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An Analysis of Cyclogenesis for Mid-Latitude and Tropical Storms Using the Petterssen-Sutcliffe Development Equation

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AN ANALYSIS OF CYCLOGENESIS FOR MID-LATITUDE AND TROPICAL STORMS USING THE PETTERSSEN-SUTCLIFFE DEVELOPMENT EQUATION

By

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The members of the Committee approve the thesis of William M. Hession defended on June 16, 2004.

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### LIST OF SYMBOLS

- $c_p$: Specific heat at constant pressure
- $\alpha$: Specific volume
- $F$: Frictional forces
- $\Gamma_a$: Adiabatic lapse rate ($K \ Pa^{-1}$)
- $f$: Coriolis force
- $\Gamma_d$: Dry adiabatic lapse rate ($K \ Pa^{-1}$)
- $g$: Gravitational constant
- $\Gamma_s$: Saturated adiabatic lapse rate ($K \ Pa^{-1}$)
- $\dot{H}$: Diabatic heating
- $\gamma_a$: Adiabatic lapse rate ($K \ km^{-1}$)
- $L$: Latent heat of vaporization
- $\gamma_d$: Dry adiabatic lapse rate ($K \ km^{-1}$)
- $p$: Pressure
- $\kappa$: Saturated adiabatic lapse rate ($K \ km^{-1}$)
- $\dot{Q}$: Diabatic heating
- $Z$: Absolute vorticity
- $q$: Mixing ratio
- $\zeta_g$: Geostrophic vorticity
- $q_s$: Saturation mixing ratio
- $\theta$: Potential temperature
- $R$: Gas constant for dry air
- $\rho$: Density
- $R_v$: Gas constant for water vapor
- $\sigma$: Stability parameter
- $S$: Stability parameter
- $\Phi$: Geopotential height
- $T$: Temperature
- $\omega$: Vertical motion
- $u$: East-west wind component
- $\nabla$: Del operator
- $V$: Horizontal wind
- $\nabla^2$: Laplacian operator
- $V_g$: Geostrophic horizontal wind
- $v$: North-south wind component
- $W$: Work due to diabatic heating
- $z$: Height
ABSTRACT

In this study, the Petterssen-Sutcliffe development equation is used to examine cyclogenesis. In the past, several other methods have been used to study cyclogenesis and calculate vertical motion, such as the kinematic and adiabatic methods, quasi-geostrophic theory as well as the approaches derived from them. However, there is little documentation on the application of the historical Petterssen-Sutcliffe method, and hence the motivation for this study.

The forcing terms of the Petterssen-Sutcliffe development equation are calculated using GEMPAK software. These forcing terms include vorticity advection, temperature advection, stability, and diabatic heating.

Two mid-latitude storms and two tropical systems were analyzed to see if this method could recognize cyclogenesis in both baroclinic and barotropic environments. The first mid-latitude storm occurred in late January 2000. It formed off the coast of the Carolinas and traveled up the East Coast over the Atlantic Ocean. The second storm spent its life cycle over land in the Mid-Atlantic and New England regions during March 1999. Both tropical systems originated in the Gulf of Mexico: Hurricane Earl (1998) and Hurricane Gordon (2000).

This method of analysis was shown to have general success in identifying cyclogenesis of mid-latitude cyclones and somewhat limited success with tropical storms. It is hoped that this method will benefit both educational and operational environments where students and forecasters can use this additional analysis to supplement their understanding of the atmosphere.
CHAPTER ONE
INTRODUCTION

Various equations and methods have been formulated to determine vertical motion in the atmosphere, which can lead to cyclogenesis. The reason for the whole host of methods is due to the fact that synoptic scale vertical motion cannot be measured directly. Ascent and descent must be calculated by indirect techniques. The earliest of these methods were developed during the first half of the 20th century. It was recognized that the basics of physics and thermodynamics could be applied to atmospheric motion (Houghton 1985). From this recognition came the kinematic and adiabatic methods.

The kinematic method is perhaps the simplest approach to calculating vertical motion. This makes use of the equation of mass continuity. The measure of vertical movement can be determined using the observed horizontal wind field and the idea that all mass flowing in and out of a volume is conserved. The kinematic method is given as

\[
\omega(p) = \omega(p_0) - \int_{p_0}^{p} \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) dp
\]

where the divergence is integrated between the surface \( p_0 \) and a desired level \( p \) to solve for the vertical motion. This method is advantageous due to its simplicity, but it does have a disadvantage. The calculated vertical motion is sensitive to any small errors in the observed horizontal wind field. However, correction schemes by researchers such as O’Brien (1970) and Pedder (1981) have been implemented to remedy this problem.

The other early method for determining vertical motion is the adiabatic method. This method is based on the first law of thermodynamics for an adiabatic atmosphere. According to the given equation,

\[
\omega = \frac{\partial T/\partial t + \mathbf{V} \cdot \nabla T}{(-T/\theta) \partial \theta/\partial p}
\]
ascent would occur with conditions of local cooling and/or warm air advection. While this method can be useful, it could not be applied in situations where strong diabatic heating is present, such as conditions of convective rainfall, since these effects are ignored in (2). Another problem occurs from the fact that atmospheric data aloft is only measured every 12 hours. The local temperature tendency \( \frac{\partial T}{\partial t} \) must be replaced by a 12-hour temperature change while the other values used in the adiabatic method are instantaneous.

The adiabatic method (summarized by Moore 2002) also provides a process where vertical motion may be obtained diagnostically, from

\[
\omega = \left( \frac{\partial p}{\partial t} \right)_\theta + \mathbf{V} \cdot \nabla p + \left( \frac{\partial p}{\partial \theta} \right) \left( \frac{\partial \theta}{\partial t} \right)
\] (2a)

where the first term involves the local pressure tendency on the isentropic surface, the second term is the negative of isentropic pressure advection, and the final term involves diabatic heating or cooling. Although in general vertical motion for most synoptic systems appears to be related to the pressure advection term, there are numerous pitfalls associated with using this method universally, particularly in mature translating systems with active precipitation (see Moore 2002).

The next innovation in calculating vertical motion was the introduction of quasi-geostrophic theory. This comes from the idea that synoptic-scale flow will always tend to be in hydrostatic and geostrophic equilibrium (Phillips 1963). Quasi-geostrophic theory gives rise to a diagnostic omega equation and a height tendency equation. These equations were used operationally during the late 1950’s and into the 1960’s. The increase in forecast skill was reasonable, but forecasts were quantitatively unreliable. In 1966 these equations were abandoned for models that used the full set of primitive equations (Houghton 1985). Holton (1992) defines the omega equation as

\[
\left( \sigma \nabla^2 + f_o \frac{\partial^2}{\partial p^2} \right) \omega = f_o \frac{\partial}{\partial p} \left[ \mathbf{V}_g \cdot \nabla \left( \frac{1}{f_o} \nabla^2 \Phi + f \right) \right] + \nabla^2 \left[ \mathbf{V}_g \cdot \nabla \left( - \frac{\partial \Phi}{\partial p} \right) \right]
\] (3)

where differential absolute vorticity advection and the Laplacian of thickness advection (proportional to temperature advection) are the two elements that determine vertical motion. Even though quantitative results are sometimes erroneous, the qualitative interpretation of quasi-geostrophic equations has proved to be most useful in providing a physical understanding of atmospheric motions.
In depth discussion of the kinematic method, adiabatic method, and quasi-geostrophic theory can be found in Bosart’s summaries in Handbook of Applied Meteorology (Houghton 1985).

