POLAR LOWS

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1. INTRODUCTION

The existence of mesoscale vortices embedded in cold air over open water has been known by western European meteorologists for a long time. The Bergen school meteorologists did not use the term polar lows, but described cases with regeneration on the occlusion (Bergeron 1949, 1954). According to T. Bergeron, a special, thermodynamic, nonfrontal kind of cyclone regeneration seems to take place in true tropical hurricane manner in certain extratropical regions (hurricane wind all round a small but deep low, the diameter of which is generally less than 500 km and the pressure less than 980 hPa at the centre). The Scandinavian meteorologists referred to the vortices as instability lows, indicating convective activity as the driving mechanism (Rabbe, 1975). The expression polar low was first introduced by British meteorologists to describe cold air depressions affecting the British Isles (Harley, 1960; Lyall, 1972). After 1960 the first satellite images clearly uncovered cyclones in the cold air behind or poleward of the polar front, (Anderson et al., 1969).

These early findings were followed up with papers on the nature and structure of the polar lows. Harrold and Browning's paper on "The polar low as a baroclinic disturbance" in 1969 is the first one in this sequence and marks a turning point in the research of polar lows. Their view was followed up by Mansfield (1974), Reed (1979) and Duncan (1977, 1978) who also discovered that polar lows often develop in a reversed shear flow, which in the Norwegian Sea means northerly winds and southerly thermal winds at the surface. The "baroclinic" view of these authors was challenged by Rasmussen (1977,1979) and Økland (1977), who both supported the older idea that polar lows are driven by organized, deep convection. In their arguments they made use of the conditional instability of second kind (CISK) introduced by Charney and Eliassen (1964).

In more recent years significant research efforts have been made, directed toward improving our knowledge of polar lows. Field experiments with detailed measurements from an aircraft was made in February 1984 (Shapiro and Fedor, 1986) and in February/March 1985 (Alaska Storms Experiment). Selected mesoscale storms were simulated (Seaman, 1983; Sardie and Warner, 1985, Grønås et al., 1986a, 1986b, 1987a) and in 1985 a fine-scale operational numerical model for polar low forecasting was introduced by the Norwegian meteorological institute, (Grønås et al., 1987b). Furthermore, special projects on polar lows were carried out in USA (Kellogg and Twitchell, 1986) and in Norway (Rasmussen and Lystad, 1987).

This review starts with a list of case studies made from different observation types, two examples are briefly described. Then a classification of polar lows into two types is attempted. Attention is laid on the type called "Arctic-front type", where the concept of potential vorticity thinking (Hoskins et al., 1985) is introduced to explain potentials for genesis of polar lows in the Norwegian Sea. This is followed up with a summary of theoretical studies and results from numerical simulations. Finally, the problem of forecasting polar lows is delt with. For further details, the recent book on polar and arctic lows edited by Twitchell, Rasmussen and Davidson (1989) is recommended.

2. CASE STUDIES

A relatively broad definition of a polar low has been given by Businger and Reed (1989): Any type of small synoptic- or subsynoptic-scale cyclone that forms in a cold air mass poleward of major jet streams or frontal zones and whose main cloud mass is largely of convective origin. Rasmussen (1989) discussed this definition and proposed to exclude small-scaled cyclones, with many features similar to extratropical cyclones, which are formed on secondary frontal zones north of the polar front.

The polar lows take place over data-void sea areas and the scale is mainly subsynoptic. In this way significant polar lows might not be verified by conventional observations. Nevertheless, several case studies are found in the literature. They may be divided according to the kind of observation systems used.

Most studies are based on conventional observations and satellite information. Such cases are often described as the low makes landfall and enters an area with a relatively dense network of surface observations. Examples are cases hitting Scandinavia with severe weather (Rabbe, 1975; 1987, Rasmussen, 1985a and b; Økland 1987), the Netherlands and Britain causing heavy snowfall (Harrold and Browning, 1969; Seaman et al., 1981) and the western coast of North America (Reed, 1979; Reed and Blier, 1986a and b; Forbes and Lottes, 1985; Locatelli et al., 1982). A few studies have satellite images as the primary source of information (Zick, 1983; Fett, 1989).

To demonstrate how conventional synoptic observations can describe a small-scale polar low, an example is taken from Økland (1987). In figure 1 two ordinary surface maps, with a time interval of six hours, show a polar low hitting Western Norway. Figure 2 shows a barograph record from a coastal station which was passed by the centre of the low. The graph indicates an inner core superimposed on a depression of larger dimension. The core passes the observing station in two hours, corresponding to a diameter of 130 km, reasonably in agreement with the maps. Radiosonde ascents from stations along the path of the low indicate deep convection to at least 500 hPa.

In a few case studies special mesoscale observation systems have been utilized, essentially with an aircraft as platform. Nearly all such studies on polar lows, using a research aircraft, have been made by scientists from NOAA, USA (Shapiro et al., 1987; Douglas and Shapiro, 1989; Shapiro and Fedor, 1989; Shapiro et al. 1989). The first polar low observed by an aircraft was the so-called ACE case from the American Cyclone Expedition in 1984 described by Shapiro et al. (1987). Figure 3 shows the background height field as given by the ECMWF analyses at 700 and 1000 hPa. An analysis of the surface pressure and winds 300 m above sea surface based on the aircraft data is given in Figure 4. The time is centred at one time level just after the time of the ECMWF analyses. A small cyclone with central pressure 980 hPa - 15 hPa less than in the largescale analysis - and surface winds up to 35 ms^{-1} is situated in the rear of the synoptic-scale low. The measurements showed a warm core structure with convection to 700 hPa, and the satellite picture Figure 5 shows a comma-shaped cloud pattern with the suggestion of an eye-like structure in the head. As pointed out by Shapiro (1986), this low was one out of several being formed in the area during this particular day. This is clearly shown by the cloud structures given in Figure 6.

Case studies based on numerical simulations have been made by Seaman (1983), Sardie and Warner (1985), Grønås et al. (1986a, 1986b, 1987 a and b), Nordeng (1987, 1990), Nordeng and Rasmussen (1992), Nordeng et al. (1989), Dell'Osso (1992). During the last decade successful simulations have been used for describing the development of extratropical cyclones (see e.g. Shutts, 1990). When an acceptable numerical simulation is obtained, the simulated storm evolution is taken to present the best estimate of the true development. However, the mesoscale structures of polar lows are a greater challenge to numerical models and unfortunately, simulations do not yet normally capture details of the developments. Especially, the pressure deficit connected to the inner core of the disturbance might be difficult to obtain.

All cases involved in the mentioned studies took place at high latitudes in the Northern Atlantic Ocean (including the North Sea, the Norwegian Sea and the Barents Sea) and the North Pacific. Similar mesoscale cyclones have been found in the Labrador Sea (Woetman Nielsen, personal communication) and the Beaufort Sea (Parker, 1989). Mesoscale lows, but with less convective activity, have also been reported from Antarctica (Heinemann 1990; Turner and Row, 1989; Bromwich, 1989). In addition to these high latitude developments, comparable mesoscale disturbances with extensive convective activity have been reported at lower latitudes (the

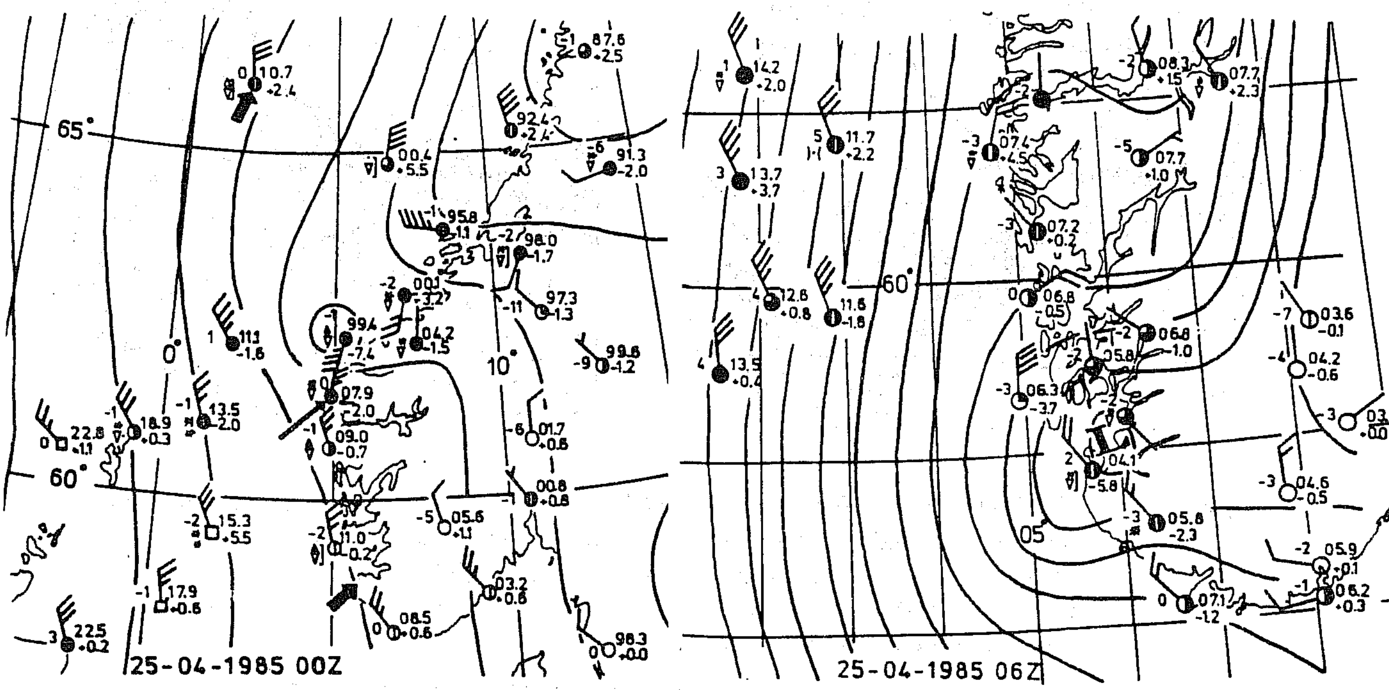

Figure 1. Left: Surface weather map for 25 April 1985 00 UTC. Bold arrows give position of RS-stations. Thin, longer arrow shows position of barograph record shown in Figure 2. Right: Surface weather map for 25 April 1985 06 UTC. The contour interval of MSLP is here 1 hPa. (After Økland 1986).

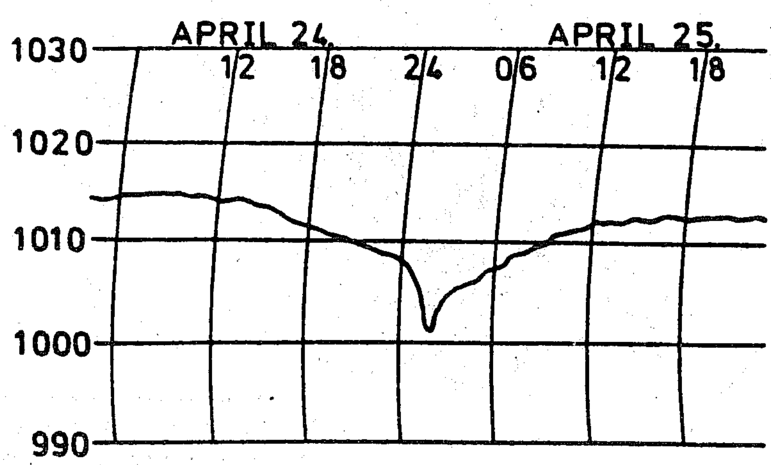

Figure 2. Barograph record showing the passage of the polar low shown in Figure 1. The location is shown in Figure 1. (After Økland 1986).

