The multiscale aspect of cirrus cloud dynamics

A dissertation submitted to the
ETH ZURICH

for the degree of
Doctor of Sciences

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2011
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Abstract

Cirrus clouds like all other clouds are important modulators of the Earth’s radiative budget and cover up to 30% of the Earth’s surface. For low (liquid) clouds, the albedo effect (i.e. cooling) is much larger than the greenhouse effect (i.e. warming), but for cirrus clouds, both effects are of comparable magnitude. The transition between net warming or cooling cirrus depends on microphysical properties such as the shape or size of ice crystals. It is currently believed that the warming effect dominates on a global scale. However, it is rather difficult to provide global estimates of the radiative effect of cirrus clouds because very little is known about their life cycle and potential formation regions. Therefore, the variety of large scale models that are used for such estimations show uncertainties in the representation of high-cloud amounts.

Generally, ice crystal formation at low temperatures ($T < 235\,\text{K}$) depends crucially on local dynamics. These dynamical processes act on various scales and influence the efficiency of cirrus cloud formation substantially. Until now, cloud resolving studies usually used the stratified (large scale) flow evaluated in almost steady state conditions, i.e. after a long time until a quasi-equilibrium has been reached. However, atmospheric flow conditions often change with time. Thus, in order to treat the multiscale aspect of cirrus formation properly, also time dependent large scale dynamics must be taken into account. For this purpose transient environmental or ambient fields have been implemented into the non-hydrostatic, anelastic model EULAG in order to control large scale dynamics and to investigate the impact on smaller scales in detail. The model includes a bulk ice microphysics scheme and a fast two-stream radiative transfer code.

In the first part, the influence of radiative cooling and small eddies on cirrus formation was investigated. Using idealized profiles with high ice supersaturations of up to 144% and weakly stable stratifications with Brunt-Vaisala frequencies down to $0.0018\,\text{s}^{-1}$ within a supersaturated layer, the influence of radiation on the formation of cirrus clouds is quite large. Due to the radiative cooling at the top of the ice supersaturated layer with cooling rates down to $-3.5\,\text{K}/\text{d}$, the stability inside the ice supersaturated layer decreases with time. During destabilization, small eddies induced by Gaussian temperature fluctuations start to grow and trigger first nucleation. These first nucleation events then induce the growth
of convective cells due to the release of lantent heat and the continuing radiative destabilization. The effects of increasing the local relative humidity by cooling due to radiation and adiabatic lifting lead to the formation of a cirrus cloud with IWC up to 33 mg/m$^3$ and mean optical depths up to 0.36. In a more stable environment, radiative cooling is not strong enough to destabilize the supersaturated layer within 8 hours; no nucleation occurs in this case.

In the second part of the thesis, the influence of time dependent environmental flows on orographic waves, and subsequently on the formation of orographic cirrus clouds was investigated. For this purpose, an idealized stratified flow over a single mountain ridge in the center of the domain was investigated. As a reference, non-transient simulations using three constant horizontal flow profiles (i.e. low, medium and high-wind) were performed. The transient simulations contain an increasing wind velocity (i.e. medium to high wind) and a decreasing wind velocity (medium to low wind) case, changing from one flow regime to the other after $t = 180$ min within the transition time $\tau = 60$ min. In all simulations, the generated orographic wave propagates up to an ice supersaturated region (ISSR) placed between 8000 and 11000 m altitude containing high supersaturations up to 125%. The adiabatic cooling in the ascending branches of the wave leads to the formation of orographic cirrus clouds with ice water contents of up to 140 mg/m$^3$. Whereas the waves generated by a non-transient flow are stationary, the waves generated in a transient regime show significant horizontal and vertical displacements of the updraft/cloud regions due to an imbalance between the conserved horizontal phase velocity and the changing environmental flow. This results in a different vertical velocity regime inside the ISSR, namely, horizontal displacement of updrafts and downdraft and also a change from an updraft dominated to a downdraft dominated regime (and vice versa) occurs due to a changing vertical wavelength. The impact on cirrus cloud formation is as follows: The transient simulations show a larger spatial distribution of cloud ice due to the horizontal displacement of updraft regions, whereas the non-transient simulations are dominated by stationary clouds directly over the mountain ridge.

The achieved knowledge from this idealized study was used to describe and understand transient orographic cirrus formation in a much more realistic framework. The model was set up with realistic terrain data from cross-sections through the Andes and reanalysis-data for wind, temperature and pressure (source: ECMWF) and the results are compared with aircraft measurement data from the INCA campaign. Again, a set of transient simulations with time dependent ambient states for potential temperature and horizontal wind (both are constantly increasing with time) were performed, in addition to the non-transient simulations, in order to classify the benefit of time dependent ambient fields in EULAG. In all simulations, the generated orographic gravity waves propagate up to an ice supersaturated region...
between 8500 and 9800 m altitude containing high supersaturations up to 130\% and lead to the formation of orographic cirrus clouds. The simulated vertical velocities are in a good agreement with the measurements, but the microphysical properties show some offsets (i.e. slight overestimation of IWC and underestimation of ICNC in the simulations). Nevertheless, the horizontal position of the formed cirrus clouds represents the measurements very well. The transient simulations show a remarkably better accuracy than the non-transient ones, by containing a larger horizontal variability of cloud ice. This was validated in a series of subsequently performed idealized simulations, using a single mountain ridge with spatial extensions comparable with the Andes. Additionally, the idealized simulations give a deeper insight into the flow regime over the realistic terrain. It follows that the transient INCA simulations start with environmental conditions that favor wave breaking, but after some time the increasing wind leads to more stable conditions. During the transition between the non-linear and the linear regime, dynamical features caused by the non-linear flow tend to be advected downstream. This transition between two different regimes is completely ignored by the non-transient simulations.

The presented results lead to following important conclusions: The kind of textbook knowledge in the cirrus community that radiation might not be important for the initial formation of cirrus clouds should be revised under the comprehension of the discussed destabilization process. On the other hand, the introduction of time dependent large scale dynamics to a cloud resolving model leads to changes in mesoscale/smallscale dynamics and microphysics, which are not predictable by using quasi-steady state conditions. Therefore, the multiscale aspect of cirrus formation is much better represented by transient simulations. This could be verified by comparing transient and non-transient simulations of orographic cirrus clouds with aircraft measurement data.
Zusammenfassung


Im ersten Teil der Arbeit wird der Einfluss von Strahlungskühlung und kleinen Wirbeln auf die Zirrusbildung untersucht. Dazu werden idealisierte Atmosphärenprofile mit hohen Eisübersättigungen von bis zu 144% und schwach stabile Schichtungen mit Mindestwerten für die Brunt-Väisälä Frequenz von $0.0018$ s$^{-1}$ (innerhalb der übersättigten Schicht) verwendet. Unter diesen Voraussetzungen ist der Einfluss der Strahlung auf die Zirrusbildung markant. Bedingt durch eine Strahlungskühlung am Ober-
rand der übersättigten Schicht von bis zu −3.5 K/Tag verringert sich die Stabilität der Region mit der Zeit. Während des Destabilisierungsvorganges wachsen kleine Wirbel, welche durch anfängliche Gauss-Fluktuationen der Temperatur entstanden sind, an und lösen erste Eiskristall-Nukleation aus. Diese ersten Nukleationsereignisse führen dann durch das Freigeben von latenter Wärme und die weitere Destabilisierung der Schicht durch Strahlungskühlung zu einem Anwachsen von kleinen konvektiven Zellen. Die Erhöhung der lokalen relativen Feuchte durch Strahlungskühlung sowie durch adiabatische Kühlung (d.h. Anhebung von Luftpaketen; dominant) führt zur Bildung von Zirruswolken mit einem Eiswassergehalt von bis zu 33 mg/m³ und einer optischen Dichte von bis zu 0.36. In einer stabileren Atmosphäre ist die Strahlungskühlung nicht stark genug, um die übersättigte Schicht innerhalb der 8 Stunden Simulationszeit zu destabilisieren; in einem solchen Fall bilden sich keine Zirruswolken durch den beschriebenen Effekt.


Verwendung von konstanten Umgebungsvariablen nicht prognostiziert werden können. Deshalb wird der multiskalene Aspekt der Zirrusbildung durch transiente Simulationen deutlich besser berücksichtigt, was auch der Vergleich mit Messdaten bekräftigt.
Chapter 1

Introduction

1.1 Cirrus clouds

The role of clouds is crucial for our understanding of the current and the changing climate (IPCC, 2007). Unfortunately, our knowledge of the contribution of clouds to radiative forcing is still limited. Because of the insufficient representation of cloud processes in existing climate models it is difficult to predict the role of clouds in a changing climate and a changing hydrological cycle. Generally, clouds warm and cool the atmosphere depending on their properties such as water content, droplet sizes, cloud height etc. For low level (liquid) clouds, the reflection of shortwave radiation at the cloud top (i.e. albedo effect; cooling) is much larger than the absorption of long wave radiation (i.e. greenhouse effect; warming), but for cirrus clouds, both effects are of comparable size (but with different signs, see e.g. Fig. 1.1, Zhang et al., 1999). Generally, optically thin cirrus are more leading to a warming, where for higher optical depths a cooling is more appropriate. The transition between a warming and a cooling regime is very sensitive on microphysical properties such as shape or size of ice crystals (e.g. Fusina et al., 2007; Wendisch et al., 2007) and furthermore these properties depend a lot on the interaction of dynamical and thermodynamical processes (i.e. cirrus formation mechanisms). Therefore, it is rather difficult to provide global estimates of the radiative effect of cirrus clouds because very little is known about the life cycle of cirrus clouds and their potential formation regions, i.e. ice-supersaturated regions (see e.g. Gierens et al., 1999; Spichtinger et al., 2003a,b). Until now it is estimated that the global net effect of cirrus clouds tends to warm the earths atmosphere (Chen et al., 2000). However, the variety of large scale models that are used for such estimations show uncertainties in the representation of high cloud amounts (Zhang et al., 2005).

Cirrus clouds are high tropospheric clouds, purely consisting of ice crystals and covering approximately 30% of the Earth’s surface (Wylie and Menzel, 1999). The formation of cirrus clouds requires high supersaturations with respect to ice, what becomes obvious when comparing the spatial distribution of
cirrus and ISSRs (see Spichtinger et al., 2003b). The existence of cloud-free air masses that are super-saturated with respect to ice in the upper troposphere or lowermost stratosphere has long been known. More than 60 years ago, Glückauf (1945) obtained values of relative humidity with respect to ice (RHi) up to 160% over Southern England. Such measurements have been termed as errors for a long time. However, during the last two decades the existence of ice supersaturated airmasses has been proven by many measurements using a variety of different measurement techniques (e.g. Jensen et al., 1998; Vay et al., 2000; Ovarlez et al., 2000; Krämer et al., 2009). The properties and global distributions of ISSRs have been determined over recent years (e.g. Spichtinger et al., 2003a,b; Gettelman et al., 2006). They occur 20 to 30% of the time in cloud free air masses in the upper troposphere over the North Atlantic. The large horizontal extent of ISSRs, with mean path lengths of about 150km, and in some rare cases of a few thousand kilometers (Gierens and Spichtinger, 2000) means that there is a substantial amount of cloud free air masses with enhanced water vapor content.

The amount of supersaturation needed for cloud formation in the cold troposphere (i.e. $T < -38^\circ$C) depends on the freezing process. There are two different freezing processes for ice crystals in the upper troposphere, namely homogeneous and heterogeneous nucleation. For homogeneous nucleation small super-cooled droplets consisting of aqueous solutions (e.g. sulfuric acid/water) freeze spontaneously if the relative humidity surpasses a threshold humidity. Koop et al. (2000) showed that this threshold only depends on temperature (and marginally on the droplet size) and is independent on the nature of the solute. For heterogeneous nucleation a solid aerosol particle (so-called ice nucleus) initiates freezing. Here the critical supersaturation required for the initiation of the freezing process strongly depends on
the properties of the ice nucleus. Several different heterogeneous freezing processes are distinguished (e.g. immersion freezing, deposition freezing, contact freezing, see DeMott et al., 2003). Generally, the freezing threshold is lower for heterogeneous nucleation. But due to the lack of enough efficient ice nuclei in the upper troposphere, the frequently measured high ice crystal number concentrations and the occurring high supersaturations can not be explained by heterogeneous freezing theory. Therefore, homogeneous freezing can be considered as the dominant freezing mechanism for cirrus clouds (e.g. DeMott et al., 2003; Haag et al., 2003). Nevertheless, recent studies showed that additional heterogeneous ice nuclei could influence the homogeneous freezing event and modify the properties of the formed cirrus clouds (see Gierens, 2003; Spichtinger and Gierens, 2009b; Spichtinger and Cziczo, 2010).

Beside moist advection, the temporal increase of RHi can be described by following equation, assuming a vertical updraft:

$$\frac{dRHi}{dt} = \left[ \frac{\partial RHi}{\partial T} \frac{dT}{dt} + \frac{\partial RHi}{\partial p} \frac{dp}{dt} \right]_{\text{source}} + \left[ \frac{\partial RHi}{\partial q} \frac{dq}{dt} \right]_{\text{sink}}$$

where $\frac{dq}{dt} = -\frac{dq_c}{dt}$ and $q_c$ denotes the cloud ice mixing ratio. Because the pressure dependent part in the source term is small compared to the temperature dependent part, the main source for RHi can be stated as cooling due to adiabatic expansion (i.e. lifting of the air parcel) or radiation (i.e. temporal imbalance between absorption and emission). The sink term is dominated by the depletion of water vapor due to diffusional growth of ice crystals. Before the critical supersaturation for freezing is reached, the RHi linearly increases. After the nucleation, the source (cooling) and sink (growth) term start to compete, until the supersaturation is depleted and a new equilibrium is reached (see figure 1.2). The most effective source for the adiabatic expansion are synoptic-scale vertical motions, gravity waves, convection and turbulence.

1.2 Multiscale aspect of cirrus formation

In the atmosphere, microphysical processes act within a few seconds (or even milliseconds) to initiate ice crystal nucleation and growth, whereas the dynamical processes (from the turbulence on the small spatial and temporal scales to mesoscale waves and large-scale atmospheric flows) occur on longer timescales and influence the subsequent evolution of clouds (i.e. provide adiabatic cooling to reach the required supersaturations). Therefore, dynamical processes acting on various scales influence the efficiency of
Figure 1.2: Timeline of a homogeneous nucleation event.

cirrus cloud formation substantially, as the magnitude of the vertical wind speed and the ambient temperature determines the produced ice crystal number concentrations (Kärcher and Lohmann, 2002). The multiscale nature of both, dynamical and microphysical processes determining the formation and evolution of cirrus clouds can be described as follows:

- Large-scale processes: One example is the slow upward motion in the warm sector of a cyclone, i.e. along the fronts. Also in warm conveyor belts air parcels can rise up to the tropopause level (Spichtinger et al., 2005b). These phenomena act on long time scales ranging from several minutes to hours. Synoptic-scale processes induce vertical wind speeds in the range from 1 to 10 cm/s, resulting in cirrus clouds with low ice crystal number concentrations (Kärcher and Lohmann, 2002) and, probably, with a rather homogeneous spatial distribution of ice crystals. Another large-scale process is the slow change of the temperature due to radiative cooling/heating. Cooling rates down to $-2\, {\text{K/d}}$ at the top of ice supersaturated regions can change the stratification of weak stable layers and trigger the formation of a cirrus cloud (Fusina et al., 2007; Fusina and Spichtinger, 2010).

- Mesoscale processes: Large scale motions are often superimposed by mesoscale processes. Prominent sources for these mesoscale motions are orographic gravity waves (e.g. Joos et al., 2008). Other sources for internal gravity waves are geostrophic adjustment (Spichtinger et al., 2005a) or convection (e.g. Bretherton and Smolarkiewicz, 1989). These mesoscale processes occur on much smaller temporal (few minutes) and spatial scales (Gary, 2010); the associated vertical wind speeds are much larger and therefore the ice crystal number densities are much higher compared to those of the large-scale flows. Thus, mesoscale processes can trigger the formation of ice crystals in a
1.2. Multiscale aspect of cirrus formation

crucial way (see e.g. Kärcher and Ström, 2003; Haag and Kärcher, 2004; Hoyle et al., 2005).

- Microphysical processes: The actual microphysical processes inside cirrus clouds (nucleation, growth etc.) act on much shorter timescales; for instance, the freezing of supercooled droplets to ice crystals occurs within milliseconds. In numerical models usually these processes cannot be resolved in space and time, but they are parameterised according to process studies and (field and/or laboratory) measurements.

Additionally, the internal dynamics of cirrus clouds during their life cycle is characterised by latent heat release and turbulence (see e.g. Demoz et al., 1998; Smith and Jonas, 1996). In a weakly unstable or even neutral environment, small convective cells can be formed by latent heat release which trigger the formation of numerous ice crystals within a very narrow range. Hence, large horizontal gradients of ice crystal number densities are generated, which results in inhomogeneities.

Turbulence inside cirrus clouds is very common at mid-latitudes. There, a coexistence with internal gravity waves has been observed (see Quante, 2006). Breaking gravity waves seem to be the most important source for turbulence in cirrus clouds (‘no turbulence without waves’, see McIntyre (2008)). The highest intensity of turbulence inside cirrus clouds was observed with the simultaneous strong wind shear. Unfortunately, in atmospheric measurements, the impacts of different atmospheric motion scales cannot easily be separated - the motion of the observed air parcels is produced by superposition of large-scale, mesoscale and small-scale motions. This makes it very difficult to quantify, which atmospheric scales are most effective for the formation and evolution of cirrus clouds in a distinct situation.

In general, the question arises how processes on different spatial and temporal scales change the formation and evolution of cirrus clouds. In order to investigate cirrus clouds, a cloud resolving model is essential. Due to the high resolution of such a model and due to limited computational power, the maximum model domain size is somehow restricted to several hundred kilometers. This domain size is too small to derive large scale dynamics by itself. This fact constrained the investigation of multiscale relations of cirrus formation for a long time.

A good example is the impact of large scale dynamics on the formation of orographic cirrus clouds. Until now, cloud resolving model investigations used the stratified large scale flow evaluated in almost steady state conditions, i.e. after a long time until a quasi-equilibrium has been reached. However, in nature flow conditions often change with time, i.e. wind fields are time-dependent. Several synoptic observations near a mountain have showed that a changing large scale flow influences the time-dependent component of the incident flow, leading to a change in the wave forcing by the mountain and subsequently changing
Chapter 1. Introduction

the conditions under which clouds are formed (see e.g. Smith, 1982; Smith and Broad, 2003; Egger and Kühnel, 2010). Beside the measurements, some rare model investigations of transient flows over mountains have been performed in the last years, investigating basic wave mechanics (Lott and Teitelbaum, 1993) up to the interaction of linear and highly non-linear cases with large scales (Chen et al., 2006). Most of these investigations focus on changes in the flow on long time scales. However, none of the model investigations considered the impact of a changing large scale flow on cold ice microphysics yet. Changing environmental conditions are important for the amount of cirrus clouds formed by orographic waves over the continents (Dean et al., 2005). Regarding the contribution of cirrus clouds on Earth’s radiative budget, their formation mechanism should be treated as accurately as possible. In current GCMs, orographic waves are handled as stationary waves, using linear theory to describe their temperature perturbations (Dean et al., 2007) or vertical velocities (Joos et al., 2008). Transient wave phenomena like horizontal displacement of wave parcels due to a changing large scale flow (see Lott and Teitelbaum, 1993) are not considered in GCMs. In other words, even if a large scale model introduces its changing large scale dynamics into the parameterization of a mesoscale/small-scale phenomenon, the parameterization is not able to handle the transient information and to calculate an accurate response. In order to improve these parameterizations, investigations using cloud resolving models must be done. For this purpose, the non-hydrostatic, anelastic model EULAG (Prusa et al., 2008), extended by time-dependent environmental or ambient states (see section 3.2.1) is applied in this thesis. This improved model opens new possibilities for cloud resolving modeling, by including the changing large scale dynamics.