As an extension of the quasi-geostrophic problem, Krishnamurti (1968) developed a general balance model to solve for omega. This equation includes the ageostrophic contributions of both the differential vorticity advection and the Laplacian of thermal advection. The other contributions in this model are divergence, vertical advection of vorticity, deformation, a beta term, a twisting term, friction, latent heat, and sensible heat transfer. This makes a total of 12 forcing functions. Krishnamurti (1968) provides a study of cyclone development using a 5-level general balance model. He shows that differential geostrophic vorticity advection and the Laplacian of geostrophic thermal advection initiate cyclone development. The latter stages of development require the previous two forcing functions as well as the effects of latent heat and friction (which quasi-geostrophic theory does not consider). Pagnotti and Bosart (1984) conducted a similar study where they used a 10-level general balance model. They showed that for weak cyclone development the effects of latent heat are as important as thermal advection. In the case of strong development, differential vorticity advection is needed along with the previous thermal effects.

Unfortunately, studies that compare kinematic, quasi-geostrophic, and general balance methods do not always agree. Baumhefner (1968) concludes that the inclusion of latent heat and friction alter vertical motion fields considerably when compared to kinematic calculations. However, Smith and Lin (1978) provide a comparison study where the kinematic method produced better results than both the quasi-geostrophic and general balance models. Since researches were not able to establish the superiority of one method over another, it became apparent that more factors need to be considered in determining vertical motion.

Seeing that the classic quasi-geostrophic approach was not fully reliable to determine vertical motion, Zwack and Okossi (1986) developed a method to solve for geostrophic vorticity tendency \( \partial \zeta_g / \partial t \) by using the same set of equations used to derive the quasi-geostrophic omega equation. This became an important new approach to the study of cyclone development since Petterssen (1956) defined cyclone development as
being proportional to the rate of production of absolute vorticity. As shown in their article, the Zwack-Okossi development equation is derived as

\[
(p_{BL} - p_T) \left( \frac{\partial \zeta}{\partial t} \right)_0 = f_0 \omega_{BL} + \int_{p_T}^{p_{BL}} - \nabla \cdot \nabla (\zeta g + f) dp
\]

\[
- \frac{R}{f_0} \int_{p_T}^{p_{BL}} \nabla^2 \left( - \nabla \cdot \nabla T + S \omega + \frac{\dot{Q}}{c_p} \right) dp dp^*
\]

(4)

where the subscripts \( BL \) and \( T \) refer to the top of the boundary layer and the top of the atmosphere, respectively; \( p^* \) is an arbitrary pressure level; \( S \) is a stability parameter equal to \( -(T/\theta)(\partial \theta / \partial p) \); and \( \dot{Q} \) is the diabatic heating rate. According to Zwack and Okossi, the following factors contribute to increasing surface geostrophic vorticity or cyclogenesis: descending motion due to orographic effects or negative Ekman pumping \( (\omega_{BL} > 0) \), positive geostrophic vorticity advection, and a maximum of warm air advection and diabatic heating. However, in areas of maximum vertical upward motion, the static stability term \( (S \omega) \) will oppose the positive forcings to cyclogenesis due to the fact that air will cool as it rises.

Lupo (1992) provides a study that compares the Zwack-Okossi development equation to an extended form of Zwack-Okossi in order to diagnose extratropical cyclones. Lupo includes ageostrophic effects since the original equation was developed in quasi-geostrophic form. His derivation yields

\[
(p_0 - p_T) \left( \frac{\partial \zeta}{\partial t} \right)_0 = \int_{p_T}^{p_0} \left( - \nabla \cdot \nabla Z \right) dp + \int_{p_T}^{p_0} \left( k \cdot \nabla \times \mathbf{F} \right) dp
\]

\[
- \frac{R}{f} \int_{p_T}^{p_0} \int_{p}^{p_0} \nabla^2 \left( - \nabla \cdot \nabla T + S \omega + \frac{\dot{Q}}{c_p} \right) dp dp^*
\]

(5)

Absolute vorticity advection has replaced the geostrophic advection in (4). The effect of friction is also part of the extended version. The contributions of temperature advection, static stability, and diabatic heating are the same. In Lupo’s study, he shows that the contributions from adiabatic cooling and diabatic heating were often the same order of magnitude as vorticity advection and temperature advection. While friction has a negative influence on cyclogenesis, it did not contribute as greatly as the other four
forcings. Lupo also shows that the extended form provides better results than Zwack and Okossi’s original form. 

One final approach to diagnosing cyclogenesis that has received very little attention in the recent literature is the original Petterssen-Sutcliffe development equation. Sutcliffe (1947) defined the rate of development of cyclones simply as convergence. Petterssen (1956) took this idea one step further by relating convergence to the rate of production of absolute vorticity. With the help of the concepts of thermal wind and thermodynamics and previous work by Sutcliffe, Petterssen derived the vorticity equation to produce the Petterssen-Sutcliffe development equation. It states that

\[
\frac{\partial Z_o}{\partial t} = A_Z - \frac{R}{f} \nabla^2 \left( \frac{g}{R} A_T + S + H \right) + \mathbf{V}_0 \cdot \nabla Z_o \tag{6}
\]

where

\[
A_Z = -\mathbf{V} \cdot \nabla Z \\
A_T = -\mathbf{V} \cdot \nabla (\Delta z) \\
S = \ln \left( \frac{p_0}{p} \right) \omega (\Gamma_o - \Gamma) \\
H = \ln \left( \frac{p_0}{p} \right) \frac{1}{c_p} \frac{dW}{dt}.
\]

Here, \(A_Z\) is the term for absolute vorticity advection (at the LND), \(A_T\) is the term representing thickness advection, \(S\) is the term for static stability, and \(H\) denotes the effects of diabatic heating. Essentially, the equation shows that positive vorticity advection at the LND and at the surface, and a maximum of the Laplacian of combined thermal effects contribute to cyclone development near sea level. One will note that this equation (6) is quite similar to both (4) and (5). All three equations show that vorticity advection, temperature advection, static stability, and diabatic heating contribute to vorticity production at the surface. Since cyclogenesis is often characterized by the deepening of sea level pressure of a cyclone, perhaps it is more insightful to diagnose cyclogenesis in terms of vorticity tendency at the surface instead of vertical motion throughout the atmosphere.

Even though the Zwack-Okossi development equation is quite similar to the Petterssen-Sutcliffe equation, no quantitative studies using the Petterssen-Sutcliffe approach have been conducted to diagnose cyclogenesis. This lack of attention is the main reason for this study. It is hoped that the use of this method may lend itself to
pedagogical weather analysis and interpretations suitable for use in training and education. The rest of this thesis describes the use of display software on gridded model data to calculate the terms of the Petterssen-Sutcliffe development equation.
CHAPTER TWO

METHODOLOGY

Referring again to (6), the Petterssen-Sutcliffe development equation (Petterssen 1956) is given as

\[ \frac{\partial Z_o}{\partial t} = -\mathbf{V} \cdot \nabla Z + \mathbf{V}_0 \cdot \nabla Z_0 \]

\[ -\frac{g}{f} \nabla^2 \left[ - \mathbf{V} \cdot \nabla (\Delta z) \right] \]

\[ -\frac{R}{f} \nabla^2 \left[ \ln \left( \frac{p_0}{p} \right) \omega (\Gamma_a - \Gamma) \right] \]

\[ -\frac{R}{f} \nabla^2 \left[ \ln \left( \frac{p_0}{p} \right) \frac{1}{c_p} \frac{dW}{dt} \right] \] (7)

where the time rate of change of absolute vorticity at the surface is partitioned into five forcing terms. The first two terms are vorticity advection at the level of non-divergence and at the surface (the latter is often neglected). The other three include thickness advection, stability, and diabatic heating. The terms contributing to temperature change use data within a layer bounded by the level of non-divergence (LND) and the surface or 1000 hPa.

Since the LND is variable, a model needed to be chosen that would provide flexibility in choosing a level. The NCEP Eta model was chosen because it provides data at 50 hPa intervals in our received grids. This was chosen over the AVN model, which contains data only at mandatory levels. When applying Dine’s compensation principle to the model data used in this study, the level of strongest upward vertical motion (and hence the level of non-divergence) was most often found at 600 hPa and was therefore the level chosen for the LND. Figure 1 shows a typical vertical profile of maximum rising motion near the center of a cyclone.
Fig. 1. Vertical profile of omega near storm center, from Eta model data.