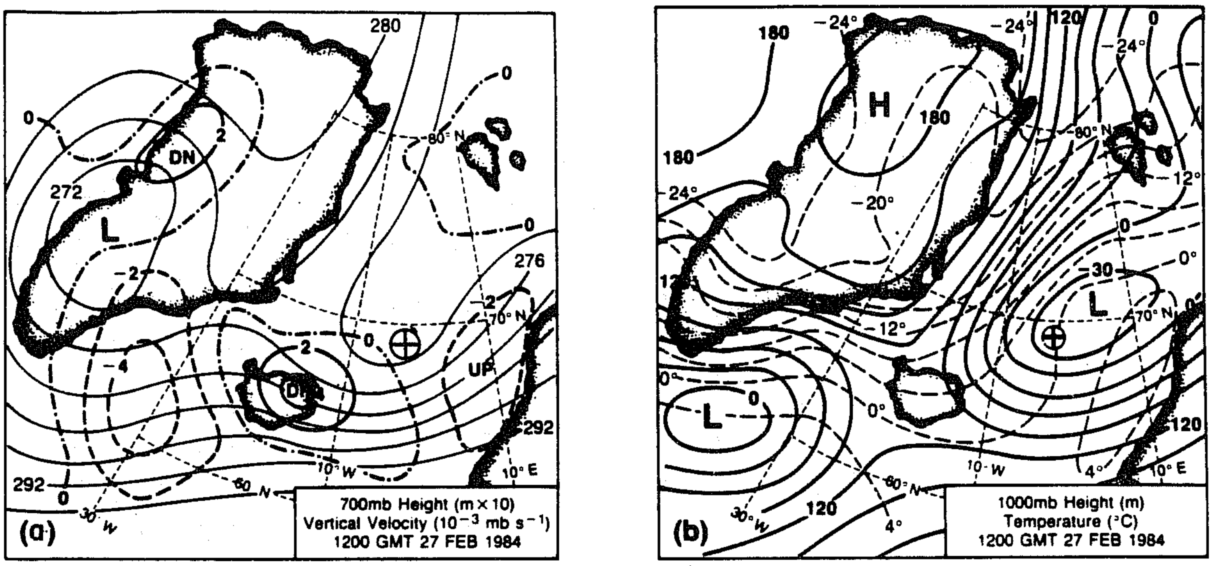

Figure 3. ECMWF analyses for 1200 UTC 27 February 1984. (a) 700-mb heights (thin solid lines in dm); and vertical velocity 10^{-3} mb s^{-1} (heavy dashed lines are positive, heavy dot-dash lines are negative). (b) 1000-mb heights (solid lines in m) and isotherms (dashed lines in degrees C). Circled cross marks position of the polar low (from Shapiro et al. 1987).

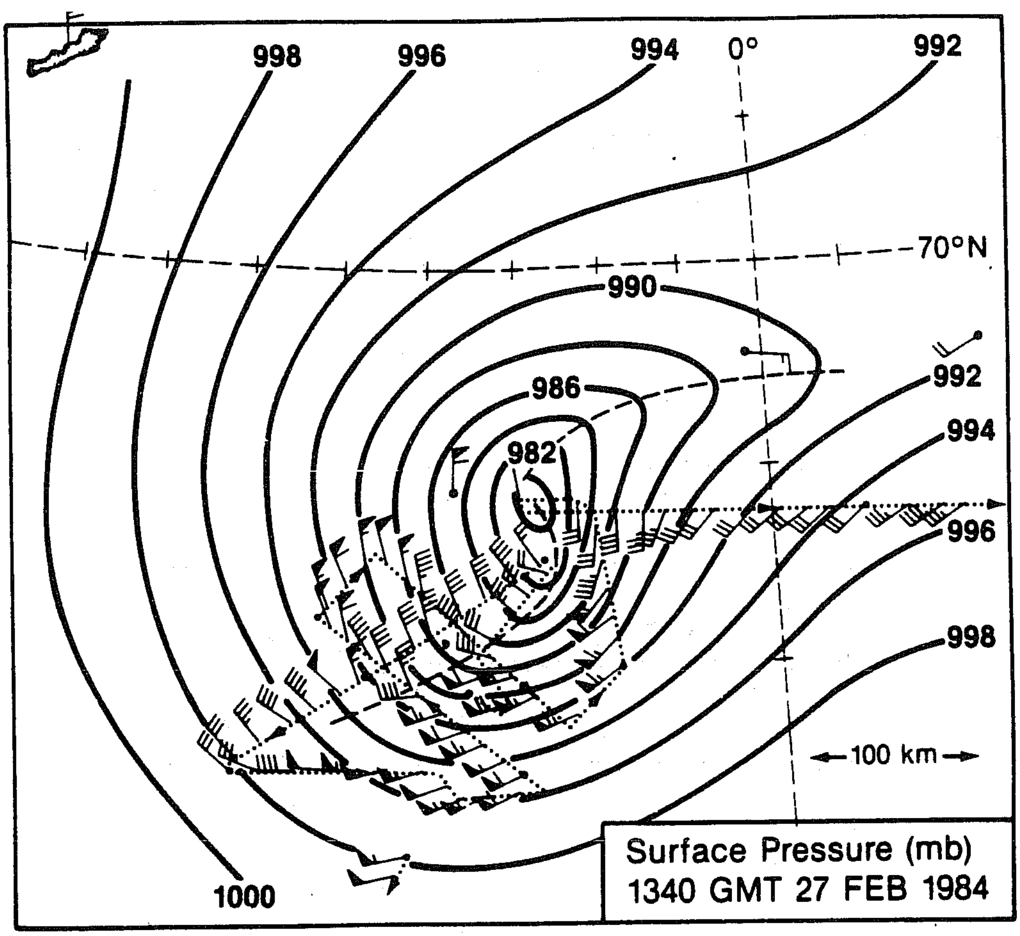

Figure 4. Surface pressure (hPa) and 300-m winds at 1340 UTC 27 February 1984. Full barb 5 ms^{-1}, penant 25 ms^{-1} (from Shapiro et al., 1987).

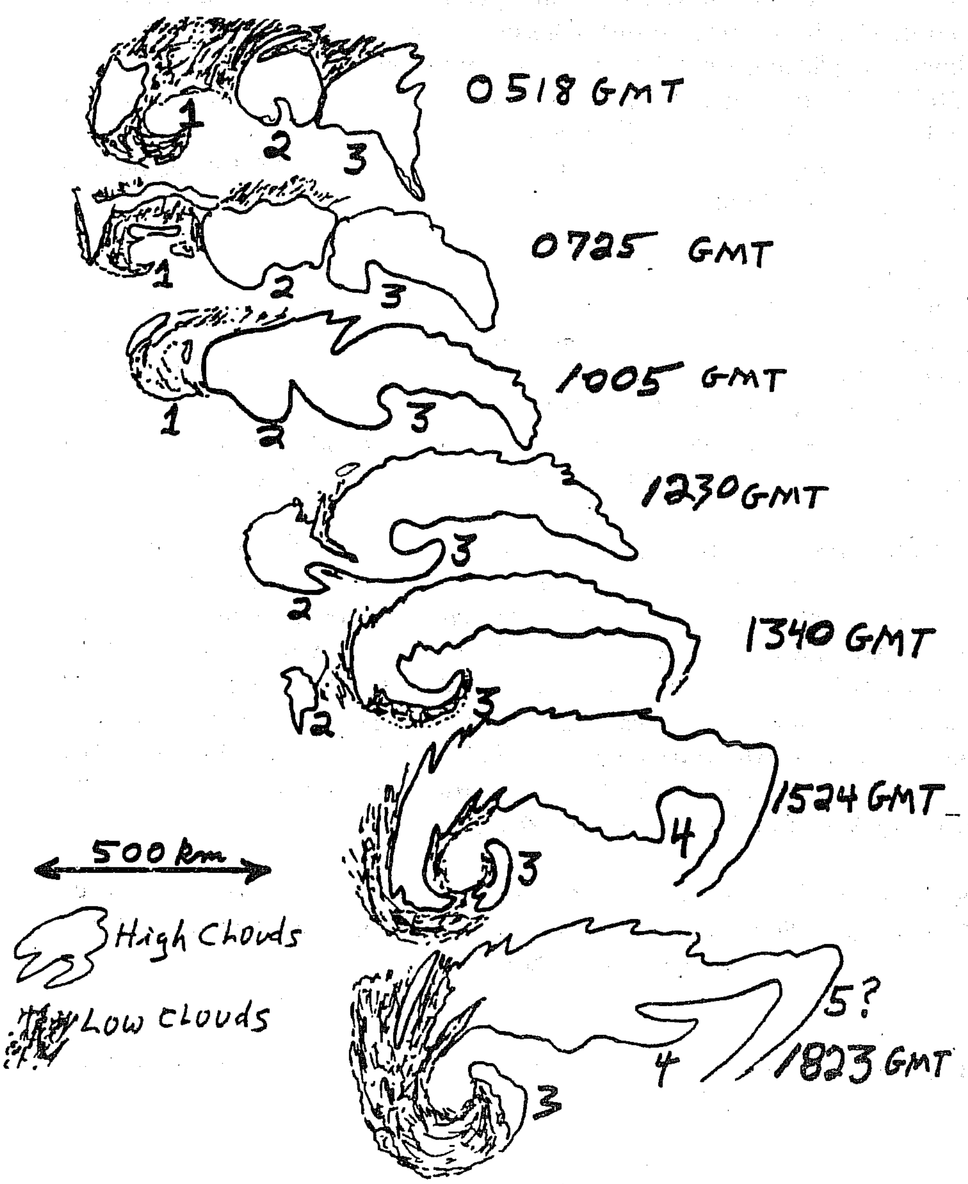

Figure 5. NOAA-satellite infrared image for 1340 UTC 27 February 1987 (from Shapiro et al., 1987).

Figure 6. Cloud structures for the period 0518 to 1823 UTC 27 February 1984 in the area of the polar low shown in Figure 4 and 5. (After Shapiro, 1986).

Mediterranean Sea) (Ernst and Matson, 1983; Billing et al., 1983; Rasmussen and Zick, 1987).

3. CLASSIFICATION OF POLAR LOWS

Polar lows span scales from hundred km to more than 1000 km in diameter, and a wide range of intensities, from strong breezes to hurricane force winds. They might be formed within few hours and last from hours to several days. Several attempts on a classification of polar lows have been made, but no widely accepted method exists. Some schemes have been proposed based on the lows's appearance on satellite images (Carleton, 1985; Forbes and Lottes, 1985), other schemes have been based on the synoptic situation (Rasmussen and Lystad, 1987; Reed, 1979). Rasmussen (1986) and Businger and Reed (1989) classified polar lows according to the underlying physical instability mechanisms.

The starting point in the present paper is the three types proposed by Businger and Reed, based the distribution of baroclinicity, static stability and surface fluxes of latent and sensible heat: 1. Short-wave/jet-streak type, characterized by a secondary vorticity maxima and positive vorticity advection aloft, deep, moderate baroclinicity, and modest surface fluxes. 2. Arctic-front type, associated with ice boundaries and characterized by shallow baroclinicity and strong surface fluxes. 3. Cold-low type, characterized by weak baroclinicity, strong surface fluxes and deep convection.

An additional simplification to this classification is here made, and just two main types are defined. The first type is the same as the first type above. It covers the so-called comma cloud polar low and a short overview is given below. The second type is an extension of the Arctic-front type to also include contributions from upper air disturbances, which probably always trigger the development of polar lows (e.g. Økland, 1989; Nordeng and Rasmussen, 1992), and deep convection (to about 500 hPa). In this way this type covers nearly all polar lows in the Norwegian and Barents Seas and polar lows in Northern Pacific where flows of cold air from the arctic sea ice play a role. The third type in the paper by Businger and Reed, with examples of deep convection in the Mediterranean, is not discussed in this paper.

3.1 Short-wave/jet-streak type

Polar lows develop a comma- or spiral-shaped pattern as they mature. The shape might be understood as a superposition of a positive vorticity centre and a moderate background current. If the magnitude of the background flow is weak, a spiral cloud is formed (Businger and Reed, 1989). American meteorologists (Anderson et al., 1969) found characteristic comma-shaped cloud patterns in cold air masses. The associated low is defined as a polar low and is often just called a "comma cloud", although such clouds also might be connected to extratropical cyclones.

A fine review on this type has been given by Businger and Reed (1989). Examples are described by Reed (1979), Mullen (1979), Reed and Blier (1986a, 1986b), Businger and Walter (1988). The disturbance is formed in a region of enhanced convection with positive vorticity advection ahead of an upper level short-wave trough moving westward. A surface trough is present in the rear of the comma trail. Weak to moderate baroclinicity with the warm air in south, exists within the cold air mass throughout the depth of the troposphere. This kind of polar lows has a relatively large scale and is not always distinguished from extratropical cyclones. Figure 7 shows a satellite image of a well-developed comma cloud in western North Pacific. A schematic representation of the comma cloud in relation to the major frontal cyclone is given in Figure 8.

