On the development of orographic cyclones

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SUMMARY

Certain characteristics of cyclone development are compared and, besides the two types (type A and type B) described by S. Petterssen and S. J. Smeye, orographic cyclone development of a third type (type C) is identified. The analysis of dynamic features, thermal structure and kinetic energy change of cyclones which developed in the lee of the Alps during the special observing period of the ALPEX support the view that orographic cyclogenesis should be considered as a specific mechanism of extratropical cyclone development.

1. INTRODUCTION

When studying various synoptic aspects of cyclone development, Petterssen and Smeye (1971) identified two types of extratropical cyclones:

(i) Type A, the well-known amplifying frontal wave, which is known to produce kinetic energy through a reduction of baroclinicity within its own domain.

(ii) Type B, initiated by a finite disturbance in the upper troposphere; its intensification is accompanied by an augmentation of the baroclinicity and the import of kinetic energy, mainly from the jet stream region.

Detailed descriptions of these two types of extratropical cyclone development will not be given here.

Evidence derived from many studies (e.g. Radinović 1965a; Speranza 1975; Buzzi and Tibaldi 1978; Illari et al. 1981; McGinley 1982) and from synoptic experience has gradually accumulated to support the view that the mechanism of lee cyclogenesis is somewhat specific and should be considered as a different type of cyclogenesis. For convenience the notation introduced by Petterssen and Smeye will be used, and we shall refer to this mechanism as type C. This categorization, it is believed, will help clarify the mechanism of cyclogenesis, since a single theory of extratropical cyclone development still does not exist.

During the last two decades, numerous case studies of orographic cyclones have revealed some of their specific features: the orographic blocking of the cold air mass; the deformation of the upper trough as it approaches the mountains; the particular evolution of the baroclinicity and vorticity, as well as the energy transformations; the various stages of development. The most comprehensive case study has been carried out by McGinley (1982), from which a clear conceptual model of orographic cyclogenesis has emerged.

If we assume that the specific features of orographic cyclones, described in most case studies, are present and pronounced in the majority of individual cases of cyclogenesis, then they should be evident in the mean fields derived from the ensemble of cyclones encountered in the ALPEX period. The main purpose of this study is to use this methodology to show that the mechanism of lee cyclogenesis is somewhat different from that responsible for types A and B.

2. STATISTICS

During the ALPEX special observing period (March-April, 1982) eight cases of

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cyclogenesis in the lee of the Alps occurred. All these cases have been briefly reviewed by Buzz and Tosi (1982), while some cases have been studied in more detail (Jansa and Ramis 1982; Dell’Osso and Tibaldi 1982; Pham 1982; Mattucks 1982; Lüdecke 1982).

Using the international ALPEX data bank, a series of hand-analysed charts for each of the eight cases of cyclogenesis has been produced. These have been used to study the statistical behaviour of the orographic cyclones in the western Mediterranean. The radiosonde and synoptic data were not subject to further quality control checks or modified prior to use in the study.

Using the criterion that the initiation of cyclogenesis takes place three hours before a closed isobar is observed, and the time of disappearance is three hours after the last closed isobar vanishes, it was found that the cyclones which occurred during the ALPEX special observing period (SOP) were relatively long lasting. Table 1 shows that the shortest duration was two days and the longest five days, while the average duration was 3.8 days.

<table>
<thead>
<tr>
<th>Case No.</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
<th>8</th>
</tr>
</thead>
<tbody>
<tr>
<td>Duration (days)</td>
<td>4.6</td>
<td>3.6</td>
<td>2.8</td>
<td>5.0</td>
<td>5.0</td>
<td>4.4</td>
<td>3.2</td>
<td>2.0</td>
</tr>
</tbody>
</table>

By following the positions of the centres of the cyclones, it is found that the majority cross the Mediterranean and disappear over the Middle East. Only in one case did a cyclone develop and decay entirely in the western Mediterranean.

There is great variation in the speed of movement of the cyclones, but during the first 24 hours of development there appears to be a characteristic behaviour, as shown in Table 2. During the six hours after cyclogenesis the cyclones are almost stationary. Thereafter they begin to move and reach a maximum speed of about 35 km h⁻¹ during the period 12-18 hours; later the speed decreases.

<table>
<thead>
<tr>
<th>Period of cyclogenesis (hours)</th>
<th>0-6</th>
<th>6-12</th>
<th>12-18</th>
<th>18-24</th>
</tr>
</thead>
<tbody>
<tr>
<td>Average distance moved (km)</td>
<td>25</td>
<td>175</td>
<td>205</td>
<td>140</td>
</tr>
</tbody>
</table>

The movement of the cyclones is closely related to the mechanism of lee cyclogenesis. During the first six hours the development is closely connected with the mountain range, but in the next six hours the cyclone develops aloft, above the mountains, and the steering effect becomes strong enough to move the cyclone with the upper flow. After 12 hours of cyclogenesis the cyclones usually develop vertically, up to the middle of the troposphere, and reach their maximum speed. Later the cyclones continue to develop vertically into the upper troposphere and expand horizontally in the lower layers. This causes a decrease of the steering effect and the cyclones slow down.

