Dynamical Perspectives on the Formation and Intensification of Polar Lows

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Abstract

This thesis consists of a collection of scientific publications addressing dynamical aspects of the formation and intensification of polar lows. Polar lows are small scale, intense, short-lived cyclones developing over ice-free oceans at high latitudes.

In the first paper, environmental atmospheric conditions during polar low genesis are identified. These environmental conditions are classified based on the direction between the thermal wind and the mean flow in the lower troposphere. If the thermal wind and mean flow are in opposite direction the environment is classified as reverse shear, if they are in the same direction the environment is classified as forward shear. The two types of pre-polar low environments exhibit distinctly different features in terms of synoptic scale patterns, baroclinicity, configurations of sea surface temperature, depth and stratification of the troposphere, polar low propagation directions, and surface fluxes. The ambient polar low environment during forward shear conditions resembles typical mid-latitude baroclinic cyclogenesis, whereas the reverse shear conditions are characterized by the presence of a synoptic scale, occluded low over the genesis location, and a strong low-level jet.

In the second paper, a method to initialize baroclinic channel models is devised. The initial conditions are hydrostatically and geostrophically balanced, which makes them particularly suitable for studying rapid cyclogenesis. Furthermore, the method allows the definition of an arbitrary windfield, hence the initialization method exhibits a large degree of freedom in defining the initial conditions. We performed an unperturbed simulation which features minimal gravity wave activity, illustrating the balanced initial conditions. We also performed a perturbed simulation to examplify the utility of the setup. In this simulation baroclinic instability is triggered resulting in perturbation growth with time- and length-scales in agreement with observed polar low development.

In the third paper, an idealized baroclinic channel model is utilized to investigate the role of latent heating during polar low development in forward shear conditions. Within this idealized setup, a low-level, weak, cyclonic perturbation is able to amplify in absence of upper-level forcing, radiation, or surface fluxes. Crucial for rapid development is sufficient latent heat release in the north-eastern quadrant of the cyclone. The potential for latent heat release to occur depends on the environmental relative humidity, baroclinicity and static stability. Due to the Arctic configuration of the setup, the perturbation depth is relative shallow, which in turn improves the effectiveness of latent heat release on cyclone amplification.

List of publications

1. Annick Terpstra, Clio Michel, Thomas Spengler, *Atmospheric environments associated with polar low genesis in the North-East Atlantic*, *in preparation*
2. Annick Terpstra, Thomas Spengler, *A novel method to initialize idealized channel simulations*, Montly Weather Review, *under review*
3. Annick Terpstra, Thomas Spengler, Richard Moore, *Idealised simulations of polar low development in an Arctic moist-baroclinic environment*, Quarterly Journal of the Royal Meteorological Society, *under review*

Prologue

Polar lows are a specific type of meso-scale cyclones developing in relative cold-air masses at high-latitudes. These rapidly developing severe storms are accompanied by gale force wind-speeds, heavy precipitation, and rough sea states.

While explorers and inhabitants of areas subjected to polar lows have been aware of their possible devastating effects for many centuries, scientific interest in polar lows only commenced after the beginning of the satellite era. The observed cloud structures of polar lows show features of both mid-latitude baroclinic systems and tropical cyclones, albeit in both cases the diameter and life-time of polar lows are small compared to their analogues. Inspired by these structural resemblances, different theories were proposed to explain polar low development including baroclinic instability and convective development, either as stand-alone mechanism, combined, or sequential. However, despite several decades of research, none of the proposed theories adequately describe the evolution of polar lows. Consequently, our current understanding of the dynamical processes involved in the formation, intensification, and maintenance of polar lows are still under debate, and our forecasting skills for polar lows are still rather limited.

The aim of this thesis is to enhance understanding in dynamical mechanisms associated with polar low development, and to contribute towards a more unifying perspective on polar low development. Therefore, instead of focussing on individual polar low cases, a more general approach is adopted.

Key to understand polar low genesis is knowledge about the environment in which polar lows develop. To gain understanding in these environments we used an existing polar low database, and reanalysis data to determine distinctive features of the environments in which polar lows formation takes place. From this analysis it became clear that the two ends of a spectrum of polar lows are occupied by so-called reverse and forward shear conditions. The latter is similar to classical mid-latitude cyclogenesis environments, wheras the reverse shear version features a different configuration.

To investigate potential forcing mechanisms we developed an initialisation method suitable for idealised channel simulations of polar lows. As a starting point, we utilized this tool to investigate the potential of low-level cyclogenesis driven by latent heat release. Results from these simulations indicate, that when certain thresholds of moisture, baroclinicity and static stability are fulfilled, rapid intensification by latent heat release is possible in a typical Arctic environment.

The remainder of this thesis is structured as follows, starting with background information on polar lows (Chapter 1), followed by a short summary of the scientific papers developed during my PhD period (Chapter 2). In Chapter 3, I consolidate the main conclusions, followed by a brief discussion, and suggestions for future directions of polar low research. Finally, the scientific papers are presented (Chapter 4).

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Chapter 1

Introduction

1.1 Characteristics of polar lows

Suggested definitions for polar lows (e.g., *Reed*, 1979; *Carleton and Song*, 1997; *Rasmussen and Turner*, 2003; *Zahn and von Storch*, 2008) introduce various restrictions on size, intensity, cloud structure, propagation direction, etc. However, the majority of the definitions agree on two distinctive features: polar lows are maritime cyclones and develop poleward of the main baroclinic zone. Thus, the term polar low is typically applied to cyclones developing in ice-free regions located on the cold side of the mid-latitude extratropical jet-stream. These ice-free regions, and thus polar lows, prevail in both hemispheres (*Rasmussen and Turner*, 2003), with major polar low regions in the Arctic being the Bering Sea, Gulf of Alaska, and Nordic Seas. The latter is the only ice-free region located at high-latitudes, i.e. poleward of 60°N and extending as far as 80°N, and will be the main focus of the remainder of this thesis.