The comma cloud might interact with polar front disturbances and several papers are devoted to such phenomena (Anderson et al., 1969; Reed, 1979; Carleton, 1985; Browning and Hill, 1985; McGinnigle et al. 1988; Mullen, 1979; Businger and Walter, 1988).

3.2 Arctic-front type

Arctic fronts are found near the ice edges between the extremely cold low-level air masses above

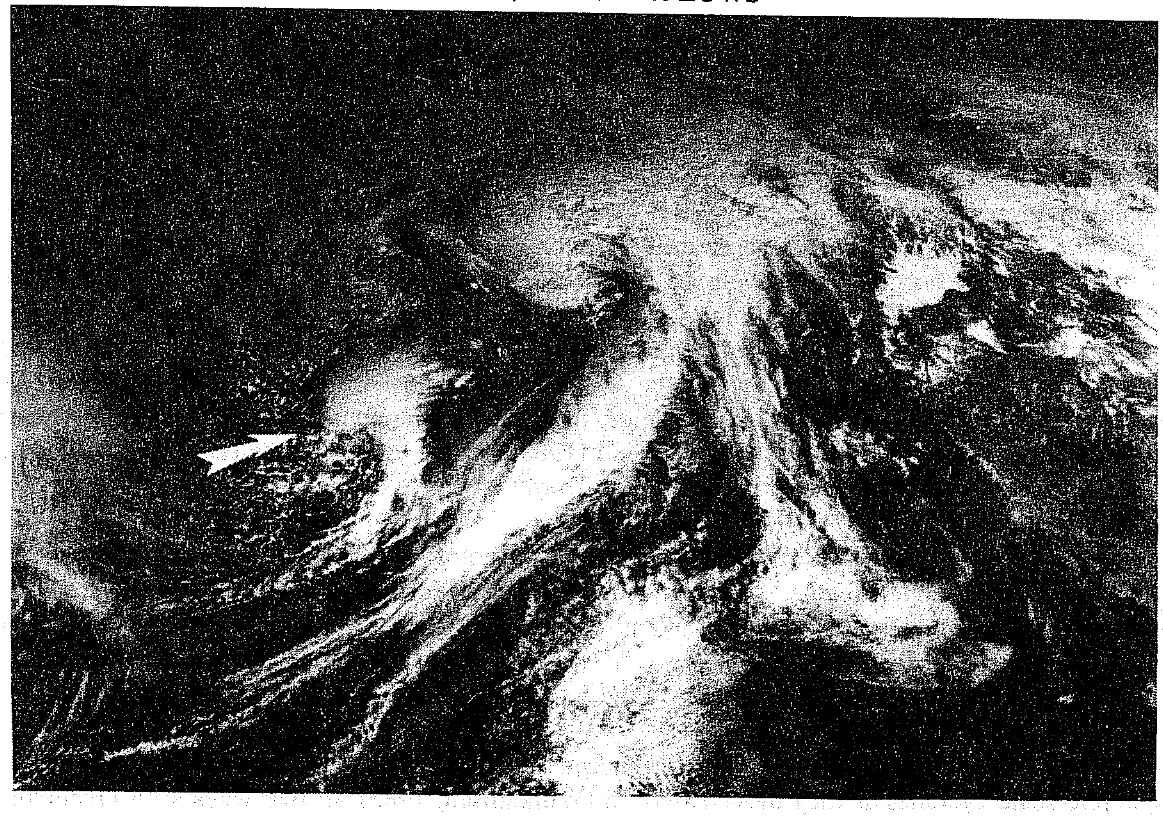

Figure 7. GOES visible satellite image of small comma cloud (see arrow) located to the west of a synoptic-scale frontal cyclone. The length of the arrow is 400 km. The west coast of the United States is visible on the right-hand side of the image (from Reed, 1979).

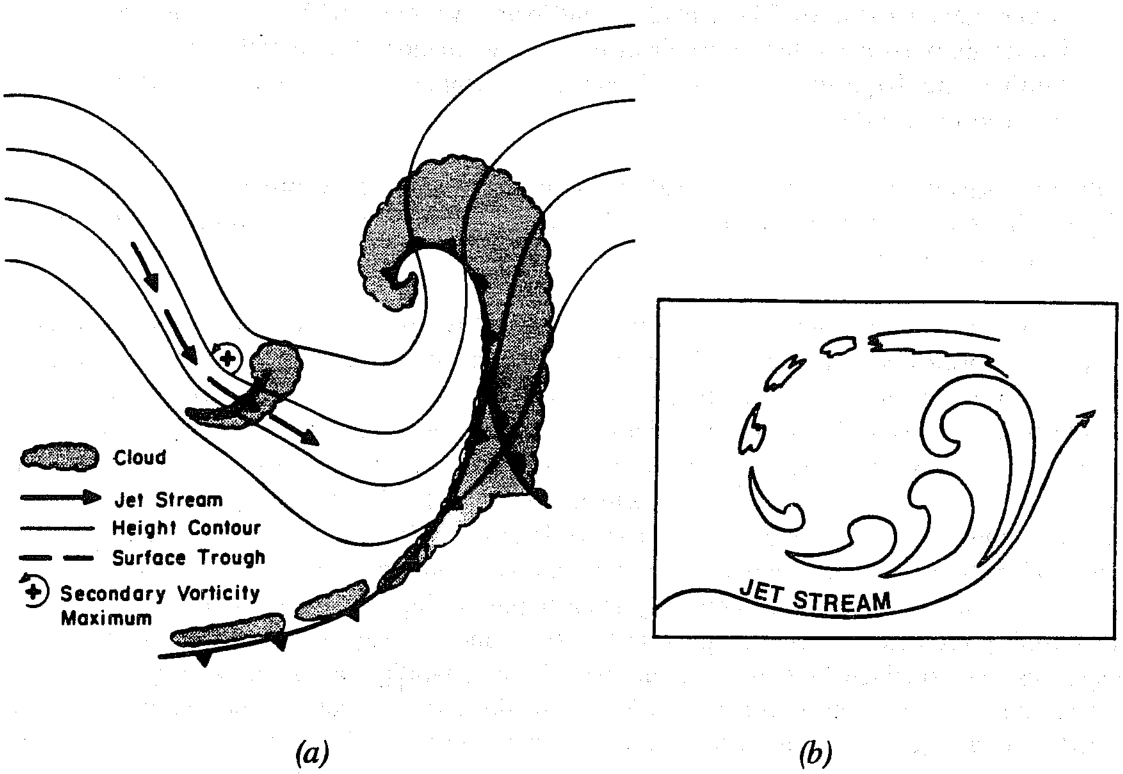

Figure 8. (a) Schematic diagram showing typical relationship of comma cloud to major cyclone and upper level jet. (b) Schematic direction of the principal life cycle of a comma cloud from the incipient stage to the mature vortex (from Zick, 1983).

the ice and warmer low level air masses over the ocean. When the cold air is entering the warm ocean, the low level inversion is gradually eroded by a convective layer, which increases in thickness and temperature downstream (e.g. Økland, 1983). The surface separating the convective boundary layer from the stable air above is here called the arctic front. As will be clear from the following, the polar lows do not necessarily form on the leading edge of the front. In fact, it is believed that they normally are developed further north. An example of an arctic front, as analyzed by Shapiro and Fedor (1989), is shown in Figure 9. The front surface follows the top of a convective ocean boundary layer, which in this case grows to a height of about 1500 m from the ice to the leading edge bending down to the surface 300 km further south. The inversion over the ice is reaching heights up to 700 hPa. Asymptotic heights of the convective layer of approximately 4000 m were estimated theoretically by Økland (1983). A real case estimate (Økland, 1989), gave a height of 3100 m after a trajectory of 700 km from the ice.

Arctic-fronts were introduced by the Bergen school meteorologists, but they are usually not analyzed on operational weather maps. Favoured regions of formation are the Greenland, Norwegian, and Barents Seas. Arctic fronts also commonly occur over the Bering Sea and northern Gulf of Alaska (e.g. Businger, 1987).

3.2.1 Synoptic conditions

Arctic fronts and polar lows of this type are normally developed in the rear of high latitude synoptic-scale cyclones as they move eastward (Wilhelmsen, 1985). In two ways such precursors create appropriate synoptic conditions for polar lows to develop:

> The synoptic-scale cyclone is characterized by a dry intrusion of stratospheric air leaving surpluses of high potential vorticity at relative low levels in the rear of the low.
>
> Low-level cold air-masses over the ice are advected out over the warm sea and Arctic-fronts move southward. These cold air outflows give zones with very strong low-level baroclinicity (beneath the reservoir of positive anomalies of potential vorticity). Furthermore, large fluxes of sensible and latent heat are present in the boundary layer over the warm sea surface.

Radiative flux-divergences at the low-level inversion over the ice concentrate high values of low level potential vorticity (Shapiro et al. 1989). However, so far no polar low formation has been connected to such structures.

The above statements are inspired by "the potential vorticity thinking" reviewed and extended in the paper by Hoskins et al. (1985), which also gives practical PV ideas of cyclogenesis. A fluid parcel conserves PV following its motion in adiabatic and inviscid flow. Intrusion of stratospheric PV could be treated as such a flow since diabatic processes here have rather long time scales compared to advective processes. Hoskins et al. introduced the "invertibility principle", saying that given some kind of balance approximation which eliminates high frequency gravity and sound waves, the wind, temperature and pressure can be determined from a knowledge of the full three-dimensional distribution of PV coupled with appropriate boundary conditions. The specification of potential temperature on the lower boundary is an important input to the mathematical solution of the invertibility problem in meteorological applications and is a key element in PV description of cyclogenesis. As explained by Hoskins et al., baroclinic instability can be viewed as the mutual interaction and enhancement of upper level PV anomalies and near-surface thermal anomalies. Air is forced to ascend just ahead of a PV anomaly so that the ascent of warm air in the thermal ridge provides the necessary energy conversion for development. The static stability can be shown to be weak beneath a PV anomaly so that organized convection in the cold air over the ocean is often enhanced in its proximity. Diabatic processes with relatively short time scales, primarily release of latent heat, will play a substantial role in concentration of low level PV and must be considered

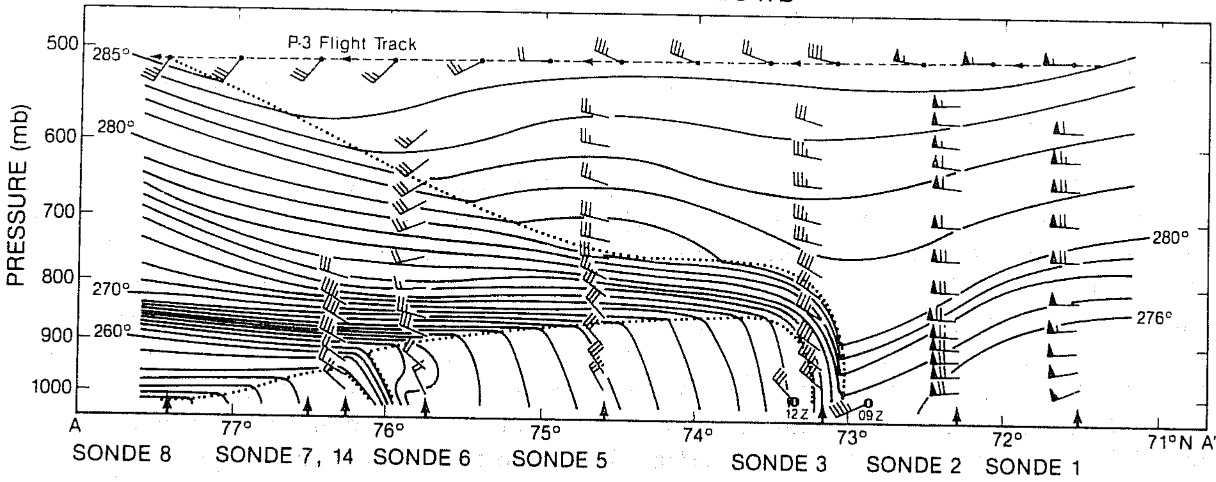

Figure 9. Cross-section analysis of potential temperature (K, solid lines) trough the arctic front on 14 February 1984 along a line east of Spitsbergen toward Northern Norway. Dropwindsonde locations are indicated by heavy arrows with identifying numbers. The dashed line with flight direction arrows and selected flight-level wind vectors shows the research aircraft flight track. Wind vectors without dotted heads indicate dropwindsonde wind profiles; wind vectors with flag=25 ms^{-1}, at full barb=5ms^{-1}, and at half barb=2.5 ms^{-1}. Dotted lines show frontal boundaries. (After Shapiro and Fedor, 1989).

