The Munich Hailstorm of July 12, 1984: A Discussion of the Synoptic Situation

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Abstract:

The severe hail storm which hit Munich and other parts of Southern Bavaria on July 12, 1984 is investigated in respect to its synoptic scale environment. The analyses revealed a potentially unstable situation at the northern side of the Alps, where (within a frontal zone) relatively cool but very moist air from the Mediterranean had flown under warm and drier air masses. An inversion on top of the moist air hindered the release of the instability and could be dissolved neither through heating from the ground nor through vertically different temperature advection. The observed destabilisation resulted rather from a synoptic-scale upward motion – caused by the combined effect of warm air advection and positive vorticity advection, increasing with height, in front of a weak trough migrating from France northeastwards. Within a large field of middle and high clouds the first convective cells formed over the Swiss Alps where the release was facilitated through orographic effects. Due to the inclusion of the moist air laying over Southern Bavaria these cells experienced a rapid amplification when they travelled northeastwards. They developed into hail producing meso-scale convective systems from which the most intense one crossed Munich.

Zusammenfassung: Das Münchener Hagelunwetter vom 12. Juli 1984: Diskussion der synoptischen Situation

Das schwere Hagelunwetter, das am 12. Juli 1984 München und andere Teile Südbayerns heimsuchte, wurde hinsichtlich seiner synoptischen Entstehungsursachen untersucht. Die Analysen ergaben, daß an diesem Tag am Alpennordrand eine hochgradig potentiell instabile Schichtung herrschte, nachdem sich im Bereich einer Frontalzone vergleichsweise kühle, aber sehr feuchte Mittelmeerluft unter warme und trockene Luftmassen in der Höhe geschoben hatte. Eine Inversion an der Obergrenze der feuchten Luft verhinderte zunächst die Auslösung der Instabilität und konnte weder durch Aufheizung vom Boden noch durch vertikal unterschiedliche Temperaturadvektion beseitigt werden. Die beobachtete durchgreifende Labilisierung ergab sich vielmehr als Ergebnis einer großräumigen Hebung – ausgelöst durch die kombinierte Wirkung von Warmluftadvektion und mit der Höhe zunehmender positiver Vorticityadvektion auf der Vorderseite eines schwachen Troges, der von Frankreich nordostwärts schwenkte. Die Hebung führte zunächst zur Bildung mittelhoher und hoher Bewölkung. Die ersten Konvektionszellen in dieser Wolkenzone formierten sich im Bereich der Schweizer Alpen, wo durch orographische Effekte die Auslösung der Instabilität erleichtert wurde. Durch Einbeziehen der nördlich des Gebirges lagernden feuchten Luft erfuhren diese Zellen bei ihrer Nordostverlagerung eine starke Intensivierung und entwickelten sich zu hagelproduzierenden konvektiven Systemen. Das am stärksten ausgeprägte dieser Systeme überquerte den Münchner Raum.

Résumé: Chute de grêle à Munich le 12 juillet 1984: Discussion de la situation synoptique.

On étudie, en fonction de son environnement synoptique, la forte chute de grêle qui a affecté Munich et la Bavière méridionale le 12 juillet 1984. L'analyse a révélé une situation potentiellement instable au nord des Alpes où, au sein d'une zône frontale, de l'air relativement froid mais très humide venant de la Méditerranée s'était glissé sous une masse d'air chaud et plus sec. Une inversion au sommet de la couche d'air humide empêchait le déclenchement de l'instabilité. Par ailleurs, cette inversion ne pouvait être résorbée ni par un réchauffement du sol ni par advection différentielle de température suivant la verticale. La déstabilisation

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observée résulte plutôt d'un mouvement ascendant d'échelle synoptique causé par l'effet combiné d'une advection d'air chaud et d'une advection de vorticité positive, croissant avec l'altitude, à l'avant d'un faible creux barométrique se déplaçant vers le nord-est à partir de la France. Les premières cellules convectives se sont formées sur les Alpes suisses, avec apparition d'un champ étendu de nuages moyens et élevés, où l'instabilité a été favorisée par des effets orographiques. En raison de la présence d'air humide sur la Bavière méridionale, ces cellules se sont rapidement amplifiées au cours de leur déplacement vers le NE. Elles ont atteint le stade des systèmes convectifs de mésoéchelle produisant de la grêle. C'est la partie la plus active qui a affecté Munich.

1 Introduction

On July 12, 1984, between 18 and 22 CEST, 16 and 20 UTC, respectively, parts of South Bavaria were hit by a severe hail storm which caused heavy damage to buildings, goods, and vegetation. The highest density of destruction was reported from Munich, which was right in the middle of the 250 km long and 5 to 15 km wide hail swath. The size of the hail stones ranged around 5–6 cm diameter, but stones up to 9,5 cm diameter were also found (BERZ, 1985). The recorded amount of precipitation came up to 40 l/m^2 (in Munich).

The great damage, of which around 500 Mio $ were covered by insurances, initiated a public discussion whether this meteorological phenomenon had been predictable or not. Surely, not all harms can be avoided by issuing a hail alert. But, as we think, the mechanism which led to this storm should be investigated in order to better understand this kind of hazards; a knowledge which may be useful for local weather prediction in the future.

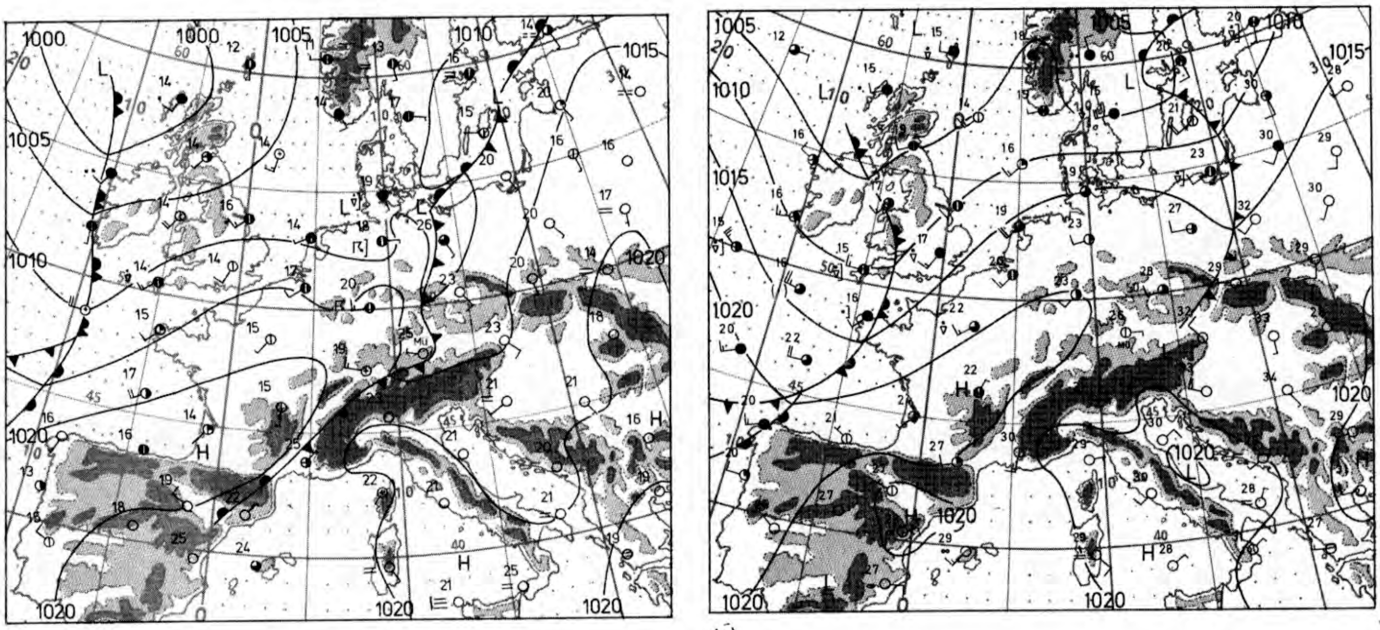
This paper has specially been prepared to display briefly the temporal variation and spatial distribution of relevant thermodynamic and kinematic-dynamic variables. Since this study had to be based on instantly available data, it should be noted that the results are provisional. Further investigations are necessary to gain a full understanding of the reasons for this remarkable event.

2 The Weather Situation on July 11, 1984 and during the first Half of July 12, 1984

On July 11, 1984 southern and eastern parts of Germany were influenced by hot and dry tropic air masses (cT according to „Berliner Wetterkarte"). It had come into this area between a weak low pressure zone over France and Spain, and a high pressure ridge, extending from the Mediterranean to the Baltic Sea. The maximum day temperatures reached values between 33 °C and 37 °C (Munich and Nuremberg).

