The diurnal cycle of midlatitude, summertime moist convection over land in an idealized cloud-resolving model

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LINDA BABETTE RICARDA SCHLEMMER
Dipl. Phys., University Hamburg
born 9 March, 1983
citizen of Germany
accepted on the recommendation of
Prof. Christoph Schär, examiner
Prof. Björn Stevens, co-examiner
Dr. Jürg Schmidli, co-examiner

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Summertime moist convection over land is an essential process of the mid-latitude climate due to its impact on the water and energy cycles. The underlying processes are very complex and difficult to understand resulting from the associated diurnal cycle and the non-linear and multiscale nature of convection. These difficulties impede the development of credible climate change scenarios, for instance regarding projections of the European summer climate with potentially significant shifts in temperature variability, precipitation extremes and droughts. Only in recent years have numerical weather and regional climate prediction models undergone increases of the grid resolution employed to the cloud-resolving kilometer-scale allowing an explicit resolution of convective processes. Simulations over time scales of decades and spatial scales of $O(10,000 \text{ km}^2)$ are nevertheless still computationally too expensive, but will become feasible within a few years.

This thesis aims at investigating basic characteristics of midlatitude summertime moist convection within an idealized context. To this end, an idealized cloud-resolving modeling framework is developed by modifying an existing regional weather forecasting and climate model. The unique features of the framework developed are, firstly, the use of the full set of physical parameterizations, allowing for the complete range of soil-atmosphere interactions in an otherwise highly idealized framework and secondly, the relaxation of predicted variables towards prescribed atmospheric and soil profiles with a weak (strong) relaxation in the lower (upper) troposphere and upper (lower) soil, so as to allow the development of a distinct diurnal boundary layer. The model is run over several diurnal cycles into a state of "diurnal equilibrium". In this state the diurnal cycle of moist convection approximately repeats itself every day, with growth of the boundary layer, development of clouds, triggering of deep convection, formation of precipitation, and subsequent decay of cloud structures. The framework is used to explore the sensitivity of the diurnal convection to changes in key environmental conditions.

In chapter 2, the roles of the moisture content and the static stability of the atmosphere on the diurnal convection are investigated. It is found that the moisture content
has negligible influence on the equilibrium convection, as moisture is redistributed in
the atmosphere by convective and boundary-layer activity and thereby controlled by the
convection itself. The static stability determines the depth of the evolving convection
and the timing of the precipitation onset and peak. Precipitation amounts are to a large
degree controlled by the amount of water evaporated from the surface into the atmos-
phere and almost independent of the static stability.

In chapter 3, the role of the soil water content as a control parameter on evapo-
ration is investigated. In our idealized framework, an amplification of soil water always
leads to an enhancement of evaporation that in turn increases precipitation. This im-
plies a positive soil moisture-precipitation feedback. For unstably stratified atmospheres
this positive feedback is strongest. For stably stratified atmospheres, more clouds de-
velop over wetter soils, reducing incoming radiation. This reduction is, however, offset
by increases in longwave radiation, leading to a near-constant net surface radiation
across experiments with increased soil moisture. Associated with the daily triggering of
deep convection for all cases considered, moisture is effectively transported out of the
boundary layer, increasing the near-surface saturation deficit, and thereby counteract-
ing a potential negative soil moisture-precipitation feedback.

Having gained insights into key aspects of present-day convection, the developed
framework is in chapter 4 applied to study, in a process-based manner, the response of
convection to changes in ambient temperature, that could e.g. result from anthropogenic
climate change. In addition, consideration is given to a stabilization of the atmosphere
that could for instance originate in a stronger upper-tropospheric warming, and to a
destabilization from a summer-drying. The idealized simulations yield an increase of
near-surface temperatures and cloud cover in a warmer climate. Increases of high pre-
cipitation intensities are observed for warmer temperatures, which are stronger if a sta-
bilization of the atmosphere is included. Over drier soils, increases in high precipitation
intensities observed for warmer climates are reduced. At the same time, decreases of
precipitation amounts at low intensities are simulated for drier soil. The results from our
idealized, cloud-resolving simulations are in accordance with theoretical considerations
which could so far not be fully reproduced in models using parameterized convection.

Overall, the developed framework gives deeper insights into summertime mid-latitude
diurnal convection over land. Its idealized design simultaneously eases its application
and its interpretation. It is transferable to a vast number of further sensitivity studies as
well as portable to different regions and seasons.
Zusammenfassung


bedeutenden Umgebungsbedingungen zu erforschen.


In Kapitel 3 wird die Bedeutung der Bodenfeuchte als Kontrollparameter für Verdunstung untersucht. In unserem idealisierten Modellierungsrahmen führt eine Zunahme des Bodenwassers immer zu einer Zunahme der Verdunstung, was wiederum den Niederschlag erhöht. Es existiert demnach eine positive Rückkopplung zwischen Bodenfeuchte und Niederschlag. Für instabil geschichtete Atmosphären ist diese Rückkopplung am stärksten. Für stabil geschichtete Atmosphären entwickeln sich mehr Wolken über feuchteren Böden, welche die einfallende Strahlung reduzieren. Diese Reduktion wird jedoch durch eine Zunahme der langwelligen Strahlung kompensiert, was zu einer beinahe konstanten Nettooberflächenstrahlung in Experimenten mit vergrößelter Bodenfeuchte führt. Zusammen mit der täglichen Auslösung von tiefer Konvektion in allen betrachteten Fällen wird die Feuchtigkeit effektiv aus der Grenzschicht heraustransportiert, womit einer potentiell möglichen negativen Rückkopplung zwischen Bodenfeuchte und Niederschlag entgegengewirkt wird.


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Chapter 1

Introduction

Moist convection plays a key role in weather and climate through its impact on the water and energy balance of the Earth. Convection is the main physical process that generates kinetic energy in the atmosphere (Steinheimer et al., 2008) and latent heat of water vapor accounts for most of the upward heat transport in the atmosphere. Convection transports water vapor vertically in the atmosphere which is a crucial process in the hydrological cycle as it links its terrestrial and atmospheric branch (Peixoto and Oort, 1992). Through formation of precipitation over the continents it is essential for life on Earth as a source for drinking water.

First, the phenomenon of moist convection will be introduced with a focus on mid-latitude convection over land. Next, different approaches to numerically modeling convection are presented. Projected changes of the hydrological cycle in climate change are reviewed. Finally, an outline of the thesis will be presented.

1.1 The nature of moist convection

A short phenomenological summary focusing on the most important processes for the current study is given here. An in depth description of moist convection can e.g. be found in chapter 5 of Houze (1993), Emanuel (1994), Stevens (2005), Bechtold (2008) or Cotton et al. (2011).

Common to all forms of moist convection is that some specific air parcels are warmer and lighter than their surrounding and thus start to rise. The rising air parcel cools as it expands and at some point reaches the same temperature as its environment and it slows down. After overshooting it returns to its level of neutral buoyancy (LNB). If moisture is contained in the raising parcel it eventually reaches saturation and releases latent heat of condensation, fueling the parcel with additional heat for further ascent. Even more energy is available as soon as freezing sets in and latent heat of freezing is released. Rising parcels will however not reach their LNB undisturbed but undergo mixing with surrounding air. In this entrainment process dry air is mixed into the ascending parcel, slowing it down. Likewise moist air is detrained into the surrounding air moistening the environment. At the tropopause level undiluted parcels originating from the planetary boundary layer (PBL) are virtually absent (Romps and Kuang, 2010). Clouds with a larger horizontal extent experience less lateral entrainment and convection is able to become deeper Kuang and Bretherton (2006). Buoyant parcels rising undiluted in so-called “hot towers” are a key element for the tropical heat balance and also play a major role in the intensification of tropical cyclones (e.g. Guimond et al., 2010).
1 Introduction

Figure 1.1: Meteosat image showing tropical convection both over land and ocean over the Southern Atlantic and southern Africa, a tropical cyclone east of Madagascar, stratocumulus west of the coast of Westafrika and convection in mid-latitude weather systems both in the South and North Atlantic.

If the rate of temperature decrease with height is large, the warm, rising parcel has a temperature that is considerably larger than its environment and experiences thus large buoyancy forces. In an atmosphere, where temperature decreases slowly with height the buoyancy forces are smaller and the atmosphere is more stable towards occurrence of convection.

After some time of ongoing moist convection condensed water droplets have acquired a large mass, their gravity force becomes larger than the friction forces holding them in the air and precipitation sets in. Above the freezing level graupel and snow forms. A large fraction of the falling hydrometeors evaporates again when falling through relatively dry air, thereby consuming latent heat of evaporation and cooling the environment. Particles stemming from the ice phase additionally consume latent heat of freezing. This cooled air falls towards the ground, where it can be recognized as cold downdrafts. The cold air spreads along the surface and leads to an uplift of unstable air in the surrounding, thereby triggering new convection (e.g. Simpson, 1980). As a result convection organizes on spatial scales on the order of 100 km. Convective organization plays a fundamental role in the formation of so-called “squall-lines" or mesoscale convective systems. Wind shear (e.g. Rotunno et al., 1988), atmospheric moisture (e.g. Tompkins,
1.1 The nature of moist convection

2001), sea-surface temperature (e.g. Graham and Barnett, 1987) and cloud-radiative interactions (e.g. Sherwood, 1999, Bergman and Hendon, 2000) furthermore play an important role in the organization of convection.

Both the heating of the upper troposphere in convective updrafts and the cooling of the lower tropopause by evaporation of precipitation decrease the lapse rate of temperature, whereby the atmosphere is stabilized. If the generation of convective available potential energy (CAPE) by large-scale processes is approximately balanced by the consumption by convection, the convection is said to be in equilibrium (Arakawa and Schubert, 1974, Emanuel et al., 1994). In this case, CAPE remains small. In the case of non-equilibrium convection, CAPE builds up over extended time, followed by a violent outbreak of convection (Done et al., 2006).

Convection can be found all over the globe in a variety of appearances and scales: at low latitudes in all seasons and at mid-latitudes over land in the warm season and over ocean (especially as cold continental air blows over warmer water in “cold-air out-breaks”) in winter (Bretherton, 2007). It comprises of the smallest drops and crystals over turbulence and clouds to its organization in mesoscale and synoptic-scale dynamical systems. Fig. 1.1 shows a Meteosat image of the globe containing both tropical and mid-latitude convection over land and water surfaces. Two different modes of convection should be distinguished:

**shallow convection**: shallow vertical extent (≤ 250 hPa), mostly non-precipitating, typically reaching the boundary-layer inversion depth

**deep convection**: intense precipitation, ice-processes are involved, possible thunder and lightning, commonly reaching the tropopause.

with a range of types occurring in-between those extremes.

Due to the difference in appearance of tropical and mid-latitude convection they will be considered separately in the following. Tropical convection is mentioned, even if the focus of this thesis is on mid-latitude convection. There are nevertheless numerous similarities between them and also knowing the differences is important.

1.1.1 Tropical Convection

Most of the convection occurs in the tropical belt where also major amounts of the global precipitation are observed. The most prominent meridional circulation is the Hadley cell. Its ascending branch is situated in the low pressure zone of the Inter-Tropical Convergence Zone (ITCZ), that is typically situated between 5 and 10° N over the Oceans but changes it location in the seasonal cycle over land-masses where it follows the zenith
angle of the sun. The Hadley cell plays a key role in the global circulation through its transport of heat from the tropics to mid-latitudes. The descending branch of the ITCZ is situated over the subtropics, where over the cold sub-tropical oceans stratocumulus forms. In the returning branch between the subtropics and the ITCZ, trade-wind cumuli prevail. These trade-wind cumuli are essential in the uptake of moisture from the sea-surface into the atmosphere to provide the moisture required to fuel the deep convection to occur in the ITCZ. The uptake of moisture by shallow convection over the subtropical oceans in the returning branch of the Hadley circulation actually constitutes the main source of water vapor for the atmosphere (Peixoto and Oort, 1992).

Tropical convection over the oceans is driven by the radiative cooling of the atmosphere which yields together with relatively constant sea-surface temperatures an atmosphere that is unstable to the displacement of moist air. The diurnal cycle of oceanic convection exhibits its maximum in the early morning, in contrast to continental convection that generally peaks in the late afternoon or evening (e.g. Hendon and Woodberry, 1993). Convection over oceans furthermore shows a considerably smaller amplitude of the diurnal cycle than over land. The heat capacity of land surfaces is smaller than of oceans yielding a stronger heating of the ground over the course of the day.

A strong departure from zonal symmetry of convection in the tropical region is observed. Convection tends to occur in organized systems such as African easterly waves (see e.g. Mohr and Thorncroft, 2006) or equatorially-trapped waves. Among the latter there is a variety of types, for which the theoretical basis has first been derived by Matsuno (1966). The most prominent representative of equatorially trapped waves is the Madden-Julian Oscillation (MJO Madden and Julian, 1971, 1972), a traveling pattern over the warm parts of the Indian and Pacific oceans. It results from the coupling of a Rossby wave and a Kelvin wave and manifests itself in an interplay of convection and sea-surface temperatures leading to anomalously high precipitation with a period of 30-60 days (e.g. Zhang, 2005).

1.1.2 Mid-latitude moist convection
Convection occurring in mid-latitudes is more limited in time and space than in the tropics. The fundamental difference between tropical- and mid-latitude weather systems is the increase of Coriolis-forces towards higher latitudes. The Rossby radius of deformation ($R_o$, the radius of influence over which perturbations spread) is accordingly considerably smaller ($\mathcal{O}(1\ 000\ km)$) for mid-latitudes than in the tropics (where it tends to infinity). Temperature gradients between different air masses are hence notably larger. As a result, convection often occurs together with synoptic eddies, where
the large-scale lifting of air e.g. along cold fronts triggers convection. If a destabilization of the atmosphere occurs by synoptic-scale systems, convection can provoke or enforce heavy-precipitation events (e.g. Doswell et al., 1996). This forced or triggered convection shows a fundamentally different picture than air-mass convection in flat-pressure gradient synoptic situations, where convection occurs randomly along small convergence lines.

The focus of this thesis is on events falling into the second category. In summertime under weak synoptic forcing, periods of non-equilibrium convection are frequently observed. Under these flat-pressure distributions the sun warms the surface, enabling latent and sensible heat fluxes that lead to a moistening and growth of the PBL. Buoyant parcels rise from the surface and get diluted by entrainment. At some point the growing PBL will reach the height of the lifted condensation level (LCL) and the first shallow-convective clouds form. The moistening of the air by detrainment continues and lowers the level of free convection (LFC) until the depth of the PBL coincides with the LFC and deep convection occurs. Until the atmosphere is sufficiently moistened a considerable amount of time passes. Therefore, the peak of convection is typically shifted by many hours with respect to the peak in surface heating (e.g. Bechtold et al., 2004). The picture can be substantially altered if large-scale advection is present that modifies the moisture content of the atmosphere, or in the presence of water surfaces or topography, that induce local thermal circulations. Triggering of deep convection is often advanced over mountain slopes, that are heated by the sun.

Observational studies of the diurnal cycle of convection over land in mid-latitudes have for instance been performed at the ARM site in the U.S. and during the COPS campaign in southern Germany. Modeling studies, partly connected to those observations have e.g. been performed by Xu et al. (2002), Bechtold et al. (2004), Chaboureau et al. (2004), Guichard et al. (2004).

The transition from shallow to deep convection is regulated by the vertical stability and the humidity content of the atmosphere. The humidity content regulates how dry the air is, that is entrained into ascending plumes and thereby the time necessary to sufficiently premoisten the atmosphere for deep convection to be triggered (Chaboureau et al., 2004, Derbyshire et al., 2004, Kuang and Bretherton, 2006, Wu et al., 2009). The static stability determines the buoyancy force acting on unstable plumes, the height of the LFC and LNB and thereby the depth of the evolving convection (Houston and Niyogi, 2007, Wu et al., 2009, Zhang and Klein, 2010). Therefore the atmospheric composition has a strong influence on the timing of convection.
It is not only the state of the atmosphere that regulates moist convection but also the condition of the soil, both on time-scales of single days, and climatic ranges. Over short timescales the amount of soil moisture has a significant influence on the partitioning of the surface heat fluxes and thereby alters the properties of the PBL. Numerous different interactions among the land-surface and boundary layer are observed that either increase or decrease the amount of clouds at the PBL top and later of precipitation. Concerning longer timescales the soil-moisture precipitation feedback can adopt different magnitudes and signs, depending on the soil and vegetation characteristics, the state of the atmosphere and the amount of incoming radiation. A positive feedback is thought to have major impacts on summertime droughts and heatwaves (e.g. Fischer et al., 2007a). Studies on the land-surface atmosphere coupling have however mostly focused on individual processes over short timescales (e.g. Findell and Eltahir, 2003a,b, Ek and Holtslag, 2004, van Heerwarden et al., 2009), case studies (e.g. Hohenegger et al., 2009), or longer-term simulations with global climate models using a coarse resolution of the computational grid (e.g. Koster et al., 2004, Cook et al., 2006).

Moist convection over land is hence subject to a variety of different mechanisms all interacting among themselves. Observational studies that aim to improve our understanding of them are limited in time and space. Therefore numerical modeling of the manifold complex processes has gained more and more importance. The most common concepts of the numerical modeling of moist convection are hence summarized in the following.

### 1.2 Modeling of moist convection

The numerical representation of convection and its influence on the environment depends critically on the mesh size of the numerical model used. Fig. 1.2 gives an overview over the scales at which selected atmospheric processes act. Organized convection is situated at spatial scales of 1-100 km. If the grid-size is larger than the relevant scales of the process itself a parameterization needs to be employed, if the grid-size is smaller (typically $O(1 \text{ km})$) the process can be simulated explicitly.

#### 1.2.1 Parameterization of moist convection

The aim of the parameterization scheme is to determine the effect of an ensemble of convective clouds on the heat- and moisture budget of the surrounding as a function of grid-scale variables. These terms are often called the apparent heat source $Q_1$ and the apparent moisture sink $Q_2$. Some schemes additionally take into account the ap-
Figure 1.2: Overview over the time- and spatial scales of selected atmospheric phenomena. Figure taken from Fuhrer (2005), after Orlanski (1975).

Parent momentum source $Q_3$. Typically a scheme acts to: (1) determine if and where convection occurs (trigger function), (2) determine the vertical heating, moistening and momentum change profile (tendency function) and (3) determine the intensity of the ongoing convection (closure).

Sub-grid convective activity was first included into the global models of Manabe and Strickler (1964), Kuo (1965) or Ooyama (1971). Today there is a huge variety of different schemes (see e.g. Stensrud, 2007). Only a few important ones acting in different ways are mentioned here: adjustment schemes (e.g. Manabe et al., 1965) that adjust the atmospheric state back to a reference profile if the atmosphere is unstable to a parcel lifted from the boundary layer and if there is a deep moist layer; mass-flux schemes (e.g. Tiedtke, 1989, Kain and Fritsch, 1990) that determine the magnitude of convective heat-release by the mass-flux at the cloud base. A cloud-model is then used to distribute the heat vertically. The cloud-model can be a spectral model (used by e.g. Arakawa and Schubert, 1974) or a bulk cloud model (used by e.g. Tiedtke, 1989). Furthermore there are different ways to “close” the scheme, e.g. by relating the cloud-base mass-flux to the sub-cloud moisture convergence (e.g. Tiedtke, 1989) or to the CAPE (e.g. Kain and Fritsch, 1990).

Convective parameterizations have however well-known problems in realistically representing today’s climate (e.g. Yang and Slingo, 2001, Arakawa, 2004, Hohenegger et al., 2008b). On weather timescales, convection is considered the prime reason for the large
difficulties with warm-season short-term quantitative precipitation forecasting (Fritsch and Carbone, 2004). On climate timescales, the inability of current climate models to represent organized tropical convection is viewed as a major stumbling block in providing confident projections of future climates (Yang and Slingo, 2001, Randall et al., 2003, Arakawa, 2004, Slingo et al., 2009). For the diurnal cycle of convection over land they have problems resembling the delay between the maximum in heat fluxes and precipitation (Bechtold et al., 2004). Even for key aggregate parameters such as climate sensitivity, cloud feedbacks are the largest source of inter-model differences (Bony et al., 2006). One therefore wishes to omit parameterizations of convection by explicitly modeling it.

1.2.2 Explicit simulations of convection

Early work to explicitly simulate moist convection reaches back into the late 1950s. The first 1-dimensional numerical simulations of moist convection were performed by Malkus and Witt (1959). Simpson et al. (1965) later performed simulations with a very simple 1 dimensional entraining bubble model. Two-dimensional anelastic cloud-model were introduced in the early 60s and were either run using axial symmetry (r-z) (e.g. Ogura, 1962), or slab symmetry (x-z) (e.g. Lilly, 1962). As soon as enough computational power was available the first three-dimensional model was developed by Steiner (1973). Numerical studies using 3D geometry to study for instance the development of super-cell thunderstorms (see e.g. Klemp and Wilhelmson, 1978a,b, Wilhelmson and Klemp, 1978) were performed later on.

The advance in computer power allowed to increase the resolution of numerical weather prediction (NWP) models and regional climate models (RCMs) until they are now acting in a similar domain where research cloud-models have been operating. For limited-area models it has become possible to perform simulations at grid-spacings in the order of 1-4 km. This increase in resolution demanded furthermore to relax the hydrostatic assumption and to improve the dynamical cores of the models. These models acting on scales of $O$(1 km) are often termed “cloud-resolving” models. This does not mean that single clouds are simulated adequately but rather that the bulk heat- moisture- and momentum fluxes of deep, organized convection are adequately represented (Weisman et al., 1997). An improvement of the quality of weather forecasts going to higher resolution could be stated (Mass et al., 2002, Done et al., 2004). Moreover CRMs start to become feasible for climate simulations. Hohenegger et al. (2008b) have for instance applied a CRM on a monthly timescale. These simulations showed very encouraging results. The overall precipitation distribution and evolution was well captured and im-
proved as compared to simulations using parameterized convection.

CRMs are still expensive to run and therefore mostly restricted to limited areas. A month-long integration using a mesh size of 3.5 km spanning the whole globe has however been performed by Satoh et al. (2008). A novel hybrid method to parameterize convection in global model is the so-called “superparameterization” approach, where the effects of convective clouds is simulated by 2D CRMs that are embedded within each grid-cell of the global model. Such attempts have e.g. been developed and used by Grabowski and Smolarkiewicz (1999), Grabowski (2001), Khairoutdinov and Randall (2001). Using superparameterizations the MJO can for example be reproduced more realistically (Randall et al., 2003).

Another class of models used to explicitly simulate the development of clouds and (mostly shallow) convection are large-eddy simulations (LES) (e.g. Siebesma et al., 2003, Stevens et al., 2005) which have their roots in the turbulence-modeling community. LES models typically employ a grid-spacing of $O(10\text{-}100 \text{ m})$ and resolve the largest eddies of a three-dimensional turbulent field explicitly. Smaller sub-grid eddies can then be derived from the resolved eddies by use of the known scaling behaviour in the inertial subrange (Kolmogorov theory). LES simulations of deep convection have only become possible in the last few years (Petch et al., 2002, Bryan et al., 2003, Khairoutdinov and Randall, 2006, Moeng et al., 2009).

1.3 Future projections of the hydrological cycle

The occurrence of a human-induced climate change is nowadays widely accepted in the scientific community, especially after the publication of the fourth assessment report (AR4) of the Intergovernmental Panel on Climate Change (IPCC, 2007b). Theoretical considerations claim that the increase of tropospheric temperatures goes together with an increase of the water-holding capacity of the atmosphere due to the Clausius-Clapeyron relation at roughly 7 % per Kelvin and a near-constant relative humidity (e.g. Trenberth, 1999). This increase of water in the atmosphere is projected to lead to an intensification of the hydrological cycle (e.g. Allen and Ingram, 2002) with increases of precipitation extremes leading to societal impacts such as increased flooding or soil erosion (IPCC, 2007a).

Future projections of precipitation predict an increase of precipitation on the global scale and an increase in precipitation extremes that is greater than changes in mean precipitation (Kharin and Zwiers, 2005). For mid-latitude land areas during summertime a decrease of mean precipitation amounts and at the same time an enhancement of in-
tense precipitation events are projected (Christensen and Christensen, 2003, Frei et al., 2006, Beniston et al., 2007). The raise of intense events can be understood by the increase in atmospheric moisture content by the non-linearity involved in the Clausius-Clapeyron relation. Observational studies show that short-term extreme precipitation events scale even stronger than the Clausius-Clapeyron relation (Lenderink and van Meijgaard, 2008, Allan et al., 2010). It is however not clear if this increase is caused by a transition from large-scale, stratiform to more convective precipitation (Haerter and Berg, 2009) or by increased latent heat release in storms as the temperature rises (Lenderink and van Meijgaard, 2009, 2010). Studies indicate that this amplification of the precipitation extremes is strongly underestimated in climate models (Allan and Soden, 2008).

Regarding the extreme precipitation events much larger changes are expected in their recurrence frequency than in their magnitude (Huntingford et al., 2006, Barnett et al., 2006, Frei et al., 2006).

For global mean precipitation on the other hand, it is the energy budget of the atmosphere, or more precisely the atmospheric radiative cooling (e.g. Allen and Ingram, 2002, Allan, 2006) or the surface energy balance (Wild et al., 2008) that constrain precipitation increases. Thus we expect mean precipitation amounts to increase at a slower rate than anticipated from the Clausius Clapeyron equation. An increase of temperature furthermore enhances evaporation from the surface thereby increasing the risk of droughts.

The regional trends of precipitation are however greatly modified by advection that redistributes evaporated moisture in the atmosphere, and the availability of soil moisture to evaporate. An increase of the potential evaporation and a resulting increase in evapotranspiration (ET) are further effects of warmer temperatures. This is projected to lead to a drying of the soil caused by the lack of available soil moisture. An increased frequency of droughts are anticipated to be the consequence (e.g. Schär et al., 2004, Seneviratne et al., 2006, Fischer et al., 2007a, Sheffield and Wood, 2008). Following Rowell and Jones (2006) thermodynamic effects such as an earlier and more rapid decline of soil moisture in spring caused by an earlier snowmelt and a positive soil moisture-precipitation feedback in summer are important processes that dry out the soil.

Fractional changes of water vapor are larger in the upper troposphere, where it is dry, than in the lower troposphere. Due to the strong positive water vapor feedback (e.g. Soden et al., 2005) this results in a non-uniform warming over the troposphere with
larger increases in the upper troposphere. Thereby, the atmospheric profile is stabilized. A more stable atmospheric profile provides a stronger radiative cooling to space. This lapse rate feedback is thought to be slightly negative (e.g. Colman, 2003, Soden and Held, 2006, Dessler and Sherwood, 2009). The combination of radiative and lapse rate feedback is however positive amplifying the global mean temperature response by a factor of 40-50% (Bony et al., 2006). A further effect influencing convection is a projected rise of the tropopause height (IPCC, 2007b, chapter 9.4.4.2).

The dynamical effects of climate change due to circulation-type changes are more difficult to assess (e.g. Ulbrich et al., 2009). Considering the European summer climate, van Ulden and van Oldenborgh (2006) found a strong impact of changes in the flow direction on precipitation amounts in global climate models, whereas Rowell and Jones (2006) identified thermodynamic factors as the main cause for decreases in summer precipitation in continental and southeastern Europe. Regarding extreme events, the picture is even more diffuse. In many of the most devastating flooding events in recent years (e.g. the 1997 Oder flooding (Fuchs and Rapp, 1998), the 2002 Elbe flooding (e.g. Ulbrich et al., 2003) and the 2005 Alpine flooding (e.g. Hohenegger et al., 2008a, Langhans et al., 2011))) anomalous circulation patterns of a \( \nabla b \) type cyclone played a crucial role controlling the moisture convergence. Beniston (2006) however argues that the 2005 Alpine flood was not representative for a “greenhouse climate”.

Modeling studies projecting future impacts of an intensified hydrological cycle are difficult, as different sources of uncertainty need to be accounted for. The future development of greenhouse gases and aerosols (Nakićenović and Swart, 2000) or land-use change (e.g. Feddema et al., 2005) comprise ambiguities, the processes of the climate system are not correctly representable in numerical models (e.g. Palmer et al., 2005) and internal climate variability (e.g. Räisänen, 2001) imposes uncertainties. Hawkins and Sutton (2009) demonstrated that for lead times of decades the governing uncertainties for the prediction of temperature are internal variability and model uncertainty, where the former becomes more important for smaller spatial scales.

