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ABSTRACT

The primary goal of this study is to examine the formation mechanisms for the intense North American anticyclone that developed in late January and early February 1989 and to assess the relative importance of adiabatic versus diabatic processes. The complete height tendency equation is used as the diagnostic tool in this study. Results of this analysis show that diabatic processes are relatively unimportant and that the principal forcing mechanisms in the anticylogenesis are vorticity advection and differential thermal advection. The cold low-level air over Alaska enhances the anticylogenesis by promoting a positive contribution from the differential thermal advection. The cold low-level temperatures preclude strong cold-air advection at low levels in the anticyclone; strong upper-level cold-air advection thus results in a large positive contribution from the differential thermal advection. The vertical advection of static stability opposes the other forcings, acting to slow the development. The ageostrophic vorticity tendency term makes a substantial contribution; during some portions of the anticylogenesis, the ageostrophic vorticity tendency enhances the anticylogenesis and at other times opposes the development. A comparison is made with the model results of a purely cold-core anticyclone forced solely by radiative cooling. It is determined that the forcing mechanisms are substantially different for the intense North American anticyclone when compared with the cold-core anticyclone driven by radiative cooling, even though the thermodynamic structures of the two anticyclones are similar.

1. Introduction

For decades, the development and evolution of anticyclones have received considerably less attention than that of cyclones. One reason for this relative neglect may be that anticyclones are usually regarded as "fair weather" systems, with most of the interest reserved for storms associated with severe weather. In this paper we describe an example of a North American anticyclone that was associated with very severe weather and discuss the evolution of this system.

During late January and early February of 1989, an intense anticyclone formed in Alaska. Surface pressures reached a North American record of 1078 hPa in Alaska, and the Federal Aviation Administration banned night and instrument flights in Fairbanks because altimeters could not be accurately calibrated to give altitude readings. Ice fog was so severe in Fairbanks that it was dangerous to drive. As the anticyclone moved southward, cold temperature records were set from Alaska to Alabama, with below-freezing temperatures all the way down to the Gulf of Mexico. Rare snowfalls occurred in southern California. According to newspaper reports, the cold weather was blamed for over 20 deaths, substantial crop damage in California and Texas, and the shutting down of roads, schools, and government offices.

In the early meteorological literature (summarized by Wexler 1951), anticyclones have been classified as cold, warm, or mixed. Cold anticyclones are associated with cold, dense air near the surface and a warm, low stratosphere. Low-level radiative cooling has been postulated to force the cold anticylogenesis. Warm anticyclones are characterized by a warm, deep troposphere and a cold, high stratosphere. The cold temperature axis in a warm anticyclone typically tilts west or northwest with height. Warm anticyclones have been hypothesized to be driven by upper-level cold-air advection and by upper-level convergence. A mixed anticyclone is characterized by a cold surface and a deep troposphere, with both low-level diabatic processes and upper-level dynamics being important in their formation. Mixed anticyclones may result from the transformation of a cold anticyclone as it moves over a warm
surface. Alaskan anticyclones in particular have been the focus of papers by Bodurtha (1952) and Bowling et al. (1968).

Several recent studies have increased our understanding of the physical mechanisms occurring in anticyclones. Using a numerical model, Curry (1983, 1987) addressed the problem of the formation of cold anticyclones and showed that the radiative cooling from condensate in the atmosphere (principally low-level ice crystals that form in the cooling air) makes an important contribution to the formation of low temperatures that drives the anticyclogenesis. A positive feedback loop was shown to exist between the formation of condensate in the cooling air and the anticyclogenesis, whereby radiative cooling from the condensate enhances anticyclogenesis and the large-scale meridional circulation associated with anticyclone replenishes the moisture in the layer of condensate, thus enhancing the radiative cooling. Toward improving our understanding of the formation of warm and mixed anticyclones, studies by Dallavalle and Bosart (1975) and Boyle and Bosart (1983) have shown the importance of strong negative vorticity advection at upper levels in producing upper-level convergence above the surface anticyclone. By using the quasigeostrophic height tendency equation, Boyle and Bosart (1983) found that the differential vorticity advection is the principal mechanism for transforming the cold anticyclone into a warm high.