The first forcing term of (7) is easily calculated in GEMPAK (Koch et al 1983, Unidata 1999) simply by advecting the vorticity field at the LND by the observed wind field at the same level. Vorticity advection at the surface was shown by Petterssen to be small and generally contributes little to vorticity production, but is retained in this study for the sake of completeness and is calculated in the same manner as the first term.

The remaining thermal terms require information within the 1000-600 hPa layer. As with similar studies, data that are used for calculations are taken at every 100 or 50 hPa. However, in order for GEMPAK to perform operations with different fields, the parameter that specifies the boundaries of a layer must include either the top or bottom of the layer (in this study, either 1000 hPa or 600 hPa). Therefore, in order to have the most realistic representation of the atmosphere while using GEMPAK, two intermediate layers are defined: 1000-800 hPa and 800-600 hPa. If a 1000-600 hPa layer average is required, the average of both intermediate layers is calculated using their respective boundaries and the two intermediate averages are then averaged. If a 1000-600 hPa
vertical gradient is required, the gradient of both intermediate layers is again calculated using their respective boundaries and the two intermediate gradients are then averaged.

Sutcliffe defines thickness advection using the mean wind within the layer. For this study, the thickness advection is calculated at each boundary (1000, 800, 600 hPa) and the mean is found using the averaging procedure described above. Next, the Laplacian is taken of the thickness advection field and then multiplied by \(-\frac{g}{f}\) in order to complete the forcing term.

Both the stability and diabatic terms require an average value throughout the 1000-600 hPa layer. The GEMPAK software can calculate layer averages, but only using data at the top and bottom of the layer. Therefore data from the 800 hPa level was used in order to acquire a more representative layer average.

Before calculating the stability term, one must refer to Petterssen’s notation. Petterssen denotes the lapse rate \(\Gamma\) in terms of pressure rather than in terms of height. If \(\gamma\) is defined as the actual lapse rate in terms of height, the hydrostatic equation \((-\frac{\partial p}{\partial z}) = \rho g \frac{\partial z}{\partial \rho}\) can be used to define \(\Gamma\) as the lapse rate in terms of pressure. The third term can now be written as

\[
-\frac{R}{f} \nabla^2 \left[ \ln \left( \frac{p_0}{p} \right) \frac{1}{\rho g} (\gamma_a - \gamma) \right]. \tag{8}
\]

To calculate this term, the first step is to obtain the average of the product of omega and a departure of the actual lapse rate from the adiabatic (dry or saturated) lapse rate. The lifting condensation level was chosen to be 900 hPa as suggested by Krishnamurti et al. (1973) and Edmon et al. (1976). The two intermediate layers for the stability term are 1000-900 hPa (dry ascent) and 900-600 hPa (saturated ascent). In the lower layer, the 1000-900 hPa lapse rate is compared to the dry adiabatic lapse rate \((\gamma_d = 9.8 \text{ K km}^{-1})\). In the upper layer, the 900-600 hPa lapse rate is compared to the saturated adiabatic lapse rate approximated by Fleagle and Businger (1980) to be

\[
\gamma_s \equiv \frac{g}{c_p} \left(1 + \frac{L q_s}{RT} \right) \left(1 + \frac{L^2 q_s}{c_p R_v T^2} \right)^{-1} \tag{9}
\]

where \(q_s\) is the saturation mixing ratio. These comparisons are then multiplied by their respective intermediate averages of \((\omega/\rho g)\). A weighted average is then applied so that
the lower layer provides 25 percent of the total stability and the upper layer provides 75 percent. An adjustment still needs to be made to this part of the third term: unit conversion. In GEMPAK, omega is assigned units of mb s\(^{-1}\) and lapse rate is assigned units of K km\(^{-1}\); therefore, the omega must be converted to Pa s\(^{-1}\) and the lapse rate must be converted to K m\(^{-1}\). Once this is calculated, the Laplacian is taken and then multiplied by \(-\left(\frac{R}{f}\right)\ln\left(p_0/p\right)\). These calculations are illustrated in Appendix A.

The term that contains \(\frac{dW}{dt}\) represents the forcing due to diabatic effects (analogous to \(\dot{Q}\) in the other systems surveyed in the previous chapter). Petterssen defines this in terms of energy where \(dW = c_p dT - \alpha dp\). Diabatic heating can be partitioned into sensible and latent heating where one form of the first law of thermodynamics yields \(\dot{H} = c_p d\theta - Ldq\) (Houze 1993). So theoretically, 

\[
\frac{dW}{dt} = c_p \frac{d\theta}{dt} - L \frac{dq}{dt} .
\]

However, the three heating terms (temperature advection, stability, and diabatic heating) are interrelated (Petterssen 1956). The effect of latent heat release (second term on the right hand side of (10)) is inherently part of the stability term since we are comparing the saturated lapse rate to the environmental lapse rate above 900 hPa. Therefore, in this study,

\[
\frac{dW}{dt} = c_p \frac{d\theta}{dt}
\]  

and

\[
\frac{dW}{dt} = c_p \left( \frac{\partial \theta}{\partial t} + \nabla \cdot \nabla \theta + \frac{\partial \theta}{\partial p} \right) .
\]

The local time derivative of potential temperature is calculated using the method of backwards differencing. A time difference is calculated from each model data and its previous 12-hour model data. The horizontal and vertical advection quantities of potential temperature are once again found by averaging the intermediate values. The partial components are summed to obtain the total derivatives and multiplied by the specific heat constant. The Laplacian is taken and then multiplied by \(-\left(\frac{R}{f}\right)\ln\left(p_0/p\right)\left(1/c_p\right)\) to produce the complete fourth term.
Upon initial testing of the GEMPAK script (available in Appendix B), it was found that both the stability and diabatic forcing terms were one order of magnitude greater than synoptic scaling would suggest. Upon analyzing the two terms, it was discovered that vertical advection in the diabatic term (12) was dominating strongly over the horizontal advection and time tendency terms. It was discovered that the Eta model was producing large variations of vertical motion in at scales smaller than the synoptic scale. A nine-point smoother is applied twice on both the stability and diabatic terms before the Laplacian is taken since both terms contain the term for vertical motion.

Finally, an area average was calculated using data points around each initialized low pressure center and along its projected path using 600 hPa streamlines. An elliptical area was chosen for the mid-latitude cyclones while a more circular area was used for the hurricanes. The averaging area is approximately 200,000 km$^2$ (essentially 4$^\circ$ latitude X 4$^\circ$ latitude). Figure 2 depicts a sample of the averaging areas for both types of storms.

Fig. 2. Averaging area depicted with dashed lines. Isobars (hPa) in solid lines.
A) mid-latitude system, 1200 UTC 24 January 2000.
B) tropical system, 0000 UTC 02 September 1998.
CHAPTER THREE
RESULTS

A. Mid-Latitude Systems

In this study, the Petterssen-Sutcliffe method is used to evaluate two mid-latitude cyclones. The two storms chosen were both examples of explosive cyclogenesis. The strength of these two particular storms was poorly forecast. Perhaps this is due to the fact that the models cannot handle strong ageostrophic forcing; therefore, using quasi-geostrophic methods may prove to be unreliable here. Studies of explosive cyclogenesis (Uhl et al. 1992, Lupo et al. 1992, Rolfson et al. 1996) using the Zwack-Okossi development equation and its extended form have shown that effects of diabatic heating and adiabatic cooling play important roles in these situations. The Petterssen-Sutcliffe method would include these types of forcings.