In the Northern hemisphere the majority of polar lows develops between November and April (e.g., *Wilhelmsen*, 1985; *Blechschmidt*, 2008; *Noer et al.*, 2011), designating them as a winter-time phenomena. Due to their spatial location at high-latitudes and their seasonal occurrence, a diurnal cycle induced by daily variance in incoming solar radiation is nearly absent during polar low development. Compared to mid-latitude or tropical cyclones, polar lows exhibit a short life-span, usually around a day (*Blechschmidt*, 2008), though sporadically they last more than 2 days (for example see, *Claud et al.*, 2004). Horizontal scales, estimated from satellite imagery, disclose that the diameter of meso-scale cloud structures in polar low regions ranges between 100-1000 km, with scales of ∼300 km being the most dominant (*Condron and Bigg*, 2006). While some of these cloud-structures are not associated with gale force wind-speeds, a commonly adopted low-level minimum wind-speed of 15 m s^{-1} is applied to classify a cyclone as polar low, with wind-speeds over 30 m s^{-1} reported (*Shapiro et al.*, 1987; *Føre et al.*, 2011).

Only a handful of targeted aircraft observations of polar lows are available (*Shapiro et al.*, 1987; *Bond and Shapiro*, 1991; *Douglas et al.*, 1995; *Brümmer et al.*, 2009; *Føre et al.*, 2011), most of them taken at the end of the polar low season (i.e., end of February or March). Therefore they might not be representative for the majority of the polar lows. However, due to the remote location and sparse observational network in polar low regions, these flights provide valuable insights in the structures associated with polar lows. Horizontal scales of the polar lows targeted by research aircrafts range between 300-500 km, and cloud-top observations indicate that most of these polar lows are relative shallow (cloud-top heights around 3 km), with observed maximum wind-speeds between 20-35 m s^{-1}. The core of polar lows is relative warm compared to the environments (+3-5 K), and in most cases frontal structures similar to mid-latitude cyclones accompany the polar lows. Estimated latent and sensible heat fluxes range between 200-500 W m^{-2}, where the ratio between these two fluxes is often close to unity.

1.2 Environmental configuration

The Nordic Seas area is prone to polar low development (e.g., *Wilhelmsen*, 1985; *Businger*, 1985; *Forbes and Lottes*, 1985; *Harold et al.*, 1999; *Blechschmidt*, 2008). Both the geographic configuration of the area, and the synoptic scale environment play a role in setting up favorable conditions for polar low development.

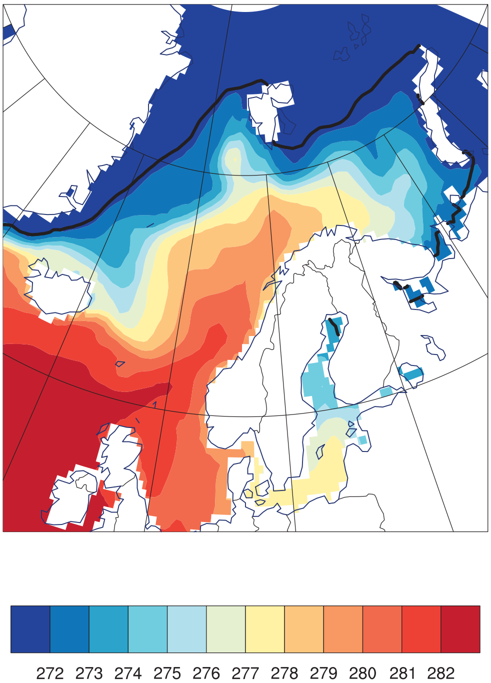

Figure 1.1: Nordic Seas area. Average sea surface temperature (shaded, units: K) and approximate ice-edge (solid line, ice-cover=50%) during 131 polar low cases.

Figure 1.1 provides an overview of the Nordic Seas region, and displays the mean sea surface temperature (SST) and the sea ice edge during 131 polar low events. The western part of the area is characterized by Greenland's high topography. The winter-time sea-ice-cover extends from Iceland up to the northern tip of Svalbard. East of

Svalbard the ice-edge continues from the southern tip of Svalbard up to Nova Zembla. This particular sea-ice configuration leaves a triangle-shaped wedge just west of Svalbard, as well as the remainder of the Norwegian and Barents Sea, ice-free ocean. Within the Nordic Seas region, this ice-free zone accommodates the development of polar lows. This relative large ice-free zone is due to surface ocean currents transporting warm water far into the Nordic Seas, resulting in an ice-free zone characterized by a strong SST gradient, ranging from colder SSTs at the ice-edge, towards warmer surface waters along the Norwegian coast. Typical for polar low events is a flow-pattern where cold air, originating from the ice-covered area, is advected over the relatively warm ocean (e.g., *Businger*, 1985; *Mallet et al.*, 2013), a condition commonly referred to as marine cold air outbreak (MCAO, *Kolstad*, 2011).

With regard to the atmospheric configuration, *Forbes and Lottes* (1985) examined both developing and non-developing cloud structures over the North-East Atlantic during an anomalous cold winter month, concluding that the relative location of the initial disturbance to the large scale environment is the distinguishing factor for development. By using superposed epoch-analysis, *Businger* (1985) laid out the synoptic scale pattern associated with polar low development over the Nordic Seas. The large scale configuration during polar low events is characterized by a trough located over Northern Scandinavia and a ridge located west of Greenland. This setup implies a geostrophic flow with a northerly component over the Nordic Seas, hence advection of cold air from ice-covered areas over the open ocean. Just west of the trough, over the Nordic Seas and polar low genesis region, a cold anomaly is found at upper levels, indicative for reduced static stability in the area below.

More recently, the increased resolution of reanalysis datasets made it possible to distinguish between several key regions of polar low development within the Nordic Seas. The results of these studies confirm the general pattern of off-ice flow and reduced static stability for different polar low genesis regions (*Blechschmidt et al.*, 2009; *Mallet et al.*, 2013). In addition, these studies point out several other parameters associated with polar low genesis, such as a positive potential vorticity (PV) anomaly at 300 hPa and high values of ‘temperature potential’, which is the difference between the SST and the temperature at 500 hPa. *Kolstad* (2011) suggested that the reduced tropopause height dictates the frequency of polar lows, whereas MCAOs determine the spatial location of polar low development.

Already recognized by *Businger* (1985), and confirmed and refined by *Mallet et al.* (2013), the synoptic scale environment conducive for polar low development consists of a build up period of approximately 4 days. During this pre-polar low period the corresponding height and temperature anomalies over the genesis area intensify, with a maximum just before the polar low occurrence, whereafter the anomalies decrease in intensity. Furthermore, the area of polar low genesis is preceded by the intensification of upper-level potential vorticity anomalies (*Blechschmidt et al.*, 2009; *Mallet et al.*, 2013), indicating enhanced stratospheric intrusions prior to polar low genesis.