The purpose of this study is to document the formation and evolution of the intense North American anticyclone of winter 1989 and to assess the relative importance of diabatic and adiabatic processes. Heat budgets for this period have been presented by Tanaka and Milkovich (1990). The complete height tendency equation developed by Tsou et al. (1987) is used as the diagnostic tool in this study.

\[
\begin{align*}
\nabla^2 + \left( \frac{\xi + f}{\sigma} \right) & \frac{\partial^2}{\partial p^2} \frac{\partial \Phi}{\partial t} - \frac{1}{f} \nabla f \cdot \nabla \frac{\partial \Phi}{\partial t} + f \frac{\partial \xi}{\partial t} \\
& = -f \nabla \cdot \nabla (\xi + f) + \frac{fR(\xi + f)}{\sigma} \frac{\partial}{\partial p} \left( \frac{1}{p} \nabla \cdot \nabla T \right) - \frac{fR(\xi + f)}{\sigma} \frac{\partial}{\partial p} \left( \frac{q}{p} \right) - \frac{f(\xi + f)}{\sigma} \frac{\partial q}{\partial p} \\
& - f \left( \frac{\partial q}{\partial x} - \frac{\partial q}{\partial y} \right) - f \frac{\omega}{\partial p} \frac{\partial \xi}{\partial p} + f k \cdot \nabla \times F. 
\end{align*}
\]

The components in (1) are described as follows:

<table>
<thead>
<tr>
<th>Term</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>A'</td>
<td>Laplacian of geopotential tendency</td>
</tr>
<tr>
<td>A''</td>
<td>(\beta) term</td>
</tr>
<tr>
<td>A'''</td>
<td>Ageostrophic vorticity tendency</td>
</tr>
<tr>
<td>B</td>
<td>Vorticity advection</td>
</tr>
<tr>
<td>C</td>
<td>Vertical differential thermal advection</td>
</tr>
<tr>
<td>D</td>
<td>Vertical differential heating</td>
</tr>
<tr>
<td>E</td>
<td>Vertical advection of static stability</td>
</tr>
<tr>
<td>F</td>
<td>Tilting effects</td>
</tr>
<tr>
<td>G</td>
<td>Vertical advection of vorticity</td>
</tr>
<tr>
<td>H</td>
<td>Frictional effects</td>
</tr>
</tbody>
</table>
In Tsou et al. (1987), a scale analysis showed that term $A''$, the ageostrophic vorticity tendency term, was small with respect to the larger terms in (1), and was thus neglected. Since we are using the complete height-tendency equation and wish to make no a priori assumptions about the magnitudes of terms for the anticyclone, the ageostrophic vorticity tendency term must be evaluated. In Tsou and Smith (1990), the ageostrophic vorticity tendency was evaluated as $\Delta \zeta_{ag}/\Delta t$. Here we propose an alternative method to evaluate the ageostrophic vorticity tendency, where the ageostrophic vorticity tendency term ($A''$) is determined from the difference between the total vorticity tendency and the geostrophic vorticity tendency:

$$\frac{\partial \zeta_{ag}}{\partial t} = -\mathbf{V}_g \cdot \nabla \zeta_{ag} - \mathbf{V}_{ag} \cdot \nabla (f + \zeta) \quad \text{(A)}$$

$$+ (f - f_0 + \zeta)^2 \frac{\partial}{\partial p} \left( \frac{\omega}{f - f_0 + \zeta} \right) \quad \text{(B)}$$

$$- \left( \frac{\partial \omega}{\partial x} \frac{\partial v}{\partial p} - \frac{\partial \omega}{\partial y} \frac{\partial u}{\partial p} \right) + \mathbf{k} \cdot \nabla \times \mathbf{F}. \quad \text{(C)}$$

Equation (2) can then be used to determine term $A''$ in (1).