This explanation is supported by the data contained in Table 3, which shows the time lag in the vertical development. Note that the first closed contour at 850 mb occurs on average six hours after the first one appears at the surface; the lag at 300 mb is 27 hours.
TABLE 3. **Time Lag (to the nearest three hours) in the Lee Cyclone Vertical Development**

<table>
<thead>
<tr>
<th>Level (mb)</th>
<th>850</th>
<th>700</th>
<th>500</th>
<th>300</th>
<th>1000–500</th>
</tr>
</thead>
<tbody>
<tr>
<td>Time lag (hours)</td>
<td>6</td>
<td>9</td>
<td>18</td>
<td>27</td>
<td>24</td>
</tr>
</tbody>
</table>

This study has also shown that the tilt and the form of the axis of the lee cyclones vary with time. In the growing stage the axis from the surface up to the mountain top is nearly normal to the surface. This is because of the fall in the heights of the standard pressure levels, mainly caused by a surface pressure fall and the absence of cold air advection in that layer. The position of the centre of the closed contours at 500 mb and above depends primarily on the cold air advection and the horizontal advection of vorticity. Therefore the axis is usually oriented towards the centre of cold advection and is more tilted in the upper layers than in the lower layers. However, when the cyclone is moving, the axis in the lower layer becomes more tilted. In the mature stage there is hardly any tilt in the axis.

As a measure of cyclone intensity, the geostrophic circulation around the cyclone centres has been calculated (Radinović and Lalic 1959). The circulation was evaluated using the finite difference form of the Laplacian of geopotential height on a 500 km grid. Table 4 shows the intensity of all eight cyclones at five isobaric levels when the cyclones reach their maximum development.

TABLE 4. **Maximum Cyclone Intensity, as Measured by (1/\(\alpha\)) \(\nabla^2 Z (10^{2}s^{-1})\) Reached During the Cyclone Activity in the ALPEX SOP**

<table>
<thead>
<tr>
<th>Level (mb)</th>
<th>1000</th>
<th>850</th>
<th>700</th>
<th>500</th>
<th>300</th>
</tr>
</thead>
<tbody>
<tr>
<td>Case No.</td>
<td>1</td>
<td>2</td>
<td>3</td>
<td>4</td>
<td>5</td>
</tr>
<tr>
<td></td>
<td>20-8</td>
<td>12-0</td>
<td>16-8</td>
<td>11-2</td>
<td>16-0</td>
</tr>
<tr>
<td></td>
<td>24-0</td>
<td>16-0</td>
<td>17-6</td>
<td>14-4</td>
<td>16-0</td>
</tr>
<tr>
<td></td>
<td>12-8</td>
<td>12-0</td>
<td>11-2</td>
<td>6-4</td>
<td>6-4</td>
</tr>
<tr>
<td></td>
<td>14-4</td>
<td>12-0</td>
<td>10-8</td>
<td>11-2</td>
<td>11-2</td>
</tr>
<tr>
<td></td>
<td>9-2</td>
<td>11-2</td>
<td>10-4</td>
<td>16-8</td>
<td>6-6</td>
</tr>
<tr>
<td></td>
<td>12-8</td>
<td>11-2</td>
<td>14-0</td>
<td>12-4</td>
<td>19-2</td>
</tr>
<tr>
<td></td>
<td>18-0</td>
<td>13-2</td>
<td>15-2</td>
<td>26-8</td>
<td>32-8</td>
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<tr>
<td></td>
<td>15-2</td>
<td>16-0</td>
<td>16-0</td>
<td>16-0</td>
<td>16-8</td>
</tr>
</tbody>
</table>

Following the criteria described by Radinović (1965b), cyclones with a Laplacian of 16 dam/(500 km)² or vorticity of \(6.4 \times 10^{-6}s^{-1}\) are described as small low-pressure systems; those with values in the range 17–32 dam/(500 km)² corresponding to vorticity of \(6.5–12.8 \times 10^{-6}s^{-1}\) are moderate; cyclones with larger values are described as having great intensity. Table 4 shows that according to these criteria there are three moderate cyclones and five with great intensity. Furthermore, in the first four cases the greatest intensity occurs at the surface, while in the other cases the intensity is largest aloft.