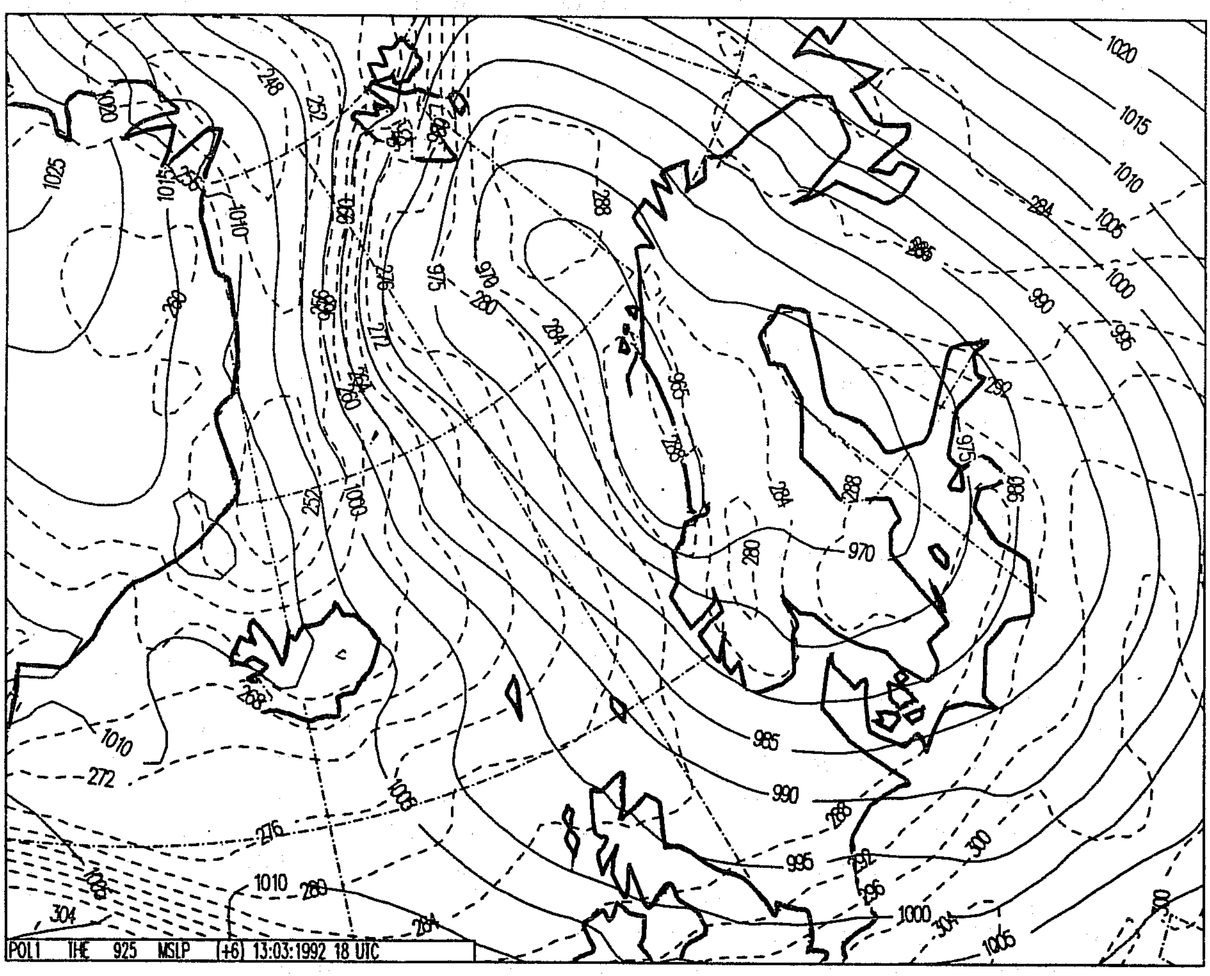

Figure 10. 6 h prediction valid at 3 March 1992 18 UTC of MSLP (solid lines) with a contour interval of 5 hPa and potential temperature at 925 hPa (stippled lines with 4 K contour intervals).

(e.g. Grønås et al., 1992). Friction will for time scales involved in the development of a polar low probably play a minor role (see Hoskins, 1990).

The ideas above will now be exemplified showing weather maps from a 6 hour simulation valid at 13 March 1992 at 18 UTC. Figure 10 shows a synoptic-scale low situated over Scandinavia and a northerly surface flow in the Norwegian Sea. The equivalent potential temperature at 925 hPa shows cold air over the ice being advected toward the sea. Strong horizontal gradients are present. The flow at the surface is southward, but the surface thermal wind is northward. This structure is referred to as the reversed shear, which is very common for this type of polar lows. In Figure 11 the heavy lines show the height (pressure) of the 280 K surface indicating a large scale intrusion. On the figure is also shown the remaining potential vorticity attached to the intrusion, with more than 4 PVU as low as 500 hPa north of Iceland. This situation is even more clearly demonstrated on the N-S cross-section Figure 12 from Western Greenland showing potential temperature and potential vorticity. The intrusion extends down to the top of the convective boundary layer which reaches 700 hPa. In accordance with the theory the stability is increased within the PV anomaly, and this time the stable layer extends toward the ice where the stable layer reaches about 750 hPa. Here the tropopause height is 400 hPa. Beneath a PV anomaly the stability will be lowered and this might explain the extension of the convective boundary layer to 3 km only 200 km from the ice.

Three sources are contributing to cyclogenesis: surface temperature anomalies, a positive PV anomaly aloft at relatively low levels and convection over an isolated area which gives possibilities for concentration of a low level PV maximum below the levels of released latent and sensible heat. The low levels involved ensure small scales of any development.

In the present case two polar lows were developed. The first sign of the formation east of Iceland was given by the updraft toward the positive IPV anomaly as this was advecting eastwards. In this case the two lows seem to develop independent of each other. Polar lows in similar cases have been observed as a train of lows. Four disturbances formed during two days in a case described by Reed and Duncan (1987).

The two examples described earlier in section 2 are of the Arctic front type. A case described by Rasmussen (1985a, 1989) and further analyzed dynamically by Strunge Pedersen (1989) - the so-called Bear Island Case from 11-12 December 1982 - is a well known case with data that have been used by many authors.

3.2.2 Synoptic climatology

Seasonal occurrence of polar lows in the Norwegian, Greenland and Barents Seas has been studied by Wilhelmsen (1985) and in the Norwegian Polar Low Project (Lystad et al. 1986). The period investigated was from 1970 to 1985. The more significant polar low tracks hitting Norway in the period 1978-82 are shown in Figure 13. The most favourable areas for polar lows to develop are the Bear Island - Spitsbergen areas and the sea east of Jan Mayen. Moreover, several developments are also found between Iceland and Scotland and in the central part of the Norwegian Sea. The season is from late September to the end of April. A frequency distribution for the period 1982-1985 is shown in Figure 14. A typical yearly amount in these years is close to 30. Some similarities between these polar lows and tropical cyclones have been found: The lows are intensified over the sea and decay quickly when entering land. Cyclonic inflow at low levels, ascending motion close to the centre and anticyclonic outflow close to the tropopause are observed in most cases. The core has a warm structure with a similar vertical distribution of equivalent potential temperature. Examples of polar lows with an eye have been observed. However, exceptionally low surface pressures have seldom been observed. Polar lows in relation to tropical hurricanes have been studied by Rasmussen (1989) and Nordeng and Rasmussen (1992).

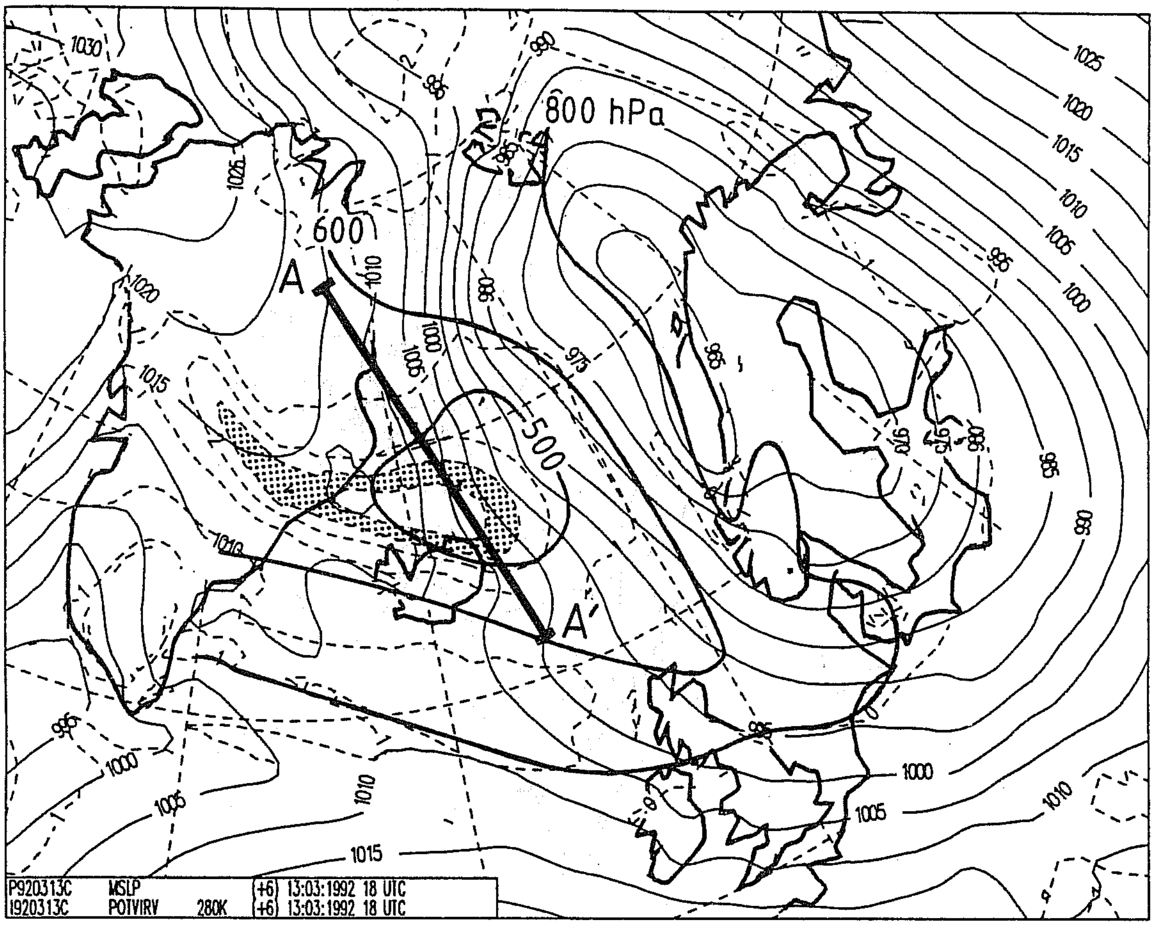

Figure 11. 6 h prediction valid at 3 March 1992 18 UTC. Solid lines: MSLP (contour intervals 5 hPa). Heavy lines: Height in pressure of the potential temperature surface 280 K. Stippled lines: Isentropic potential vorticity at the surface of 280 K with contour intervals of 1 PVU. Shaded area IPV between 3 and 4 PVU. The heavy line AA' gives the position of cross-section shown in Figure 12.