In the course of the day a foehn-like descending motion set in at the northern flank of the Alps and caused a further warming of the lower troposphere. At the Hohenpeissenberg – almost 1000 m high – a temperature of 32.4 °C was measured, the highest value ever observed during the more than 200 years of observation. Combined with this warming a drying-up of the air occurred.

During the evening the low pressure zone and the embedded cold front approached Bavaria from the west. During its passage thunderstorms were observed in Baden-Wurttemberg, near Lake of Constance and in northern Bavaria, but further in the east the front lost its effectiveness rapidly. In the night between the 11th and 12th of July (at 22 UTC = 24 CEST) the front reached Munich (Figure 1a), causing there only wind shift from south to northwest, a 3/8 stratocumulus cloud cover and later some gusts.



a) b)

Figure 1 Surface maps from July 12, 1984, 00 UTC (a) and 12 UTC (b), with isobars (in hPa), fronts and some observations; Mü Munich

Connected with rising pressure a widespread descending motion set in during the second half of the night and in the morning. As a result the thundery activity and precipitation west and north of Bavaria ceased very soon and the sky cleared up. At the same time the cooler but moister air behind the front covered Bavaria up to the Alps preventing a further motion of this air mass. During the forenoon the air was heated from the ground and convection was released. But only a few amounts of shallow cumulus cloud developed until early afternoon.

Due to the pressure rise a surface ridge of high pressure developed from France in northeasterly direction. Between this ridge and a quasistationary high pressure area over the Mediterranean a channel of low pressure remained just over the Alps in which some heat lows formed in the big valleys. Over southern Germany, a weak northerly current set in near the ground which reached up to the Alps. On top of the mountains it changed into a partly strong southerly flow.

The surface front which had reached the northern flank of the Alps in the morning was completely destroyed after that time. No real front-like discontinuity of wind or pressure was detectable in the weather charts of the noon (Figure 1b). Only the distribution of temperature and dewpoint roughly showed the position of the remaining 100–200 km wide frontal zone.

In higher levels the frontal zone extended along the line Rhone-valley – southwestern Germany – eastern Germany and Poland (comp. Figure 2). The jetstream in the upper troposphere had the same position. It is noticeable that the frontal zone was best developed between 850 and 700 hPa and vertically reached up to 600 hPa only. Above that level the baroclinicity was much weaker than below and no defined frontal zone existed. Consequently the biggest vertical wind shear occurred in the levels between 850 and 600 hPa. Remarkable is the great temperature gradient on the northwestern slope of the Alps at 850 hPa. It resulted from the additional heating which the air experiences within a mountainous area during a clear summer day. From the observations of the synoptic stations we can deduce that in our case the air in the alpine region was in 850 hPa more than 5 K warmer than at the same level over the neighbouring plains.

Considering the distribution of geopotential and wind of the various pressure levels in Figure 2, we find already in 850 hPa a very different picture compared with the surface map. Instead of the ridge

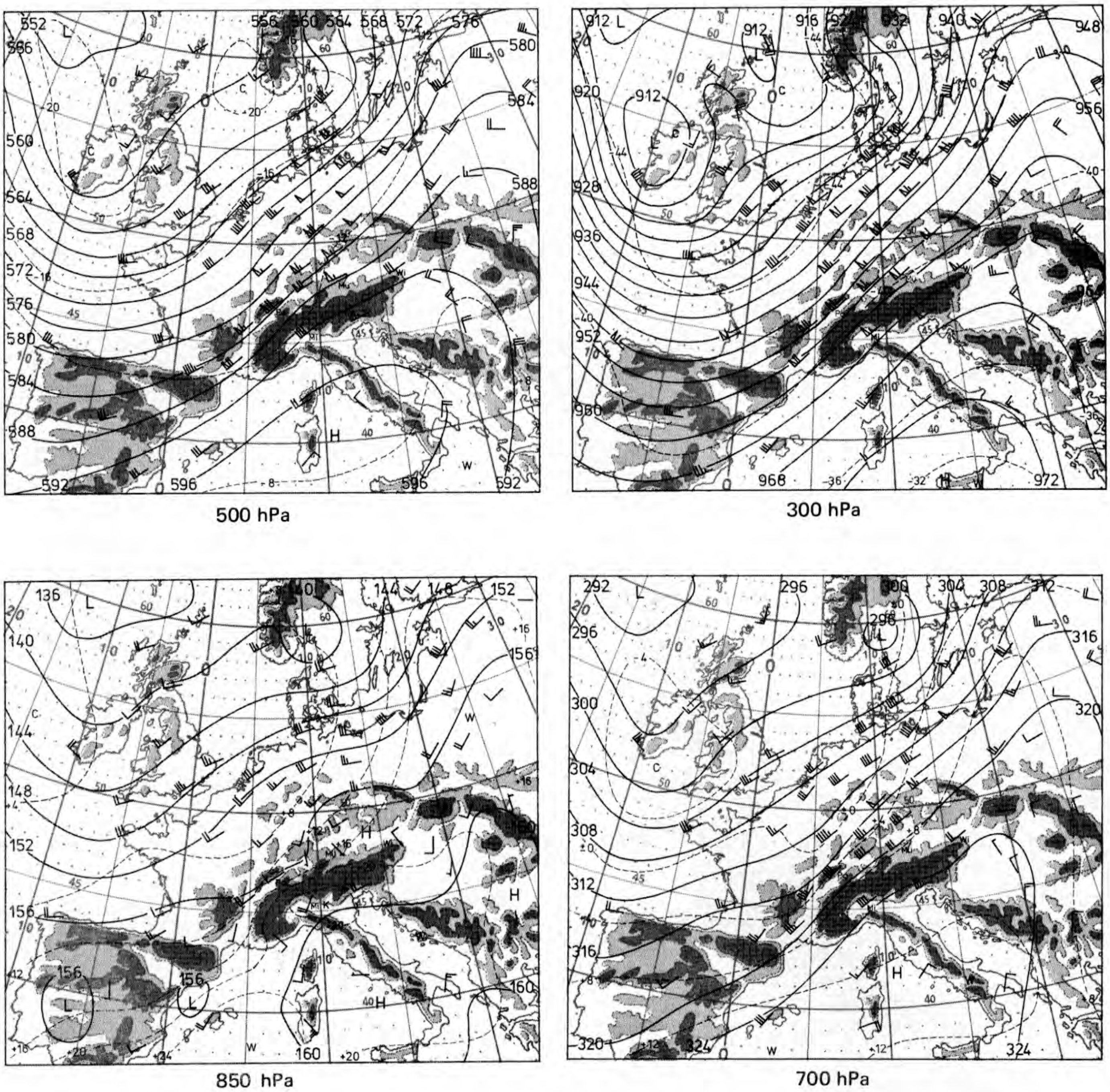


Figure 2 Absolute topographies from July 12, 1984, 12 UTC with isohypses (solid lines, in gpdm), isotherms (dashed lines, in °C) and observed winds; W warm, C cold; Mü Munich, Pa Payerne, Mi Milano

of high pressure across Germany and the low pressure channel over the alpine region there was a generally southwesterly flow over Central Europe with an accentuated wavy structure near the Alps. A ridge over eastern Austria and the CSSR was followed by a trough with axis over the eastern parts of Switzerland. The development of this trough possibly had its origin in an orographic effect initiated by the strong southerly winds blowing over the mountains – as described in KURZ (1982). Ahead of the trough which moved along the Alps eastwards, the warmer air was again transported a little further to the north whereas behind it weak cold air advection took place.

In higher levels (700–300 hPa) the flow showed another structure: Embedded in the vertically increasing southwesterlies we find two troughs – a well pronounced one over southern Scandinavia and another rather weakly developed one over eastern France, both separated by an area with anticyclonic curvature. In 700 and 500 hPa another trough was visible over the eastern parts of the Alps and north-east of it, respectively.

3 Vertical Profiles of Temperature and Dewpoint

A comparison of the upper air data of Munich of 00 and 12 UTC (Figure 3) very slearly shows the combined effect of the above described air mass change, the simultaneously working descending motion and the vertical mixing within the boundary layer. The nightly radiosounding had been started just after the passage of the front. Besides a shallow layer near the surface the sounding represented the hot and dry tropic air mass. The vertical temperature profile still showed a value of 30 °C above the surface inversion. Above, the temperature dropped with an average lapse rate of 7.8 K/km to − 60.7 °C in the tropopause level at 12.6 km MSL.

The intrusion of the new air mass behind the front caused a cooling by 10–12 K within a mixed layer between the surface and 2200 m MSL until noon, where a well marked inversion with a 4 K temperature rise had formed. Above the inversion the cooling decreased to 5 K and less and became zero at about 5 km MSL. Comparisons with other ascents show that the inversion did not lay horizontally but rose with a mean slope of 1 : 300 to the northwest decreasing its stability in doing so. This means that the inversion was a part of the sloping frontal zone mentioned above (comp. Figure 2).