On the side of model uncertainties a major deficiency of most current climate prediction models is the use of parameterization schemes to predict convection. As described above, these schemes have shortcomings even in present-day climate. The credibility of precipitation projections are therefore limited. Soden and Held (2006) stated that cloud feedbacks provide the largest source of uncertainty in current predictions of climate sensitivity. Going to regional scales uncertainties of predictions generally increase. A set of RCMs used in the PRUDENCE project (Vidale et al., 2007) showed
for instance large differences in simulated summer-climate between different models, when synoptic scale forcing is weak and the model physics have a great influence. Simulations using CRMs are however still expensive and not yet affordable for climate simulations over decades. RCMs of mesoscale extent furthermore rely on lateral boundary data that stems from a driving model using a parameterization for convection. As a result a future projection of convective summer precipitation is still an issue discussed without reliable outcomes.

1.4 Aims and outline of this study

The aim of this PhD thesis is to understand the key parameters that determine the characteristics of moist convection in the European summer climate and to investigate the sensitivity of these parameters to changes in future climates. To this end an idealized cloud-resolving modeling framework is introduced. As stated above we believe that the results obtained by explicitly modeling convection are more reliable than results stemming from a parameterization scheme for convection.

The outline of the thesis is as follows:

1. chapter 2: In a first step the framework and the model used are introduced. The control simulations of the diurnal cycle of summertime moist convection over land is evaluated. On this reference simulation systematic sensitivity tests on different parameters will be performed. First, the influence of atmospheric conditions, such as the moisture content and the static stability are evaluated. Next, the impact of the specific formulation of the framework will be evaluated.

2. chapter 3: The framework is then extended to include not only variations of atmospheric parameters but also variations of the soil moisture. Impacts of soil moisture on the evolution of the PBL and eventually on the amount and timing of precipitation are investigated. The changes in soil moisture are combined with a variation of the static stability of the atmosphere. Long-term interactions and feedbacks between soil moisture and precipitation are thereby investigated.

3. chapter 4: The framework is applied to study implications of climate change on mid-latitude diurnal convection over land. Here we try in an idealized setting to assess the importance of key parameters rather than giving exact future projections.

4. Appendix A: Work focusing on the impact of a large-scale disturbance on an Alpine heavy-precipitation event are presented. In the type of events examined
both the large-scale moisture advection and the destabilization of the atmosphere have a crucial impact on the amount and location of the precipitation. The author of this thesis performed this work mainly as part of her Diploma thesis but work was continued during the PhD.

5. Appendix B: To gain a deeper understanding of the model a simulation without diurnal cycle of radiation is performed. This simulation is used to evaluate heat- and moisture budgets of the framework.

6. Appendix C: A large set of sensitivity studies focusing on both numerical and physical aspects of the model employed are documented.

7. Appendix D: Details of the model formulation are specified at length.
Chapter 2

An idealized cloud-resolving framework for the study of summertime midlatitude diurnal convection over land

Linda Schlemmer¹, Cathy Hohenegger², Jürg Schmidli¹
Christopher S. Bretherton², Christoph Schär¹

¹:Institute for Atmospheric and Climate Science, ETH Zurich, Switzerland
²:Department of Atmospheric Sciences, University of Washington, Seattle, USA

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Abstract

We introduce an idealized cloud-resolving modeling (CRM) framework for the study of mid-latitude diurnal convection over land. The framework is used to study the feedbacks between soil, boundary-layer and diurnal convection. It includes a set-up with explicit convection and a full set of parameterizations. Predicted variables are constantly relaxed towards prescribed atmospheric profiles and soil conditions. The relaxation is weak in the lower troposphere and upper soil, such as to allow the development of a realistic diurnal planetary boundary layer. The model is run to its own equilibrium (30 days).

The framework is able to produce a realistic timing of the diurnal cycle of convection. It also confirms the development of deeper convection in a more unstably stratified atmosphere.

With our relaxation method, the simulated “diurnal equilibrium convection” determines the humidity profile of the lower atmosphere, and the simulation becomes insensitive to the reference humidity profile. However, if a faster relaxation time is used in the lower troposphere, the convection and rainfall become much more sensitive to the reference humidity, consistent with previous studies.
2 An idealized cloud-resolving framework

2.1 Introduction

Convection plays a key role in weather and climate through its impact on the water and energy balance of the Earth. Numerical models have well-known problems with the simulation of convection (e.g. Yang and Slingo, 2001, Arakawa, 2004). On weather timescales, convection is considered the prime reason for the large difficulties with warm-season short-term quantitative precipitation forecasting (Fritsch and Carbone, 2004). On climate timescales, the inability of current climate models to represent organized tropical convection is viewed as a major stumbling block in providing confident projections of future climates (Yang and Slingo, 2001, Randall et al., 2003, Arakawa, 2004, Slingo et al., 2009). Even for key aggregate parameters such as climate sensitivity, associated cloud feedbacks are the largest source of inter-model differences (Bony et al., 2006).

While much of the current research on moist convection focuses on tropical convective systems over sea, recent studies demonstrate that the understanding and simulation of extratropical convection over land needs urgently to be improved. For instance, simulations of present-day and future climates performed with a range of regional climate models (RCMs) over Europe have revealed large differences between simulated precipitation patterns (Christensen et al., 2007). The discrepancies are especially large in summer, when synoptic-scale forcing is weak and thus the influence of the chosen physical model parameterization formulation is large (Vidale et al., 2003, Déqué et al., 2007, Beniston et al., 2007). Summer precipitation is to a large degree determined by moist convection, traditionally represented in RCMs by convective parameterization schemes. An inter-comparison of different parameterizations shows that large difficulties exist in the technical formulation of such schemes (e.g. Molinari and Dudek, 1992). Many convective schemes also exhibit a too early onset and peak of deep, precipitating convection (Dai et al., 1999, Bechtold et al., 2004, Guichard et al., 2004). Brockhaus et al. (2008) noted that, even in multi-summer averages of precipitation over Europe, a strong model shift appears in the convective precipitation peak (3 - 7 hours too early), as well as an incorrect humidity profile.

The increasing computer power allows the use of models with finer grid spacing. The increased resolution enables a more realistic representation of topography and surface fields. Most importantly, the finer grid and the non-hydrostatic model formulation render possible an explicit treatment of convective processes. In numerical weather prediction (NWP), such cloud-resolving models (CRMs) are more and more used on a routine basis (see e.g. Hohenegger and Schär, 2007). In NWP the quality of forecasts can
2.1 Introduction

generally be improved using higher resolution (Mass et al., 2002, Done et al., 2004). In a recent study, Hohenegger et al. (2008b) have applied a CRM on a monthly timescale. These simulations showed very encouraging results. The overall precipitation distribution and evolution was well captured and improved as compared to simulations using parameterized convection. They also showed in a further study that, for this case, CRMs and models using parameterized convection do not agree in their representation of the soil moisture-precipitation feedback (Hohenegger et al., 2009).

CRMs can also be used to better understand the physical mechanisms underlying moist convection and thus ultimately help improving convective parameterizations (Randall et al., 2003). Idealized studies of convective processes using CRMs in mid-latitude continental regions have mostly been forced with observations (see e.g. Xu et al., 2002, Chaboureau et al., 2004, Guichard et al., 2004, Grabowski et al., 2006). The CRMs are relaxed towards observed profiles while at the surface fluxes of latent and sensible heat are prescribed. The response of convection over sea to different idealized atmospheric humidity contents has been investigated by Derbyshire et al. (2004). They found a strong impact of mid-tropospheric humidity on the depth of convection. Dry mid-tropospheric profiles are able to suppress deep convection entirely by strong entrainment of dry air into convective plumes. Wu et al. (2009) performed two-dimensional idealized studies of moist convection over land to investigate the impact of both humidity and stability on the transition from shallow to deep convection and found deeper clouds both for more moist and more unstable soundings. The transition from shallow to deep convection also occurred earlier in these cases.

The aim of the current study is to investigate the sensitivity of moist convection over land to atmospheric stability and humidity. We extend previous work by using a full set of parameterizations including radiation, boundary-layer processes and a soil model to calculate surface heat- and moisture fluxes. This set-up allows the study of the coupled land-surface atmosphere system. Unlike Wu et al. (2009), we employ a 3D geometry and unlike most previous studies we investigate convection in the model’s equilibrium. Here the model is no longer in the spin-up phase and the upper soil layers are in equilibrium with the atmosphere. The only enforced constraints are the conditions in the upper troposphere and in the lower soil. After extended integration time, a quasi-equilibrium state is reached. Here the term equilibrium state does not mean that convection is in equilibrium with the large-scale forcing as defined by Arakawa and Schubert (1974). Rather the evolving convection shows a strong diurnal cycle that is distinct from a radiative-convective equilibrium (Manabe and Strickler, 1964) and also from equilibrium convection over land as defined by Done et al. (2006). Instead the
model is allowed to attain a state where the diurnal cycle repeats itself from day to day. We use the term “diurnal equilibrium” for this state in the following. The boundary layer is enabled to develop into a state where, integrated over the diurnal cycle, the surface latent heat flux balances precipitation and the drying from subsidence, while the surface sensible heat fluxes and subsidence warming are balanced by net radiative cooling as well as evaporative cooling from falling precipitation and warming from condensation (Betts, 2000). We will compare this diurnal equilibrium convection with the convection during the first days, when the model has not yet reached its equilibrium. Eminent for land surfaces is the strong diurnal cycle of surface fluxes and the fact that it needs a certain time until soil and atmosphere are in equilibrium. In studies of convection over prescribed sea-surface temperatures, as e.g. done in Derbyshire et al. (2004), we expect the equilibrium to be reached more quickly than over land surfaces. Convection will also be more continuous over sea surfaces as the diurnal cycle of heat fluxes is considerably weaker.

The paper is organized as follows: in section 2.2 we describe the set-up of the experiment and the model used. The control simulation and the sensitivity experiments that test the role of tropospheric humidity and stability are presented in section 2.3. Finally the results are concluded in section 2.4.

2.2 Experimental set-up

2.2.1 Model description

The simulations are performed with version 4.0 of the Consortium for Small-Scale Modeling Model in Climate Mode (CCLM). The CCLM is a nonhydrostatic limited-area model that numerically solves the fully compressible equations for a moist atmosphere using the split-explicit approach (see especially Wicker and Skamarock, 1998, Steppeler et al., 2003, Doms and Förstner, 2004, Will et al., 2008). The model has been validated both at grid-spacings of 0.22 and 0.44° (Jaeger et al., 2008) and cloud-resolving scales (Δx=2.2 km) (Hohenegger et al., 2008b).

In our simulations the horizontal grid-spacing is Δx = 2.2 km and the large time step 20 s. We refer to the grid-spacing of 2.2 km as a cloud-resolving model in the sense that it is able to adequately represent vertical fluxes of heat, moisture and momentum associated with organized, deep convection (see e.g. Weisman et al., 1997). We do however not claim to resolve single convective clouds. The computational domain comprises 100 x 100 grid points and 50 vertical levels. In the vertical, a Gal-Chen coordinate (Gal-Chen and Somerville, 1975) is used with a grid spacing of 20 m in the
lowest levels and \( \approx 400 \) m in the middle troposphere. Double-periodic lateral boundary conditions are employed. At the upper boundary a Rayleigh damping sponge layer starting at a height of 12 km is used to absorb gravity waves.

A third order Runge-Kutta scheme is utilized for the time integration (see Förstner and Doms, 2004) with a fifth-order advection scheme for horizontal and vertical winds, temperature and pressure and a second order Bott scheme (Bott, 1989) for horizontal advection of moist quantities. Coriolis forces are set to zero and the spherical geometry is neglected. Disregarding the Coriolis forces, we might miss some enhancement of convection through organization of individual cells into mesoscale convective systems (e.g. Skamarock et al., 1994), but we simplify the set-up ensuring consistent lateral boundary conditions.

The model is run with a full parameterization package including a radiation scheme after Ritter and Geleyn (1992), a prognostic turbulent energy-based scheme for turbulent diffusion of order 1.5 (see Raschendorfer, 2001) based on Mellor and Yamada (1974), a turbulent kinetic energy-based surface layer scheme (Mironov and Raschendorfer, 2001) and a five categories microphysics bulk scheme described in Reinhardt and Seifert (2006) considering prognostic cloud water, cloud ice, precipitation, graupel and snow. Furthermore a subgrid-scale cloud-scheme affecting radiation and turbulence is utilized. The model includes a subgrid-scale shallow convection scheme, but this is switched off for the current simulations. To simulate the interactions between the atmosphere and the underlying soil, the multi-layer soil model TERRA_ML after Heise et al. (2003) with 10 layers is used with a depth of the uppermost layer of 1 cm. The upper seven soil layers down to a depth of 1.47 m are hydrologically active and the lowest three layers are climatic layers. The soil type is set to loam. Vegetation is parameterized by prescribing plant-cover (0.84), leaf-area index (2.96) and root depth (0.56 m) characteristic for European conditions. The surface roughness length is set to 0.0387 m. Evapotranspiration includes bare soil evaporation, evaporation from the interception reservoir and transpiration from vegetation. The formulation of the terms closely follows the Biosphere-Atmosphere Transfer Scheme (BATS) (Dickinson, 1984).

The elevation of the surface is set to 489.0 m corresponding to the altitude of Munich. Incoming solar radiation is determined according to 48.25° N and 0° E on 12th July throughout the whole simulation. The model is initialized with a single vertical sounding prescribing temperature, specific humidity and the horizontal wind components (see section 2.2.3 for further details). White noise is applied to the initial temperature at the lowest layer with a maximum amplitude of \( \pm 0.02 \) K to break the symmetry of the initial
conditions. The model is run for 30 days.

2.2.2 Relaxation method

Our aim is to represent diurnal convection over land in a doubly-periodic computational domain. Conceptually we can think of the moist convection interacting with large-scale synoptic forcing as well as boundary-layer, radiative and land-surface processes. In our modelling framework the large-scale forcing is represented by relaxing the simulated atmosphere towards an externally prescribed profile, while the other processes are explicitly simulated.

We relax the mean model state towards the prescribed atmospheric profiles using a height-dependent strength of the relaxation with weak and strong relaxation in the lower and upper troposphere, respectively. The reason behind this choice is as follows: First, in the absence of large-scale perturbations near-surface conditions are primarily controlled by boundary-layer and convective parameters, while the upper tropospheric conditions are more nonlocally controlled due to horizontal advection by strong winds and deep gravity wave adjustment. Indeed, the general increase of the horizontal wind with height implies stronger advective tendencies at high altitudes. Second, as we include the full sequence of model parameterizations in our framework, we do not merely address the role of convection in some atmospheric environment, but rather the full interaction between the soil and the atmosphere. The relaxation of the lower-tropospheric conditions towards a prescribed profile would thus be difficult to justify, in particular as we allow for a strong diurnal cycle over continental land-surfaces. This strong diurnal cycle is in contrast to the diurnal cycle over tropical oceans, where it is much weaker and heat-fluxes from the surface are dominated by latent heating.

As our modeling framework includes a soil model, a similar relaxation strategy is also applied here. We relax the soil towards prescribed profiles of soil water content with smaller relaxation coefficients in the upper soil layers. Overall, the system studied corresponds to forcing the system by relaxing in the upper atmosphere and lower soil, while the relevant physical processes are allowed to determine the properties and diurnal cycle near the soil/atmosphere interface.

The height-dependent relaxation is implemented by an additional term that is added to the right-hand side of the prognostic equations:

\[
\frac{\partial X}{\partial t} = \left( \frac{\partial X}{\partial t} \right)_{\text{other terms}} - \frac{X - X_{\text{ref}}}{\tau} \cdot f(p), \quad X : T, Q_v, U, V \tag{2.1}
\]
2.2 Experimental set-up

\[ f(p) = 0.5 \cdot (1 + erf\left(\frac{p_0 - p}{b}\right)) \]  

(2.2)

where \(X_{ref}\) are the values of the reference-profile (see section 2.2.3), \(\bar{X}\) domain-mean values of the predicted variables, \(p\) the pressure and \(erf\) the error-function defined as:

\[ erf(x) = \frac{2}{\sqrt{\pi}} \int_0^x e^{-t^2} dt. \]

We set \(p_0\) to 500 hPa and \(b\) to 300 hPa, yielding the relaxation profile illustrated in Fig. 2.1a. A relaxation time constant \(\tau\) of 1 day is chosen. The same profiles are used as input and reference profiles towards which the simulations are relaxed.

![Figure 2.1: Functions used to control the strength of the relaxation in (a) the atmosphere (see equation 2.2) and (b) the soil.](image)

Soil moisture is relaxed in a similar way to balance gravitational and surface runoff and to restore water that was evaporated into the atmosphere. The chosen relaxation profile is shown in Fig. 2.1b, where \(\tau_{soil}\) is set to 2 days. Deep soil temperature remains approximately constant without relaxation and is thus not relaxed.

2.2.3 Input and reference profile

We use idealized profiles to drive our simulations. The profiles are motivated by radiosoundings from Munich over July 12th-15th in 2006 and zonal mean climatologies valid at 48° N for summer conditions from Peixoto and Oort (1992). For temperature (Fig. 2.2a, black lines), a linear decrease with height and constant values above 12 200 m are assumed. Relative humidity in the troposphere (Fig. 2.2b) is modeled based on the climatology of Peixoto and Oort (1996). The best fit to these profiles is obtained by
combining two tanh functions as follows:

\[
RH(p) = \begin{cases} 
\frac{RH1+RH2}{2} - \frac{RH1-RH2}{2} \cdot \tanh \left( \frac{z}{2500 \, m} - 1.8 \right), & \text{for } p > 250 \, hPa \\
\frac{RH2+RH3}{2} - \frac{RH2-RH3}{2} \cdot \tanh \left( \frac{z}{300 \, m} - 30 \right), & \text{for } p \leq 250 \, hPa 
\end{cases}
\] (2.3)
where \( z \) is the height. \( RH_1 \) and \( RH_2 \), the relative humidity in the lower and upper troposphere respectively, are the parameters to be varied in a sensitivity experiment (see below). In the stratosphere we assume a relative humidity of \( RH_3 = 10\% \). Zonal wind (Fig. 2.2a, grey solid line) is based on Oort and Peixoto (1983) and is constructed using two quadratic functions. In the troposphere westerly winds increase with height from 2 m s\(^{-1}\) at the surface to their maximum value of 17.5 m s\(^{-1}\) at 200 hPa as shown in Fig. 2.2a. In the stratosphere we assume a transition to easterly winds consistent with summer conditions. Meridional winds are set to zero.

The reference soil moisture saturation \( S \) amounts to 60 % at the surface, increasing to 100 % at a depth of 2.50 m following a quadratic function (see Fig. 2.2c, solid line). The soil temperature is constructed assuming a surface temperature of 18°C, a yearly-mean temperature of 8°C at a depth of 12 m and a linear decrease between these two values (Fig. 2.2c, dashed line).

To test how convection reacts to a changed humidity content of the lower and middle troposphere the simulations DRY, CTL and WET are performed, where \( RH_1 \) and \( RH_2 \) are changed (values are given in Table 2.1). To investigate the role of the lower troposphere a fundamentally different set of simulations (DRY_CST, CTL_CST and WET_CST) is carried out where the relaxation strength is height-independent, meaning \( f(p) = 1 \) for all values of \( p \). Otherwise, they use the same settings as DRY, CTL and WET, respectively. To investigate the effect of atmospheric stability on the diurnal cycle of convection the three simulations STABLE, CTL and UNSTABLE with changed temperature lapse-rate are performed (see Table 2.1).

Values for convective available potential energy (CAPE), convective inhibition (CIN) (see e.g. Emanuel et al., 1994) and precipitable water content for the different input/reference profiles are given in Table 2.1. As expected the WET simulations exhibit larger CAPE, smaller CIN values and more precipitable water than the DRY simulations. Note that CAPE increases with decreasing stability and increasing humidity. Values for precipitable water are largest for the stable soundings as the middle and upper troposphere is warmest in these profiles. With a given value for relative humidity this results in the largest water amounts.
Figure 2.3: Time series of domain mean values of (a) the vertical mean temperature (°C), (b) pressure (hPa) in the lowest atmospheric layer, (c) surface precipitation rate (mm h⁻¹), (d) soil saturation (%) of the uppermost 0.5 cm deep soil layer (solid line) and lowest hydrologically active layer (dashed line) and (e) soil temperature (K) of the uppermost soil layer (solid line) and the lowest hydrologically active layer (dashed line). The lowest hydrologically active layer is situated at 1.47 m depth.
Table 2.1: RH1, RH2, dT/dz, CAPE, CIN and precipitable water values for the input/reference profiles of the different simulations. The simulations DRY\_CST, CTL\_CST and WET\_CST use the same settings as DRY, CTL and WET, but are performed with a height-independent, constant relaxation coefficient.

<table>
<thead>
<tr>
<th></th>
<th>RH1 [%]</th>
<th>RH2 [%]</th>
<th>dT/dz [K/100m]</th>
<th>CAPE [J kg(^{-1})]</th>
<th>CIN [J kg(^{-1})]</th>
<th>Precipitable water [mm]</th>
</tr>
</thead>
<tbody>
<tr>
<td>DRY</td>
<td>60</td>
<td>30</td>
<td>-0.7</td>
<td>48.2</td>
<td>82.3</td>
<td>19.8</td>
</tr>
<tr>
<td>CTL</td>
<td>70</td>
<td>40</td>
<td>-0.7</td>
<td>344</td>
<td>31.1</td>
<td>23.5</td>
</tr>
<tr>
<td>WET</td>
<td>80</td>
<td>50</td>
<td>-0.7</td>
<td>821</td>
<td>8.59</td>
<td>27.2</td>
</tr>
<tr>
<td>STABLE</td>
<td>70</td>
<td>40</td>
<td>-0.8</td>
<td>0.0111</td>
<td>0.0</td>
<td>26.9</td>
</tr>
<tr>
<td>UNSTABLE</td>
<td>70</td>
<td>40</td>
<td>-0.6</td>
<td>1740</td>
<td>10.3</td>
<td>20.9</td>
</tr>
</tbody>
</table>

Table 2.2: Mean values of precipitation, vertically integrated mass flux and surface sensible and latent heat flux averaged over days 16-30 for all experiments.

<table>
<thead>
<tr>
<th></th>
<th>Precipitation [mm h(^{-1})]</th>
<th>Integrated convective mass-flux [kg m(^{-1}) s(^{-1})]</th>
<th>Sensible heat flux [W m(^{-2})]</th>
<th>Latent heat flux [W m(^{-2})]</th>
</tr>
</thead>
<tbody>
<tr>
<td>DRY</td>
<td>0.130</td>
<td>96.7</td>
<td>28.8</td>
<td>132</td>
</tr>
<tr>
<td>CTL</td>
<td>0.140</td>
<td>101</td>
<td>28.4</td>
<td>131</td>
</tr>
<tr>
<td>WET</td>
<td>0.147</td>
<td>107</td>
<td>27.9</td>
<td>130</td>
</tr>
<tr>
<td>DRY_CST</td>
<td>0.0478</td>
<td>34.0</td>
<td>35.1</td>
<td>122</td>
</tr>
<tr>
<td>CTL_CST</td>
<td>0.102</td>
<td>58.2</td>
<td>36.6</td>
<td>118</td>
</tr>
<tr>
<td>WET_CST</td>
<td>0.155</td>
<td>81.2</td>
<td>35.6</td>
<td>110</td>
</tr>
<tr>
<td>STABLE</td>
<td>0.120</td>
<td>57.2</td>
<td>24.2</td>
<td>123</td>
</tr>
<tr>
<td>UNSTABLE</td>
<td>0.142</td>
<td>180</td>
<td>32.0</td>
<td>130</td>
</tr>
</tbody>
</table>

2.3 Results

2.3.1 The control simulation

In this section we describe the characteristics of the control simulation with a lapse-rate of -0.7 K/100 m, and relative humidities of 70 % in the lower troposphere and 40 % in the upper troposphere (see section 2.2.3). Fig. 2.3 shows simulated time series of the vertical mean temperature, pressure, surface precipitation rate, soil temperature and soil saturation averaged over the computational domain. As can be clearly seen diurnal equilibrium is reached after 16 days. This conclusion also holds for other variables. Hence, days 16 to 30 of the simulation are used to evaluate the mean diurnal cycle of convection. The 15 days can be thought of as an ensemble where each day is a
Figure 2.4: Profiles of (a) specific humidity \( \text{g kg}^{-1} \), (b) temperature \( ^\circ\text{C} \), (c) zonal wind (solid) and meridional wind (dashed line) \( \text{m s}^{-1} \) and (d) soil moisture saturation \( S \) (%) as a function of soil depth. Black lines show the prescribed profiles, grey shade simulated hourly values for the CTL simulation averaged over the computational domain.

realization of the same experiment with slightly varied initial conditions.

The ability of the relaxation to constrain the profiles in the upper troposphere and the deeper soil is demonstrated in Fig. 2.4 for specific humidity, temperature, wind and soil-water content. The simulated values are close to the reference profile in the upper troposphere and stratosphere but show a strong diurnal cycle in the lower troposphere. Convection furthermore transports low zonal momentum into the upper troposphere. Zonal winds are therefore reduced between 200 and 400 hPa and vice versa below (Fig. 2.4c). Soil moisture saturation (Fig. 2.4d) shows a strong diurnal cycle in the up-
permost layer and nearly constant values in the lowest hydrologically active layer.

Fig. 2.5 shows clouds (grey shading), surface precipitation (colored shading), upward velocity (red contour lines) and horizontal winds (black arrows) at different times of the day to illustrate the structure of the ongoing convection. Narrow updraft regions with the

Figure 2.5: Clouds (grey shading), surface precipitation (colored shading, mm h\(^{-1}\)), upward velocity at 860 hPa (red contour lines) and horizontal wind field at 977 hPa (black arrows) at (a) 1530, (b) 1830 and (c) 2130 UTC at day 23 of the simulation. A cloud is identified by \( QC > 10^{-6} \text{ kg kg}^{-1} \) somewhere in the vertical column above. Upward velocity was determined by \( w > 0.2 \text{ m s}^{-1} \).

Figure 2.6 (following page): Mean diurnal cycle of (a) specific cloud water content (\( \text{kg kg}^{-1} \), shaded area), specific cloud ice content (\( \text{kg kg}^{-1} \), contour lines) and surface precipitation (\( \text{mm h}^{-1} \), solid black line, the variability of the domain mean value over the 15 days is indicated by the black shading showing minimum and maximum values), (b) convective mass-flux (shaded area, \( \text{kg m}^{-2} \text{ s}^{-1} \)), (c) surface net shortwave radiation (\( \text{SW, solid line} \)), longwave net radiation (\( \text{LW, dashed line} \)), sensible heat flux (\( \text{H, dotted line} \)) and latent heat flux (\( \text{LE, dash-dotted line} \)) in \( \text{W m}^{-2} \). (d) Mean diurnal cycle of CAPE (J kg\(^{-1}\)) and (e) CIN (J kg\(^{-1}\)) with domain mean values in black and mean of cloudy profiles in grey. Mean values are shown in solid lines while the 10th and 90th percentile are dashed. The 10th and 90th percentile were calculated by considering all grid points at all 15 days at each time of the day. (f) Height of the domain mean value (solid line) and 10th and 90th percentile (dashed lines) of the LCL (black line) and the LFC (grey line). All panels are for the CTL simulation. Averages are taken here and in the following figures over day 16-30 and over the computational domain except noted differences.
An idealized cloud-resolving framework

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2.3 Results

vertical velocity $w > 0.2 \text{ m s}^{-1}$ and convergence of horizontal winds are visible. Clouds are mostly collocated with those updraft regions. The location of surface precipitation is shifted with respect to the clouds. Regions with divergent wind field caused by the downdrafts due to evaporative cooling of falling precipitation are furthermore visible. Downdraft regions occupy a considerably larger space than updraft regions. At 1530 UTC deep convection is in an early stage with a large number of updraft regions and small precipitation amounts reaching the surface. At 1830 UTC convection is very active with organized updraft and downdraft regions, whereas at 2130 UTC the number of updrafts is decreased but precipitation continues to fall from remaining passive clouds.

Specific cloud water and cloud ice content (Fig. 2.6a) show a distinct diurnal cycle. In the early morning there is a thin mid-level cloud cover and some thin ice-clouds (domain mean value $< 10^{-6} \text{ kg kg}^{-1}$) remaining from convection of the previous day presumably amplified by radiative cooling. In the early afternoon (1430 UTC) deep convection and precipitation is simulated. Deep convection continues for 5-6 h until clouds decay in the evening. Convective mass-flux (Fig. 2.6b) shows a comparable picture. Convective mass-flux $M_c$ was calculated (see e.g. Robe and Emanuel, 1996):

$$M_c = \frac{1}{A} \int_{Qc + Qi > 10^{-6} \text{ kg kg}^{-1}} \rho w \, dx \, dy$$

(2.4)

Where $A$ is the domain and $\rho$ the density of moist air. As a result of the convection a pronounced diurnal cycle of precipitation is simulated (Fig. 2.6a, black shaded area). The onset of precipitation coincides with the point in time when clouds reach their maximum depth and ice formation sets in. The precipitation peaks at 1730 UTC and daily area-mean precipitation sums are 3.4 mm. The day to day variability is small. Precipitation peak time varies by 2 hours and precipitation amounts by a factor of 1.6. The domain mean hourly peak rate of 0.9 mm h$^{-1}$ is at the upper end of the range of simulated values of Guichard et al. (2004) who performed idealized simulations of the diurnal cycle of deep precipitating convection over land with a range of CRMs and single column models (SCMs).