The first cyclone discussed here occurred during late January 2000. Figures 3 and 4 illustrate the evolution of this storm from 1200 UTC 24 January through 0000 UTC 26 January. An upper level trough with an axis extending from Ontario and Quebec down toward the central Gulf Coast moved over a surface frontal boundary located along the Gulf Coast. As the trough moved eastward, a cut-off low formed which moved the developing surface low over the waters of the Gulf Stream. The rapidly deepening surface low then traveled north along the Atlantic coastline with central low pressures dropping below 990 hPa. The storm began to weaken during the occlusion process as it approached Maine and the Canadian Maritimes. Synoptic studies of this cyclone appear in Buizza and Chessa (2002), Langland et al. (2002), and Zhang et al. (2002).

Figure 5 depicts the contributions of each of the five forcing terms toward cyclogenesis during each 12-hour period from 1200 UTC 24 January to 1200 UTC 26 January. The magnitude of the forcings is shown as \(10^{-9}\) with units of s\(^{-2}\). Table 1 shows the numerical results of the total and individual forcings as well as the central sea level.
Fig. 3. Height contours at 600 hPa, from Eta model initialization.
D) 0000 UTC 26 January 2000. E) 1200 UTC 26 January 2000
Fig. 3 - continued.
Fig. 4. Isobars of sea level pressure, from Eta model initialization.
D) 0000 UTC 26 January 2000. E) 1200 UTC 26 January 2000
Fig. 4 - continued.
Fig. 5. Contribution of Petterssen-Sutcliffe terms toward cyclogenesis, from 1200 UTC 24 January 2000 to 1200 UTC 26 January 2000.

pressure of the cyclone. On 1200 UTC 24 January, the time of the initial stage of the cyclone, the terms contributing most to cyclogenesis are diabatic heating and thickness advection with diabatic heating contributing the most. The remaining terms are not producing significant amounts of forcing at this time. Twelve hours later (Time 2), the storm undergoes dramatic deepening. On 0000 UTC 25 January the energy associated with an upper level vorticity maximum had moved over the cyclone as the storm center moved over the waters of the Gulf Stream off the coast of the Carolinas. Forcing from both vorticity advection and thickness advection has significantly increased. The effects due to diabatic heating have decreased, but the term is still positive. The stability term is producing a negative contribution, and the surface vorticity advection term is still negligible. As the time advances to 1200 UTC 25 January (Time 3) the central surface pressure of the storm continues to deepen; however, the only term offering significant forcing is thickness advection. Normally, the stability term contributes negatively to cyclogenesis since air will cool when it rises. At this time frame the stability term is positive. This is due to the fact that the environmental lapse rate in the region is greater.
Table 1. Contribution of Petterssen-Sutcliffe terms, sum of terms, and sea level pressure at storm’s center from 1200 UTC 24 January 2000 to 1200 UTC 26 January 2000.

<table>
<thead>
<tr>
<th></th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
</tr>
</thead>
<tbody>
<tr>
<td>Vorticity Adv</td>
<td>-0.4</td>
<td>3.8</td>
<td>0.6</td>
<td>2.6</td>
<td>4.8</td>
</tr>
<tr>
<td>Thick Adv</td>
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<td>9.0</td>
<td>9.7</td>
<td>6.3</td>
<td>4.6</td>
</tr>
<tr>
<td>Stability</td>
<td>0.8</td>
<td>-3.9</td>
<td>1.9</td>
<td>-2.6</td>
<td>-7.2</td>
</tr>
<tr>
<td>Diabatic</td>
<td>4.9</td>
<td>2.9</td>
<td>-5.2</td>
<td>1.2</td>
<td>4.4</td>
</tr>
<tr>
<td>Sfc Vort Adv</td>
<td>0.1</td>
<td>-0.4</td>
<td>-1.3</td>
<td>-0.4</td>
<td>-0.6</td>
</tr>
<tr>
<td>Total</td>
<td>9.1</td>
<td>11.4</td>
<td>5.7</td>
<td>7.1</td>
<td>6.0</td>
</tr>
<tr>
<td>SLP</td>
<td>1007</td>
<td>992</td>
<td>985</td>
<td>981</td>
<td>990</td>
</tr>
</tbody>
</table>

than the saturated adiabatic lapse rate. This will cause the sign to change from negative to positive in the stability term. Contributions from surface vorticity advection and diabatic heating have decreased, where the contribution from vorticity advection is close to zero. It is interesting, and perhaps puzzling, to see that vorticity advection is negligible and diabatic heating is negative, especially since the storm is still gaining strength. Diabatic heating less than zero could be a sign of evaporational cooling. The next time frame (Time 4) shows the storm at its strongest during 0000 UTC 26 January with a central low pressure of 981 hPa. Forcing due to vorticity advection and diabatic heating have now increased and are providing positive contributions. Positive forcing due to thickness advection begins to decline and the stability term is now working against cyclogenesis. Forcing from surface vorticity advection is now negligible and continues in this manner through the rest of the storm’s life cycle. In the last analyzed stage of the cyclone on 1200 UTC 26 January (Time 5), the storm enters the occlusion stage and the central surface pressure begins to rise. The thickness advection term has decreased. However, it is curious to see that vorticity advection has increased and is now at its greatest contribution of all five time frames. The diabatic term has also increased considerably. Negative forcing from the stability term has increased and is also at its greatest (negative) contribution of all five time frames. It is also the term of greatest intensity over the other individual terms at this last time frame.

Figure 6 shows the sum of all the forcings for each time frame. As expected, there is a sharp increase in forcing between the first and second time periods. This
corresponds to the rapid deepening of the cyclone at the beginning of its life cycle. Even though the storm continues to deepen during its first 48 hours, the total forcing decreases in an inconsistent manner after its first 24 hours, partially due to the surprising lack of vorticity advection and the negative contribution from the diabatic heating term. At the occlusion stage (Time 5), the results here would suggest further intensification, but this could be attributed to the unanticipated increase in vorticity advection and diabatic heating. Also, the method lacks a term for frictional dissipation as in (5), which would help to weaken the cyclone.

![Total Forcing](image)

**Fig. 6.** Net sum of Petterssen-Sutcliffe forcing terms toward cyclogenesis, from 1200 UTC 24 January 2000 to 1200 UTC 26 January 2000.

The next mid-latitude storm analyzed in this study primarily affected the northeastern United States in early March 1999. The life cycle of this storm is shown in Figures 7 and 8. The upper level charts during this time show a short wave rotating off of a large trough located over eastern Canada. The wave deepens into a trough with an axis extending from Ontario down toward Mississippi and Alabama. At this point a surface low develops over the Ohio Valley. As the bottom of the axis of the trough swings toward the Carolinas, a new low appears to develop over the Mid-Atlantic states along
Fig. 7. Height contours at 600 hPa, from Eta model initialization.
Fig. 8. Isobars of sea level pressure, from Eta model initialization.
the same surface boundary. The wave then lifts northward and can be seen as a cut-off low where the surface low continues to deepen and then finally weakens as it moves into Quebec.

Figure 9 depicts the contributions of each of the five forcing terms for this storm during each 12-hour period from 0000 UTC 03 March to 0000 UTC 05 March. Table 2 again shows the calculated data. At the time that the low pressure center first develops (Time 1), the terms contributing positively to cyclogenesis are thickness advection and diabatic heating. The stability term provides a negative contribution. All remaining terms essentially give no contribution. The central sea level pressure of the cyclone deepens slightly over the next 12 hours (Time 2); however, cyclogenetic forcing has decreased. There is only a small amount of forcing from vorticity advection and diabatic heating. Forcing due to thickness advection is now negligible as well as forcing from stability. Surface vorticity advection contributes negatively at this time, but its forcing is negligible. On 0000 UTC 04 March (Time 3), the low has redeveloped along the
Table 2. Contribution of Petterssen-Sutcliffe terms, sum of terms, and sea level pressure at storm’s center from 0000 UTC 03 March 1999 to 0000 UTC 05 March 1999.