If every term on the right-hand side of the complete height tendency equation (1) can be obtained from observational data or from parameterization schemes, we can determine the height tendency by inverting the left-hand side of the equation. Terms $D$ and $H$, the diabatic and friction terms in (1), respectively, require a parameterization for their evaluation. The method used for computing the friction term ($H$) in (1) is similar to those applied by Manabe et al. (1965) and Smagorinsky et al. (1965), which is basically a mixing-length formulation that includes a dependence on atmospheric stability. The diabatic-warming term ($D$) in (1) includes radiation, latent, and sensible heating effects. The radiative heating is calculated by using the radiation parameterization described by Harshvardhan et al. (1987). Both the terrestrial and solar radiation are considered in this scheme. Water vapor, carbon dioxide, and ozone are the gaseous absorbers for longwave radiation and water vapor and ozone are the gaseous absorbers for shortwave radiation. The radiative effects of clouds are also included. Surface albedo is parameterized after Marchuk et al. (1979) to include the effects of variable snow depth. The latent heat released at each level is the summation of convective latent heat release and stable latent heat release, which are calculated following Edmon and Vincent (1976). The sensible heat flux is computed after Manabe et al. (1965) and Smagorinsky et al. (1965).

In order to solve (1), the equation must be elliptic, that is, $(f + f)/\sigma$ must be positive. If total vorticity is negative, we let $(f + f)/\sigma$ equal a small number ($5.0 \times 10^{-3} \text{ m}^2 \text{ s}^{-4} \text{ kg}^{-1}$). If $\sigma$ is negative and total vorticity is positive, we let $\sigma$ equal the smallest positive static stability parameter in the analysis domain. The sequential overrelaxation method is used to solve the extended height tendency equation. The iteration is terminated when the difference between two successive solutions is less than $1.0 \times 10^{-12} \text{ m}^2 \text{ s}^{-3}$. To solve the height tendency equation, we interpolate and extrapolate the data to 21 layers with 50-hPa intervals from 1050 to 50 hPa. For numerical simplicity, solutions are obtained even when the surface pressure is lower than 1050 hPa; however, values at grid points below the surface are not included in the analysis.

The boundary conditions utilized to solve the height tendency equation must be continuous and regular. A zero boundary condition is commonly used at the boundaries to satisfy this requirement. The height tendency, however, particularly at the lower boundary, may be substantially different from zero especially during the anticyclonicogenesis period. An improved lower boundary condition is described by Tsou et al. (1987), which consists of applying the relaxation scheme twice. The zero boundary condition is used for the first iteration, and then the calculated height tendencies of some higher level obtained from the first iteration are assigned to the lower boundary for the second iteration. We follow this approach, and the choice of the level to use for the lower boundary condition in the second iteration is guided by comparing the calculated with the observed ($\Delta Z/\Delta t$) height tendencies for different choices of levels. The correlation coefficients are found to be highest when the 700-hPa level is chosen for the lower boundary condition in the second iteration. Because of the terrain in this region, 700 hPa is the lowest level that is entirely above the surface. The choice of the lower boundary condition affects the results only for levels below 700 hPa. In keeping with the guidelines used by Tsou et al. (1987), we let the 700-hPa results from the first time relaxation be the lateral and the bottom boundary conditions for the second relaxation. The upper boundary is chosen to be 50 hPa, with a value of zero used during both relaxations.

3. Data and synoptic overview

A general overview of the large-scale circulation patterns in the Alaska region during January and February 1989 is given by Tanaka and Mikkovich (1990). The period of interest for this study is from 0000 UTC 23 January to 1200 UTC 7 February 1989. The data domain is from $20^\circ$ to $90^\circ$N and $142.5^\circ$ to $52.5^\circ$W, which includes part of the Pacific Ocean and most of North America. Upper-air data that are utilized in this study are the European Centre for Medium-Range Weather Forecasts (ECMWF)–World Meteorological Organi-
zation (WMO) twice-daily (0000 and 1200 UTC) global analyses, as obtained from the National Center for Atmospheric Research (NCAR). Surface temperatures are obtained from the National Meteorological Center (NMC) analyses. Cloud, fog, and snow cover data are obtained from the Techniques Development Laboratory (TDL) dataset at NCAR.

a. Data

The ECMWF-WMO global analysis data include horizontal and vertical winds, temperature, humidity, and geopotential height at levels of 1000, 850, 700, 500, 300, 200, and 100 hPa on a 2.5° × 2.5° latitude-longitude grid. The dataset is generated by a four-dimensional assimilation system that uses a three-dimensional multivariate optimum interpolation analysis and automatic data checking. A nonlinear normal-mode initialization scheme is then applied to eliminate high-speed gravity waves. Finally, a numerical model produces a first guess for the subsequent analysis. A description of the assimilation and analysis schemes are described by Lorenc (1981), Bengtsson et al. (1982), and Hollingsworth et al. (1985). Recent revisions in the mass and wind analysis are discussed by Shaw et al. (1987), in which the analysis and the first guess are in better agreement with the observation. Diabatic initialization is performed after Wergen (1988). Tiede et al. (1988) describe the recently modified cloud and convection schemes.

