally efficient, produces a realistic vortex, and is highly configurable.

A series of simulations were conducted to analyze the sensitivity of several of the key tunable parameters in the initialization methodology. To facilitate comparison, a nature simulation was conducted and sensitivity experiments were initialized during the rapid intensification phase. Removal of the vortex from the nature simulation results in dynamic adjustment as the prognostic model variables are perturbed because of the insertion of synthetic data. The dynamic adjustment dominates the early evolution such that any sensitivity in the intensity evolution to boundary layer eddy diffusivity or initial divergent flow is lost. However, differences in the structural evolution are observed. Enhancing the moisture in the eyewall significantly reduces the adjustment time. By accelerating saturation in the mesoscale core, sustained deep convection produces a deep secondary circulation as opposed to a shallow boundary layer secondary circulation, which acts to spin down the boundary layer top circulation. The sensitivity experiments were then repeated with the addition of the moisture enhancement in each case. In these cases, sensitivity to boundary layer eddy diffusivity was small and only manifested itself for unrealistically small values of the eddy diffusion coefficient.

Inclusion of a secondary circulation at the initial time produced minimal differences from an initial balanced flow regime. Only during the first few hours was the convective response to divergence notable in the near-surface tangential wind evolution. It is clear that the adjustment period is necessary to generate flow characteristics consistent with the model physics.

Two additional stages of the tropical cyclone life cycle in the nature simulation were used to test the ability of the synthetic vortex algorithm to reproduce the intensity evolution of the nature simulation: at preintensification and steady state. At both times an adjustment period was observed similar to that seen in the intensification stage sensitivity simulations. In all three cases the post-adjustment intensity evolutions of the synthetic vortex simulations closely match those of the nature simulations.

In any numerical simulation in which new data are introduced, adjustment is a reality. That being said, it is possible that the adjustment time may be reduced without the need for a moisture enhancement by balancing the inserted vortex with the environmental flow. Ideally, this would be completed using a nonlinear balance equation based on the numerical model’s equations of motion to ensure consistency. However, the vortex itself is balanced and the analysis, away from where the vortex is added, is balanced, so that rebalancing may be unnecessary. Bender et al. (2007), in 100 test cases, found no difference in evolution when rebalancing was applied.

Although several of the adjustable parameters do not appear to produce a noticeable sensitivity in the early simulated TC intensity evolution, we believe that the methodology is useful for several reasons. First, it provides new, flexible, and physically based vortex initialization software to the community for their own investigations.
of TC predictability, initialization, and sensitivity to different configurations of the TC not demonstrated in this paper. Second, it can be used to test new observing strategies in observing system simulation experiments (OSSEs). Third, a variety of configurations can be used in initialization of ensemble forecasts, or in combination with data assimilation. And finally, this methodology serves as a simple benchmark against which other more advanced initialization and assimilation methodologies can be compared.

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