<table>
<thead>
<tr>
<th></th>
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<th>5</th>
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</thead>
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<td>Vorticity Adv</td>
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<td>1.7</td>
<td>1.0</td>
<td>5.2</td>
<td>1.5</td>
</tr>
<tr>
<td>Thick Adv</td>
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<td>-0.1</td>
<td>10.0</td>
<td>4.3</td>
<td>12.0</td>
</tr>
<tr>
<td>Stability</td>
<td>-3.4</td>
<td>-0.2</td>
<td>2.1</td>
<td>-2.7</td>
<td>-4.0</td>
</tr>
<tr>
<td>Diabatic</td>
<td>3.8</td>
<td>1.2</td>
<td>1.9</td>
<td>8.1</td>
<td>3.8</td>
</tr>
<tr>
<td>Sfc Vort Adv</td>
<td>-0.2</td>
<td>-0.6</td>
<td>-0.7</td>
<td>-0.8</td>
<td>-2.7</td>
</tr>
<tr>
<td>Total</td>
<td>6.1</td>
<td>2.0</td>
<td>14.3</td>
<td>14.1</td>
<td>10.6</td>
</tr>
<tr>
<td>SLP</td>
<td>1003</td>
<td>999</td>
<td>990</td>
<td>985</td>
<td>993</td>
</tr>
</tbody>
</table>

Surface frontal boundary and cyclogenetic forcing has returned. Thickness advection is the dominant forcing term at this time. However, there is only a small amount of forcing due to vorticity advection. The stability and diabatic terms contribute positively. The surface vorticity advection term contributes negatively, but its contribution is very small. The storm is at its strongest on 1200 UTC 04 March (Time 4) with a central low pressure of 985 hPa. Forcing due to vorticity advection and diabatic heating have both increased significantly with diabatic heating being the dominant term at this time. The contribution from thickness advection has decreased but provides a contribution similar to vorticity advection. Now, at the height of the storm, the stability term begins to contribute negatively to cyclogenesis. Surface vorticity advection still produces an insignificant negative contribution. As the storm moves over Canada, the cyclone finally begins to weaken (Time 5). Forcing due to thickness advection is still positive, but showing a surprising increase. The vorticity advection and diabatic terms have both decreased. The stability term is now working harder against cyclogenesis as the cyclone fills.

The sum of all forcings is depicted in Figure 10. At first, there is a moderate amount of forcing at the initialization of the storm, but this forcing decreases 12 hours later even though the storm has strengthened in intensity. Hope returns for proponents of cyclogenesis 24 hours after the first time frame from the redeveloping surface low. At this point (Time 3) forcing has increased dramatically and stays approximately at the same level at the height of the storm (Time 4). Finally, at the last stage of the cyclone,
the surface low is occluded and is situated underneath the upper level low. Even though the total forcing is positive (which would imply cyclogenesis), a qualitative analysis shows total forcing decreases as the storm weakens.

B. Tropical Cyclones

The next two cyclones studied were probably not the intent of either Petterssen or Sutcliffe to be analyzed using this method. However, since the equation analyzes cyclone intensification at the surface and includes diabatic and stability effects, we were curious to see how this approach would interpret tropical cyclones. Generally vorticity effects are thought to be negligible for tropical cyclones, but are dominated by diabatic effects. Also, hybrid systems exhibiting baroclinic characteristics are also common as some tropical systems move into higher latitudes. The two storms analyzed in this section are both minimal hurricanes that originated in the region of the Gulf of Mexico. They were selected because both hurricanes developed hybrid baroclinic characteristics during their life cycles. They were also chosen because the locations of their entire life cycles were contained within the grid area of the Eta model.
The first hurricane presented here, Hurricane Earl, formed in early September 1998 and Figure 11 shows the track of this storm. It originated in the Bay of Campeche and traveled northward toward Louisiana as a tropical storm. It was expected to make landfall in that state, but it made an abrupt turn toward the northeast. As it headed toward the Florida panhandle, it gained hurricane status and reached Category 2 on the Saffir-Simpson scale. Hurricane Earl made landfall in the Florida Panhandle as a Category 1 hurricane during the early morning hours of 09 September and quickly lost its strength as the eye of the storm moved over land and into southern Georgia.

The forcings of Hurricane Earl were analyzed over four time frames ranging from 0000 UTC 02 September through 1200 UTC 03 September. Figure 12 depicts the five separate forcings. As a comparison to the mid-latitude cyclones, one can see the overall forcing for this cyclone is weaker. Also, the diabatic term is generally the dominant term during the storm’s life cycle. Table 3 contains the forcing values and central sea level pressure of the storm. On 0000 UTC 02 September, the first time frame analyzed, Earl has reached tropical storm status. The term providing the largest forcing is diabatic.
heating, but it is negative. Again, a negative diabatic term could be caused by the effect of evaporational cooling. All other terms are relatively small, approximately on the order of $1.0 \times 10^{-9}$ s$^{-2}$ or less. Twelve hours later (Time 2) the diabatic term has now become positive giving a moderate contribution toward cyclogenesis. Little significant change has occurred with the other four terms. Intensification of the cyclone is shown at the third time frame during 0000 UTC 03 September. At this time the storm has reached

<table>
<thead>
<tr>
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<th>1</th>
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</thead>
<tbody>
<tr>
<td>Vorticity Adv</td>
<td>-0.3</td>
<td>-0.4</td>
<td>-0.4</td>
<td>2.9</td>
</tr>
<tr>
<td>Thick Adv</td>
<td>-1.5</td>
<td>0.9</td>
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<td>5.9</td>
</tr>
<tr>
<td>Stability</td>
<td>1.1</td>
<td>0.4</td>
<td>0.0</td>
<td>0.1</td>
</tr>
<tr>
<td>Diabatic</td>
<td>-8.5</td>
<td>6.8</td>
<td>2.1</td>
<td>18.5</td>
</tr>
<tr>
<td>Sfc Vort Adv</td>
<td>-1.6</td>
<td>-1.7</td>
<td>-2.8</td>
<td>-1.3</td>
</tr>
<tr>
<td>Total</td>
<td>-10.8</td>
<td>6.0</td>
<td>-1.9</td>
<td>26.1</td>
</tr>
<tr>
<td>SLP</td>
<td>998</td>
<td>996</td>
<td>987</td>
<td>989</td>
</tr>
</tbody>
</table>
hurricane status. The only terms providing significant forcing are diabatic heating (positive) and surface vorticity advection (negative). Twelve hours later (Time 4) the center of the storm is over land and has weakened; however, the forcing terms do not reflect this. Forcing due to diabatic heating has increased dramatically with upper level vorticity advection and thickness advection providing positive contributions. This could be due to the storm moving into higher latitudes and gaining hybrid characteristics. Negative forcing due to surface vorticity advection has been slightly reduced and forcing due to stability is negligible.

The sum of all forcings is depicted in Figure 13. Between the first two time frames there is some intensification of the storm; however, the total forcing is negative.

![Total Forcing Graph](image)

**Fig. 13.** Net sum of Petterssen-Sutcliffe forcing terms toward cyclogenesis, from 0000 UTC 02 September 1998 to 1200 UTC 03 September 1998.

Again, this is due in large part to the unexpected results of the diabatic term. On 1200 UTC 02 September (Time 2) the forcing is positive and over the next 12 hours the storm did strengthen to ‘Hurricane’ status. At the height of the storm (Time 3), the total forcing becomes negative and the hurricane does in fact weaken over the next 12 hours.
In the final stage, when the storm weakens and is over land, cyclogenetic forcing is now positive. If the storm had re-intensified in the 12 hours following this time, this positive forcing might have made sense. However, the central pressure of the storm continued to increase after this point.

The second tropical cyclone analyzed is Hurricane Gordon from September 2000. Figure 14 shows the path of the hurricane. This storm originated as a tropical depression over the Yucatan Peninsula and moved into the Gulf of Mexico as a tropical storm. As it

moved in a northeast direction it achieved hurricane status on 0000 UTC 17 September. It made landfall approximately 24 hours later as a tropical storm on the Florida peninsula near Cedar Key.