1.3 Dynamical mechanisms

Controversy on the dynamical mechanisms associated with polar low development, emanated the widely accepted view of a spectrum of polar lows, with baroclinic systems on one end of the spectrum and convective systems on the other end (*Rasmussen and Turner*, 2003). Furthermore, during its life-cycle a polar low can 'shift' along the spectrum, e.g., starting as a baroclinic system and becoming convectively dominated as it matures (see e.g., *Nielsen*, 1997; *Nordeng and Rasmussen*, 1992).

1.3.1 Baroclinic instability

Baroclinic instability is a theoretical mechanism developed to explain the origin of wave-like patterns of cyclones and anticyclones in the mid-latitudes (*Charney*, 1948; *Eady*, 1949). In a baroclinic troposphere, density depends on both the temperature and pressure, yielding an uneven distribution of potential energy in this part of the atmosphere. Perturbation growth derived from the asymmetrical potential energy distribution, is denoted as baroclinic instability. During perturbation growth, simultaneous ascent of warm air and descent of cold air results in conversion of potential energy into kinetic energy, and thus a relaxation of the basic state potential energy gradient. In a 'dry' atmosphere, the growth rate of the perturbation is a function of the wavelength and depth of the perturbation, and the distribution of the potential energy.

By analysing vertical velocities derived from radar observation of a polar low crossing Britain, *Harrold and Browning* (1969) provided the first hints that baroclinic instability might be active during polar low development. In this case study, the derived vertical velocities were consistent with slantwise ascent along isentropic surfaces, as opposed to much stronger vertical motions usually accompanying deep convection. Their results were thus indicative for baroclinic instability during polar low development.

Motivated by these findings *Mansfield* (1974) applied a simple dry linear model to calculate the growth rates and wavelengths associated with polar lows. Static stability and shear parameters from the Harrold and Browning case served as input parameters for the model, resulting in predicted wavelengths and growth rates that agree well with the observed values. Extensions of the model with parameterizations of surface sensible heating and surface friction show that both processes significantly reduce the growth rate. Therefore, Mansfield concludes that polar lows develop in areas of weak low-level winds, and that dry baroclinic instability theory is able to explain the development of polar lows. Caveat in the Mansfield study is the assumed depth of the perturbation. As the planetary boundary layer (PBL) was capped by a strong inversion, Mansfield assumed that polar low development occurred within this layer, whereas vertical cross-sections of the case studied by Harrold and Browning clearly show that the polar low extended well above the PBL. Thus, the match between observed and predicted perturbation characteristics could be attributed to an unrealistic assumption of the perturbation depth. Another approach applying dry baroclinic instability to polar low development included analysis of normal mode solutions using a quasi-geostrophic model (*Duncan*, 1977). The growing perturbations were confined to the lowest 300 mb of the atmosphere and the bulk of baroclinic energy conversion occurred within this layer. Furthermore, predicted propagation speeds and wavelengths matched only with observations for 2 of the 3 considered cases. *Reed and Duncan* (1987) applied the same model to other cases and concluded that predicted wavelengths of the perturbation are in line with observations, but phase speeds and growth rates did not match well with observed values.

Overall, dry baroclinic instability theory is not very successful in predicting polar low development, and most studies attribute the failure to the exclusion of moisture in their analysis. Moisture affects the baroclinic instability mechanism mostly when condensation or evaporation occurs. The effect of these processes is a local heating or cooling of the atmosphere, resulting in a local redistribution of the potential energy, hence the generation of available potential energy. A necessary condition for condensation to take place is saturation. Thus, key for moisture to be effective in modifying the baroclinic mechanism is the cooling of air, usually achieved by lifting of air, such that the air adiabatically cools and eventually becomes saturated.

Sardie and Warner (1983) applied a three-layer, two-dimensional quasi-geostrophic model, and included simple parameterizations for latent heating to evaluate the possibility of polar low development by dry baroclinicity, moist baroclinicity, and Convective Instability of the Second Kind (CISK, *Charney and Eliassen*, 1964), including combinations of these mechanisms. The authors concluded that dry baroclinicity cannot be the sole mechanism explaining polar low development, and moist baroclinicity could explain polar low development in the Pacific area. However, for polar lows in the Atlantic area moist baroclinicity needs to be accompanied by CISK for realistic polar low development. Further evidence for the crucial role of condensation for realistic polar low development is provided by *Nordeng* (1987), who utilized a numerical model with explicit parameterizations for slantwise and deep convection, to simulate two polar low cases. Although two cases studied were not sufficient to draw general conclusions on the relative importance of both convection parameterizations, the study showed that realistic polar low development was not possible without latent heating. Furthermore, Nordeng noted the importance of the distribution of latent heating, with slantwise convection spreading the heating more horizontally, and deep convection increasing the height of latent heating. *Craig and Cho* (1988) utilized an Eady-model of baroclinic instability incorporating wave-CISK to investigate the role of latent heating during polar low development. Results from this study identify a threshold value for latent heating to become effective in modifying the baroclinic mechanism, this threshold depending on the vertical distribution of latent heating. Furthermore, lowering of the height of latent heating resulted in smaller wavelengths. For relative low values of latent heating the main modification of the baroclinic mechanism was by reducing the static stability, thus enhancing the growth rates, whereas for high values of latent heating the development becomes dominated by generation of available potential energy by latent heating with a minor role for baroclinic generation of eddy available potential energy.

These studies established the prominent role of latent heating in conjunction with baroclinic instability for polar low development, where latent heat acts to enhance growth rates and reduce wavelengths, in accordance with the small scales and rapid development of polar lows.