A detailed analysis of these situations also reveals that in seven cases there is a lag with height of the appearance of maximum intensity—the lag at 850 and 700 mb is about six hours, and about 12 hours at 500 mb. In case 7 the reverse happens, the maximum intensity first appears aloft and then at the surface. This case was completely different from the others because there was a cold pool on the north-west slope of the Alps which formed before cyclogenesis occurred.
In order to have a better understanding of the dynamics of lee cyclogenesis, the pressure and geopotential changes in different sectors of the cyclones have been examined. The changes were measured at the centre of the cyclone (denoted by O) and at four other points which were 500 km towards the west (W), south (S), east (E) and north (N) of the cyclone centre. Two 12-hour periods from the beginning of cyclogenesis were used, the periods being taken between the nearest synoptic hours. The results for all eight cases are presented in Tables 5 and 6.

TABLE 5. AVERAGE GEOPOTENTIAL CHANGES DURING THE FIRST 12 HOURS OF CYCOGENESIS (gpm)

<table>
<thead>
<tr>
<th>Level (mb)</th>
<th>O</th>
<th>W</th>
<th>S</th>
<th>E</th>
<th>N</th>
</tr>
</thead>
<tbody>
<tr>
<td>1000</td>
<td>-64</td>
<td>34</td>
<td>-20</td>
<td>-32</td>
<td>36</td>
</tr>
<tr>
<td>850</td>
<td>-52</td>
<td>2</td>
<td>-24</td>
<td>-28</td>
<td>-10</td>
</tr>
<tr>
<td>700</td>
<td>-53</td>
<td>-16</td>
<td>-19</td>
<td>-28</td>
<td>-29</td>
</tr>
<tr>
<td>500</td>
<td>-76</td>
<td>-44</td>
<td>-19</td>
<td>-38</td>
<td>-69</td>
</tr>
</tbody>
</table>

TABLE 6. AVERAGE GEOPOTENTIAL CHANGES DURING THE PERIOD 12–24 HOURS AFTER CYCOGENESIS STARTS (gpm)

<table>
<thead>
<tr>
<th>Level (mb)</th>
<th>O</th>
<th>W</th>
<th>S</th>
<th>E</th>
<th>N</th>
</tr>
</thead>
<tbody>
<tr>
<td>1000</td>
<td>-40</td>
<td>30</td>
<td>-19</td>
<td>-28</td>
<td>45</td>
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<tr>
<td>850</td>
<td>-52</td>
<td>2</td>
<td>-30</td>
<td>-21</td>
<td>30</td>
</tr>
<tr>
<td>700</td>
<td>-68</td>
<td>-14</td>
<td>-44</td>
<td>-29</td>
<td>14</td>
</tr>
<tr>
<td>500</td>
<td>-95</td>
<td>-39</td>
<td>-65</td>
<td>-39</td>
<td>-8</td>
</tr>
</tbody>
</table>

Table 5 shows that the greatest fall of geopotential occurs at the centre of the cyclone at all levels. Also the magnitude of the surface geopotential fall at the centre is about twice the rise in the northern and western sectors, showing that the pressure fall due to cyclogenesis is more pronounced than the effect of the cold air accumulation on the windward side of the mountain barrier. Further, the greater fall of geopotential at 500 mb compared with that at 1000 mb illustrates that during the first 12 hours of cyclogenesis cold advection takes place aloft over the centre of the cyclone.

From Table 6 it may be seen that in the period 12–24 hours after the cyclone first forms, the geopotential at the centre of the cyclone is still falling. However, the fall increases with height in the troposphere, showing that in this period both cold air advection and cyclonic vorticity advection play an important role in the vertical development of the cyclone. In the western and eastern sectors of the cyclone the geopotential changes are similar in the two periods. This is not so in the southern sector, where there is a marked enhancement with time of the geopotential fall aloft and this shows that cold air and cyclonic vorticity in the middle troposphere have already moved into this region. A very marked increase in surface potential in the northern sector of the cyclone, which extends up to nearly 500 mb, indicates that the cold air advection on the windward side of the Alps decreases or stops, while pressure aloft continues to rise dynamically (i.e. due to mass convergence aloft).

An analysis of cross-sections through the cyclone in various directions has revealed some interesting features; some of these are summarized in Table 7.
TABLE 7. AVERAGE MAXIMUM WIND SPEED IN THE ZONAL AND MERIDIONAL CROSS-SECTIONS AT (a) THE TIME OF APPEARANCE OF CYCLOGENESIS AND (b) 12 HOURS LATER. ZONAL CROSS-SECTIONS MADE ALONG 47°N AND MERIDIONAL ONES ALONG 10°E