Because surface data was unavailable from ECMWF, the surface data used in this study is produced by NMC’s Global Data Assimilation System. A detailed description of the system is given by McPherson et al. (1979), Dey (1989), and Kanamitsu (1989). The system uses a 6-h intermittent assimilation method, which takes a 6-h forecast from the previous analysis as the first guess and is applied on the model sigma levels. The surface temperature and pressure from this analysis are employed in this study. It is noted here that the use of separate analyses for upper air and the surface has little impact on the evaluation of the height tendency since the surface data is utilized only in the radiation calculations.

The cloud, fog, and snow cover data are obtained from the TDL dataset, comprised of hourly reports from United States and Canadian stations. These data are used in this study for the radiation calculations. The grid-scale cloud fraction is determined from the TDL data where available. When there are TDL observations in the 2.5° × 2.5° latitude-longitude region, the average observed cloud fraction value is assigned to that point. The observed cloud amounts are distributed vertically in a grid box by assigning low, middle, and high cloud amounts to the appropriate pressure level based on consideration of the sea level pressure. For grid boxes where there are no TDL cloud observations (e.g., over the ocean), cloud fraction is estimated from the layer relative humidities after Mitchell and Hahn (1990) as applied to the ECMWF analyses by Sheu and Curry (1992). Reports of fog are assigned as cloud at the level just above the surface. To account for lower-tropospheric ice crystals which are not included in surface cloud observations (Curry et al. 1990), we assume that low-level ice crystals are present at temperatures less than −15°C and at relative humidities exceeding ice saturation.

b. Synoptic overview

The investigation of the anticyclonic evolution is focused on the period from 0000 UTC 26 January to 1200 UTC 7 February in 1989. During this period the anticyclone moved from the Chukotsk Highland to Alaska, became stationary in Alaska and northwest Canada for several days, and then turned southeastward toward the United States. The evolution of the height field for the anticyclone throughout the period is shown in Fig. 1, where a time series (every 12 h) of the average heights at 850, 500, and 300 hPa are given in the anticyclonic region as defined by Table 1. The evolution of the anticyclone is divided here into four stages, with a plot shown for each of the four stages of surface pressure (Fig. 2), 850-hPa temperatures (Fig. 3), and 300-hPa heights (Fig. 4).

Stage 1 comprises the period from 0000 UTC 26 January to 1200 UTC 27 January, during which the relatively weak surface anticyclone slowly moved eastward from the Chukotsk Highland toward the Bering Sea. Prior to this period, the anticyclone had been relatively weak and approximately stationary for several days. At the beginning of this stage, the maximum sea level pressure in the anticyclone over Chukotsk is 1026 hPa, which is associated with a surface temperature minimum. A ridge is apparent over the surface anticyclone at levels as high as 500 hPa. The tropopause level is at 300 hPa over Chukotsk and stratospheric temperatures are relatively warm. The above features are typical of a cold-core anticyclone, with the exception of this anticyclone being somewhat deeper than is typical of cold-core anticyclones. At the same time, a weak and shallow but relatively warm surface anticyclone can be seen in the midlatitudes of the eastern Pacific Ocean.

Stage 2 covers the period from 0000 UTC 28 January to 0000 UTC 30 January. During this period, the cold anticyclone moves eastward from Chukotsk Highland toward the ocean and merges with the Pacific anticyclone, forming an elongated anticyclone with increasing horizontal extent. During this period low-level temperatures warm from the very cold values of the Chukotsk anticyclone as the lower levels are modified by the warm underlying ocean. The geopotential height at 1000 hPa increased 111 m in this 48-h period. Large
Fig. 1. Time series (every 12 h) of the average heights at 850, 500, and 300 hPa in the anticyclonic region as defined by Table 1.
values of subsiding vertical motion and negative vorticity are observed throughout the troposphere (up to 300 hPa), showing the deep development of the anticyclone. A trough at 200 hPa is evident directly above the surface anticyclone. The tropopause above the high pressure system is at 200 hPa and the stratosphere is relatively warm. Both the cold temperature and ridge axes tilt westward with height. By the end of this period, the anticyclone possesses most of the characteristics of a warm anticyclone. During this period the anticyclone continues to migrate toward the west coast of North America and at the end of this period the easternmost edge of the surface anticyclone reaches Alaska. The lowest surface temperature in Alaska is 217 K at the end of this stage.