The main objective of this thesis is to gain new insights on the formation and life cycle of tropospheric cirrus clouds allowing for a broad range of atmospheric motion scales. Especially, the influence of the competing large-scale and mesoscale dynamical processes on cloud formation and evolution will be investigated by a new approach of multiscale numerical modelling: Cloud-resolving numerical simulations are driven by slowly varying environmental or ambient states representing the larger scale forcing.

1.3 Dissertation overview

In this thesis, the impact of processes on different scales on cirrus formation and evolution is determined, using the cloud resolving model EULAG. The focus is on two different large scale effects, which are leading to cloud formation in a distinct situation.

- Radiation: In chapter 2, the impact of radiation (i.e. in situ radiative cooling/warming) around
ISSRs on the stability of the upper troposphere is investigated. In the cirrus community it is usually assumed that radiation is not important for the initial formation of cirrus clouds. The main argument is that radiative cooling would result into very low cooling rates or equivalently in very low vertical updrafts on the order of millimetres per second. However destabilization due to radiation (i.e. a large scale process) could trigger small scale convection and lead to sufficiently large vertical motions for cirrus formation. This should revise the former point of view. This topic is discussed extensively in this chapter.

- Orographic waves: In the next chapter, basic mechanics of transient orographic waves are discussed. For this purpose, a time dependent large scale flow generates waves over an idealized mountain and subsequently leads to cirrus formation. The manifold transient scenarios are compared with non-transient cases (i.e. with a constant flow over a mountain) in order to understand and extract the impact of transient flows on orographic cirrus formation.

In the following chapter 4, the achieved knowledge from chapter 3 is used to describe and understand transient orographic cirrus formation in a much more realistic framework. The model is set up with realistic terrain data and reanalysis data for wind, temperature and pressure (source: ECMWF) and the results are compared with aircraft measurement data from the INCA campaign (Gayet et al., 2004). As in the chapter before, transient and non-transient simulations are compared. Whereas the idealized simulations in chapter 3 provided an overview over basic wave mechanics and their influence on cirrus clouds, chapter 4 is more dedicated to the question, what the benefits are from using transient simulations, which regard the multiscale aspect of cirrus formation more precisely.

Chapter 5 presents a summary of all results, followed by a short outlook on possible future work that could benefit from the current work and augment the current conclusions.
Chapter 2

Cirrus Clouds triggered by Radiation, a Multiscale Phenomenon

In this study, the influence of radiative cooling and small eddies on cirrus formation is investigated. For this purpose the non-hydrostatic, anelastic model EULAG is used with a recently developed and validated ice microphysics scheme (Spichtinger and Gierens, 2009a). Additionally, we implemented a fast radiative transfer code (Fu et al., 1998). Using idealized profiles with high ice supersaturations up to 144% and weakly stable stratifications with Brunt-Vaisala frequencies down to 0.0018 s\(^{-1}\) within a supersaturated layer, the influence of radiation on the formation of cirrus clouds is remarkable. Due to the radiative cooling at the top of the ice supersaturated layer with cooling rates down to −3.5 K/d, the stability inside the ice supersaturated layer decreases with time. During destabilization, small eddies induced by Gaussian temperature fluctuations start to grow and trigger first nucleation. These first nucleation events then induce the growth of convective cells due to the radiative destabilization. The effects of increasing the local relative humidity by cooling due to radiation and adiabatic lifting lead to the formation of a cirrus cloud with IWC up to 33 mg/m\(^3\) and mean optical depths up to 0.36. In a more stable environment, radiative cooling is not strong enough to destabilize the supersaturated layer within 8 hours; no nucleation occurs in this case.

Overall triggering of cirrus clouds via radiation works only if the supersaturated layer is destabilized by radiative cooling such that small eddies can grow in amplitude and finally initialize ice nucleation. Both processes on different scales, small-scale eddies and large-scale radiative cooling are necessary.

Chapter 2. Cirrus Clouds triggered by Radiation, a Multiscale Phenomenon

2.1 Introduction

The existence of cloud-free air masses that are supersaturated with respect to ice in the upper troposphere or lowermost stratosphere has long been known. More than 60 years ago, Glückauf (1945) obtained values of relative humidity with respect to ice (RHi) up to 160% over Southern England. The fact that high ice supersaturations can occur in the upper troposphere has been neglected for many years and has often been termed erroneous. During the last two decades the existence of supersaturated airmasses has been proven by many measurements using a variety of different measurement techniques (e.g. Jensen et al., 1998; Vay et al., 2000; Ovarlez et al., 2000; Krämer et al., 2009). These measurements are consistent with theoretical considerations, that ice crystals form at very high supersaturations, where the exact freezing threshold depends on the nucleation mechanism (homogeneous freezing of solution droplets or heterogeneous nucleation, see e.g. Koop et al., 2000; De Mott et al., 2003, respectively). For homogeneous nucleation, which is probably the dominant mechanism for ice crystal formation at low temperatures ($T < -38^\circ$C) relative humidities in the range 140 to 170% RHi, depending on temperature (Koop et al., 2000), are required.

The properties and global distributions of ice supersaturated regions (ISSRs) have been determined over recent years (e.g. Spichtinger et al., 2003a,b; Gettelman et al., 2006). They occur 20 to 30% of the time in cloud free air masses in the upper troposphere over the North Atlantic. The large horizontal extent of ISSRs, with mean path lengths of about 150km, and in some rare cases of a few thousand kilometers (Gierens and Spichtinger, 2000) means that there is a substantial amount of cloud free air masses with enhanced water vapor content. However, only a few parametrizations exist that correctly (i.e. physically) describe the formation of cirrus clouds (from ISSRs) driven by synoptic scale dynamics (CAM3, see Lin et al, 2007; ECHAM5, see Kärcher and Lohmann, 2002). The impact of meso- and small-scale motions is not explicitly included, although some approximations are used to obtain meso-scale motions (Kärcher and Lohmann, 2002).

The radiative impact of ISSRs (with RHi up to 130%) has been investigated by Fusina et al. (2007). These regions of enhanced water vapor can reduce the total outgoing longwave radiation by more than 1W/m$^2$ and imply a significant cooling at their upper boundary (i.e. due to thermal emission). In a stably stratified atmosphere, these diabatic heating rates are too small to significantly influence the local dynamics. But if potential stability of the environment is weak or neutral, radiative cooling can decrease the Brunt-Vaisala frequency to a critical value. This finally leads to a change (increase) of the extent and amplitude of preexisting small scale motions. The combination of the locally amplified vertical velocity...
and the radiative cooling can trigger the first nucleation. Regions with weak stratification (i.e. small vertical gradients of potential temperature) have been observed in the upper troposphere, using radiosonde data obtained over the meteorological observatory in Lindenberg, Germany. In these regions dynamical instabilities could also occur: The Miles theorem describes that a stable atmosphere is given by a Richardson number $Ri > 0.25$,

$$Ri = \frac{N^2}{\left(\frac{du}{dz}\right)^2}, \text{using } N^2 = \frac{g}{\theta} \frac{\partial \theta}{\partial z}$$

(2.1)

where $N$ denotes the Brunt-Vaisala frequency. Moderate and strong windshears are often observed in the upper troposphere, as described by Birner (2006).

In this study, we investigate the possibility of cirrus cloud formation due to radiative cooling in weakly stable layers within the upper troposphere. Sensitivity studies are carried out for the most important initial parameters such as potential stability, windshear and RHi within the ISSR. The main purpose is to investigate the sensitivity to environmental conditions, for which radiative cooling (a large scale process) at the top of an ISSR can destabilize the stratification. During the destabilization, the amplitude of preexisting small scale eddies increase and trigger cloud formation. Only the superposition of effects on different scales (i.e. large-scale radiative cooling and small scale motions) can finally lead to cirrus formation. The main properties of the formed cirrus clouds are investigated, such as ice water content (IWC), ice crystal number density (ICNC), cloud lifetime and their impact on the radiation. For this purpose, the non-hydrostatic anelastic model EuLag (Eulerian, semi-Lagrangian Model) is used.

![Figure 2.1: Heating-rates [K/d] of an ISSR with RHi = 130% due to emission and absorption of radiation.](image)
Chapter 2. Cirrus Clouds triggered by Radiation, a Multiscale Phenomenon

(e.g. Prusa et al., 2008), using a two stream radiative transfer code (Fu, 1996; Fu et al., 1998) and an ice microphysics scheme (Spichtinger and Gierens, 2009a).

The paper is organized as follows: In the next chapter, the model is described briefly. In chapter 3 we define the experimental setup for the reference simulations. In chapter 4, results of the reference and sensitivity simulations are presented. We end with discussions and conclusions.

2.2 Model Description

As a basic dynamical model, the anelastic non-hydrostatic model EULAG is used (see Prusa et al., 2008). The dry anelastic equations solved in the model are presented in Smolarkiewicz and Margolin (1997).

2.2.1 Ice Physics

A recently developed bulk ice microphysic scheme is used, which can treat an arbitrary number of ice classes. These ice classes are distinguished by their formation mechanism (e.g. heterogeneously vs. homogeneously formed ice). The following processes for cold cirrus clouds are parameterized: nucleation (homogeneous), deposition (growth, evaporation) and sedimentation (Spichtinger and Gierens, 2009a). It provides a consistent double moment scheme with terminal velocities for ice number and mass concentration. Aggregation is not yet implemented in the microphysics scheme. However, aggregation is of less importance for the cold temperature regime ($T < -38^\circ$C) and/or for moderate vertical velocities (Kajikawa and Heymsfield, 1989).

For the parametrization of homogeneous freezing of aqueous solution droplets, sulfuric acid solution droplets are assumed as a background aerosol, using a lognormal distribution for the $H_2SO_4$ droplet size (geometric standard deviation $\sigma_r = 1.4$, geometric mode radius $r_m = 25$ nm, total number concentration $300\text{ cm}^{-3}$). The freezing rates are calculated using a temperature based parametrization, based on water activity (see Koop et al., 2000). We use a modified Koenig ansatz (König, 1971) to parameterize the diffusional growth or evaporation for small ice crystals, including corrections for the kinetic growth regime and ventilation respectively. For all simulations in this study, only homogeneous nucleation is taken into account. For details of the model, the reader is referred to Spichtinger and Gierens (2009a).
2.2.2 Radiation Transfer Model

A two stream radiative transfer code, i.e. a representation of forward and backward streams, has been implemented into the EuLag model. It contains 6 bands in the solar and 12 bands in the thermal infrared regime. For a detailed description, the reader is referred to Fu (1996) for the shortwave and to Fu et al. (1998) for the longwave parametrizations, respectively.

The required parameters for ice microphysics are ice water content (IWC) and effective radius, which is derived following Slingo (1989), Dobbie et al. (1999) and Fu (1996). In the microphysical model ice crystals are assumed to be small hexagonal columns. The ice crystal size is lognormally distributed with a geometric standard deviation of $\sigma_L = 1.5$. The effective radius is calculated as described in Fusina et al. (2007) under the assumption of randomly oriented columns (Ebert and Curry, 1992).

The radiative transfer code uses following constant trace gas concentrations: CO$_2$: 330ppmv; CH$_4$: 1.6ppmv; N$_2$O: 0.28ppmv, respectively (default values of the radiative transfer model). The ozone profile depends on altitude, including the stratospheric ozone layer. The model domain of the radiative transfer code has a maximum altitude of $L_{cr} = 50$km, which is not necessarily equivalent to the top level of the EULAG model ($L_e$). The vertical resolution within the EULAG model domain is set by the model setup. If $L_{cr} > L_e$ (this is the case for our simulations), the vertical resolution of the additional layers exceeding the EULAG model domain is set to 1km, the water vapor mixing ratio is set to $q_v = 10^{-11}$kg/kg, the pressure profile is interpolated using the US standard atmosphere; the temperature profile is built using a constant lapse rate up to a final temperature of $T(50km) = 275$K (according to the US standard atmosphere (COESA, 1976)). The solar zenith angle is set to 60° and the solar surface albedo is 0.3. The infrared surface emissivity is assumed to be 1.

2.3 Experimental Setup

For the simulations, an idealized framework is used, including a 2D model domain with a horizontal extent of $L_x = 12.8$km and a vertical extent of $L_z = 15$km for dynamics and ice microphysics, respectively. An ISSR is placed between $lb = 9500$m and $ub = 10500$m with a constant relative humidity over ice of 140%, which fully occupies the whole horizontal extent of this layer. RHi below and above the ISSR is set to 60% and 5%, respectively (see Fig. 2.2, left panel). Using a grid resolution of $dx = 100$m and $dz = 50$m, in the horizontal and the vertical, respectively, the domain is discretized with $nx \times nz = 128 \times 301$ grid points.

A constant-stratification ambient profile of potential temperature, as described by Clark and Farley
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Figure 2.2: Vertical profiles used for the reference simulation. Left: RHi profile; middle: potential temperature profile; right: horizontal velocity profile.

(1984), is used with a surface temperature $T = 293.15\,\text{K}$ and a Brunt-Vaisala frequency $N = 0.0094\,\text{s}^{-1}$. Within the layer between 9250 m and 10750 m altitude, the stability of the stratification is decreased to a constant vertical gradient of the potential temperature of $\frac{\partial \theta}{\partial z} = 0.4\,\text{K/km}$ (which corresponds to a Brunt-Vaisala frequency $N = 0.0035\,\text{s}^{-1}$). The tropopause is placed at an altitude of 12000 m (see Fig. 2.2 middle). The stability above the tropopause is set to $N = 0.02\,\text{s}^{-1}$. The profile of the horizontal velocity contains a constant wind below and above and a weak windshear of $\frac{\partial u}{\partial z} = 1 \cdot 10^{-3}\,\text{s}^{-1}$ (for the reference case) within the ISSR (Fig. 2.2, right panel):

$$u(z) = \begin{cases} 1 & \text{for } 0 \leq z < lb, \\ 1 + du_c \cdot (z - lb) & \text{for } lb \leq z \leq ub, \\ 1 + du_c \cdot (ub - lb) & \text{for } ub < z. \end{cases} \quad (2.2)$$

The initial potential temperature field is distributed by a Gaussian noise with standard deviation $\sigma_T = 0.01\,\text{K}$. All simulations run for a total time of 8 h with 1 s increments for the dynamics, optional 0.1 s increments for the ice physics in case of homogeneous nucleation and 10 s increments for the radiation, respectively. We assume a calm troposphere, i.e. no vertical wind in our small 2D domain.

2.4 Results

2.4.1 Reference Case

At the beginning of the reference case simulation, only radiative processes (emission and absorption of LW radiation) can change dynamical and microphysical properties in the domain. The obtained heating rates (see Fig. 2.3) have an impact on two major characteristics within the ISSR. First, the temperature
2.4. Results

Figure 2.3: Vertical radiative heating rates (split into solar, infrared and total) at the initial time step, containing an ISSR with maximum RHi of 140%.

decrease (strongest at the top) tends to increase RHi by $\sim +1\%$ per hour (see Fig. 2.4 d). Second, the stability of the stratification decreases with time, reaching its lowest neutral/unstable state at $t = 200\text{min}$ (see Fig. 2.4 a), where the first unstable grid cells (i.e. with a negative squared Brunt-Vaisala frequency) occur at $t = 150\text{min}$. Within this first 200min, small eddies (induced by initial temperature fluctuations) grow due to the destabilization and start to lift up air parcels, temporally increasing their RHi. Patterns within the vertically averaged turbulent kinetic energy enhance the formation and growth of these small cells. However, the vertical velocities do not increase significantly before the stratification becomes unstable, but the amplitude of the temperature deviations increases slightly.

After $t = 200\text{min}$, the eddies start to grow rapidly in size (favored by the unstable environment) and therefore, the first air parcels reach the threshold of homogeneous nucleation of 156% (see Fig. 2.4 d). This effect happens only in several small isolated cells, surrounded by downwelling regions. Latent heat release due to the growth of the ice crystals amplifies the motions inside the ISSR, increasing the vertical velocity from its preliminary $w_{\text{max}} = \pm 0.03\text{m/s}$ up to $w_{\text{max}} = \pm 0.5\text{m/s}$, forming convective cells. The formed cirrus cloud is persistent over the rest of the simulation time, containing ice crystal number concentrations up to $\text{ICNC} = 300\text{L}^{-1}$ and a maximum mean ice water path of $\text{IWP} = 2.95 \cdot 10^{-3}\text{kg/m}^2$ (averaged over all columns). The peak with the highest ice water content corresponds to a local maximum of the vertical velocity. The whole time evolution for $\text{IWP}$ is showed in Fig. 2.4c.

At the time of maximum $\text{IWP}$, an optical depth of $\tau = 0.1$ is derived. $\tau$ is averaged over all columns, observing a peak-value of $\tau = 0.36$, which is significantly higher than the mean value due to the patchiness of the cloud. Due to the high supersaturation inside the layer, the formed ice crystals grow rapidly to larger sizes and start to sediment. During the downward motion, they continue to grow and deplete
water vapor, reducing the RHi in the lower part of the ISSR. In the downwelling regions, some entrainment of very dry air from above the ISSR can be observed, making the 2D field of RHi more patchy than before. Within the upper part of the ISSR, at the top of the upwelling regions, the RHi remains at the homogeneous freezing threshold for a longer time, continuously forming new ice crystals (Fig. 2.5, \( t = 360\) min).

### 2.4.2 Variation of potential stability

For sensitivity tests, the initial vertical gradient of the potential temperature \( \theta \) is changed within the ISSR. For a small wind shear, as used in the reference case and in this set of sensitivity simulations, this value represents the stability because of high Richardson numbers. Cases with stronger wind shear will be discussed in the next section. A set of value is chosen as: \( \frac{\partial \theta}{\partial z} = 0.1/0.2/0.3/0.4/0.5/0.6/0.8 \) K/km (corresponding to Brunt-Vaisala frequencies \( N = 0.0018/0.0025/0.003/0.0035/0.0039/0.0043/0.005 \) s\(^{-1}\) respectively). These values have been compared with radiosonde records obtained from routine measurements over the meteorological observatory Lindenberg, Germany (see e.g. Spichtinger et al., 2003a). The dataset covers the time period from 01/02/2000 to 30/04/2001 with measurements every 6 hours.
2.4. Results

Figure 2.5: Cirrus cloud formation in the reference simulation - Colors indicate RHi in %; Contour lines represent IWC $[10^{-6} \text{kg/m}^3]$ increments. Time evolution: 1. Destabilization at ISSR top, first nucleation. 2. Development of the cloud/convective cells. 3. Entrainment.