<table>
<thead>
<tr>
<th>Cross-section</th>
<th>Time</th>
<th>Direction</th>
<th>Speed (m s⁻¹)</th>
<th>Position</th>
<th>Height (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Zonal</td>
<td>(a)</td>
<td>S</td>
<td>16</td>
<td>10.5°E</td>
<td>7.5</td>
</tr>
<tr>
<td></td>
<td>(b)</td>
<td>S</td>
<td>15</td>
<td>13.0°E</td>
<td>6.8</td>
</tr>
<tr>
<td>Zonal</td>
<td>(a)</td>
<td>N</td>
<td>30</td>
<td>2.5°W</td>
<td>7.1</td>
</tr>
<tr>
<td></td>
<td>(b)</td>
<td>N</td>
<td>40</td>
<td>1.0°E</td>
<td>7.2</td>
</tr>
<tr>
<td>Meridional</td>
<td>(a)</td>
<td>W</td>
<td>37</td>
<td>46.5°N</td>
<td>7.5</td>
</tr>
<tr>
<td></td>
<td>(b)</td>
<td>W</td>
<td>32</td>
<td>42.2°N</td>
<td>7.5</td>
</tr>
</tbody>
</table>

The zonal cross-sections show that at the onset of cyclogenesis there is a southerly wind component over the Alps. The maximum speed occurs at about 7 km, but its magnitude varies greatly from case to case (2–30 m s⁻¹). At the same time the winds at the same level about 1000 km to the west have a northerly component of approximately twice the speed. The maximum northerly component of wind over the Alps usually occurs 12–24 hours after cyclogenesis has started.

From the meridional cross-sections it may be seen that during the initial stages of cyclogenesis, winds with a westerly component are dominant over the Alpine region. The maximum speed occurs at about 7 km above the northern slopes of the Alps with an average speed of about 37 m s⁻¹ in all eight cases. This westerly jet shifts about 500 km southwards during the next 12 hours; during this process the altitude of the jet remains the same, but the speed decreases by about 5 m s⁻¹. Finally, it should be mentioned that the representativeness of the means contained in the tables shown in this section is not very high. As a result of the small number of cases, eight, in most of the tables the variational coefficient amounts to one third of the averages. Further, it should be noted that representativeness of the mean values contained in the tables is to some extent limited by the coarseness of the grid (500 km) as well as by the smoothing imposed by the subjective analysis of the synoptic charts. These limitations could hardly be estimated since during the lee cyclone development, synoptic pattern dimensions changed rapidly.

3. DYNAMICAL FEATURES

The horizontal wind components at the beginning of cyclogenesis and 12 hours later have been analyzed for all cases. The mean values of the zonal and meridional components, as well as their 12-hour differences, are shown in Figs. 1–4 for the 1000, 850, 700 and 500 mb levels. Figures 5–8 show the corresponding charts of divergence and vertical velocity. We will now discuss the pertinent features of the fields.

(a) Zonal wind components

At the beginning of cyclogenesis, as seen in Fig. 1, the mean zonal wind component at the surface is generally weak (1–4 m s⁻¹). The highest speeds of 2–4 m s⁻¹ are found only over France and north-east Spain, while the negative values (easterly direction) over central and northern Italy show that convergence already exists at the surface. At 850 mb the belt of high zonal wind components, reaching 12–14 m s⁻¹, extends westward and northward of the Alps, while there is a minimum of less than 2 m s⁻¹ in the northern Adriatic. This distribution, with a strong gradient over the mountain range, shows the real effect of the Alps. Higher up, the wind speed gradient become less sharp over the
Figure 1. Mean zonal wind component (m s\(^{-1}\)) at the beginning of cyclogenesis in the lee of the Alps observed during the ALPEX SOP.

Figure 2. As Fig. 1 but for 12 hours after the beginning of cyclogenesis.
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mountains. However, the speed over and to the north of the Alps increases, while in the central and southern parts of the western Mediterranean basin there is little change.

Figure 2 shows that 12 hours later the belt of high zonal wind component shifts southwards; the maximum speed at lower levels is now located over Corsica, and at higher levels over the Gulf of Lyon.

The difference between the zonal wind components at the beginning of cyclogenesis and 12 hours later shows that the centre of the maximum positive change of speed is located near Corsica, and appears to be the result of the intensification of the zonal wind in the southern sector of the low-level cyclone in the lee of the Alps. The negative values in the northern part of the area are considered to be the result of two effects: at lower levels the zonal wind component decreases due to the blocking of cold air, while at higher levels the decrease is caused by the meridional shift of the belt of high winds.

(b) Meridional wind components

The meridional wind components at the beginning of the cyclogenesis are shown in Fig. 3, and this clearly shows the diffusent flow pattern. At the surface the meridional flow is very weak everywhere, less than 1 m s\(^{-1}\). Higher up, at 850 mb and 700 mb, there are two centres of high meridional component: a positive (southerly direction) centre to the north of the Alps and a negative one over the Tyrrenhian Sea. These centres correspond to the northern and southern branches of the diffusent flow pattern. However, at 500 mb, due to strong zonal wind the northern branch of the diffusent flow is absent, and the positive centre of the meridional component disappears.