Stage 3 comprises the period between 1200 UTC 30 January and 1200 UTC 3 February. During this period, the anticyclone center moves into Alaska and becomes stationary at the border between Alaska and Canada. The anticyclone intensifies; sea level pressure reaches a maximum value of 1078 hPa and the geopotential height at 1000 hPa reaches its maximum value of 502 m at the end of this period. The surface anticyclone is shallower than in stage 2, being evident only up to 500 hPa, since the ridge tilts westward with height. The upper-level ridge remains stationary, forming a blocking pattern. Strong downward motion is present over the center of the surface anticyclone. Lower-tropospheric temperatures are cold in the anticyclone. At the beginning of this stage, the lower-tropospheric cold center breaks into two parts, one center staying in western Alaska while the other moves southward toward the United States. By the end of this stage, this second cold center is found in northern Montana. During this period the lower-tropospheric temperatures increase over Alaska. The center of the anticyclone remains over the low-level temperature minimum in Alaska. At the end of the period the upper-level ridge moves over the surface anticyclone.

Stage 4 is the final time period, from 0000 UTC 4 February until 1200 UTC 7 February. The high pressure system leaves Alaska and slowly migrates southward along the west coast of Canada. Concurrently, the surface anticyclone decays as the upper-level ridge remains over the surface high. The magnitude of the divergence and vertical motions also decreases during this stage. The tropopause is still around 200 hPa. Low-level temperatures in the anticyclone continue to increase, although record low temperatures occur in the southwestern United States.

4. Height-tendency results

To assess the performance of the height tendency equation in diagnosing the life cycle of the anticyclone, we compare the height tendency calculated from (1) with the observed height tendencies determined from (\( \Delta Z / \Delta t \)). In presenting this comparison, we have chosen one analysis period from each of the four stages, focusing on the region of the anticyclone as delineated in Table 1. This is shown in Fig. 5 by comparing the observed with the calculated vertical profiles of mean height tendencies and by evaluating the vertical profiles of the linear correlation coefficients between the calculated and observed height tendencies. In Fig. 6 we show maps of both the observed and calculated height tendencies for 0000 UTC 31 January (stage 3).

As seen from Fig. 5, the correlation coefficients between the observed and calculated height tendencies are positive and very high at all levels except for the stage 1 profile, where the correlations are positive but not as large as for the other stages. A comparison of the vertical profiles of observed and calculated mean height tendencies shows that the magnitude of the calculated height tendencies is systematically larger than the observed height tendency. In evaluating the comparisons between the calculated and observed height tendencies, it is important to keep in mind that the height tendencies calculated from (1) are instantaneous height tendencies, while the observed height tendencies represent averages over a 24-h period. Thus we would expect the magnitudes of the calculated height tendencies to be higher than the observed tendencies. Also, it is not possible to get a perfect correlation between the observed and calculated height tendencies because of the nonlinearity of the height tendency over a 24-h
Fig. 2. Maps of mean sea level pressure during the four stages of anticyclone evolution: (a) stage 1 (0000 UTC 12 January 1989); (b) stage 2 (0000 UTC 12 January 1989); (c) stage 3 (0000 UTC 1 February); (d) stage 4 (0000 UTC 5 February). Isobathic interval is 8 mb.
Fig. 3. Same as Fig. 2 except for 850-hPa temperatures. Isopleth interval is 10 K.
A general assessment of the magnitude of each forcing term in (1) around the anticyclonic region (as delineated in Table 1) is shown in Table 2. We see that the tilting term (F) and the vertical advection of vorticity term (G) in (1) are at least one order of magnitude smaller at all levels than the largest terms. Tsou et al. (1987) also found very small contributions arising from terms F and G. The friction term (H) is seen to be negligibly small at all levels above 950 hPa; at the lowest levels the magnitude of H is of the same order as the largest terms. The diabatic term (D) is shown to be important at the lower levels but diminishing in importance above 600 hPa. The vorticity advection term (B), the vertical differential thermal advection term (C), the ageostrophic vorticity tendency term (A"), and the vertical advection of static stability term (E) are relatively large at all levels. It is noted here that the ageostrophic vorticity tendency term (A"), which is seen to be one of the largest terms, was neglected by Tsou et al. (1987).