(i.e. 00, 06, 12, 18 UTC). 14.7% of all profiles show a layer of at least 500m thickness, containing a potential stability between 0.1 and 0.8 K/km (5% for stabilities lower than 0.4 K/km; reference case). 3.26% of all profiles contain the same stability, but within a layer of at least 1km thickness (0.64% for stabilities lower than 0.4 K/km). These cases are filtered for temperatures lower than 235 K for comparison in potential cirrus regions. It follows that the values used in this sensitivity study are in the range of atmospheric relevance.

All cases in this sensitivity simulation fulfill the Miles theorem for shear stability (i.e. $Ri > 0.25$). All other simulation parameters are equal to the reference case.

When we reduce the vertical gradient of the potential temperature of the reference case (0.4 K/km) to $\partial \theta/\partial z = 0.2$ K/km, the maximum obtained amount of IWP increases from $IWP = 3 \cdot 10^{-3}$ kg/m$^2$ to $IWP = 6.8 \cdot 10^{-3}$ kg/m$^2$. This peak value appears 100 min earlier than in the reference case and corresponds to the moment of maximum kinetic energy within the ISSR (i.e. the vertical wind speeds...
are strongest; see Fig. 2.6 b). One can conclude that in simulations with a lower stability inside the ISSR, nucleation occurs earlier than in the reference case due to two reasons. First, the amplitude of the temperature variation inside the small eddies is larger due to the lower stability. Second, and more important, by the destabilization of the layer due to radiation is faster, starting at lower Brunt-Vaisala frequencies. The obtained IWP is higher for cases with weaker initial stabilities (Fig. 2.6 a)). The largest IWP = 8.9 \cdot 10^{-3} \text{kg/m}^2 (with a maximum value of IWC = 33 \text{mg/m}^3) can be observed for the simulation with the weakest initial potential stability, 170 min after initialization.

The main reason for the higher IWP are the higher vertical velocities obtained in the simulations with weaker initial stability, resulting in higher ice crystal amounts (Fig. 2.7). In the simulation with the weakest stability (0.1 K/km), vertical updrafts of up to w = 1.6 m/s are observed in some isolated cells. If the initial stability is reduced further, then convective cells can be generated due to the initial temperature perturbations and the role of the radiation in the cloud building process is no longer dominant.
2.4. Results

Figure 2.7: a) Histogram of vertical velocity for all time steps (for 3 different thermal stratifications \( \frac{\partial \theta}{\partial z} = 0.1/0.4/0.8 \text{ K/km} \)). The red line represents the reference case; b) Histogram of number concentration for all time steps (for 3 different thermal stratifications \( \frac{\partial \theta}{\partial z} = 0.1/0.4/0.8 \text{ K/km} \)). The red line represents the reference case.

(i.e. the superposition of these two effects, radiation and small eddies, is no longer required). For higher initial stabilities the destabilization event takes more time. For all simulations with a stronger stability than the reference case, the simulation time of 480 min is not sufficient to deplete the whole supersaturation inside the layer. In these cases, layer-clouds are formed in the upper part of the ISSR. If the initial stability exceeds the threshold of 0.8 K/km, no nucleation occurs within the 8 h simulation time.

2.4.3 Variation of wind shear

In a second set of sensitivity simulations, the impact of different strengths of wind shear is determined while the thermal stratification is set to the reference value of \( \frac{\partial \theta}{\partial z} = 0.4 \text{ K/km} \). We use values of \( \frac{\partial u}{\partial z} = 0/1/2/4/6/8 \cdot 10^{-3} \text{ s}^{-1} \) in the altitude range \( 9500 \leq z \leq 10500 \text{ m} \) (see Fig. 2.8 a). The chosen values are weak, but in the range of observations in the upper troposphere (see, e.g., the statistics pre-
sent by Birner, 2006). For the strongest wind shear $\partial u / \partial z = 8 \cdot 10^{-3} \text{s}^{-1}$, the initial Richardson Number $Ri = 0.19 < 0.25$ is below the threshold for shear instability. In this case, we expect the formation of Kelvin-Helmholtz instabilities. Values of wind shear above this threshold are not of interest for this study, because the mixing effect of shear instability would superimpose with the destabilization effect of radiative cooling in an unpredictable way. All other simulation parameters are equal to the reference case.

For this set of sensitivity simulations, the highest $\text{IW}P = 4.15 \cdot 10^{-3} \text{kg/m}^2$ was observed for zero wind shear. In this scenario, the formation and evolution of the isolated convective cells is not affected by wind shear, so the cells have a larger vertical extent than in the cases with windshear. The statistics in Birner (2006) show that for the mid-latitudes, wind shear free scenarios are very unlikely at cirrus altitudes. Increasing the wind shear to $\partial u / \partial z = 2 \cdot 10^{-3} \text{s}^{-1}$ reduces the vertical extent of the upwelling regions and the vertical updraft velocity inside. The cells start to tilt with height. Due to the increased mixing inside the ISSR, fewer regions with RH close to the homogeneous threshold are observed. This leads to reduced nucleation, whereby the number of ice crystals $\text{ICNC}$ decreases. Increasing the wind shear more will amplify these further effects, leading to lower IWPs (see Fig. 2.8 b). For windshears of $\partial u / \partial z \geq 4 \cdot 10^{-3} \text{s}^{-1}$, shear instability is generated with time. This leads to more turbulent motions inside the ISSR and results in stronger vertical velocities followed by an increased homogeneous nucleation rate. For the case with the highest windshear of $\partial u / \partial z \geq 8 \cdot 10^{-3} \text{s}^{-1}$, a pronounced Kelvin-Helmholtz instability develops after $t = 440 \text{min}$, containing vertical updraft velocities of $w > 2 \text{m/s}$. The cirrus clouds formed by turbulence due to shear instabilities appear even, when the radiation code is disabled. For this reason, they are not of interest for this study.

It can be concluded that, after the destabilization of the stratification due to radiative cooling, isolated eddies are a key feature of cirrus clouds triggered by radiation, supported by the latent heat release after the first nucleation event. Their evolution and persistence is suppressed by stronger wind shears. Wind shear can also block the destabilization of the layer with the highest radiative cooling rate, by mixing it with the surrounding air. However, this happens only for values close to the threshold for shear instability ($Ri \leq 0.25$).

### 2.4.4 Variation of RH

A third set of sensitivity simulations uses different amounts of water vapor inside the ISSR. By increasing the RH, the distance to the homogeneous freezing threshold is decreased, so earlier nucleation can be
expected. Decreasing the RHi leads to the opposite effect. The values used for RHi are 132 / 136 / 140 / 144 %, respectively. Wind shear and thermal stability are set to the values in the reference case ($\partial u/\partial z = 10^{-3} \text{s}^{-1}, N = 0.0035 \text{s}^{-1}$).

For simulations with a maximum RHi lower than in the reference case (i.e. $< 140\%$), it takes more time to reach the homogeneous freezing threshold (due to cooling from emission and adiabatic lifting). On the other hand, the radiative cooling inside the ISSR is a function of the RHi (i.e. the water vapor mixing ratio) as discussed in Fusina et al. (2007). Thus, for lower RHi values at the same temperature, the cooling rates decrease. This results in a slower destabilization of the upper part of the ISSR. The differences between the updraft velocities are small, no significant change in the numbers of ice crystals can be observed between the different simulations. It follows that the enhanced water-vapor content increases the size of the ice crystals but not their number concentration. This can modulate the cirrus life time, as larger crystals fall faster down to subsaturated regions and sublime.
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Figure 2.9: Mean ice water path $IWP$ (averaged over all columns) of simulations using different initial values for RHi inside the ISSR (RHi = 132 to 144%). The black line represents the reference case.

For the simulation with $RHi = 144\%$, the nucleation starts 15 min earlier than in the reference case, obtaining a maximum $IWP$ that is 30% higher. A comparison of the $IWP$ between the different sensitivity simulations is given in Fig. 2.9.

In the simulations with $RHi = 136 / 140 / 144\%$, a second nucleation peak at the end of the simulation can be observed. This nucleation event is possibly triggered by the latent heat release during the earlier ice crystal forming, superimposed with destabilization due to radiative cooling. If the initial RHi is reduced to less than 132%, the cooling will again get weaker and no destabilization occurs (i.e. formation of an updraft) and for this reason, no nucleation can be observed during the simulation time of 8h. If the initial RHi is increased to 150% or more, the small scale motions due to the initial temperature fluctuations are strong enough to trigger nucleation without any prior change in the local stability due to radiative cooling (not shown).

2.4.5 Variation of RHi Gradient at ISSR boundary

Here we discuss the influence of the water vapour gradient at the top of the ISSR on local heating rates. Radiosonde profiles show a variety of different shapes of the RHi profile in the upper troposphere. In Fig. 2.10, two different radiosonde profiles with different water vapor gradients are shown. We use the same setup as in the reference case simulation (maximum RHi = 140%) and increase the vertical distance in which the RHi decreases from its maximum value down to RHi = 5% near the tropopause, using following values: $dz = 250/500/750/1000/1500m$, where $dz = 250m$ stands for the reference case (see Figure 2.11 a).
2.4. Results

Figure 2.10: Profile 1: steep upper RH\textit{i} gradient (Lindenberg 01/07/00 06UTC); Profile 2: weak upper RH\textit{i} gradient (Lindenberg 04/03/01 06UTC)

In Figure 2.11 b, the heating rates for the different RH\textit{i} gradients can be seen. It is clear, that for weaker gradients, the cooling area is distributed over a larger vertical layer than for stronger gradients. Therefore, the maximum cooling peak is lower, resulting in a weaker destabilization of the upper layer. Thus, only the profiles with the strongest two initial RH\textit{i}-gradients were able to trigger the described cirrus formation mechanism within the simulation time of 8h (assuming all other parameters to be equal to the reference case). This might be a constraint upon this kind of cloud formation process. A sharp decrease of RH\textit{i} at the upper boundary of the ISSR is strictly required for destabilization due to radiative cooling. One example for a sufficient strong RH\textit{i} gradient is shown in Fig. 2.10, Profile 1.

2.4.6 Variation of temperature

Another way to change the water vapor content within the ISSR is to vary the vertical temperature profile. Using constant initial values for RH\textit{i} in the ISSR implies that for an increasing temperature, the water vapor mixing ratio will increase too. This influences the local heating rates and, due to radiation, also the destabilization of the stratification. The destabilization due to radiation is faster for profiles with higher environmental temperatures and the formed cirrus has a larger IWC. However, the general sensitivity of higher H$_2$O content in the upper troposphere is shown in section 2.4.4.

2.4.7 Discussion of radiative impacts

The variation of the potential stability has the biggest influence on the radiative properties of the formed cirrus cloud. Increasing the initial $\partial \theta / \partial z$ from 0.1 to 0.8K/km (i.e. $N = 0.0018$ to 0.005s$^{-1}$), the
maximum value of the mean optical depth (averaged for all columns for every time step) decreases from $\tau = 0.36$ to 0.05 (see Fig. 2.12 a - c). Note that this is a mean value over the whole domain and the values for single columns can be considerably larger (the largest optical depth for a single column can be observed in case of the smallest potential stability: $\tau_{\text{max}} = 0.82$). The smallest changes of $\tau$ can be observed for the sensitivity study of changing initial RHi ($\tau = 0.14$ to 0.08). For all cases, the calculated total outgoing radiation fluxes (the sum of shortwave and longwave radiation) decrease by a certain amount after the cloud formation, i.e. the difference between the absorbed and emitted longwave radiation inside the cloud exceeds the amount of reflected shortwave radiation at the cloud top (warming scenario, see Fusina et al. (2007)).

**Figure 2.11:** a) Vertical RHi profiles of ISSRs using different upper gradients; b) Radiative heating rates for 5 different RHi gradients (initial conditions, see a)).
Figure 2.12: Mean optical depth (as a function of time) of cirrus clouds, triggered by radiative cooling for following sensitivity studies: a) thermal stratification; b) dynamic stability (variation of vertical wind shear); c) RHi, using different peak values inside the ISSR.

2.5 Conclusions

The nonhydrostatic, anelastic model EULAG (Prusa et al., 2008) was used with a recently developed and validated ice microphysics scheme (Spichtinger and Gierens, 2009a) and an additionally implemented fast radiative transfer code (Fu et al., 1998) to investigate the influence of a superposition of radiative cooling and small eddies on cirrus formation. For this purpose, idealized profiles with high supersaturations up to 144% RHi and weak thermal stability have been used. The focus is on the multiscale aspect of cirrus formation superimposing large scale (i.e. radiative cooling) with small scale (i.e. small eddies) effects. Only the combination of these effects results in significant cloud formation.

The results can be summarized as follows: destabilization due to radiative cooling can lead to amplification of small scale eddies, which can act as an initial cloud forming mechanism. Sensitivity studies for following parameters have been performed: static and dynamic stability and the RHi within the ISSR. The values of these key factors should be between 0.1 and 0.8 K/km for the thermal stability (i.e. the vertical gradient of the potential temperature), 132 to 144% RHi and there should be no shear instability (Ri > 0.25).
Our investigations could answer some questions concerning the formation of cirrus clouds due to radiative cooling:

[1] Cooling due to thermal emission at the top of an ISSR can destabilize an initial weak stable profile within several hours (depending on the initial stability and RHi).

[2] During destabilization, the amplitude of initial small eddies increases and leads to the first nucleation of ice crystals. Supported by the subsequent latent heat release, vertical updraft velocities up to $1.6 \text{ m/s}$ can occur.

[3] Within the first 8 hours, cirrus clouds are formed with mean ice water paths up to $8.9 \cdot 10^{-3} \text{ kg/m}^2$ and ice crystal number densities up to $\text{ICNC} = 350 \text{ L}^{-1}$.

[4] In sensitivity studies it was shown that increasing the initial potential stability would delay the first nucleation event up to several hours and decrease the strength of the nucleation event, as we can see for $\text{IWP}$ and ICNC. Increasing the initial wind shear would lead to smaller cells, and therefore to nucleation of fewer crystals. Increasing the RHi within the ISSR amplifies the thermal emission and shortens the duration to the first nucleation event, but it has only a marginal effect on the vertical velocities. This will lead to cirrus clouds with larger $\text{IWP}$.

[5] To obtain sufficient radiative cooling at the top of the ISSR, a sharp decrease of RHi is required.

Using a simple phase-diagram (Fig. 2.13) we can explain the sensitivity due to certain parameters. If the initial RHi is lower than the boundary (a), the radiative cooling would be too weak to destabilize the stratification. If it is too high (boundary (c)), the random motions due to initial temperature fluctuations would be strong enough to trigger a cloud before the profile becomes unstable. If the stability (static or dynamic) becomes too weak (boundary (b)), updrafts can be triggered spontaneously due to initial small-scale motions or induced shear-instability. If it becomes too strong (boundary (d)), the radiative cooling is again not strong enough to destabilize the profile within a certain time (8h for this simulations). At least for conditions near the boundary (b), it is very hard to distinguish which of the observed effects (e.g. destabilization due to radiative cooling, shear-instability, small-scale eddies) actually is the most important. It has to be considered that we always have a superposition of different cloud-controlling effects on different scales. It must be noted here that without radiative cooling at the top of the ISSR, there would be no cloud formation for all simulations within 8 hours. This implies that other cloud building mechanisms like frontal lifting or orographic effects (i.e. gravity waves) (e.g. Spichtinger and Gierens, 2009b; Joos et al., 2009) should not occur during the simulation time.

As a kind of textbook knowledge in the cirrus community it is usually assumed that radiation might not
2.5. Conclusions

Figure 2.13: Phase-diagram of possible conditions, for which thermal emission is an important cloud-building factor in dependence of the stability (statical and dynamical) and the RHi within an ISSR. The regions outside b) and c) accord to spontaneous cloud formation without an influence of radiation, whereas outside a) and d) the conditions supress any cloud formation within the simulation time.

be important for the initial formation of cirrus clouds. The main argument is that radiative cooling would result into very low cooling rates or equivalently into vertical updraughts of the order of millimetres per second. This would lead to very thin cirrus clouds containing only a few ice crystals per litre (see e.g. Kärcher and Spichtinger, 2009).

However, from the results of the present study this position might be revised under the comprehension of the discussed destabilisation process. In presence of weakly stable profiles both effects, i.e. radiative cooling and destabilisation might lead to the formation of visible cirrus clouds. On the other hand, the impact of radiation on the stability of the upper troposphere itself, discussed in a broader sense, should be an interesting topic for future research. From this point of view the role of convective cells in ice-supersaturated regions and cirrus cloud layers might be interesting in terms of cirrus cloud inhomogeneities and patchiness of cirrus clouds, also in terms of the radiative impact of cirrus clouds. Finally, this could also be important for more physically based parameterisations of cirrus clouds in large-scale models, including also the macroscopic structure on the cloud scale.

We could not discuss in detail the issue of frequency of occurrence of environmental conditions, which allow the radiation to have a predominant impact on the stability of the upper troposphere (and therefore be of importance for cirrus formation). The radiosonde data used originates from only one measurement site and therefore does not give an insight into global distributions. This must be investigated in future studies, using meteorological analyses and maybe also output from large-scale models in order to obtain a better overview about the importance of the described mechanism. It also remains unclear how important radiative destabilisation is, when superimposed with other large- or meso-scale processes (i.e. frontal lifting, gravity waves etc.); this will be subject of future research.
Acknowledgements

We thank the European Centre for Medium-Range Weather Forecasts for computing time (special project SPCHCLAI "Cloud aerosol interactions") and Andreas Doernbrack, Ulrike Lohmann and Thomas Peter for fruitful discussions. We are grateful to Qiang Fu for sharing his radiation transfer scheme and Thierry Corti for help in using it. We also thank Declan O’Donnell for his help in proof-reading. This work contributes to the project "Impact of dynamics on cirrus clouds" (Grant: 200021-117700) supported by the Swiss National Science Foundation (SNSF).
Chapter 3

Cirrus clouds formed by a time dependent flow over a mountain

In this study, the influence of time dependent large scale flows on cirrus formation over mountains is investigated using the non-hydrostatic, anelastic model EULAG, including a bulk ice microphysics scheme. Transient environmental or ambient fields for potential temperature $\theta_e$ and horizontal wind $u_e$ were implemented in EULAG in order to represent time dependent large scale dynamical variations which drive the processes on small scales. These small-scale processes influencing cirrus formation are resolved in our model. Thus, this new method attempts to simulate the multiscale aspect of cloud formation more accurate. For this purpose, an idealized stratified flow over a single mountain ridge in the center of the model domain was investigated. As a reference, non-transient simulations using three constant horizontal flow profiles (i.e. low, medium and high-wind) were performed. The transient simulations contain an increasing wind (i.e. medium to high wind) and a decreasing wind (medium to low wind) case, changing from one flow regime to the other after $t = 180\text{min}$ within the transition time $\tau_t = 60\text{min}$. In all simulations, the generated orographic wave propagates up to an ice supersaturated region (ISSR) placed between 8000m and 11000m altitude containing high supersaturations up to 125%. The adiabatic cooling in the ascending branches of the wave leads to the formation of orographic cirrus clouds with ice water contents of up to 140mg/m$^3$. Whereas the waves generated by a non-transient flow are stationary, the waves generated in a transient regime show significant horizontal and vertical displacements of the updraft/cloud regions due to an imbalance between the conserved horizontal phase velocity and the changing environmental flow. This results in a different vertical velocity regime inside the ISSR, namely horizontal displacement of updrafts and downdraft and also a change from an updraft dominated to a downdraft dominated regime (and vice versa) occurs due to a changing vertical wavelength. The impact on cirrus cloud formation is as follows: The transient simulations show a larger spatial distribution of cloud ice due to the horizontal displacement of updraft regions, whereas the non-transient simulations are dominated by stationary clouds directly over the mountain ridge.