Another way to conceptualize baroclinic instability is by applying so-called potential vorticity-thinking (*Hoskins et al.*, 1985). Within this framework baroclinic instability is regarded as the interaction between anomalous PV blobs at different levels, where a positive feedback between the upper-level and lower-level PV anomaly results in enhancement of the low-level PV anomaly, hence an intensification of the cyclone. As PV is conserved for adiabatic, frictionless flow, PV fields can be directly associated with the temperature and velocity field by assuming a balanced state via the so-called invertibility principle (*Hoskins et al.*, 1985). An extension of this principle is piecewise PV-inversion, where, analog to electrostatics (*Bishop and Thorpe*, 1994), the relative contribution of different PV anomalies (usually with respect to a time-mean PV field) are individually assessed. Several studies applied this principle to polar low cases. However, they all indicate the complications with this approach, as PV-inversion techniques assume a balanced state, and negligible contributions from the divergent component of the flow. Polar lows are intense meso-scale cyclones, therefore they do not necessary fulfil these criteria. Keeping this in mind, these studies provide some insights in the relative contributions from individual PV anomalies for the individual polar low cases. Both *Wu et al.* (2011) and *Nordeng and Røsting* (2011) contributed the development throughout the life-cycle of the polar low in question to an intensifying positive PV anomaly at the tropopause, whereas for the polar low case investigated by *Bracegirdle and Gray* (2009) the upper-level PV anomaly only dominated in the initial phase of the development.

High resolution idealized baroclinic channel simulations provide the opportunity to systematically investigate polar low development in an idealized, and compared to case studies, easier to control environment. By utilizing such a tool, *Yanase and Niino* (2007) showed that gradually increasing the baroclinicity, resulted in a gradual transition of the developing cyclone from a more convective, spiraliform structure towards more baroclinic, comma-shaped cyclones. *Adakudlu* (2011) used a similar framework and confirmed the importance of baroclinicity and static stability on the development of polar lows.

Recognizing the enhanced growth rates and reduced wavelengths during cyclogenesis with strong latent heat release, *Montgomery and Farrel* (1992) developed a conceptual model for polar low development. During the first stage an upper-level positive PV anomaly interacts with a weaker low-level positive PV anomaly. The second stage is characterized by a dominant role for latent heat release. In this so-called diabatic destabilistation phase, the isentropes are locally stretched due to latent heating, thereby strengthening the low-level vorticity. This mechanism is self-enhancing and therefore able to sustain the polar low. Another proposed conceptual model for polar low development is type-C cyclogenesis, first identified by *Deveson et al.* (2002), and connected to polar low development by *Bracegirdle and Gray* (2008). Again, baroclinicity plays a prominent role in the initial setup of an environment favorable for rapid cyclogenesis. However, the development is dominated by latent heat release during the intensification phase. By distinguishing the ratio between upper-level and low-level contributions towards the forcing of vertical velocities, *Bracegirdle and Gray* (2008) assigned approximatelly 30% of the examined polar lows as type-C. Common denominator for both conceptual models is the relative importance of latent heat release.

1.3.2 Air-sea interactions

The majority of polar lows intensify and sustain over open water bodies (*Businger*, 1985), not over the adjacent ice-covered regions, alluding to the importance of air-sea interaction for their developments. Note that this does not imply that the initial triggering of polar lows cannot originate over ice-covered areas and/or orography in the surrounding area. During polar low development, the ocean provides an almost infinite source of moisture and serves as a large heat source due to the relative cold air during polar low events.

Mansfield (1974) was one of the first to explore the role of surface fluxes during baroclinic polar low development. His study utilized an Eady-model with simplified parameterizations for heating from the surface. The SST gradients in polar low regions are large, and for simplicity Mansfield only considered perturbations propagating along the SST gradient. Within this configuration, surface heating turned out to be detrimental to the growth rates, though wavelengths were reduced when surface heating was added. Mansfield concluded that favorable conditions for polar low development would include weak surface winds, corresponding to weak surface heating. A study by *Fantini and Buzzi* (1993) confirmed the reduced growth rates for additional surface fluxes.

Beside the effect of surface fluxes on modifying the baroclinic mechanism, surface fluxes can also serve as an energy source for maintaining and intensifying a pre-existing cyclone. To explain the maintenance of organized convective circulations, such as tropical cyclones, *Charney and Eliassen* (1964) introduced the CISK concept. CISK is based on low-level convergence, specifically moisture convergence, which provides a source for latent heat release. The release of latent heat enhances the low-level vorticity, and thus reinforces the low-level convergence, providing a positive feedback mechanism to sustain and intensify a cyclone. The effectiveness of CISK depends on the available amount of Convective Available Potential Energy (CAPE). Several studies (e.g., *Økland*, 1987; *Rasmussen*, 1979) argue for the CISK mechanism to be active in polar low development, implying the availability of a reservoir of CAPE during polar low events. Recently, an analysis of soundings from drop-sondes during polar low events, indicated that CAPE is nearly absent during polar low development (*Linders and Saetra*, 2010). Instead, the authors propose that CAPE is consumed at approximately the same rate as it is generated. Therefore, it seems unlikely that CISK plays a major role in the development of polar lows.

The validity of CISK for tropical cyclones was questioned by *Emanuel* (1986). He noticed that the atmospheric column during tropical cyclones was not able to provide a considerable resevoir of CAPE, and instead proposed a mechanism based on air-sea interaction, later renamed Wind Induced Surface Heat Exchange (WISHE), for maintenance and intensification of convective cyclones. The WISHE-mechanism is based on the direct relation between the low-level wind-speed and the surface fluxes. During WISHE the positive feedback is established by low-level moisture convergence, resulting in latent heating, which in turn enhances the low-level wind-speed, hence the surface fluxes, providing more moisture for latent heat release. *Emanuel and Rotunno* (1989) showed the applicability for WISHE during polar low development by utilizing an axisymmetric model, and concluded that at least some polar lows might be driven by WISHE. Furthermore, *Emanuel and Rotunno* (1989) emphasised that WISHE depends on an initial, sufficiently strong cyclone, in order for the mechanism to sustain the positive feedback. *Fantini and Buzzi* (1993) explored the possibility of an initial baroclinic atmosphere where surface fluxes provide the transition to more convective systems. By applying a simple 2D model, the results show that surface fluxes can transform a perturbation in a baroclinic atmosphere into a hurricane-like structure, where the energy from the heat fluxes is initially stored in the planetary boundary layer, until deep convection is initiated.

One factor complicating the assessment of CISK and WISHE for polar lows is that both theories assume a closed circulation in the vertical plane, i.e., they assume axisymmetric conditions. However, analyses of cloud structures from polar lows in satellite images, indicate that the majority of polar lows are not axisymmetric. *Blechschmidt* (2008) only attributed the symmetry property to 20% of the examined polar lows, whereas *Mitnik* (2009) identified approximately 30% as symmetric.