a. Effects of individual forcings

The contribution to the height tendency from each of the forcing terms in (1) can be determined by solving (1) for each individual forcing term. In the following discussion of the individual forcing effects, we will neglect the tilting term (F), the vertical advection of vorticity term (G), and the friction term (H), and focus on the terms that make the largest contribution. Time series of the horizontally averaged values of the individual forcing terms around the anticyclonic region (see Table 1) at 300, 500, and 850 hPa are illustrated in Fig. 7.

From the height tendency equation, it is seen that negative vorticity advection (term B) causes the height to increase. It is seen from Fig. 7 that vorticity advection contributes to height increases at all three pressure levels throughout the anticyclone life cycle (as described by Figs. 1–4). The vorticity advection makes its largest relative contribution to height increases during stage 2 when the high-level ridge gradually intensifies on the west side of the anticyclone and causes larger vorticity advection in the anticyclonic region and during stage 3 when the upper-level ridge to the west of the surface anticyclone becomes stronger. After the anticyclone moves to Alaska, both the wind speed and the upper-level ridge become stronger. During stage 4, the upper-level ridge moves over the surface anticyclone and the height tendency gradually decreases to nearly zero. The magnitude of the contribution of vorticity advection to the height tendency is largest in the upper troposphere.

From Fig. 7, we can see that the effect of the vertical differential thermal advection, term C in (1), has a similar pattern of height tendency as does vorticity advection. Basically, if cold-air advection increases with
height in the vicinity of a pressure surface, the height of that pressure surface will increase. The magnitude of this term, particularly during stage 3, indicates strong baroclinicity. During stage 1, the weak warm-air advection at low levels and cold-air advection above contributes to a slow height increase at all three levels. During stage 2, the low-level temperature advection is still relatively weak while at higher levels there is strong cold advection from the west side where the ridge is intensifying and the temperature is relatively low. After the anticyclone moves to Alaska (stage 3), the wind from the northwest side brings strong cold-air advection at 300 hPa. Because of the extremely cold temperatures at low levels over Alaska, there is no cold-air advection at low levels. As a result, strong vertical differential thermal advection becomes one of the most important factors causing the height increase during this period. During stage 4, the contribution from the vertical differential thermal advection becomes increasing smaller, with slight negative values at 850 hPa occurring by the end of this stage.

The effect of vertical advection of static stability, term (E) in (1), can be explained by using the thermodynamic equation in thickness form (Smith and Tsou 1988):

$$\frac{\partial}{\partial t} \left( - \frac{\partial \Phi}{\partial p} \right) = - \frac{R}{p} \nabla \cdot T + \frac{R \dot{q}}{p c_p} + \omega \sigma. \quad (3)$$

This equation shows that for a given value of $\omega$, larger thickness change can occur in an atmosphere that has a large static stability. Since throughout most of the
troposphere we have large static stability at high levels and smaller static stability at low levels, downward motion at a specific layer will cause the thickness of the atmosphere below the layer to decrease more than the thickness of the atmosphere above it. In the anticyclonic region, the downward vertical motion causes compression of the air below and thus a height decrease, the contribution to the height decrease being largest during stage 3 when the subsiding velocity was at maximum. Although the subsiding vertical velocities diminish during stage 4, the vertical advection of the static stability term dominates the other terms at 850 hPa, resulting in the decay of the anticyclone.

Although the contribution of ageostrophic vorticity tendency term \((A^{v})\) to the height tendency has typically been neglected, it is seen from Table 2 and Fig. 7 that this term is of similar magnitude to the dominant terms. By analyzing the component terms in (2), it is found that the most important factors determining the ageostrophic vorticity tendency are the ageostrophic vorticity adected by geostrophic wind (term A), the total vorticity adected by the ageostrophic wind (term B), and term C, which is a combination of divergence and vertical advection of relative vorticity effects. For the most part, the ageostrophic vorticity tendency makes a positive contribution to the height tendency, although during stage 2 a negative contribution to the height tendency is determined for all three levels shown in Fig. 7. During the period 28–30 January, a period of rapid increase in height at 850, 500, and 300 hPa (see Fig. 1), the ageostrophic vorticity tendency makes a negative contribution to the height, acting to diminish the anticyclogenesis. This is consistent with the occurrence of geostrophic adjustment to the mass distur-
Table 2. Average absolute value of each term (s⁻¹) in the height tendency equation (1), averaged over the anticyclone region as delineated in Table 1 for the entire period (0000 UTC 26 January–1200 UTC 7 February 1989).