Incoming net shortwave radiation (Fig. 2.6c, solid line) is the primary driver of the diurnal cycle. The heating of the surface leads to sensible and latent heat fluxes (Fig. 2.6, dotted and dash-dotted lines) which activate the growth and moistening of the PBL. Latent heat fluxes are larger than sensible heat fluxes leading to an average Bowen ratio around 0.4. The simulated values agree well with observed values from the GCSS WG4 Case 3 (continental convection over the Southern Great Plains) intercomparison.
project (Xu et al., 2002). This is valid both for the surface net radiation (shortwave plus longwave) and the sensible and latent heat fluxes.

Fig. 2.6d, e show the diurnal cycle of CAPE and CIN values, and Fig. 2.6f the height of the LFC, the LCL and the PBL. CAPE and CIN values are calculated assuming a mixed layer extending over the lowest 150 hPa. The height of the PBL was determined using the bulk Richardson method. A value of 0.33 under stable conditions (Wetzel, 1982) and of 0.22 under convective conditions (Vogelzang and Holtslag, 1996) were used as thresholds. To estimate the spatial variability we additionally include the 10th and 90th percentile of the values with a separation between cloudy profiles ($QC > 10^{-6}$ kg kg$^{-1}$) and the mean value over all points. Not surprisingly the mean value of CAPE at cloudy points is much smaller than the mean over the whole domain as active convection reduces CAPE at these spots. CAPE reaches a maximum around 1430 UTC, three hours before the precipitation peak. The maximum amounts to 1500 J kg$^{-1}$ which is much larger than the CAPE values of the input profile (see Table 2.1). Thereafter CAPE starts to decrease and becomes comparable to the CAPE of the reference profile.

CIN (see Fig. 2.6e) starts to decrease already from 0600 UTC onwards and falls below 10 J kg$^{-1}$ around 1200 UTC. The onset of convection roughly coincides with the time when the descending LFC reaches the LCL (see Fig. 2.6f). As to be expected, there is large spatial variability; already at 1130 UTC the 90th percentile of the LCL and the 10th percentile of the LFC coincide (Fig. 2.6f). Remarkably, it takes another three hours for deep convection to develop (see Fig. 2.6a). Hereby one must consider that CIN calculations do not take into account the dilution of ascending parcels which delays convection. This dilution of plumes could explain why air remains subsaturated and clouds are absent despite small CIN amounts. The increase in CIN from 1500 UTC onwards both in the environment and under cloudy points does not limit precipitation activity as for most cloudy points CIN remains zero. PBL height is shown by the dark grey line in Fig. 2.6f. Its 90th percentile approaches the LFC already at 1300 UTC, while the domain mean value of the ascending PBL height and the descending LFC never reach each other.

The simulated diurnal cycle compares well with observations and earlier studies (e.g. Chaboureau et al., 2004), the only difference is the apparent absence of shallow convection. To clarify ongoing processes we calculate the normalized saturation deficit (NSD) (see Fig. 2.7) as introduced by Chaboureau et al. (2004):

$$NSD = \frac{r_{sat} - r}{\sigma(r_{sat} - r)}$$  \hspace{1cm} (2.5)
where \( r \) is the mixing ratio, \( r_{sat} \) the saturation mixing ratio and \( \sigma(r_{sat} - r) \) the standard deviation of the saturation deficit \( r_{sat} - r \). The saturation deficit is determined by calculating for each hour of the day the mean of the saturation deficit over all points in the domain and all 15 evaluation days. The standard deviation was determined as a function of height using all values in the domain over all 15 evaluation days for each time of the day. Reduced values of the NSD indicate both a moistening of the air by diluting ascending parcels (numerator) and an increased mixing between moist boundary layer air and and dry environmental air from above (denominator).

In CTL a tongue of reduced values of NSD is present ahead of the presence of cloud condensate from about 1100 UTC (Fig. 2.7). The tongue of reduced values of NSD indicates that the top of the PBL is approaching saturation as the PBL deepens, a processes inherent to shallow convection. Shallow clouds are however not simulated in contrast to the simulations of Chaboureau et al. (2004, see their Fig. 5).

The simulated diurnal cycle corresponds to the expected convective development over mid-latitude continental area in case of weak synoptic-scale forcing. Overall we conclude that the model is able to simulate a reasonable diurnal cycle of convection.

### 2.3.2 Sensitivity experiments

In the following we present the sensitivity experiments introduced in section 2.2.3 in order to investigate how the diurnal convection reacts to changes in relative humidity

**Figure 2.7:** Mean diurnal cycle of normalized saturation deficit (cf. equation 2.5) with the 0.01 g kg\(^{-1}\) contour line of cloud condensate (black line) for CTL.
Figure 2.8: Vertical profiles of domain mean values of relative humidity (%) at (a) 0000 and (b) 1000 UTC for DRY, CTL, WET, DRY.CST, CTL.CST and WET.CST.

Sensitivity to humidity and role of relaxation
We perform three simulations to investigate the sensitivity of convection to the relative humidity: DRY, CTL and WET (see Fig. 2.2b). The three sensitivity use relative humidity values between 60 and 80% in the lower troposphere and between 30 and 50% in the upper troposphere. Fig. 2.8 shows simulated profiles of relative humidity at 0000 and 1000 UTC. Most prominently, the predicted equilibrium values of relative humidity differ only by a few percent between the CTL, WET and DRY simulations (black lines), despite large differences in the prescribed humidity profiles. Also the diurnal cycles of cloud water content, ice content and precipitation in Fig. 2.9 show strikingly similar diurnal cycles. Likewise, 2 m temperature, 2 m dew point depression and surface heat-fluxes are remarkably insensitive to changes in prescribed humidity (not shown).

Results thus show that the evolving diurnal equilibrium state in the lower troposphere is mostly independent of the prescribed moisture profile. This is in contrast to previous studies (e.g. Wu et al., 2009). Using short-term integrations, these studies documented a pronounced sensitivity of convection to environmental humidity. To understand the reasons behind our results we investigate the first phase of the simulations, when the model is not yet in a diurnal equilibrium state. Fig. 2.10 shows the evolutions of cloud water, cloud-ice and surface precipitation during the first three days for CTL and WET. The development of clouds and mass-fluxes is considerably delayed in CTL compared to WET during the first day of the simulation. This is in accordance with earlier studies (e.g. Derbyshire et al., 2004, Wu et al., 2009). The entrainment of dry air into ascend-
### 2.3 Results

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#### Figures 2.9 and 2.10

**Figure 2.9:** Top panels show the mean diurnal cycle of specific cloud water content (kg kg\(^{-1}\), shaded area), cloud ice content (kg kg\(^{-1}\), contour lines) and domain mean surface precipitation (mm h\(^{-1}\), black solid line, minimum and maximum values over the 15 days of simulation are indicated by dark grey shading). The number in the lower left corner gives the mean daily precipitation amount (mm day\(^{-1}\)). Bottom panels show the mean diurnal cycle of convective mass-flux (kg m\(^{-2}\) s\(^{-1}\), shaded area).

**Figure 2.10:** Domain mean specific cloud water content (kg kg\(^{-1}\), shaded area), specific cloud ice content (kg kg\(^{-1}\), contour lines) and surface precipitation (mm h\(^{-1}\), solid black line) for (a) CTL and (b) WET during the first three days.
An idealized cloud-resolving framework

ing plumes delays the deepening of convection. In addition cloud bases are lower in WET. Already at day three, differences become small as convection has redistributed the moisture in the atmosphere. Latent and sensible heat fluxes explain how the different simulations approach each other. CTL has larger latent heat fluxes than WET which help to moisten the lower atmosphere (not shown). It is further interesting to note that during the first day of the simulation, shallow convection is simulated in both settings, while on subsequent days the atmosphere is already sufficiently moistened by the convection of the previous day and an immediate transition to deep convection is possible.

The results show that the simulations are able to “build” their own PBL structure and cloud fields, irrespective of the prescribed humidity profiles. We test this by limiting the freedom of the model in a set of simulations with height-independent relaxation \( f(p) = 1 \) everywhere, see section 2.2.3. This height-independent relaxation corresponds to a strong external forcing throughout the whole troposphere.

With uniform relaxation the humidity is forced to stay close to the given profile, also in the lower layers. Figs. 2.11 shows the resulting clouds, precipitation and convective

![Figure 2.11: Same as Fig. 2.9 but for simulations using height-independent relaxation: (a, d) DRY_CST, (b, e) CTL_CST and (c, f) WET_CST.](image-url)
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mass-fluxes. The resulting diurnal-equilibrium state now differs between WET_CST, CTL_CST and DRY_CST. With the dry profile, the onset of cloud formation is delayed by half an hour, the cloud base is shifted upwards (from ≈ 850 hPa in CTL to ≈ 820 hPa in DRY_CST), the cloud top is lowered, convective mass-fluxes (Figs. 2.11d-e) start later (1330 UTC in DRY_CST, 1300 UTC in CTL_CST and 1230 UTC in WET_CST) and are less intense, the onset of precipitation (Figs. 2.11a-c, black shaded area) is delayed and the precipitation amounts reduced.

The domain mean vertical profiles of relative humidity at 0000 UTC and 1000 UTC are shown in Fig. 2.8 by the grey solid lines. In agreement with Fig. 2.11 and in opposition to WET, CTL and DRY, they differ between WET_CST, CTL_CST and DRY_CST and are largest for WET_CST. The unforced profiles (WET, CST and DRY) show a peak in relative humidity higher up than the forced simulations. The low-level humid layer that has evolved in both the forced and unforced simulations at 1000 UTC is at a similar height in all simulations. It corresponds to the top of the PBL. This humid layer will eventually become moister and reach saturation (not shown). Saturation will first be reached in the WET_CST simulation.

In summary, the simulations demonstrate that the influence of humidity depends upon relaxation strategy when considering the feedback between the land and the atmosphere. With weak relaxation in the lower troposphere, it is soil moisture and the PBL that ultimately control low-level humidity and thus convective activity.

We also performed additional simulations with height-dependent relaxation and profiles where the relative humidity is changed only in the upper troposphere. The change of humidity in the upper troposphere has negligible influence on the convection (not shown). Humidity is transported upwards from the soil into the lower and then upper troposphere. This transport dominates the moistening/drying by relaxation.

Sensitivity to the vertical stability

As detailed in section 2.2.3 we perform three simulations to investigate the sensitivity of convection to the static stability: STABLE, CTL and UNSTABLE (see also Fig. 2.2a). All simulations use the same relative-humidity profile with RH1=70 % and RH2= 40 % but different temperature profiles.

Fig. 2.12 shows the diurnal cycle of cloud water and cloud ice content, precipitation and convective mass-fluxes for the three simulations. The stability of the atmosphere has a decisive influence on the ability of convection to develop. Cloud tops are con-
Figure 2.12: Same as Fig. 2.9 but for (a, d) STABLE, (b, e) CTL and (c, f) UNSTABLE.

Figure 2.13: (a) Mean diurnal cycle of surface precipitation (mm h\(^{-1}\), solid lines) and the variability of the domain mean value over the 15 days (shading indicating the range between minimum and maximum values). (b) Vertical profiles of domain mean potential temperature (K) for day 16-30 of the simulation for the STABLE (black), CTL (blue) and UNSTABLE (red) simulation.
2.3 Results

Figure 2.14: Vertical profiles of domain mean values of relative humidity (%) at (a) 0000 and (b) 1200 UTC for STABLE, CTL and UNSTABLE.

 considerably higher in UNSTABLE (reaching up to $\approx 130$ hPa) and shallower (reaching up to $\approx 210$ hPa) in STABLE. In STABLE, deep clouds start to develop already at 1230 UTC, convective mass-fluxes set in at 1230 UTC (Fig. 2.12f), cloud bases are shifted downward and the onset and peak of precipitation are slightly advanced. Surface precipitation for the three simulations is shown in Fig. 2.13a. The day to day variability of the peak time varies between 1.5 hours in STABLE and 2 hours in CTL and UNSTABLE. In UNSTABLE the onset of convective mass-fluxes is delayed to 1430 UTC (1400 UTC in CTL), mass-fluxes reach up higher and the base is shifted upwards.

Figure 2.15: Skew-T log-p diagram of the domain mean values at 0600 UTC averaged over the 15 days of the simulation of STABLE (black) and UNSTABLE (red). The dashed lines indicate a parcel ascent computed from the values at the lowest atmospheric level.
Vertical profiles of domain mean relative humidity at 0000 and 1200 UTC are shown in Fig. 2.14 and a skew-T log-p diagram at 0600 UTC is shown in Fig. 2.15. In STABLE, boundary-layer moisture is larger leading to a lower LCL and LFC. The higher PBL-moisture content in STABLE reduces outgoing longwave radiation and leads to warmer early-morning near-surface temperatures. As a consequence the LCL coincides with the LFC earlier which explains the earlier onset of convection (see Fig. 2.6f for CTL). The level of neutral buoyancy (LNB) is also lower in STABLE and the freezing level is shifted upwards due to the smaller lapse rate. Therefore only relatively small amounts of ice clouds are simulated. Due to the thicker mid-level cloud cover in STABLE in the morning, the maximum sensible heat flux is reduced by $\sim 35 \text{ W m}^{-2}$ and maximum latent heat fluxes is reduced by $20 \text{ W m}^{-2}$ in comparison to CTL (not shown). This reduction in evapotranspiration can explain reduced precipitation amounts in STABLE.

As can be seen in Fig. 2.12 UNSTABLE exhibits the deepest clouds with most ice formation. This is in agreement with the findings of Wu et al. (2009) and Derbyshire et al. (2004). The build-up of clouds is delayed compared to the CTL simulation as the LCL and LFC are considerably higher in the morning. This upward shift of LCL and LFC are caused by drier low-level conditions. The deepening of the PBL is however faster in the UNSTABLE simulation as a smaller temperature gradient needs to be overcome and sensible heat fluxes are larger than in CTL (not shown). Convection starts when the growing PBL coincides with the LFC and it is more intense in UNSTABLE when it finally starts. Despite pronounced differences in convective activity, the overall amount of precipitation is comparable between CTL and UNSTABLE. The reduction in STABLE can be explained by the reduction in evapotranspiration.

Our simulations are able to replicate a significant time delay between maximum net radiation at the surface (at local noon) and the development of convective precipitation (in the afternoon or evening) (e.g. Bechtold et al., 2004). While a detailed analysis would require additional studies, the current simulations are suggestive of the following hypothesis: The simulations confirm that convection is initiated within the diurnal cycle once the descending level of free convection meets the ascending lifting condensation level (Fig. 2.6f). Both the LFC and LCL changes depend in complex manners upon the accumulation of sensible and latent surface heat fluxes in the PBL. As the accumulation of heat and moisture plays the decisive role (rather than merely the instantaneous fluxes), it is not surprising that a time delay between peak surface radiation and convection occurs. The complexity of the underlying interactions (which in particular involves the growth of a well-mixed boundary layer) allows for a considerable case-to-case sensitivity. Indeed, the different simulations exhibit very different delays. For instance,
UNSTABLE has peak precipitation at 2100 UTC, much later than STABLE where it occurs at 1600 UTC (see Fig. 2.13a).

In opposition to the role of humidity the evolving state of diurnal equilibrium is dependent on the stability of the atmosphere. The relaxation however acts both on humidity and temperature to maintain the desired profile. Why do the simulations preserve the different temperature profiles but not the different moisture profiles?

Fig. 2.13b shows simulated vertical profiles of potential temperature for STABLE, CTL and UNSTABLE. The simulations have the same surface temperatures at the initial time, but the different temperature gradients imply different temperatures at the tropopause level. During the simulation temperatures are fixed at the tropopause level through the relaxation. Diurnal mean temperatures in the lower troposphere warm by 7.6 K in STABLE relative to the reference profile, but only 2.4 K in UNSTABLE. The difference in stability between simulations throughout the whole troposphere thereby becomes smaller during the spinup phase but still remains.

Despite the modified stability the colder upper-tropospheric temperatures in UNSTABLE certainly also have an influence on the precipitation intensity due to the larger amount of ice clouds and the latent heat of freezing that is released (e.g. Houze, 1993, Chapter 8).

From a moisture budget perspective the moisture sink due to the relaxation is small in the simulations with different reference relative humidity profiles but the same temperature profiles (≈ 17% of the evapotranspiration in CTL). The upper troposphere where the adjustment is active is dry, while the temperature adjustment is working similarly in all three cases. Furthermore, all simulations except the _CST cases have similar mean rainfall rates, corresponding in energy units to about 100 W m⁻². This is also understandable from a budget perspective, because the latent heat flux is about 130 W m⁻², and the adjustment sink of moisture is small, so the rainfall has to nearly balance the surface evaporation. The surface evaporation rate in turn is controlled by the net radiative flux into the surface and the soil moisture. The former is roughly the same in all cases despite some differences in clouds (most pronounced in STABLE). The latter will respond to the mean precipitation, which is similar in all cases. Thus we don’t expect much difference in the mean precipitation between the cases, even those in which static stability is changed. The exception is the _CST cases for which the moisture adjustment sink can be much stronger because it includes a large lower-tropospheric contribution.
2.4 Summary

We have presented a novel method to study the interaction of the land surface with the atmosphere in convective weather regimes lasting over several days. These regimes are characterized by weak pressure gradients, weak advection in the lower and middle troposphere while stronger advection prevails in the upper troposphere. We simulated such convective regimes using a cloud-resolving model (CRM) in an idealized setting. The effect of the large-scale forcing was prescribed by relaxing temperature, relative humidity and horizontal wind fields towards a steady-state background profile. The specified relaxation is weak in the lower troposphere, allowing the model to simulate its own boundary layer, land atmosphere exchange and convection; and strong in the upper troposphere and stratosphere, representing the strong forcing due to upper-level advection. Unlike previous studies we simulated the long-term behavior of the system by coupling a soil model to the atmospheric model and including all relevant parameterizations. The soil water content was relaxed more strongly in the deeper layers to balance gravitational runoff. We let the model run into its equilibrium and evaluated the diurnal convection in this state of diurnal equilibrium.

Our idealized CRM produces a realistic timing of the diurnal cycle of precipitation, convective mass fluxes, clouds and heat fluxes. Sensitivity tests on the response of convection to variations in static stability show deeper convection in a more unstable environment. This is in agreement with buoyancy principles and earlier studies. The onset and peak of precipitation are also shifted to later times in a more unstable environment, but precipitation amounts remain nearly unchanged.

A change of the prescribed humidity profile however has only negligible impacts on convection. In the model equilibrium the convection (determined by the static stability) and evaporation determine the moisture content of the lower atmosphere. This moisture content then regulates the timing and intensity of the diurnal convection. As a result, the external specification of the humidity profile is not necessary. If, however, the lower troposphere is constrained to the large-scale profile by using a height-independent relaxation, a strong sensitivity of convection to humidity is observed, with a later onset of convection, a shift of cloud-bases to higher levels, and less precipitation in a drier atmosphere. Dry air entrainment delays the transition of shallow into deep convection considerably. This sensitivity has been described in earlier studies, but we consider it as an internal element of the interactively coupled system.

The idealized cloud-resolving framework developed here applies to multi-day episodes of diurnal convection over large-scale landmasses under weak synoptic forcing. We
conclude from our simulations that the stability of the atmosphere is then a decisive factor regulating the strength of convection and the timing and peak amount of precipitation, while the moisture content of the atmosphere should be considered as an interactive variable rather than a control parameter.

Acknowledgments. This project has been partially funded by the Swiss National Science Foundation through NCCR Climate. The necessary computer resources for the simulations were provided by the Swiss National Supercomputing Center (CSCS) in Manno, partially in the framework of the Swiss-Alps grant. We furthermore wish to thank Daniel Lüthi and Oliver Fuhrer for giving advice using the CCLM.
Chapter 3

Diurnal equilibrium convection and land surface-atmosphere interactions

Linda Schlemmer\textsuperscript{1}, Cathy Hohenegger\textsuperscript{2}, Jürg Schmidli\textsuperscript{1}, Christoph Schär\textsuperscript{1}

\textsuperscript{1}:Institute for Atmospheric and Climate Science, ETH Zurich, Switzerland
\textsuperscript{2}:Max-Planck-Institute for Meteorology, Hamburg, Germany

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Abstract

The influence of soil moisture and atmospheric stability on mid-latitude diurnal convection and land-atmosphere exchange is investigated in an idealized cloud-resolving modeling framework using a full set of parameterization schemes. In each member of a series of month-long experiments, the model attains a state where deep, precipitating convection is triggered every day. This state is referred to as equilibrium diurnal convection. It resembles the equilibrium behaviour of the land-atmosphere system and could be thought of as a substitute for climate simulations. The triggering occurs via different mechanisms depending on the atmosphere-soil setting. In our framework latent heat fluxes comprise the primary control over the precipitation amounts. We find that evaporation is regulated by the availability of energy on the one hand and the availability of soil moisture and the near-surface saturation deficit of the atmosphere on the other hand. Increased cloud cover over wet soils reduces net shortwave radiation but increases net longwave radiation leading to a near-compensation of the two effects on available energy. Increased boundary-layer moisture is removed by deep convection thus reducing the near-surface saturation deficit and preventing a negative feedback of boundary-layer moisture content on the latent heat fluxes. We also find that there is a spatial correlation between soil moisture and precipitation anomalies, suggesting that the soil-moisture precipitation feedback acts on a scale of 10-50 km. Overall, evaporation from the soil increases with increasing soil moisture for all simulations, leading to a positive soil-moisture precipitation feedback. This finding is robust among different soil types, wind speeds as well as different atmospheric humidity and stability profiles. The simulated predominance of a positive soil-moisture precipitation feedback applies to horizontally homogeneous quasi-steady conditions, while further work will be needed
to address more transient and inhomogeneous situations.

3.1 Introduction

The interaction of the land surface with the atmosphere is an important element of the climate system and includes multiple processes and feedback mechanisms. It is well known that drought or precipitation events may be amplified by persistent soil moisture anomalies (Enthekabi et al., 1992, Seneviratne et al., 2006, Fischer et al., 2007a,b). Whether a certain state of the soil can persist over longer time scales is controlled by the strength of the land-atmosphere coupling. In regions where soils are close to saturation, evapotranspiration (ET) is controlled by the available net radiation. In dry regions in contrast, the availability of soil moisture controls ET and anomalies are able to persist over longer timescales (Koster et al., 2004, Dirmeyer et al., 2009). Concerning the availability of net radiation, on the one hand, an increase in cloud cover over wet soils reflects incoming radiation, thereby acting as a negative control on net shortwave radiation. On the other hand, net outgoing longwave radiation is trapped by larger cloud and water vapor amounts, acting as an amplifying mechanism. The overall effect is thought to be slightly positive (see e.g. Eltahir, 1998, Schär et al., 1999).

The question, whether the occurrence of precipitation is favored over wet or dry soils is of outermost importance for the climate. Using data from Illinois, D’Odorico and Porporato (2004) showed that soil moisture is linked to the occurrence of subsequent precipitation but not subsequent precipitation amount, and that the land-atmosphere system tends to be locked in either a wet, or a dry phase, whereas Alfieri et al. (2008) detected a relatively weak feedback between soil moisture and the frequency of occurrence of precipitation, that could both be positive or negative. From a water budget perspective one would nevertheless argue that more ET must lead to larger precipitation amounts on large space- and time scales. On small scales precipitation occurrence and amount is however not only controlled by the amount of water that is evaporated into the atmosphere but also by large-scale processes and numerous boundary-layer processes and interactions among them. A stabilization of the atmosphere over wet soils can for instance result in a negative soil moisture precipitation feedback. Over Southern Africa it was found that wet soils can lead to a stronger atmospheric stratification and to the formation of anticyclonic circulations that induce subsidence and divergence at the surface. The subsiding motion suppressed convective activity and precipitation, representing a negative soil moisture precipitation feedback (Cook et al., 2006). For summertime moist-convection over land in Europe a similar stabiliziation of the atmospheric profile over wet soil was described in Hohenegger et al. (2009). They furthermore found differ-
ent signs of the soil moisture-precipitation feedback in simulations using parameterized convection and simulations using explicitly resolved convection. A peculiar feature of their simulation was the presence of a stable layer that developed in the convection-permitting simulations over wet soils and inhibited the triggering of deep convection, resulting in reduced precipitation amounts over wet soils.

On seasonal time scales mostly a strong positive correlation between soil-moisture and evaporation and a positive correlation between soil-moisture and atmospheric recycling ratio, implying a positive soil-moisture precipitation feedback was found for Western Africa (van den Hurk and van Meijgaard, 2010).

Whether precipitation occurs or not is often decided by the triggering of deep convection. As described in Schär et al. (1999), the growth of the PBL will be slower over wet soils than over dry soils, concentrating the moist entropy flux into a shallower layer leading to higher values of convective available potential energy. Findell and Eltahir (2003a) explored different triggering mechanisms over dry soils, where boundary layer growth will occur, and wet soils, where a moistening of the atmosphere leads to a fall of the level of free convection (LFC). The stability and the humidity content of the atmosphere determine if convection is triggered. Strongly sheared winds or winds that show a backing with height, leading to cold air advection in the lower troposphere can furthermore suppress the triggering of convection (Findell and Eltahir, 2003b).

At small spatial scales inhomogeneities in the land cover (Brown and Arnold, 1998) or soil moisture inhomogeneities can have a decisive influence on the development of convective events (Emori, 1998). Taylor et al. (2010) investigated the importance of anomalies of the land surface in the generation of convergence zones that can lead to the initiation of storms in the Sahel zone.

A further important aspect for precipitation to occur is the formation of clouds. Ek and Mahrt (1994) and Ek and Holtslag (2004) focused on the development of clouds at the top of the planetary boundary layer (PBL) and found a more rapid increase of the relative humidity at the top of the PBL over drier soils and weak atmospheric stratifications due to the rapid deepening of the PBL. In the case of a strong stratification of the atmosphere, wet soils lead to larger relative humidities at the PBL top via a direct moistening of the PBL.

Turbulent fluxes from the surface, which link the soil to the atmosphere can moreover strongly be modified by several boundary-layer processes. The importance of freetropospheric humidity content and dry-air entrainment on the evolution of surface heat-
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fluxes and the evolution of the PBL has been pointed out by Santanello et al. (2007) and van Heerwarden et al. (2009). The entrainment of dry air promoted by a growth of the PBL acts to dry the PBL and thereby to enhance latent heating. The importance of a developing nocturnal residual layer under dry free-tropospheric conditions in the formation of droughts is pointed out by Santanello et al. (2007). Most of these studies focused on a short time-scale and investigated merely whether cloud formation occurred or precipitation was triggered. They did not investigate resulting impacts upon precipitation amounts or the behavior over several days.

In our study we analyze the interplay between the surface, the convective boundary layer and the deep atmosphere over numerous diurnal cycles in an idealized cloud-resolving model with a full set of parameterization. We investigate the longer-time behavior of the system, where the atmosphere is in equilibrium with a specific soil moisture distribution. Our focus lies on the evolution of the diurnal cycle of moist convection with different soil moisture contents and atmospheric stabilities. Our modelling framework has been introduced in Schlemmer et al. (2011b) and used to investigate the sensitivity of the diurnal cycle of moist convection to atmospheric moisture and stability. It was found that in a certain regime, referred to as diurnal equilibrium, the prescribed moisture content of the atmosphere has negligible influence, as it is determined by the convection itself. The stability of the atmosphere determines the depth of the evolving convection and the timing of the precipitation peak. The total amount of precipitation however is largely unchanged as it is controlled by ET.

In the current study we expand the work of Schlemmer et al. (2011b) by additionally varying the soil moisture content and thereby ET. The simulations mimic the conditions during convectively active flat-pressure synoptic situations, that may persist over several days in mid-latitude regions, as for example in July 2006 over mid-Europe (Hohenegger et al., 2008b) or in summertime in the Southeastern United States (Garrett, 1982). Sensitivity of precipitation to soil moisture appears to be large in such conditions.