†Fusina, F., P. Spichtinger and A. Dörnbrack, 2010. Cirrus clouds formed by a time dependent flow over a mountain. J. Atmos. Sci., to be submitted.
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Non-transient simulations of orographic cirrus clouds can give a good first estimate, but are not able to represent the impact of time dependent large scale dynamics. The use of transient environmental states provides new possibilities to look deeper inside the formation and evolution of orographic cirrus clouds under a more realistic multiscale aspect.
3.1 Introduction

Cirrus clouds as all other clouds are important modulators of Earth’s radiative budget. For low (liquid) clouds, the albedo effect (i.e. cooling) is much larger than the greenhouse effect (i.e. warming), but for cirrus clouds, both effects are of comparable size (with different signs). The transition between a net warming or cooling cirrus depends on microphysical properties as shape or size of ice crystals (see e.g. Zhang et al., 1999; Fusina et al., 2007). The global net effect of cirrus clouds tends to warm Earth’s atmosphere (Chen et al., 2000), but this is not confirmed up to now. Ice crystal formation at low temperatures \( T < 235 \text{K} \) depends crucially on local dynamics (e.g. Kärcher and Lohmann, 2002; Spichtinger and Gierens, 2009a). Thus, for a better estimate of the radiative impact of cirrus clouds we have to investigate the interaction of (local) dynamics and ice processes in more detail. Orographic waves are a possible source for updrafts and are able to form cirrus clouds (Joos et al., 2009). Cloud resolving model investigations of orographic cirrus clouds usually used the stratified flow evaluated in almost steady state conditions, i.e. after a long time until a quasi-equilibrium was reached. However, atmospheric flow conditions often change with time, i.e. wind fields are generally time-dependent. Time dependent large scale flows can influence the characteristics of orographic waves, such as amplitude, phase and generated wind or pressure. Several in-situ observations near mountains have shown, that a changing large scale flow influences the magnitude and direction of the incident flow and leads to a change in the wave forcing by the mountain (see Smith, 1982; Smith and Broad, 2003; Egger and Kühnel, 2010). There were some model investigations of transient flows over mountains in the last years, investigating basic wave mechanics (Lott and Teitelbaum, 1993) up to the interaction of linear and highly non-linear cases with large scales (Chen et al., 2006). But none of these investigations considered the impact of transient orographic waves on cold ice microphysics yet, but focussed on changes in the flow on long time scales.

In current GCMs, orographic waves are considered to be stationary waves, using linear theory to describe their temperature perturbations or vertical updrafts (Dean et al., 2007; Joos et al., 2008, 2010). Transient wave phenomena like horizontal displacement of wave parcels due to a changing large scale flow (see Lott and Teitelbaum, 1993) are not considered in GCMs. Regarding the contribution of cirrus clouds on Earth’s radiative budget and the substantial amount of cirrus clouds formed by orographic waves over the continents (Dean et al., 2005), their formation mechanism should be treated as accurate as possible. In this study, we investigate the influence of vertically propagating mountain waves rather than the effect of wave breaking on cirrus cloud formation. Therefore, only scenarios with a characteristic vertical
wavenumber $N/u_0$ that is large compared to the mountain height $h_0$, i.e. with small inverse Froude number, are considered:

$$Fr^{-1} = \frac{Nh_0}{u_0} \ll 1$$

(3.1)

Here, $N$ denotes the Brunt-Vaisala frequency and $u_0$ the velocity of the environmental flow. For scenarios with $Fr^{-1} \geq 1$, windward blocking and low-wave breaking would occur, leading to a highly non-linear solution (Schär and Durran, 1997; Rotunno et al., 1999). In the upper troposphere, we assume a uniform wind speed to avoid the interaction of waves with a critical level (e.g. Dörnbrack and Nappo, 1996). Based on linear theory (Dörnbrack and Nappo, 1997), the vertical wavelength $\lambda_z$ and the group velocity $C_{gz}$ can be expressed by:

$$\lambda_z = \frac{2\pi u_0}{Nh_0} ; C_{gz} \approx \frac{k u_0^2}{N}$$

(3.2)

where $k = 2\pi/\lambda_h$ denotes the horizontal wavenumber.

The main goal of this study is to investigate the spatial and temporal response of propagating mountain waves due to a transient ambient flow. Furthermore, the impact of these wave packets on cirrus cloud formation will be studied. To estimate the influence on cirrus formation and evolution, the change of the updraft regime at cold temperature regions ($T < 235$ K) is an important issue. Therefore, detailed statistics of the vertical velocity regime in these regions are investigated. For this purpose, the non-hydrostatic anelastic model EULAG (Eulerian, semi-Lagrangian Model) is applied (Prusa et al., 2008), extended by time-dependent ambient states and by an up-to-date ice microphysics scheme (Spichtinger and Gierens, 2009a). The dynamical core of EULAG participated in an intercomparison of 11 different models to simulate quasi-linear and strongly non-linear wave regimes. The results suggest that EULAG provides consistent and stable solutions for the various wave regimes (Doyle et al., 2010).

The paper is organized as follows. In section 3.2, the modifications to the standard EULAG model are briefly introduced and the method to simulate cirrus cloud formation under varying ambient states is described. Section 3.3 covers the simulation setup for the transient and non-transient reference simulations, followed by the results of the reference and sensitivity simulations in section 3.4. Conclusions and discussions are presented in section 3.5.
3.2 Model Description

As a basic dynamical model, the anelastic non-hydrostatic model EULAG is used (see Prusa et al., 2008). The dry anelastic equations solved in the model are presented in Smolarkiewicz and Margolin (1997). In this study, modified anelastic equations are used, as described in the following section.

3.2.1 Time dependent environmental states

We implemented a new framework for variable environmental states in the model EULAG. Here, we briefly derive the anelastic equations, which are used for the multiscale formulation. The physical basis is a dry, rotating fluid under constant gravity acceleration \(-g \mathbf{k}\) (see e.g. Durran, 1998). This leads to the following equations of general mass conservation (eq. 3.3), momentum (eq. 3.4) and inertial (eq. 3.5) energy, with the additional assumption of an ideal gas (eq. 3.6):

\[
\frac{Dp}{Dt} + \rho \nabla u = 0 \quad (3.3)
\]
\[
\frac{Du}{Dt} = -\frac{1}{\rho} \nabla p - g \mathbf{k} - 2\Omega \times u + \mathcal{F} \quad (3.4)
\]
\[
\frac{D\Theta}{Dt} = H \quad (3.5)
\]
\[
p = \rho RT \quad (3.6)
\]

Where \(\frac{D}{Dt} = \frac{\partial}{\partial t} + u \cdot \nabla\) denotes the total derivative, \(u\) the velocity vector and \(\Omega\) is Earth’s angular velocity. The terms \(H\) and \(\mathcal{F}\) summarize all external forces associated with diabatic processes \((H)\), diffusion and others \((\mathcal{F})\). \(R\) is the ideal gas constant and \(\Theta\) denotes potential temperature.

In a first step, pressure and density are split into a horizontally homogeneous hydrostatic background state (denoted by overbar) and perturbations (denoted by primes):

\[
p = \bar{p}(z) + p' \quad , \quad \rho = \bar{\rho}(z) + \rho' \quad , \quad \frac{\partial \bar{p}}{\partial z} = -g \bar{\rho} \quad (3.7)
\]

By linearizing the pressure gradient as discussed in Lipps and Hemler (1982,1985) and others as follows:

\[
\frac{1}{\bar{\rho}} \nabla p' \simeq \nabla \left( \frac{p'}{\bar{\rho}} \right) \quad (3.8)
\]
we obtain the generic form of the anelastic equations (see equations 1.52, 1.64 in Durran, 1998):

$$\nabla \cdot (\bar{\rho} \mathbf{u}) = 0$$  (3.9)

$$\frac{D \mathbf{u}}{Dt} = -\nabla \left( \frac{p'}{\bar{\rho}} \right) + g \frac{\Theta'}{\bar{\Theta}} \mathbf{k} - 2\Omega \times \mathbf{u} + \mathcal{F}$$  (3.10)

$$\frac{D \Theta}{Dt} = H$$  (3.11)

Here, we additionally make use of the approximation:

$$\frac{\bar{\rho}'}{\bar{\rho}} \approx \frac{\rho'}{\bar{\rho}} \approx -\frac{\Theta'}{\bar{\Theta}}$$  (3.12)

In a next step, we generalize this procedure by subtracting an environmental (or ambient) state, denoted by the subscript $e$ for the variables $p$, $\Theta$ and $\mathbf{u}$.

In general, the variables $p_e$, $\Theta_e$, $\mathbf{u}_e$ are not independent and the specific form of the partial differential equation derived by the subtraction procedure depends on the class of allowed environmental states. In a first derivation we postulate an inertial ambient state determined by a balance between pressure gradient, buoyancy and Coriolis forces:

$$0 = -\nabla \left( \frac{p_e - \bar{p}}{\bar{\rho}} \right) + g \frac{\Theta_e - \bar{\Theta}}{\bar{\Theta}} \mathbf{k} - 2\Omega \times \mathbf{u}_e$$  (3.13)

together with implied compatibility conditions.

By subtracting the environmental state from equations (3.10) and (3.11) in a similar way as done before for $\bar{\rho}$ and $\bar{\rho}$, the modified anelastic equations read as follows:

$$\nabla \cdot (\bar{\rho} \mathbf{u}) = 0$$  (3.14)

$$\frac{D \mathbf{u}}{Dt} = -\nabla \left( \frac{p'}{\bar{\rho}} \right) + g \frac{\Theta'}{\bar{\Theta}} \mathbf{k} - 2\Omega \times \mathbf{u}' + \mathcal{F}$$  (3.15)

$$\frac{D \Theta'}{Dt} = -\mathbf{u} \cdot \nabla \Theta_e + H$$  (3.16)
This form of the equations is used in various numerical models simulating atmospheric flows (e.g. Clark and Farley, 1984; Grabowski and Smolarkiewicz, 2002; Prusa et al., 2008). The variables indicated by primes denote deviations from the environmental state, i.e. for a variable $\psi$ the deviation is determined by $\psi' = \psi - \psi_e$.

It should be noted here that although the ambient state is chosen to be independent in spatial variables and time, this is not necessary a priori. In a further generalization we postulate the existence of an ambient state satisfying the generic anelastic equations:

\begin{align*}
\nabla \cdot (\bar{\rho} \mathbf{u}_e) &= 0 \quad (3.17) \\
\frac{De\mathbf{u}_e}{Dt} &= -\nabla \left( \frac{p_e - \bar{p}}{\bar{\rho}} \right) + g \frac{\Theta_e - \bar{\Theta}}{\Theta} \mathbf{k} - 2\Omega \times \mathbf{u}_e + \mathbf{f}_e \quad (3.18) \\
\frac{De\Theta_e}{Dt} &= H_e \quad (3.19)
\end{align*}

with $\frac{D}{Dt} = \frac{\partial}{\partial t} + \mathbf{u}_e \cdot \nabla$.

This postulated environmental state or ambient state ($\mathbf{u}_e, p_e, \Theta_e$) can be prescribed analytically, determined numerically by solving the equations or even adopted from meteorological analysis or measurements. Now, we expand the adiabatic equations by obtaining the differences (3.10) - (3.18) and (3.11) - (3.19); this procedure provides the following set of equations:

\begin{align*}
\frac{D\mathbf{u}}{Dt} &= -\nabla \left( \frac{p'}{\bar{\rho}} \right) + g \frac{\Theta'}{\Theta} \mathbf{k} - 2\Omega \times \mathbf{u}' + \mathbf{f} + \frac{De\mathbf{u}_e}{Dt} \quad (3.20) \\
\frac{D\Theta'}{Dt} &= -\mathbf{u} \cdot \nabla \Theta_e + H - \frac{\partial \Theta_e}{\partial t} \quad (3.21)
\end{align*}

The crucial new feature of these equations is the appearance of the forcing terms on the right sides of the equations. They act as "perturbations" in the whole model domain. Therefore the ambient but prescribed dynamics of the environmental states influences the solution of the equations (3.20) and (3.21).

In summary, we use a known 4D solution of the anelastic equations for each time step and each grid point as a forcing term in the equation solving the evolution of the dynamics under the influence of the (slowly) changing ambient state. Hence, we control the larger scale dynamical field (i.e. the ambient state) and obtain higher accuracy in solving the equations on the smaller scale, i.e. in the model domain.

In our simulations, we exclusively use of time-dependent environmental states in two dimensions. How-
ever, the model framework is prepared for the more general setup as described above.

### 3.2.2 Ice physics

A recently developed state-of-the-art bulk ice microphysical scheme is used (Spichtinger and Gierens, 2009a). Arbitrary many ice classes can be treated, distinguished by their formation mechanisms (e.g. heterogeneously vs. homogeneously formed ice). In our simulations, only homogeneous freezing of aqueous solution droplets (Koop et al., 2000) is considered; for high vertical velocities as given by mesoscale orographic gravity waves, we can assume that homogeneous freezing is dominant in terms of formed ice crystal number concentrations (e.g. Kärcher and Ström, 2003). Nevertheless, we want to mention that additional heterogeneous ice nuclei could influence and modify homogeneous freezing events (Spichtinger and Cziczo, 2010). The following processes for cold cirrus clouds are parameterized: Nucleation (homogeneous), deposition (growth, evaporation) and sedimentation (Spichtinger and Gierens, 2009a). The scheme is a consistent double moment scheme with terminal velocities for ice number and mass concentration. Aggregation is not yet implemented in the microphysics scheme. However, aggregation is of less importance for the cold temperature regime \( T < -38^\circ\text{C} \) and/or for moderate vertical velocities (Kajikawa and Heymsfield, 1989).

### 3.3 Experimental setup

In order to determine the influence of time dependent environmental flows on cirrus clouds, we use both transient and non-transient simulations, corresponding to situations with and without a changing large scale flow. An idealized framework was used for the simulations, including a 2-D model domain with a horizontal extent of \( L_x = 307\text{km} \) and a vertical extent of \( L_z = 20\text{km} \). Using a grid resolution of \( dx = 200\text{m} \) and \( dz = 50\text{m} \), in the horizontal and the vertical, respectively, the domain is discretized with \( nx \times nz = 1536 \times 401 \) grid points. The temporal resolution is set to \( dt = 2\text{s} \) increments for dynamics and \( dt = 0.1\text{s} \) sub steps for ice physics, respectively. For potential temperature, a constant-stratification background and ambient profile is used (see Clark and Farley, 1984) with a surface temperature \( T = 285\text{K} \) and a Brunt-Väisälä frequency \( N = 0.0115\text{s}^{-1} \). In the center of the domain, a Gaussian-shaped mountain is placed. The mountain shape can be described as

\[
H(x) = h_0 \cdot \exp\left(-\frac{x^2}{b^2}\right)
\]  
(3.22)
3.3. Experimental setup

where $h_0 = 400 \text{m}$ is the maximum ridge height and $b = 12.5 \text{km}$ stands for the half width of the mountain. The basic-state velocity field is a zonal wind as follows:

$$u_e(t,z) = \begin{cases} 
  u_g + \frac{z}{z_b} \cdot (u_0(t) - u_g) & \text{for } 0 \leq z < z_b, \\
  u_0(t) & \text{for } z \geq z_b.
\end{cases} \quad (3.23)$$

where $z_b = 2000 \text{m}$ denotes the upper limit of the boundary layer wind shear and $u_g = 5 \text{m/s}$ stands for the horizontal wind velocity at the frictionless surface, i.e. $z = 0 \text{m}$. The initial wind is extracted from the basic state using $u(z) = u_e(t=0,z)$. This study covers 5 different scenarios, including 2 transient and 3 non-transient simulation setups, using the following expressions for $u_0(t)$:

$$u_0(t) = \begin{cases} 
  5.5 \text{m/s} & \text{case 1, non-transient,} \\
  10 \text{m/s} & \text{case 2, non-transient,} \\
  14.5 \text{m/s} & \text{case 3, non-transient,} \\
  10 \cdot \tilde{U}_i(t) \text{ m/s} & \text{case 4, transient} \\
  10 \cdot \tilde{U}_d(t) \text{ m/s} & \text{case 5, transient.}
\end{cases} \quad (3.24)$$

For case 4, the transient part of the increasing wind $\tilde{U}_i$ is described as:

$$\tilde{U}_i(t) = \begin{cases} 
  1 & \text{for } t \leq 180 \text{min}, \\
  1 + 0.5 \cdot \sin \left( \frac{2\pi (t - 180)}{\tau} \right) & \text{for } 180 < t < 180 + \tau, \text{min}, \\
  1.45 & \text{for } 240 \text{min} \leq t.
\end{cases} \quad (3.25)$$

and in case of a decreasing wind (case 5), $\tilde{U}_d$ stands for:

$$\tilde{U}_d(t) = \begin{cases} 
  1 & \text{for } t \leq 180 \text{min}, \\
  1 - 0.5 \cdot \sin \left( \frac{2\pi (t - 180)}{\tau} \right) & \text{for } 180 < t < 180 + \tau, \text{min}, \\
  0.55 & \text{for } 240 \text{min} \leq t.
\end{cases} \quad (3.26)$$

where both cases use $\tau = 240 \text{min}$ and a transient period of $\tau_t = 60 \text{min}$. In other words, both transient scenarios start with $u_0 = 10 \text{m/s}$ (i.e. the basic state of case 2) and after 3 hours simulation time, they change their wind profile to a state as used in case 1 (decreasing wind scenario) or case 3 (increasing wind scenario) within $\tau_t$, respectively. The sensitivity on the length of the transition time $\tau_t$ is investigated in
Chapter 3. Cirrus clouds formed by a time dependent flow over a mountain

section 3.4.3.

The three initial wind profiles (i.e. at \( t = 0 \)) are shown in figure 3.1. All following simulations, the transient as well as the non-transient use horizontal wind conditions, which are within an upper and lower limit, defined by case 1 (“constant” wind \( u = 5.5 \text{ m/s} \)) and case 3 (“constant” wind \( u = 14.5 \text{ m/s} \)). The non-dimensional mountain height (i.e. the inverse Froude number \( \hat{h} = Fr^{-1} \)) is given by \( \hat{h} = Nh_0/u_0 \), thus using \( u_0 = 5.5/10/14.5 \text{ m/s} \) leads to \( \hat{h} = 0.84/0.46/0.32 \), which is below the limit for supercritical flows (i.e. \( \hat{h} < 1 \)). In all simulations, the effect of Earth’s rotation is not taken into account. Nevertheless the Coriolis force is important especially for transient flows over large mountain ridges (Bannon and Zehnder, 1985). For our simulation setup, the Coriolis force would damp the vertical velocity only slightly.

In order to allow cirrus cloud formation in the upper troposphere above the mountain ridge, an ice supersaturated region (ISSR) is placed between 8000m and 11000m with an initially constant relative humidity over ice (RHi) of 125%. This ISSR fully occupies the whole horizontal extent of this layer and is surrounded by 60% RHi below and 5% RHi above. A vertical absorber is placed above \( z = 14 \text{ km} \), using a time scale of 240s to avoid reflections of gravity waves at the upper boundary of the domain. To avoid any non-linear interaction before the waves enters the vertical absorber, there is no tropopause and no critical layer (i.e. a high altitude wind shear) placed in the model domain. Since we are interested in investigation of the impact of transient background wind fields on wave characteristics and cirrus formation, this idealization is justified.

