Cyclogenesis in the lee of the Alps: A case study*

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SUMMARY

A case of deep and rapid cyclogenesis in the lee of the Alps is analysed by means of cross-sections, isentropic maps and trajectories based on synoptic data, to investigate the three-dimensional structure of the phenomenon and the nature of the processes that are responsible for it. The results of the analysis are also compared with available theoretical models. Considerations of the various scales involved in the development lead to a description of the phenomenon in terms of two distinct phases: a very rapid 'trigger' phase due to interaction between the frontal layer and the Alps, and a more usual "baroclinic development" phase. Through a selection process typical of baroclinic instability, the lee cyclone acquires the observed horizontal and vertical scales and undergoes the normal life of a mid-latitude depression. The insufficiency of the present synoptic network for a satisfactory analysis of such a meteorological phenomenon is also stressed.

1. INTRODUCTION

Cyclogenesis in the lee of the Alps has always been a challenging problem for the European forecaster. The problem has been attacked several times in the past (CENFAM 1961–1965; Radinović 1965; Egger 1972; Buzzi and Rizzi 1975); a more recent paper by Speranza (1975) includes a detailed review of the available literature on the subject.

It is well known that the whole area south of the Alps is a highly cyclogenetic region; depressions in the Mediterranean area are generally of small scale compared with those of higher latitudes and their development is characteristically rapid.

It is generally admitted that the orography surrounding the area plays a fundamental role in the generation and evolution of the depressions. Operational numerical models do not satisfactorily incorporate the effects of obstacles like the Alps, mainly because of the limited horizontal extent of these steep mountains compared with the spacing between grid points. Numerical experiments have been carried out to investigate this problem; they have clarified some aspects of the phenomenon, but it has not yet been possible to apply their results to operational numerical forecasts.

Theoretical models relating to rotating, stratified and unsheared flow over isolated obstacles, which have recently appeared in the literature, can explain some observed characteristics of the local interaction between the large-scale flow and the Alps, as is described in the following sections.

The high–low pressure perturbation, with the associated anticyclonic vorticity over the mountain and cyclonic vorticity in the lee, is described in the results of Huppert and Bryan (1976) and Buzzi and Tibaldi (1977). The anticyclonic vorticity is produced in both models by vortex tube compression over the obstacle; the lee cyclonic vorticity is, in the former, a result of the time dependency of the problem (the flow is started from rest), and in the latter is produced by means of Ekman pumping at the lower boundary. Both models are able to account for the small-scale and shallow depressions that have often been observed in the lee of the Alps (CENFAM 1961–1965) and sometimes confused with cyclogenesis. The use of this term should be restricted to those cases, like the one described in this paper, in which baroclinic development leading to a deep cyclone is evident. In the models mentioned above baroclinic instability is, on the contrary, not represented.

The trajectories at low levels near the Alps (see later) indicate strong horizontal deflection of the flow round the mountain range; this deflection is more pronounced on the...
eastern side. The anticyclonic curvature over the obstacle is a result common to all theoretical models of rotating and stratified flow over topography. Nevertheless, up to now there has been no analytical study able to answer in a satisfactory way the question: How much flow goes over and how much goes round? The only noteworthy approach in the literature about the definition of 'over and round' is due to Huppert and Bryan. In our case study, although the spatial resolution of radiosoundings is not sufficient to define the flow just over the mountain, trajectories indicate that complete overflow, in the sense that air parcels near the ground on the upwind side reach the top of the mountain, is not likely to have happened.

Another feature that seems to play an important role in this kind of lee cyclogenesis is the retardation of advective processes caused by the Alps. All analytical models of flow past isolated obstacles produce, to a certain extent, a deformation of the motion field that could partially account for this retardation process. The possibility that this retardation acts as a cyclogenetic factor has already been put forward by Radinović (1965) and Egger (1972). Radinović relates the formation of the lee cyclone primarily to the deformation of the thickness (or temperature) field in the lowest layers, produced when a cold front impinges on the Alps on their convex side. The blocking of cold air exerted by the mountain range, together with its arc shape, are responsible, in Radinović's description, for the production of negative values of $V_\phi^2 h$ in the lee (where $h$ is the 1000–500 mb thickness). If a pressure trough is moving over the region in the upper layers, a corresponding pressure minimum should be induced near the ground. The process of graphical integration proposed by Radinović to forecast cyclogenesis involves rather crude assumptions, and modern numerical models have made it obsolete, but at least in the initial stage of the cyclone development the essential features contained in his diagnosis appear to be realistic.