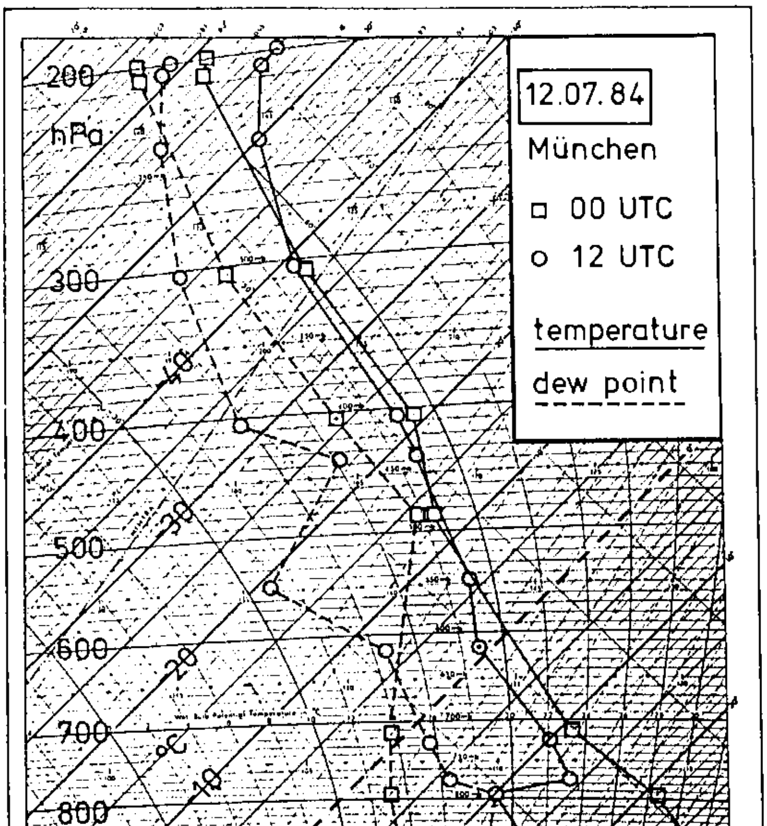


Figure 3
Soundings of Munich from July 12, 1984, 00 UTC and 12 UTC, with temperature (solid lines) and dewpoint (dashed lines) in a tephigram

Particularly important for the further development is the fact, that the incoming cooler air was substantially moister than the hot and dry air masses ahead and over the surface front. Thus the dewpoints increased by 5 K or more up to a height of 4 km, so that a drastic increase of the relative humidity resulted with the decrease of the temperature. Just below the inversion the air was saturated. 12 hours before a relative humidity of only 30 % had been measured in that level!

With the inversion the lapse rate was stable with respect to convection released by heating of the air near the ground; accordingly only some shallow cumulus developed until noon at the top of the moist layer. Nevertheless, merely four to six hours later, numerous thunderstorms occurred in southern Bavaria, among them also very intensive convective processes like the Munich hailfall.

Since there are no obvious explanations it seems useful to investigate the destabilizing effects which might be responsible for the heavy convective activities. Three possible mechanisms of destabilizing will be discussed in order to find a satisfactory answer:

Possibility 1: Dissolution of the inversion by surface heating

Convective processes could develop up to higher levels if the air below the inversion were heated up to a potential temperature matching the value just above the inversion. Near the ground a temperature increase of up to more than 30 °C would have been necessary, meaning an energy input of about 16 MJ/m^2. Consider the few hours between the sounding and the first thunderstorms in Bavaria this value presumes an energy flux of 900 W/m^2, i.e. 2/3 of the solar radiation income. According to heat balance measurements near Munich in a comparable season (BERZ, 1969), the maximum values of the sensible heat flux into the atmosphere range only between 250 and 300 W/m^2. Thus, the possibility of local dissolution of the inversion by heating from below can be rejected. In fact maximum temperatures of 26.9 °C and 27.8 °C were measured at the airport and the center of Munich, respectively.

Possibility 2: Advective modification of the stratification

A warm air advection below the inversion as well as a cold advection above it could have another way to destabilize the atmosphere. Considering the wind and temperature distribution in the absolute topographies of Figure 2, we find indeed warm air advection in 850 hPa – ahead of the above mentioned trough north of the Alps – and some cold air advection in 500 hPa. This statement is confirmed by the inspection of the wind sounding of Munich of 12 UTC which showed a marked veering between 1 and 3 km altitude and some backing in heights about 6 km. But with an amount of at most 0.2 Kh^{-1} the vertically different temperature transports were so small, that a significant change of the lapse rate could not have taken place within a reasonable time. The 18 UTC wind sounding of Munich still showed weak warm air advection up to a height of 4 km – thus as well below as above the inversion layer. Altogether the described horizontal advection could have contributed to a destabilization of the mid troposphere but not to the removal of the inversion.

Possibility 3: Destabilizing by forced lifting of the air column

A forced lifting of a whole air column can also lead to an unstable stratification. Often just a small lift of a few hundred meters destabilizes the air as soon as the stratification is potentially unstable (PALMEN and NEWTON, 1969). This condition occurs if the relative humidity is high in the lower levels and low in the upper levels. During a lift of the entire air column the dynamic condensation level will be reached earlier in the lower layers as it will be reached in the higher layers, i.e. more latent heat will be released in the lower portion. As a consequence the vertical gradient of the potential temperatures decreases, and can even become negative. This process allows a total destabilizing of the air up to high altitudes within a short time.

A necessary and sufficient condition for the potential instability is the decrease of the equivalent-potential temperature with the height. This measure of both sensible and latent heat energy is used in the present paper to check the stratification upon potential instability.

Figure 4 shows the vertical profiles of the potential and the equivalent-potential temperature of the Munich sounding at 12 UTC. The equivalent-potential temperature decreases by 12 K between the surface and 5100 m MSL. Thus, the lower half of the troposphere was potentially highly unstable. It is remarkable that the inversion, which is well marked in the profile of the potential temperature, does almost not appear in the equivalent-potential temperature profile.

By graphic construction in the tephigram (comp. Figure 5) it may be assumed that in the case of the Munich sounding a lifting of about 100 hPa was sufficient to remove the blocking inversion as well as to reach saturation in all levels. Beyond it the lapse rate would be unstable for saturated air between 3 and 6 km height, so that the primarily working lifting – forced by any mechanism – could pass into an accelerated convective flow and the cloud air coming up from the lower moist layers could ascent freely up to high altitudes.

Some different possibilities may be considered as forcing mechanisms for the lifting of the potentially unstable air – macro-scale-lifting connected with synoptic features like baroclinic waves, meso-scale-

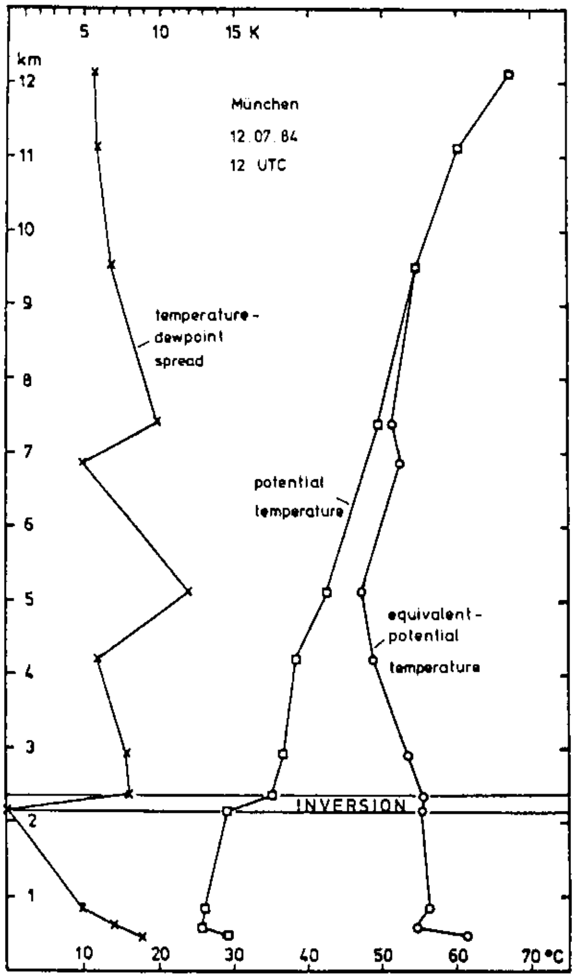


Figure 4 Vertical profiles of potential temperature, equivalent-potential temperature and temperature-dewpoint spread over Munich on July 12, 1984, 12 UTC