The model data studied from this storm include four time frames ranging from 1200 UTC 16 September through 0000 UTC 18 September. Figure 15 shows the calculated results of each of the five forcing terms. Table 4 also includes the associated
Fig. 15. Contribution of Petterssen-Sutcliffe terms toward cyclogenesis, from 1200 UTC 16 September 2000 to 0000 UTC 18 September 2000.

Table 4. Contribution of Petterssen-Sutcliffe terms, sum of terms, and sea level pressure at storm’s center from 1200 UTC 16 September 2000 to 0000 UTC 18 September 2000.

<table>
<thead>
<tr>
<th></th>
<th>1</th>
<th>2</th>
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<th>4</th>
</tr>
</thead>
<tbody>
<tr>
<td>Vorticity Adv</td>
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<td>0.4</td>
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<td>1.0</td>
</tr>
<tr>
<td>Thick Adv</td>
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<td>-0.1</td>
<td>2.5</td>
<td>2.0</td>
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<tr>
<td>Stability</td>
<td>0.1</td>
<td>1.0</td>
<td>1.9</td>
<td>-1.1</td>
</tr>
<tr>
<td>Diabatic</td>
<td>2.1</td>
<td>8.0</td>
<td>8.4</td>
<td>4.7</td>
</tr>
<tr>
<td>Sfc Vort Adv</td>
<td>0.0</td>
<td>-0.8</td>
<td>-1.2</td>
<td>-0.5</td>
</tr>
<tr>
<td>Total</td>
<td>2.2</td>
<td>8.5</td>
<td>11.8</td>
<td>6.1</td>
</tr>
<tr>
<td>SLP</td>
<td>997</td>
<td>985</td>
<td>981</td>
<td>988</td>
</tr>
</tbody>
</table>

data values as well as the total forcing and central sea level pressure of the cyclone. During the first time frame the cyclone has the status of Tropical Storm Gordon and is situated over water. The only contribution to cyclogenesis is a small amount from the diabatic term. All other terms are negligible. Twelve hours later (Time 2) the storm has reached hurricane strength. The diabatic term is again providing the greatest contribution toward cyclogenesis. The only other terms of relative significance are stability and
surface vorticity advection; however, they have opposite signs and cancel each other. At 1200 UTC 17 September (Time 3), the cyclone is still a hurricane and is at its peak strength. The diabatic term is approximately the same as the previous 12-hour time period. There are positive contributions from both the thickness advection and stability terms. There is a relatively small negative contribution from surface vorticity advection and upper level vorticity advection is negligible. During the final time frame the cyclone is classified as a tropical storm and is about to make landfall. The diabatic heating term has decreased, but it is still positive. There are also positive contributions from vorticity advection and thickness advection. The stability term provides a negative contribution and the surface vorticity advection term is negligible.

The sum of all forcings for each time frame is shown in Figure 16. The results from Hurricane Gordon appear to have had better results than the other tropical system analyzed in this study. At the first two time frames there is positive forcing. The tropical storm did grow stronger after each of these two time frames. At the third and fourth time frames one would expect to see negative forcing since the storm continuously weakens after 1200 UTC 17 September (Time 3), but this was not the case. However, a qualitative

![Total Forcing](image)

**Fig. 16.** Net sum of Petterssen-Sutcliffe forcing terms toward cyclogenesis, from 1200 UTC 16 September 2000 to 0000 UTC 18 September 2000.
analysis is more insightful. There is an increase in forcing as the storm grows stronger (Time 1 – Time 3) and a decrease in forcing as the storm weakens (Time 3 – Time 4).

C. Validation of Method

Finally the calculated values of forcing need to be compared to the observed values of surface vorticity tendency ($\partial Z_0/\partial t$). This is shown in Tables 5 through 8. The calculated and observed forcings are in units of $10^{-9}$ s$^{-2}$. The central sea level pressure is in units of hPa. Table 5 also compares observed $\partial Z_0/\partial t$ to what its value was forecast to be 12 hours before. The calculated values tend to be a bit larger than the observed ones.

Table 5. Surface vorticity tendency (Petterssen-Sutcliffe, observed, and 12-hour forecast) along with central sea level pressure of storm, from 1200 UTC 24 January 2000 to 1200 UTC 26 January 2000.

<table>
<thead>
<tr>
<th></th>
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<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
</tr>
</thead>
<tbody>
<tr>
<td>Calculated</td>
<td>9.1</td>
<td>11.4</td>
<td>5.7</td>
<td>7.1</td>
<td>6.0</td>
</tr>
<tr>
<td>Observed</td>
<td>0.9</td>
<td>3.4</td>
<td>4.4</td>
<td>3.2</td>
<td>2.9</td>
</tr>
<tr>
<td>12-Hour Forecast</td>
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<td>2.3</td>
<td>3.2</td>
<td>3.2</td>
<td>1.9</td>
</tr>
<tr>
<td>SLP</td>
<td>1007</td>
<td>992</td>
<td>985</td>
<td>981</td>
<td>990</td>
</tr>
</tbody>
</table>

Table 6. Surface vorticity tendency (Petterssen-Sutcliffe and observed) along with central sea level pressure of storm, from 0000 UTC 03 March 1999 to 0000 UTC 05 March 1999.

<table>
<thead>
<tr>
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<th>4</th>
<th>5</th>
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<tbody>
<tr>
<td>Calculated</td>
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<td>2.0</td>
<td>14.3</td>
<td>14.1</td>
<td>10.6</td>
</tr>
<tr>
<td>Observed</td>
<td>2.4</td>
<td>1.8</td>
<td>3.4</td>
<td>3.2</td>
<td>3.8</td>
</tr>
<tr>
<td>SLP</td>
<td>1003</td>
<td>999</td>
<td>990</td>
<td>985</td>
<td>993</td>
</tr>
</tbody>
</table>
Table 7. Surface vorticity tendency (Petterssen-Sutcliffe and observed) along with central sea level pressure of storm, from 0000 UTC 02 September 1998 to 1200 UTC 03 September 1998.

<table>
<thead>
<tr>
<th></th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
</tr>
</thead>
<tbody>
<tr>
<td>Calculated</td>
<td>-10.8</td>
<td>6.0</td>
<td>-1.9</td>
<td>26.1</td>
</tr>
<tr>
<td>Observed</td>
<td>0.4</td>
<td>0.9</td>
<td>2.9</td>
<td>1.3</td>
</tr>
<tr>
<td>SLP</td>
<td>998</td>
<td>996</td>
<td>987</td>
<td>989</td>
</tr>
</tbody>
</table>

Table 8. Surface vorticity tendency (Petterssen-Sutcliffe and observed) along with central sea level pressure of storm, from 1200 UTC 16 September 2000 to 0000 UTC 18 September 2000.

<table>
<thead>
<tr>
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<th>1</th>
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<th>3</th>
<th>4</th>
</tr>
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<tbody>
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<td>8.5</td>
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<td>6.1</td>
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<tr>
<td>Observed</td>
<td>2.1</td>
<td>2.9</td>
<td>4.2</td>
<td>2.7</td>
</tr>
<tr>
<td>SLP</td>
<td>997</td>
<td>985</td>
<td>981</td>
<td>988</td>
</tr>
</tbody>
</table>

This is could be due to the smoothing method when calculating the diabatic and stability terms. It could also be due to the lack of a frictional term as seen in the extended form of the Zwack-Okossi development equation (5). Naturally the surface vorticity tendency should increase as a storm grows stronger. This relationship is seen more often looking at the observed $\partial Z_0/\partial t$. The calculated forcing does show this trend with the two mid-latitude storms, but only in a more general fashion.
CHAPTER FOUR
CONCLUSIONS

Many of the citations used in this study have already established the timing of the forcing terms in relation to the life cycle of a mid-latitude cyclone. Generally vorticity advection, thickness advection, and diabatic heating are instrumental in the beginning stages of cyclogenesis and become less effective as the storm continues. Stability becomes more important as the cyclone matures and finally dominates at the occlusion stage. This final section describes the effectiveness of the Petterssen-Sutcliffe approach while using GEMPAK as compared to the established understanding of cyclogenesis.