![Figure 3.1: Three different initial profiles of the horizontal velocity of a flow over a mountain ridge.](image)
3.4 Results

3.4.1 Non transient simulations

In order to compare the effects of time dependent flows, three reference simulations using quasi steady state conditions (case 1/2/3) were carried out and for these cases, a stationary hydrostatic gravity wave was excited at the surface with an intrinsic horizontal phase velocity of $c_I = -u_0$. In this case, the total horizontal phase velocity $c_h = c_I + u_0$ is zero; thus the wave propagates with a vertical group velocity of $C_{gz} \approx 2\pi u_0^2 / (N\lambda_x)$, where the dominant horizontal wavelength can be approximated as $\lambda_x \approx 4b = 50\text{km}$ (see Nappo, 2002). Vertical propagation velocities of $C_{gz} \approx 0.33/1.09/2.30\text{m/s}$ can be derived for cases 1/2/3, respectively. This results in a propagation time up to the level of the ISSR of $t_p = 404/122/58\text{min}$. Since the horizontal wavenumber is small compared to the vertical wavenumber (i.e. quasi-hydrostatic state), we can use linear theory to estimate the vertical wavelength as $\lambda_z \approx 2\pi u_0 / N$. This leads to vertical wavelengths of about $\lambda_z \approx 3005/5465/7920\text{m}$ for cases 1/2/3, respectively. Compared to the model simulations, these values derived from linear theory applicable strictly only to infinitesimal small surface modulations are slightly underestimating the simulated vertical wavelengths. The values extracted from the simulations are: $\tilde{\lambda}_z \approx 3100/5800/8450\text{m}$. A possible explanation for this might be that the waves simulated by our model are not perfectly linear and hydrostatic (i.e. their wave axis is skewed downstream by up to $5^\circ$, see figure 3.2). Therefore, the vertical wavelength is slightly larger than the linear wavelength. Eventually, for these scenarios, non-linear effects only play a minor role.

The updraft regions over the mountain contain vertical velocity up to $w_{max} = 0.3/0.49/0.69\text{m/s}$ at $t = 6\text{h}$ for cases 1/2/3, respectively. If one of these updraft regions is located within the ISSR between $z = 8$ and $11\text{km}$ for sufficient time, the lifted air parcels can reach the threshold of homogeneous nucleation. The formed cirrus clouds show maximum ice water contents (IWC) up to $140\text{mg/m}^3$ for case 1, $115\text{mg/m}^3$ for case 2 and $80\text{mg/m}^3$ for case 3, respectively (see figure 3.3 a). It has to be taken into account that the frequencies are normalized over the number of grid boxes with $IWC > 0$.

Normalized frequency distribution of the vertical velocity inside the ISSR (i.e. between $8 \leq z \leq 11\text{km}$) are crucial to understand cloud formation (see figure 3.3 b). For the low-wind case (case 1), the velocities are symmetrically distributed and show a maximum of $w_{max} = \pm0.2\text{m/s}$. This implies that for this case, up- and downdrafts have an equivalent impact on the vertical flow regime. Nevertheless, the low-wind case shows a large amount of cloud ice. The other two cases show a highly asymmetric distribution of vertical velocities. In case 2, the ISSR is dominated by updrafts, whereas in case 3 downdrafts are
predominant. This results from the different vertical wavelengths of the gravity wave, and therefore from the different vertical positions of down- and updraft regions in the particular cases. It follows that the downdraft dominated high-wind case shows significantly lower ice water contents than the medium-wind case. It can be concluded, that not only the total maximum vertical velocity induced by the wave is an important factor for cirrus formation over mountains, but also the degree of overlapping between the updraft regions and the ISSR. The spatial and temporal evolution of cirrus clouds formed in the 3 non-transient (i.e. stationary) simulations are shown in a time series of the ice water path (IWP) (see figure 3.4 a - c). In the 2D representation (x-axis = horizontal extension, y-axis = time) contour lines of IWP are shown. The position of the mountain is additionally masked (peak height at $x = 0$ km).

In all cases, a maximum IWP can be observed near the mountain ridge (with a small lee-ward displacement). Only case 2 shows some significant cloud ice on the lee-ward side of the mountain within distances from the ridge of more than 100km, with maximum values of $IWP$ up to 0.35 g/m$^2$. This is

Figure 3.2: Vertical velocity and lines of constant potential temperature for three non-transient gravity waves, generated by three different flow profiles using $u_0 = 5.5/10/14.5$ m/s at $t = 300$ min after initialization.
3.4. Results

Figure 3.3: An orographic wave is generated by three different non-transient horizontal flow profiles using \( u_0 = 5.5/10/14.5 \text{ m/s} \) (i.e. case 1/2/3, respectively) and triggers cloud formation due to sufficient vertical updrafts inside an ISSR. This figure shows: a) Normalized frequency distribution of IWC for cases 1/2/3; b) Normalized frequency distribution of vertical velocity inside the ISSR between \( 8 \leq z \leq 11 \text{ km} \) for cases 1/2/3 due to advection of ice crystals formed in the hydrostatic wave. In contrast to case 3, there is no strong downdraft in the lee of the mountain, thus, in case 2 the ice crystals survived while they evaporated in case 3. In all cases, the main part of the cirrus is stationary over the mountain from its formation until the end of the simulation time. The time of the first cloud formation corresponds well with the arrival time based on the vertical group velocity of the gravity waves. This is most restrictive in case 1. Due to the small vertical group velocity (\( C_{gt} \approx 0.33 \text{ m/s} \)), the wave needs about 6.5h to reach the ISSR. This limits cloud formation to the last 1.5 hours of the simulation time. The cloud formed in this short time is quite dense (i.e. with large IWC) as compared to the other cases.

3.4.2 Transient simulations

Until \( t = 3h \) after initialization, the results of the two transient simulations are equivalent with case 2, because the wind profile in all three cases is constant, i.e. the same. After \( t = 3h \) the background wind is changing; thus, the relation for a stationary wave (i.e. \( c_h = 0 \)) is not fulfilled anymore, because the
Figure 3.4: Ice water path IWP (contour lines with 0.01 kg/m³ increments) of cirrus clouds triggered by non-transient orographic gravity waves in a 2D representation (x-axis = horizontal extension, y-axis = time) for: a) low-wind (case 1, $u_0 = 5.5$ m/s); b) medium-wind (case 2, $u_0 = 10$ m/s); c) high-wind (case 3, $u_0 = 14.5$ m/s). The position of the mountain is additionally marked.

vertical wavenumber $m$ of a wave, emitted at time $t$ is assumed to be conserved (i.e. the waves are generated on the surface and propagate upwards without changing their properties). But the background velocity $u_0$ changes with time, thus, the intrinsic phase velocity for already emitted waves is $c_I \neq -u_0$ and therefore $c_h > 0$ for case 4, or $c_h < 0$ for case 5, respectively. The imbalance of the intrinsic phase velocity and the background flow during the transient phase of the simulation can be observed as a horizontal dislocation of the wave away from its original position over the mountain ridge (see figure 3.5). This displacement corresponds well with the estimates done by Lott and Teitelbaum (1993), under the use of ray tracing for non-stationary topographic waves.

After $t = 4$ h $u_0$ remains constant for the rest of the simulation. Thus wave packets excited at the surface after this time, are stationary again (i.e. $c_I = -u_0$). Nevertheless, it has to be noted that for cases with an
3.4. Results

increasing wind, the new waves will interfere with the old ones, due to their higher propagation velocity. In this case, a new stationary solution is obtained only when all waves with smaller wavelengths than the new ones have left the domain (i.e. entered the absorber at $z = 14\text{km}$). Using the estimated vertical propagation velocity from above, this will need another 3.67h.

In order to explain the impact of transient waves on cirrus clouds, the time evolution of vertical velocities at $z = 10\text{km}$ is shown in figure 3.6. Whereas in the non-transient case 2 (figure 3.6 a) the updraft regions are stationary, cases with increasing (case 4) and decreasing wind (case 5) (see figure 3.6 b and c) show a significant horizontal displacement after the change in the horizontal flow profile.

In case of increasing wind (case 4), the flow regime strongly changes from an updraft dominated (like case 2) to a downdraft dominated regime, similar to the non-transient high-wind case 3. A strong asymmetric distribution (i.e. with larger negative branch) emphasizes the downdraft domination of the flow regime (see figure 3.7). Additionally also the sequence of the vertical motion from a Lagrangian perspective changes to the opposite case (i.e. from "updraft followed by downdraft" to "downdraft followed by updraft"). This can be seen directly in the sequence of updrafts and downdrafts in figure 3.6.

According to the fast vertical propagation of the wave emitted at $u_0 = 14.5\text{m/s}$, the new flow regime starts to dominate only about 1.5 hours after increasing the background flow. In case of decreasing wind (case 5), the change in the flow regime at $z = 10\text{km}$ is dominated by a decay of vertical motions, while the wave packets are advected upstream. In this case, the simulation time is not long enough to capture
Figure 3.6: Vertical velocity $w$ triggered by orographic waves on a $z = 10\text{km}$ cross section with $dw = 0.1\text{m/s}$ increments for a): non-transient medium wind case (case 2); b): increasing wind case (case 4); c): decreasing wind case (case 5). Red lines represent $w > 0$, blue lines: $w < 0$.

The full impact of the new wave inside the ISSR, because its vertical propagation is too slow (estimates for vertical propagation velocities are explained in section 3.4.1). The maximum vertical velocities can be estimated as $w_{\text{max}} = 0.4/0.45\text{m/s}$ for cases 4/5, respectively.

The impact of the different flow conditions on the formation and evolution of cirrus clouds is as follows. For both cases (increasing/decreasing wind), the horizontal dislocation of the wave patterns leads to a displacement (in the same direction) of the preexisting cloud as well. In the increasing wind case (case 4), the preexisting cloud moves downstream until it leaves the domain or its ice crystals sediment down to regions with subsaturated air (with respect to ice) and evaporate. Shortly after the transient phase, the new flow regime triggers a new stationary cloud directly over the mountain ridge (see figure 3.8 a).
3.4. Results

Figure 3.7: Normalized frequency distribution of vertical velocity triggered by orographic waves inside an ISSR between $8 \leq z \leq 11\text{ km}$ for increasing (case 4) and decreasing (case 5) wind case, additionally steady state conditions before the transient phase are shown (i.e. $t \leq 180\text{ min}$).

Figure 3.8: Ice water path (contour lines with $0.01\text{ kg/m}^3$ increments) of cirrus clouds triggered by orographic gravity waves in a 2D representation ($x$-axis = horizontal extension, $y$-axis = time) for: a) increasing wind case (case 4); b) decreasing wind case (case 5). The position of the mountain is additionally marked.

Maximum ice water path of the new cloud is $0.105\text{ kg/m}^2$, which is 84% higher than in the non-transient reference case (case 3, $IWP_{\text{max}} = 0.065\text{ kg/m}^2$). The decreasing wind case (case 5) shows a upstream displacement of the cirrus cloud, until it disappears at $t \approx 400\text{ min}$. While a small amount of cloud ice still exists over the mountain ridge, the formation of a new cloud (triggered by the new flow regime) can not be observed (see figure 3.8 b). This is because the "new" wave has a very slow propagation velocity and does not reach the ISSR before the end of the simulation. Thus, after the faster propagating "old" wave has left the ISSR, no sufficient updraft can be observed until new wave packets would arrive.

In the frequency distribution of IWC (see figure 3.9 a), we can see that the two transient scenarios show about the same maximum IWC of about $110\text{ mg/m}^3$, but with a lower occurrence of high ice water contents for the decreasing wind case (case 5). These values are in the same range than the non-transient
medium-wind case (case 2), where the low-wind case shows high IWC of up to about 140 mg/m$^3$ and the downdraft dominated high-wind case (case 3) has a maximum IWC of only about 80 mg/m$^3$. The maximum ice crystal number density follows the expectations one can infer from the vertical velocity distribution inside the ISSR (Kärcher and Lohmann, 2002). The increasing wind case (case 5) shows ice crystal number concentrations of up to 1500 L$^{-1}$ (see figure 3.9 b). Here, the interference of different wave packets leads to high vertical velocities. Numbers of about 900 L$^{-1}$ are derived in the decreasing wind (case 5) and in the non-transient medium wind case (case 2). This is not surprising because both cases have about the same maximum vertical velocity of about 0.4 m/s within the ISSR. Cases with significantly lower maximum vertical velocities (i.e. cases 1 and 3; $w_{\text{max}} \approx 0.2$ m/s) show much smaller ICNC up to 200 L$^{-1}$. This is not only due to a lower generation of supersaturation-peaks (Kärcher and Lohmann, 2002), but also because downdrafts are a possible sink of ice crystals (i.e. evaporation of ice crystals).

The spatial distribution of cloud ice (i.e. the total IWP) shows some significant differences between the transient and the non-transient cases (figure 3.9 c). The non transient cases show a strong maximum of IWP over the mountain and some a much weaker secondary maximum downstream (cases 2 and 3). The total cloud ice of the transient simulations shows a wider horizontal distribution without a strong maximum. This means that the area, covered by clouds, is larger in the transient cases.

### 3.4.3 Sensitivity simulations

In the reference simulations, the transient period was set to $\tau_t = 60$ min (i.e. the transition time from one flow state to the other). This was approximately the time scale $t_p$, that the fastest generated waves (i.e. at $u_0 = 14.5$ m/s) needed to reach the ISSR. In order to investigate the impact of $\tau_t$ on the dynamics and the microphysics of orographic cirrus clouds, a set of sensitivity simulations was performed using different values for $\tau_t$ ($\tau_t = 30/120/180$ min) for both increasing and decreasing wind scenarios as done in cases 4 and 5. These simulations cover scenarios, where the ratio between $\tau_t$ and the time scale $t_p$ is smaller or larger than 1. In other words, for cases with $\tau_t/t_p < 1$ the wind has completely changed from one state to another without any new waves reaching the ISSR during this time. In cases with $\tau_t/t_p > 1$, waves generated under new conditions already reach the ISSR, during the transient phase. A larger value of $\tau_t$, compared to the reference case, results into a slower horizontal displacement of wave parcels and therefore also in a slower displacement of the preexisting cirrus cloud. The slower displacement leads to an increased lifetime of the preexisting cirrus, especially pronounced for the decreasing
Figure 3.9: Properties of orographic cirrus clouds, generated by all transient and non-transient horizontal flows (case 1-5) over a mountain for the time period \( t = 1 \rightarrow 8 \)h. a) Normalized frequency distribution of ice water content IWC; b) Normalized frequency distribution of ice crystal number densities ICNC; c) Total ice water path IWP.

wind scenarios. This is shown in time series of IWP (see figure 3.11). In the case with \( \tau_t = 180 \)min, the upstream traveling cloud stays 80min longer (i.e. until the end of the simulation), than in the case using \( \tau_t = 30 \)min.

On the other hand, the impact of the new quasi equilibrium on the ISSR is delayed for larger \( \tau_t \). This is pronounced in the increasing wind scenarios, where formation of the new post-transient cirrus over the mountain ridge starts approximately 100min later for \( \tau_t = 180 \)min compared to the case with \( \tau_t = 30 \)min. This has an impact on the horizontal distribution of the total ice water formed during the simulation time. The increasing wind scenarios show a maximum of total IWP on the leeward side for large \( \tau_t \) (i.e. the preexisting cloud contributes more to the total IWP than the post-transient one), where for small \( \tau_t \) more cloud ice is formed directly over the mountain ridge (see figure 3.10 a). The decreasing wind scenarios show a slight increase in total IWP for larger \( \tau_t \), due to the larger cloud lifetime. The maximum total IWP for \( \tau_t = 180 \)min is 29% larger than for \( \tau_t = 30 \)min (see figure 3.10 b).
Figure 3.10: Total ice water path IWP of orographic cirrus clouds for sensitivity simulations using different transition times $\tau_t$ for the time period $t = 1 \rightarrow 8\, \text{h}$: a) increasing wind scenarios; b) decreasing wind scenarios.

3.5 Conclusions

The nonhydrostatic, anelastic model EULAG (Prusa et al., 2008) with a recently developed and validated ice microphysics scheme (Spichtinger and Gierens, 2009a) was used to investigate the influence of transient orographic waves on cirrus formation and evolution. Additionally introduced time dependent environmental states are used to prescribe transient environmental conditions on a larger scale. For this purpose, idealized profiles with supersaturations up to 125% RHi has been used for 3 nontransient and 2 transient scenarios. The focus is on comparing the impact of characteristic transient and nontransient orographic wave patterns on cirrus formation and evolution. This has been achieved by using constant and time dependent background flows, with increasing and decreasing wind, which represent the large scale dynamics and drive the processes on a smaller scale, which are resolved in our 2D domain.

The results can be summarized as follows:

Whereas the waves generated by non-transient runs are stationary, the transient simulations show significant horizontal displacements and vertical rearrangements of updraft regions due to an imbalance between the conserved horizontal phase velocity and the changing background flow, influencing the formation and evolution of cirrus clouds.

Following statements concerning the impact of transient orographic waves on cirrus formation can be done:
1. Time depending large scale flows cause a temporal imbalance between the intrinsic phase velocity of an orographic gravity wave and the background flow, leading to a horizontal displacement of the wave parcels.

2. In a transient environment, the vertical velocity regime inside an ISSR can change significantly due...
Chapter 3. Cirrus clouds formed by a time dependent flow over a mountain

to the horizontal displacement of wave parcels, the changing vertical wavelength and the different intensity of the background flow.

3. The transient simulations show a larger spatial distribution of cloud ice, because of the horizontal displacement of the updraft regions. The non-transient simulations are dominated by a stationary cloud directly over the mountain ridge.

4. Decreasing the background flow can result into a cutoff of cloud formation.

5. An increasing background flow does not necessarily result into a higher ice crystal number concentration, because even if the maximum vertical velocity increases, the vertical velocity regime (i.e. down- or updraft dominated) inside the ISSR can change on a contrary way.

6. If the wind and thus the dynamical regime changes slower, the lifetime of preexisting clouds is extended.

Non-transient simulations of (non-braking) gravity waves can give a good first estimate if cloud formation is possible under the given conditions and about its properties. But questions about the evolution and lifetime of the clouds remain open. The use of cloud resolving simulations with transient environmental states gives us new possibilities to look deeper inside the lifetime of an orographic cirrus cloud. An interesting fact is that the impact of a changing wind over a mountain on cirrus formation and evolution has two different types of effects.

1. Immediate effect: The horizontal displacement of wave parcels occurs at the same time as the environmental flow changes. This has an immediate effect on the location of the cloud.

2. Delayed effect: The change of the updraft regime due to the new wavelength of the orographic wave is delayed by the time the new wave needs to propagate through the troposphere up to the ISSR. This leads to a delayed impact on cirrus formation and evolution.

In other words, the impact of a sudden flow change is dispersed over several hours. This implies that for a strong changing environment, the impacts of different flow regimes are superimposing and interfering all the time. This makes it hard to distinguish the impact of the different flow regimes.