1.3.3 Complementary mechanisms

Baroclinic and convective mechansims constitute the majority of research on polar low dynamics, yet other processes might contribute to polar low development. One such mechanism is barotropic instability. In contrast to baroclinic instability, perturbation growth during barotropic instability is by conversion of basic state kinetic energy to eddy kinetic energy (*Vallis*, 2006). Studies assuming a barotropic troposphere concluded that barotropic instability probably does not provide a mechanism for polar low development (e.g., *Mullen*, 1979; *Reed*, 1979; *Fantini and Buzzi*, 1993). On the contrary, as noted by *Rasmussen and Turner* (2003), barotropic instability might be able to initiate small scale vortices along a shear line, thereby providing initially small cyclones which potentially develop into polar lows. Another factor in polar low development is the role of the ice-edge and orography in setting up favorable conditions for polar low development, highlighted in several case studies (e.g., *Martin and Moore*, 2006; *Adakudlu and Barstad*, 2011; *Nordeng and Røsting*, 2011). An additional, less explored contribution to polar low development is the role of radiative processes. *Craig* (1995) showed that, in an axisymmetric model, enhanced radiative cooling of the polar low environment, compared to lower cooling rates at the relative warm polar low core, can contribute to enhanced growth rates and increase the maximum intensity.

1.4 Numerical modelling

A detailed analysis of specific polar low cases provides insights in precursors, and dynamical mechanisms contributing to the development of individual polar lows. High resolution numerical models often complement those case studies, providing an additional tool to evaluate and investigate the respective dynamical mechanisms.

Numerical simulations can complement observations by providing a tool to generate information which is not available from observations. Furthermore, numerical models provide a platform to investigate the sensitivity of the development towards different components of the model. An example of such a sensitivity, is switching off latent heat release, which in the majority of the polar low cases resulted in weaker, less realistic polar low development (e.g., *Nordeng and Rasmussen*, 1992; *Bresch et al.*, 1997; *Claud et al.*, 2004; *McInnes et al.*, 2011). However, the different model-components are interacting with each other, and thus changes in the development cannot solely be attributed to the investigated parameter.

Another, more recently applied contribution of numerical models is the possiblity to perform ensemble simulation, thereby creating probabilistic forecasts (*Kristiansen et al.*, 2011).

Chapter 2

Compendium of scientific publications

Paper I: *Atmospheric conditions associated with polar low genesis in the North-East Atlantic*

As of now, a climatology of the meso-scale ambient polar low environment has not been compiled, rendering it difficult to assess the role of environmental shear on polar low development. In this paper, we provide an overview of the meso-scale environments in which polar low genesis takes place. Furthermore we identify two distinctive ambient polar low environments based on the lower tropospheric windshear configuration. The angle between the thermal wind and the mean wind-speed between 925-700 hPa is used to define the two different pre-polar low environments. The environment is classified as reverse shear, if the thermal wind and mean wind are in opposing directions, and it is classified as forward shear if they are in the same direction (Fig. 2.1).

The sub-synoptic configuration of the forward shear environment is similar to typical mid-latitude environments conducive to cyclogenesis. The lower-level and upper-level baroclinic zones are parallel, and a lower-level wave in the temperature field evolves in conjunction with an intensifying and migrating upper-level positive PV perturbation. The configuration of the reverse shear environment is dominated by an occluded, synoptic scale low, centered over the polar low genesis area. The circulation associated with the synoptic scale low results in strong cold air advection on the right side of the genesis region. Futhermore, the genesis region in reverse shear conditions exhibits by a more stationary upper-level PV anomaly. The reverse shear environment features a low-level jet to the right of the genesis location. Due to this low-level jet, the surface fluxes for the reverse shear environments are on average twice the surface fluxes of the forward shear environment.

Typical differences between the SST and the 2-m air temperature are around 7 K for both cases. However, the difference between the SST and the 500-hPa temperature is only > 43 K for the reverse shear cases, mainly due to the lower tropopause during reverse shear condtions, compared to forward shear conditions.

Forward and reverse shear conditions feature different synoptic scale settings. The reverse shear synoptic scale is characterized by a trough over Scandinavia, whereas the forward shear conditions are characterized by a ridge over Scandinavia. This results in a mean propagation direction which is southward for reverse shear cases and eastward for forward shear cases. However, in both setups cold air gets advection from the ice-covered regions over open ocean, indicative for a cold air outbreak.

(a) (b)

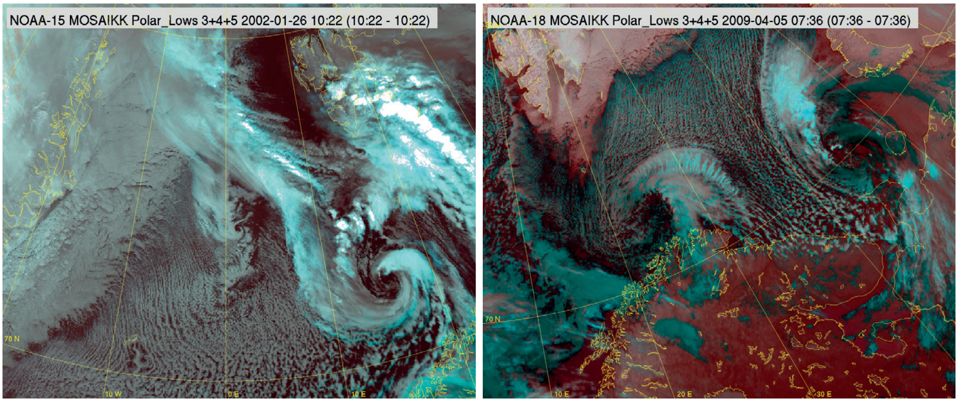

Figure 2.1: Satelite image for a reverse shear (a) and forward shear (b) polar low. Source: http://polarlow.met.no.

Paper II: *A novel initialization method for idealized channel simulations*

Due to the specific configuration associated with a polar low case study, it remains difficult to draw general conclusions valid for all polar lows. To gain understanding in the dynamical mechanisms underlying polar low development, one would need to study many polar low cases to identify common features. However, each polar low will still have its own specific characteristics in terms of ambient atmospheric environment and development.

One way to overcome these issues is by designing an idealization. Being similar to the envisioned situation, but stripped away of unnecessary complications the idealization provides a tool to study basic interactions in an idealized polar low environment. Furthermore, by utilizing a numerical model, the sensitivity of the idealized setup can be explored systematically. In this paper we devise a tool to initialize idealized baroclinic channel models suitable for polar low simulations.