We concentrate on the role of the availability of soil moisture for ET versus the role of available radiation and near-surface humidity in the soil moisture-precipitation feedback. The paper is organized as follows: in section 3.2 we review the framework and the model used. In section 3.3 we give a short theoretical consideration on what we expect for the soil moisture-precipitation feedback in the diurnal equilibrium and present simulations with varied parameters. We furthermore conduct sensitivity studies on the specific model setup used. In section 3.4 we discuss the limitations of our framework and in 3.5 we summarize the work performed.
3.2 Experimental set-up

The numerical framework used has been introduced in Schlemmer et al. (2011b) and we briefly summarize the most important concepts here.

The framework applies to situations where the soil-atmosphere system reaches a state of diurnal equilibrium. In the real world flat-pressure synoptic situations, where a state of diurnal equilibrium evolves can be present over a period of 1-2 weeks and our framework can be thought to describe an asymptotic limit to these situations. The framework is designed in such a way, that the water balance is maintained by the system. This implies that any increase of ET ultimately increases the atmospheric moisture content. Therefore, it can not represent situations, where a replacement of the humidified air masses by large-scale processes occurs.

3.2.1 Model description

The simulations are performed with version 4.0 of the COSMO-CLM (Consortium for Small-Scale Modeling Model in Climate Mode, hereafter CCLM). The CCLM is a versatile limited-area atmospheric modeling system including a whole suite of model parameterizations (Steppeler et al., 2003, Doms and Förstner, 2004, Baldauf et al., 2011). It is based on the non-hydrostatic compressible atmospheric equations, uses the split-explicit time-stepping scheme (Klemp and Wilhelmson, 1978a, Wicker and Skamarock, 2002), and is suited for applications with horizontal grid-spacings from about 100 m to 100 km. The model is run with a full set of parameterizations, except for convection which is explicitly resolved. For more details on the employed parameterizations, the reader is referred to Schlemmer et al. (2011b). We use a horizontal grid-spacing of $\Delta x = 2.2$ km with $100 \times 100$ grid points in the horizontal and 50 vertical levels. Double-periodic lateral boundary conditions are employed. The Earth’s rotation is neglected as in many other cloud-resolving modeling studies (e.g. Su et al., 1999).

3.2.2 Framework

Atmosphere

We simulate diurnal convection over land, where moist convection interacts with the large-scale synoptic forcing as well as boundary-layer, radiative and land-surface processes. In our modeling framework the large-scale forcing is represented by relaxing the simulated atmosphere towards an externally prescribed profile, while the other processes are explicitly simulated.

The elevation of the surface is set to 489.0 m corresponding to the altitude of Munich. Incoming solar radiation is determined according to $48.25^\circ$ N and $0^\circ$ E on 12th July.
throughout the whole simulation. The model is initialized with a single vertical sounding prescribing the variables pressure, temperature, specific humidity and the horizontal wind components. White noise is applied to the initial temperature at the lowest layer with a maximum amplitude of ±0.02 K, to break the symmetry of the initial state. The model is run for 30 days, where day 16-30 of the simulation are used for the evaluation.

We relax the mean model state towards the prescribed atmospheric profiles using a height-dependent strength of the relaxation with weak (strong) relaxation in the lower (upper) troposphere. Similarly, soil moisture is relaxed weakly (strongly) in the upper (lower) soil (see below). This set-up enables the soil-atmosphere interface to develop freely i.e. the diurnal boundary layer and convection develop in response to the solar forcing and are not strongly affected by the relaxation.

The height-dependent relaxation is implemented by an additional term that is added to the right-hand side of the prognostic equations.

\[
\frac{\partial X}{\partial t} = \left(\frac{\partial X}{\partial t}\right)_{\text{physical terms}} - \frac{X - X_{ref}}{\tau} \cdot f(p),
\]

(3.1)

where \(X\) stands for temperature, specific humidity or zonal or meridional wind, \(X_{ref}\) are the values of the reference-profile, \(\bar{X}\) domain-mean values of the predicted variables and

\[
f(p) = 0.5 \cdot (1 + erf\left(\frac{p_0 - p}{b}\right)),
\]

(3.2)

where \(p\) is pressure and \(erf\) the error-function defined as: \(erf(x) = \frac{2}{\sqrt{\pi}} \int_0^x e^{-t^2} dt\). We set \(p_0\) to 500 hPa and \(b\) to 300 hPa. A relaxation parameter \(\tau\) of 1 day is chosen implying relaxation times of 2 days and 61 days at the 500 hPa and 950 hPa level, respectively. The same profiles are used as input and reference profiles.

The profiles used to initialize the atmosphere are shown in Fig. 3.1a. To investigate the role of the temperature stratification for the land-atmosphere coupling, three different profiles for temperature are used:

**STABLE:** \(dT dz^{-1} = -0.6\) K (100 m)\(^{-1}\)

**CTL:** \(dT dz^{-1} = -0.7\) K (100 m)\(^{-1}\)

**UNSTABLE:** \(dT dz^{-1} = -0.8\) K (100 m)\(^{-1}\)

Values for convective available potential energy (CAPE) and convective inhibition (CIN) for the input profiles are given in Table 3.1.
Since the relaxation is strong in the upper troposphere and tropopause region, upper-tropospheric temperature stays distinct between simulations using different stabilities, but equal for simulations with differing soil moisture but the same atmospheric stability. In the lower troposphere the model is able to simulate its own state, resulting in diurnal cycles depending upon soil moisture content and upper-tropospheric stability and humidity (compare Fig. 4 in Schlemmer et al., 2011b).

**Soil**

To simulate the interactions between the atmosphere and the underlying soil, the multi-layer soil model TERRA_ML after Heise et al. (2003) with 10 layers is used with a total depth of 11.5 m. Layer thickness varies from 1 cm at the surface to 7.5 m in the deep soil. The upper seven soil layers down to a depth of 1.47 m are hydrologically active. The lowest three layers are climatic layers. The soil type is set to loam. Vegetation is prescribed by specifying a leaf area index of 2.96, a plant cover of 0.84 and a root-depth of 0.56 m. The surface roughness is set to 0.04 m. These settings represent typical values for European conditions. The land-surface scheme considers infiltration,
3 Land surface-atmosphere interactions

<table>
<thead>
<tr>
<th>Simulation name</th>
<th>dT/dz</th>
<th>RH_{surf}</th>
<th>RH_{ut}</th>
<th>CTP</th>
<th>HI_{low}</th>
<th>CAPE</th>
<th>CIN</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>K/100 m</td>
<td>%</td>
<td>%</td>
<td>J kg(^{-1})</td>
<td>K</td>
<td>J kg(^{-1})</td>
<td>J kg(^{-1})</td>
</tr>
<tr>
<td>STABLE</td>
<td>-0.6</td>
<td>70</td>
<td>40</td>
<td>67.7</td>
<td>9.88</td>
<td>0.0111</td>
<td>0.00</td>
</tr>
<tr>
<td>CTL</td>
<td>-0.7</td>
<td>70</td>
<td>40</td>
<td>132</td>
<td>9.81</td>
<td>344</td>
<td>31.1</td>
</tr>
<tr>
<td>UNSTABLE</td>
<td>-0.8</td>
<td>70</td>
<td>40</td>
<td>196</td>
<td>9.75</td>
<td>1740</td>
<td>10.3</td>
</tr>
<tr>
<td>STABLE_WET</td>
<td>-0.6</td>
<td>90</td>
<td>50</td>
<td>68.0</td>
<td>2.57</td>
<td>311</td>
<td>2.58</td>
</tr>
<tr>
<td>STABLE_DRY</td>
<td>-0.6</td>
<td>50</td>
<td>30</td>
<td>67.3</td>
<td>19.4</td>
<td>0.0115</td>
<td>0.00</td>
</tr>
</tbody>
</table>

Table 3.1: Lapse rate of temperature (K/100 m), surface relative humidity (RH_{surf}, %), upper tropospheric relative humidity (RH_{ut}, %), convective triggering potential (CTP, J kg\(^{-1}\)), humidity index (HI_{low}, K), CAPE (J kg\(^{-1}\)) and CIN (J kg\(^{-1}\)) for the input/reference profiles used.

Percolation, capillary movement, melting and freezing of snow, ET and runoff. ET, which is of key importance in controlling the feedback between soil moisture and convection, takes into account three source terms: bare soil evaporation, evaporation from the interception reservoir and transpiration from vegetation. The formulation closely follows the Biosphere-Atmosphere Transfer Scheme (BATS) (Dickinson, 1984). Potential evaporation is determined using a drag-law approach:

\[
E_{\text{pot}}(T_{sfc}) = \rho C_d^d q |v_h| (Q(T_{sfc}) - q),
\]  

where \(T_{sfc}\) is the skin temperature, \(\rho\) the density of the atmosphere at the lowest atmospheric layer, \(C_d^d\) the bulk-aerodynamic coefficient for turbulent moisture transfer at the surface, \(q\) the specific humidity at the lowest atmospheric layer and \(Q(T_{sfc})\) the saturation specific humidity at the surface. Bare soil evaporation \(E_b\) is then determined using a supply and demand approach, where the smaller term of either the maximum flux of moisture \(F_m\) that the soil can sustain, or the potential evaporation is used. \(F_m\) is mainly a function of the average soil water content.

Evaporation from the interception reservoir is determined according to

\[
E_i = \min\left[\frac{\rho_w}{\rho_w - \rho q} W_i; f_i E_{\text{pot}}(T_{sfc})\right],
\]

where \(\rho_w\) is the density of water, \(W_i\) the water depth of the interception store and \(f_i\) the partial coverage of the surface by interception water.

The biophysical control of plants on transpiration is considered via the stomatal resistance. For the formulation of the stomatal resistance \(r_s\), the multiplicative formulation by Jarvis (1976) is utilized. It takes into account the influence of radiation, water, temperature and atmospheric humidity by the \(F\) functions as

\[
r_s^{-1} = r_{\text{max}}^{-1} + \left( r_{\text{min}}^{-1} - r_{\text{max}}^{-1} \right) \left[ F_{\text{rad}} F_{\text{wat}} F_{\text{tem}} F_{\text{hum}} \right],
\]

50
with the parameters $r_{\text{min}} = 150\, \text{sm}^{-1}$ and $r_{\text{max}} = 4000\, \text{sm}^{-1}$. The functions $F$ are 1 for optimal conditions and decrease to 0 for unfavorable conditions for plants. Most important for the following studies are the radiation function:

$$F_{\text{rad}} = \min \left( 1, \frac{\text{PAR}}{\text{PAR}_{\text{crit}}} \right),$$

where $\text{PAR}$ is the photosynthetically active radiation and $\text{PAR}_{\text{crit}} = 100\, \text{W m}^2$ is a tuning parameter and the dependency on soil water:

$$F_{\text{wat}} = \max \left[ 0, \min \left( 1, \frac{w_{\text{l,root}} - w_{\text{PW}P}}{w_{\text{TLP}} - w_{\text{PW}P}} \right) \right],$$

where $w_{\text{l,root}}$ is the liquid water content fraction of the soil averaged over the root depth and $w_{\text{PW}P}$ the permanent wilting point. The turgor loss point $w_{\text{TLP}}$ is parameterized after Denmead and Shaw (1962):

$$w_{\text{TLP}} = w_{\text{PW}P} + (w_{\text{FC}} - w_{\text{PW}P}) \cdot (0.81 + 0.121 \arctan(E_{\text{pot}}(T_{\text{sfc}}) - E_{\text{pot,norm}})),$$

where $E_{\text{pot,norm}} = 4.75\, \text{mm day}^{-1}$. In the 60CTL simulation $E_{\text{pot}}(T_{\text{sfc}}) \approx 13\, \text{mm day}^{-1}$ averaged over the diurnal cycle which is relatively far away from $E_{\text{pot,norm}}$. The turgor loss point of plants is therefore close to the field capacity.

In a regional climate model setup the CCLM was found to adequately simulate latent heat fluxes, but to underestimate sensible heat fluxes, leading to a bias in the Bowen ratio. This bias has partly been attributed to an underestimation of incoming solar radiation, due to an over-prediction of the cloud cover (Jaeger et al., 2009, Davin et al., 2011) and an unrealistically high aerosol optical depth (Zubler et al., 2011).

To initialize soil moisture an idealized profile is used, which increases quadratically from the surface value to saturation at a depth of 2.50 m. To assess the impacts of soil moisture on the diurnal cycle of convection, simulations with relative surface soil moisture of 20, 40, 60 and 80 %, respectively, are performed. The different input profiles for soil moisture saturation are shown in Fig. 3.1. For the loam soil $S=20\%$ lies below the permanent wilting point, $S=40\%$ and $60\%$ are situated between the permanent wilting point and the field capacity and $S=80\%$ is above the field capacity. To restore water that is evaporated to the atmosphere or lost by surface and groundwater runoff, a relaxation equivalent to that used in the atmosphere is performed on soil moisture with a time constant of $\tau_{\text{soil}} = 2$ days in the deep soil. The relaxation is weakest close to the surface and increases to full strength at the lowest hydrologically active layer (cf.
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Figure 3.2: (a) Simulated soil moisture saturation (%) and (b) soil temperature (K) for 40_CTL (dashed-line), 60_CTL (solid line) and 80_CTL (dash-dotted line) at the uppermost layer (black lines) and the lowest hydrologically active layer at a depth of 1.47 m (grey lines).

Schlemmer et al., 2011b). The concept here is comparable to that in the atmosphere: we wish to keep the model in the deep soil close to a specified state but allow the soil-atmosphere system to interact via surface fluxes of latent and sensible heat, boundary-layer processes and deep convection. The soil moisture profile attains an equilibrium state with the atmosphere. Soil moisture and soil temperature for day 16-30 of the simulations using the CTL atmospheric stability are shown in Fig. 3.2. The upper soil layers show a pronounced diurnal cycle, the lower layers a considerably weaker one. Soil temperature shows a very weak trend of about 1 K per 15 days in the lowest active layer.

3.3 Results

Our goal is to understand the feedbacks and relevant processes in the coupled land surface - atmosphere system for different soil moisture contents and atmospheric stabilities. Therefore we perform simulations with combined changes in atmospheric stability (STABLE, CTL and UNSTABLE, cf. section 3.2.2) and volumetric soil water contents (20, 40, 60 and 80 %, cf. section 3.2.2), resulting in a set of 12 simulations. The naming of the simulations is as follows: the soil moisture is combined with the atmospheric stability, 60_STABLE for example means, that $S=60\%$ and the STABLE atmospheric profile is used. The simulations are summarized in Table 3.2. To infer, how long it takes the system to reach equilibrium we perform a set of simulations, where the system is first run to diurnal equilibrium, and then an instantaneous perturbation is applied. The
3.3 Results

Perturbations consist of a change from one to another atmospheric reference stability profile, or from one to another soil moisture content (e.g. S=60% to S=40 or 80% saturation). Spatial inhomogeneities are thereby preserved across this transition. It takes the disturbed simulations about 5-10 days to recover a state of diurnal equilibrium, depending on the variable considered and the change applied. Precipitation adapts over time scales of 5-10 days to the new state, whereas near-surface temperature and humidity requires 8-10 days. The system adapts more quickly to changes in soil moisture than to changes of the atmospheric stability.

3.3.1 Budget considerations

We shortly review what we expect from a theoretical perspective for the soil moisture-precipitation feedback in the diurnal equilibrium in our idealized framework. The energy balance at the surface reads

\[ SW + LW + E + H + G = 0, \]  

(3.7)

where \( SW \) is the surface net shortwave radiation, \( LW \) the surface net longwave radiation, \( E \) the latent heat flux, \( H \) the sensible heat flux and \( G \) the ground heat flux. \( G \) is approximately zero averaged over the diurnal cycle, as soil temperatures show only

<table>
<thead>
<tr>
<th>Simulation name</th>
<th>precipitation sum [mm d(^{-1})]</th>
<th>( Q ) [W m(^{-2})]</th>
<th>( E ) [W m(^{-2})]</th>
<th>( H ) [W m(^{-2})]</th>
<th>( E/L_{\text{v}} - P ) [mm d(^{-1})]</th>
</tr>
</thead>
<tbody>
<tr>
<td>20_STABLE</td>
<td>0.00</td>
<td>82</td>
<td>17</td>
<td>65</td>
<td>0.58</td>
</tr>
<tr>
<td>40_STABLE</td>
<td>1.94</td>
<td>149</td>
<td>106</td>
<td>42</td>
<td>1.74</td>
</tr>
<tr>
<td>60_STABLE</td>
<td>2.88</td>
<td>147</td>
<td>123</td>
<td>24</td>
<td>1.36</td>
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<tr>
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<td>150</td>
<td>134</td>
<td>16</td>
<td>1.21</td>
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<tr>
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<td>12</td>
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<tr>
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<td>104</td>
<td>41</td>
<td>1.58</td>
</tr>
<tr>
<td>60_CTL</td>
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<td>129</td>
<td>30</td>
<td>1.11</td>
</tr>
<tr>
<td>80_CTL</td>
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<td>159</td>
<td>141</td>
<td>18</td>
<td>1.06</td>
</tr>
<tr>
<td>20_UNSTABLE</td>
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<td>9</td>
<td>87</td>
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<tr>
<td>40_UNSTABLE</td>
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<td>97</td>
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<tr>
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<td>165</td>
<td>144</td>
<td>21</td>
<td>1.00</td>
</tr>
</tbody>
</table>

Table 3.2: Mean daily precipitation amount (mm day\(^{-1}\)), available energy (\( Q \), W m\(^{-2}\)), latent (\( E \), W m\(^{-2}\)), sensible heat flux (\( H \), W m\(^{-2}\)) and difference between evaporation and precipitation (mm day\(^{-1}\)) for the main set of simulations.
a negligible trend after the first 15 days (see Fig. 3.2b). For the water budget of the atmosphere $\frac{\partial W_{atm}}{\partial t}$ we can write:

$$\frac{\partial W_{atm}}{\partial t} = \frac{E}{L_v} - P + R,$$

(3.8)

where $L_v$ is the latent heat of vaporization, $P$ is the surface rain rate and $R$ the moisture tendencies due to the relaxation. In the equilibrium, averaged over the diurnal cycle and the entire domain, the budget of the atmosphere is approximately closed ($\frac{\partial W_{atm}}{\partial t} \approx 0$). The values for the relaxation tendencies $R$ amount to roughly 20% of ET and the relaxation acts to dry the atmosphere. In the equilibrium, the remaining 80% of the evaporated water needs to be converted into precipitation, as in our framework there must be an approximate balance between ET and P.

For low soil moisture content, ET is below its potential rate $E_{pot}$ and a decrease of soil moisture leads to a decrease of ET. Therefore, a decrease of soil moisture must decrease precipitation, leading to a positive feedback. For high soil moisture content (where the soil meets atmospheric demand and ET is at its potential rate) the only possible mechanism to end up in a negative soil moisture precipitation feedback in diurnal equilibrium is to decrease potential evaporation. This could happen via two different pathways. The first one is to reduce the available energy at the surface $Q = R_n - G$, e.g. through cloud processes. The second pathway is to increase the moisture content of the boundary layer, thereby reducing the saturation deficit and $E_{pot}$. We will elucidate the role of the mentioned mechanisms for our idealized framework in the following section. It should be stressed that the current simulations can not be directly compared to the simulations of Hohenegger et al. (2009). An increase (decrease) of soil moisture in their cloud-resolving simulations did lead to larger (smaller) latent heat fluxes, but reduced (enhanced) precipitation amounts (see their Figure 6), which is characteristic for a transient phase.

Relaxation terms for the atmosphere are largest for stable atmospheric conditions and for the soil for a soil-moisture saturation of 80% since most runoff occurs here. Without the relaxation, that keeps the soil-atmosphere system in a specified state, the system changed most quickly to another state under such conditions. To infer, how long the system retains the equilibrium state we conducted additional simulations, where the relaxation is switched off after 30 days both in the atmosphere and in the soil. The diurnal cycle on the two following days is compared to the mean diurnal cycle during the phase including relaxation. The result is very similar if the relaxation is switched off. Thus, the additional terms affecting the heat- and moisture budget are small correction
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terms that are applied in order to keep the model from drifting into a different state. They do, however, not affect the dynamic of the system.

3.3.2 Diurnal equilibrium

Figure 3.3 shows the mean diurnal cycle of cloud water, cloud ice and surface rain rate for the set of simulations, averaged over day 16-30 of the simulation. Remarkably, all simulations with $S \geq 40\%$ show deep, precipitating convection. This means that the model gets into a diurnal equilibrium state where enough soil moisture is available to evaporate, trigger convection, condense and precipitate. Reducing $S$ further to 20\% leads to an almost complete shut-off of clouds and precipitation, as the soil simply cannot evaporate enough moisture to form precipitation.

Going from more stable to more unstable atmospheric profiles, the period of rainfall is prolonged and the peak precipitation is shifted to later times. This phenomenon

![Figure 3.3: Mean diurnal cycle of cloud water content (kg kg$^{-1}$, shaded area), cloud ice content (kg kg$^{-1}$, contour lines) and domain mean precipitation (mm h$^{-1}$, black solid line, minimum and maximum values over the 15 days of simulation are indicated by dark grey shading) for the set of 12 simulations. Numbers in the lower left corner indicate mean precipitation amount averaged over day 16-30 (mm day$^{-1}$).](image-url)
has been explained in Schlemmer et al. (2011b) and can be understood by the accumulation of humidity in the PBL under stable conditions. Over wetter soils, precipitation occurrence is additionally shifted to later times, and most importantly, precipitation amounts are increased, which constitutes a positive soil moisture-precipitation feedback. Cloud bases and cloud tops are shifted to lower elevations, as expected from thermodynamic considerations. Also, persistent mid-level clouds (around 700 hPa) begin to appear, especially in the simulations 60_STABLE, 60_CTL, 80_STABLE, 80_CTL and 80_UNSTABLE. Their thickness increases with soil moisture and stability.

In comparison to stable conditions, unstable atmospheric profiles shift clouds tops to higher altitudes, more ice clouds are simulated, precipitation starts to fall later, and the time of the precipitation peak is shifted to later times, as found in Schlemmer et al. (2011b). Changes in stability do not seem to impact the overall sign of the simulated soil moisture-precipitation feedback.

Q, ET and sensible heat fluxes are shown in Fig. 3.4. Q is of comparable magnitude in all simulations with $S \geq 40\%$, with a slight decrease in 60_STABLE and 80_STABLE, where the mid-level clouds reflect incoming radiation, while increased longwave downward radiation offsets this reduction only partly. For $S=20\%$ increased longwave radiation in the dry atmosphere leads to a reduction of the available energy. The first potential

**Figure 3.4:** Upper row: Mean diurnal cycle of available energy ($Q$, W m$^{-2}$) and lower row: mean diurnal cycle of surface sensible (black lines) and latent (grey lines) heat fluxes (W m$^{-2}$) for (a, d) STABLE, (b, e) CTL and (c, f) UNSTABLE for 20 (dotted line), 40 (dashed line), 60 (solid line) and 80 % (dash-dotted line) soil moisture saturation.
pathway into a negative soil moisture-precipitation feedback mentioned in section 3.3.1 is as a consequence not realized, as a significant reduction of available energy occurs over dry, but not over wet soils.

All simulations with \( S \geq 40\% \) have latent heat fluxes that are larger than sensible heat fluxes. The midday Bowen ratios lie 0.2 in 80_STABLE and 0.75 in 40_UNSTABLE. As seen in Fig. 3.4, latent heat fluxes increase with increasing soil moisture as there is more moisture available for ET. Sensible heat fluxes decrease with increasing soil moisture as a larger portion of the available net radiation is passed into latent heating.

Since precipitation must approximately balance evaporation in the diurnal mean in our simulations (see section 3.3.1), the increase in latent heating leads to an increase in precipitation. Mean daily precipitation amounts and latent heat fluxes, averaged over the evaluation period are summarized in Fig. 3.5a and together with the sensible heat fluxes shown in Table 3.2. For all atmospheric profiles, mean daily precipitation amounts increase with increasing soil moisture, indicative of a positive soil moisture-precipitation feedback. The increase is smallest for STABLE, due to the more pronounced cloud cover. The experiments have been repeated with an even more stable atmosphere (\( \Delta T \Delta z^{-1} = -0.5 \text{ K (100 m)}^{-1} \)). Still more clouds build up in these simulations over wet soils, but the feedback remains positive (not shown).

The difference between precipitation and evaporation is largest for \( S=40\% \) and decreases both for higher and lower soil moisture values. Furthermore, it decreases for decreasing atmospheric stability. This means that the atmospheric moisture forcing (i.e. relaxation) is stronger over drier soils and stable atmospheric conditions. In the soil, the relaxation is however largest for wet soils, for which runoff is largest. It increases also

![Figure 3.5: Mean precipitation amount (black, mm day\(^{-1}\)) and evaporation (grey, mm day\(^{-1}\)) for day 16-30 as a function of soil moisture for simulations with (a) different stability, (b) different atmospheric humidities and (c) for additional sensitivity studies. Panel (d) shows mean precipitation amount (black, mm day\(^{-1}\)) and evaporation (grey, mm day\(^{-1}\)) averaged over day 1-3 for the simulations CTL_UNSTABLE.](image)
marginally for stable compared to unstable atmospheric conditions. Figure 3.5d shows mean ET and precipitation averaged over the three days following the transition from CTL to UNSTABLE in transient experiments. ET increases with soil moisture, similar to what is observed in the state of diurnal equilibrium. For precipitation however, mean amounts are reduced over wet soil. This reduction primarily results from a delayed trigger of moist convection over wet soils. Thus, during such transient phases our framework can produce a negative soil-moisture precipitation feedback.

Fig. 3.6 shows domain mean profiles of relative humidity at 0000 and 1200 UTC. The boundary layer is considerably drier over dry soils than over wet soils and the height of the two maxima in relative humidity (the lower one coinciding with the top of the boundary layer, the upper one coinciding with the cloud layer visible in Fig. 3.3) are situated higher for drier soils. This upward shift is also reflected in the upward shift of cloud bases. The increase of PBL moisture content between $S=60$ and 80% reduces the saturation deficit most prominently for STABLE. This reduction is however not sufficient to reduce evaporation and to enter the second pathway into a negative soil moisture precipitation feedback (see section 3.3.1).

### 3.3.3 Spatial pattern of feedback

Looking at spatial patterns of the precipitation amounts, large differences between the simulations are visible. Figure 3.7 shows precipitation accumulated over days 16-30 and soil saturation at a depth of 2.5 cm averaged over the same period. Precipitation shows an inhomogeneous, spotty distribution of elongated patches of increased values with a size of about 20 grid points ($\equiv 44$ km) in the prevailing zonal flow direction. This indicates that locations where precipitation fell during the previous days favor the occurrence of new precipitation. Locations with high precipitation amounts furthermore coin-

**Figure 3.6:** Domain mean profiles of relative humidity at 0000 (black lines) and 1200 UTC (grey lines) averaged over the 15 days of the simulations.
3.3 Results

Figure 3.7: Spatial distribution of precipitation accumulated over day 16-30 of the simulation (color shade, mm) and soil moisture saturation averaged over day 16-30 at a depth of 2.5 cm (contour lines) for the main set of simulations. The blue contours correspond to the 40% (left column), 60% (middle column) and 75.5% (right column) level and the black contour to the 35, 58 and 75% level respectively. There is a correlation between positive soil moisture and precipitation anomalies.

cide with spots of wet soil. Calculated Pearson correlation coefficients between mean precipitation and mean soil-moisture range from 0.67 for 80_CTL to 0.84 for 40_CTL. As described in several studies (e.g. Emori, 1998, Pielke, 2001, Baker et al., 2001, Taylor et al., 2010) soil moisture gradients play an important role in the triggering of convec-
Emori (1998) found a negative soil moisture-precipitation feedback at small spatial scales ($O(10 \text{ km})$) in idealized 2D experiments. Over dry soils, moisture gradients induced local thermal circulations leading to maximum precipitation over dry soil patches. Baker et al. (2001) found a preferred occurrence of heavy precipitation over existing wet soil. In the later stage of a convective cell, the amount of soil moisture furthermore regulates the growth of the cell by reducing the height of the LFC (e.g. Clark et al., 2004, Taylor et al., 2010). (Taylor and Ellis, 2006) in contrast documents a dependency of the feedback on the length-scale of the anomaly with a negative feedback for wet patches with an elongation of at least 74 km. In our simulations possessing a strong pattern, we find that thermal circulations develop with subsidence over negative soil moisture gradients along the flow direction and rising motion over positive soil moisture gradients. Over wet patches, the stratification is more stable and CIN values are larger. The triggering of convective cells occurs in conjunction with the rising motion along the positive soil moisture gradients. The triggered convective cells are advected with the flow and precipitate, when they pass over a wet patch. In our simulations, the aggregation of water in preferred locations is most pronounced in the 40.STABLE simulation, where the spatial standard deviation of precipitation is almost as large as the mean value. The spatial variability decreases both for more unstable atmospheres and for wetter soils. This indicates that the described mechanisms are especially important in semi-arid regions and under a strong stratification. Overall Fig. 3.7 shows that there is not only a domain-mean positive precipitation feedback, but that spatial variations develop that support the existence of a local positive feedback acting at scales of $O(40 \text{ km})$.