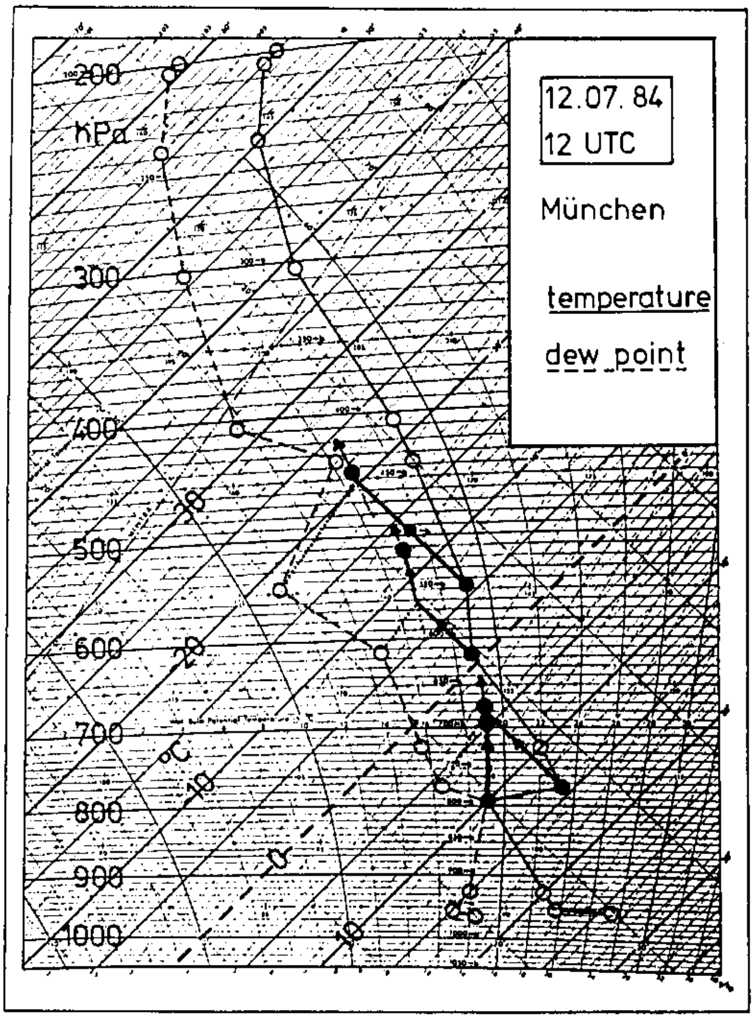


Figure 5 Behaviour of lifted air starting from four different levels based on the vertical profile over Munich on July 12, 1984, 12 UTC. Starting and final points of lifts by 100 hPa are indicated by full circles.

lifting in the vicinity of fronts or frontal zones, forced upslope motion on mountains or finally the strong ascending motion within the updraft of a pre-existing and well-developed convective system or complex. Since the pressure interval of 100 hPa corresponds to a height interval of about 1000 m in the inversion level of the Munich sounding, a duration of about six hours would have been necessary for the release of the instability with a macro-scale ascent of, say, 5 cm/s. The same is valid for orographically forced vertical motions. With a frontal lifting of 50 cm s^{-1}, a little more than half an hour would suffice, and in a convective updraft with 5 m s^{-1} vertical velocity it would take only some minutes to destroy the inversion and reach saturation in the levels above.

To prove whether the described process has been responsible for the surprising formation of the hail storm or not, we proceed from the one-dimensional consideration to horizontally two-dimensional field analysis.

4 Horizontal Distribution of Thermodynamic Variables

In this section the horizontal distribution of some thermodynamically relevant variables will be discussed, as they have been derived from the 12 UTC radiosoundings and from synoptic stations in the Alps and other mountains.

The Figures 6a and 6b show the dewpoint depression $T - T_D$ as a measure of relative humidity. At the 850 hPa level we find a great contrast between both sides of the Alps. This reflects the already mentioned existence of two air masses, the moist and relatively cool air coming in behind of the front and lying now north of the Alps and in the Rhone-valley, and the dry and hot air in the alpine region and south of the mountains. In 700 hPa the distribution of humidity was somewhat different: At this level the tongue of the moistest air extended from the Western Alps northeastwards and was situated just inside the frontal zone. The high values of dewpoint spread over northern Germany at this level were a result of the large-scale subsidence at the rear of the Scandinavian trough.

The equivalent-potential temperatures at 850 hPa (Figure 7a) have their highest values within the Alps. They result from relatively high air temperatures combined with lower moisture (comp. Figures 2 and 6a) and reflect the above mentioned additional heating of the air in a mountainous region during a summer day. The same is valid for the high values over the Apennines. Besides of that the distribution clearly shows the air mass differences. In contrast to the real temperatures the moist air north of the Alps had a much higher equivalent-potential temperature than the very dry air over Italy. This high equivalent-potential temperature indicates the tropical origin of the moist air mass. Comparisons with the days before show that this mass stemmed from the western Mediterranean (xT after „Berliner Wetterkarte"). It was originally warmer and drier but had been cooled and moistened through lifting and evaporation of falling rain in the active parts of the frontal zone.

The drop of the equivalent-potential temperature further northwestwards points to the existence of a third air mass, namely cold air from the Atlantic (mPs after „Berliner Wetterkarte") which had been incorporated in the frontal zone during the day before.

A deep potentially unstable stratification would be indicated by a decrease with the height of the equivalent-potential temperatures between 850 and 500 hPa. Since at 500 hPa this temperature showed relatively small variations according to the weak baroclinicity there and reached only maximal values around 50 °C, the difference map in Figure 7b largely reflects the 850 hPa distribution. We find negative differences within a zone covering the northern part of the western Mediterranean, the Rhone-valley, the Alps, Switzerland, Southern Germany, Austria and Czechoslovakia. In that area deep convection should have been possible if the instability would have been released through lifting.

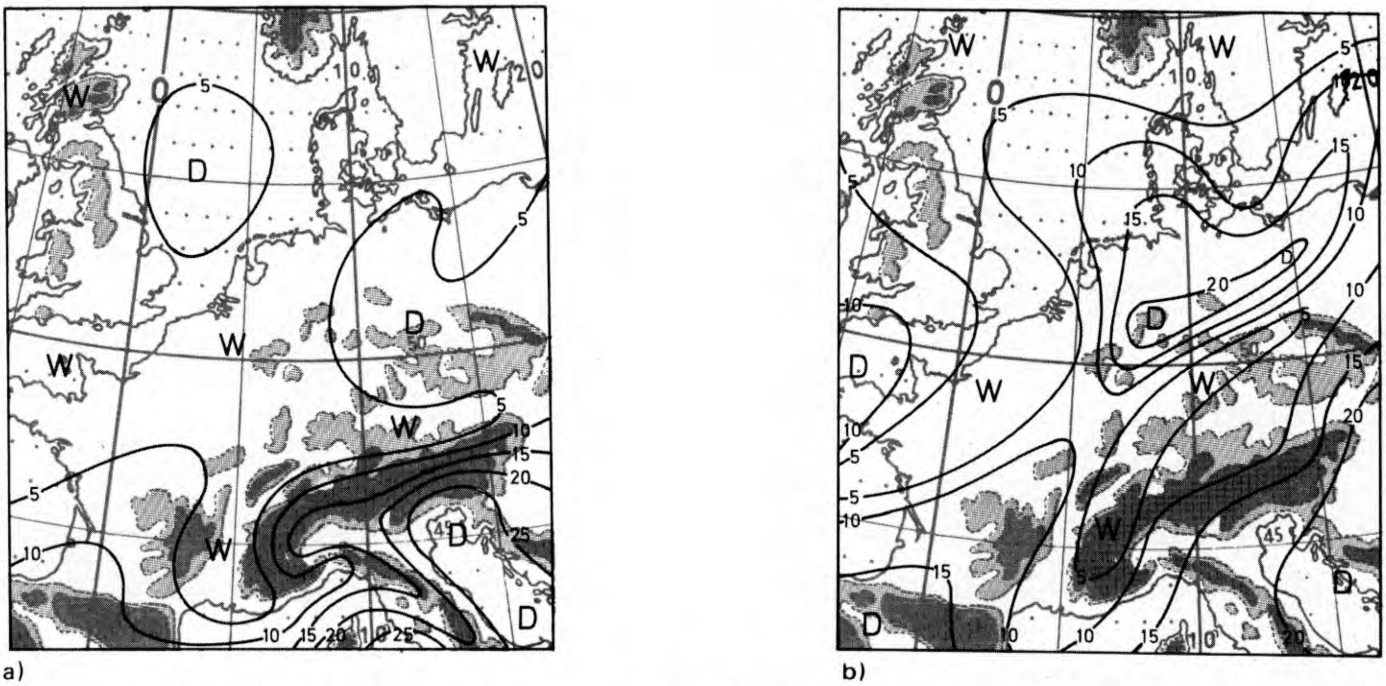


Figure 6 July 12, 1984, 12 UTC: Temperature-dewpoint spread in 850 hPa (a) and 700 hPa (b) in K; W wet, D dry