Preferably, the initial state of an idealized baroclinic channel model is in geostrophic and hydrostatic balance. These requirements prevent the emergence of spurious gravity wave activity during the initial simulation time, which could interfere with the developing disturbance. The requirement of balanced initial conditions becomes more important for cyclogenesis with reduced timescales, such as polar lows, as less model spin-up time is allowed.

The method outlined in this paper, is based on analytically defining the wind field, whereafter the atmospheric fields are derived. These atmospheric fields are hydrostatically and geostrophically balanced, which is highlighted by performing an unperturbed simulation. This simulation features low gravity wave activity and minimal adjustments to the initial fields after a 5-days simulation period. In another simulation, we perturb the balanced fields by introducing a weak, surface based cyclonic vortex, to examplify the application of the setup to perturbation growth. In this simulation baroclinic instability is triggered and the developing cyclone exhibits length- and time-scales in agreement with polar low development.

Paper III: *Idealised simulations of polar low development in an Arctic moist-baroclinic environment*

Previous research addressing the influence of diabatic heating on cyclogenesis in a moist baroclinic environment identified the concept of a Diabatic Rossby Vortex (DRV), a mechanism associated with rapid intensification of marine cyclones in mid-latitudes. In this paper we explore the potential of DRV-like growth in a typical high-latitude environment.

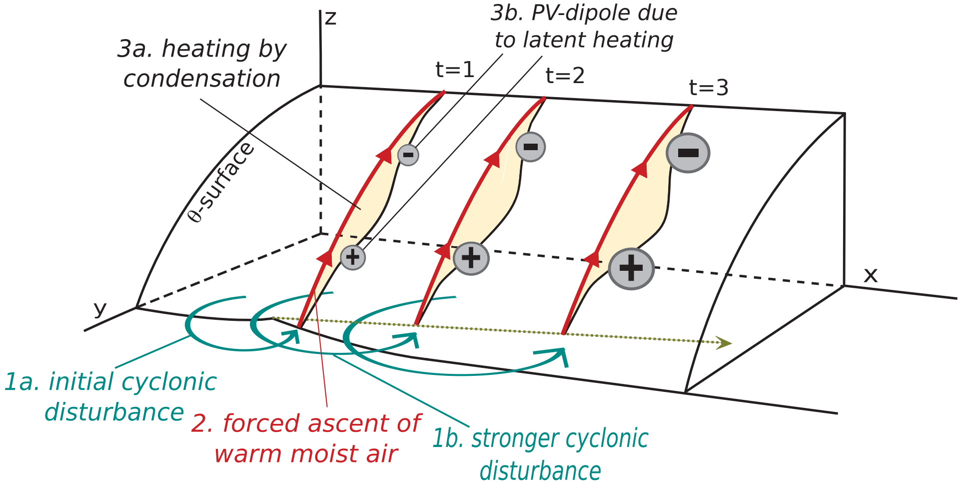

Figure 2.2: Schematic DRV mechanism.

The DRV refers to a solitary low-level cyclone vortex in a moist baroclinic environment, which is able to intensify and propagate independent of far field influences. Figure 2.2 provides a schematic of the mechanism characterizing the DRV. An initially weak, low-level cyclonic vortex features poleward flow on its eastern side, inducing upward transport of moist air along isentropic surfaces. As the moist air becomes saturated, condensation occurs, with latent heating locally stretching the isentropes, enhancing the vorticity at low levels, hence intensifying the low-level cyclone. This positive feedback reinforces the supply of moisture and thus the latent heating at mid-tropospheric levels providing an intensification mechanism for a low-level cyclonic disturbance in a moist baroclinic environment.

By utilizing an idealized baroclinic channel setup, we explore DRV-like growth in an Arctic moist-baroclinic environment. The initial setup consists of a baroclinic jet in thermal wind balance, and we mimic typical Arctic conditions by applying a low tropopause height and low surface temperatures. The setup is perturbed with a weak, surface based cyclonic vortex, which triggers the DRV-like growth. First, we contrast the development during dry and moist conditions. For dry conditions the growth rates are considerably weaker than for moist conditions. Analysis of the intensity and spatial distribution of the contributions to eddy available potential energy, and conversion of eddy available potential energy in kinetic energy indicates that the enhanced growth rates in the moist simulation are due to latent heat release in the north-eastern quadrant of the cyclone. Structurally, the developing disturbance in the moist simulation resembles typical DRV growth. However, compared to mid-latitude DRV development, the depth of the DRV is more shallow in our simulations. Due to lower temperatures in Arctic environments, the specific humidity is lower than for mid-latitude environments, hence the maximum values of latent heat release in an Arctic environment are smaller. However, the effects of latent heat release, i.e., increasing the vorticity due to stretching of the isentropes, is dependent on the vertical structure of latent heating. Therefore we surmise that the shallowness of the DRV structure in our Arctic simulations, is key for DRV-like growth in Arctic environments.

A suit of sensitivity experiments addresses the relative importance of moisture, baroclinicity, and static stability for the potential of DRV-like polar low development. All explored parameters exhibit a threshold value, which needs to be overcome before rapid cyclone growth is established. However, it should be noted that these threshold values are interdependent and thus should not be interpreted as absolute values.

Chapter 3

Discussion

3.1 General conclusions

The conclusions from the publications can be summarized as follows:

- The angle between the lower tropospheric thermal wind and mean wind in the ambient pre-polar low environment is suitable to distinguish between two distinct polar low environments, i.e., forward and reverse shear. Similarities and differences between these two types are summarized in Table 3.1.
- Given an arbitrary 2D wind field in a z-x plane, a method for deriving the associated hydrostatically and geostrophically balanced atmospheric fields is devised. These fields are suitable for initialization of idealized channel models.
- In an idealized Arctic moist-baroclinic environment, low-level perturbation growth is possible in absence of surface fluxes, friction, radiation, and upper-level disturbances. Crucial for rapid perturbation growth is latent heat release in the north-eastern quadrant of the cyclone.

Table 3.1: Comparison of forward and reverse shear conditions. Hereafter, LL means lower-level and UL means upper-level.