Egger produces some successful numerical experiments with a primitive equation model. The Alps are simulated as a vertical wall blocking the normal flow in the lowest layers up to about 3000 m, so that the effects of the slope of the terrain are neglected. Horizontal advection varying in the vertical because of the blocking effect of the mountain, as well as horizontal deformation of the thermal field induced by differential advection in the horizontal, seem to be responsible for the pressure decrease in the lee during the first stages of Egger's experiment. Then a proper baroclinic development takes place in the region. In Egger's work the arc shape of the mountain seems to be important in producing cyclogenesis, because an experiment with a straight barrier does not reproduce the phenomenon. Nevertheless, in a more recent numerical experiment by Trevisan (1976) lee cyclogenesis has been reproduced modelling a straight mountain aligned from west to east, with a gap in it of about 900 km, simulating the gap between the Alps and the Pyrenees. This gap gives rise to a 'channeling' effect deforming the cold advection from the north and inducing again a negative Laplacian in the temperature field, where later a cyclone forms.

We present here a detailed case study based on synoptic data. We aim to describe the three-dimensional structure of the phenomenon, identifying the local influence of the orography on the development of the depression.

The nature of the data forces our analysis to be limited to rather large time–space scales compared with those involved in the cyclogenesis. Nevertheless, existing synoptic data are sufficient to reveal typical features, and to indicate the preferred direction of further more detailed studies.

2. THE ANALYSED CASE: PRE-EXISTING SYNOPTIC SITUATION AND INITIAL STAGE OF DEVELOPMENT

We will describe the cyclogenesis in the lee of the Alps that occurred on 3 April 1973. We chose this case for analysis because it is representative of rapid cyclonic development in
the lee of the Alps with evident orographic influences on the flow and temperature fields. Furthermore, it occurred in a season when sensible heat release from the Mediterranean can be excluded as an important contribution to cyclogenetic mechanism. Two winter cases have already been analysed with similar techniques (Buzzi and Rizzi 1975) and reference is made to that paper for a more complete picture.
On 2 April a very strong cyclone had swept from the British Isles to the Baltic Sea (see Richards and Stubbs 1974). We analyse here the secondary cyclogenesis that developed when the cold northwesterly airstream associated with this primary cyclone impinged on the Alps. As is typical in these situations, the new cyclone grew independently, accompanied by a
rapid deepening of the trough aloft which ultimately developed into a cutoff mid-tropospheric low.

The general features of the event can easily be recognized in Figs. 1(a)–(d) and 2(a)–(b). In Fig. 3 the strong primary cyclone, moving eastwards on 2 April from the British Isles
to central Europe, can be observed with its well-defined cloud pattern. Very deep clouds are revealed by their shadows near the centre, and the cold front in the rear is pronounced. Fig. 4 contains the first of the cross-sections presented here, all of which are almost perpendicular to both the surface cold front and the Alps, and along the approximate direction of propagation of the cold outbreak. In this cross-section (corresponding to the line A–B of Fig. 1(a)) the cold front appears as a unique and diffuse baroclinic region in the layer between the surface and 350 mb, from Aughton to Trappes. At the surface the cold front can be identified by the presence of a low-level jet near Trappes. The frontal structure, particularly in the mid and upper troposphere, is more pronounced to the west of the cross-section; the maximum normal velocity at about 350 mb on the cross-section corresponds to the exit region of the main upper-level jet (see also Fig. 2(a)).

At 00 GMT of the third the surface cold front has reached the Alps having been retarded in the mountain region, as can be seen in Fig. 1(b). In the same figure we notice the high–low pressure perturbation in the Alpine region. This can be ascribed to the local influence of the mountain. The frontal zone does not seem to play a fundamental role in the formation of this feature in the pressure field near the Alps. From the surface maps at intermediate times (not shown), it is evident that the high–low pattern builds up before the frontal region reaches the Alps, persists while the front approaches the massif and intensifies when the front crosses the obstacle.