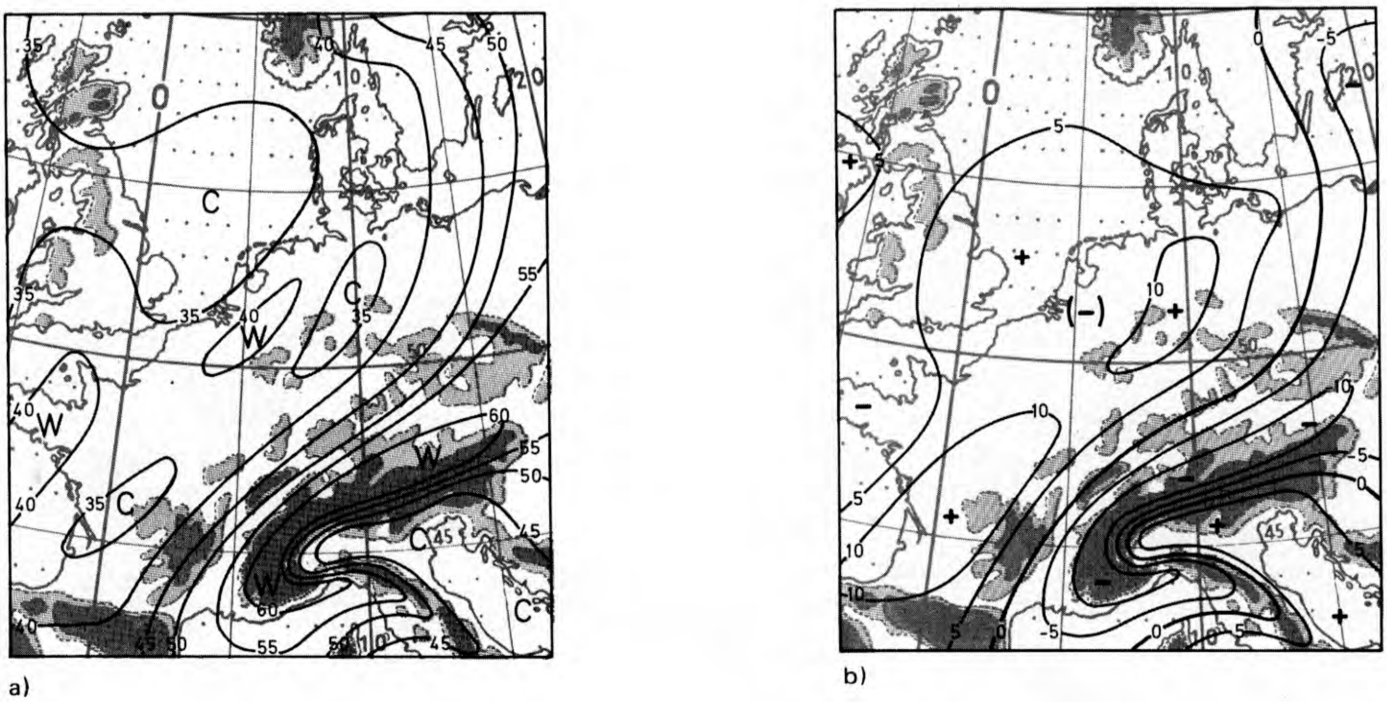


Figure 7 July 12, 1984, 12 UTC: (a) Equivalent-potential temperature in 850 hPa (in °C); W warm, C cold
(b) Difference of the equivalent-potential temperatures in 500 hPa and 850 hPa (in K)

The detailed inspection of the various soundings confirms these results but shows also important differences. In all available soundings a stable layer was present – in the east (Vienna) a well developed inversion as over Munich, in the west an isothermal layer only (Payerne, comp. Figure 8). Further to the west (Rhone-valley) the colder air from the Atlantic had penetrated into the lower layers, so that the

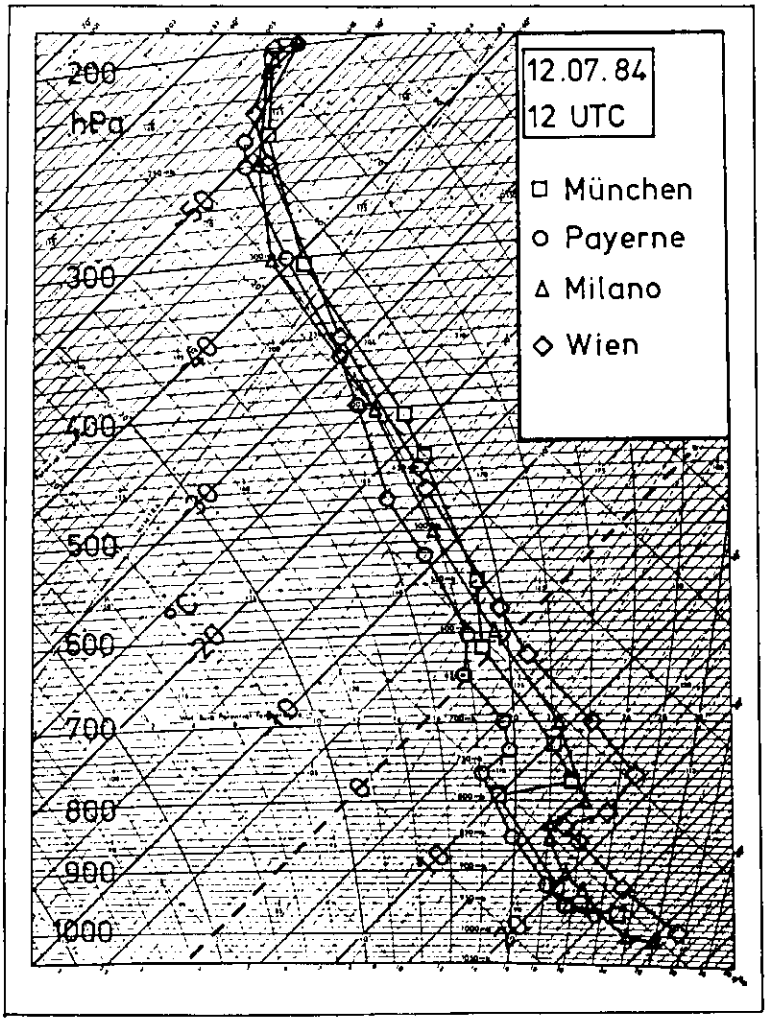


Figure 8
Soundings of Munich, Payerne, Milano and Vienna on July 12, 1984, 12 UTC with temperature profiles

potential instability was restricted to levels in the mid troposphere. Moreover the release of instability was rendered more difficult there through the existence of layers with vertically increasing equivalent-potential temperatures between unstable layers. Really similar conditions like over Munich showed the soundings of Neuhausen ob Eck (just north of the Lake of Constance) and Vienna, only.

As a result we can conclude that the best conditions for severe convective activities prevailed in the area north of the Alps between the mid of Switzerland and eastern Austria bordered at the north by the Danube. This distribution is by no means fortuitous but clearly shows the influence of the Alps. The mountain barrier prevented the further movement of the moist air invading behind the front. This mass laid in the lower levels north of the Alps and formed with its high energy supply the ideal source for convective overturnings – if the instability would have been released.

As the influence of the Alps is concerned, another important point is remarkable. The mountainous region acts – as repeatedly mentioned – during a clear summer day as an elevated heat source, so that the air around the mountains becomes warmer than the air in the free atmosphere further away. Since the heating decreases with height a destabilizing effect results. The above mentioned stable layer which hindered the release of instability was therefore surely weaker developed in the neighbourhood of the mountains as over the plains, or even dissolved. For that reason the best conditions for a quick beginning of the convective overturnings were given directly at the northern flank of the Alps. But in any case a trigger for the release was necessary.

5 Kinematic and Dynamic Fields

What mechanism could have been the trigger for the release of potential instability in the afternoon of July 12, 1984? Since no surface front existed, frontal lifting could not have been the case. The same applies to lifting within a pre-existing convective system. Some orographic effects might be the cause - the ascending motion at isolated mountain slopes, within the valley wind system or – on a bigger scale – within the northerly current against the Alps and into the above mentioned heat lows. Forced lifting could also result from the strong southerly winds in height of the mountain tops. But all these possibilities do not suffice as triggering effects, since the subsequent release of instability occurred at one place only near the Alps and not everywhere along it. Therefore we have to suppose that a macro-scale lifting was responsible for the initial formation of deep convection – may be in interaction with orographic effects.