### 3.3.4 Precipitation intensity

Not only mean precipitation amounts but also precipitation intensities are affected by the stability and the soil moisture. Figure 3.8 shows the corresponding histograms of hourly precipitation sums. Since the simulations over dry soils trigger precipitation less often (independently of the atmospheric stability), we normalized Fig. 3.8 by the total number of rainy grid points ($\geq 0.5 \text{ mm h}^{-1}$). Simulations over domain-mean wet soils show strongest intensities. Over wet soils more water is stored in the atmosphere to form precipitation (see Table 3.3). Over dry soils, precipitation more commonly falls through dry air and more precipitation may evaporate. This is also reflected in stronger convective downdrafts over dry soils (not shown). With respect to the different stabilities, precipitation intensities are highest for the STABLE profile. CAPE values are increased and CIN values are decreased, thus facilitating the higher rain rates.
Table 3.3: Equilibrium values of convective triggering potential (CTP, J kg\(^{-1}\)) at 0600 UTC, humidity index (HI\(_{\text{low}}\), K) at 0600 UTC, CAPE (J kg\(^{-1}\)) at 1100 UTC, CIN (J kg\(^{-1}\)) at 0600 UTC and precipitable water averaged over the diurnal cycle (mm). The first two characters of the simulation names refer to the volumetric soil-moisture profiles depicted in Fig. 3.1b.

### 3.3.5 Triggering of convection

Figures 3.3-3.7 underline the existence of a positive soil-moisture precipitation feedback in our system as expected from balance considerations (see section 3.3.1). It is interesting to compare this against predictions based on thermodynamic indices. As pointed out by Findell and Eltahir (2003a), deep convection is triggered when the descending LFC meets the top of the PBL. This has been confirmed by Schlemmer et al. (2011b). Over wet soils this is achieved by a direct moistening of the atmosphere through latent heat fluxes, leading to an increase of the equivalent potential temperature \(\theta_e\) in the PBL and a resulting lowering of the LFC. Over dry soils, the soil cannot meet the atmospheric demand and the available energy will be passed into sensible heat flux, leading to a warming and subsequent deepening of the PBL. Whether LFC fall or PBL growth is more effective in triggering deep convection depends on the stability and moisture content of the lower troposphere.

The height of the LFC, lifted condensation level (LCL) and PBL and CIN values of our simulations are shown in Fig. 3.9. For each soil moisture and stability class, the model
Figure 3.8: Logarithmic histogram of hourly precipitation sums collected at each grid point of the domain over the time period of day 16-30 for (a) STABLE, (b) CTL and (c) UNSTABLE for \( S = 40\% \) (red), 60\% (black) and 80\% (blue). Bins are 0.5 \( \text{mm h}^{-1} \) wide. Results are normalized with the number of rainy grid points.

attains a state, where deep convection is triggered. Looking at the time evolution of LFC and PBL height, we see that for drier soils boundary layer growth is favored, whereas for more moist soils the descent of the LFC dominates. In a more unstable atmosphere, the LFC is brought down more efficiently than in a stable atmosphere. The triggered convection then redistributes accumulated PBL moisture in the atmosphere, rendering a decrease of ET with increasing soil moisture impossible in our framework as described in section 3.3.1.

To distinguish between different regimes of soil-atmosphere interaction, Findell and Eltahir (2003a) introduced the convective triggering potential (CTP) humidity index (HI<sub>low</sub>) framework. The CTP is a measure of the stratification of the atmosphere between 100 and 300 hPa above ground and HI<sub>low</sub> is a measure of the dryness of the lower atmosphere. Large CTP values indicate an atmosphere that is close to dry adiabatic and areas showing high sensible heat flux are favorable for triggering convection. Smaller but still positive CTP values indicate a lapse rate closer to moist adiabatic and areas of high latent heat flux are advantageous for triggering of convection. Calculated values for the input profiles used in this study are given in Table 3.1. The CTP is in all cases positive, indicating that the atmosphere is sufficiently unstable for deep convection to occur. The HI<sub>low</sub> values are \( \approx 9.8 \) K, putting the soundings just at the border between the regimes where wet and dry soils dominate. Corresponding values for the diurnal equilibrium phase, computed at 0600 UTC, are given in Table 3.3. CTP values increase with stability and decreasing soil moisture while HI<sub>low</sub> increases with drier soils and decreasing stability. In all cases, the obtained values would predict a positive soil-moisture feedback, in agreement with our simulations.
3.3 Results

Figure 3.9: Height of the domain mean value (solid line) and 10th and 90th percentile (dashed lines) of the LCL (light grey line), the LFC (black line) and the boundary layer height (dark grey line) for the set of 9 simulations.

3.3.6 Additional sensitivity experiments

Above we have found that our simulation strategy yields a positive soil moisture precipitation feedback in all cases investigated. Here we study the sensitivity of this result. To further assess the importance of the evapotranspiration in determining the precipitation response in our diurnal equilibrium framework (see section 3.2), we perform simulations with altered ET as well as different wind speed and soil type. The additional experiments are only performed for the simulations using the STABLE profile, as in these experiments the triggering of convection is most critical. Values for mean precipitation amounts and ET are shown in Fig. 3.5b and c and together with surface fluxes of sensible and latent heat summarized in Table 3.4.

Wind speed has a strong influence on the evaporation of water from the surface into the atmosphere, as it controls turbulence and thereby affects potential evaporation. Hence a set of simulations with the prescribed background wind speeds decreased to 10% of the original value over the whole atmospheric column is performed, named STA-
Table 3.4: Same as Table 3.2 but for the additional sensitivity tests.

3 Land surface-atmosphere interactions

<table>
<thead>
<tr>
<th>Simulation name</th>
<th>precipitation sum [mm d$^{-1}$]</th>
<th>Q [W m$^{-2}$]</th>
<th>E [W m$^{-2}$]</th>
<th>H [W m$^{-2}$]</th>
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<td>148</td>
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<tr>
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<tr>
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<td>147</td>
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<td>15.3</td>
</tr>
</tbody>
</table>

BLE_CALM. Resulting precipitation amounts as a function of soil moisture are shown in Fig. 3.5c by the dashed line. Figure 3.5 also shows the situation for our reference simulations. The new simulations still sustain a positive feedback, but of slightly weaker strength. This is not surprising: the reduction in wind speed reduces ET for a given soil moisture content by transferring the available energy into H, but cannot shut down ET (see section 3.2). Looking at the modeled Bowen ratio, there is hardly any difference between the simulations using different wind speed (not shown). The reduction in modeled boundary-layer wind speed is moreover small (maximum -2 m s$^{-1}$ during the time of maximum convective activity), as it becomes independent of the prescribed background profile due to the height-dependent formulation of the relaxation.

In the next set of simulations, called STABLE_SAND, we change the soil type from loam to sand. For sandy soils, water is evaporated more easily since the hydrologic conductivity parameter controlling bare soil evaporation is a factor 9 larger in sand than loam soil. Furthermore the field capacity $\Theta_{fc}$ of sand is smaller than for loam ($\Theta_{fc,sand}=54\%$, $\Theta_{fc,loam}=75\%$ saturation), meaning that the dependency of transpiration on the available soil water reaches its maximum at smaller volumetric soil water
contents (described by the $F_{\text{wat}}$ function in the Jarvis (1976) approach, cf. eq. 3.4).
Comparing the diurnal cycles of convection, the change from loam to sand soil roughly equals a moistening of the loam soil by 20%. In addition, over sand the simulations with $S=60\%$ and $S=80\%$ look nearly identical, as evaporation is at the potential rate here and cannot be increased any further thus preventing an even further positive soil moisture-precipitation feedback. Precipitation nevertheless increases marginally going from $S=60$ to 80%. For low soil moisture amounts ET decreases and, as expected, precipitation amounts decrease. ET and the precipitation amounts remain nevertheless larger than over loam, as evaporation is limited by the availability of water, which is higher over sand soil.

The formation of mid-level clouds observed in Fig. 3.3 influences transpiration by a reduction of photosynthetically active radiation (PAR). In the formulation of transpiration, this is considered in the function $F_{\text{rad}}$ (cf. eq. 3.4). For PAR values below 100 $\text{W m}^{-2}$ a linear decrease of $F_{\text{rad}}$ towards zero at PAR=0 is assumed (cf. eq. 3.5). To investigate how sensitive our results are to this threshold, we increase the threshold $PAR_{\text{crit}}$ below which transpiration is sensitive to PAR to 200 $\text{W m}^{-2}$ in experiment STABLE\_PAR. Resulting ET and precipitation (Fig. 3.5c, dash-dotted line) are qualitatively in line with the previous results.

In a further simulation, called STABLE\_crsmin we lower the parameter for the minimum stomatal resistance $r_{\text{min}}$ in the formulation of transpiration (cf. eq. 3.4) from 150 s m$^{-1}$ to 50 s m$^{-1}$. This change in model physics is also one of the perturbations applied in the project “shortrange ensemble forecasting system” (SREPS) using the COSMO model (Marsigli, 2009). Lowering $r_{\text{min}}$ should allow for more ET. ET and precipitation for 60\_STABLE\_crsmin and 80\_STABLE\_crsmin are very close. Precipitation increases for $S=60\%$ from 2.88 to 3.35 mm day$^{-1}$, while for $S=80\%$ the increase in precipitation is very small (not shown). The positive soil moisture-precipitation feedback remains, however.

### 3.3.7 Role of atmospheric humidity

As a last sensitivity experiment we conduct two additional sets of simulations using drier (simulations STABLE\_DRY) and wetter (simulations STABLE\_WET) atmospheric humidity profiles. Findell and Eltahir (2003a) pointed out that the humidity content of the lower atmosphere has a strong influence on whether a dry or a wet soil is more efficient in triggering convection. Values for lower-tropospheric humidity were decreased (increased) by 20%, upper-tropospheric humidity by 10% resulting in relative humidity profiles with 90 and 50% for the simulation STABLE\_WET and 50 and 30% for...
3 Land surface-atmosphere interactions

STABLE\_DRY (see Table 3.1). These atmospheric profiles fall in the categories “wet soil advantage” and “dry soil advantage”, respectively, as defined in Findell and Eltahir (2003a). As seen in Fig. 3.5b and Table 3.2, diurnal equilibrium values for precipitation are slightly decreased (increased) for STABLE\_DRY (STABLE\_WET), but the overall increase of precipitation with soil moisture is consistent with the previously observed positive soil moisture-precipitation feedback. ET is nearly identical and nearly independent of the atmospheric humidity. A slight reduction due to the decrease of the saturation deficit in STABLE\_WET can be seen. This relatively low sensitivity of ET to the prescribed relative humidity of the atmosphere has been described in Schlemmer et al. (2011b). It results from the redistribution of moisture in the atmosphere in the state of diurnal equilibrium by ongoing convection.

3.4 Discussion

As seen in the set of simulations, both the atmospheric stability and the moisture content of the soil have a strong influence on the diurnal cycle of convection. In a stable atmosphere, convection tends to be suppressed, but ultimately results in more violent outbreaks of convective activity and higher precipitation rates. In a weakly stratified environment, convection can develop more easily and moisture is quickly redistributed in the atmosphere. Over dry loam soils ($S \leq 60\%$) the formation of clouds and precipitation is limited as the soil cannot meet atmospheric demand. Over wet soil, more moisture is evaporated from the soil, resulting in more clouds and precipitation. An increase of precipitation with increasing soil moisture, i.e. a positive soil moisture-precipitation feedback, resulted in all experiments, and is a robust feature for the current simulation strategy. Puzzling to us is our inability to reproduce a negative soil moisture-precipitation feedback with the current setup, which has been reported in several previous studies.

The current study includes no topography, thus we miss some important processes in the build-up of convection and precipitation. As described in Hohenegger et al. (2009), valley circulations may be amplified or damped by dry and wet soils, respectively. In addition we do not include large-scale water bodies as e.g. in Cook et al. (2006), that modify the patterns of moisture convergence. In our idealized framework we are moreover unable to span the full range of possible interactions. Most importantly we cannot reproduce the influence of large-scale processes such as subsidence that suppresses convection and favors the development of stable layers. For example this is frequently observed in the subsiding branch of the Hadley circulation where over cold ocean sur-
faces the formation of stratocumulus clouds is favored and over continental regions arid regions appear.

A possibility to increase PBL moisture in our framework would be to suppress the triggering of convection by the presence of such stable layers. It is however difficult to realize such a mechanism in the current set up. We did put considerable effort into including subsidence in different ways. In a first approach, we imposed low-level convergence and upper-level divergence onto the horizontal wind field, resulting in subsiding motion with a downward velocity of up to $\approx 2 \text{ mm s}^{-1}$. This subsidence however dominated over the surface fluxes, dried and warmed the atmosphere and reduced precipitation amounts over both wet and dry soils. In a second approach we added vertical advective tendencies resulting from a prescribed subsidence-profile on the tendencies of temperature, specific humidity and horizontal winds directly (cf. e.g. Siebesma et al., 2003). These experiments delivered no directly interpretable answers. In both approaches a stable cloud layer did not evolve, as clouds broke up after a few days and precipitation started to fall. The constant input of energy into the system through radiation and the strong turbulent fluxes from the land surface counteract a strong stabilization.

Moreover, the presented framework addresses diurnal-convection situations in summertime only. Changing weather situations, characterized by transience, are not captured. Soil moisture shows variations only in the upper soil layers and is not fully interactive. Experiments including specific time-dependent advection or temperature gradients are however beyond the scope of this paper, but some pseudo-transience was investigated in some additional experiments.

### 3.5 Summary

The influence of soil moisture and atmospheric static stability on the diurnal cycle of convection and precipitation has been investigated in an idealized cloud-resolving modeling framework in a state of diurnal equilibrium. We think that the framework presented can give helpful insights into the longer-term behavior of the coupled summertime land-atmosphere system. We performed a set of simulations with relative soil moisture saturations of 20, 40, 60 and 80% using atmospheric profiles of different stability (stable, control and unstable). In all simulations with $S \geq 40\%$ deep, precipitating convection is triggered every day.

- In weak and intermediate stable atmospheres very deep convection develops. For these conditions, an increase of soil moisture causes an increase of latent heat...
fluxes and precipitation.

- For strong atmospheric stabilities, an increase of soil moisture leads to the formation of clouds in the morning hours that shield incoming radiation and trap outgoing longwave radiation. The overall influence on the net available surface radiation is a slight increase over wet soils. Thus, latent heat fluxes increase for increased soil moisture leading to larger precipitation amounts. In both cases, the soil moisture-precipitation feedback is therefore positive. Sensitivity experiments on the moisture content of the atmosphere, the soil type used, the wind speed and the parameterization of transpiration confirm the increase of diurnal equilibrium precipitation with increasing soil moisture.

The spatial distribution of precipitation shows very inhomogeneous patterns that are most pronounced over dry soils and for stable stratifications. The occurrence of rainfall is favored over wet soil patches, indicating also a positive feedback at small spatial scales.

Overall, the framework developed gives deeper insights into the longer-term behavior of the coupled summertime land-atmosphere system. The main limitation is the restriction to quasi-steady configurations. Further work will be needed to address the atmospheric response to soil moisture in more transient settings.

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Chapter 4

Sensitivity of mid-latitude diurnal convection to ambient temperature and climate warming

Linda Schlemmer, Jürg Schmidli, Christoph Schär

Institute for Atmospheric and Climate Science, ETH Zurich, Switzerland

(Publication in preparation)

Abstract

Climate models suggest that climate change may imply increases in heavy precipitation events, in some regions even despite reductions in mean amounts. This result has been obtained from models using parameterized convection. Here, the response of mid-latitude, extratropical diurnal convection over land to climate-change induced changes in environmental atmospheric profiles is investigated in an idealized cloud-resolving modeling framework. Specific changes addressed include a mean warming of 3 and 6 K, a stabilization of the atmosphere, and a drying of the soil. The modeling framework mimics the thermodynamic effects on cloud, convection and precipitation that could result from anthropogenic warming, but neglects potential changes in large-scale synoptic climatology.

Under these circumstances, results indicate that changes in the incidence of heavy precipitation depend upon the combined role of the three factors considered. In general, cloud cover increases in warmer climates. Mean precipitation amounts remain unchanged under a warming, whereas they are strongly reduced by a drying of the soil. Concerning high precipitation intensities, a significant increase is simulated for a warming combined with a stabilization of the atmosphere, an effect that may be partly or fully compensated by the presence of a drier soil.
4 Sensitivity to ambient temperature

4.1 Introduction

Concerning precipitation changes accompanying a change of earth’s climate, both thermodynamic and dynamic factors need to be considered. With respect to the thermodynamic factors, the hydrological cycle is expected to undergo an intensification together with the increase of the saturation vapor pressure with temperature of approximately 7 % K\(^{-1}\), leading together with near-constant values of relative humidity to a non-linear increase of the specific humidity with temperature.

Future projections of precipitation predict an increase of precipitation on the global scale and an increase in precipitation extremes that is greater than changes in mean precipitation (Kharin and Zwiers, 2005). For mid-latitude land areas during summertime, a decrease of mean precipitation amounts and at the same time an enhancement of intense precipitation events are projected (Christensen and Christensen, 2003, Frei et al., 2006, Beniston et al., 2007). Observations indicate that hourly precipitation extremes can even exceed expectations from the Clausius-Clapeyron equation for daily mean surface temperatures above 12\(^\circ\)C (Lenderink and van Meijgaard, 2008, Allan et al., 2010).

It is discussed whether this effect stems from the excess of latent heat release in extreme showers (Trenberth et al., 2003, Lenderink and van Meijgaard, 2009, 2010), or from the shift of large-scale, stratiform precipitation to more convective precipitation in this specific temperature range (Haerter and Berg, 2009). Christensen and Christensen (2003) demonstrated by using a regional climate model covering Europe, that although large parts of Europe will undergo reductions in precipitation amounts, increases at high intensities are nevertheless to be expected. Studies indicate that this amplification of the precipitation extremes is strongly underestimated in climate models (Allan and Soden, 2008).

The increase of the potential evaporation in warmer climates and a resulting increase in evapotranspiration (ET) are projected to lead to a drying of the soil caused by the lack of available soil moisture. An increased frequency of droughts are anticipated to be the consequence (e.g. Schär et al., 2004, Seneviratne et al., 2006, Fischer et al., 2007a, Sheffield and Wood, 2008). Following Rowell and Jones (2006) thermodynamic effects such as an earlier and more rapid decline of soil moisture in spring caused by an earlier snowmelt and a positive soil moisture-precipitation feedback in summer are further important processes that dry out the soil.

Projected temperature increases are however not homogeneous across the whole vertical column but are thought to be greater in the upper troposphere, where fractional changes of water vapor are larger and the strong positive water-vapor feedback thus has a larger impact (e.g. Soden et al., 2005). This inhomogeneous warming causes a
stabilization of the atmosphere.

One major factor leading to low reliability of future model projections of summertime mid-latitude convection over land is the use of parametrization schemes for moist convection both in the general circulation models (GCMs) and regional climate models (RCMs) used. In addition, inter-model differences of projections are large (e.g. Frei et al., 2006). The aim of the current study is to get an estimate of the behavior of summertime mid-latitude moist convection over land under increased temperature that is derived independently of forcing boundary data and a parameterization scheme for deep convection. Using an idealized cloud-resolving framework we are able to simulate possible thermodynamic effects of climate change on convection from a process-based perspective. The high-resolution grid-spacing allows an explicit modeling of deep convection. We construct the background profile, in which convection develops, in a way it could look like in an atmosphere affected by anthropogenic temperature change.

4.2 Setup of the experiments

4.2.1 Framework

The setup employed follows the framework introduced in Schlemmer et al. (2011b) and is only shortly reviewed here. The aim is to represent convection interacting with large-scale and radiative forcing as well as land-surface and boundary-layer processes. Convection is modeled in a doubly-periodic domain. The large-scale advective forcing is represented by a relaxation of the simulated atmosphere towards externally specified profiles. The relaxation is thereby weak in the lower troposphere to allow for the full coupling of land-surface and boundary-layer processes. In the upper troposphere the relaxation is stronger, representing the increased synoptic forcing due to stronger advection. Radiation, land-surface and boundary-layer processes are simulating by employing the full set of physical parameterization packages. As described in Schlemmer et al. (2011b) we let the model run into a state of diurnal equilibrium and evaluate day 16-30 of the simulation. In the state of diurnal equilibrium precipitation occurs each day at roughly the same time with small variations of the domain mean rain rates from day to day.

4.2.2 Experiments

The background profile of the reference simulation (“CTL”) uses an atmospheric stability of $\frac{dT}{dz}=-0.7\,\text{K}/100\,\text{m}$ and a relative humidity of 70% in the lower and 40% in the upper troposphere, respectively. Soil moisture saturation is set to 60% at the surface increasing to saturation at a depth of 2.5 m. The background wind profile exhibits a
tropospheric wind shear of about 15 m s$^{-1}$, representative for extratropical mid-latitude summer conditions. The simulation is identical to the “CTL” simulation in Schlemmer et al. (2011b) and the “60_CTL” simulation in (Schlemmer et al., 2011a).

In order to represent potential impacts of climate change, adjustments are made to the environmental profile, which are motivated by large-scale projections of the IPCC (2007b) valid for the northern mid-latitude summer conditions over land. Temperature is increased by +3 K and +6 K throughout the whole atmospheric column in the input/relaxation profile. Simulations are termed “3K” and “6K”. Soil moisture saturation is kept as in the CTL simulation but soil temperatures are increased by 3 respectively 6 K throughout the whole soil column to represent the overall shift of the climate. Note that these changes are effective in the deep soil only.

Observations show, that the lower troposphere has warmed by about 0.16 K and the upper troposphere by about 0.1-0.2 K per decade since 1979 (Karl et al., 2006). To estimate the impact of these changes in the lapse-rate, we perform a set of experiments, where the lower troposphere is warmed by +3 and +6 K as in the first set of sensitivity experiments, but where the upper troposphere (at 200 hPa) is warmed by 3.75 K. A linear interpolation of the imposed warming in-between is assumed. The differential warming corresponds to a reduction of the lapse-rate by 0.0064 K (100 m)$^{-1}$. This is also in the range of predicted lapse-rate changes of Lorenz and DeWeaver (2007), where changes vary however stronger with height. The two simulations with altered lapse rates are termed “3K_lr” and “6K_lr”. Except for a change of the lapse-rate the setup is identical to 3K and 6K.

To reproduce the effect of a possible future summer drying, soil moisture saturation $S$ is decreased by 10% from 60 to 50% in the two simulations called “3K_lr_dry” and “6K_lr_dry”. The value of 10% is for example expected for central North America in summertime for the period 2070-2099 under the “business as usual” emission scenario A2 (Nakićenović and Swart, 2000) in Sheffield and Wood (2008). The setup is otherwise identical to 3K_lr respectively 6K_lr.

The water vapor feedback of the atmosphere in GCMs can be estimated assuming constant relative humidity in all models (Soden and Held, 2006). Thus, in our set of simulations prescribed relative humidity is kept constant. Resulting from the non-linearity of the Clausius Clapeyron equation, near-surface values of specific humidity are increased by a factor 1.45 in the 6K simulation. However, low level humidity is mostly controlled internally by the interaction between land-surface, boundary layer, convective and radiative processes, while the influence of the externally specified humidity profiles is restricted mostly to upper tropospheric levels. In a previous study we have
shown that - for the adopted modeling framework - the diurnal cycle of convection is only weakly affected by the prescribed humidity profile (Schlemmer et al., 2011b). Simulated domain mean values of temperature in the state of diurnal equilibrium show a strong diurnal cycle, especially in the lower troposphere. Yet both temperature and specific humidity remain considerably larger for the warmer climates than for the CTL climate (see additional material).

We emphasize that we neither include the effect of aerosols, nor do we explicitly alter the CO$_2$ concentrations for the calculations of radiative transfer functions. This is consistent with the idealized nature of the current study and the fact that the atmospheric profile in limited area simulations is mostly determined by the large-scale environment rather than local greenhouse-gas forcing (Seneviratne et al., 2002). We merely focus on the effects that changes in background temperature, soil moisture and stratification exert on diurnal convection.

4.2.3 Model

The COSMO-CLM (hereafter CCLM) model version 4.0 is employed for the simulations (Steppeler et al., 2003, Doms and Förstner, 2004). It is based on the non-hydrostatic compressible atmospheric equations, uses a split-explicit time-stepping scheme (Klemp and Wilhelmson, 1978a, Wicker and Skamarock, 2002), and is suited for applications with horizontal grid-spacings from about 100 m to 100 km. The model has increasingly been used for cloud-resolving simulations, both for research and numerical weather prediction purposes (e.g. Hohenegger et al., 2009, Baldauf et al., 2011). The Runge-Kutta time integration scheme with a time step of 20 s is used for this study. The domain spans 100 × 100 grid points in the horizontal and 50 vertical levels. A horizontal grid spacing of 2.2 km is employed. Doubly-periodic lateral boundary conditions are used and the Coriolis force is switched off. The modeling domain includes no topography and homogeneous surface properties. In the initialization, a random temperature perturbation of ± 0.02 K is applied on the temperature at the lowest atmospheric level. A detailed description of the physical parameterizations used can be found in Schlemmer et al. (2011b) and Schlemmer et al. (2011a). One should keep in mind that the resolution employed is still too coarse to fully resolve moist convection and that the simulated precipitation intensity is linked to details of model-specific formulations and grid spacing (Bryan et al., 2003).
Figure 4.1: Mean diurnal cycle of domain mean specific water content (kg kg$^{-1}$, shaded area), specific ice content (kg kg$^{-1}$, contour lines with logarithmic interval $10^{-2.5}$ kg kg$^{-1}$), surface precipitation (mm h$^{-1}$, black solid line) and minimum and maximum values of domain mean precipitation over the 15 days of the simulation (dark grey shade) for the set of simulations. The number in the lower left corner gives mean daily precipitation amounts (mm day$^{-1}$).

4.3 Results

Fig. 4.1 shows the mean diurnal cycle of simulated cloud water, cloud ice and surface rain rate for the set of simulations. The mean diurnal cycle of deep convection and precipitation is remarkably similar in all simulations. Deep convection starts around 15 LT and mean precipitation peaks between 17 and 19 LT. A slight increase of early-morning mid-level clouds is visible and clouds persist until later in the evening in the
warmer climates. Deep convection occurs at the same time in CTL and 3K and about 1 h earlier in 6K. The time of the mean precipitation peak is shifted to slightly later times. Mean rain rates increase to a small extent in 3K but decreases in 6K as compared to CTL. However, the day-to-day variability of the domain mean precipitation (see width of the dark gray band) increases considerably for warmer climates leading to higher peak rates. Ice clouds are reduced in the warmer climate and the maximum height of water clouds is shifted to higher altitudes due to the shift of the freezing level to higher altitudes (located at ≈ 740 hPa in CTL, 700 hPa in 3K and 660 hPa in 6K), thus leading to a larger ratio of water to ice clouds. Precipitation production is shifted to more warm-rain processes (collision-coalescence of water droplets) and reduced ice processes in the warmer climates. In addition, the release of latent heat of freezing and reabsorbed heat at lower levels by the melting ice is reduced. The effect of increased warm-rain processes on precipitation is however presumably small since the ice phase exists in all simulations and the heat released by freezing is moreover considerably smaller than that by condensation. To test the impact of this effect we conduct a simulation, where the lapse-rate is stabilized in the CTL simulation. The differential warming is once applied at the surface and once at the tropopause level (CTL_{lr} and -0.75K_{lr}). Resulting simulations are remarkably similar (not shown). Thus we conclude, that changes of boundary-layer processes under warming dominate the impact of ice-cloud reductions. A stabilization of the atmosphere by the lapse-rate feedback increases the mid-level cloud cover in the morning but decreases clouds in the evening (see second row in Fig. 4.1). Mean rain rates are roughly unchanged compared to the CTL climate and decreased in 3K_{lr} compared to 3K and increased for 6K_{lr} compared to 6K. The day-to-day variability becomes even larger. It results from variations with periods >24 h.