Twelve hours after cyclogenesis started the distribution of meridional wind component has completely changed (Fig. 4). At low levels there now exist two centres of high speed: a positive centre over the central part of Italy and a negative one over the Rhone Valley. Both these reflect the well-developed circulation of the low-level cyclone. Higher up, these centres are located further north and are connected with the upper trough whose axis has just reached the Alpine massif. It is worthy of note that at all levels the northerly component is much stronger than the southerly one. Finally, it should be mentioned that the meridional/zonal wind components are approximately the components along/across the Alps.

(c) Divergence

The divergence of the wind field was computed by finite differences

\[
\nabla \cdot \mathbf{V} = \frac{1}{2} \frac{\Delta u}{\Delta x} + \frac{1}{2} \frac{\Delta v}{\Delta y} - \left( \frac{v}{u} \right) \tan \phi
\]

where \( \mathbf{V} \) signifies the vector of horizontal wind velocity, \( u \) and \( v \) horizontal components of wind, \( \Delta x \) and \( \Delta y \) zonal and meridional spacing between grid points, \( a \) the earth's radius and \( \phi \) the latitude.

Initially (Fig. 5) there are two distinct features in the divergence pattern at low levels (1000 and 850 mb): the high negative (convergence) values over the western parts of the Alps and the high positive (divergence) values over the eastern part. In mid-troposphere (700–500 mb) the belt of negative values of divergence spreads over most of the western Mediterranean basin, while the area with positive values covers the whole western and north-western part of the Alpine region. After 12 hours there is a marked change in the divergence pattern, as seen in Fig. 6. The centre of negative values in the lower levels is now located in the Gulf of Genoa, while the northern and north-western parts of the Alps are covered by positive values. In the middle troposphere the negative centre shifts towards the Alps, and a belt of positive divergence develops in the western
Figure 3. Mean meridional wind component (m s\(^{-1}\)) at the beginning of cyclogenesis in the lee of the Alps observed during the ALPEX SOP.

Figure 4. As Fig. 3 but for 12 hours after the beginning of cyclogenesis.
Figure 5. Mean wind divergence (10^(-8)s^-1) at the beginning of cyclogenesis in the lee of the Alps observed during the ALPEX SOP.

Figure 6. As Fig. 5 but for 12 hours after the beginning of cyclogenesis.
Mediterranean. Divergence in the upper levels of the troposphere at this stage of development is not very uniform.

\[ (d) \ \text{Vertical velocity} \]

Vertical velocities for the same grid points on a given isobaric surface with pressure \( p_n \) were obtained from the equation of continuity by the formula

\[
\omega_{p_n} = \omega_{p_{n-1}} + \int_{p_n}^{p_{n-1}} \nabla \cdot \mathbf{V} \, dp.
\]

The omega vertical velocity at the onset of cyclogenesis is shown in Fig. 7. There are three main features:

(i) the area of descent (positive values) covering the Adriatic and the western part of Yugoslavia which strengthens from the surface to 500 mb;

(ii) the area of ascent (negative values) over the western part of the Alps, increasing up to 500 mb;

(iii) the area of ascent over the north-western part of the Alps, weak at low levels and intensifying from 700 mb upwards.

The most prominent feature in the omega field 12 hours after cyclogenesis has started is the development of ascending motion over the Ligurian Sea (Fig. 8). This reflects the intensification and vertical development of the cyclone in the Gulf of Genoa. Also note that from 700 mb upwards the three areas of descent surrounding the ascending air intensify.

The change in omega during the first 12 hours of cyclogenesis gives the strong tendency for there to be ascent over the Ligurian Sea, northern Italy and the Adriatic, as well as descent over the Alps and the southern parts of the west Mediterranean basin.

4. THERMAL STRUCTURE

Synoptic evidence, as well as various investigations (Radinović 1965a; Tibaldi 1980), show that orographic cyclones have a rather specific thermal structure. In order to give an indication of the evolution of the thermal structure during cyclone development in the lee of the Alps, we will consider a few well-known features.

Due to the surface conditions, the air mass over the western Mediterranean during the colder half of the year is very different from that over the surrounding land; the air over the sea is warmer, more humid and less stable. When cyclogenesis starts, the cyclonic circulation develops in a relatively homogeneous air mass. However, once rapid intensification takes place there is stronger ascent and condensation, particularly over the southern slopes of the Alps, which causes the temperature, moisture and stability of the air over the western Mediterranean to change. As the cyclonic circulation enlarges, very warm dry air from north Africa is drawn into the forward side of the cyclone. At the same time there is cold air advection above the mountain level over the cyclone, while below that level advection is obstructed. Later, the cold air flows round the mountain range and penetrates into the western Mediterranean from the south-west (through the Rhone Valley) and the north-east (over the western part of Yugoslavia). Since these air masses have different tracks, their characteristics are slightly different by the time they meet in the lee of the Alps (Bergeron 1928). Besides, many local effects of the mountains (such as wind channelling, cold air lakes, warm air islands, extensive cloud cover, heavy precipitation zones, pressure gradient intensity zones, etc.) cause difficulties in making a frontal analysis. Since there is a lack of a reliable theory about
Figure 7. Mean vertical velocity ($10^{-4}$mb s$^{-1}$) at the beginning of cyclogenesis in the lee of the Alps observed during the ALPEX SOP.