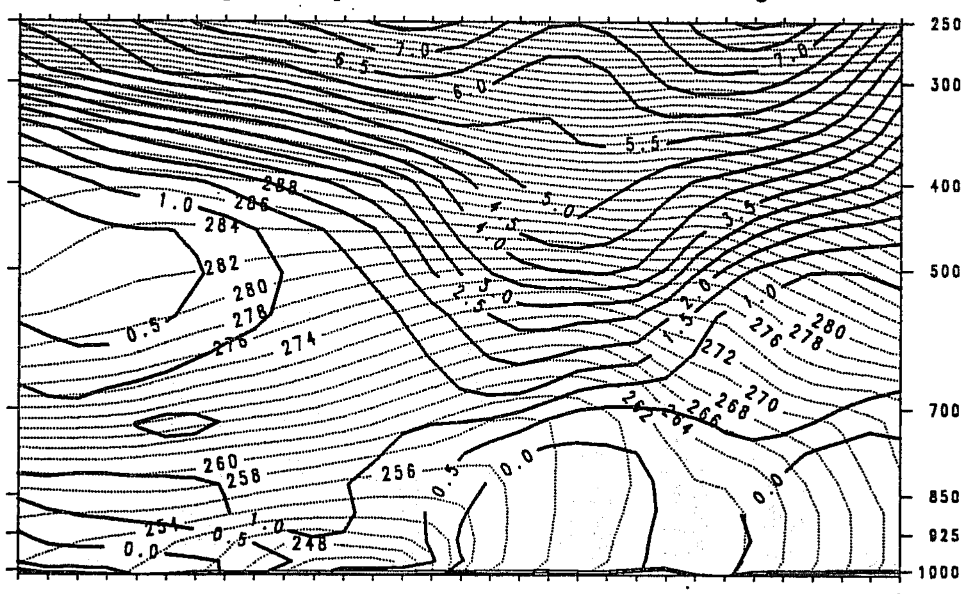

Figure 12. Cross-section of potential temperature (dotted lines) with a contour interval of 2 K and potential vorticity (solid lines) with a contour interval of 0.5 PVU. The distance between two tick-marks horizontally is 50 km. The time and position of the cross-section is given in Figure 11.

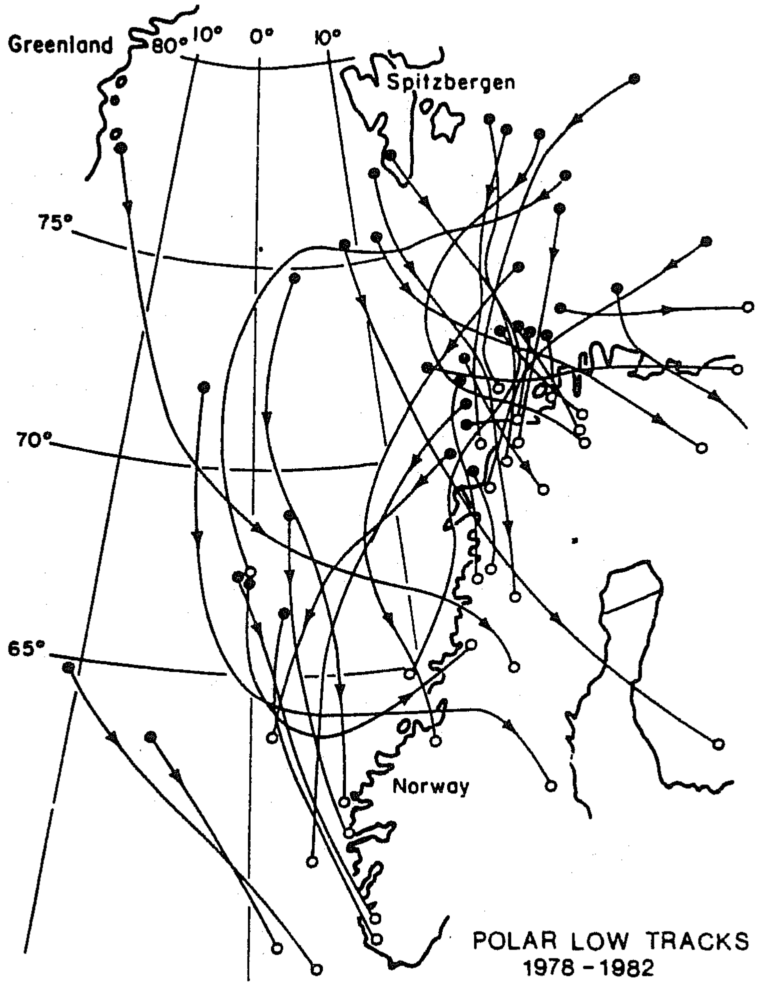

Figure 13. Trajectories of the position of significant polar lows close to Norway for the period 1978-82 (33 cases). (After Lystad, 1986).

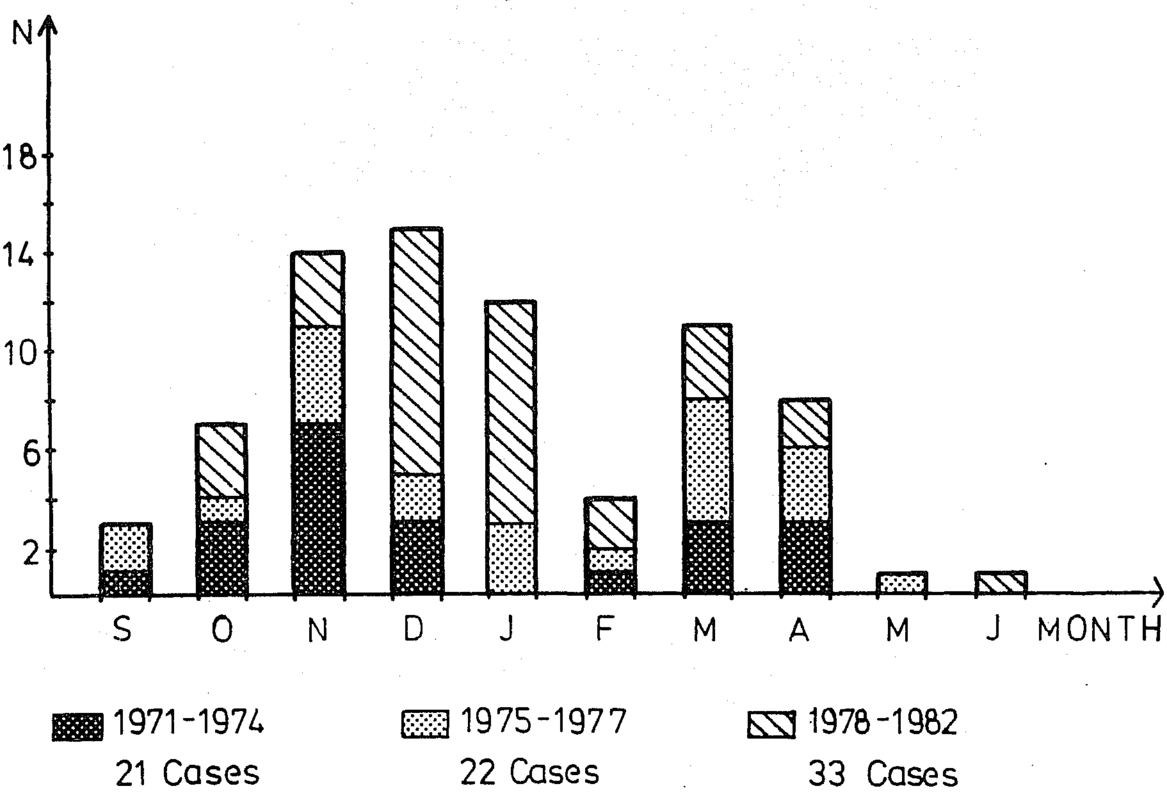

Figure 14. Frequency distribution of polar lows in the Norwegian and Barents Sea for the period 1971-82 (76 cases). (After Lystad, 1986).

4. THEORETICAL STUDIES

In order to create a polar low one or more instability mechanisms must be present. Since diabatic processes seems to play a role and conditional instability of the second kind (CISK) or air-sea interaction instability will often be involved. These instabilities theories, which are competing theories on the same topic, cannot cause the genesis of a polar low, but only contribute to the amplification in a later stage of the development. Disturbances of significant amplitudes have to be present before diabatic amplification takes place. Baroclinic instability is the obvious candidate for the development of this first stage, because baroclinicity is nearly always present in the early stages of the cyclogenesis.

4.1 Baroclinic instability

In a case with a polar low passing England, Harrold and Browning (1969) showed that the precipitation occurred mainly in slantwise ascent as in a typical baroclinic disturbance. This paper was followed by studies dealing with baroclinic instability as a mechanism of polar low formation. Mansfield (1974) applied Eady's perturbation model (dry model without friction) to a shallow layer using the observed mean state in the Harrold and Browning case. The most unstable normal modes found had e-folding times of 1-2 days for sizes of 600-800 km. Duncan (1977) used a linear quasi-geostrophic (dry) model to find normal mode solutions for unstable disturbances with small static stability near the surface. The presence of small static stability together with low level baroclinicity resulted in the formation of shallow disturbances that looked like the observed disturbances in size, growth rate and propagation speed. Duncan (1978) later used a numerical model of Stalley and Gall (1977) to a case of baroclinic instability in a reversed-shear flow and demonstrated the structure of the reversed-shear disturbance. Reed and Duncan (1987) applied the model to an observed background state in a case of trains polar lows in the reversed-shear flow. Baroclinic instability was suggested to originate the train (wavelength 500-600 km). The observed low moved with less speed than in the model, which also gave too small growth rates, suggesting a significant influence of diabatic heating.

Blumen (1980) studied unstable two-dimensional Eady-waves in a two-layer model characterized by smaller static stability in the lower layer. The model gave a maximum baroclinic instability to low wavelengths when the lower layer has a stratification close to the adiabatic lapse rate. Wiin-Nielsen (1989) used a three-layer model in a similar manner as Blumen. He found minum e-folding times for wavelengths of 100-200 km, when the lapse-rate in the lower layer is close to the adiabatic lapse rate. The wavelength was found to be sensitive to the lapse rate and slightly smaller lapse rates increased the wavelength to 500 km.

Moore and Peltier (1987, 1989a, 1989b) have investigated the stability of frontal zones induced by deformation and discovered a new cyclone-scale mode of baroclinic instability using the primitive equations. They show that this mode filters out in both quasi-geostrophic and geostrophic approximations to the primitive equations. They used their methodology (Moore and Peltier, 1989c) in identifying dynamical processes responsible for the development of polar low wave trains. The predicted doubling and wavelength of the most unstable wave in the branch were found to be in agreement with the observations made by Reed and Duncan (1987). The wave was found to have its maximum amplitude in regions where the static stability field had its minimum. A dry model was used and they argue that the rapid initial growth of such disturbances need not involve a strong feedback with moist diabatic processes.

The theoretical investigations mentioned above are based on normal mode solutions. By taking nonmodal initial conditions which correspond to an upper level trough overtaking a surface depression, Farrell (1982, 1984) shows that growth follows, with a scale of cyclogenesis is that of the upper level trough, and that this scale is independent of the most rapidly growing normal mode. As a short-wave trough approaches a surface depression, under favourable conditions rapid deepening to 3-5 times the initial pressure deficit in less than 12 hours might occur. Orlanski

(1986) applied the initial value approach in a two-dimensional model on mesoscale cyclogenesis in a mean baroclinic state. As others, he is stressing the importance of low static stability at low levels and moist processes in obtaining strong development. Montgomery and Farrel (1991) suggested that some polar low developments consist of an initial baroclinic growth phase followed by a prolonged slow intensification due to diabatic effects.

4.2 Conditional instability of second kind

Diabatic heating, primarily latent heat release and warming by sensible heat, plays an important role in the formation of polar lows. Studies of this are mainly built on the concept of the conditional instability of second kind (CISK) introduced by Charney and Eliassen (1964). This theory explains the growth of the hurricane depression as the result of a cooperative interaction between cumulus convection and the largescale circulation. Suppose that cumulus convection has been sufficiently organized, then CISK predicts a positive feedback between the cloud scale and the developing vortex scale: the vortex provides moisture convergence in the boundary layer for the convective process and the cumulus scale provides latent heating that intensifies the vortex. As the frictional convergence enhance the diabatic heating of the core, the gradient wind balance is disturbed in such a way that the vortex has to grow, in spite of the spin-down effect of boundary layer pumping. The theory of CISK has never been generally accepted as a theory of tropical cyclones (e.g. Emanuel, 1986, 1987). Van Delden (1989a) suggests that some confusion is rooted in the fact that CISK was presented as a linear theory. He reminded that Charney and Eliassen recognized that CISK is not a theory for the genesis of a hurricane depression in the same way as static thermal instability. They realized that a pre-existing depression of at least small amplitude is necessary before CISK can start and give further growing. Van Delden **(1989a and b) developed a** finite amplitude analysis of CISK and concluded that CISK cannot cause the genesis of polar lows, but only contributes to the amplification in a later stage of the development.