In this study, only cases with an approximately linear solution have been taken into account. The impact of transient environmental states on non-linear wave solutions (i.e. wave breaking, hydraulic jump, etc.) and the interaction with cirrus formation has to be investigated in the future. In a next step, more realistic simulations should be performed, using a more complex topography and realistic profiles for the
environmental states. The question, if this new method leads to an improved accuracy of 2D simulations can only be answered by comparing realistic simulations with measurement data. On the other hand, the impact of transient potential temperature $\theta_e$ has to be investigated.

**Acknowledgment**

We thank the European Center for Medium-Range Weather Forecasts for computing time (special project SPCHCLAI ”Cloud aerosol interactions”). This work contributes to the project ”Impact of dynamics on cirrus clouds” (Grant: 20021-117700) supported by the Swiss National Science Foundation (SNSF) and associated to the priority program 1276 MetStröm: ”Multiple Scales in Fluid Mechanics and Meteorology” of the German Research Foundation (DFG). We also thank Piotr Smolarkiewicz for his major contribution to the development and implementation of transient environmental states in EULAG. A successful implementation would not have been possible without his great knowledge of model architecture and numerics.
Chapter 4

Orographic cirrus clouds generated by a time dependent flow over the Andes

In this study, the influence of time dependent large scale flows on cirrus formation over mountains is investigated, using the non-hydrostatic, anelastic model EULAG including a bulk ice microphysics scheme. Realistic topography of two zonal cross sections of the Andes at 52.3°S and 53°S latitude are used together with ECMWF reanalysis data for vertical profiles of potential temperature \( \theta \), pressure \( p \) and horizontal wind \( u \), respectively. As a reference, non-transient simulations were performed, using constant environmental states. A set of transient simulations with time dependent environmental states (denoted by subscript \( e \)) for potential temperature \( \theta_e \) and horizontal wind \( u_e \) (both are constantly increasing with time) are compared with the non-transient simulations. Additionally, the results of all simulations are verified with aircraft measurement data obtained from the INCA campaign (Gayet et al., 2004). In all simulations, the generated orographic gravity waves propagate up to an ice supersaturated region (ISSR) between 8500m and 9500m altitude containing high supersaturations up to 130% and lead to the formation of orographic cirrus clouds. The simulated vertical velocities are in a good agreement with the measurements, but the microphysical properties show some offsets (i.e. slight overestimation of IWC and underestimation of ICNC in the simulations). Nevertheless, the horizontal position of the formed cirrus clouds represents the measurements very well. The transient simulations show a remarkably better agreement with the measurements than the non-transient ones, by containing a larger horizontal variability of cloud ice. This can be shown in a series of subsequently performed idealized simulations, using a single mountain ridge with spatial extensions comparable with the Andes. Additionally, the idealized simulations could give a deeper insight into the flow regime over the realistic terrain. It follows that the transient INCA simulations start with environmental conditions that favor wave breaking, but after some time the increasing wind leads to more stable conditions. During the transition between the non-linear

\[ \text{Fusina, F., P. Spichtinger, H. Joos and A. Dörnbrack, 2010. Orographic cirrus clouds generated by a time dependent flow over the Andes. J. Atmos. Sci., to be submitted.} \]
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and the linear regime, dynamical features caused by the non-linear flow tend to be advected downstream. This transition between two different regimes is completely ignored by the non-transient simulations. Thus, using transient environmental states improves the agreement with measurements of cloud resolving 2D simulations, also in realistic setups.

4.1 Introduction

Cirrus clouds, or clouds in general, have a crucial impact on Earth’s radiative budget. For low (liquid) clouds, the albedo effect (cooling via scattering/reflection of solar radiation) is dominating the greenhouse effect (warming via absorption and re-emission of terrestrial infrared radiation). But for cirrus clouds, both effects are of comparable size (with different signs), thus net warming or cooling is possible. Microphysical properties as shape or size of ice crystals are responsible for the transition between a net cooling or warming cirrus (see Zhang et al., 1999; Fusina et al., 2007). In the upper troposphere ice crystals form mainly via homogeneous freezing of aqueous solution droplets (Koop et al., 2000) or via heterogeneous nucleation at pre-existing insoluble aerosol particles (DeMott et al., 2003). For both mechanisms high ice supersaturation (up to $\sim 160 - 170\%\text{RHi}$) is required. A generic way for forming ice supersaturation is cooling of the air via adiabatic expansion, i.e. in air parcels lifted by upward motion. Updrafts can be generated by many different atmospheric processes on a variety of scales. For instance, orographic waves are a known source of updrafts, sufficient for cirrus formation (Joos et al., 2009). Following Dean et al. (2005), the amount of cirrus clouds formed by orographic waves over the continents is quite substantial. Therefore, their formation mechanism should be treated as accurate as possible. In current general circulation models (GCMs), orographic waves are handled as stationary waves, using linear theory to describe their temperature perturbations (Dean et al., 2007) or vertical velocity (Joos et al., 2008). Transient wave phenomena like horizontal displacement of wave parcels due to a changing large scale flow (see e.g. chapter 3 or Lott and Teitelbaum, 1993) are not considered in GCMs. From several synoptic observations near mountains it follows that changing large scale flow has an influence to the time-dependent component of the incident flow. This subsequently leads to a change in the wave forcing by the mountain (Smith, 1982; Bannon and Zehnder, 1985; Smith and Broad, 2003; Egger and Kühnel, 2010). However, before GCM parameterizations can be improved, accurate cloud resolving studies must be performed in order to investigate the potential impact of changing environmental conditions. Cloud resolving model investigations of orographic clouds usually use the stratified flow evaluated in almost steady state conditions, without regarding flow conditions that change with time.
There were some model investigations of transient flows over mountains in the last years, investigating basic wave mechanics (Lott and Teitelbaum, 1993) up to the interaction of linear and highly non-linear cases with large scales (Chen et al., 2006). Most of these investigations focus on changes in the flow on synoptic time scales. In chapter 3 the impact of transient orographic waves on cold ice microphysics was discussed for quasi-hydrostatic waves in an idealized 2D model domain, showing a significant horizontal cloud displacement for changing large scale flows. How important this effect is in a realistic setting has not been answered yet.

For this purpose, we use the non-hydrostatic, anelastic model EULAG (Prusa et al., 2008) with additionally introduced time dependent ambient states (see chapter 3), in order to represent transient large scale flows over mountains. For validation of the simulations, aircraft measurement data from the INCA (Interhemispheric differences in cirrus properties from anthropogenic emissions) campaign (Gayet et al., 2004) are used. The main goal of this study is to understand the differences between the simulations using time dependent environmental states to represent large scale dynamics and simulations using quasi steady-state conditions. Subsequently, questions about the advantages of transient states in 2D simulations will be answered. Additional idealized simulations are performed to better understand some of the processes in the realistic case.

4.2 Model description

As a basic dynamical model, the anelastic non-hydrostatic model EULAG is used (see Prusa et al., 2008). The dry anelastic equations solved in the model are presented in Smolarkiewicz and Margolin (1997). To represent time dependent large scale flows, modified anelastic equations are used, as described in the following section.

4.2.1 Transient environmental states

In the anelastic non-hydrostatic model EULAG, modified anelastic equations are solved in the perturbation form, using a background state. This set of equations can be further generalized by subtracting an environmental or ambient state (denoted by subscript $e$) for variables pressure $p$, potential temperature $\theta$ and velocity $\mathbf{u}$, respectively. We postulate that this environmental state is a previously known 4D solution of the generic anelastic equations. The environmental state can be spatially and temporally changing, in contrast to the usual assumption of a steady environmental state. After the expansion of the anelastic equations for different classes of environmental states, a new set of anelastic equation is derived:
\[ \nabla \cdot (\bar{\rho}u) = 0 \]  

\[ \frac{Du}{Dt} = -\nabla \left( \frac{p'}{\bar{\rho}} \right) + g \frac{\Theta'}{\bar{\Theta}} + 2\Omega \times u' + F + \frac{D_e u_e}{Dt} \]  

\[ \frac{D\Theta'}{Dt} = -u \cdot \nabla \Theta_e + H - \frac{\partial \Theta_e}{\partial t} \]

where \( D_e/Dt \equiv \partial/\partial t + u_e \cdot \nabla \). The crucial new feature is the appearance of the forcing terms on the right hand sides representing the change of the prescribed environmental state with time and space. These act as "perturbations" for the whole model domain. Thus, the environmental state dynamics influence the solution of the equations. Further insights are provided in section 3.2.1.

### 4.2.2 Ice physics

A recently developed bulk ice microphysic scheme is used, which can treat an arbitrary number of ice classes. These ice classes correspond to different nucleation processes (e.g. heterogeneous and homogeneous nucleation). The parameterization includes nucleation (homogeneous, heterogeneous), deposition (growth, evaporation) and sedimentation (Spichtinger and Gierens, 2009a) and provides a consistent double moment scheme with terminal velocities for ice number and mass concentration. In our simulations, only homogeneous freezing of aqueous solution droplets (Koop et al., 2000) is taken into account because for high vertical velocities and supersaturations in the low temperature range at \( T < -38^\circ \), it can be considered to be the dominant freezing mechanism (see e.g. Kärcher and Ström, 2003). Nevertheless, additional heterogeneous ice nuclei could influence the homogeneous freezing event and modify the properties of the formed cirrus clouds (Spichtinger and Cziczo, 2010). However, because of the investigation of orographic cirrus clouds in the Southern Hemisphere, the restriction to homogeneous nucleation only, assuming "clean air" is justified (see e.g. Minikin et al., 2003; Haag et al., 2003). Aggregation is not yet implemented in the microphysics scheme. However, aggregation is of less importance for the cold temperature regime (\( T < -38^\circ \)) and/or for moderate vertical velocities (Kajikawa and Heymsfield, 1989). A more detailed description of the ice microphysic scheme is given in Spichtinger and Gierens (2009a).
4.3 INCA scenario

Our main question is how important it is to consider time dependent large scale flows in cloud resolving simulations. Therefore, transient and non-transient simulations of a realistic scenario are compared with measurements. In our case, aircraft measurements from the INCA (Interhemispheric differences in cirrus properties from anthropogenic emissions) campaign (Gayet et al., 2004) are used as a realistic reference state. The INCA campaign took place in the year 2000 at two different places, namely Punta Arenas, Chile (April) and Prestwick, Scotland (October), respectively. For the comparison with our simulations, the measurements from Chile were chosen. This choice has mainly two reasons: First, during the campaign one day (5th of April 2000) was dedicated to measure orographic cirrus clouds. Second, in the Southern hemisphere the additional loading of heterogeneous ice nuclei can be neglected, thus homogeneous nucleation should be the dominant freezing mechanism. The corresponding flights took place between 17:30 and 19:30 UTC on zonal tracks between 69.2°W to 76°W longitude at 52.3°S and 53°S latitude, respectively. The aircraft’s altitude was between 9150m to 9440m. The following measured parameters are of importance for our study: Ice water content (IWC), ice particle number concentration (ICNC) and vertical velocity (w), respectively. The vertical velocities were measured with a five-hole probe (only during sections of constant altitude), estimating an accuracy in the order of 0.1 m/s (Bögel and Baumann, 1991). Ice particle number concentrations were derived, first, using a residual particle measurement with the Counterflow Virtual Impactor CVI (see Noone et al., 1993) and, second, using a combination of two instruments, the FSSP-300 and 2DC-C optical probe mounted onboard the DLR Falcon research aircraft (Gayet et al., 2002). The corresponding particle size range (diameter) used for this study is $3 - 800 \mu m$.

To compare the measurements with model simulations, a 2D EULAG model domain is initialized using a digital elevation profile from the National Geographical Data Centre (NGDC, Hastings et al., 1999) at 52.3°S and 53°S latitude (see figure 4.1), respectively. The surface is assumed to be frictionless. The

Figure 4.1: Terrain of a zonal cross section across the Andes at 52.3°S (upper panel) and 53°S (lower panel) latitude, between 65.25°W and 75.25°W longitude.
spatial extent of the model domain is $L_x = 1000\text{km}$ in the horizontal and $L_z = 20\text{km}$ in the vertical, using a horizontal resolution of $dx = 1000\text{m}$ and a vertical resolution of $dz = 50\text{m}$. The dynamical time step is set to $dt = 2.5\text{s}$ and the increments used for microphysics are $dt_m = dt/10 = 0.25\text{s}$. The simulation time for all scenarios is 13h. The Coriolis force is neglected in all simulations.

As input profiles for $p$, $\Theta_e$, and $u_e$, data from the ECMWF (European Center for Medium Range Weather Forecasts) reanalysis dataset are used for the 5th April 2000 at 12:00 / 18:00 / 24:00 UTC, according to the different simulation setups (see figure 4.2). Figure 4.3 shows a zoom on the upper troposphere potential temperature. Because the reanalysis data is only available in $0.5^\circ$ resolution, the profiles for $52.3^\circ S$ have been linearly interpolated using data from $52.5^\circ S$ and $53^\circ S$. The wind direction is approximately $260^\circ$, but we assume a pure west wind for all simulations, which is a reasonable assumption.

For all simulations, an ice supersaturated regions (ISSR) with an initial supersaturation of 130% RHi is
placed at 8500 – 9500 m altitude. The location is in a good agreement with the aircrafts altitude during the measurements. The simulations are split in two categories: a) non-transient scenarios, b) transient scenarios. A simulation overview is given in table 4.1. The specific simulation setups are as follows.

4.3.1 Non-transient scenarios

The non-transient simulations use the 18:00 UTC reanalysis data as a source of the initial $\theta_e$ and $u_e$ profiles (i.e. represented by the dashed lines in figure 4.2) at 52.3°S and 53°S latitude, respectively. These simulations correspond to a scenario without a changing large scale flow and are denoted by “nt”.

4.3.2 Transient scenarios

For the transient simulations, the time dependent environmental states are enabled in EULAG. The adequate forcings for $u_e$ and $\theta_e$ (see equations 4.2 and 4.3) are calculated at every time step by linear interpolation between the 12:00 / 18:00 / 24:00 UTC reanalysis data for both geographical latitudes. For every simulation setup transient simulations using time dependent $u_e$ and $\theta_e$ (these simulations are denoted by “tdlu”) and transient simulations using time dependent $u_e$ but constant profile for $\theta_e$ (using the 18:00 UTC ECMWF $\theta$ profile; denoted by “tdluθ”) are performed. These simulations correspond to a much more realistic scenario, which satisfies in a better way the multiscale aspect of cirrus cloud formation due to dynamical processes.
Table 4.1: INCA simulation overview. All simulations are performed for 52.3°S and 53°S latitude, respectively.

<table>
<thead>
<tr>
<th>name</th>
<th>( u_e )</th>
<th>( \theta_e )</th>
</tr>
</thead>
<tbody>
<tr>
<td>nt</td>
<td>18:00 UTC</td>
<td>18:00 UTC</td>
</tr>
<tr>
<td>tdlu</td>
<td>12:00 → 24:00 UTC</td>
<td>18:00 UTC</td>
</tr>
<tr>
<td>tdl( \theta )</td>
<td>12:00 → 24:00 UTC</td>
<td>12:00 → 24:00 UTC</td>
</tr>
</tbody>
</table>

Table 4.2: Time, latitude and altitude of three different aircraft measurement events over the Andes during the INCA campaign at 5th April 2000.

<table>
<thead>
<tr>
<th>name</th>
<th>time</th>
<th>latitude</th>
<th>altitude</th>
</tr>
</thead>
<tbody>
<tr>
<td>INCA521</td>
<td>17:30-18:00 UTC</td>
<td>52.3°S</td>
<td>9450m</td>
</tr>
<tr>
<td>INCA53</td>
<td>18:15-18:45 UTC</td>
<td>53°S</td>
<td>9450m</td>
</tr>
<tr>
<td>INCA522</td>
<td>19:00-19:30 UTC</td>
<td>52.3°S</td>
<td>9150m</td>
</tr>
</tbody>
</table>

4.4 INCA results

The zonal flow generates orographic wave packets at the surface. These waves are propagating upward through the whole troposphere inducing maximum vertical velocities of \( \pm 8 \text{m/s} \) at an altitude between 4 and 6 km. Once the wave packets reach the ISSR, they will lead to cloud formation if the induced vertical velocities are high enough to trigger high ice supersaturations above the homogeneous freezing thresholds. For this study, we are specially interested in parameters, which are important for cloud formation. More precisely, we investigate the histograms of vertical velocities \( w \), IWC and ICNC, respectively. Furthermore, the temporal evolution of the vertical velocity and the ice water path (IWP) is investigated, in order to get a better understanding of the cloud cover over the measurement site. The results of the simulations are compared with three different measurement events as described in table 4.2. For comparison, all non-transient simulations use a 30min slot between 4.5 and 5h simulation time as a reference period. At this time, the wave is fully established in the domain and represents an almost steady state solution. The transient simulations are brought into relation with a real time axis (UTC Time), trying to represent the measurement period as good as possible.

There are some restrictions for comparing 2D simulations with aircraft measurements. First, a 2D simulation setup tends to overestimate vertical velocities (see Dörnbrack, 1998). On the other hand, an aircraft cruising with \( \approx 170 \text{m/s} \) is not necessarily able to capture the highest peaks in vertical velocity. If we take this constraints into account, we can expect slightly higher vertical velocities in the simulations (Joos et al., 2009).
The normalized frequency of vertical velocities shows a quite good agreement with the measurements with maximum/minimum velocities of $\pm 2.5 \text{ m/s}$ (figure 4.4). All non-transient and transient simulations at 53$^\circ$S (INCA53) show about the same degree of overestimation (see figure 4.4b). The simulations at 52.3$^\circ$S (corresponding to INCA52$_1$ and INCA52$_2$) seem to produce higher vertical velocities in the transient scenarios ($\text{tdl}_u, \text{tdl}_\theta$) than for the non-transient (nt) ones (see figure 4.4a and 4.4c). Especially the non-transient INCA52$_2$ simulation shows significantly weaker downdrafts, compared with the measure-
ments. No significant difference can be observed between the two transient simulations $tdl_u$ and $tdl_{u\theta}$.

In order to get a good agreement of the simulated cirrus clouds with the measurements, not only the strength of down/updrafts must be equivalent but also their spatial distribution. For this purpose, time series of the corresponding flight level cross sections have been compared with the measured horizontal vertical velocity profiles.

For all scenarios, the transient simulations show a larger horizontal variability in vertical velocity than the non-transient simulations (see figures 4.5a to 4.5c). The horizontal distribution of updrafts and downdrafts corresponds better with the observations than the dynamics of the non-transient simulations which are limited to much smaller areas over the mountains. However, we cannot expect a perfect agreement between our 2D simulations and the measurements. Although none of the model simulations is able to precisely represent the measured vertical motions, an advantage of the transient simulations is obvious. The horizontal wavelength of the vertical velocity fluctuation is in the same range as the measurements with values of about $80 - 100\,\text{km}$ (derived from figure 4.5c). A significant difference between the two transient simulations $tdl_u$ and $tdl_{u\theta}$ can not be observed in any scenario, thus the impact of transient $\theta_e$ on dynamics is small in this cases. The evolution of the vertical updraft regions (INCA53 scenario) for the whole simulation time is shown in figure 4.5d. The non-transient simulation has a relatively steady horizontal distribution of the updraft regions, but compared with the transient simulation, the amplitude of the updrafts is oscillating with time. This is somehow contradictory to what we would expect from a non-transient simulation and suggests that some resonance effects could play a role. This makes it hard to find the right time period at which the simulation should be compared with measurements and points to some weakness of the use of steady state simulations.