Figure 8. As Fig. 7 but for 12 hours after the beginning of cyclogenesis.
the thermal structure of circulation systems in this region, it is the practice of analysts to apply the simple Bergen school theory, and when the situation is very complicated they usually draw an occlusion in place of a complex front. As a result analyses over southern Europe seem to be rather inconsistent and have a large number of occlusions.

On the basis of the analysis of the eight cases of cyclogenesis which occurred during the ALPEX SOP, we will now describe some additional features of the thermal structure of orographic cyclones. For this purpose the thickness changes in different layers for various sectors of the cyclones during their development have been calculated; the results are given in Tables 8 and 9.

**TABLE 8. AVERAGE THICKNESS CHANGES DURING THE FIRST 12 HOURS OF CYCLOGENESIS (gpm)**

<table>
<thead>
<tr>
<th>Layer (mb)</th>
<th>O</th>
<th>W</th>
<th>S</th>
<th>E</th>
<th>N</th>
</tr>
</thead>
<tbody>
<tr>
<td>850-1000</td>
<td>2</td>
<td>-12</td>
<td>2</td>
<td>-3</td>
<td>-16</td>
</tr>
<tr>
<td>700-850</td>
<td>-9</td>
<td>-19</td>
<td>-3</td>
<td>0</td>
<td>-19</td>
</tr>
<tr>
<td>500-700</td>
<td>-38</td>
<td>-42</td>
<td>-11</td>
<td>6</td>
<td>-47</td>
</tr>
<tr>
<td>500-1000</td>
<td>-55</td>
<td>-61</td>
<td>-25</td>
<td>-6</td>
<td>-78</td>
</tr>
</tbody>
</table>

**TABLE 9. AVERAGE THICKNESS CHANGES DURING THE 12 TO 24 HOURS AFTER THE START OF CYCLOGENESIS (gpm)**

<table>
<thead>
<tr>
<th>Layer (mb)</th>
<th>O</th>
<th>W</th>
<th>S</th>
<th>E</th>
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</tr>
</thead>
<tbody>
<tr>
<td>850-1000</td>
<td>-8</td>
<td>-9</td>
<td>-7</td>
<td>-2</td>
<td>-4</td>
</tr>
<tr>
<td>700-850</td>
<td>-13</td>
<td>-15</td>
<td>-6</td>
<td>-9</td>
<td>-13</td>
</tr>
<tr>
<td>500-700</td>
<td>-36</td>
<td>-27</td>
<td>-24</td>
<td>-17</td>
<td>-21</td>
</tr>
</tbody>
</table>

From Table 8 it may be seen that during the first 12 hours of cyclogenesis there is slight warm advection at the centre of the cyclone only in the lowest layer. In the other layers there is cold advection and this intensifies upwards. The strongest advection of the cold air is in the northern and western sectors of the cyclone, i.e. on the windward side of the mountain range. Also notice that cold advection is present even in the southern sector except in the lowest layer. In the eastern sector the situation is the opposite—slight advection of warm air aloft and cold advection in the lowest layers. This can be explained by the fact that often during cyclogenesis in the western Mediterranean a shallow cold air mass penetrates to the Balkan peninsula, while the south-westerly airflow remains aloft.

During the 12-24-hour period of cyclogenesis, cold advection takes place in all layers (Table 9). The greatest intensity is in the centre and the lowest in the eastern sector. It is also important to notice that there is a large difference in the intensity and volume of the cold and warm advections: compared with the cold advection, the warm advection is practically negligible. As a consequence, it is almost impossible to analyse a warm front caused by warm air 'pulled up' from the south. However, a warm front is usually found as part of a wave formed on the cold front coming from the north or north-west after entering into the cyclone's circulation, but even then the warm front often has the characteristics of a cold front if the cyclone is fast moving.
5. **Kinetic energy change**

For a detailed description of the mechanism of lee cyclogenesis, various terms in the equation for change of kinetic energy in a specified region have been computed using a similar approach to that followed by Palmen and Holopainen (1962) in their study of intensive cyclogenesis over the U.S.A., and by Radinović and Mesinger (1970) who investigated a rapid cyclone development over the western Mediterranean.

From these studies it has been seen that the total kinetic energy change at the beginning of cyclogenesis is higher than 12 hours later. Further it is apparent that nearly all this kinetic energy enhancement comes from the import of kinetic energy by the horizontal flux. In comparison, the terms representing the net energy generation or transformation of available potential energy into kinetic energy are small and some even negative. This suggests that the main source of kinetic energy for lee cyclones is not the local baroclinicity, as might be expected, but rather the existing kinetic energy of the zonal flow. The role of the local baroclinicity seems to be to secure the necessary conditions for a specific redistribution of existing zonal flow kinetic energy.