In the cross-section of 00 GMT, 3 April (Fig. 5, along line C–D in Fig. 1(b)) two separate and more intense frontal structures appear: one from the ground to approximately 550 mb,
and one in the mid and upper troposphere well above the mountain. While the low-level cold front is still on the north side of the Alps, the mid-tropospheric one is already above the lee region and they are separated by a layer of low static stability. The mid-tropospheric front has entered this cross-section from the west, moving along the southwestern side of the advancing upper air trough underneath the associated jet stream.
Figure 6. Isentropic analysis, $\theta = 285\,\text{K}$, 3 April 1973, 00 GMT. Thin lines: pressure in mb; thick lines: Montgomery streamfunction isolines at $60 \times 10^5 \text{cm}^2\text{s}^{-2}$; wind in knots.

The tilting of the line that connects the low-level secondary wind maximum with the jet maximum aloft has been reversed in 12 hours. At the same time a pool of air with very low static stability appears over Milan at about 700 mb (approximately the level of the top of the mountain). This allowed thunderstorms to develop in the region south of the Alps approximately six hours later, in advance of the low-level front, at a time when direct solar radiation cannot be effective. Thunderstorms were not previously observed along the front.

The isentropic map shown in Fig. 6 ($\theta = 285\,\text{K}$) represents the low-level frontal layer in the cross-section of Fig. 5; the intersection of the 285 K isentropic surface with the ground marks the position of the surface front within a few tens of kilometres in the region between the Pyrenees and Czechoslovakia. A pronounced diffuence in the wind field where the intersection coincides with the upwind side of the Alps is evident, confirming local retardation of the southeastward displacement of the front at low levels. All these facts indicate that the obstacle is responsible for a vertical variation of the velocity of propagation of the system, retarding the lower layers, and for a horizontal deformation of the low-level front. The possibility that these effects act as a trigger for baroclinic instability has already been put forward in the literature and will be discussed in more detail later.

3. THE CYCLONE LIFE

The secondary cyclone develops rapidly between 00 and 12 GMT on 3 April over the Po valley. It then moves slowly southeastwards while increasing in horizontal extent, as can be observed in Fig. 1(b), (c) and (d). In the meantime the primary cyclone dies out, moving northeastwards and becoming more separate from the secondary one. The latter shows up as a wave in the surface front as can be observed on the surface charts and on the isentropic map at $\theta = 285\,\text{K}$ at 12 GMT of the third (Fig. 7).

This vortex is characterized by distinct cyclonic vorticity; cyclonic horizontal shear is striking on the cross-section in Fig. 8, drawn along line E-F of Fig. 1(c). The horizontal velocity gradient is concentrated along the front that extends from about 400 mb over the Alps to the surface over central Italy. A maximum of northeasterly wind appears south of the Alps in a layer from the surface up to the front. This pronounced baroclinic region can be identified with the one that intersected the ground in the cross-section of Fig. 5. Trace of
the upper-level front present in the latter can be partly recognized in the cross-section at 12 GMT between Rome and Brindisi, even though data from Rome are missing above 400 mb, and the cross-section there is not parallel to the maximum temperature gradient. The separation between the two frontal layers and the wind maxima associated with them, already apparent at 00 GMT on the third, is seen to have increased in the next 12 hours.

The vertical depth of the cyclogenesis is larger than the vertical penetration of the frontal region: a closed geopotential minimum is pronounced also at 500 mb (Fig. 2(b)), and the absolute maximum of geopotential decrease between 00 and 12 GMT is actually located at 300 mb over the same region. Apparently the entire troposphere above the region south of the Alps experiences cyclonic development, but the westward tilting of the
trough line is less pronounced than the average for baroclinic disturbances in the initial stage of development. These aspects suggest that different mechanisms could act at the same time at different levels to produce a deep vortex, as discussed in section 5.