Økland (1977) investigated CISK for surroundings typical for polar lows using an analytical model. He showed that release of comparative small amounts of latent heat may be sufficient for the spin-up of polar lows. He also found that the static stability must be below a certain limit for deepening to occur. In two later papers (Økland 1987, 1989) he gave a simple conceptual CISK model for the polar low formation. First a convective boundary layer is formed with the cold air outflow. Beneath an upper layer disturbance deep convection is organized in an isolated horizontal area and the CISK starts.
Økland stresses that it is not sufficient to show that theories of tropical cyclones can be adopted to polar low genesis. In addition, it must be explained why similar lows do not develop in middle atmosphere. For instance at the coast of west Norway there is maximum of thunderstorms in winter when winds are blowing from the Atlantic. Troughs of small dimensions are frequent, but polar lows are never observed. This is so because the low-level inversion is missing in such flows.

Rasmussen (1977, 1979) used a three-layer, linear, quasi-geostrophic model to study normal modes for CISK-driven disturbances under typical polar low conditions. The maximum growth rate occurred at wave lengths comparable to the size of polar lows. Computed e-folding times of approximately 12 hours were large enough to account for the fast development.

Bratseth (1985) showed, using a linear analytic model built on the Boussinesq approximation in the primitive equations, that CISK may be important in polar air if the heating takes place in a shallow layer at low levels at the top of the boundary layer. The layer must be elevated above the surface in order for a short-wave cut-off to exist.

4.3 Air-sea interaction instability

CISK is a conceptual model used as an explanation for the development of tropical cyclones and polar lows. Emanual (1986) challenges the CISK theory which he believes is an energetically inconsistent concept. He makes use of the observation that in the tropical atmosphere, clouds have mean temperatures very close to those of their immediate environment. Thus, no amount of moisture convergence and associated heating will lead to energy production, which depends on a positive correlation of heating and temperature perturbation. In order to generate actual temperature perturbations aloft, one needs to have elevation of θ_e in the boundary layer. In tropical cyclones, this elevation arises from transfer of latent and possibly sensible heat from the sea surface. Emanuel proposes that both tropical cyclones and polar lows are formed as a consequence of a two-stage dynamical process. In the first stage, a broad-scale cyclonic circulation is created at the surface, e.g. by migration of an upper tropospheric anomaly of PV. This is similar to what is needed for CISK. If a cyclone of sufficient amplitude is created, a finite amplitude sea-air instability kicks in where anomalous sea surface heat fluxes induced by strong surface winds and falling pressure lead to increased temperature anomalies and increased surface wind and pressure deficit. An axisymmetric analytical nonlinear model of the mature storm is developed, leading to the prediction of the central pressure drop in a one-to-one relationship to the increase of θ_e in the centre. In this explanation cumulus convection clouds are viewed as agents for redistributing aloft heat acquired from the sea surface. It is not the organization of convection which is required, but rather the organization of surface sensible and latent heat fluxes. It is shown by Emanuel and Rotunno (1989) both analytically and by numerical models that disturbances may amplify in a conditionally neutral atmosphere providing that they begin with a sufficiently large amplitude. Their numerical experiments show that air-sea interaction can produce polar lows in consistence with the observations, but with a stronger intensity. They argue that the weaker intensities observed might be because polar lows remain over warm water for too short a period.

Data indicate that combined fluxes of sensible and latent heat may have been as high as 1000 W m^{-2} over the Norwegian Sea during an intense low (Shapiro et al., 1987). However, it was pointed out by Rasmussen and Lystad (1987) that CAPE is large during cold air outbreaks when polar lows develop and therefore arguments applied to the tropical hurricane may not apply to polar lows.

5. NUMERICAL PREDICTION EXPERIMENTS

Polar low NWP experiments have been made for two main reasons: with the aim to validate and improve operational models predicting polar lows, and as case studies on the genesis and life-cycle of polar lows. We have seen that inner cores of polar lows of the arctic type might have a diameter down to 100 km. This means that any hope of detailed result requires a numerical model with extremely high resolution.

NWP experiments, starting with coarser scale numerical analyses from ECMWF or NMC, have been made by Seaman (1983), Sardie and Warner (1985), Grønås et al. (1986a, 1986b, 1987a and b), Nordeng (1987, 1990), Nordeng and Rasmussen (1992), Nordeng et al. (1989), Langeland and Miller (1989), Dell'Osso(1992). The hydrostatic mesoscale limited area models used have been the Norwegian operational model, the PSU/NCAR model or the ECMWF limited area model. The horizontal resolution has mainly been 50 km or less. The highest resolution, with a horizontal mesh of about 20 km, has been used by Dell 'Osso

In the seven cases studied by Grønås et al., disturbances were predicted that could be associated with the observed polar lows, but the strength of the disturbances was generally too weak. Nordeng and Rasmussen (1992) successfully simulated a case in the Norwegian Sea. However, in this case the measured pressure deficit in the core was modest. Del Osso (1992) managed to simulate fairly well the large pressure deficit observed by Shapiro et al. (1987) in the ACE case. This result was only obtained by using the best resolution. The Bear Island case (Rasmussen, 1985a) is an example of a case where the models do not capture the intensification of the inner core of the cyclone.

In order to illustrate how a successful simulation can give information on the genesis and structure of typical polar lows, a case study by Grønås and Kvamstø (yet unpublished) will be presented.

The Norwegian model, with the extension on the hydrological cycle according to Sundqvist et al. (1989), has been used with 18 layers below 100 hPa and a horizontal mesh of 50 km. This resolution correspond to the current resolution in the ECMWF-model. The starting analyses and the lateral boundary conditions are ECMWF analyses. The results after six hours have already been shown.

After 24 h two polar lows have developed. Figure 15 shows the sea level pressure and IPV at 272.5 K at this time. Streamers of IPV are stretching towards the two polar lows. Notice also a third branch toward a low level (unreversed shear) temperature wave further south. We concentrate on the polar low near S-Norway. The streamer around the low on the southern side is as expected in a reversed-shear baroclinic development. However, the minimum band of IPV northwestward from the low is very apparent. This minimum is a result of redistribution of PV toward lower levels caused by release of latent and sensible heat. Figure 16 a and b shows a W-E cross-section of equivalent potential temperature, potential temperature and vorticity through the low. A low level cyclonic vortex has been formed with anticyclonic vorticity above. The 3-dimensional wind field shows strong inflow at low levels, ascending motion around the central part and anticyclonic outflow at the tropopause (550 hPa). The warm core has a vertical structure in equivalent potential temperature as in tropical cyclones, however, no clear subsidence is found in the centre. The redistribution of PV leads to increased upper level outflow and surface pressure fall. When the model is run without release of latent heat the two polar lows are replaced with surface troughs and more shallow, tilting disturbances. The pressure differences between the two runs make up for the surface pressure deficit for the two polar lows in the complete run. These experiments indicate that CISK has to be activated in order to get closed isobars and a circular shape of the surface disturbance. However, baroclinic instability (moist) caused by interaction of the upper level positive PV anomaly and the surface temperature anomaly seem to create the disturbance.

6. FORECASTING POLAR LOWS

The arctic front type of polar lows presents a difficult forecast problem due to the small size and lack of a sufficiently dense observing network in the regions of interest. During the Norwegian Polar Low Project some forecasting experience was gained for disturbances in the Norwegian and Barents Seas (Lystad, 1986). The forecasting problem was divided into the problem of monitoring and the problem of making the actual forecast. The monitoring involves the synthesis of all available data in the representation of the atmospheric state. Mainly three kinds of information are available, analyses and short range numerical forecasts; individual reports from land, ship and drifting buoys and radiosonde stations; and interpretation of infrared images from polar-orbiting satellites.

A method to predict the track of already identified polar lows was tried by Midtbø (1986). Winds from 850 and 700 hPa winds of numerical models were used. Average position error of 200 km for 18 h forecasts was found for average path length of 800 km.

At the Norwegian meteorological institute a mesoscale model covering the Norwegian and Barents Seas was made operational in 1985 (Grønås et al., 1987). During the polar low season 1987/88 a verification test (unpublished) of the models ability to forecast polar lows was made. All available satellite images for this winter were searched for any mesoscale disturbances in the area with a diameter less than 500 km. Surges of cold air from the ice were frequent this winter and a little more than 100 polar low cases were found. In more than 80% of the cases the short range forecasts (less than 24 hours) included disturbances which could be identified with the polar low indications at the images. Position errors were measured and found to increase in the mean from 60 km in the six hour forecast to 130 km in the 24 hour forecast. The investigation did not say anything of the model's ability to forecast the intensity of the disturbances and did not estimate the rate of "false alarm" predicted by the model. In 1990 the resolution of the Norwegian operational model for this area was increased to 30 levels below 100 hPa and a 25 km horizontal mesh

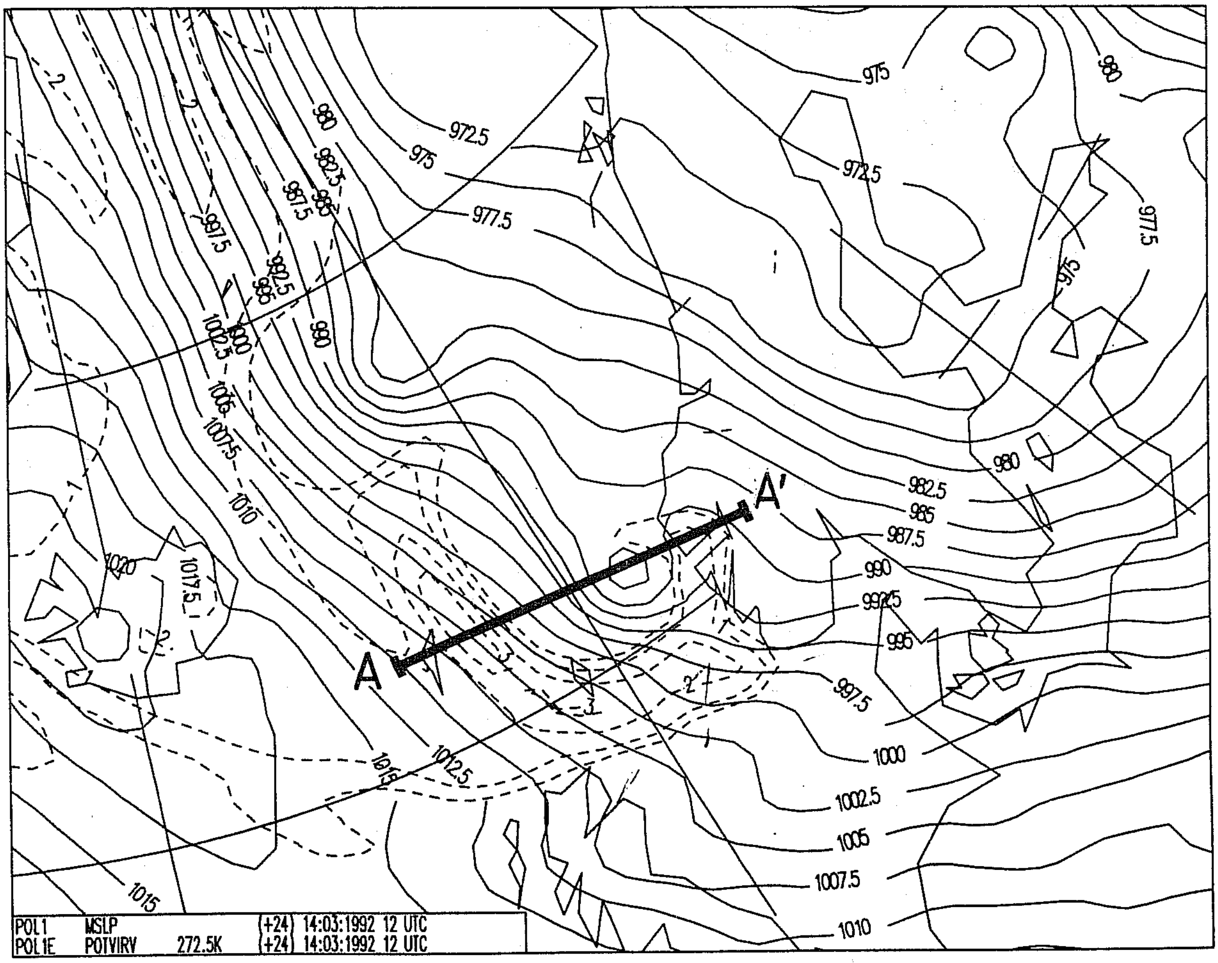

Figure 15. 24 h prediction valid 14 March 1992 12 UTC of MSLP (solid lines) with a contour interval of 2.5 hPa and isentropic potential vorticity (stippled lines) at 272.5 K with lines drawn for 0.5,1,2,3 PVU. The heavy line AA' gives the position of the cross-section shown in Figure 16.