The drying of the soil reduces the amount of water clouds but increases ice clouds. This is consistent with results from Schlemmer et al. (2011a), where an upward shift of clouds to higher altitudes over drier soils is observed. Mean precipitation amounts are considerably decreased as compared to CTL. A direct comparison of the diurnal cycle of mean rain rates is shown in Fig. 4.2a. The onset of precipitation is equal in all simulations but 6K and 6K_{lr}, where it occurs 1 h earlier. The precipitation peak occurs at about the same time for warmer climates. Fig. 4.2b-d show the change in the precipitation rate versus the change in temperature for mean amounts and the 10-day and 100-day return level including 95% confidence bounds (details concerning the calculation procedure are given in the additional material). Additional simulations, where only a drying of the soil without the stabilization is considered as well as the lapse-rate and soil-moisture changes are applied to the
CTL climate are included into the graph. Almost no changes are visible in the mean amounts for warmer climates. Drying the soil decreases mean amounts. At high intensities amounts increase under a warming. The stabilization of the atmosphere in 3Klr increases precipitation stronger than temperature increases alone, whereas in 6Klr increases are comparable to 6K. A drier soil in simulation 3Klr,dry and 6Klr,dry offsets these increases again. Over the drier soil intensities are again increased in a more stable atmosphere. Thus our idealized setup shows that increases of high precipitation intensities result if the warming of the atmosphere goes together with a stabilization. A drying of the soil in summertime could, however, partly or fully offset this increase in the precipitation extremes. Note that resulting precipitation increases are all considerably smaller than expected from Clausius-Clapeyron scaling. This is in contrast to previous findings for tropical precipitation (see Allan and Soden, 2008) but also mid-latitude areas (Lenderink and van Meijgaard, 2008, 2010), where an increase of high-intensity precipitation beyond Clausius-Clapeyron scaling was found.

Further analysis shows that maximum 2 m temperatures warm less than minimum temperatures in the warmer climates (not shown). The associated decrease of the diurnal temperature range (DTR) becomes stronger for more stably stratified atmospheres. DTR reductions caused by a stronger increase of minimum temperatures than maximum temperatures have been observed (Easterling et al., 1997). They are partly due to an increase in cloud cover (e.g. Dai et al., 2006) which agrees well with simulated cloud cover in our simulations (see Fig. 4.1). However, observations are also affected by changes in aerosol concentration (Makowski et al., 2009). The drying of the soil increases 2 m temperature and shows stronger increases of maximum temperatures which again compensates the reduction of the DTR. Near-surface humidity increases to some extent during daytime in the warmer climates, but strongly decreases over drier soils where ET is reduced by the availability of water.

Convective Available Potential Energy (CAPE) values increase with increased ambient temperature (see additional material). The consumption of CAPE by convection in the afternoon is not as effective in the warmer climates as in the CTL climate leading to larger night-time values. This effect is more pronounced if the atmospheric stability is increased. The day-to-day variability increases due to an emerging longer-day periodicity. The drying of the soil reduces CAPE over the whole diurnal cycle, which can be understood by the reduction of boundary-layer moisture over drier soil. A decrease of the lapse-rate leads to a slight decrease in Convective Inhibition (CIN) and thus enables the earlier onset of convection. This is consistent with the findings of Schlemmer et al. (2011b). The drier soil on the other hand increases CIN again.
4.3 Results

Figure 4.2: (a) Mean diurnal cycle of domain-mean precipitation (mm h$^{-1}$, solid line). (b-d) Precipitation increases vs temperature increases on a log-linear plot for the set of simulations for (b) mean values, (c) 10-day return levels and (d) 100-day return levels of hourly precipitation sums. Precipitation sums for each simulation is normalized by the respective amounts of the CTL simulation. “ref” indicates simulations, where only a warming is applied, “lapse-rate” simulations, where the atmosphere is stabilized, “dry” simulations, where soil-moisture saturation is decreased and “lapse-rate-dry” simulations, where both a drying and stabilization is applied. The black solid lines indicates theoretical precipitation increases expected from Clausius-Clapeyron scaling.

Convective mass-fluxes are active over a longer period of time, but have a smaller magnitude in the warmer climates (see additional material). This agrees well with the picture in the cloud field (Fig. 4.1). Thus the instability is consumed over a longer period of time but convection is less intense. The decrease of the lapse rate in 3K lr and 6K lr intensifies and shortens the period of convection again. These stronger updrafts in the more stable atmosphere can partly explain the larger rain rates seen in Fig. 4.2.
drying of the soil leads to a further shortening of the convective period and to higher convective mass-fluxes. Since ET from the surface is reduced in the simulations with the dry soil, precipitation intensities are not further increased in these simulations.

4.4 Discussion and Conclusions

The sensitivity of the diurnal cycle of summertime moist convection in mid-latitudes to changes of the ambient temperature profile was assessed in a set of idealized cloud-resolving simulations with temperature increases of +3 and +6 K. Values for upper-tropospheric relative humidity were kept constant, resulting in increased values for specific humidity in the warmer simulations. Moreover the role of an increase of the atmospheric stability was assessed by stabilizing the background profile in accordance with observed changes. Finally, the role of a summertime drying of the soil was investigated for the stable atmosphere by reducing soil moisture saturation by 10%.

In summary, we find that the above changes to the ambient atmospheric profiles, that are projected to result from anthropogenic climate change, have a decisive influence on the diurnal cycle of convection. In our idealized framework, cloud cover increases under a warming. The drying of the soil again reduces cloud cover and mean precipitation amounts.

Concerning precipitation, a significant increase at high intensities is found if a stabilization of the atmosphere is taken into account. An increase of temperature alone does not lead to significant increases. Increases at high intensities remain however smaller than expected from Clausius-Clapeyron scaling in all experiments, in contrast to previous studies and theoretical consideration. A drying of the soil strongly reduces the intensities for a given return period. Regarding mean precipitation amounts an increase of the background temperature shows small changes of both signs, whereas a decrease of soil moisture leads to a clear decrease of mean precipitation amounts.

It should be noted that the idealized framework represents, however, only stationary weather situations and cannot reproduce the full summertime European climate, which is marked by the occurrence of transient weather events. Our framework cannot generate situations with strong large-scale moisture convergence, that have the power to produce devastating extreme precipitation events. Moreover, it captures no topography, where increased moisture fluxes in warmer climates can produce larger orographic precipitation amounts.

On the other hand, the drying of the soil is a feature that can affect large parts of continental Europe over timescales of months as e.g. in the summer 2003 heat waves (e.g.
4.4 Discussion and Conclusions

Black et al., 2004, Fischer et al., 2007b). As confirmed using our model, mean precipitation amounts and boundary-layer moisture decrease and temperatures increase under such conditions. In the current study these two effects, namely the increase of precipitation in the high intensities and decreases at the low intensities (cf. Christensen and Christensen, 2003) could both be confirmed in separate experiments. The framework used is unable to realize a transition between these two extremes. Overall the idealized convection-permitting framework gives useful insights in the behavior of moist convection under increased temperature from a process-based perspective.
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Auxiliary Material

4.5 Calculation of precipitation statistics

To deduce whether the distributions of precipitation are significantly different for the different climates, a two-sample Kolmogorov-Smirnov test is performed on hourly precipitation intensities collected over all rainy grid points over day 16-30. A threshold value of 1 mm h\(^{-1}\) is used to distinguish between rainy and non-rainy points. The Kolmogorov-Smirnov test states that all distributions are different at the 5\% significance level.

To infer more information about the behavior at large intensities, return levels for intensities using a maximum-likelihood fit of the generalized extreme value distribution using the R (R Development Core Team, 2008) package `ismev` (see Coles, 2001, for details) are calculated. Block maxima were constructed by sub-dividing the domain into 9 equally large subdomains. For each subdomain the daily maximum of hourly precipitation intensity is selected resulting in a dataset holding 135 values.

4.6 Additional figures: profiles of temperature and humidity, CAPE and convective mass-fluxes

Figure 4.3: Input profiles for (a) temperature (°C) and (b) specific humidity (g kg\(^{-1}\)) and simulated profiles of minimum and maximum values over day 16-30 of domain mean (c) temperature (°C) and (d) specific humidity (g kg\(^{-1}\)). Solid lines indicate the mean value. Simulations shown are CTL (black), 3K (blue solid line), 3K_{lr} (blue dashed line), 6K (red solid line) and 6K_{lr} (red dashed line).
Figure 4.4: Mean diurnal cycle of CAPE (J kg\(^{-1}\)) with domain mean values in black and cloudy points in grey for (a) CTL, (b) 3K, (c) 6K, (d) 3K\_lr, (e) 6K\_lr, (f) 3K\_lr\_dry and (g) 6K\_lr\_dry. Mean values are shown by the solid lines while the 10th and 90th percentile are shown by the dashed lines. The 10th and 90th percentile were calculated by considering all grid points on all 15 days.
Figure 4.5: Mean diurnal cycle of convective mass-flux (kg m\(^{-2}\) s\(^{-1}\)) for the set of simulations.
Chapter 5

Conclusions and Outlook

5.1 Conclusions

This thesis investigated the characteristics of summertime moist convection over continental Europe and its sensitivity to modifications of key environmental conditions. To this end an idealized cloud-resolving framework was developed by modifying an existing weather prediction and regional climate model. The model is run over one month, where a state of diurnal equilibrium develops. In this state convection and precipitation is observed every day with a small day-to-day variability.

It was demonstrated that the framework is able to produce a realistic diurnal cycle of heat-fluxes, clouds and precipitation. Moist convection over land is a complex system with many interactions. Looking at convective weather situations lasting over several days it was found, that the main controls on the diurnal cycle are both the energy balance at the surface and the stratification of the atmosphere. The availability of soil water and the available net radiation determine, how much water is evaporated into the atmosphere which again regulates the amount of precipitation. For the simulations in diurnal equilibrium throughout a positive soil moisture precipitation feedback was observed.

The stratification of the atmosphere on the other hand controls how easily convection can be triggered and how effective moisture and heat are redistributed vertically. The vertical stability moreover regulates the amount of moisture that accumulates in the PBL and thereby the timing of the triggering of convection and precipitation. Earlier precipitation peaks are observed in a more stably stratified environment in diurnal equilibrium, which is counterintuitive from daily experience. The positive feedback was found not only to work over time, but also on spatial scales of 10-50 km. The spatial feedback is strongest for stably stratified atmospheres over semi-arid soils.

The prescribed moisture content of the atmosphere is an important ingredient regulating convective activity over single days. It was confirmed that the entrainment of dry air into ascending plumes is able to considerably delay the transition from shallow to deep convection. Over the course of multiple days, when moisture is redistributed in the atmosphere by ongoing convection, this feature vanishes and the system gets almost independent of the prescribed background humidity. It is then the evaporation from the soil and convection itself that regulate the moisture content of the atmosphere.

The above mentioned environmental conditions (the energy balance at the surface, the availability of soil moisture and the static stability of the atmosphere) are all subject to an anthropogenic climate change. The developed idealized framework was therefore
used to investigate the response of mid-latitude diurnal convection to changes in the ambient temperature, a stabilization of the atmosphere resulting from a stronger upper-tropospheric warming and a destabilization through a drying of the soil in a process-based manner. Our framework holds the advantage of being autonomous from external driving data, and the cloud-resolving formulation of convection delivers results derived independently of parameterizations for moist convection. We confirm an increase of near-surface temperatures, the formation of more clouds and a decrease of the diurnal temperature range in warmer climates. Regarding precipitation, an increase at the high intensities is observed for an increased background temperature. A significant increase is however only observed, if the temperature increase is accompanied with a stabilization of the atmosphere. A drying of the soil reduces precipitation amounts both in the mean but also at the high intensities considerably.

The modeling framework was tested across a large range of parameters. It was found that it is nearly insensitive against changes in the parameterization of ice microphysics, whereas the formulation of moisture terms in the calculation of turbulence, that mostly influences PBL moisture has a strong influence. The numerics of the model are implemented reasonably well, showing minor sensitivities to the time step used as well as a reasonable convergence increasing the horizontal resolution. The vertical grid spacing impacts the amount of simulated shallow clouds, but the overall result only to a small degree.

The developed framework is well suited to further investigate the characteristics of moist convection. Because of its design that includes a strong relaxation in the upper troposphere resembling the increase of advection with height, it is mostly applicable to mid-latitudes. With slight modifications to this formulation it is however also portable to the tropics. It proofed to be a very useful tool to get deeper insights into involved processes by reducing the complexity that a usual climate model brings about. The idealized nature reduces computational cost, and eases interpretation of processes.

### 5.2 Outlook

A further process to investigate in future work could be some disturbance to the diurnal equilibrium state. By including a time-variant relaxation profile one could mimic a transition from one weather regime to a slightly different one. The time scales needed for the model to adapt to the new state could be one issue for discussion. In the soil a change of the soil moisture content could be performed to investigate how long it takes
for the coupled system to adapt to the new state.

The incorporation of irregularities to the very symmetric state as it is now would definitely be interesting. The inclusion of orography presumably changed the triggering of convection completely, leading to a different onset time of clouds and precipitation in different areas of the domain. For the surface it would furthermore be interesting to add inhomogeneities by e.g. changing the vegetation or the soil type across the domain, or even by including waterbodies into the model. One might see a pattern in the locations, where convection tends to be triggered and where it favorably develops.

Future work focusing on the model formulation could include an increase of the resolution. Together with this increase in resolution one would like to change to a 3D turbulence scheme and replace the numerical diffusion with a more physically meaningful diffusion. One would hope to resolve processes, especially associated with shallow convection better. It would be interesting to see if shallow convection developed in the state of diurnal equilibrium, to what degree shallow convection modified the later stage of deep convection and how long the transition took in a more sophisticated model.

To be able to better investigate the coupling of the land surface with the PBL and resulting convection it would be useful to couple the land surface scheme with the atmosphere by passing the fluxes of sensible and latent heat instead of passing the boundary conditions skin temperature and skin moisture as done now. This resulted in a more consistent treatment of the fluxes throughout the whole model. The inclusion of further processes in the vegetation could be of further interest. Coupling a more advanced land-surface scheme to the CCLM model as for instance done by Davin et al. (2011) could be one aim to consider.

So far the model has only been applied to convection over land. It could however be intriguing to transfer it to maritime regions and simulate the development and break-up of stratocumulus clouds. A starting point could be to employ data for test cases as for example the GCSS RICO case (http://www.knmi.nl/samenw/rico/index.html) as e.g. done in Helmert et al. (2008) to infer to which degree the model is able to reproduce a stable stratocumulus deck.

A further issue of interest would be the inclusion of aerosols into the model that influence microphysics and radiation. The developed framework is the ideal testbed for the direct effects of aerosols on radiation and temperature (e.g. Haywood and Boucher, 2000), the semi-direct effects of a heating of the troposphere by aerosols (Hansen et al., 1997) and the indirect effects on albedo (Twomey, 1974) and cloud life-time (Albrecht,
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1989).

To infer, whether the received cloud-resolving results of this thesis outperform diurnal convection simulated using a parameterization scheme it would be tempting to repeat some of the simulations using different parameterization schemes. Firstly, the degree of difference between various schemes and the CRM result could be analyzed and secondly, the cloud-resolving results could be used to improve parameterizations.
Appendix A

Disentangling the forcing mechanisms of a heavy precipitation event along the Alpine south side using potential vorticity inversion

Linda Schlemmer¹, Olivia Martius¹, Michael Sprenger¹
Cornelia Schwierz², Arwen Twitchett²

¹:Institute for Atmospheric and Climate Science, ETH Zurich, Switzerland
²:Institute for Climate and Atmospheric Science, University of Leeds, United Kingdom

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Abstract

Extreme precipitation events along the Alpine South side (AS) are often forced by upper-level positive potential vorticity (PV) anomalies over Western Europe. These so-called PV streamers go along with a dynamical forcing for upward motion, a reduction of the static stability in the troposphere, hence facilitating convection, and are associated with low-level winds that transport moisture towards the Alps.

A case of heavy precipitation is examined using the ERA-40 reanalysis data. Piecewise PV inversion (PPVI) and the CHRM limited area model are used to assess the influence of meso-scale parts of the streamer on the precipitation event. The impact on the vertical stability is quantified by the convective available potential energy (CAPE) and an index of static stability. Very sensitive areas in terms of the stability are located beneath the southern tip of the streamer, smaller changes in the stability are observed in the Alpine region.

The moisture transport towards the Alps is sensitive to the amplitude of the streamer which influences the amount of water that can be transported along its eastern flank. The impact of the topography on the flow is assessed by calculating an average inverse Froude number. Whether or not the air parcels are blocked by or lifted over the barrier (going along with suppressed and enhanced precipitation, respectively) depends on
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the vertical stability and the impinging wind velocity, i.e. two parameters which are inherently linked to the PV streamer and its substructure.

A.1 Introduction

Heavy precipitation (HP) events occurring in highly structured and steep terrain (here the southern flanks of the European Alps) often cause secondary natural hazards (flooding, landslides), that represent serious threats to the population and infrastructure. This study focuses on improving our understanding of synoptic- and meso-scale atmospheric mechanisms causing such events. As a long-term objective, it aims to advance the prediction of the location and intensity of similar HP events.

Recognizing the importance of the topic, the Mesoscale Alpine Programme (MAP) (e.g. Bougeault et al., 2001) focused on the factors influencing precipitation in the complex topography of the Alpine region. Both meso-scale and large-scale flow dynamics impact the precipitation intensity and its spatial structure. On the meso-scale, the flow strongly interacts with the local topography (Gheusi and Davies, 2004) and thus triggers convection (e.g. Buzzi et al., 1998, Fuhrer and Schär, 2005). On the synoptic-scale, high potential vorticity (PV) intrusions over Western Europe play an important role in forcing HP along the Alpine south side (Massacand et al., 1998, 2001, Martius et al., 2006, Hoinka and Davies, 2007). Generally these intrusions adopt the form of narrow (∼ 500 km), deep (∼ 4 km) and meridionally elongated (∼ 2000 km) filaments of stratospheric air, termed PV streamers (Appenzeller and Davies, 1992) and reach from the British Isles southwards to the Iberian Peninsula. These PV structures are reflected in geopotential troughs located over Western Europe (e.g. Jacobite, 1987, Rudari et al., 2005).

A PV streamer, being a positive upper-level PV anomaly, has several dynamical characteristics that can trigger and/or enhance precipitation:

(a) The static stability in the troposphere is reduced underneath a streamer (Hoskins et al., 1985) and the initiation of convection is thus facilitated.

(b) In a Lagrangian sense, air is forced to ascend northwards along the PV streamer’s downstream flank following the sloping isentropes (e.g. Hoskins et al., 1985). From a Eulerian perspective the horizontal movement of the streamer leads to a lifting of isentropes as it passes by (the so-called “vacuum cleaner effect”). The induced vertical motion can initiate additional convection along the downstream flank of the streamer (e.g. Funatsu and Waugh, 2008).

(c) The flow induced by an elongated streamer over western Europe can penetrate to
the ground and transport moisture from the Atlantic and/or the Mediterranean Sea towards the Alps (e.g. Reale et al., 2001, Koch, 2004, Turato et al., 2004). The strong southerly flow along the eastern flank of the PV streamer can act like a conveyor belt and bring large amounts of moisture towards the Alps. This phenomenon is known as a warm conveyor belt (Eckhardt et al., 2004) or as atmospheric rivers (Newell et al., 1992) in other parts of the world.

(d) In the particular setting of the streamer being located over Western Europe, the Alps act as an additional forcing factor. The streamer influences both the low-level wind field and the static stability and therefore has a crucial impact on the way the flow interacts with the orography, with major implications for the location and intensity of the precipitation. Further on the orography slows down the eastward progression of the streamer (Morgenstern and Davies, 1999). Both factors have a crucial impact on the overall amount of precipitation, which is determined by the duration of an event multiplied by its rainfall rate (Doswell et al., 1996).

The meso-scale parts of the streamer influence points (a), (c) and (d) listed above. Indeed, Fehlmann et al. (2000) show that the location, extension and effective amplitude of meso-scale PV substructures substantially influence precipitation patterns on the Alpine south side (AS). This study builds upon and extends their work by quantifying effects of the substructure of the streamer on forcing factors (a), (c) and (d) and by addressing their relevance for one specific precipitation event.

The correct representation of the meso-scale substructure of a streamer still poses a challenge to state-of-the-art weather prediction models. As demonstrated in Fehlmann and Quadri (2000), these meso-scale substructures are relevant for the forecast quality. They find that forecasts of HP along the Alpine south side are sensitive to the location and amplitude of the southern part of the streamer. Grazzini (2007) found that the skill in predicting the strong moisture fluxes associated with HP events has improved at a faster rate than the skill for average conditions since 1986. Nevertheless, errors in the forecasted PV field in amplitude, phase and substructure still occur in present-day forecasts (e.g. Dirren et al., 2003, Didone, 2006).

This study focuses on one HP event which occurred on 13 November 1996 and it addresses the following questions concerning the factors favoring HP: What is the PV streamer’s dynamical role in the tropospheric destabilization? How does the meso-scale substructure influence the northward advection of water vapor? Finally, to what degree does it influence the orographic lifting and enhancement of convection? To answer these questions, we modify the streamer’s substructure using piecewise PV in-
version (PPVI) (Davis, 1992). The design of the experiments is guided by an analysis of the typical structure and amplitude of forecast errors for streamer induced HP events.

The paper is organized the following way. A detailed description of the PPVI method and the data used is given in section A.2. An overview over the synoptic situation as well as a short discussion of typical forecast errors is presented in section A.3, followed by an in-depth discussion of the PPVI application to the meso-scale substructure of the PV streamer (section A.4) and implications for the precipitation (section A.5). Finally, the main results are summarized in section A.6.

A.2 Data and Methodology

A.2.1 Data
The meteorological fields used in the case study stem from the ERA-40 reanalysis (Uppala et al., 2005) and are interpolated onto a 1° x 1° latitude-longitude grid. The verification fields for the precipitation are taken from an observation-based Alpine precipitation climatology (Frei and Schär, 1998). Finally, for the assessment of the forecast errors we use ECMWF operational forecast and analysis data.

A.2.2 Piecewise PV Inversion and model runs
The PPVI technique used for this study was originally developed by Fehlmann (1997) and the approach slightly differs from that of Davis (1992). In a nutshell, this PPVI approach is based upon the quasi-geostrophic approximation, but takes into account the nonlinearity of the Ertel-PV inversion problem by using an iterative technique. For

<table>
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<th>angle [°]</th>
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<th>x_max [km]</th>
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</tr>
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<td>1200</td>
<td>-1800</td>
<td>1800</td>
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</tbody>
</table>

Table A.1: Parameters used to define the filtering box for the PV anomaly. The center of rotation is given in longitude and latitude as well as the angle by which the box is rotated. The extent of the box in the cartesian grid is given by the maximum and minimum values for the x- and y- direction in km.
A.2 Data and Methodology

Further details refer to Fehlmann (1997).

The first step in the inversion is the definition of a PV anomaly, which is done using a simple spatial-filtering technique. A three-dimensional box is placed around the PV structure to be modified. Within this box the PV is smoothed by building the zonal mean within the box. The filtered field is then subtracted from the original field. The size and the amplitude of the PV anomaly are dependent on the size of the box and the number of iterations of the filter. The location and size of the box used for each experiment are listed in Table A.1. In the vertical the box extends from 2 km up to 15 km. 5 iterations of the filter were done. This approach allows modifying meso-scale substructures within a synoptic-scale PV anomaly.

The resulting anomaly is separated into its positive and negative part and the positive part is the PV anomaly to be inverted. Boundary values for potential temperature (upper and lower lid) and horizontal wind (lateral boundaries) are set to zero. With this choice of boundary conditions an ambiguity is introduced (Hakim et al., 1996) because multiple states can share the same PV distribution. Hence the new balanced state is a realistic and a correct solution of the inversion problem, but is not necessarily a unique solution. Finally, we developed a simple algorithm that allows to stretch and weaken/strengthen a PV anomaly. Using this tool the anomaly is scaled to its desired form and added to the background field.

After the PPVI, the inverted fields undergo an implicit normal mode initialization (Temperton, 1988, Temperton and Roch, 1991). This step removes spurious fast-moving gravity waves, which can arise from the PPVI.

The output of the PPVI serves as initial condition for numerical weather prediction (NWP) simulations. We use the Climate High Resolution Model (CHRM), which is a modification of the High Resolution Model (HRM) version 1.6. of the German Weather Service (for more details see Majewski, 1999, Lüthi et al., 1996, Vidale et al., 2003). The hydrostatic model is run with 40 vertical levels, the variables are interpolated onto a latitude-longitude grid with $0.5^\circ \times 0.5^\circ$ horizontal resolution. The parameterized physical processes include a vertical diffusion scheme of order 1.5 after Mellor and Yamada (1974) and a surface-layer formulation after Holtslag and Boville (1993), Kessler-type grid-scale microphysics (Kessler, 1969) including a parameterization of the ice-phase (Lin et al., 1983) and a mass flux convection scheme after Tiedtke (1989). Initial fields are taken from the ERA-40 data for the control run and from the corresponding output of the PPVI experiments. Boundary conditions are taken from the ERA-40 data. The model is integrated for 24 hours.
A Alpine heavy precipitation event

A.2.3 Moisture treatment
The humidity field remains unchanged during the PPVI procedure. This leads to discrepancies between the unchanged moisture and the modified wind and temperature fields. First, the temperature can fall below the calculated dew-point temperature at some grid points and the PPVI hence leads to artificial super-saturation (thermodynamic inconsistency). Second, the moisture distribution is determined by the wind and temperature fields associated with the original PV distribution. If the latter is altered, with associated changes in temperature and wind, the unchanged moisture is hence no longer dynamically consistent with the flow (flow evolution inconsistency). This is of particular importance if the PV alterations are carried out in the tropics and if the inverted fields are used to initialize model runs (McTaggart-Cowan et al., 2003).

We present three approaches to address these problems by adapting the moisture distribution after the PPVI. In a first approach (QI), the moisture field is corrected to 100% relative humidity at points where over-saturation occurs and the specific humidity is kept constant elsewhere. In a second approach (RH), the relative humidity is kept constant. In a third, more radical approach, the humidity field is filtered on constant height surfaces inside the inversion domain (FI). In this approach we apply 5 iterations of a median filter of size 10 x 15 grid-points onto the humidity field. The filter conserves the planetary-scale structure of the moisture field, but the synoptic-scale structures are removed. This third approach is closest to the solution chosen by Funatsu and Waugh (2008), who use a climatological moisture distribution.

Finally note that we apply only meso-scale changes to the original PV field and that the integration time of the NWP model is short (24 h). Under the assumption of approximate linearity this leads to correspondingly small changes in the temperature and humidity field.

A.2.4 Stability and flow-regime diagnostics
The convective potential is assessed by means of most-unstable CAPE and CIN (convective available potential energy and convective inhibition; see e.g. Emanuel et al., 1994). CAPE and CIN are calculated using the post-processing routine of the Consortium for Small-scale Modeling (COSMO, Steppeler et al., 2003) adjusted for ECMWF fields. A classical parcel ascent is calculated for the most-unstable parcel in the lowest 300 hPa.

In addition the static stability is assessed using four traditional indices and combining them into a new combined threshold index (CTI): Totals-index (Miller, 1972), K-index
A.2 Data and Methodology

(George, 1960), squared dry and wet Brunt-Väisälä frequency at 850 hPa. The dry Brunt-Väisälä frequency $N$ was calculated using $N = \sqrt{\frac{g}{\frac{\partial \theta}{\partial z}}}$, the wet one $N_w$ accordingly using the equivalent potential temperature $\theta_e$ instead of potential temperature $\theta$. $\theta_e$ was determined using the formula by Bolton (1980). These four indices are combined into a new index in the following way. For each individual index, a threshold is set which divides stable from unstable stratification. The combined index is then simply the number of individual indices that surpass their instability threshold, i.e. it varies between 0 (stable) and 4 (unstable). The following thresholds are used: Totals-index $> 44$ K, K-index $< 24$ K, squared dry Brunt-Väisälä frequency at 850 hPa $< 10^{-4}$ s$^{-2}$ and squared wet Brunt-Väisälä frequency $< 0$ s$^{-2}$. In contrast to $N$, $N_w$ exceeds the set threshold for particularly wet regions located primarily in the Atlantic and the Mediterranean. The K-index and the Totals index are not independent since they both evaluate temperature and dew point at the same levels. However, in many cases the two thresholds are exceeded at different locations. A related procedure has been used to identify unstable regions from satellite images (see e.g. Mecikalski and Bedka, 2006).

The vertical stability, the speed of the incident flow, and the shape and height of the topography determine whether the air impinging upon the Alps is blocked by the latter or carried over it. The inverse Froude number (or dimensionless mountain height) $\epsilon = NH_0/U$ (where $H_0$ is the mountain height and $U$ the incident flow speed) is a widely used measure to distinguish between the two flow configurations or to decide which part of the impinging flow is blocked and which is able to surpass the Alpine barrier.