<table>
<thead>
<tr>
<th>Level (mb)</th>
<th>Term B</th>
<th>Term C</th>
<th>Term D</th>
<th>Term E</th>
<th>Term F</th>
<th>Term G</th>
<th>Term H</th>
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b. Influence of the total forcing

Throughout stage 1, the height increase is small at all levels (Fig. 1), the dominant forcing mechanisms (Fig. 7) being vertical differential thermal advection (C) and ageostrophic-vorticity tendency (A⁺⁺). During stage 2, when the anticyclone moves over the Pacific Ocean, the height increases rapidly, most notably at the upper levels. The forcing during stage 2 is complex; toward the beginning of this stage increasing negative vorticity advection becomes dominant. The vertical differential heating term increases at all levels during this stage, caused principally by latent heating in the midtroposphere. Toward the end of the second stage the height increase is very sharp, particularly at 500 and 300 hPa; this is principally associated with the decrease in static stability that results in a decrease in the damping associated with the vertical advection of static stability. This results in a strong ridge over the Pacific Ocean. At the onset of stage 3, the anticyclone moves over Alaska where there is a preexisting pool of low-level cold air. Heights decrease at the upper two levels during the beginning of this stage (see Fig. 1). The upper-level ridge over the Pacific contributes to strong cold-air advection at upper levels over the surface anticyclone and the low-level cold air precludes large cold-air advection at low levels. The resulting vertical gradient in cold-air advection results in strong forcing of the anticyclone due to the differential vertical thermal advection term (B). Both the vorticity tendency (A) and the ageostrophic vorticity tendency (A⁺⁺) also provide substantial forcing of the anticyclone during this stage. As a result of the temperature decreases throughout the troposphere, the troposphere becomes compressed with a shallow and relatively warm tropopause. At this stage the anticyclone possesses all of the characteristics
c. Comparison with development of cold-core anticyclone

The height tendency equation (1) is applied to the gridded output of the cold-core anticyclone from the modelling study of Curry (1987). In summary, Curry (1987) used a nonlinear axisymmetric numerical model, with a radial model domain of 2000 km consisting of a central ice pack with radius 1000 km that is surrounded by open water. The initial conditions of

![Diagram](image-url)

**FIG. 7.** Time series of the horizontally averaged values of the individual forcing terms in the height tendency equation (m day$^{-1}$) around the anticyclonic region as defined by Table 1 at 300, 500, and 850 hPa. Terms are illustrated as: —— vorticity advection; --- vertical differential thermal advection; —— vertical differential heating; ····· vertical advection of static stability; ······ ageostrophic vorticity tendency.

of a cold anticyclone, even though low-level radiative cooling plays a very minor role in the intensification of the anticyclone.

During stage 4 the heights decrease at all levels, the most rapid decrease occurring at 850 hPa. The upper-level ridge overtakes the surface anticyclone, resulting in the diminution of the forcing at all levels. The decay of the surface anticyclone results directly from the vertical advection of static stability (E).

![Diagram](image-url)

**FIG. 8.** Vertical profiles of contributions to the diabatic $q$ (K day$^{-1}$) for (a) stage 2 and (b) stage 3. Terms are illustrated as: —— total heating; --- shortwave radiative heating; —— longwave radiative heating; ····· latent heating.
this calculation were that there are no winds and no horizontal temperature gradients. The anticyclone developed as the air over the ice packs cooled radiatively. After an integration period of 5 days, the central surface pressure had increased by 10 hPa and low-level anticyclonic velocities reached 10 m s\(^{-1}\) near the ice edge.