6. **Mechanisms of development**

One of the mechanisms for cyclogenesis in the western Mediterranean basin, which seems to be dominant, was described several decades ago by Radinović and Lalic (1959), and Radinović (1965a). They suggested that lee cyclogenesis is associated with the existence of a trough in the 1000–500 mb thickness, situated to the west of the Alps, and its pronounced deformation as it approaches the mountain range. This deformation is indicated by a concentration of the thickness lines and intensification of the thermal wind which, according to the thermal development term in Sutcliffe's development theory, produces cyclonic vorticity at the surface. Several recent studies (Egger 1972; Bleck 1977; Buzzi and Tibaldi 1978; McGinley 1982) have supported this basic type of mechanism, though they have also shown that it is considerably more complex than first thought (see in particular McGinley (1982)).

A detailed investigation of the cases of cyclogenesis which occurred during the ALPEX SOP have shown that there are some additional effects involved in the mechanism of lee cyclogenesis which should be taken into consideration. In all eight cases the following features are, more or less, apparent:

(i) A diffluent upper-level flow over the Alps and western Mediterranean immediately precedes lee cyclogenesis (see Figs. 9 to 12).

(ii) During maximum development there is cold advection over the sea-level cyclone (see Figs. 13 and 14).

(iii) Two zones of concentrated thickness lines are evident during maximum development: one extending along the Alpine mountain range and the other along the north African coast (see Figs. 15 and 16).

The results of this study indicate that there may be compensating factors at different levels: generation of vorticity by thermal advection at low levels and advection of vorticity by the large-scale flow at upper levels. This confirms the findings of Buzzi and Tibaldi (1978).

Finally, the cold advection above mountain top level and its blocking below that level during the first stage of cyclogenesis, as described earlier, lead to the formation of two thermal jets in the 1000–500 mb thickness.
Figure 9. Mean sea-level pressure at 00 GMT 2 March 1982, with isopleths at 2.5 mb intervals and isallobars at 2 mb/3 h (dashed lines).
Figure 10. 500 mb contour and temperature at 0000 Z March with isotherms at 3° intervals.
Figure 11. Mean sea-level pressure at 12 GMT 2 March with isopleths at 2.5 mb intervals and isallobars at 2 mb/3 h (dashed lines).
Figure 12. Change in mean sea-level pressure between 12 and 06 GMT 2 March with isopleths at 2 mb intervals.
Figure 13. Change in the 1000–850 mb thickness between 12 and 00 GMT 2 March with isopleths at 20 m intervals.
Figure 14. Change in the 700-500 mb thickness between 12 and 0000 2 March with isopleths at 20 mb intervals.
Figure 15. 1000–850 mb thickness at 12 GMT 2 March with isopleths at 1 dam intervals.
Figure 16. 1000-500 mb thickness at 1200 GMT 2 March with isopleths at 2 dam intervals.
7. COMPARISON WITH OTHER TYPES OF EXTRATROPICAL CYCLONES

The difference in the mechanism for the three types of cyclogenesis is revealed by a comparison of their dynamical and synoptic characteristics which appear on synoptic charts. On the basis of the work of Petterssen and Smeyhe (1971), the present study, and a number of other studies concerned with lee cyclogenesis, the pertinent characteristics of the three types of extratropical cyclone development will be summarized. In this comparison the different types of cyclogenetic mechanisms are called types A, B and C: type A—amplifying frontal waves; type B—disturbance in the upper troposphere; type C—orographic disturbance.

(a) Initial synoptic conditions

A — baroclinic instability of frontal wave

B — disturbance in the upper troposphere accompanied by an augmentation of the baroclinicity (medium and upper troposphere)

C — orographic blocking of cold air mass accompanied by an augmentation of the baroclinicity in the lower troposphere (Radinović 1965a; Buzzi and Tibaldi 1978; McGinley 1982)

(b) Energy source

A — kinetic energy produced through a reduction of the baroclinicity locally within the domain of the frontal wave

B — kinetic energy imported into the cyclone domain, mainly from the jet-stream region

C — kinetic energy produced through a reduction of the baroclinicity locally within the domain of the orographic obstacle and redistribution of kinetic energy of the zonal flow (see section 5 in this paper; Tibaldi et al. 1980; McGinley 1982)

(c) Commencement of development

A — development commences under a more or less straight upper current (without appreciable vorticity advection) in the zone of maximum baroclinicity (frontal region)

B — development commences when a pre-existing upper trough, with strong vorticity advection on its forward side, spreads over a low-level area of warm advection (or near absence of cold advection) in which fronts may or may not be present