The normalized frequencies of the ice water content show an unclear tendency (see figure 4.4d,e,f). Mainly because IWC strongly depends on the given initial amount of water vapor in the simulation, which is difficult to adjust for realistic cases, because no reliable measurements in the luv of the mountain exist. Whereas for the INCA52$_1$ scenario, all simulations underestimate the amount of ice water, scenario INCA53 shows a strong overestimation and scenario INCA52$_2$ a slight overestimation. On the other hand, it can be observed that the non-transient simulations always produce a slightly higher IWC than the transient ones. The difference is strongest for INCA53, where the nt-simulation reaches about $IWC_{\text{max}} \approx 76\,\text{mg/m}^3$, but the two transient simulations do not exceed $IWC_{\text{max}} \approx 60\,\text{mg/m}^3$. The transient simulation using both transient flow and potential temperature (i.e. $tdl_{u\theta}$) shows a remarkably lower
4.4. INCA results

Figure 4.5: Time series of horizontal vertical velocity cross sections generated by an orographic gravity wave over the Andes for non-transient (nt) and transient simulations using time dependent $u_e$ with a constant $\theta_e$ ($tdl_u$) or time dependent $u_e$ and $\theta_e$ ($tdl_{u\theta}$) compared with aircraft measurements (meas) for a) INCA52$_1$ scenario at 9450m; b) INCA53 scenario at 9450m; c) INCA52$_2$ scenario at 9150m; d) Comparison of non-transient with transient simulation ($w > 0$) of INCA53 scenario at 9450m ASL for the whole simulation time, using 0.5 m/s increments.

maximum for the ice water content in scenario INCA52$_2$, compared to $tdl_u$. The lower maximum IWC of $tdl_{u\theta}$ correlates better with the measurements than the other simulations, nevertheless, the accuracy of IWC between $25 - 55$ mg/m$^3$ is slightly overestimated. One reason for this difference could be that this transient simulation ($tdl_{u\theta}$) starts with lower temperatures than the nt and $tdl_u$ simulations (see figure
4.3), before it reaches a relatively steady temperature profile at 18:00 UTC. Because the initial relative humidity (i.e. at the left boundary of the model) is constant for the whole simulation, the specific humidity available for cloud formation was lower during the first 6h of the simulation and therefore less cloud ice has been produced. But it has to be considered, that for scenario INCA52$_1$ (which is located 1.5h earlier over the same terrain), the difference between those two simulations is smaller. The specific humidity also has an influence on the propagation of gravity waves (i.e. damping for higher specific humidity; see Dörnbrack, 1998). Considering the very similar updraft velocities and distributions for $tdl_u$ and $tdl_{u\theta}$, the effect of moisture on dynamics does not seem to be significant in this study. The absolute humidity in the domain must be substantially larger to have a significant impact (i.e. wave damping).

If we compare the simulated ICNC with the data from the combined optical measurement method (meas), we can see that the model can not precisely reproduce the high numbers derived by the instruments (see figures 4.4g,h,i). Overestimation of measured ICNC due to shattering of ice crystals can be excluded as different techniques have been used (Gayet et al., 2006). The data derived by the CVI fits much better to the simulations, but the use of this method for cirrus is not advisable because it is known that this method underestimates high ice crystal number concentrations (Kärcher and Ström, 2003). Concerning the ICNC, the non-transient and transient simulations do not show any significant differences, except in scenario INCA52$_2$, where the $tdl_{u\theta}$ simulation can reproduce the high number concentrations of the measurement. The reason for that might also be the lower temperatures in the first part of the transient simulation. For such conditions a higher ICNC can be expected (see Kärcher and Lohmann, 2002). Combined with the accurate IWC, simulation setup $tdl_{u\theta}$ seems to be the most realistic of all scenarios in order to represent the measurement as good as possible.

By looking at the temporal evolution of the ice water path (IWP) during the measurement periods, we can estimate the zonal position of the formed cirrus clouds. Using the IWP gives us a better overview over the whole cirrus formation above the terrain, without having a limitation to a specific altitude. Therefore, we have to keep in mind, that the horizontal spread of the IWP might be larger than the spread of the measured IWC on the compared 1D flight track, but we have the advantage that no ice water is missing in the comparison just due to an inaccurate cross section altitude in the simulation data. Additionally it has to be noted, that there are no measurements for $x > 645$ km. The 2D representation ($x$-axis = spatial extension; $y$-axis = time) of the IWP shows the time period corresponding to the different aircraft measurements (figure 4.6). Thus, a virtual flight track through this 2D representation would start in the lower left and end in the upper right part of the plot. The comparison with the aircraft measurement can
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Figure 4.6: Time series of the ice water path (IWP) of orographic cirrus clouds for non-transient (nt) and transient simulations using time dependent $u_e$ with a constant $\theta_e$ (tdl$_u$) or time dependent $u_e$ and $\theta_e$ (tdl$_{u\theta}$) compared with aircraft measurements (meas) for a) INCA52$_1$ scenario; b) INCA53 scenario; c) INCA52$_2$ scenario, using 0.01 kg/m$^2$ increments.

be described as follows:

- INCA52$_1$ scenario (figure 4.6a): The aircraft measurement shows a bimodal distribution of cloud ice, with a distance of $\sim$ 100km between the peaks. All simulations show a bimodal cloud structure, but the two transient simulations (tdl$_u$ and tdl$_{u\theta}$) reproduce the distance between the peaks more precisely. The cirrus clouds in the non-transient simulation (nt) show a lower horizontal variability with a strong peak at $x = 500$km. Thus, the spatial distribution of cloud ice of the transient simulations fits better to the measurements than the non-transient one.
• INCA52 scenario (figure 4.6c): Different to the earlier INCA52\textsubscript{1} measurement, the spatial distribution of the cloud ice shows a more uniform shape with three local peaks between $x = 500$ and $650\text{km}$. This shape is best reproduced by the $tdl_{u\theta}$ simulation. But nevertheless, $tdl_u$ performs also very well and shows about the same horizontal cloud extension. The non-transient case shows the same bi-modal structure than described before, which does not exactly fit to the measurements. On the other hand, the simulated cloud extension is not larger than $100\text{km}$, where the extension in the transient simulations is larger than $150\text{km}$. Thus, the transient simulations show (as in INCA52\textsubscript{1}) a significantly larger horizontal variability of cloud ice.

• INCA53 scenario (figure 4.6b): The measured cirrus cloud is located between $x = 550$ and $650\text{km}$. The horizontal dimension of the cloud is well reproduced by all simulations, but for all cases the horizontal position is $\sim 50\text{km}$ more in the upstream direction. Additionally, the spatial distribution of cloud ice in the nt case is very unsteady during the reference period. This is mainly due to the unsteady amplitude of vertical velocity during the whole nt-simulation. It is unclear, why the simulations are able to reproduce the horizontal cloud extension more accurate for the INCA52 scenarios than for the INCA53 scenario. One reason could be, that at $53^\circ\text{S}$, three-dimensional effects (which are not resolved in our simulation) are more important than at $52^\circ\text{S}$.

### 4.5 Idealized cases

A typical mountain dimension from the Andes cross section would be around $h_0 = 800\text{m}$ in altitude and $2b = 50\text{km}$ in width. Those values are taken to perform a set of idealized simulations, in order to try to understand some processes that happen in the realistic case in a more isolated framework, modeling only one mountain ridge. Mainly, we try to answer questions about the flow regime and the spatial distribution of cirrus clouds in the INCA case, which can not be answered by just looking at the realistic simulations. Because of the manifold numbers of different wavelengths generated over the realistic topography, strong wave interference inhibits detailed wave analysis. We use a 2D domain with a horizontal extent of $L_x = 204.6\text{km}$ and a vertical extent of $L_z = 20\text{km}$ for dynamics and ice microphysics, respectively. The domain is discretized with $nx \times nz = 1024 \times 401$ grid points, using a grid resolution of $dx = 200\text{m}$ and $dz = 50\text{m}$, in the horizontal and the vertical, respectively (using open lateral boundaries). The used temporal resolution is $dt = 1\text{s}$ for the dynamics and $dt = 0.1\text{s}$ for the microphysics, respectively. The simulations last for $7\text{ hours}$. A Gaussian shaped mountain is placed in the middle of the domain, using
4.5. Idealized cases

Figure 4.7: Idealized profiles of left: horizontal wind with $u_{\text{max}1} = 17.5 \text{ m/s}$ (solid line) and $u_{\text{max}2} = 22.5 \text{ m/s}$ (dashed line); right: potential temperature profile.

The following expression for the height:

$$H(x) = h_0 \cdot \exp \left( -\frac{x^2}{b^2} \right)$$

The $\theta_e$ and $u_e$ profiles are idealized using the shape of the realistic profiles from the ECMWF reanalysis data of the INCA campaign (see section 4.3) for two distinct scenarios (i.e. $u_1$ and $u_2$). After a strong wind shear below $z = 2 \text{ km}$, the two wind profiles increase to constant velocities of $u_{\text{max}1} = 17.5 \text{ m/s}$ (for $u_1$, relating to the 12:00 UTC INCA profile) and $u_{\text{max}2} = 22.5 \text{ m/s}$ (for $u_2$, relating to the 18:00 UTC INCA profile). A jet stream is placed between $8 \leq z \leq 11 \text{ km}$, showing maximum horizontal velocities of $u_{\text{jet},1} = 25 \text{ m/s}$ and $u_{\text{jet},2} = 35 \text{ m/s}$ for $u_1$ and $u_2$, respectively (see figure 4.7 left). The mean tropospheric stability is $N^2 = 0.012 \text{s}^{-1}$ ($N$ denotes the Brunt-Vaisala frequency) and a tropopause is placed at $z = 10 \text{ km}$ (see figure 4.7 right). According to the realistic INCA scenarios, an ice supersaturated region is placed between $8500 \leq z \leq 9500 \text{ m}$, containing an RHi of 130% within, 20% below and 5% above the ISSR.

We discriminate again between simulations using a transient horizontal flow and simulations with a constant horizontal flow.

1. Non-transient simulations:

Two simulations, using a maximum horizontal velocity of $u_1$ ("nt 17.5" = 17.5 m/s) and $u_2$ ("nt 22.5" = 22.5 m/s), respectively (referring to the value of $u$ in the altitude range between ~ 2 and
Chapter 4. Orographic cirrus clouds generated by a time dependent flow over the Andes

4.5.1 Results

Investigating the evolution of the generated vertical motion above the mountain ridge, it becomes obvious, that two different types of flow regimes play a roll. The non-transient simulation "nt 17.5" shows a typical non-linear solution (i.e. wave breaking occurs, see figure 4.8a). On the other hand, the solution of "nt 22.5" can be stated as quasi-linear, generating a hydrostatic gravity wave with some minor reflections (i.e. at the tropopause), showing a vertical wavelength of $\sim 11$ km (see figure 4.8b). The maximum ve-
4.5. Idealized cases

Locilities of both simulations do not necessarily lead to a non-linear solution, because their corresponding non-dimensional mountain height \( \hat{h} = Nh_0/u \) is below the threshold of 1 (i.e. \( \hat{h}(17.5 \text{ m/s}) = 0.55 \) and \( \hat{h}(22.5 \text{ m/s}) = 0.43 \)). The critical horizontal velocity that would lead to non-linearities (i.e. \( \hat{h} \geq 1 \)) is \( u_{\text{crit}} \leq 9.6 \text{ m/s} \). Due to the strong windshear below \( z = 2 \text{ km} \), this critical velocity is reached in both idealized horizontal velocity profiles. However, only the low-wind case "nt 17.5" shows significant wave breaking. The same two regimes occur also in simulations without a jet stream (not shown), thus, any sort of shear instability near the jet can be excluded as a main source for non-linearities in our case.

The transient simulations show a transition between the two flow regimes. The increasing case (case "tdl-inc") starts under non-linear flow conditions, but with increasing wind, the wave above the mountain becomes linear and the induced non-linearities from the first part of the simulations are advected downstream. This can be seen at \( t = 420 \text{ min} \) in figure 4.8c at \( x \approx 60 \text{ km} \) where a small hydraulic jump travels downstream and generates vertical motions. The decreasing wind scenario (case "tdl-dec") starts with linear flow conditions, but with decreasing horizontal wind, the flow is about to generate wave breaking just at the end of the simulation. This can be seen in figure 4.8d by looking at the lines of constant \( \theta \). However, the growing non-linearities do not lead to a substantial increase of vertical velocity until the end of the simulation.

Due to the wave breaking, "nt 17.5" shows significantly higher vertical velocities (up to \( w_{\text{max}} = 14.5 \text{ m/s} \)) than the other non-transient but linear case (\( w_{\text{max}} = 2.5 \text{ m/s} \)). Both transient simulations show maximum vertical velocities of \( w_{\text{max}} = 4 - 5 \text{ m/s} \) (see figure 4.9a). Even though the increasing transient case starts with a potential non-linear solution, no wave breaking occurs, because the increasing horizontal velocity rapidly generates a quasi-hydrostatic wave similar to case "nt 22.5".

The high vertical velocities in case "nt 17.5" generate large ice crystal number densities of up to 100 cm\(^{-3}\), whereas the other three simulations show maximum ICNC of about 10 cm\(^{-3}\). The formed ice water content is in the same range for the simulations "nt 17.5", "tdl-inc" and "tdl-dec" and a bit lower for the non-linear case "nt 22.5". The spatial evolution of the formed cirrus clouds is shown in figure 4.10. The time series of the ice water path show a typical horizontal displacement of the cirrus core (i.e. downstream for "tdl-inc" and upstream for "tdl-dec"). This is due to a temporal imbalance between the horizontal phase velocity of the orographic wave (conserved after emission) and the background flow (changes with time) as discussed in chapter 3. In the non-transient cases, the center of the cirrus cloud is stationary over the mountain ridge. This transient effect might be a reason for the larger downstream IWP of the transient INCA simulations. Additionally, the downward advected wave disturbances from
the non-linear start-phase of the increasing wind scenario (case "tdl-inc") are strong enough to influence the formation and evolution of cirrus clouds even at a distance of $\sim 75\,\text{km}$ away from the mountain ridge. This can be seen by comparing the ice water path of "tdl-inc" with the non-transient "nt 22.5" scenario at $t = 420\,\text{min}$ (see figure 4.11). At this time, both cases have the same horizontal velocity profile and similar updrafts directly above the mountain. This effect could also contribute to the differences in cloud cover between the transient and the non-transient INCA simulations.

The idealized simulations also enable us to say something about the flow regime in the INCA scenarios. According to the idealized simulations, the transient INCA simulation starts with conditions that favor wave breaking, but with time (i.e. increasing horizontal flow) the stability of the solution increases. Besides the contribution to the understanding of the INCA case, the idealized simulations give an idea how orographic waves change from a linear to a non-linear regime (and vice versa).

4.6 Conclusions and summary

A two-dimensional version of the non-hydrostatic anelastic model EULAG with recently introduced transient environmental states has been used for different formation scenarios of orographic cirrus clouds. The main task of this study was to investigate the difference between transient and non-transient simulations using atmospheric profiles and topography related to the INCA campaign and to compare the
Figure 4.10: Time series of the ice water path (IWP; using 0.01 kg/m$^2$ increments) generated by gravity waves over an idealized mountain located at $x = 0$ km, derived from the non-transient simulations a) $u_{\text{max}} = 17.5$ m/s (case "nt 17.5"); b) $u_{\text{max}} = 22.5$ m/s (case "nt 22.5"); transient simulations c) case "tdl-inc" with increasing horizontal velocity from 17.5 m/s to 22.5 m/s and d) case "tdl-dec" with decreasing horizontal velocity from 22.5 m/s to 17.5 m/s.

Figure 4.11: Ice water path of cirrus clouds generated by orographic gravity waves in two different simulations at $t = 420$ min. The dashed line stands for a non-transient reference case with $u_{\text{max}} = 22.5$ m/s and the solid line represents a transient case with $u_{\text{max}}$ increasing from 17.5 to 22.5 m/s. At this time, both simulations have the same horizontal flow profiles.

results with aircraft measurement data. The transient simulations are distinguished between cases using transient horizontal wind $u_e$ and cases using both, transient environmental wind $u_e$ and potential temperature $\theta_e$. In the second part, simulations using idealized topography and environmental states (on the basis of the INCA scenarios) have been performed, in order to investigate the impact of transient horizontal winds over a mountain on cirrus formation in a more isolated framework. The reduction of strong wave interference and reflection, which is typical for a realistic scenario like INCA, allows a better view on the different involved processes. The idealized simulations could help to answer some questions about
the flow regime and the horizontal cirrus distribution of the INCA case.

The results can be summarized as follows:

The zonal flow over the mountains generates orographic waves, which propagate upward through the atmosphere. The induced vertical motions are in the same range as the measurements, but their spatial distribution is hard to compare. The measured horizontal wavelength of the vertical velocity perturbations is well reproduced by the model. The measured and simulated microphysical properties of the formed cirrus clouds are also in the same range. The scenarios at 52.3°S latitude (i.e. INCA52₁ and INCA52₂) reproduce the horizontal cirrus position and extension very well, whereas the scenario at 53°S shows a small upstream displacement compared to the measurement. Regarding the spatial distribution of cloud ice, the transient simulations show a better agreement with the measurements than the non-transient ones. With respect to normalized frequencies of the vertical velocity ($w$) and the ice crystal number concentration (ICNC), no significant difference could be observed between transient and non-transient simulations, whereas the ice water content (IWC) is slightly higher in the non-transient cases. Simulations using both transient $u_e$ and $\theta_e$ show a slight advantage in producing a more realistic amount of cloud ice and number concentrations, compared to simulations with constant $\theta_e$ profile. In our case, a changing $\theta_e$ has no significant influence on the stability of the troposphere, therefore its impact on dynamics is neglectable (i.e. only marginal changes to the flow regime).

The performed idealized simulations used a single mountain ridge with dimensions comparable to the INCA terrain and similar horizontal flow profiles. It could be shown that the zonal wind at 12:00 UTC in the INCA data ($u_{max} = 17.5 \text{ m/s}$) favors wave breaking above the single ridge, but the wind at 18:00 UTC ($u_{max} = 22.5 \text{ m/s}$) shows a more linear solution (i.e. a quasi-hydrostatic gravity wave). For the INCA case, this means that the transient simulation shows a transition from a non-linear to a linear flow regime, but the non-transient simulations are always in the linear regime. The stronger downstream variability of cloud ice in the transient simulation could also be reproduced by the idealized simulation. The main reason for this is probably the horizontal displacement of the formation regions of cirrus clouds, caused by a temporal imbalance between the emitted horizontal phase velocity (conserved) of the wave and the changing environmental flow. On the other hand, the idealized simulations showed that in the case of a transition from a non-linear to a linear flow regime, non-linearities are advected downstream and possibly influence cirrus formation and evolution. It is hard to separate the impact of these two processes because they happen at the same time and interfer strongly.

In summary:

1. The transient simulations show a larger horizontal variability of vertical velocities and IWC than
the non-transient ones. This correlates better with measurements, thus the use of transient environmental states in small scale models can be recommended to increase the accuracy of the results.

2. Using transient potential temperature profiles beside the transient flow has mainly an influence on the microphysical properties of the formed clouds, rather than on dynamics. Note that in our simulations, the changing temperature profile has not induced a change of the atmospheric stability (i.e. the rate of change was small and almost vertically uniform).