The field of motion has been analysed with particular attention, in order to reveal the orographic influences. Twelve-hour trajectories have been computed from isentropic maps with Danielsen's (1961) technique which is based on conservation of energy and dry entropy. The 285 K surface is representative of the flow, in the Alpine region, at heights between 850 and 650 mb, i.e. around and below the top level of the mountain range. It can be observed in Figs. 7 and 9(a) that the flow at these levels is essentially diffuent round the obstacle. The splitting into two branches of the northwesterly low-level stream caused by the upstream influence of the Alps is evident over central Europe. Particular evidence of orographic influence is seen near the eastern edge of the mountain range, where strong anticyclonic curvature, with substantial ageostrophic component, is experienced by air coming from central Europe. This air is responsible for the cold outbreak over the Po valley and the northern Adriatic Sea. This local flow pattern is a common feature in other cases of cyclo-
genesis in the lee of the Alps (Buzzi and Rizzi 1975) and can account for the occurrence of bora wind under these circumstances.

Fig. 9(b) shows trajectories at \( \theta = 295 \text{ K} \), representative of the warm flank of the cold front on the eastern side of the lee cyclone. These trajectories are very different from those in Fig. 9(a). Between 00 and 12 GMT of the third, warm air flows from southwest to northeast over the Alps, experiencing strong ascent (see also later); farther west a branch of descending cold air flows rapidly southeasterwards, penetrating over the western Mediterranean.

A splitting of the flow northwest of the Alps is seen also at \( \theta = 305 \text{ K} \) (not shown), which is far above the mountain top. Direct orographic influence at this level is doubtful, but the deformation of wind and temperature fields in the lowest layers could account also for a deviation of the air aloft, as discussed in the next section.

The nephanalysis (Fig. 10), derived from a satellite cloud photograph taken at about 11 GMT on the third (not shown), and from station reports, gives a clearer synoptic picture of the system. The comma shape of the clouds over northeastern Europe, near the centre of the primary cyclone, is still visible. The cloud belt between central Italy and Spain is associated with the western part of the cold front that has penetrated over the western Mediterranean and which was over the Bay of Biscay the day before (Fig. 3). In the intermediate area, occupied by the secondary cyclone, the cloud distribution has undergone the most important changes, losing the form of one compact narrow belt. Widespread and deep clouds cover northern Yugoslavia, Austria and northeastern Italy, associated with the main precipitation region. In particular there are several cumulonimbi and thunderstorms in the belt extending between the Ligurian Sea and the eastern Alps, as indicated by surface and 'atmospherics' reports. This belt is readily associated with the intense front in the low troposphere, visible about one hour later in the cross-section of Fig. 8.

This nephanalysis can be compared with the trajectories computed on the 295 K surface (Fig. 9(b)) and with the associated pattern of vertical velocity (Fig. 11), averaged over twelve hours, that has been drawn by plotting the value of the vertical displacement at the mid-time point of each trajectory.

The correlation between the distribution of intense ascending vertical motion and clouds is fairly good.

Strong ascent, associated with southwesterly flow streaming over the Alps, takes place

![Figure 10. Nephanalysis; 3 April 1973, about 11 GMT. Light shading: uniform cloud cover; heavy shading: region of continuous precipitation. Areas of scattered clouds have not been considered.](image-url)
in the middle troposphere over northern Italy. The vertical velocity pattern is evaluated on a $\theta$ surface (295 K) that lies on the warm side of the low-level frontal layer (see Figs. 5 and 8) where strong upward motion is typical. Nonetheless the non-uniformity of the pattern along the front and the shape of the zero vertical velocity contour that follows the mountain chain suggest that this motion can be an important aspect of the readjustment process taking place during the interaction of the front with the orography (see section 5).

The geopotential deficit at the centre of the vortex does not increase further at any level after 18 GMT. A cutoff process takes place in the middle troposphere from this time, with warm advection over central Europe. This process ultimately leads to a separate vortex over the Mediterranean which moves slowly southeastwards, gradually decreasing in intensity from 4 April.

4. THE SCALES OF MOTION OF THE PHENOMENON

In order to discuss in more detail the processes involved, we have evaluated from our analysis the basic scales of the motion. We define the horizontal scale $L$ of a system as $\lambda/2\pi$, where $\lambda$ is the dominant horizontal wavelength, subjectively estimated.

The basic flow which interacts with the Alps is characterized by a variety of horizontal scales. It is possible, however, to distinguish two important scales: that of the primary cyclone (400–1000 km); and the transverse scale of the associated low-level and mid-troposphere fronts (about 50 km – the longitudinal scale is similar to that of the primary cyclone).