5.1 Theoretical Background

The diagnosis of the macro-scale vertical motions can be derived from the distribution of the geopotential using the omega-equation, which is described by HOLTON (1972) for instance.
The omega-equation

$$\left(\nabla_h^2 + \frac{f_0^2}{\sigma_0}\,\frac{\partial^2}{\partial p^2}\right)\omega = \frac{f_0}{\sigma_0}\,\frac{\partial}{\partial p}\,[\mathbb{V}_g \cdot \nabla \eta_g] + \frac{1}{\sigma_0}\nabla^2\left[\mathbb{V}_g \cdot \nabla_h\left(-\frac{\partial\phi}{\partial p}\right)\right] \qquad (1)$$

is a three-dimensional diagnostic Poisson-equation, where ω is the vertical velocity in the pressure coordinate system

$$\omega \equiv \frac{dp}{dt} \qquad (2)$$

The right hand side of (1) depends only on the three-dimensional variation of the geopotential ϕ.
Here

$$\mathbb{V}_g = \frac{1}{f_0}\,\mathbb{K} \times \nabla_h \phi \qquad \text{with} \qquad f_0 = 10^{-4}\ s^{-1} \qquad (3)$$

is the geostrophic wind vector,

$$\eta_g = \frac{1}{f_0}\nabla_h^2 \phi + f \qquad \text{with} \qquad f \text{ Coriolisparameter} \qquad (4)$$

is the absolute geostrophic vorticity, and $(-\partial\phi/\partial p)$ the thickness which is a measure for the local temperature. σ_0 is a stability factor which is defined to be

$$\sigma_0 = \overline{\frac{1}{\theta}\,\frac{\partial\theta}{\partial p}\,\frac{\partial\phi}{\partial p}} \qquad (5)$$

within the particular column.
As forcing functions for vertical motions the vertical variation of vorticity advection and the horizontal distribution of thickness advection act on the right-hand side of the omega-equation. A qualitative interpretation yields that ascending motion is favoured by either a vertical increase of positive vorticity advection or relatively strong warm air advection or both. Descending motion is to be expected in a region with vertically increasing negative vorticity advection or maximum cold air advection or both.

The solution of the omega-equation at a given point of time requires the knowledge of the three-dimensional variation of the geopotential. The data base was restricted to the reported geopotential heights of four main levels, 850, 700, 500, and 300 hPa. In order to process them numerically they have been interpolated to a horizontal grid on each of the main pressure levels. The grid was chosen as a stereographic projection plane with 15×18 grid points with a spacing of 120 km in each direction. The whole domain covers an area of $1800 \times 2160\ km^2$ (see Figures 9 to 12).

All horizontal derivatives can now be approximated by central differences over two grid points, i.e. 240 km. Vertical gradients must be calculated in the pressure system in a similar manner, i.e. using central differences over two layers (350 or 400 hPa).

The interpolation from station data to the grid was made by hand. In this way it was possible to obtain the grid values in an efficient manner, where additional information (like temperature and wind) could be incorporated in order to avoid the infiltration of incorrect geopotential values into the grid, and to fill out sparse data regions with high reliability.

The solution of the omega-equation in form (1) is not possible with the reduced data base. The knowledge of ϕ on just four pressure levels is insufficient to apply realistic lower and upper boundary conditions. Therefore a shortended form of the omega-equation has been used in this paper. A wave solution for (1) described by HOLTON (1972)

$$\left(\nabla_h^2 + \frac{f_0^2}{\sigma_0}\frac{\partial^2}{\partial p^2}\right)\omega = \left[-(k^2 + l^2) - \frac{1}{\sigma_0}\left(\frac{f_0\,\pi}{p_0}\right)^2\right]\omega; \quad p_0 = 10^5\ \mathrm{Pa} \tag{6}$$

suggests that a condition

$$\frac{\partial^2 \omega}{\partial p^2} = 0 \tag{7}$$

can be considered as far as

$$k^2 + l^2 \gg \frac{1}{\sigma_0}\left(\frac{f_0\,\pi}{p_0}\right)^2 \quad . \tag{8}$$

k and l are the wave numbers in x- and y-direction. The relation (8) is fulfilled, if $\sigma_0 > 10^{-6}\ m^2\ s^{-2}\ Pa^{-2}$, and the horizontal wave length does not exceed 2000 km. Since the actual value of σ_0 is $2.6\ 10^{-6}\ m^2\ s^{-2}\ Pa^{-2}$, and the domain extension of 2000 km does not allow the resolution of longer waves anyway, the condition (7) is fairly acceptable. Now it is possible to gain the horizontal distribution of the vertical motion at least on the 700 and 500 hPa-level. But we have to keep in mind that (7) actually implies a constant vertical gradient of ω on these levels. Nevertheless, Equation (1), reduced to a twodimensional Equation by (7), was applied to both the 700 hPa and the 500 hPa level. The problem was solved by using the iterative Gauss-Seidel-procedure. All calculations could be performed on a personal computer.

5.2 Discussion of the Results

5.2.1 Vorticity and vorticity advection

Starting with the distribution of the geopotential (Figure 2) the geostropic vorticity was calculated first. The fields of this property are presented in Figure 9. They very clearly show the wave structure of the currents at the various levels, which was described in Section 2.

In 850 hPa we find a SW-NE orientated dipol of positive and negative vorticity associated with the small trough near the Alps. It is noticeable that the calculated figures inside the Alps are highly fictitious since the winds in the mountainous region are not at all geostrophic.

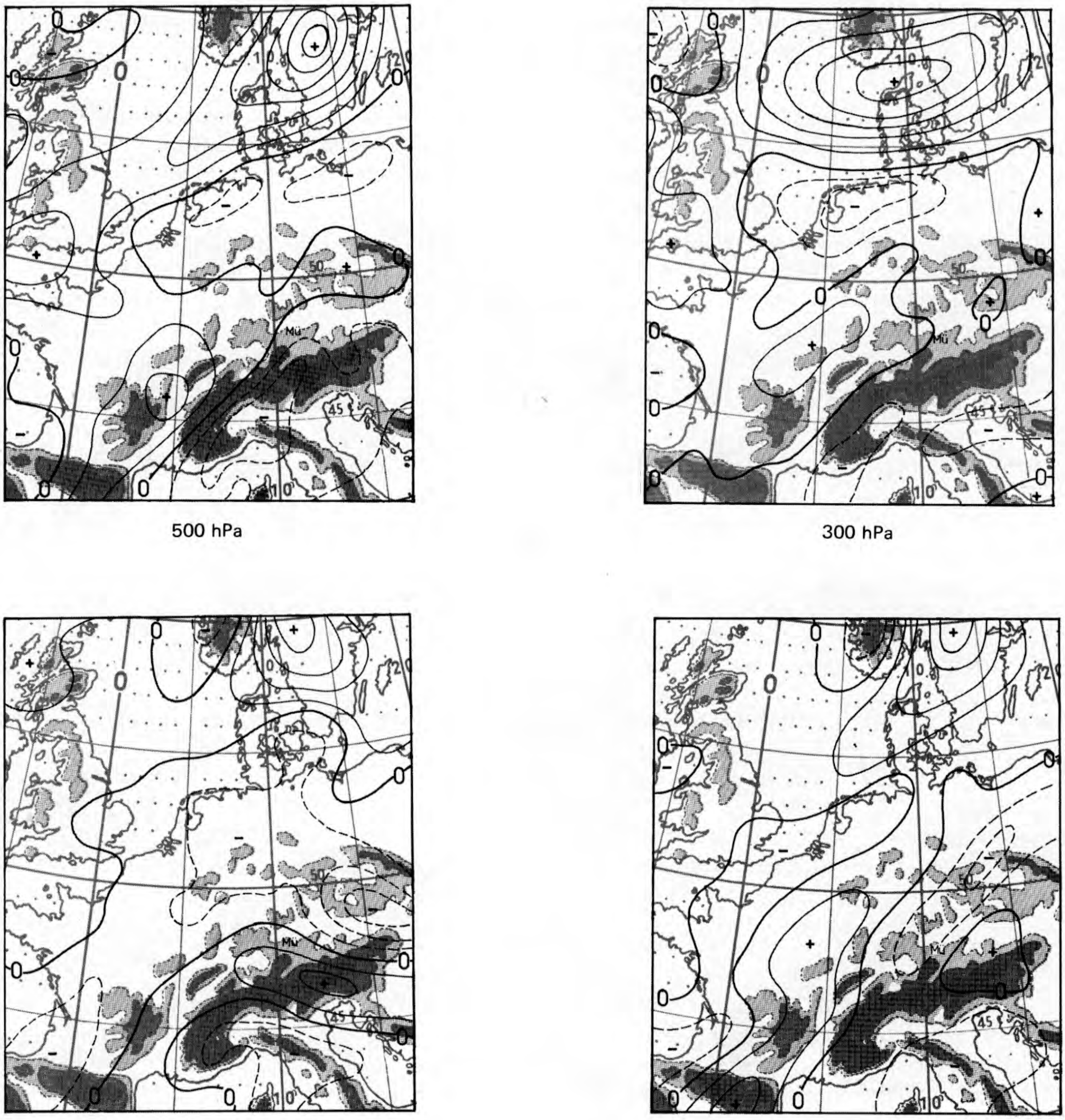
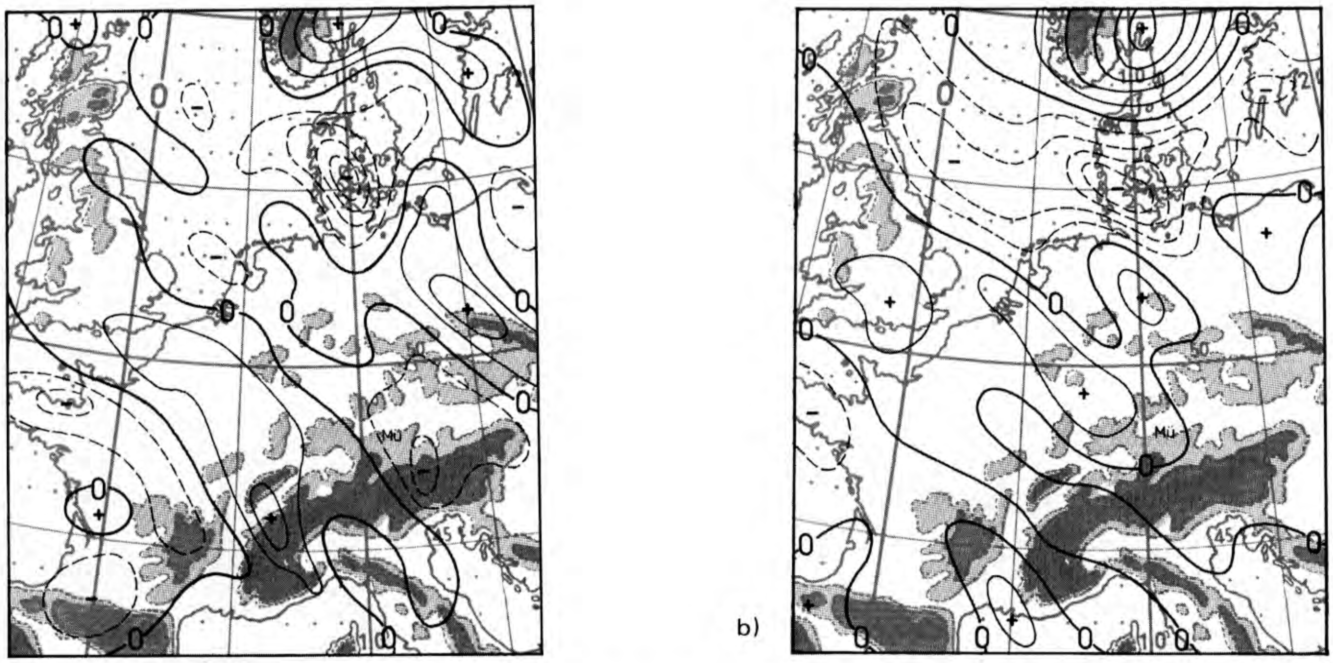