3. When the flow changes from a non-linear to a linear regime (by increasing the horizontal wind), pre-existing non-linearities (i.e. hydraulic jump, wave breaking) tend to be advected downstream. The induced vertical motions are strong enough to influence the evolution of cirrus clouds. This could be another reason for the larger horizontal variability of cloud ice in the transient INCA simulations. Thus, the time-dependent simulation tends to smooth the non-linearities and leads to a more "linear" behaviour of the flow.

One of the most important questions that arises is, how accurate the results can be, by using a 2D cloud resolving model to reproduce a detailed, realistic scenario. This question remains somehow unanswered, because we did not perform 3D simulations for comparison. But we have seen that using transient environmental states can lead to a further improvement of 2D simulations. The multiscale aspect of cirrus formation is much better represented when time dependent large scale flows are embedded in cloud resolving simulations. On the one hand, there is an impact on dynamics, because flow regimes over mountains change under different wind conditions, and secondly changing temperature regimes have an impact on the microphysical properties of the formed clouds. Generally, the results of the 2D simulations agree well with the measurements.

The cloud cover of cirrus clouds formed by orographic waves over the continents is quite substantial (Dean et al., 2005) and due to their strong impact on the radiative budget (see Zhang et al., 1999; Fusina et al., 2007), the cirrus formation should be treated as accurate as possible. A future task would be that such transient effects in orographic cirrus formation and evolution would be a part of the orographic cloud parameterization in GCMs. As a first estimate, a more realistic treatment of transient orographic waves would increase the cirrus cloud cover due to the increased horizontal variability.

Acknowledgment

We thank the European Center for Medium-Range Weather Forecasts for computing time (special project SPCHCLAI "cloud aerosol interactions"). This work contributes to the project "Impact of dynamics on
cirrus clouds” (Grant: 200021-117700) supported by the Swiss National Science Foundation (SNSF) and associated to the priority program 1276 MetStröm: "Multiple Scales in Fluid Mechanics and Meteorology” of the German Research Foundation (DFG). We also thank Piotr Smolarkiewicz for his major contribution to the development and implementation of transient environmental states in EULAG. A successful implementation would not have been possible without his great knowledge of model architecture and numerics.
Chapter 5

Summary and Outlook

5.1 Summary

In this thesis, the non-hydrostatic, anelastic model EULAG (Prusa et al., 2008) was used to investigate the influence of the competing large-scale and mesoscale dynamical processes on cirrus formation and evolution. For this purpose, a modified version of the model was used, including a recently developed and validated ice microphysics scheme (Spichtinger and Gierens, 2009a) and a fast radiative transfer code (Fu, 1996; Fu et al., 1998). Additionally, time dependent ambient states for horizontal velocity and potential temperature have been added in order to represent large scale forcing (see section 3.2.1). The following two topics were discussed:

5.1.1 Impact of a changing environment due to radiative processes on cirrus formation

For this part of the thesis, idealized simulation setups with high supersaturations up to $RHi = 144\%$ and weak potential stabilities are used. The cooling due to thermal emission at the top of the ice supersaturated region (ISSR) destabilizes the initially weak profile within several hours (depending on the initial stability and RHi). During the destabilization the amplitude of pre-existing small eddies increases and leads to the first nucleation of ice crystals. In the reference case, the destabilization process lasted for 4h until first nucleation occurred. Supported by the subsequent latent heat release, small convective cells are formed with vertical velocities of up to $1.6\text m/s$. The formed cirrus clouds are persistent over many hours and show mean optical depths up to $\tau = 0.35$. Sensitivity simulations showed that increasing the initial potential stability results in a delay of the first nucleation by up to several hours and decreases the strength of the nucleation event (i.e. lower ice crystal number concentrations and ice water contents). The radiative destabilization becomes subordinate, when another (faster) destabilization process (i.e. shear instability) takes place at the same time.
5.1.2 Impact of a changing environment using time-dependent ambient states on cirrus formation

In a first step, the impact of time dependent large scale flows on basic orographic wave mechanics and cirrus formation was investigated. This was achieved by using constant and time dependent background flows with increasing and decreasing horizontal wind velocity, respectively. This background flows represent the large scale dynamics and drive the processes on a smaller scale, which are resolved in our 2D EULAG domain. In all simulations, an orographic wave is generated at the surface and propagates vertically towards an ISSR. Whereas the waves generated by non-transient simulations are stationary, the transient simulations (i.e. decreasing or increasing wind scenarios) show significant horizontal displacements of updraft regions due to an imbalance between the conserved horizontal phase velocity and the changing background flow. On the other hand, a vertical rearrangement of the updraft and downdraft regions is induced by the changing wavelength of the wave. This modifies the vertical motion regime inside an ISSR and subsequently has a crucial impact on cirrus cloud formation and evolution. Regarding the impact of transient flows on orographic cirrus clouds, two different effects can be distinguished:

1. Direct effect: The horizontal displacement of wave parcels occurs at the same time as the environmental flow changes. This has an immediate effect on the location of the cloud.
2. Delayed effect: The change of the updraft regime due to the new wavelength of the orographic wave is delayed by the time the new wave needs to propagate through the troposphere up to the ISSR. This leads to a delayed impact on cirrus formation and evolution.

In other words, the impact of a sudden flow change is dispersed over several hours. This implies that for a fast changing environment, the impacts of different flow regimes are superimposing and interfering. This makes it hard to distinguish the impact of the different flow regimes.

Thus, the transient simulations showed an increased horizontal distribution of cloud ice, compared to the non-transient reference simulations, which formed strong stationary IWC peaks directly above the mountain ridge. The total amount of IWC formed in the simulations is not very sensitive to transient flows, but the changing vertical velocity regimes influence the ICNC in the transient simulations. In increasing wind scenarios, the new waves (fast propagation) interfere with old waves (slow propagation), leading to high vertical velocities (and therefore high ICNC; Kärcher and Lohmann, 2002) in case of constructive interference. Many of these transient effects are neither linear nor have the same time scale, therefore it is not possible to predict their impact on cloud formation using non-transient simulations. However, non-transient simulations can give a good first estimate about the basic properties of orographic clouds. Another comparison with measurements is discussed in section 5.2.2.
5.1. Summary

After these idealized simulations that give a first overview of the impact of transient flows on orographic waves, more realistic simulations were performed. The results were compared to aircraft measurement data, in order to classify the benefit of using transient ambient states. For this purpose, atmospheric profiles and topography representing the environment of the INCA campaign were used for different transient and non-transient scenarios. The transient simulations are using a time dependent, constantly increasing background flow $u_e$, in some cases together with time dependent potential temperature $\theta_e$ profiles. The zonal flow over the mountains generates orographic waves, which propagate upwards through the atmosphere. The induced vertical motions are in the same range as the measurements, but their spatial distribution shows some inaccuracies. However, the measured horizontal wavelength of the vertical velocity perturbations is well reproduced by the model. The measured and simulated microphysical properties of the formed cirrus clouds are comparable, but with some small inaccuracies. Regarding the spatial distribution of cloud ice, the transient simulations show a larger horizontal distribution, which is in better agreement with the measurements than the results of the non-transient simulations. As for the normalized frequencies of $w$ and ICNC, no significant difference could be observed between transient and non-transient simulations, but the IWC is slightly higher in the non-transient cases. Simulations using both transient $u_e$ and $\theta_e$ show a slight advantage in producing a more realistic amount of cloud ice and number concentrations, compared to simulations with a constant $\theta_e$ profile. However, the influence of a time dependent $\theta_e$ on dynamics (i.e. wave propagation, vertical velocity) was marginal. Nevertheless, it has to be considered that in our simulations the changing temperature profile has not induced a change of the atmospheric stability (i.e. the rate of change was small and almost vertically uniform).

In a next step, simulations using idealized topography and ambient states similar to the INCA scenario have been performed, in order to answer some remaining questions concerning the observed cloud evolution and the flow regime in a better controlled, thus isolated framework. The reduction of strong wave interference and reflection, which is typical for a realistic scenario like INCA with multiple mountains and different mountain widths, gives a better and clearer view on the different involved processes. These simulations showed a non-linear flow regime for slow horizontal velocities ($u = 17.5 \text{ m/s}$), but a linear regime for a stronger horizontal wind ($u = 22.5 \text{ m/s}$). Therefore, a transient scenario with decreasing wind ($u = 22.5 \rightarrow 17.5 \text{ m/s}$) starts under linear conditions and becomes non-linear after some time. In the performed simulation, the flow generated a small hydraulic jump at the very end of the simulation. Due to the limited simulation time, the growing non-linearities did not lead to a substantial increase of
vertical velocity. On the other hand, a transient scenario with increasing wind \((u = 17.5 \rightarrow 22.5 \text{ m/s})\) includes the transition from a non-linear to a linear flow regime. Non-linearities (i.e. wave breaking, hydraulic jump, etc.) caused by the early flow regime are advected downstream when the horizontal flow increases and the flow regime becomes linear. In our simulations, such displaced disturbances had a crucial influence on cloud formation and evolution even at distances of more than \(\sim 75 \text{ km} \) downstream from the mountain ridge. Following this idealized representation of the INCA case, downward propagation of non-linear effects could also be a reason (beside the horizontal displacement due to non-steady waves) for the larger horizontal variability of cloud ice in the transient INCA simulations.

The presented results lead to the following conclusions: Introducing time dependent large scale dynamics to a cloud resolving model induces changes in mesoscale/smallscale dynamics and microphysics, which are not predictable by using quasi-steady state conditions. Therefore, the multiscale aspect of cirrus formation is better represented by transient simulations. This could be verified by comparing transient and non-transient simulations of orographic cirrus clouds with aircraft measurement data.

5.2 Discussion and Outlook

5.2.1 Cirrus clouds triggered by radiation

In the cirrus community, it was assumed as a kind of textbook knowledge that radiation might not be important for the initial formation of cirrus clouds (see e.g. Kärcher and Spichtinger, 2009). Under the comprehension of the discussed destabilization process this position should be revised. In the presence of weakly stable profiles both effects, i.e. radiative cooling and destabilization might lead to the formation of visible cirrus clouds. On the other hand, the impact of radiation on the stability of the upper troposphere itself, discussed in a broader sense, should be an interesting topic for future research. From this point of view the role of convective cells in ice-supersaturated regions and cirrus cloud layers might be interesting in terms of cirrus cloud inhomogeneities and patchiness of cirrus clouds, but also in terms of the radiative impact of cirrus clouds. Finally, this could also be important for more physically based parameterizations of cirrus clouds in large-scale models, including the macroscopic structure on the cloud scale.

We could not discuss in detail the issue of frequency of occurrence of environmental conditions, which allow the radiation to have a predominant impact on the stability of the upper troposphere and therefore be of importance for cirrus formation. The used radiosonde data originates from only one measurement...
5.2. Discussion and Outlook

site and therefore does not give an insight into global distributions. This should be investigated in future studies, using meteorological analyses and possibly the output from large-scale models in order to obtain a better overview on the importance of the described mechanism. It also remains unclear how important radiative destabilization is when superimposed with other large- or meso-scale processes (i.e. frontal lifting, gravity waves, etc.); this will be subject of future research.

5.2.2 Cirrus clouds triggered by transient large scale dynamics

A very common question that arises while performing any kind of simulations of atmospheric flows is how representative the results are. Generally, simulations can be validated by comparing them to measurements or other reference models. A combination of both methods was performed in a recent study by Doyle et al. (2010), where the results of nine different models were compared to measurements from the T-REX campaign. The presented results suggest that EULAG provides consistent and stable solutions for the various wave regimes. A direct comparison to aircraft measurement data was discussed in chapter 4, with simulation results that correlate well to the measurements. However, as only few measurements of orographic cirrus clouds exist, it is sometimes necessary to broaden ones horizon to get more opportunities to compare and validate the simulations. In our case, we could look beyond the tropopause and enter the stratospheric realm. In a study by Reichardt et al. (2004), polar stratospheric clouds (PSC) over a mountain were investigated with a ground-based lidar instrument (i.e. on the leeward side). In this case, a PSC is formed due to vertically propagating orographic waves that generate elongated stratospheric temperature anomalies parallel to the mountain ridge. During the observation, the horizontal flow (perpendicular to the ridge) increased with time. Subsequently, the lidar observed the PSC increasing its altitude and moving downstream for several hours due to a displacement of the temperature anomaly (see figure 5.1, lower panel). These horizontal and vertical displacement processes could be explained by transient orographic waves, as we have seen in our study. Our simulations (figure 5.1, upper panel; corresponds to the increasing wind case in chapter 3) show a similar vertical displacement of wave packets (i.e. adjustment to a longer wavelength) due to an increased horizontal flow leading to an upward propagation of the updraft regions, where clouds should be formed. These processes can not be observed by using a model that does not consider time dependent large scale dynamics.

Taking into account that the amount of orographic cirrus clouds over continents is quite substantial (Dean et al., 2005), their formation should be treated as accurate as possible. However, most GCMs handle orographic waves as stationary waves using linear theory to describe their temperature perturbations
Figure 5.1: Upper panel: Vertical cross section of vertical velocity at the leeward side of a mountain, showing vertical altering of up- and downdraft regions due to an increased horizontal flow. The contour line level steps are 0.5 m/s; lower panel: Temporal evolution of 355-nm backscatter ratio R of a polar stratospheric cloud (PSC) on 16th January 1997 over Esrange (Reichardt et al., 2004).

(Dean et al., 2007). The global climate model ECHAM5 for example assumes an upward motion of the whole column over a mountain to calculate the amount of orographic clouds (Joos et al., 2008). But one has to consider that beside regions of cloud formation (updraft), an orographic wave also generates regions that evaporate clouds (downdraft). The location of down- and updrafts is mainly determined by the vertical wavelength of the wave. The importance of the interference between updraft and downdraft regions and ISSRs on cloud formation has been discussed in chapter 3. Therefore, I recommend that the vertical velocity should depend preferably on the wave phase in current and future GCMs. On the other hand, none of the transient wave processes discussed in this thesis is considered in current GCMs. A more physically based parameterization of orographic cirrus clouds in large-scale models should include the following processes:

1. The rate of change of the ambient flow to derive the horizontal displacement and lifetime of pre-existing clouds.

2. The time, a new wave needs to propagate to an ISSR must be taken into account. This effect is inversely proportional to the wavelength.
3. The impact of the new vertical wavelength on the overlapping of the updraft and downdraft regions with an ISSR. This effect is not instantaneous, thus the wave propagation time must be considered. As a first estimate, a more realistic treatment of transient orographic waves would increase the orographic cirrus cloud cover due to the increased horizontal variability of cloud ice. Please regard that in this thesis, only the impact of different transient processes on various scales on cirrus cloud formation has been investigated and discussed. Nevertheless, the feedback of transient orographic waves back on the large scale (i.e. impact on wave drag, potential vorticity, etc.) is also very important. This process is discussed by Chen et al. (2006) and is subject of current investigations.

In chapter 4, a time dependent potential temperature $\theta_e$ was used, but our scenarios did not lead to a significant change of the potential stability of the atmosphere. Thus, future studies could investigate the impact of changing atmospheric stability on orographic waves and the formation and evolution of cirrus clouds in more detail.
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<td>$C_{gc}$</td>
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<tr>
<td>COESA</td>
<td>US Commission/Stand Atmosphere (Collaborators: NASA, NOAA, USAF)</td>
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<tr>
<td>CVI</td>
<td>Counterflow Virtual Impactor</td>
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<td>DFG</td>
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<td>DLR</td>
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<td>ECHAM5</td>
<td>global climate model, Version of MPI Hamburg</td>
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<td>ECMWF</td>
<td>European Centre for Medium-Range Weather Forecasts</td>
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<tr>
<td>EULAG</td>
<td>EULERian, semi-LAGrangian model for fluids</td>
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<td>GCM</td>
<td>General Circulation Model</td>
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<tr>
<td>INCA</td>
<td>Interhemispheric differences in cirrus properties from anthropogenic emissions</td>
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<td>ISSR</td>
<td>Ice Supersaturated Region</td>
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<td>LW</td>
<td>longwave</td>
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<td>NASA</td>
<td>National Aeronautics &amp; Space Administration</td>
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<td>NCAR</td>
<td>National Center for Atmospheric Research</td>
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<td>NOAA</td>
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<td>NGDC</td>
<td>National Geographical Data Center</td>
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<td>PSC</td>
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<td>SNSF</td>
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<td>SW</td>
<td>shortwave</td>
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<td>TOA</td>
<td>top of atmosphere</td>
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<tr>
<td>USAF</td>
<td>United States Air Force</td>
</tr>
<tr>
<td>UTC</td>
<td>coordinated universal time</td>
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5.1 Upper panel: Vertical cross section of vertical velocity at the leeward side of a mountain, showing vertical altering of up- and downdraft regions due to an increased horizontal flow. The contour line level steps are 0.5 m/s; lower panel: Temporal evolution of 355-nm backscatter ratio *R* of a polar stratospheric cloud (PSC) on 16th January 1997 over Esrange (Reichardt et al., 2004).
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Curriculum Vitae

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**PUBLICATIONS**


Danksagung

Ich möchte mich als erstes ganz herzlich bei Prof. Dr. Ulrike Lohmann bedanken, dass sie mir die Möglichkeit gegeben hat, die letzten 3 Jahre als Doktorand in ihrer Gruppe zu arbeiten. Die diversen Gespräche sowie ihre Hinweise darauf, dass man ab und zu auch über den Tellerrand des eigenen Fachgebiets schauen muss, haben mir neue, hilfreiche Inputs für meine Arbeit gegeben.

An Prof. Dr. Peter Spichtinger geht ein weiteres riesengrosses Dankeschön. Er war für mich während der letzten Jahre von unschätzbarem Wert, als äusserst fachkompetenter wie auch geduldiger Lehrer und Mentor. Seine Unterstützung hat massgebend zum Gelingen dieser Arbeit beitragen. Es war eine ganz grosse Freude, mit ihm zusammen arbeiten zu dürfen! Prof. Dr. Ulrich Achatz möchte ich herzlich danken, dass er sich die Zeit und die Mühe nimmt, meine Arbeit zu begutachten. Vielen Dank auch an Dr. Andreas Dörnbrack für die gute Unterstützung bei Modell-Fragen sowie für das Einführen in die Genüsse der Bayrischen Küche.

Außerdem möchte ich mich bei allen anderen PostDocs, Doktoranden und Mitarbeitern des Instituts bedanken, welche mich auf dem steinigen Weg meiner Dissertation begleitet haben. Ein besonderer Dank geht hierbei an Dr. Hanna Joos für ihre Unterstützung und Zusammenarbeit, an Dr. Dani Lüthi für konstant funktionsfähige Linux Systeme sowie Datenautobahnen zum ECMWF, an Eva Choffat für ihre Navigation durch den administrativen Dschungel einer Hochschule, an Peter Isler für den technischen Support sowie für die diversen exzellenten Aperos und an die Kaffee Maschine im P-Stock für die stetige und meist zuverlässig gewährleistete Koffeinversorgung. Zusätzlich bedanke ich mich bei allen Bürokrämeraden/-innen für die tolle Zeit (besonders Francesco hat durch seinen Tessiner Flair für gute Stimmung gesorgt)!


Diese Arbeit ist meinem Onkel Bruno gewidmet, welcher zu Beginn meines Doktorats so plötzlich und unerwartet verstorben ist.