The 24-26 January 2000 storm is a perfect example of a type of explosive cyclogenesis known as a Carolina low. This occurs when an existing surface low moves off the coast of the Carolinas, thermal effects are enhanced by the contrast of cold air and the warmer Gulf Stream waters, and a trough advances into the region, which brings vorticity advection. The vorticity advection was not a factor during the first stage analyzed; however, it could be seen that a vorticity maximum was traveling along the jet stream and heading toward the surface low. At the time of rapid deepening the vorticity maximum had reached the surface low, which was also over the open waters. So it is not surprising that the results show forcing due to vorticity advection at this time. The storm continues to deepen over the following 24 hours; however, vorticity advection dropped to a nearly negligible level and then continued to rise again. It is also interesting to see, as the storm reached the occlusion phase, that the forcing from vorticity advection is positive and at its highest level throughout the life cycle of the storm.

The pattern of forcing due to thickness advection is more typical of what one would expect. Positive forcing exists at the beginning of the storm. As explosive cyclogenesis occurs the forcing dramatically increases and continues to increase as the storm grows stronger. At the height of the storm the forcing begins to wane and continues to decrease as the storm weakens, but there is a slight increase in forcing during
the occlusion phase. The pattern of the diabatic term had unexpected results. At the beginning of the storm there is positive forcing (as one would expect to see), but the forcing becomes negative just before the height of the storm. Also, as the storm weakens, the forcing becomes positive and at a value as high as it was in the beginning of the storm. The stability term is generally consistent with the life cycle of the storm. During the open stage of the cyclone there is consistent negative forcing (except for one time frame) ending with a stronger negative contribution at the occlusion stage. The unexpected results at 1200 UTC 25 January (Time 3) of vorticity advection, stability, and diabatic heating could be correlated to poor resolution of the storm by the Eta model.

It appears this approach worked fairly well with the January 2000 storm with a few exceptions. The exceptions are the erratic patterns from the third and fifth time frames. It is possible that these erratic patterns could be due to the model’s poor resolution of extreme events. However, the terms of vorticity advection, thickness advection, and diabatic heating were all instrumental in the initial stage of explosive cyclogenesis and afterwards there was a (general) gradual decrease of forcing. The stability term provides negative forcing as the cyclone matures since dryer, cooler air is continuously drawn into the system and finally becomes the dominant term at the end of the life cycle.

The March 1999 storm provided interesting results due to the regeneration of the surface low along the frontal boundary. Apparently this is why there was very little forcing at the second time frame analyzed. When cyclogenesis begins with the secondary low, forcing from vorticity advection is very low, as thickness advection dominates. Vorticity advection does provide positive forcing, but not until the height of the storm. The progression of the thickness advection term is perhaps more typical. There is a generous amount of forcing at the initial cyclogenetic stage of the secondary low and then decreases afterward, but the forcing resurges during the occlusion stage. This increase in thickness advection forcing occurred during the final stage of the January 2000 storm as well. Also, while diabatic heating was not a strong factor initially, it did provide plenty of forcing during the height of the storm and then decreased as the storm weakened. Finally the stability term again shows a positive contribution just before the height of the storm (as in the January storm) and then becomes more negative as the storm weakens.
Perhaps it is not surprising that this approach did not work as well with tropical systems in this study. Again, this could be due to the fact that tropical systems originate with the lack of upper level dynamics. This study failed to relate surface vorticity tendency to cyclogenesis with Hurricane Earl; however, qualitatively, did a good job with Hurricane Gordon. During Hurricane Earl, there was an alternating pattern of negative then positive forcing during the storm’s life cycle. During Hurricane Gordon, the total forcing increases as the storm gains strength then decreases as the storm weakens.

At this point, further testing should be done on more storms to check the validity of this method. Also if there is a way to include a frictional term, this method would provide a more inclusive analysis of cyclogenesis. However, a more immediate application could be used in an educational setting as a comparison to other theories of cyclogenesis that are typically presented at a collegiate level. While no method is found to be perfect, there are a few advantages to using the Petterssen-Sutcliffe approach as compared to others. For example, Q-G theory does not have a direct relationship to surface data, hence possibly missing out on pertinent information. Also, if a jet streak at upper levels is missed because winds aloft are not available, the Petterssen-Sutcliffe approach could still be used since data is only needed from the surface up to the LND. Finally, the application in the operational field could prove quite instrumental. Meteorologists could supplement and possibly improve their forecasts with a better understanding of the atmosphere during the life cycle of a cyclone.
APPENDIX A

VARIABLES USED IN GEMPAK SCRIPTING

\[ T_1 > - \mathbf{V} \cdot \nabla Z \]

\[ T_2 > - \frac{R}{f} \nabla^2 \left[ \frac{g}{R} (- \mathbf{V} \cdot \nabla (\Delta z)) \right] \]

\[ - \frac{g}{f} \nabla^2 [- \mathbf{V} \cdot \nabla (\Delta z)] \]

\[ T_3 > - \frac{R}{f} \nabla^2 \left[ \ln \left( \frac{p_0}{p} \right) \omega (\Gamma_a - \Gamma) \right] \]

\[ - \frac{287}{f} \nabla^2 \ln \frac{1000}{600} \omega (\Gamma_a - \Gamma) \]

\[ - \frac{146.6}{f} \nabla^2 \left( \omega (\Gamma_a - \Gamma) \right) \]

\[ \text{OLRV} > \frac{\omega (\Gamma_a - \Gamma)}{2} = \frac{\text{OLR1} + \text{OLR2}}{2} \]

\[ \text{OLR1} > \bar{\omega}_{1000:900} \left( \frac{100 \text{ Pa}}{1 \text{ mb}} \right) \text{GAM1} \]

\[ \text{OLR2} > \bar{\omega}_{900:600} \left( \frac{100 \text{ Pa}}{1 \text{ mb}} \right) \text{GAM2} \]

\[ \text{GAM1} > \left( \frac{\gamma - \gamma}{\rho g} \right)_{1000:900} \left( \frac{1 \text{ km}}{1000 \text{ m}} \right) \]

\[ \text{GAM2} > \left( \frac{\gamma - \gamma}{\rho g} \right)_{900:600} \left( \frac{1 \text{ km}}{1000 \text{ m}} \right) \]

\[ \text{GSAT} \approx \gamma_s \approx \frac{g}{c_p} \left( 1 + \frac{Lq_s}{RT} \right) \left( 1 + \frac{L^2 q_s}{c_p R T^2} \right)^{-1} > \frac{G600 + G900}{2} \]
A900 > \left(1 + \frac{Lq_s}{RT}\right)_{900}

B900 > \left(\frac{L^2 q_s}{c_p R_v T^2}\right)_{900}

G900 > \frac{g}{c_p} (A900)(1 + B900)^{-1}

A600 > \left(1 + \frac{Lq_s}{RT}\right)_{600}

B600 > \left(\frac{L^2 q_s}{c_p R_v T^2}\right)_{600}

G600 > \frac{g}{c_p} (A600)(1 + B600)^{-1}

DEN1 > \bar{\rho}_{1000:900}

DEN2 > \bar{\rho}_{900:600}

T4 > \frac{R}{f} \nabla^2 \ln \frac{p_0}{p} \frac{1}{c_p} \frac{d\bar{W}}{dt}

- \frac{287}{f} \nabla^2 \ln \frac{1000}{600} \frac{1}{1004} \frac{d\bar{W}}{dt}

- \frac{0.1460}{f} \nabla^2 \frac{d\bar{W}}{dt}

DWDT > \frac{d\bar{W}}{dt} = c_p \frac{d\bar{\theta}}{dt} > 1004 \text{ J K}^{-1} \text{ kg}^{-1} (\text{DTA} + \text{DTB} + \text{DTC})