The scales of the perturbation induced directly by the mountain are determined by the transverse and longitudinal scales of the Alps, which range respectively from 80 to 100 km and from 200 to 250 km. Evidence of disturbances in the pressure field on these scales is given by the 'high–low' pattern near the Alps (Fig. 1(b)) already described, and by the concentrated gradient along the Alps still visible in Fig. 1(d). Perturbation on the same scales can be recognized also in the deformation of the thermal field (Fig. 7) and of the velocity field (Figs. 9(a) and 11) near the obstacle.

In the initial stage, i.e. between 00 and 12 GMT, 3 April, the secondary cyclone grows with a horizontal scale of about 250 km (evaluated on the surface map at 06 GMT, not shown). Then it undergoes an increase of scale, reaching 500 km after 00 GMT on the fourth. 500 km is also the scale of the upper-level development, even in the initial stage.
To define usefully the vertical penetration of the perturbation we must distinguish between the quasi-stationary orographic disturbance and the deformation of the cold front resulting from its interaction with the Alps. The former has a vertical penetration of about 2 km, evaluated by comparing the 850 and 700 mb maps of this and other cases (not shown) of flow over the Alps (see also Buzzi and Tibaldi 1977).

The deformation of the cold front is obvious up to about 4 km, but thermal distortion in the lower layers may modify the velocity field at higher levels through the thermal wind. In particular, the penetration from northwest of the upper-level jet over the western Mediterranean is favoured by the enhancement of the low-level cold advection west of the Alps over the Rhone valley and the Gulf of Lions: a sharp wedge of cold air can be observed in this region on the 1000–500 mb thickness map (not shown). Such orographic distortion of the upper flow has already been suggested to be fundamental for the generation of lee cyclones on the basis of numerical experiments (Trevisan 1976).

Evaluation of the vertical scale of the secondary cyclone presents some difficulties. The geopotential decrease at different levels is coherent in time throughout the depth of the troposphere, but in the initial stage the horizontal scales are different. The cross-section of Fig. 8 shows two separate regions of cyclonic vorticity; the low-level one has the smaller horizontal scale, while the cyclonic region aloft is much the broader. In the later stage, however, the cyclone becomes more coherent at all levels, taking the typical aspect of a mid-latitude cutoff cyclone, where it occupies the depth of the troposphere.

5. PROPOSED MECHANISMS

Fig. 12 shows the growth in the intensity of the lee cyclone. This growth has been isolated using the following technique, also used by Mansfield (1974). The 'undisturbed' situation, i.e. with no secondary cyclogenesis and no mountain effect, has been subjectively reconstructed on surface maps every six hours. Then pressure differences between the reconstructed undisturbed state and the actual state have been calculated.

The most striking feature of Fig. 12 is the violent growth during the initial phase labelled A in the figure. The e-folding time is about 6 hours until 06 GMT of the third, then it becomes much longer (25–30 hours, phase B) until 18 GMT. No appreciable growth is

![Figure 12. Plot versus time of the natural logarithm of the difference between the ideal reconstructed pressure (see text) and the actual pressure in the centre of the cyclone.](image-url)
measured subsequently (phase C). Now the front crosses the Alps during phase A and the growth of the pressure perturbation at the surface has been evaluated in other cases of cold fronts crossing the Alps, and the same typical values have been observed. Thus we ascribe phase A merely to the front crossing the Alps.

During phase B typical values of $Ro$ (Rossby number) and $Ri$ (Richardson number) for the region occupied by the lee cyclone are respectively 0.3 and 7. These values are reasonably compatible with those in the quasi-geostrophic baroclinic instability theory (Eady 1949), for which $Ro \approx Ri^{-1} \ll 1$.

The e-folding time, computed by inserting into Eady’s theory the average values of static stability and vertical shear measured in phase B of the cyclone life, is about 30 hours. In this calculation it has been assumed also that the ratio between horizontal and vertical scales is that for maximum growth rate (see later). The computed value agrees satisfactorily with that observed (25–30 hours), indicating that baroclinic instability theory is able to account for the second phase of the cyclogenesis.