Figure 9 July 12, 1984, 12 UTC: Relative geostrophic vorticity in 850, 700, 500, and 300 hPa; contour intervals: $2.5 \cdot 10^{-5}\,s^{-1}$



a) b)

• **Figure 10** July 12, 1984, 12 UTC: Mean vertical gradient of geostrophic advection of absolute geostrophic vorticity between 850 hPa and 500 hPa (a) and between 700 hPa and 300 hPa (b); contour intervals: $5 \cdot 10^{-14} s^{-2} Pa^{-1}$

At higher levels there are main maxima of the vorticity over southern Scandinavia and eastern France separated by a region of negative vorticity over northern Germany. This reflects the two troughs and the region of anticyclonic curvature between them embedded in the southwesterly basic current. Around the trough lying over France a big portion of the vorticity is generated by the horizontal shear at the flanks of the jetstream. This is the reason for the large gradient of the vorticity over the western parts of the Alps especially in 500 hPa but in 300 hPa, too.

Ahead of the above mentioned troughs positive advection occurred, behind it negative advection of vorticity. Due to the vertical increase of the southwesterly current also the vorticity advection increased with the height (comp. Figure 10). This is typical for baroclinic waves with nearly upright axes. In our case the vertical shear of the wind was strongest in the layers between 850 and 600 hPa, so that the increase of the vorticity advection had also its maximum there.

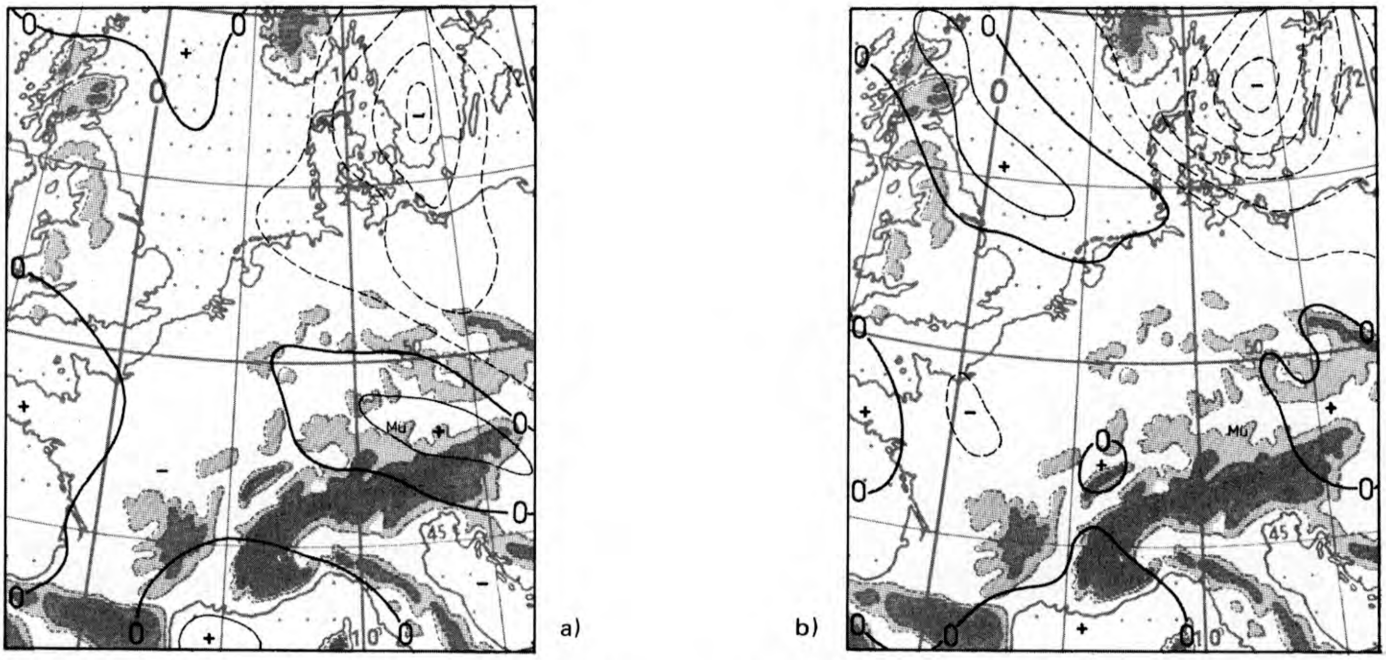
A contribution to ascending motions resulted over the western part of the Alps, Switzerland and eastern France, whereas a drive for descending motions was found especially over the southwestern half of France and southern Scandinavia.