Studies determining the flow response have mainly focused on idealized flow over idealized topography (e.g. Schär and Davies, 1988, Smith, 1989, Olafsson and Bougeault, 1997). Reinecke and Durran (2008) investigated different methods to characterize non-uniform flows using the inverse Froude number. We follow their approach and calculate the inverse Froude number based on the bulk value $N_b = \sqrt{\frac{g}{\frac{\partial \theta_{H_0}}{\theta_{H_0}}}}$, where $\theta_{900}$ is the reference potential temperature at the ground and $\theta_{H_0}$ the potential temperature at $H_0$. Following Reinecke and Durran (2008) the bulk method is the better predictor of the low-level flow diversion. $U$ and $N$ values are evaluated along a cross-section across the Alps, parallel to the wind field at 850 hPa. $H_0$ is set to 2000 m and $U$ at $H_0$ as well as $N$ are identified one Rossby radius of deformation $L_R = NH_0/f_0 \approx 90$ km away from the barrier (Pierrehumbert and Wyman, 1985). The exact distance of the identification point from the Alps is not very critical because the wind and stability are relatively uniform for grid points farther than approximately 80 km away from the Alps. Typically the transition from blocked to cross-mountain flow occurs at $\epsilon_{crit} = 1.0 - 1.2$. 

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Figure A.1: Evolution of the PV at 320 K (shaded) in the ERA-40 dataset. The bold black line indicates the 2 PVU isoline, the thin black line the 8 PVU isoline. The following dates are shown: a) 1800 UTC 11 November 1996, b) 0600 UTC 12 November, c) 1800 UTC 12 November, d) 0600 UTC 13 November. The arrows represent the wind field at 850 hPa. The black bold arrow in the lower-left corner corresponds to a wind speed of 20 m s\(^{-1}\).

Figure A.2: Specific humidity in g kg\(^{-1}\) at 850 hPa at 0600 UTC 12 November 1996. Panel a) shows the ERA-40 field. Panel b) shows the specific humidity field after the application of a median filter. The solid black line depicts the 2 PVU contour at 320 K.
A.3 Synoptic overview and forecast error assessment

A.3.1 Synoptic situation

A deep trough in the 500 hPa geopotential was located over the eastern Atlantic and Western Europe on 13 November 1996 (not shown). This trough went along with a meridionally elongated pattern of high PV (a PV streamer) located over Western Europe which extended south to $\approx 32^\circ$ N (Fig. A.1). It will be shown that this synoptic-scale PV streamer acted as the main upper-level forcing of the precipitation. The streamer formed on 11 November 1996, became progressively longer and thinner and reached its maximal elongation on 13 November. At this time, the dynamical tropopause was considerably lowered. The minimum height of the 2 PVU iso-surface was found at 550 hPa over northern Germany at 1800 UTC 13 November 1996 (not shown).

An elongated meso-scale structure of high PV air ($\geq 8$ PVU), located inside the synoptic-scale PV streamer, moved northwards along the eastern flank of the streamer between 1800 UTC 11 November and 1800 UTC 12 November. This structure was linked to a local wind maximum at upper-level with a considerable PV cross-gradient component that lead to PV advection and influenced the temporal evolution of the streamer (not shown). The streamer followed a typical anti-cyclonic lifecycle. At 0600 UTC 13 November, a hook-like structure formed in the tip of the streamer indicating local cyclonic wrapping and the beginning of barotropic decay.

A strong wind field was co-aligned with the rim of the streamer and extended all the way through the troposphere. The low-level wind maximum along the eastern flank was associated with a band-like structure of very humid air that extended northward from the subtropics towards the Alps prior to the onset and during the HP period (Fig. A.2a). The precipitation accumulated between 0600 UTC November 12 and 0600 UTC November 13 1996 is shown in Fig. A.3. The 24 h accumulated precipitation exceeded 120 mm in the French Alps (label A), 25 mm in the Piedmont (label B) and 100 mm in southern Switzerland (label C).

The mid-tropospheric isentropes exhibit the characteristic displacement in the vertical associated with a positive upper-level PV streamer (see for example Fig. 3 in Funatsu and Waugh, 2008, Hoskins et al., 1985). They slope upward both towards the center of the anomaly in the east-west direction and towards the north in the north-south direction (not shown). The changes in the vertical temperature structure are reflected in the dry stability. This can be seen in a vertical cross-section through the tip-region showing $N^2$ (Fig. A.4a). Stratospheric air intrudes into the troposphere and the isentropes are bent upwards underneath the streamer, indicating a cold anomaly in the lower troposphere.
A Alpine heavy precipitation event

Figure A.3: Precipitation in mm accumulated over 24 hours (0600 UTC 12 November to 0600 UTC 13 November). The bold black lines indicate the areas used in the discussion in section A.5. Label A denotes the French Alpes, label B the Piedmont and label C the Ticino.

A “bubble” of reduced stability (small $N^2$) that is most likely the result of moist processes is located directly underneath the streamer (between 380 - 600 hPa and 15° W - 10° W, label A). Below, a stable layer separates the bubble from the unstable surface-layer between 850 hPa and the ground. A band of relatively high stability extending from the surface up to the tropopause is located to the west of the streamer (see Fig. A.4a, label B). This band is closely linked to the upper-level intrusion and the associated lifting of the isentropes. The high spatial variability of the stability pattern is illustrated further in a plan view of $N^2$ at 450 hPa (Fig. A.4b). Areas of low stability are found underneath the center part of the streamer (where we observe the bubble in panel a) and a band of high stability is located along the streamer’s western edge.

The observed distribution of CAPE is closely tied to the upper-level PV field and is characterized by high values (up to 930 J kg$^{-1}$) in the tip region of the streamer (Fig. A.5a). The tip is situated over the ocean where the lower troposphere is moist and hence favors convection. Values of up to 300 J kg$^{-1}$ are found over the French Alps and values of $\approx$ 500 J kg$^{-1}$ are present along the streamer’s eastern side. In these areas convective activity can be seen in the infrared satellite image (Fig. A.6). CIN values are large along the eastern flank of the streamer in the Mediterranean (up to 300 J kg$^{-1}$) and over southern France (up to 50 J kg$^{-1}$) (not shown). The areas identified as highly unstable by the CTI are in general co-located with elevated CAPE values (Fig. A.5b). The different components of the CTI respond to instabilities in different regions. In contrast to $N$, $N_w$ exceeds the set threshold in particularly moist regions located in the Atlantic and the Mediterranean. In the Mediterranean and over Northern Europe.
Figure A.4: All panels show the situation at 0600 UTC 12 November 1996: a) Cross-section of $N^2$ (shaded), the 2 PVU isoline (bold black line) and isentropes (black lines) in the ERA-40 data; b) $N^2$ in $10^{-4} \text{s}^{-2}$ at 450 hPa in the ERA-40 data, the bold black line is 2 PVU isoline at 320 K, the bold white line indicates the position of the cross-section; c) differences in $N^2$ (experiment - ERA-40, shaded) along the cross-section for the TA experiment, the bold black line indicates the location of the 2 PVU isoline in the ERA-40 data. d) same as c) but for the TR experiment.

only the Totals-index exceeds the instability threshold.

A.3.2 Medium range forecast quality

Owing to the streamer’s influence on the HP event it is worthwhile to show how accurately the PV streamer at 1200 UTC 12 November 1996 was forecasted. Such an error analysis will also provide us with good estimates of typical mis-representations for the modifications in the subsequent sensitivity study. To this end, the upper-level PV field at the 320 K isosurface of the 96-hour ECMWF deterministic forecast is compared with
Figure A.5: a) CAPE field in J kg\(^{-1}\) at 0600 UTC 12 November 1996 (shaded) calculated from the ERA-40 data b) CTI calculated from ERA-40 data for the same date. The bold black line shows the 2 PVU isoline at 320 K.

Figure A.6: Meteosat infrared satellite image at 0600 UTC 12 November 1996 (colour). Black contour lines show PV in PVU at 315 K. The red shaded areas indicate CAPE values exceeding 100 J kg\(^{-1}\), and green shaded areas indicate CAPE values exceeding 300 J kg\(^{-1}\).

The corresponding ERA-40 data (Fig. A.7a). The forecasted PV streamer is in the right position (no phase error), but the streamer is broader and shorter than in the ERA-40 data (structural error). As a consequence, the largest errors in the upper-level PV field are located in the tip area and along the flanks of the streamer. The amplitude of these errors amounts to several PVU (amplitude error).
A.3 Synoptic overview and forecast error assessment

Figure A.7: a) Difference (96 h forecast - ERA-40 data) of the PV at 320 K in PVU at 1200 UTC 12 November 1996 (shaded). The bold black line depicts the 2 PVU at 320 K in the ERA-40 data, the dashed black-grey bold line indicates the 2 PVU isoline at 320 K in the forecast. b) Difference (analysis-114 h forecast) of the PV at 320 K in PVU at 1800 UTC 26 October 2004 (shaded). The bold black line depicts the 2 PVU at 320 K in the ECMWF analysis, the dashed black-grey bold line indicates the 2 PVU isoline at 320 K in the forecast.

It might be argued that this discrepancy between forecast and reanalysis is due to the relatively un-matured forecast system in 1996. Yet, a more recent example from 26 October 2004 (Fig. A.7b) shows a small phase error but in general a very similar error pattern and underlines the possibility of considerable PV errors in the most sophisticated forecast systems.

These observations in terms of amplitude, phase and structural errors are corroborated by a more systematic study of an ensemble of thirteen Alpine HP cases\(^1\). The error assessment was undertaken using a newly developed methodology that is outlined in more detail by Twitchett and Schwierz (2010, in preparation).

The main conclusions drawn from all the thirteen case studies investigated suggest that typical forecast errors for streamers are: (i) in the forecast lead time of 6 hours streamer structures closely match the analysis counterparts; (ii) for forecast lead times of 30 to 54 hours a larger spread of errors is found and there is a tendency towards a too early development of the streamer; (iii) with lead times of 102 and 106 hours forecasted

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A Alpine heavy precipitation event

streamers are generally shorter and wider than the analysis streamers and break later. PV modifications of about +/- 2-3 PVU seem justifiable from this limited error climatology. These modifications should be predominantly applied to the edges and to the tip of the PV streamer to mimic the tendency of the forecast models to produce too broad, weak and short streamers.

A.4 Alterations of the substructure of the streamer

A.4.1 Overview on the experiments

To assess the potential impact of forecast errors described in section A.3.2 on this HP event, a series of sensitivity experiments were conducted using the ERA-40 data. Distinct meso-scale substructures of the streamer were altered and the sensitivity of HP to these changes tested.

The following streamer modifications are discussed in the paper (Fig. A.8a and b): a) a strengthening (EA) and weakening (ER) of the eastern flank of the streamer (the co-location of the observed meso-scale PV maximum and area of PPVI modifications along the eastern flank is a coincidence), b) a strengthening (TA) and weakening (TR) of the tip and c) a shorter and broader streamer with a structure that mimics the form

![Figure A.8](image-url)

**Figure A.8:** PV at 320 K at 0600 UTC 12 November 1996 in the ERA-40 data (shaded). The contour lines in panel a) indicate the difference between the ERA-40 data and the modified PV field for experiments TA, TR (solid black line), EA and ER (solid white line). Panel b) shows the PV modifications used for the SB experiment, black lines indicate added PV structures and white line indicate PV structures that were removed. The contour interval is [1,2,4] PVU in both panels.
A.4 Alterations of the substructure of the streamer

of the streamer in the 96h forecast (SB). The amplitude of these modifications are, in terms of PV and change in total energy (see Fita et al., 2007, for details), comparable to the errors found in the 48 h forecast (cf chapter A.3.2). All modifications were made at 0600 UTC 12 November, 36 h prior to the mature stage of the streamer and 24 h prior to the HP event. Table A.2 provides an overview of the modifications and the resulting changes in PV. We are now interested in how sensitive the static stability, the low-level moisture transport towards the Alps and the orographic lifting react to the meso-scale alterations of the streamer.

<table>
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<th>Δ(v_⊥) [% CTL]</th>
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</table>

Table A.2: Modifications of the PV and the resulting changes in the humidity flux. Changes in PV are thereby quantified giving both the area on the 320 K isosurface where the magnitude of the PV change exceeds 0.1 PVU and the spatial average of the change in this area. For the modification SB the negative part and the positive part of the changes are given separately. The humidity fluxes q · v_⊥ are measured at the cross-section indicated in Fig. A.9a where v_⊥ is the wind speed measured normal to the cross-section. Values for the humidity flux are given for the three different moisture treatments RH, QI and FI. ε_b denotes the bulk estimate of the inverse Froude number. All quantities are valid at 0600 UTC 12 November at the time of the modifications.

A.4.2 Overview on the analysis methods

Stability
The alteration of an upper-level PV structure has an effect on the temperature stratification in the troposphere below (e.g. Hoskins et al., 1985). The question addressed here is whether these temperature changes have a crucial impact on the static stability. Both temperature and moisture are essential for a destabilization of the environment. The analysis of the moist effects on the stability is complicated by the fact that the
PPVI method does not include an adaption of the moisture field to the altered temperature and flow fields. We therefore discuss the impact of the PPVI on the dry and the moist stability separately. All experiments are carried out with three different moisture treatments (QI (absolute moisture content is kept constant), RH (relative humidity kept constant) and FI (strong spatial filtering; see section A.2A.2.3 and Fig. A.2b for more details) allowing a thorough assessment of the sensitivity to moisture effects.

**Moisture flux**

As stated in the introduction, ample and sustained moisture supply is crucial for long-lasting heavy precipitation events (see e.g. Doswell et al., 1996). Large amounts of water vapor and cloud droplets were transported within the low-level jet along the eastern side of the PV streamer towards the Alps between 0600 UTC and 1200 UTC 12 November (Fig. A.9a and b). In order to quantify the sensitivity to the PV alterations, the humidity flux at 0600 UTC 12 November is determined along a vertical cross-section located on the eastern side of the streamer (see Fig. A.9). More specifically, the prod-

**Figure A.9:** Panel a) shows the humidity distribution in g kg\(^{-1}\) (shaded) and the wind field at 850 hPa for the ERA-40 data at 0600 UTC 12 November. The black line is the 2 PVU contour at 320 K. The bold white line indicates the position of the cross-section used for the humidity flux calculations. The bold black and white line indicates the cross-section used to determine the inverse Froude number. The reference wind vector in the lower left corner indicates a wind speed of 20 m s\(^{-1}\). Panel b) shows the humidity flux in (kg kg\(^{-1}\)·m s\(^{-1}\)) (shaded) at 0600 UTC 12 November 1996 at the white cross-section shown in panel a. The solid black lines show the wind field perpendicular to the cross-section. The bold grey line indicates the 2 PVU isoline.
A.4 Alterations of the substructure of the streamer

uct of water vapor and cloud water times velocity perpendicular to the cross-section is integrated over the area of the cross-section:

\[
\frac{1}{\rho \cdot g} \int \int m \cdot v_\perp \, dp \, ds
\]  \hspace{1cm} (A.1)

where \( g \) is the acceleration of gravity, \( \rho \) is the density of water, \( m \) the mixing ratio, \( ds \) and \( dp \) are the horizontal, respectively vertical dimensions of the cross-section and \( v_\perp \) denotes the velocity component perpendicular to the cross-section. In the control run the water flux amounts to 213 000 \( m^3 \text{H}_2\text{O s}^{-1} \). As stated in Newell et al. (1992), atmospheric rivers can transport water amounts equivalent to the Amazon (\( \approx 165 \, 000 \, m^3 \text{ s}^{-1} \)), or even exceed them.

**Orographic effects**

A mountain range as high as the Alps exerts a significant impact upon precipitation processes via forced lifting. The inverse Froude number \( \epsilon = \frac{N H_0}{U} \) (Smith, 1989) is used here to determine whether or to what degree the southerly flow towards the Alps was blocked by the latter or directed over them. By doing so, many processes which are important for a detailed assessment of the interaction of the flow with the topography are neglected. For instance, Reeves and Lin (2006) discuss the relevance of stable layer formation over the Po valley for the development of convection, and Rotunno and Ferretti (2001) discuss how low-level barrier-jets interacting with the complex topography of the Alps are associated with convergence-driven lifting and precipitation. Here we simply aim to determine the streamer’s impact on the details of the large-scale flow.

![Figure A.10: Vertical wind velocity in Pa s\(^{-1}\) for the control run at 0600 UTC 12 November (shaded). Contour lines depict horizontal wind speed in m s\(^{-1}\) parallel to the cross-section (perpendicular to the Alps as indicated in Fig. A.9a).](image)
past elements of the Alpine terrain.

An intensification of the precipitation occurs if the air is forced to surmount the Alps and consequently substantial lifting condensation takes place (inverse Froude number being smaller than 1). The reverse is true for flow around the Alps (inverse Froude number being larger than 1). A direct link exists between the inverse Froude number and an upper-level PV anomaly. If the PV anomaly is orientated perpendicular to the mountain ridge, as it is in this case, the associated increase of $U$ and decrease in $N$ will facilitate overflow.

We used the bulk estimate for the inverse Froude number as proposed by Reinecke and Durran (2008, see section 2 for more information). In the simulations the bulk inverse Froude number $\epsilon_b$ varies between $\epsilon_b = 0.61 - 1.52$. A cross-section showing the lifting of the air masses in the control run is shown in Fig. A.10. The upper-level wind speed approximately perpendicular to the Alps exceeds 40 m s$^{-1}$.

### A.4.3 Modifications of the streamer’s tip

**Stability**

Modifications of the PV field have a considerable impact on the dry stability. These changes in $N^2$ have a complex 3-dimensional structure. This is illustrated exemplarily in Fig. A.4 for the streamer’s tip. In Fig. A.4a the upper-level stratospheric high-PV intrusion, i.e. the streamer, and the troposphere underneath the western flank of the streamer (label B) are areas of high static stability. An area of reduced stability is present in the upper troposphere directly underneath the streamer (label A). The atmospheric boundary layer is highly unstable between 25$^\circ$ W and 10$^\circ$ W. Between these two unstable layers lies a layer of relatively stable stratification.

The effect of the PV modifications in the tip on this structure is illustrated in Figs. A.4c and d. They show the differences $\Delta N^2 = N^2_{\text{modified}} - N^2_{\text{orig}}$ for an enhanced (TA) and reduced (TR) tip amplitude. Overall the modifications (TA, Fig. A.4c and TR, Fig. A.4d) have opposite effects on $N^2$. Changes in the stability inside the streamer (TA, increased stability; TR, reduced stability) are accompanied by changes in the stability of opposite sign near the surface and in the low stability region A. The stability changes in the western flank area (B) and on the eastern flank between 750 and 600 hPa follow those inside the streamer. A complete switch of the stratification from unstable (TA and ERA-40) to stable (TR) occurs in the boundary layer area.

The effects of the upper-level PV modifications on the stability become even more com-
Figure A.11: CAPE values in J kg$^{-1}$ for the experiment with an enhanced tip (TA, shaded) at 0600 UTC 12 November 1996 using moisture treatment a) RH, b) QI and c) FI respectively. Panel d) shows CAPE values in J kg$^{-1}$ for the experiment with a reduced tip (TR, shaded) at 0600 UTC 12 November 1996 for the moisture treatment RH. The bold black line is 2 PVU contour at 320 K.

plex if moist effects are taken into account. CAPE, CIN and CTI are used here to characterize changes in the moist stability. All indices take moist and dry effects into account. CAPE and the CTI are sensitive to changes in the moisture and temperature stratification at different levels and therefore complement each other. CAPE is an integral measure of the convective potential of an entire air column and is quite sensitive to the distribution of near-surface moisture; the simple stability indices, on the other hand, are more robust in this respect, and provide a rough characterization of the low-to mid-tropospheric stability above the boundary layer. CIN provides information about the amount of energy that needs to be released before convection is initiated.
Here CAPE reacts particularly sensitive to alterations of the upper-level PV over the oceans (see Figs. A.11a and d). The temperature changes strongly influence the vertical stability (CAPE) because of the high relative humidity in the near surface layers. This is in contrast to $N^2$ at 850 hPa which decreases both over the ocean and over land (not shown). The changes in $N^2$ are located underneath the streamer and the stability along the eastern side of the streamer and in the Alpine region remains almost unchanged by tip PV-modifications. CIN is increased to 10 J kg$^{-1}$ in the tip region and to 200 J kg$^{-1}$ along the eastern flank of the streamer in the TR experiment. Hence while a misrepresentation of the tip will not directly affect the intensity of convection in the target area through destabilization it can still have an impact through the amount of moisture that is transported towards the Alps (see Table A.2).

The CAPE distribution for TA is very similar for the QI and RH approach (Fig. A.11a and b). CIN is increased to 20 J kg$^{-1}$ in the tip region and to 150 J kg$^{-1}$ along the eastern side of the streamer in the TR experiment using the QI moisture treatment (not shown). In the FI approach the low-level moisture is redistributed by the filter and the CAPE pattern looks different with considerably higher CAPE values over the Atlantic (Fig. A.11c).

**Moisture flux**

In the TA experiment the moisture fluxes are larger than in the control run (Table A.2), but the precipitation averaged over the two control areas is slightly lower in the French Alps and almost zero in the Swiss Alps (Table A.3). The reason for these discrepancies is a cyclonic wrap-up of the enhanced streamer tip that leads to a westward shift of the moisture flux maximum. In the TR setting the streamer does not start to roll-up cyclonically and moisture fluxes towards both the French and the Swiss Alps, albeit of reduced magnitude, are sustained (not shown).
Table A.3: Observed and modelled rain sums in mm accumulated between 0600 UTC 12 November until 0600 UTC 13 November averaged over the two boxes shown in Fig. A.3. The model-derived precipitation is separated into convective and large-scale precipitation. Values are given for the moisture treatments RH and QI.
A.4.4 Modifications of the streamer’s eastern flank

Stability

The effects on the stability are complex in the area of the French Alps, where significant amounts of precipitation fell. Dry stability (N^2) increases (decreases) for EA (ER) down to about 700 hPa. Between 700 hPa and the model topography at approximately 750 hPa the stability decreases (increases) for EA (ER) (not shown).

Looking at the moist stability using the CTI, an increase (decrease) of the stability is observed for ER (EA) under the streamer tip, under the eastern side and along the eastern side of the streamer. The stabilization for ER is robust for all three moisture treatments with the strongest amplification observed for the FI approach and the weakest response for the RH approach. The destabilization in case EA is most pronounced when using the QI approach. In the area of the French Alps the CTI indicates reduced (enhanced) stability for EA (ER).

Moisture fluxes

The PV modifications along the flanks of the streamer have a large effect on the low-level wind field and thereby the moisture flux. More quantitatively, a reduction ER (enhancement EA) of the PV along the eastern flank is accompanied by a reduction by 43 % (enhancement by 35 % ) of the wind field at the cross-section (Fig. A.9a) at 0600 UTC 12 November. Consequently, the moisture flux through the cross-section indicated in Fig. A.9a varies between 141 000 m^3 s^{-1} (66 %, ER) and 260 000 m^3 s^{-1} (122 %, EA) for the RH moisture treatment (Table A.2). If the specific humidity is kept constant (moisture treatment QI) the range of recorded values is larger and varies between 116 000 m^3 s^{-1} (54 %) for the ER case and 296 000 m^3 s^{-1} (139 %) for the EA case (valid at 0600 UTC 12 November). The range is larger than for the RH cases because of an additional temperature that effects the RH treatment. A positive upper-level PV anomaly induces colder temperatures at the surface in the EA setting. If the relative humidity is kept constant (RH), the colder temperatures will lead to a decrease in specific humidity. Hence in the RH setting, the positive contribution to the flux by the stronger low-level winds is partially cancelled by this negative humidity effect.

Using the FI approach the moisture fluxes amount to approximately 50 % of the values recorded for QI and RH for both experiments.

Inverse Froude number and the flow in the vicinity of the Alps

Changes in the amplitude of the upper-level PV along the eastern flank of the streamer affect the flow regime in the vicinity of the Alps substantially. The southerly upper-level
flow which impinges on the Alps from the south, varies between approximately 40 m s\(^{-1}\) in the control run (Fig. A.10) and more than 50 m s\(^{-1}\) for an enhanced (EA) respectively approximately 20 m s\(^{-1}\) for a reduced (ER) eastern flank of the streamer (Fig. A.12). This southerly flow is strongest (weakest) in the EA (ER) experiment throughout the troposphere. As a consequence ER is the only PV configuration where \(\epsilon_b > \epsilon_{\text{crit}}\) and a transition of the flow regime occurs between ER (\(\epsilon_b = 1.52\), blocked flow) and the control run (\(\epsilon_b = 0.87\), flow over the Alps).

A stronger streamer leads to a destabilization of the atmosphere and to an increase of the cross-barrier wind speed, both of which contribute to a reduction in \(\epsilon\) and enhanced flow over the mountains. This lifting in its turn enhances orographic precipitation. Strong maxima in the vertical wind velocity, that exceed the intensity of the vertical winds in the control run, are found in the EA experiment (Fig. A.12a). On the upstream side of the Alps the air lifted whereas sinking motion occurs on the downstream side. Coming back to the simple model proposed by Doswell et al. (1996), a strong eastern flank of the streamer influences both the moisture flux and the orographic lifting in a reinforcing way.

**A.4.5 A broader and shorter streamer**

In this experiment the structure of the streamer is modified using the stretching tool described in section A.2A.2.2. The SB streamer resembles the structure of the streamer in the 96 h forecast but it is located about 5° further east over the Alps (cf. Fig.A.13a).

**Stability**

Both CAPE and the CTI indicate that the atmosphere is less stable directly underneath the extended eastern flank of the SB streamer (see Fig. A.13a). CAPE reacts again very
Figure A.13: Panel a) shows CAPE in J kg$^{-1}$ for the experiment SB at 0600 UTC 12 November 1996. The bold line indicates the 2 PVU line at 320 K. Panel b) shows the vertical wind velocity in Pa s$^{-1}$ for the experiment SB at 0600 UTC 12 November 1996 (shaded). Contour lines depict horizontal wind speed in m s$^{-1}$ parallel to the cross-section (perpendicular to the Alps as indicated in Fig. A.9a).

sensitively in regions with high low-level moisture concentrations. In the CAPE field the atmosphere is particularly unstable over the Mediterranean off the coast of France. CIN values are increased in these areas. The land-sea contrast is less pronounced in the CTI. Along the Alpine southside both CAPE and the CTI register a labilization of the atmosphere, which is strongest in the QI moisture treatment but also substantial in the RH moisture experiment.

Moisture flux
The southerly wind maximum along the eastern flank of the streamer is shifted by about 4° to the east in the SB experiment at upper levels as well as at the surface but the amplitude is slightly higher (about 2-3 m s$^{-1}$ larger) in the latter compared to the ERA-40 data (not shown). The integrated humidity flux through the cross-section is slightly larger (13%) in the SB experiment if the QI moisture treatment is used and reduced by 8% if the RH moisture treatment is applied compared to the ERA-40 data. The reduction in the moisture flux using the RH moisture treatment is due to the previously described temperature effect. The temperature in the lowest levels is approximately 4 to 5° colder in the SB experiment leading to a considerable reduction of the moisture content of the air if the relative humidity is kept constant.

Inverse Froude number and the flow in the vicinity of the Alps
The upper-level southerly wind component perpendicular to the Alps is about 10 m s$^{-1}$ faster in the SB experiment compared to the control run (Fig. A.13b). The lifting on
the upstream side and the sinking motion on the downstream side of the Alps is slightly stronger (0.2 m s$^{-1}$) than in the control run. The eastward shift of the streamer positions the maximum of the incident flow directly at the cross-section. Strong upper-level winds of up to 54 m s$^{-1}$ decrease $c_b$ to 0.61. The lifting at the first mountain ridge (43N) in the experiment is comparable to the control run whereas the vertical motion at the second ridge (46N) is enhanced in the experiment.

### A.5 Precipitation

In this section the effects of the PV modifications on the precipitation are discussed based on the model simulations. To this end the wind and temperature fields from the PPVI are combined with the three different moisture fields and used as initial conditions for the CHRM model. The precipitation is validated in the two areas indicated by the boxes in Fig. A.3. To compare the different experiments we analyze precipitation fields after an integration time of 24 h (see Table A.3).

The large-scale structure of the precipitation with one maximum located along the Swiss Alpine south side (region 1) and one maximum over the French Alps (region 2) is well captured by the model. Along the Swiss Alpine south side (region 1) the rain amount is underestimated in the control run (Fig.A.14a) where an area mean of 16 mm (100 %) fell over 24 hours in the target area, while the interpolated rain gauge data set registered 36 mm. The precipitation maximum over the French Alps (region 2) is better captured and slightly overestimated by 8 mm (57mm, 100%). In the model more than 85 % of the precipitation along the Swiss Alpine south-side and more than 46 % of the precipitation over the northern French Alps is large-scale precipitation whereas the precipitation maximum over the southern French Alps is mainly convectively driven.