Results of the height tendency analysis for this cold-core anticyclone show that the dominant forcing mechanism of the low-level anticyclone is the vertical differential heating term (D) associated with the infrared radiative cooling. At higher levels, forcing principally due to vertical differential heating (D), and thermal advection (C), and the vertical advection of static stability (E) result in height decrease. In the early period of the anticylogenesis, the low-level forcing is in part counteracted by the ageostrophic vorticity tendency (A\(^{\prime\prime}\)).

The genesis of the intense North American anticyclone that is examined in this paper clearly has different forcing mechanisms than the archetypal cold-core anticyclone.

d. Comparison with cyclonic development

It is instructive to compare the development of the intense North American anticyclone in the context of the height tendency equation with the evolution of an intense extratropical cyclone system (Tsou et al. 1987). At the most obvious level, a comparison of the results for the anticyclone case with the cyclone case shows that for the most part the forcing terms are of opposite sign. Apart from this obvious difference, the following additional differences are noted. For the cyclone development, the vorticity advection (term B) was the dominant term forcing the height decrease, with the thermal advection term (term C) having a magnitude that was typically less by a factor of a half. For the anticyclone development, the thermal advection term is of equal magnitude to the vorticity advection, and during most of stage 3 the thermal advection term dominates. The ageostrophic vorticity tendency (term A\(^{\prime\prime}\)) was neglected in the Tsou et al. study; for the anticyclone case this term was shown to be of considerable importance. In both cases the vertical advection of static stability (term D) counteracted the height increase (decrease) for the anticyclone (cyclone), putting a brake on the system development. For the anticyclone the magnitude of the static stability term was relatively larger than for the cyclone. The vertical differential heating term was relatively small for both the anticyclone and the cyclone development; for the anticyclone the dominant heating mechanism was radiative cooling while for the cyclone the dominant heating mechanism was latent heat release. Overall, the vertical differential heating in the cyclone made a larger direct contribution to the intensification of the system than did radiative cooling for the anticyclone.

5. Summary and conclusions

The intense anticyclone observed during winter 1989 was examined using the complete height tendency equation to determine the factors contributing to the anticylogenesis. The results of the height tendency equation showed that the most important forcings contributing to the anticylogenesis were the vorticity advection and the vertical differential thermal advection. The ageostrophic vorticity tendency was a dominant term, in some stages contributing to the anticylogenesis and at other stages inhibiting the anticylogenesis. The vertical advection of static stability acted in opposition to the anticylogenesis.

Diabatic processes were shown to be relatively unimportant in the anticylogenesis. This is a somewhat surprising result, since the formation of cold-core anticyclones is typically associated with radiative cooling. To elucidate this result, a comparison was made with the model results of a purely cold-core anticyclone forced solely by radiative cooling. It was determined that the forcing mechanisms are substantially different for the intense North American anticyclone when compared with the cold-core anticyclone driven by radiative cooling, even though the thermodynamic structures of the two anticyclones are similar. The intensification of the anticyclone once it reached Alaska with its cold temperatures was not coincidental, however. The cold low-level air over Alaska enhanced the anticylogenesis by promoting a positive contribution from the differential thermal advection. The cold low-level temperatures precluded cold-air advection at low levels in the anticyclone; upper-level cold-air advection associated with an intense ridge over the Pacific thus resulted in a strong positive contribution from the differential thermal advection.

By using the complete height tendency equation, it was shown that the major forcings for the anticylogenesis could be determined and the results of the height tendency pattern were similar to the observed height changes during the anticylogenesis. The complete height tendency equation has been shown to be a suitable tool for examining anticylogenesis. The importance of including ageostrophic terms has been emphasized. Ageostrophic motions were significant during all stages of the anticyclone life cycle, suggesting that numerical weather prediction model initialization schemes that filter ageostrophic motions may result in poor forecasts of anticyclones. The solution of the height tendency equation was shown to be sensitive to the boundary conditions. The choice of the lower-boundary condition was complicated by the mountainous terrain; use of sigma coordinates is not a solution to the lower boundary problem since the height tendency equation is very cumbersome in sigma coordinates and difficult to interpret.

This paper has presented a case study of a very in-
tense anticyclone, with record surface pressures. This anticyclone should not be construed as a typical anticyclone, nor the formation mechanisms described as typical for anticyclogenesis. Investigation of additional anticyclone events using the methodology described here would improve our understanding of this often overlooked baroclinic phenomena. The formation of anticyclones has been shown to be very complex and this subject deserves more attention than it has received.

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