Variable	Forward Shear	Reverse Shear
Propagation direction I	parallel to LL baroclinic zone	perpendicular to LL baroclinic zone
Propagation direction II	Eastward	Southward
Temperature advection	negligible outside the perturbation	strong cold air advection on the right side
Tropopause	strongly sloping	nearly flat
Main jet location	UL jet on the right side	LL jet on the right side
Development location	warm side of the LL baroclinic zone	warm side of the LL baroclinic zone
UL PV+ anomaly	intensifying, migrating	broadening, stationary
Sub-synoptic scale	deep baroclinic zone	occluded low
Surface fluxes LH/SH	$\sim$70/70 W m^{-2}	$\sim$140/140 W m^{-2}
SST-T$_5$00	$<$ 43 K	$>$ 43 K
SST-T$_2m$	6-7 K	6-7 K
Synoptic scale pattern	Scandinavian Blocking	Atlantic Ridge

3.2 Brief discussion

The separation between reverse and forward shear environments is based on a dynamical configuration of the ambient pre-polar low environment. The two configurations are distinctly different and provide a starting point for examining the dynamical mechanisms involved in forward and reverse shear polar low development. Note that this classification does not separate between convective and baroclinic systems. Both environments exhibit a similar intensity of baroclinicity. However, the effect of the baroclinicity might be different for both cases. If one extrapolates findings for meso-scale convective systems to polar lows, the storm relative windshear could be key in polar low development (*Markowski and Richardson*, 2006). The development in forward shear cases seems very similar to typical mid-latitude cyclogenesis, whereas for the reverse shear cases our hypothesis is that the vertical shear is able to organize the convection, leading to subsequent meso-scale growth.

By exploring the posibility of DRV-like development for polar lows, we investigated the forward shear configuration. Potentially, this mechanism might not be applicable to polar lows developing in reverse shear environments. However, in paper III, we showed that despite the relative low absolute values of moisture, latent heating can play a dominant role in Arctic areas. Due to the shallowness of the polar lows, the vertical gradients of latent heating can be large, even for relative low absolute values of latent heating. This finding is independent of forward of reverse shear, and thus applicable to both cases.

3.3 Future outlook

One valuable way of advancement in science is the development of conceptual models and meaningful classifications. A well-known example of a conceptual model in meteorology is the Norwegian cyclone model (*Bjerknes*, 1919). Although improvements and adjustments have been made, it still serves as an archetype for cyclone development. Similarly, the Beaufort scale (*Garbett*, 1926) is an example of a classification, so powerful that it is still used around the world to describe wind strengths. These examples emerged from a vast body of experience and insight in the respective phenomena. More recent examples, such as the development of the RKW-theory (describing the influence of shear on squall lines, *Weisman and Rotunno*, 2004), or the concept of warm and cold conveyor belts (*Harrold*, 1973), indicate that major advancements in understanding atmospheric phenomena can be achieved by developing conceptual models. However, with respect to polar lows we are still lacking a useful conceptual model, rendering it difficult to understand these phenomena. In my opinion this is the major challenge for future polar low research, to identify the dominant mechanisms for polar low development and develop an in-depth understanding of the relation between the different processes to construct appropriate conceptual model(s).

Although observations, including reanalysis data, and high-resolution numerical studies provide a starting point for polar low exploration, the idealized framework provides a valuable test-bed for unraveling the dynamics and shedding light on the sensitivity of possible mechanisms. Below, I outline and discuss some possible avenues to identify the major atmospheric processes and their interactions during polar low development.

3.3.1 Cold air outbreaks

Polar low development is often associated with MCAOs (e.g., *Grønås and Kvamstø*, 1995; *Bond and Shapiro*, 1991; *Kolstad*, 2006; *Brümmer et al.*, 2009; *Kolstad*, 2011). Even though, it is unclear whether polar low development takes place in the cold air masses themselves or along the frontal zone associated with the MCAO. Some research addresses the development of the PBL during MCAOs (e.g., *Claud et al.*, 1992; *Hartmann et al.*, 1997), but we still know little about the frontal structure associated with the MCAO. These frontal structures might be crucial for polar low development by enhancing surface fluxes due to along frontal jets, or by providing environmental shear to organize convection, or other currently unknown mechanisms.

The results in Paper I (Chapter 4) indicate that polar low development takes place on the warm side of the cold front structure associated with the MCAO, both for forward and reverse shear environments. Another indicator for the potential role of fronts during polar low genesis is provided by *Noer et al.* (2011). The authors examplified polar lows development at the edge of a MCAO. They also indicated that some meso-scale cyclones develop within the cold-air masses, i.e., not at the frontal zone. However, these structures are generally weak and unable to penetrate the inversion capping the PBL and should therefore not be called polar lows.

To get a first notion of the spatial location of polar lows relative to frontal zones, we applied a frontal detection algorithm (*Schemm et al.*, 2014) to detect frontal zones in the vicinity of the polar lows examined in Paper I (Chapter 4). Examples of polar low tracks and the detected frontal structures are shown in Figure 3.1. From a first inspection it seems that indeed the majority of the polar lows develop in the vicinity of the MCAO cold front.

The above indicates the potential importance of cold fronts associated with MCAO for polar low development. Therefore, understanding of the structure of these fronts

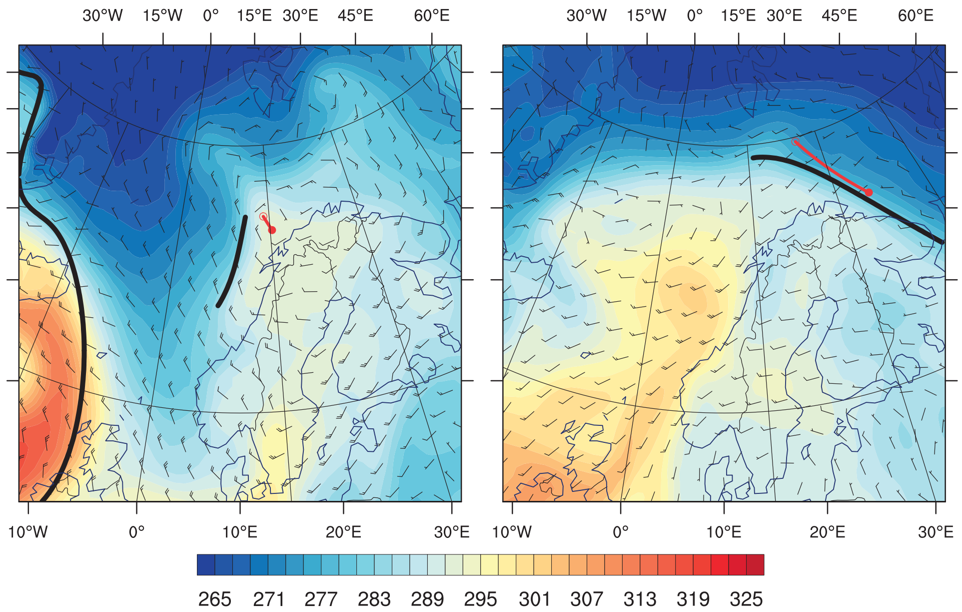

Figure 3.1: Examples of detected fronts (thick black lines) and polar low tracks (red line, open circles represent the start location and closed circles the end loaction of the track). Shown are the equivalent potential temperature (shaded) at 850 hPa and the windbarbs at 850 hPa.