On the other hand, Eady’s theory does not describe the early stage of the development (phase A). In this period, the interaction at low level between the cold front and the Alps is fundamental, and the local values of $Ro$ and $Ri$ have to be considered. $Ro$ can be evaluated from trajectories passing near the Alps: typical values of around unity occur below 600 mb near the eastern side of the mountain range. The Richardson number for the flow interacting with the Alps during phase A is about 1–2.

The stability problem for values of $Ro$ and $Ri$ of order unity has been investigated by Orlanski (1968). Baroclinic and different types of barotropic instability are simultaneously involved in this range of scales, and the computed variability of the growth rate is high. Nevertheless, the growth rates for $Ro$ near 1 and $Ri$ of a few units, as in our case, never exceed the value for quasi-geostrophic baroclinic instability. The conclusion is therefore that the rapid growth of the depression at low level during phase A (Fig. 12) is not to be related to some known form of internal instability of the flow.

The speed of propagation of the low-level front is about 40 km h$^{-1}$; this implies roughly six hours for the front to cross the Alps. We therefore suggest that during phase A a readjustment of the frontal circulation takes place, as a consequence of the impact of the front on the mountain. This disturbance provides a ‘trigger’ for the baroclinic development that dominates phase B, when the deepening depression reaches a better organization, increasing its horizontal scale and becoming a proper cyclone. The nature of this front-readjustment process, which is essentially three dimensional and involves scales typical of inertia–gravity waves, will be the subject of future investigations.

A geopotential decrease, simultaneous with the low-level cyclogenesis, is observed also in the middle and upper troposphere. This deepening takes place, between 00 and 12 gmt, 3 April, in the region centred over northern Italy, corresponding at that time to the southern part of the upper-level trough and a cutoff process is initiated.

Rapid cyclonic development at upper levels is not yet a well-known process; it has often been observed in association with a strong jet moving southeast along the west part of a trough in the middle and upper troposphere (see, for example, Palmén and Newton 1969). In this case geopotential decrease occurs on the left side of the exit region of the jet. A process of readjustment of the mass field to the velocity field has been proposed as a qualitative explanation, but the mechanism is still essentially unknown.

The role played by the jet and by the upper-level development in cyclogenesis in the lee of the Alps has already been analysed and discussed (Reiter 1963; Danielsen 1973; Buzzi and Rizzi 1975). In particular, a correlation between cyclogenesis and penetration of the jet over the western Mediterranean, flowing over the region between the Alps and the Pyrenees, has been pointed out. If the jet has a strong northerly component, it splits into two
branches over central Europe: one branch turns cyclonically to the east, north of the Alps; the other branch moves south towards the Mediterranean. All these phenomena have also been observed in our case study. We have seen that it is not unreasonable to hold the Alps responsible for a deflection of the jet aloft, via the thermal deformation induced at low levels; as a consequence, if upper-level cyclogenesis is related to the position of the jet, the Alps can play a role in localizing, if not causing, upper-level cyclogenesis.

Our basic hypothesis is that strong cyclones, extending through the entire troposphere, form in the lee of the Alps when upper-level cyclogenesis (in the sense of a deepening trough) takes place over low-level cyclogenesis. In the initial phase, A (Fig. 12), the two processes are essentially uncorrelated, in the sense that different mechanisms act at different levels: at low levels the readjustment process due to the Alps–cold front interaction; at high levels baroclinic–barotropic instability associated with the upper front and jet system.

In the later phase, B, the vortex gradually organizes itself, acquiring vertical coherence through the whole troposphere. Baroclinic instability can satisfactorily account for the growth and emerging scale during this interval. Neither the upper-level cyclogenesis nor the low-level mountain-induced perturbation is separately sufficient to give the observed vertically coherent cyclone that develops in about 12 hours.

Numerical forecasts have been run for this and other cases either without orography or including it only after it has been smoothed. These suggest that, without the Alps, upper cyclogenesis would have occurred, but to the east of its actual position, and that the surface cyclone either would have not appeared or would have appeared later farther east, as a consequence of the upper-level development.