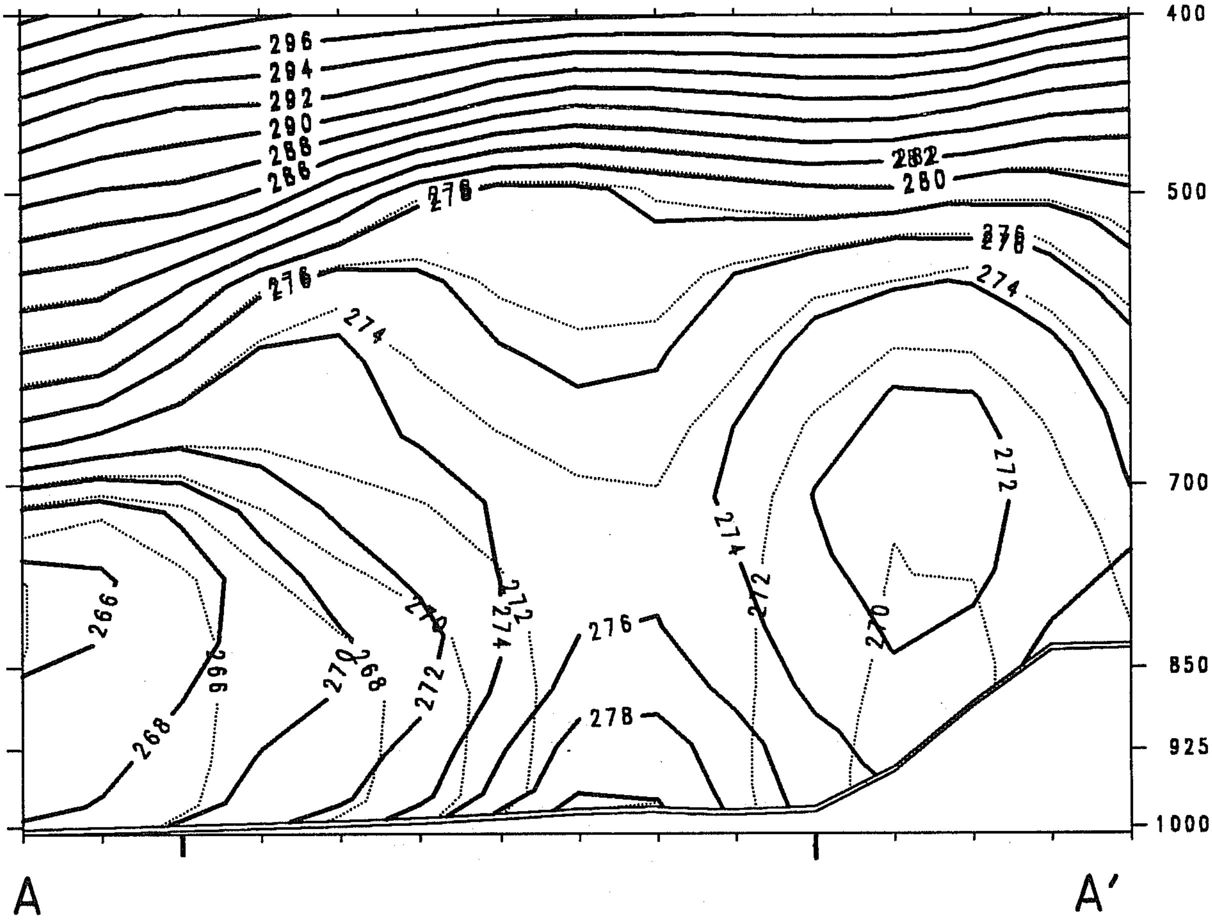

Figure 16. a) Cross-section of potential temperature (dotted lines) and equivalent potential temperature (solid lines). Contour steps are 2 K. The time and position of the cross-section is given in Figure 15.

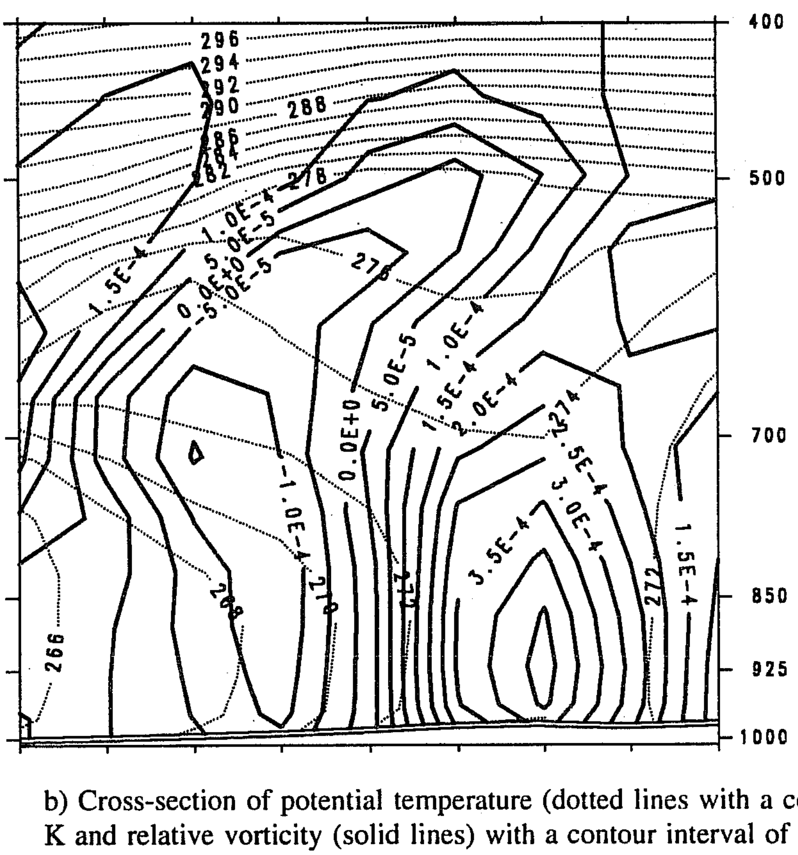

Figure 16. b) Cross-section of potential temperature (dotted lines with a contour interval of 2 K and relative vorticity (solid lines) with a contour interval of 0.5E-4 (s^{-1}). The time and position of the cross-section is given in Figure 15. Only a part of the cross-section AA' is shown.

(Grønås, 1990).

The general experience is that the mesoscale numerical models give considerable forecasting guidance in the prediction of polar lows. The positioning errors are relatively large and the intensity of inner cores are often too weak. Disturbances caused by CISK instabilities might be considered as an independent disturbance in contrast to mesoscale disturbances forced by the larger scales. Such independent disturbances will probably have low predictability compared to forced mesoscale features. Because of this, relatively low predictability should be expected for the forecasting of polar lows.

Improvement of polar low forecasting will result from improvements of the numerical models through the advancement of relevant observation systems, assimilation systems, physical parametrizations and higher resolution. The radiosonde ascents in the area should be increased from two to four drops a day. Assimilation of radiances, scatterometer winds, etc. should be tried using variational analysis with adjoint models. Parametrization of the microphysics connected to the hydrological cycle should be further validated and improved. The up-dating of SST in the models should be more frequent.

The forecaster will continue to play a major role in polar low prediction for the foreseeable future. He/she has to monitor the models output together with other kinds of observations to make the best prediction of polar lows over ocean.

REFERENCES

Anderson, R.K., J.P. Ashman, F. Bittner, G.R. Farr, E.W. Ferguson, A.H. Smith, 1969. Application of Meteorological Satellite Data in Analysis and Forecasting. ESSA Tech. Rep. NESC51, Goverment Printing Office, Washington, D.C. (NTIS AD-697033).

Bergeron, T., 1949. De tropiska orkanernas problem (swedish). Svenska Fysikersamfundets publikation Kosmos, Vol 27.

Bergeron, T., 1954. Reviews of tropical hurricanes. Quart. J. Roy. Meteor. Soc., 80, 131-164.

Billing, H.I., I. Haupt, W. Tonn, 1983. Evolution of a hurricane-like cyclone in the Mediterranean Sea. Beitr. Phys. Atm., 56, 508-510.

Blumen, W., 1980. On the evolution and interaction of short and long baroclinic waves in the Eady type. J. Atmos. Sci, 37, 1984-1993.

Bratseth, A.M., 1985. A note on CISK in polar air masses. Tellus, 37A, 403-406.

Bromwitch, D.H., 1989. Subsynoptic-scale cyclone developments in the Ros Sea sector of the Antarctic. Polar and arctic lows. A. Deepak Publishing, Hampton, Virginia, USA, 331-346.

Browning, K.A., F.F. Hill, 1985. Mesoscale of a polar low trough interacting with a polar front. Q.J.R.M.S., 11, 445-462.

Businger, S., 1987. The synoptic climatology of polar low outbreaks over the Gulf of Alaska and Bering Sea. Tellus, 39. 307-325.

Businger, S., R.J. Reed, 1989. Polar Lows. Polar and arctic lows. A. Deepak Publishing, Hampton, Virginia, USA, 3-46.

Businger, S., B. Walter, 1988. Comma cloud development and associated rapid cyclogenesis over the Gulf of Alaska: A case study using aircraft and operational data. Mon. Wea. Rev., 116, 1103-1123.

Carleton, A.M., 1985. Satellite climatological aspects of the "polar low" and "instant occlusion". Tellus, 37, 433-450.

Charney, J.G., and A. Eliassen, 1964. On the growth of hurricane depression. J. Atmos. Sci., 21, 68-75.

Dell'Osso, L. 1992. Paper in ECMWF seminar proceeding 1992.

Douglas, M.W., M.A. Shapiro, 1989. A comparison of the structure of two polar lows observed by research aircraft. Polar and arctic lows. A. Deepak Publishing, Hampton, Virginia, USA, 291-312.

Duncan, C.N., 1977. A numerical investigation on polar lows. Q.J.R.M.S., 103, 255-268.

Duncan, C.N., 1978. Baroclinic instability in a reversed shear flow. Met. Mag., 107, 17-23.

Emanuel, K.A., 1986. An air-sea interaction theory for tropical cyclones. Part I: Steady-state maintenance. J. Atmos. Sci., 43, 585-604.

Emanuel, K.A., R. Rotunno, 1989. Polar lows as arctic hurricanes. Tellus, 41A, 1-17.

Ernst, J.A., M. Matson, 1982. A Mediterranean tropical storm? Weather, 38, 332-337.

Farrel, B.F., 1992. The initial growth of disturbances in a baroclinic flow. J. Atmos. Sci., 39, 1663-1668.

Farrel, B.F., 1984. Modal and non-modeal baroclinic waves. J. Atmos. Sci., 41, 668-673.

Fett, R.W., 1989. Polar low development assosiated with boundary layer fronts in Greenland, Norwegian and Barents Seas. Polar and Arctic Lows. A. DEEPAK Publishing, Hampton, Virginia, USA, 313-322.

Forbes, G.S., W.D. Lottes, 1985. Classification of mesoscale vortices in polar airstreams and the influence of the large-scale environment on their evolutions. Tellus, 37, 132-155.