5.2.2 Thickness advection

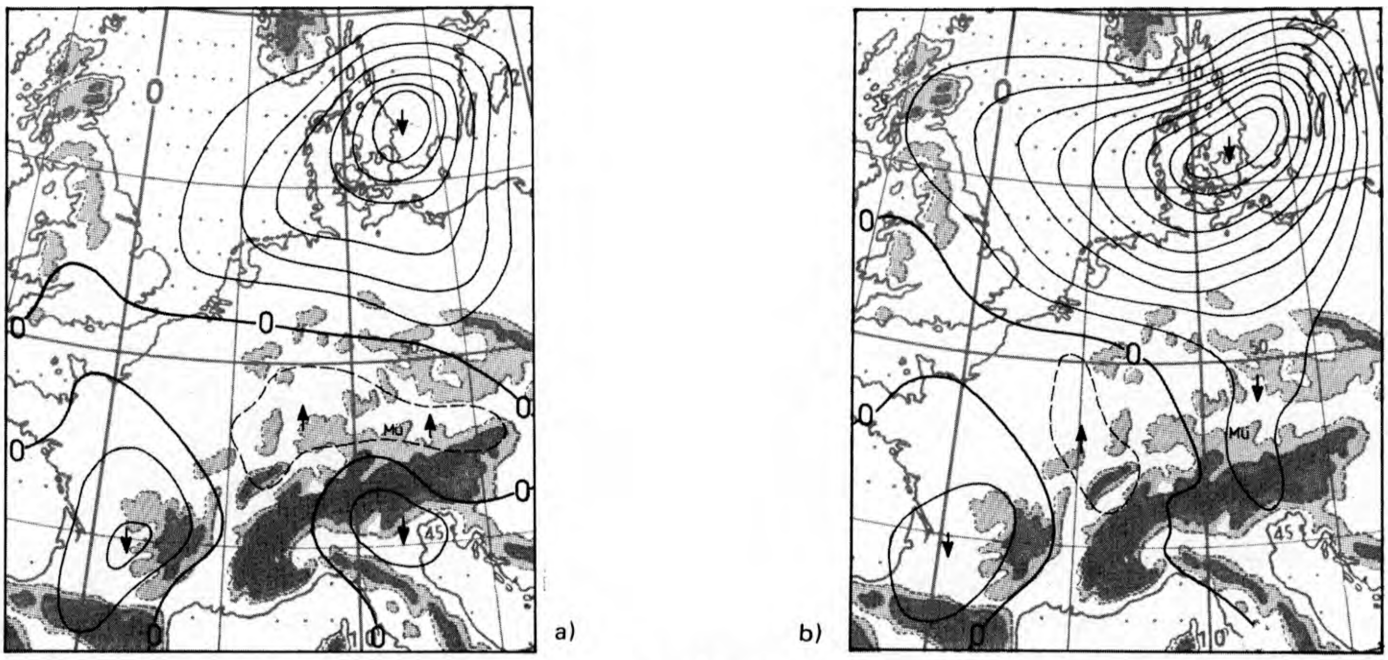
As the thickness advection between 850 and 500 hPa and 500 and 300 hPa, respectively, is concerned, the main feature was the strong cold air advection south of the low or trough over southern Scandinavia (Figure 11). We can expect a forcing for a descending motion in this region. Whereas in higher levels the thickness advection over mid Europe was generally weak, we find in the lower troposphere the already mentioned warm air advection north of the Alps and over its eastern parts. This warm air advection contributes to an upward motion.

5.2.3 Vertical velocities in 700 and 500 hPa

The combination of both forcing functions in the omega-equation yields the vertical velocities for 700 and 500 hPa presented in Figure 12. The dominant feature is the big area with partly strong descending motion over southern Scandinavia reaching southward to the mid of Germany in 700 hPa,



- **Figure 11** July 12, 1984, 12 UTC: Geostrophic thickness advection in 700 hPa (a) and 500 hPa (b); contour intervals: $2.5 \cdot 10^{-7}\,m^{-2}\,s^{-3}\,Pa^{-1}$



- **Figure 12** July 12, 1984, 12 UTC: Vertical velocity in 700 hPa (a) and 500 hPa (b); contour intervals: $5 \cdot 10^{-2}\,Pa\,s^{-1}$

but to the Alps in 500 hPa. Descending motions result also over the southwestern half of France and – within an isolated area – south of the Alps in 700 hPa.

Especially important for our reasoning are the areas with ascending motion over eastern France, Switzerland and the western Alps in both levels – there mainly generated through vertically increasing positive vorticity advection – and over the southern half of Germany in 700 hPa – there generated through the warm air advection in the lower troposphere. With this result we have found the seeked macro-scale lifting which should have been the triggering mechanism for the initial release of potential instability in the next hours.

The maximum value of ω within the zone with ascending motions north of the Alps at 700 hPa is almost 100 mPa s^{-1}. This corresponds to a vertical velocity of about 1.5 cm s^{-1}. That amount may appear too small to cause the necessary changes of the vertical distribution of temperature and moisture for the release of instability in a reasonable time. Thinking thus, we would overlook that because of the many omissions and simplifications the omega-equation in the form of (1) at best can yield a qualitatively but surely not quantitatively correct picture of the macro-scale vertical motions. In this respect the result is very satisfying.

6 The Weather Situation during the Afternoon of July 12, 1984

As a result of our diagnostic calculation we should assume a slow lifting of the air mass in the levels up to 500 hPa over Bavaria already at noon. Nevertheless, no cloud formation or even release of stronger convection was directly linked with this process at that time or during the next hours. But the warm air advection which caused the ascending motion in that area led simultaneously to a strengthening of the pressure fall near the ground, so that during the afternoon a small low developed in the region northeast of Salzburg. Combined with that one in 850 hPa a closed low formed within the trough which moved eastwards along the Alps. The 18 UTC-wind sounding of Munich still showed – as already mentioned – warm air advection in the lower troposphere.

In contrast to that, the diagnosed ascending motion ahead of the upper trough in the west made itself felt by middle and high clouds which at noon encroached into southwestern Germany from Switzerland and eastern France. According to the potentially unstable lapse rate in these levels the middle clouds showed partially castellanus-forms. The high moisture content in the middle troposphere (comp. Figure 6b) reflected the ascending motion and the cloud formation caused by it. The weaker development of the stable layer over Payerne compared with Munich or even Vienna might also be traced back to the destabilization through lifting. But until noon no precipitation or even thunderstorms were reported from the stations of the synoptic network.

Based on the described distribution of the vorticity advection the upper trough had to swing further on towards northeast. The upper wind measurements of 18 UTC allowed to recognize this movement and showed the axis of the through in 500 and 300 hPa approximately over Switzerland. Ahead of it an anticyclonic curvature of the streamline was visible – especially in 300 hPa. The jetstream ran with speeds up to 48 m/s in 300 hPa and up to 35 m/s in 500 hPa from the Rhone-valley over Switzerland and southern Germany northeastwards. Similar to noon the vertical wind shear was at most places largest between 850 and 500 hPa corresponding to the strongest baroclinicity in these levels.

As a result of the trough movement, the ascending motion over Switzerland and southern Germany probably became stronger during the afternoon, so that the possibility of the release of potential instability through lifting became better and better. However, in such a case the question of release or non-release of instability decisively depends on the amount of lifting and the span of time during which the lifting can work.

7 Development of the Storm

The hailstorm hit Munich between 17.50 and 18.30 UTC. But as we have learnt from ground observations, satellite image sequences and radar survey, the storm was not a local event.

As minutely shown by HÖLLER et al. (1986), the first precipitating cells – observed by radar – developed already a short time after noon at 12.35 UTC over the „Berner Oberland", i.e. over and on the main crest of the Swiss Alps. As discussed in Section 5 the release of instability in this region was easiest, and orographically induced lifting should have substantially contributed to the release.

It took nearly two hours until the lapse rate in the air over the plains north of the Alps was destabilized by the lifting so far that convective clouds could not only develop in the middle levels but also within the moist layer near the ground. After that the cells rapidly intensified and began to travel northeastwards with a mean speed of almost 100 km/h. At 15.13 UTC, Constance, as the first synoptic station, observed thunderstorms – but without any gusts and without hail in the precipitation. At around 16 UTC Laupheim, Memmingen, Leipheim and Ingolstadt were reporting thunderstorms, and one hour later the zone with cumulonimbus clouds, showers and thunderstorms reached up to the Bayerischer Wald (Bavarian Forest) in northeasterly direction.

After crossing the Lake of Constance, some precipitating cells changed their direction and continued propagating more towards east. This turning to the right of the mean wind direction is typical for big convective systems, so-called "supercell storms", which develop within a current with moderate to strong vertical shear and winds veering with height (see e.g. ATKINSON, 1981; comp. also HÖLLER et al., 1986). These systems consist of well developed updrafts and downdrafts. Within the downdraft rain cooled air from higher levels is transported to the ground where it pushes forward leading to convergence with the warmer air. This air is lifted and feeds the updraft. If this air is moist and potentially unstable new convective clouds can form, so that the system as a whole remains in a quasistationary state and moves – often with high speed – through the unstable air mass. The passage of such a system is at the surface accompanied by stormy wind, temperature drop, pressure rise, heavy precipitation and even hail.

All these phenomena were observed in southern Bavaria during the late afternoon and evening of July 12, 1984. Hail was reported for the first time shortly after 16 UTC from climate stations close to the Lake of Constance and east of Ravensburg. Then several hail swaths developed. The most intense one – 5 – 15 km broad and 250 km long – was connected with a super cell storm in which the hailfall propagated eastnortheastwards with a speed of 60 – 70 km h^{-1} and just crossed Munich.

During the late evening hours the convective system had become a large meso-scale convective complex (as defined by MADDOX, 1980) which travelled northeastward to the CSSR and Poland. Heavy rain intensities were reported also from these countries (65 l/m^2 in Svatouch/CSSR, 42 l/m^2 in Warsaw/Poland). During the next day the system moved towards the Leningrad area where it lost its intensity and vanished.

On July, 13 the frontal zone had propagated some hundred kilometers to the east, but the stability situation was still the same. Therefore the approach of a new trough with ascending motion ahead of it caused an almost identic convective development which also produced hail and ended in a meso-scale convective complex. The first cells were observed near Garmisch. They travelled across southeast Bavaria to Upper Austria, where precipitation amounts up to 50 l/m^2 were measured.

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