and insight in the development of instabilities along these fronts might provide insights in polar low genesis. One of the first approaches to investigate frontal instablities in relation to polar lows was performed by *Mingalev et al.* (2014). The authors conclude that convexity, i.e., a zonal variation in the shape of the Arctic front, is able to trigger polar lows.

3.3.2 Surface fluxes and SST gradients

Polar lows develop in areas with a strong SST gradients, a feature also found in breeding areas for extratropical cyclones, such as the Gulf Stream around the East Coast of the US or the Kuroshio Current to the east of Japan (*Hoskins and Valdes*, 1990). A gradient in SST can provide a source for baroclinicity by differential heating of the atmosphere (*Cione et al.*, 1993). Also meandering eddies along the frontal zone can provide a local source of heating, thereby functioning as a triggering mechanism for cyclogenesis. However, the role of the SST gradients and ocean surface eddies on the development of polar lows has not received much attention, yet.

To gain insight in the influence of SST gradients in relation to polar low development, understanding the influence of surface sensible and latent heat fluxes on the formation of polar lows is key. As a starting point we performed a number of idealized channel simulations. The initial conditions are similar to the control run in Paper III (Chapter 4). The simulations include a boundary-layer parameterization and we mimic a cold air outbreak by introducing a SST which is 9 K warmer than the surface air temperature. This results in a SST gradient similar to the SST gradient in the Nordic Seas during winter. To compare the relative influence of the latent and sensible heat flux, a simulation with both fluxes switched on (SFX), and simulations with only latent (QFX) or only sensible (HFX) heat flux are performed.

The time-evolution of the vertically integrated eddy kinetic energy within a box (dimensions 1200x1600 km) centered around the developing disturbance, indicates that the simulations with only sensible (HFX) or only latent heat fluxes (QFX) intensify at a comparable rate, with a slightly faster and stronger intensification for the simulation with only latent heat flux (Fig. 3.2a). However, the simulation with both fluxes switched on (SFX) exhibit a much larger growth rate and stronger intensity, than expected from the individual contributions of each of the fluxes. Thus, sensible and latent heat fluxes seem to be more effective for cyclogenesis when they occur in conjunction.

Another interesting feature of this set of simulations is the contributions from heating terms towards the eddy available potential energy (for the equation, see Paper III, Eq. (6)). These contributions are negative for the simulation with only sensible heat fluxes (Fig. 3.2b). Local heating in the model setup is via the cumulus, micro-physics, or boundary-layer parameterization. A closer examination shows that the negative contributions originated from the boundary-layer parameterization, where heating of the boundary layer air is co-located with negative temperature deviations from the zonal mean (not shown).

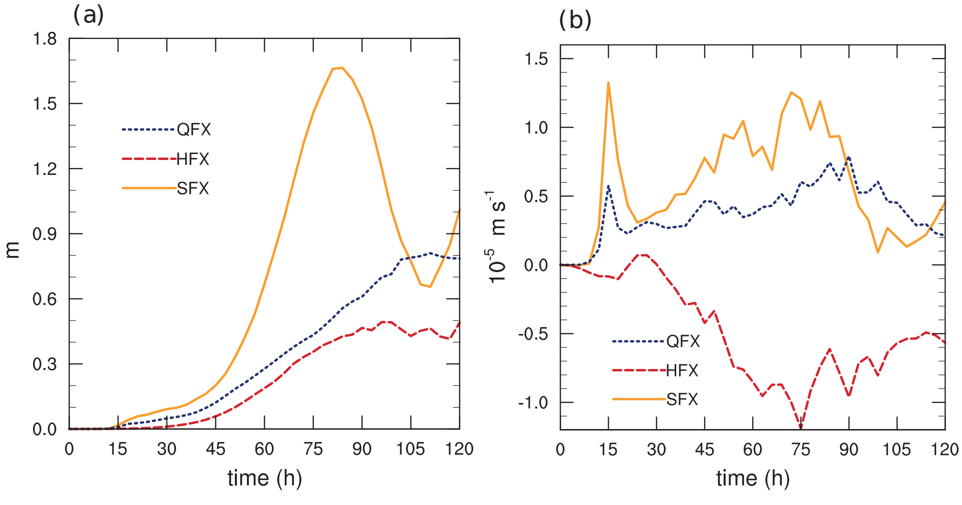

Figure 3.2: Time evolution of the vertically integrated eddy kinetic energy (a) and energy conversions of local heating to eddy available potential energy (b), for experiments SFX, QFX and HFX (see text for a description of the experiments).

These simulations are just an example of the potential of the idealized framework to gain understanding in the role of surface fluxes and SST gradients on perturbation growth. By defining appropriate initial conditions the idealized framework is suitable to identify air-ocean configurations favorable for cyclogenesis and the effects of these configurations on cyclogenesis.

3.3.3 Role of upper-level forcing

Currently, the available conceptual models for polar low development (*Montgomery and Farrel*, 1992; *Bracegirdle and Gray*, 2008) both incorporate an upper-level perturbation as the triggering mechanism for polar low development. The results in Paper I (Chapter 4) indicate that a migrating upper-level PV-perturbation is present in forward shear environments. However, for reverse shear environments the upper-level PV-perturbation is more stationary. Furthermore, we showed in Paper III (Chapter 4) that a lower-level perturbation is able to intensify in the absence of upper-level forcing. Thus, it remains unclear if upper-level forcing is always necessary for polar low development.