DTA > \frac{\partial \bar{\theta}}{\partial t} = \frac{TAVG_{10} - TAVG_{14}}{43,200 \text{ s}}

TAVG > \bar{\theta} = \frac{TAV1 + TAV2}{2}
TAV1 > \( \bar{\theta}_{1000:800} \)

TAV2 > \( \bar{\theta}_{800:600} \)

DTB > \( \frac{\nabla \cdot \nabla \theta}{\nabla \cdot \nabla} = \frac{DTB1 + DTB2}{2} \)

DTB1 > \( (\nabla \cdot \nabla \theta)_{1000:800} \)

DTB2 > \( (\nabla \cdot \nabla \theta)_{800:600} \)

DTC > \( -\frac{\partial \bar{\theta}}{\partial p} = \frac{DTC1 + DTC2}{2} \)

DTC1 > \( \frac{\bar{\omega}_{1000:800} \left( \frac{100 \text{ Pa}}{1 \text{ mb}} \right) \theta_{1000} - \theta_{600}}{200 \text{ hPa}} \)

DTC2 > \( \frac{\bar{\omega}_{800:600} \left( \frac{100 \text{ Pa}}{1 \text{ mb}} \right) \theta_{600} - \theta_{600}}{200 \text{ hPa}} \)

V0Z0 > \( -V_0 \cdot \nabla Z_0 \)

VOR0 > \( Z_0 \)

VERI > \( \left( \frac{dZ_0}{dt} \right)_{obr} \) > \( \frac{(Z_0)_{/0} - (Z_0)_{/1}}{43,200 \text{s}} \)
#! /bin/csh
## Computes 4 forcing terms of P-S development equation
set gridf1 = '99030400.gem'
set gridf2 = '99030312.gem'
set gridf3 = '99030400.gem + 99030312.gem'
set gtime1 = 'f00'
gddiag << endgddiag
GDFILE   = $gridf1
GDOUTF   = $gridf1
GFUNC    = mul(1,avg(dden(1000,tvrk@1000),dden(900,tvrk@900)))
GDATTIM  = $gtime1
GLEVEL   = 600:1000
GVCORD   = pres
GRDNAM   = den1
GPACK    = r
gfunc=mul(1,avg(dden(600,tvrk@600),dden(900,tvrk@900)))
grdnam=den2
r
gfunc=mul(1,quo(quo(quo(add(9.8,stab@900:1000),den1),gravty),1000))
grdnam=gam1
r
gfunc=mul(1,add(1,quo(quo(mul(2500000,mixs@900),287),tmpk@900)))
grdnam=a900
r
gfunc=mul(1,quo(quo(quo(mul(6.25e12,mixs@900),462844),tmpk@900),tmpk@900))
grdnam=b900
r
gfunc=mul(1,quo(mul(9.8,a900),add(1,b900))
grdnam=g900
r
gfunc=mul(1,add(1,quo(quo(mul(2500000,mixs@600),287),tmpk@600)))
grdnam=a600
r
gfunc=mul(1,quo(quo(quo(mul(6.25e12,mixs@600),462844),tmpk@600),tmpk@600))
grdnam=b600
r
gfunc=mul(1,quo(mul(9.8,a600),add(1,b600))
grdnam=g600
r
gfunc=mul(1,avg(g600,g900))
grdnam=gsat
r
gfunc=mul(1,quo(quo(quo(add(gsat,stab@600:900),den2),gravty),1000))
grdnam=gam2
r
gfunc=mul(1,lav(omeg@1000:900))
grdnam=omv1
r
gfunc=mul(1,lav(omeg@600:900))
grdnam=omv2
r
gfunc=mul(1,mul(.25,mul(mul(gam1,omv1),100)))
grdnam=olr1
r
gfunc=mul(1,mul(.75,mul(mul(gam2,omv2),100)))
grdnam=olr2
r
gfunc=mul(1,add(olr1,olr2))
grdnam=olrv
r
gfunc=mul(1,sm9s(sm9s(olrv)))
grdnam=stb1
r
gfunc=mul(1,lav(thta@1000:800))
grdnam=tav1
r
gfunc=mul(1,lav(thta@600:800))
grdnam=tav2
r
gfunc=mul(1,avg(tav1,tav2))
grdnam=tavg
r
gfunc=mul(1,avor@1000)
grdnam=vor0
r
GDFILE = gridf2
gdout=gridf2

gfunc=mul(1,lav(thta@1000:800))
grdnam=tav1
r

gfunc=mul(1,lav(thta@600:800))
grdnam=tav2
r

gfunc=mul(1,avg(tav1,tav2))
grdnam=tavg
r

gfunc=mul(1,avor@1000)
grdnam=vor0
r

GDFILE=gridf3
gdout=gridf1

gfunc=mul(1,quo(sub(tavg^f00,tavg^f00+2),43200))
grdnam=dta
r

gfunc=mul(1,quo(sub(vor0^f00,vor0^f00+2),43200))
grdnam=veri
r

GDFILE=gridf1
gdout=gridf1

gfunc=mul(-1,avg(adv(thta@1000,wnd@1000),adv(thta@800,wnd@800)))
grdnam=dtb1
r

gfunc=mul(-1,avg(adv(thta@600,wnd@600),adv(thta@800,wnd@800)))
grdnam=dtb2
r

gfunc=mul(1,avg(dtb1,dtb2))
grdnam=dtb
r

gfunc=mul(1,mul(quo(ldf(thta@1000:800),200),lav(omeg@1000:800)))
grdnam=dtc1
r

gfunc=mul(1,mul(quo(ldf(thta@600:800),-200),lav(omeg@600:800)))
grdnam=dtc2
r

gfunc=mul(1,avg(dtc1,dtc2))
grdnam=dtc
r
gfunc=add(dta,dtb)
grdnam=sum1

r
gfunc=add(sum1,dtc)
grdnam=dt.dt

r
gfunc=mul(1004,dt.dt)
grdnam=dwdt

r
gfunc=mul(1,sm9s(sm9s(dwdt)))
grdnam=dbh1

r

GDFILE   = $gridf1
GOUTF   = $gridf1
GDATIM   = $gtime1
GLEVEL   = 600:1000
GVCORD   = pres
GPACK    =

gfunc=quo(mul(-146.6,lap(stb1)),corl)
grdnam=t3

r
gfunc=quo(mul(-.1460,lap(dbh1),corl)
grdnam=t4

r
gfunc=mul(1,adv(avor(wnd@600),wnd@600))
grdnam=t1

r
gfunc=mul(1,avg(adv(ldf(hght),wnd@1000),adv(ldf(hght),wnd@800)))
grdnam=adv1

r
gfunc=mul(1,avg(adv(ldf(hght),wnd@600),adv(ldf(hght),wnd@800)))
grdnam=adv2

r
gfunc=mul(1,avg(adv1,adv2))
grdnam=adv

r
gfunc=mul(-1,quo(mul(gravty,lap(adv)),corl))
grdnam=t2

r
gfunc=mul(1,adv(avor(wnd@1000),wnd@1000))
grdnam=v0z0

r
gfunc=add(t1,t2)
grdnam=t1t2
r

gfunc=add(t1t2,t3)
grdnam=t123
r

gfunc=add(t123,t4)
grdnam=all4
r

gfunc=add(all4,v0z0)
grdnam=dzdt
r

ex
endgddiag
gpend
REFERENCES


William was born on December 27, 1969 in Wilmington, DE. He received a Bachelor of Arts for Geography from the University of Delaware in 1997. He moved to Tallahassee, FL the same year to pursue a Master’s degree in meteorology at the Florida State University. He is currently teaching middle school science in Tampa, FL.