C — development commences when a pre-existing upper trough approaching the mountain ranges starts undergoing deformation caused by cold air mass blocking and its deflection round the obstacle. The deformation is accompanied by an augmentation of the baroclinicity within the domain of the obstacle and the creation of the negative thickness vorticity in the layer below the mountain top level in the lee of the mountain Radinović 1965a; Egger 1972; Tibaldi 1980)

(d) Upper trough and low-system connection

A — no upper cold trough is present initially, but one develops as the low-level cyclone intensifies; the separation between the upper trough and the low-level cyclone remains almost unchanged until peak intensity is reached

B — the separation between the upper trough and the low-level system decreases rapidly while the cyclone intensifies, and the axis tends to a vertical position as the cyclone approaches peak intensity
C — the separation between the upper trough, with its axis on the windward side, and the low-level cyclone in the lee of the obstacle remains essentially unchanged until the cold air mass starts flowing over the obstacle and then it decreases rapidly (Radinović 1965a; Buzzi and Rizzi 1975)

(e) Vorticity advection

A — the amount of vorticity advection aloft is very small initially, and remains relatively small throughout the development

B — the amount of vorticity advection aloft is very large initially and decreases towards the time when peak intensity is reached

C — the amount of vorticity advection aloft is small initially and remains small until the cold air starts flowing over the obstacle; then a pre-existing upper trough moves into the domain of the low-level cyclone and positive vorticity advection aloft increases rapidly (see section 6 of this paper and McGinley 1982)

(f) Thermal advection

A — the thermal advection due to baroclinic instability is very pronounced and the main contribution to the intensification of the cyclone comes from the thermal advection

B — the amount of thermal advection is small initially and increases as the low-level cyclone intensifies

C — the amount of thermal advection is negligible initially since the low-level cyclone starts developing in a thermally uniform air mass; but there are two critical times when the thermal advection changes radically: first, when the cold air starts penetrating in the rear of the low-level cyclone, and second, when cold air starts flowing over the obstacle (Radinović 1962; Buzzi and Tibaldi 1978; McGinley 1982)

(g) Baroclinicity

A — the amount of baroclinicity in the lower troposphere is large initially and decreases as the wave occludes

B — the amount of baroclinicity in the lower troposphere is relatively small initially and increases as the storm intensifies

C — the amount of baroclinicity in the lower troposphere is large initially and increases until peak intensity is reached and then decreases. (Radinović 1965a; Tibaldi et al. 1980; McGinley 1982)

(h) Front evolution

A — cold and warm fronts as well as a warm sector are usually well defined throughout the development and the end result of the development is an occlusion of the classical type

B — no one front or several fronts at the same time can be identified in the domain of the cyclone; the end result of the development is a thermal structure that resembles the classical occlusion

C — no fronts initially, then warm advection on the forward side of the cyclone increases as the low-level cyclone intensities but it is difficult to identify a warm front; after cold air penetration a cold front passes from the rear of the cyclone to its forward
side; the end of the development is usually an occlusion of the orographic type (Bergeron 1928)

(i) Movement
A — moves relatively fast initially in the direction of the upper flow and as the disturbance develops aloft it slows down
B — moves relatively fast initially with the speed gradually decreasing as the cyclone approaches peak intensity, when it becomes almost stationary
C — stays stationary until the steering effect (in terms of Sutcliffe’s theory of development) exceeds the orographic effect (Radinović and Lalic 1959; Speranza 1975; McGinley 1982)

(ii) Cut-off
A — rarely cut-off in the upper troposphere and when it happens it occurs in a stage of occlusion
B — it cuts-off regularly as the cyclone approaches peak intensity; it is caused by intense cold advection in the rear and continuing warm advection in the upper troposphere on the forward side of the cyclone
C — cut-off regularly produced at the time of peak intensity, caused chiefly by intense cold advection by the sea-level pressure ridge on the windward side of the mountain obstacle, which builds up as a consequence of cold air accumulation (see section 2)

8. Conclusions

From the results described in this paper, we can conclude that the evidence drawn from the cases of lee cyclogenesis which occurred during the ALPEX SOP supports Danielsen’s ideas (1973) that “the baroclinic instability determines when, and the mountains determine where, the cyclone will form”. That is, an upper air trough consists of the cold air mass in the lower troposphere and strong cyclonic vorticity in the middle of the troposphere. As the trough progresses the effect of upper air cyclonic vorticity is usually compensated by cold advection in the lower layers. However, when it passes over the mountain range the upper air cyclonic vorticity is not adequately compensated and baroclinic instability is triggered. As a consequence, low-level cyclogenesis occurs in the lee of the obstacle.

Another conclusion from this study is that the mechanism of lee cyclogenesis is rather specific. It seems justified to treat the orographic cyclone as a separate type of extratropical cyclone.

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