Grønås, S., A. Foss, M. Lystad, 1986a. Numerical simulations of polar low in the Norwegian Sea, Part I: The model an simulations of the polar low 26-27 February 1984. Tech. Rep. No 5, Norwegian Meteorological Institute, Oslo, 55 pp.

Grønås, S., A. Foss, M. Lystad, 1986b. Numerical simulations of polar low in the Norwegian Sea, Part II: Simulations of six synoptic situations. Tech. Rep. No 18, Norwegian Meteorological Institute, Oslo, 32 pp.

Grønås, S., A. Foss, M. Lystad, 1987a. Numerical simulations of polar lows in the Norwegian Sea. Tellus, 39, 334-354.

Grønås, S., A. Foss, K.H. Midtbø, 1987b. The Norwegian mesoscale NWP system. Proc. Symp. Mesoscale Ana. and Forecasting, Vancouver, Can., ESA SP-282.

Grønås, S., 1990. Early results with the new Norwegian high resolution operational NWP models. HIRLAM Technical Report No. 8.

Grønås, S., N.G. Kvamstø, E. Raustein, 1992. Dynamcial aspects of the mesoscale storm over N-Germany, 27-28 August 1989. Submitted to Tellus.

Harley, D.G., 1960. Frontal contour analysis of a "ploar low". Meteor. Mag. 89, 146-147.

Harrold T.W. and K.A. Browning, 1969. The polar low as a baroclinic disturbance. Quart. J. Roy: Meteor. Soc., 95, 710-723.

Heinemann, G., 1990. Mesoscale vortices in the Weddell Sea Region (Antarctica). Mon. Wea. Rev. 118, 779-793.

Hoskins, B.J., M.E. McIntyre and A.W. Robertson, 1985. On the use and significance of isentropic potential vorticity maps. Q.J.R.M.S. 111, 877-946.

Hoskins, B.J., 1990. Theory of extratropical cyclones. The Eric Palmen memorial volume. American Meteorological Society, Boston.

Kellogg, T.T. and P.F. Twitchell, 1986. Summary of the workshop on arctic lows 9-10 May 1985. Boulder, Colorado. Bull. Amer. Meteor. Soc., 67, 186-193.

Langeland, R.H., R.J. Miller, 1989.0 Polar low sensitivity to sea surface temperature and horizontal grid resolution in a numerical model. Polar and Arctic Lows. A. DEEPAK Publishing, Hampton, Virginia, USA, 247-254.

Locatelli, J.D., P.V. Hobbs, J.A. Werth, 1982. Mesoscale structures of vortices in polar air streams. Mon. Wea. Rev., 110, 1417-1433.

Lyall, I.T., 1972. The polar low over Britain. Wather. 27, 378-390.

Lystad, M. (Ed.), 1986. Polar lows in the Norwegian, Greeland and Barents Seas. Final Rep., Polar Lows Project, The Norwegian Meteorological Institute, Oslo, 196 pp.

Mansfield, D.A., 1974. Polar lows: The development of baroclinic disturbances in cold air outbreaks. Q.J.R.M.S., 100, 541-554.

McGinnigle, J.B., M.V. Young, M.J.Bader, 1988. The development of instant occlusions in the Northe Atlantic. Meteor. O.15 Internal Report, Meteorological Office, London Road, Bracknell, 25 pp.

Midtbø, K.H., 1986. Polar low forecasting. Proceedings of the international conference on polar lows, Oslo, 1986. The Norwegian Meteorological Institute, 363 pp.

Montgomery, M.T., Farrel, B.F., 1991. Moist surface frontogenesis associated with interior potential vorticity anomalies in a semigeostrophic model. J. Atmos. Sci, 48, 343-367.

Moore, G.W.K, W.R. Peltier, 1987. Cyclogenesis in frontal zones. J. Atmos. Sci., 44, 384-409.

Moore, G.W.K, W.R. Peltier, 1989a. Non-separable baroclinic instability. Part I: Quasi-geostrophic dynamics. J. Atmos. Sci, 46, 57-78.

Moore, G.W.K, W.R. Peltier, 1989b. Frontal cyclogenesis and the geostrophic momentum approximation. Geophys. Astrophys. Fluid Dyn.

Moore, G.W.K, W.R. Peltier, 1989c. On the development of polar low wavetrains. Polar and Arctic Lows. A. DEEPAK Publishing, Hampton, Virginia, USA, 141-153.

Mullen, S.L., 1979. An investigation of small synoptic-scale cyclones in polar air streams. Mon. Wea. Rev., 107, 1636-1647.

Nordeng, T.E., 1987. The effect of vertical and slantwise convection on the simulation of polar lows. Tellus, 39, 354-357.

Nordeng, T.E., 1990. A model-based diagnostic study of the development and maintenance mechanism of two polar lows. Tellus 42A, 92-108.

Nordeng, T.E., A. Foss, S. Grønås, M. Lystad, K.H. Midtbø, 1989. On the role of resolution and physical parameterization for numerical simulations of polar lows. Polar and Arctic Lows. A. DEEPAK Publishing, Hampton, Virginia, USA, 217-232.

Nordeng, T.E., E.A. Rasmussen, 1992. A most beautiful polar low. A case study of a polar low development in the Bear Island region. Tellus, 44A, 88-99.

Orlanski, I., 1986. Localized baroclinity: A source for meso-α cyclones. J. Atmos. Sci., 43,2857-2885.

Pedersen, T. Strunge, 1989. On the effect of using different formulations for the forcing in the omega equation applied to polar lows. Polar and Arctic Lows. A. DEEPAK Publishing, Hampton, Virginia, USA, 233-246.

Parker, M.N., 1989. Polar lows in Beaufort Sea. Polar and Arctic Lows. A. DEEPAK Publishing, Hampton, Virginia, USA, 323-330.

Rabbe. Å., 1975. Arctic instability lows. Meteorologiske Annaler, 6, 303-329. The Norwegian Meteorological Institute, Oslo.

Rabbe., Å., 1987. A polar low over the Norwegian Sea, 29 February-1 March 1984. Tellus, 39, 326-333.

Rasmussen, E., 1977. The Polar Low as a CISK Phenomenon. Rep. No. 6, Inst. Theoret. Meteor., Københavns Universitet, Copenhagen, 77 pp.

Rasmussen, E., 1979. The polar low as an extratropical CISK disturbance. Quart. J. Roy. Meteor. Soc., 105, 531-549.

Rasmussen, E., 1985a. A case study of polar low development over the Barents Sea. Tellus, 37A, 407-418.

Rasmussen, E., 1985b. A case study of a polar low development over the Barents Sea. Tech. Rep., Polar Lows Project, The Norwegian Meteorological Institute, Oslo, 42 pp.

Rasmussen, E., 1986. Different types of polar lows affecting Scandinavia. Polar lows project. Proceedings of the international conference on polar lows. Oslo, Norway, 20-23 May 1986.

Rasmussen, E., M. Lystad, 1987. The Norwegian polar low project: A summary of the international conference of polar lows. Bull. Amer. Meteor. Soc., 68, 801-816.

Rasmussen, E., C. Zick, 1987. A subsynoptic vortex over the Mediterranean Sea with som resemblance to polar lows. Tellus, 39, 408-425.

Rasmussen, E., 1989. A comparative study of tropical cyclones and polar lows. Polar and arctic lows, 47-80. A. Deepak Publishing, Hampton, Virginia, USA.

Reed, R.J., 1979. Cyclogenesis in polar air streams. Mon. Wea. Rev., 107, 38-52.

Reed, R.J., W. Blier, 1986a. A case study of comma cloud development in the Eastern Pacific. Mon. Wea. Rev., 114, 1681-1695.

Reed, R.J., W. Blier, 1986b. A further case study of comma cloud development in the Eastern Pacific. Mon. Wea. Rev., 114, 1696-1708.

Reed, R.J., C.N. Duncan, 1987. Baroclinic instability as a mechanism for the serial development of polar lows: A case study. Tellus, 39, 376-384.

Sardie, J.M., and T.T. Warner, 1985. A numerical study of the development mechanisms of polar lows. Tellus, 37, 460-477.

Staley, D.O., R.L. Gall, 1977. On the wavelength of maximum baroclinic instability. J. Atmos. Sci, 34, 1679-1688.

Sundqvist, H., Berge, E. and Kristjansson, J.E., 1989. Condensation and cloud parameterization studies with a mesoscale NWP model. Mon. Wea. Rev. 117, 1641-1657.

Turner, J., M. Row, 1989. Mesoscale vortices in the British Artarctic Territory. Polar and arctic lows. A. Deepak Publishing, Hampton, Virginia, USA, 347-356.

Twitchell, P.F., E.A. Rasmussen and K. Davidson, 1989. Polar and arctic lows. A. Deepak Publishing, Hampton, Virginia, USA, 419 pp.

Van Delden, A. 1989a. Gradient wind adjustment, CISK and the growth of polar lows by diabatic heating. Polar and arctic lows, eds P.F. Twitchell, E. Rassmussen and K.L. Davidson. A Deepak Publishing, Vi., USA, 109-130.

Seaman, N,L., H. Otten, R.A. Anthes, 1981. A rapidly developing polar low in the North Sea of Janunary 2, 1979. First international conference on meteorology and air/sea interaction of the coastal Zone, May 10-14, 1982. The Hague, Netherlands.

Seaman, N.L., 1983. Simulation of a mesoscale polar low in the North Sea. Sixth Conference on Numerical Weather Prediction, A.M.S., Boston, 24-30.

Shapiro, M.A., 1986. The arctic cyclone expedition, 1984: Research aircraft observations over the Norwegian and the Barents Sea. Polar lows project. Proceedings of the international conference on polar lows. Oslo, Norway, 20-23 May 1986.

Shapiro, M.A., L.S. Fedor, 1986. The Arctic Cyclone Expedition, 1984: Research and Aircraft Observations of Fronts and Polar Lows Over the Norwegian an Barents Sea, Part 1. Polar Lows Project, Tech. Rep. No. 20. The Norwegian Meteorological Institute, Oslo, Norway, 56 pp.

Shapiro, M.A., L.S. Fedor, T. Hampel, 1987. Research aricraft measurements within a polar low over the Norwegian Sea. Tellus, 37A, 272-306.

Shapiro, M.A., L.S. Fedor, 1989. A case study of en ice-edge boundary layer front and polar low development over the Norwegian and Barents Seas. Polar and arctic lows. A. Deepak Publishing, Hampton, Virginia, USA, 257-278.

Shapiro M.A., T. Hampel, L.S. Fedor, 1989. Reaserch aircraft observations of an arctic front over the Barents Sea. Polar and arctic lows. A. Deepak Publishing, Hampton, Virginia, USA, 279-290.

Shutts, G.J. 1990. Dynamical aspects of the October storm, 1987: A study of a successful simulation. Q.J.R.M.S. 116, 1315-1347.

Van Delden, A. 1989b. On the deepening and filling of balanced cyclones by diabatic heating. Meteorology and Atmospheric Physics, 40.

Wiin-Nielsen, A., 1989. On the precursors of polar lows. Polar and arctic lows. A. Deepak Publishing, Hampton, Virginia, USA, 85-108

Wilhelmsen, K., 1985. Climatological study of gale-producing polar lows near Norway. Tellus, 37A, 451-459.

Zick, C., 1983. Method and results of an analysis of comma cloud development by means of vorticity fields from upper tropospheric satellite wind data. Meteor. Rdsch, 36, 69-84.

Økland, H., 1977. On the Intensification of Small-Scale Cyclones Formed in Very Cold Air Masses Heated Over the Ocean. Inst. Rep. Ser. No. 26, Institutt for Geofysikk, Universitetet, Oslo, 25 pp.

Økland, H., 1983. Modelling the height, temperature and relative humidity of a well-mixed planetary boundary layer over a water surface. Boundary-Layer Meteor., 25, 121-141.

Økland, H., 1987. Heating by organized convection as a source of polar low intensification. Tellus, 39A, 397-407.

Økland, H., 1989. On the genesis of polar lows. In: Polar and arctic lows, eds P.F. Twitchell, E. Rassmussen and K.L. Davidson. A Deepak Publishing, Vi., USA, 179-190.