The apparent increase of horizontal scale as the cyclone develops can be explained in the following way. The initial low-level perturbation, induced by the Alps, has a maximum horizontal scale of about 250 km, and a vertical penetration of 3–4 km. Simultaneously the upper-level trough deepens on a larger horizontal scale (about 500 km). Local baroclinic development begins at this stage and its vertical penetration can reasonably be assumed to be 8–10 km; in fact, broad-scale baroclinity is present throughout the troposphere, and no very stable layer exists that, acting as a lid, could separate the upper and lower circulation systems (unlike the cases described by Mansfield 1974). On the other hand, the fastest growing baroclinic waves are those characterized by horizontal and vertical scales that satisfy the relation (e.g. Green 1960) \( L/H \approx \sqrt{Nf} \), which gives \( L = 800–1000 \) km for typical values of \( N \) and \( f \), if \( H = 8–10 \) km. Thus the actual final horizontal scale of our cyclone (about 500 km) is intermediate between the initial scale near the surface and the dominant scale predicted by baroclinic instability theory for a spectrally uniform initial perturbation of infinitesimal amplitude. But in our case the amplitude of the initial perturbation is neither infinitesimal nor constant for all wavelengths; it is, on the contrary, large on scales smaller than those for maximum instability. Therefore baroclinic development, acting for a finite time, causes a selective growth on scales larger than those initially dominant, and this effect may be held to explain the increase with time of the apparent horizontal scale of the cyclone.

The last consideration concerns the possible role played by the latent heat release. Recent studies (see, for example, Gall 1976) have demonstrated that latent heat can have a strong influence on cyclonic development, acting mostly in increasing the growth rate and in modifying the kinetic energy distribution in the vertical. In our case, on the other hand, precipitation began in the region of cyclogenesis after 06 GMT of the third, i.e. when the first stage of the cyclonic development had been completed. It can therefore be concluded, in agreement with Radinović (1962), that latent heat plays an unimportant role in the initiation of the cyclone, although it may modify its characteristics in subsequent stages.
6. CONCLUDING REMARKS

The analysis of this case of cyclogenesis in the lee of the Alps, compared with previous studies and numerical experiments, has led to the conclusion that deep and rapidly growing cyclones forming south of the Alps are the results of simultaneous processes, strongly influencing each other:

(a) An orographic perturbation acts on the low-level cold front as it crosses the Alps. A readjustment process takes place at this initial stage and overlaps with the small-scale high–low pressure system generally associated with an airstream flowing over an isolated obstacle. The perturbation due to the readjustment process is responsible for the initiation of a low-level cyclonic development.

(b) A large-scale trough deepens aloft over Italy, the localization of its maximum deepening being related to the deflection of the northwesterly jet towards the Alpine region; this, in turn, is a possible consequence of the low-level mountain-induced thermal deformation.

(c) Baroclinic instability, triggered at low level by the frontal readjustment process, and sustained aloft by the deepening of the large-scale trough, accounts for the subsequent stage of the cyclone life; the cyclone increases its horizontal scale to 500 km and takes the typical feature of a cutoff cyclone of mid-latitudes.

A better physical understanding of the mechanisms involved can be achieved with the aid of theoretical and numerical investigation. The limits of analyses based on data provided by the present synoptic network are serious. This study shows that cyclogenesis in the lee of the Alps involves space and time scales which are hardly resolved by the present distribution of radiosonde stations. Not only are small-scale dynamics (gravity waves and turbulence generated by the roughness of the terrain) filtered out by the present synoptic network, but even phenomena on the same scale as the Alpine range, of the order of few hundreds of kilometres, cannot be captured with certainty. Stations that could provide upper air soundings are lacking in the mountain area. Spacing among existing stations is much larger in the Mediterranean region, where the cyclones form, than over central Europe. Moreover, data provided every 12 hours do not permit a reasonable investigation of a phenomenon whose growth rate is about six hours.

Surface data alone are inadequate for detailed three-dimensional analyses. Particularly in regions of steep orography they are strongly influenced by local conditions which prevent coherence on the synoptic scale.

Quantitative analyses, precise enough to allow clarification of the physical mechanisms of this phenomenon, can be performed only with the support of at least one ad hoc observational study (Charney 1976). Essential features having been isolated, the existing network may be adequate as a monitor.

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