Averaged over the box located to the south of the Swiss Alps (region 1) the precipitation values in the different experiments with moisture treatment RH range from 5 mm (45 %) in the TA set-up to 48 mm (300 %) in the EA set-up (see Fig. A.14). In the experiment TA the moisture fluxes and the main precipitation area are shifted to the west compared to the control run (Fig. A.14c).

Note that the precipitation maxima are located north of the Alps in the ER experiment (Fig. A.14d). This is in good agreement with the flow tendency around the mountains (see section A.4) and the fact that the main moisture flux is directed around the Alps along their western flank. In the set-up with a shorter and broader streamer the precipitation maximum along the south side of the Alps (34 mm) is higher than in the control run and extending eastward along the Alps (Fig. A.14f). The shape and amplitude of
Figure A.14: Shown is the 24 hour accumulated precipitation (0600 UTC 12 November, until 0600 UTC 13 November) in mm (shaded) for a) the control run and the b) TR, c) TA, d) ER, e) EA, f) SB experiments using moisture treatment RH.

This precipitation signature is closer to the observed structure than the control run. This is due to a shift of the location of the main humidity transport route. The second precipitation maximum over the French Alps is underestimated in this experiment compared to the observations and the control run. The model precipitation is mainly triggered by the large-scale flow in the SB experiment along the Alpine south side. The convective contribution of about 24% is however larger than in the control run. This is in good
agreement with the larger CAPE and CTI values in this area in this experiment (see Fig. A.13a).

Overall, the difference in the accumulated area mean precipitation between the QI and the RH moisture treatment does not exceed 5 mm in all experiments with the exception of the ER experiment where the precipitation in the French Alps is very sensitive to the moisture treatment (see Table A.3). In the ER experiment humidity fluxes between the moisture treatment QI and RH determined at 0600 UTC 12 November differ by only 18% but precipitation sums integrated over 24 hours differ by 42% (see Table A.2). The humidity fluxes are calculated directly after the PPVI and reflect only the instantaneous situation whereas the precipitation depends on the moisture fluxes over a time span of 24 hours. Integrated over the time span of 24 h 629 000 m$^3$ are registered at the cross-section for moisture treatment RH. In the ER experiment using the QI approach humidity fluxes become progressively weaker and integrated over the 24 h period 556 000 m$^3$ (88% of the RH treatment) are recorded.

A.6  Summary and conclusions

This study presents a detailed analysis of the forcing factors that lead to a heavy precipitation event that occurred along the Alpine South side in November 1996. A piecewise PV inversion approach is used to analyze the dynamical and physical mechanisms that lead to the event. Besides the dynamical analysis this study contains methodology-oriented parts discussing three approaches to adapt the moisture field after the PV inversion as well as a detailed comparison of various indicators of dry and moist stability.

A.6.1  Discussion of the three moisture treatment

In the Data and Methods section we propose three approaches how to treat the moisture distribution when using PPVI namely i) to apply a strong filter to the moisture field to remove all structures below the planetary scale (FI), ii) to keep the absolute humidity constant and only correct areas of supersaturation (QI), and iii) to keep the relative humidity constant (RH). A detailed discussion of the effect of the three moisture treatments on the stability and the amount of moisture that is transported towards the Alps is given in section A.4. In summary it can be stated that for the QI and RH both the stability and the moisture fluxes are dynamically very similar. For the FI approach in the other hand we find widespread and non-local effects on the stability and a significant reduction in the amplitude of the moisture fluxes towards the Alps.
We do therefore not recommend to use a strongly filtered or a climatological mean moisture field for meso-scale modifications of the PV. This is further underlined considering the partition principle of PV and using linear scale arguments, the amplitude of the moisture error introduced by the filtering is expected to be larger than the inconsistencies between an altered flow field and the original moisture field both in its spatial extent and its amplitude. In addition relevant information on the synoptic-scale moisture distribution is lost through the filtering process.

Keeping the relative humidity constant is the physically most consistent method from a stability analysis point of view. We have used this approach as reference method throughout the paper. No definitive solution to the moisture treatment problem is offered. In our results the sensitivity of the vertical stability to the moisture treatment is comparable in magnitude to the response to the actual PV alteration. This indicates how important the moisture field can be for extra-tropical analyses and that further effort needs to be put into the development of moist PV inversion techniques such as the approach proposed by McTaggart-Cowan et al. (2003).

A.6.2 Summary of the PPVI experiments
Previous climatological and case-based studies have shown that very often so-called PV streamers located over Western Europe are the key synoptic-scale drivers behind the heavy precipitation events. As illustrated here, current state NWP models still have problems with the correct prediction and representation of these PV streamers. Error analysis reveals that streamers tend to be too broad and too short in the forecasts.

Piecewise potential vorticity inversion is used to assess effects of modifications of the upper-level PV distribution, that resemble the typical forecast errors, on the precipitation for the November 1996 case. The modifications are similar both in magnitude and structure to the observed forecast errors. The modifications of meso-scale PV structures located in the tip and along flanks of the streamer are evaluated with regard to their impact on the stability, the moisture flux and the orographic precipitation forcing.

**stability:** The impact of a meso-scale PV anomaly on the stability in general has a complex three-dimensional structure and is not limited to upper-tropospheric levels. Areas of low stability both directly beneath the PV streamer as well as in the lowest layer of the atmosphere between the ground and approximately 850 hPa are very sensitive to modifications of the upper-level PV. Over the ocean, where the air is very moist, small changes in the low-level temperature due to upper-level...
PV modifications have a major impact on the CAPE. The impact of the upper-level field on the stability over the precipitation area in the Alps is relatively limited and changes over time. Destabilization is therefore a second-order effect in the formation of HP in the Alps.

**moisture flux:** changes along the eastern flank of the PV streamer have a crucial impact on the moisture transport towards the Alps influencing both the amount of moisture that is transport and the location where in the strongest moisture fluxes hit upon the Alps. The moisture treatment has a substantial influence on the magnitude of the moisture fluxes.

**orographic forcing:** The upper-level PV structure has a crucial impact on whether the air is forced to flow over or around the topographic obstacle. Precipitation is enhanced for overflow conditions with a strong positive signal in the local lifting. An upper-level PV anomaly that is situated to the west of and orientated perpendicular to the mountain chain, as is the case for this example, causes a reduction in the stability and an increase in the flow component perpendicular to the mountain. As a consequence, the flow is more likely to cross the barrier, as quantitatively assessed by the inverse Froude number.

The outcome of the PV modification experiments clearly show that the structural errors often associated with the forecast of PV streamers can have a decisive impact on the quality of the precipitation forecast.

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Appendix B

Radiative Convective Advection Equilibrium

To investigate the behavior of our idealized framework without the variations caused by the diurnal cycle we perform a “radiative-convective equilibrium” (RCE)-like simulation. Because of its simplicity, where a time-invariant forcing is applied, this is a common way to investigate moist convection.

Radiative-convective equilibrium (RCE) is defined as a state, where the divergence of the net vertical radiative flux (short-wave and longwave) is compensated by the convergence of the vertical enthalpy flux in convective clouds (Emanuel, 2007). It is a statistical rather than an actual equilibrium but the global-mean state of the atmosphere is thought to exist in a state of RCE. The concept goes back to the early GCM modeling work of Manabe (1969) and it is especially valid in the tropics over sea surface, where the diurnal cycle of surface heat fluxes is small. A discussion of RCE can be found in Emanuel (2007). In contrast to mid-latitudes, where the quasi-geostrophic equilibrium explains a major part of the large-scale dynamics, RCE is one of the only equilibrium concepts for the tropics.

The concept of RCE has been used in numerous numerical models, both using parameterized convection (e.g. Held et al., 2007) and explicitly simulated convection (e.g. Tompkins and Craig, 1999). Frequently, the radiative cooling of the atmosphere is realized by applying a constant cooling rate of typically $-4 \text{ K day}^{-1}$, that brings about a destabilization of the atmosphere enabling convection and resulting precipitation. Values for surface temperature are kept constant (e.g. Parodi and Emanuel, 2009). RCE models are especially useful to infer the statistical distribution of clouds and convective mass-fluxes. Although for mid-latitude convection over land RCE is owing to the strong variations in the diurnal cycle of surface heat-fluxes not a good approximation, we nevertheless believe that it is an interesting experiment to perform with our model. Without the diurnal cycle much of the disturbance in time is removed and vertical profiles of convection-related quantities can be evaluated in a steady state.

To this end our idealized framework introduced in chapter 2 is used neglecting the diurnal cycle of incoming radiation in the current study. The incoming radiation is set such, that the shortwave radiation integrated over 24 h is approximately equal in the simulation with and without diurnal cycle in a dry simulation (meaning that both the humidity content of the atmosphere and the soil are set to zero). Short-wave radiation at the surface and at the top of the atmosphere (TOA) for the dry runs are shown in
B Radiative Convective Adveotive Equilibrium

**Figure B.1:** Surface (solid line) and TOA (dashed line) short-wave radiation for the simulation with diurnal cycle (red line) and the simulation for the RCE experiment (blue line).

Fig. B.1. Otherwise, the model setup is equal to the 60_CTL experiment in chapter 2 and 3 meaning that a height-dependent relaxation towards a profile with 70 respectively 40 % relative humidity and the intermediate stability (dT/dz=-0.7 K/100 m) is used in the atmosphere and a soil-moisture saturation of 60 % including vegetation is applied. Lex-pcor is set to .FALSE. (see Appendix C.1) and the sgs cloud scheme is switched on. The simulation is termed "rad_cst".

The time-series of cloud-water, cloud-ice, precipitation and 2 m temperature is shown in Fig. B.2. With the time-invariant forcing the model attains a state, where cloud water, cloud ice and precipitation is first simulated continuously, whereas later in time periods with a complete absence of clouds and precipitation are seen. 2 m temperature rises in the start of the simulation to \( \approx 23\,^\circ\text{C} \). From day 10 onward it shows oscillations between 22 and 24 \(^\circ\text{C}\). The evolution of CAPE and CIN, including the variability in the domain is shown in Fig. B.3. Until day 12 of the simulations there is a large variability of CAPE and CIN in the domain, indicating the presence of convective cells in multiple stages of their lifecycle in different parts of the domain. Starting roughly at day 12, there are periods, when the variability of the domain becomes very small and the 10th and 90th percentile nearly follow the domain mean value. There are points in time when CIN is completely removed. Cloudy points acquire CAPE values of up to 800 J kg\(^{-1}\) in the mean at these times. Investigating the spatial patterns the following picture emerges: convective activity starts at some point in the domain, organizes itself into larger clusters that move through the domain, collide, build even larger clusters and consume all
accumulated instability by convective activity. As soon as all CAPE is removed convective activity ceases in the whole domain. Thereafter it takes a certain amount of time until new instability is built up again.

The presence of these quasi-periodicities indicates that the size of the domain considered is too small. Convective cells spread over the periodic boundaries and collide with themselves again. Simulations were reconducted doubling the size of the domain from 220x220 km to 440x440 km. A frequency analysis of the domain mean 2 m temperature is shown in Fig. B.4. The oscillations commence also in the larger domain, but the period between two maxima in 2 m temperature is increased. Shorter simulations (15 days) using the UNSTABLE instead of the CTL profile of temperature show an earlier development of the oscillations (from 2 day onward) and a longer period between convective inactivity. Convective instability can be reduced more effective in the UNSTABLE atmosphere and a longer timespan for recharging is necessary. The strong organization observed emerges only if moisture is considered. In a dry run (qv=0) no oscillations developed.

Tompkins (2001) tackled the issue of domain size by simulating RCE convection over constant SSTs in a large channel with an extension of 64 km in the meridional, and 1024 km in the zonal direction using a grid-spacing of 2 km. Our experiment is repeated using their grid setup with 512x32 grid points resulting in a domain of 1126 kmx70 km. Hovmoller diagrams of 2 m temperature and surface precipitation for this simulation are shown in Fig. B.5. In the early phase of the simulation, convective cells develop all over the domain. These single convective cells organize into larger clusters. All

![Figure B.2: Domain mean (a) cloud water (shade, kg kg$^{-1}$), cloud ice (contour lines, kg kg$^{-1}$) and surface rain rate (black line, mm h$^{-1}$) and (b) 2 m temperature ($^\circ$C) for rad_cst.](image)
Figure B.3: (a) CAPE (J kg$^{-1}$) and (b) CIN (J kg$^{-1}$) values with the domain mean value in black and the mean over cloudy points in blue for rad_cst. The solid lines indicates the domain mean, whereas the shaded area indicates the 10th and 90th percentile of values over the domain.

Figure B.4: Periodogramm of the domain mean 2m temperature for rad_cst (black line), the simulation with the increased domain size (blue line), the simulation using the UNSTABLE profile (green line) and for comparison the 60_CTL simulation including the diurnal cycle of incoming radiation.

cells however organize into one “supercluster” after around 6 days. This supercluster consists of many individual convective cells that are triggered, grow and decay again. The supercluster of organized cells moves against the background flow with a speed of approximately 24.2 km h$^{-1}$. No other cells away from the supercluster build up. This result is in contrast to the simulations of Tompkins (2001), who observed many separate clusters, all moving against the background current. The difference could be related
Figure B.5: Hovmöller (longitude-time) plot of (a) precipitation (mm h$^{-1}$) and (b) 2 m temperature ($^\circ$C) for the simulation in the large channel. Quantities were averaged over the meridional direction.

to either the relatively strong background wind including vertical shear supporting the organization of convection (e.g. Rotunno et al., 1988) imposed in our simulations, or by the land-surface instead of the ocean surface used, favoring organization by moistening the soil through precipitation. Over wet soils the LFC is lowered enabling the triggering of convection (e.g. Taylor et al., 2010).

Due to this strong organization and the resulting periodicities, our framework is not fully
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suitable for RCE simulations since it constitute a time-variant disturbance. Compared to the simulations with diurnal cycle the variations of e.g. 2m temperature in rad_cst are relatively small such that we are “more in equilibrium” than in the standard setup 60_CTL. In the following, the simulations performed with the small domain (rad_cst) is investigated more closely. Heat and moisture budgets of the simulation are documented in the Appendix C.1. In the mean, radiative cooling nearly offsets solar heating (see Table C.2), resulting in an atmosphere that is unstable to the displacement of moist air. The heating released in the convective updrafts on the other hand is almost removed by the relaxation. Instead of a pure radiative convective equilibrium, a radiative-convective-relaxation equilibrium develops. Speaking in more physical terms and recalling that the relaxation can be thought of an advection, we can term it Radiative-convective advection equilibrium. The additional cooling imposed by the cold air advection enables additional convection. In the tropics in contrast, where weak advection prevails, the atmosphere needs to radiate away all excess energy gained by convection to destabilize the atmosphere sufficiently for new convection to occur.

To infer the vertical distribution of clouds and precipitation, domain mean values of the simulations are averaged over time starting from day 10 onwards. Vertical profiles of cloud-fraction, clouds, precipitation, convective mass-fluxes, relative humidity and a skew-T diagram of the equilibrium state are shown in Fig. B.6. To estimate the variability in time the minimum and maximum of domain mean values from day 10 onwards is shown by thin lines. Cloud-fraction exhibits a first maximum of water clouds at around 720 hPa and a second maximum at 250 hPa caused by ice clouds. The cloud fraction of ice anvils is about ten times larger than the cloud-fraction of water cloud. This could also be related to the different threshold used for the computation (10^{-6} kg kg^{-1} for water and 10^{-10} kg kg^{-1} for ice). Water clouds show a larger variability in time than ice clouds. The amount of water and ice contained is however roughly comparable for water and

Figure B.6 (following page): Average domain mean profile of (a) cloud-fraction, (b) cloud water and cloud ice (kg kg^{-1}), (c) graupel, snow and rain (kg kg^{-1}), (d) upward and downward convective mass-flux (kg m^{-2} s^{-1}), (d) relative humidity (%) and (e) skew-T diagram with temperature in black, dew-point temperature in blue and the ascent of an undiluted parcel starting from the lowest atmospheric level in green. In (a), (b), (c) and (d) the minimum and maximum of domain mean values from day 10 onwards are indicated by the thin lines. In (e) the shaded area indicates the 10th and 90th percentile of simulated values in the domain starting from day 10 onwards.
ice, meaning that ice clouds cover a larger area than water clouds. The freezing level is situated at 700 hPa, (see Fig. B.6f), though liquid water is found up to an altitude of 300 hPa, where a temperature of about -40°C prevails, which indicates that in the updrafts cloud water remains unfrozen.

Falling hydrometeors exist in and below the clouds. Snow falls down to an altitude of 600-700 hPa, where it melts or evaporates. Graupel falls further down to 800 hPa due to its larger terminal velocity. Rain is present below 600 hPa. The variability in time of graupel is considerably larger than for snow.

Convective mass-fluxes were determined using a threshold of $w=1 \text{ m s}^{-1}$, both for upward and downward mass-fluxes. This threshold is commonly used for defining upward mass-fluxes but can also be used as a threshold for downward mass-fluxes, even if subsidence is slower than upward convection (see Tompkins and Craig, 1999, and references therein). Upward mass-fluxes show a maximum at a height of 700 hPa, coinciding with the cloud-base. Downward fluxes are considerably smaller and shifted to lower levels. They reach their maximum between 650 and 850 hPa, the height where snow and graupel are converted into rain. Heat for melting is required here, the air is thereby cooled producing cold downdrafts falling to the surface. Evaporation of falling rain leads to further cooling.

In the profile of relative humidity (Fig. B.6 e) the spatial variability is additionally included into the graph. The shaded area indicates the 10th respectively 90th percentile.
of values simulated from day 10 onwards in the whole domain. A maximum of RH between 900 and 700 hPa is visible, which is shortly below the cloud base shown in panel a). Above 600 hPa values are relatively constant with height until the tropopause is reached, where moisture is virtually absent. The near-constant mid-tropospheric values are maintained by the continuous convection that detrains water into the air.

The mean skew-T diagram (Fig. B.6f) shows a nearly dry-adiabatic profile below 750 hPa (below the cloud base), and a nearly moist-adiabatic profile above until a height of 400 hPa. Above this height the profile is more stable than moist-adiabatic again. This is probably due to the strong relaxation of temperature in the upper troposphere.

Fig. B.7 shows a histogram of upward velocities inside the PBL (950 hPa, panel a) and in the free troposphere (535 hPa, panel b). Inside the PBL the skewness is neutral, as convection is mostly dry. In the free troposphere a strong positive skewness, typical for deep, moist convection is visible. This skewness indicates that updrafts occupy a relatively small area and exhibit high velocities, whereas downdrafts are spread over considerably larger domains and have smaller velocities.

Fig. B.8 shows an instantaneous view on the convection in the RCE. The major part of clouds is collocated into one cloud cluster. Some updrafts already exhibit an ice anvil with falling precipitation, whereas other updrafts consist of liquid water only. At the surface temperatures are cooler below the cloud cluster. In other areas of the domain there are cirrus clouds remaining from older convective updrafts and some shallow low-level water clouds. Below the shallow cumuli a cooling of the surface can be seen.

The study neglecting the diurnal cycle gave us deeper insights into convective processes simulated with the idealized framework. The various different processes involved in the formation of clouds and precipitation could thereby be investigated further.
Figure B.8: Instantaneous view on the convection in the RCE after 514 h of integration. Pink volumes indicate cloud water, blue volumes cloud ice and green volumes hydrometeors (graupel, snow and rain). The skin temperature is displayed on the plane, values are given by the color-code.
Appendix C

Model sensitivity studies

This chapter contains several sensitivity studies for the numerical model used. Having a survey on the model sensitivity we get an estimate of the consistency of the results presented above. Section C.1 documents the heat- and moisture budgets of the simulations as well as the influences of the explicit moisture correction in the turbulence calculation on the simulations. This switch has a rather large influence on the moisture budget of the simulations and proofed to erroneously worsen the moisture budget considerably.

In section C.2 the control of the relaxation on the simulations is investigated by using different time scales for the relaxation constant. Section C.3 and C.4 investigate physical parameterizations, namely the subgrid-scale cloud scheme affecting turbulence and radiation, the shallow convection scheme and the parameterization of ice microphysics. Sections C.5- C.8 focus on numerical aspects such as the time-step, the horizontal and vertical resolution of the grid and the computational horizontal diffusion. In section C.9 a short summary on the sensitivity tests is given.

The default setup for the simulations is given in Table C.1, where the number in the third row indicates the section, where the specific parameter will be addressed. Except the mentioned section, variables are always set to their default values. The parameter \( \text{tur} \text{len} \) is the asymptotic turbulent length scale in the Blackadar (1962) formulation for the mixing length for turbulence. All of the above documented simulations do not perform the erroneous explicit moisture correction, and for the sensitivity tests shown in the following only the simulations concerning the vertical resolution (section C.7) and the ice microphysics (section C.4) include the erroneous correction.

<table>
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<th>SGS</th>
<th>( c^i_{au} )</th>
<th>( v_0_s )</th>
<th>( \text{agg} )</th>
<th>( \Delta t )</th>
<th>( \Delta x )</th>
<th>nz</th>
<th>adv.</th>
<th>( \alpha_u )</th>
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<td>s(^{-1})</td>
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<td>Bott</td>
<td>C.8</td>
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</table>

Table C.1: Setting of the default model setup for the parameters for the relaxation constant \( \tau \), the constants in the ice microphysics \( c^i_{au} \), \( v_0_s \) and \( \text{agg} \), the constant for computation horizontal diffusion on temperature \( \alpha_t \) and on humidity \( \alpha_q \) and the asymptotic turbulent length scale in the Blackadar (1962) formulation \( \text{tur} \text{len} \). The different parameters are addressed in the section indicated in the third line.
C Model sensitivity studies

C.1 Heat and moisture budgets and explicit moisture correction

An analysis of the heat and moisture budget of the model is performed. First, we use the simulations without diurnal cycle of radiation introduced in Appendix B. Second, we investigate the simulation 80_STABLE introduced in chapter 3 including the diurnal cycle of radiation. Furthermore we illustrate the effect of the explicit moisture correction performed in the turbulence routine on the moisture budget for these simulations.

Regarding the moisture correction, the turbulence routine acts on the moist-conserved variables liquid water potential temperature \( \theta_l = \theta \cdot (1 - \frac{L_v}{\rho_c q_c}) \) and total water content \( q_t = q_v + q_c + q_r \). The calculation of the turbulent fluxes is based on K theory, where the flux is related to the vertical gradient of the quantity: e.g. \( w'\theta'_l = -K_H \frac{\partial \theta_l}{\partial z} \) and \( w'q'_t = -K_H \frac{\partial q_t}{\partial z} \), where \( K_H \) is the turbulent diffusion coefficient for heat and water vapor, the overbar denotes an averaging of the quantities and primes denote residual anomalies subtracted from the mean state. To derive the turbulence tendencies for temperature, specific humidity and specific cloud water and to obtain the phase transitions between vapor and liquid an explicit moisture correction has to be applied to the above mentioned tendencies after the turbulence routine. The model however performs a saturation adjustment after the turbulence and the explicit correction should have no effect. An incorrect implementation of the terms however leads to a violation of the moisture budget and to artificial sources of water vapor.

The influence of the explicit moisture correction is investigated for simulations including the sub-grid scale (SGS) cloud scheme and for simulations without. The details of the used SGS cloud scheme will be explained later (see section C.3).

C.1.1 Constant radiation

In a first step the simulations with constant incoming radiation are shown with the following set of four simulations:

- **rad_cst**: lexcpor=.FALSE., with SGS clouds, identical to the simulation shown in Appendix B.
- **rad_cst_lexcpor**: lexcpor=.TRUE., with SGS clouds
- **rad_cst_nosgs**: lexcpor=.FALSE., no SGS clouds
- **rad_cst_lexcpor_nosgs**: lexcpor=.TRUE., no SGS clouds
C.1 Heat and moisture budgets and explicit moisture correction

Averaging of all displayed quantities is always performed in time from day 10 to day 30 and spatially over the domain. Mean profiles of cloud water, cloud ice and hydrometeors for the simulations are shown in Fig. C.1 and C.2. Mean values of ice clouds remain nearly unchanged between the simulations. A slight upward shift of ice clouds for lexpcor=.FALSE. can be seen. The variability of ice clouds moreover increases for lexpcor=.FALSE. The mean values of low-level clouds are significantly reduced (about a factor of two) if lexpcor is set to .FALSE.. For simulations using the SGS cloud scheme the variability of low-level clouds is considerably enlarged, whereas the mean values show rather small changes with small decreases for lexpcor=.TRUE. and slight increases for
C Model sensitivity studies

**Figure C.2:** Mean profile (bold lines) and minimum and maximum value of the domain mean value (thin line) of graupel, snow and rain for (a) rad cst, (b) rad cst lexpcor, (c) rad cst nosgs and (d) rad cst lexpcor nosgs.

lexpcor=.FALSE.

Mean graupel and snow amounts remain nearly unchanged with a small shift to higher levels for lexpcor=.FALSE. Spatial variability for snow and graupel increases in accordance with the increase in variability of the ice clouds. Together with the observed reduction in low-level cloud cover the rainfall amounts decrease.

Fig. C.3 displays mean profiles of relative humidity and potential temperature. Values are increased by roughly 10% below 700 hPa if the moisture correction is switched on. The SGS cloud scheme leads to an upward shift of 50-100 hPa of the moisture peak.
C.1 Heat and moisture budgets and explicit moisture correction

Figure C.3: Mean profiles of (a) relative humidity (%) and (b) potential temperature (K) for rad_cst (green), rad_cst_lexpcor (red), rad_cst_nosgs (blue) and rad_cst_lexpcor_nosgs (black).

Potential temperature increases below 700 hPa if the moisture correction is switched off. The use of the SGS cloud scheme leads to a stronger inversion at the top of the PBL.

Fig. C.4 shows time series of domain mean surface rain rate for the set of simulations. Striking is the occurrence of periods without rain for lexpcor=.FALSE.. This phenomenon has been discussed in Appendix B and is related to the strong organization of convection across the domain resulting in periodic appearances of cloud clusters that consume all instability present. This feature is absent if the moisture correction with its resulting larger low-level humidity values is performed.
Fig. C.5 shows vertical profiles of heating rates averaged from day 10 onwards resulting from the different parameterizations. They were calculated using a budget tool implemented into the CCLM by Wolfgang Langhans (personal communication). The term “dpdt+advfw” stems from the fast-waves solver and combines the vertical advection of the background atmosphere \( \partial p/\partial t \) and the advection in the fast-waves. The term “advection” comprises advection from the slow modes only. The Rayleigh damping at the upper boundary of the model is denoted as “sponge”.

Solar heating (orange line) is strongest in the stratosphere, where the absorption of ultraviolet radiation in the ozone layer occurs. An increase of solar heating near the surface is visible, presumably caused by the absorption of short-wave radiation by clouds and water vapor in the lower troposphere and absorption of reflected radiation from the surface. Longwave radiation (red line) acts to cool the atmosphere through radiative cooling. Values are large in the lower troposphere, where temperatures are higher and the moisture content is larger. Depending on the cloud amount simulated, the long-wave cooling around 700 hPa is more or less pronounced. It increases considerably using the SGS cloud scheme. Latent heating (blue line) shows positive values between 650 and 250 hPa, the height where the strongest heating in convective updrafts due to the release of latent heat occurs. The structure in the lower troposphere is more complex and depends strongly on the amount of clouds simulated. For simulations using the SGS cloud scheme there is maximal heating around 750 hPa, the height, where the cloud bases are visible in Fig. C.1. At around 700 hPa there is a relative minimum of latent heating. The height coincides with the freezing-line, and the cooling occurs due to the heat required to melt falling snow and graupel. Boundary-layer values are negative because of the evaporative cooling of falling precipitation. For the simulations using lexpcor=.TRUE. the peak at 800-850 hPa is more pronounced. It coincides with the peak in relative humidity seen in Fig. C.3. The relaxation acts to cool the atmosphere and is strongest between 350 and 400 hPa where it compensates the latent heating in the convective updrafts. The sponge is active above 200 hPa. It absorbs gravity waves generated by convection. Turbulence is mostly active in the lower troposphere and compensates latent heating partially. Advection redistributes heat mostly vertically.

To get an estimate of the performance of the model in conserving the heat budget the following terms are considered: solar heating, thermal cooling, latent heating, surface sensible heat flux, the heating resulting from the relaxation towards the background profile and the relaxation in the sponge. The other terms are only internal conversion terms and are thus not considered. They are moreover not appropriate to determine
the budget, since they pass possible errors on to the next routine and discrepancies are cancelled out. Domain-mean values of considered heating rates are given in Table C.2. They are derived by averaging in time between \( t_1 = 10 \text{ d} \) and \( t_2 = 30 \text{ d} \), over the zonal and meridional direction and over height:

\[
X_{\text{mean}} = \frac{1}{t_2 - t_1} \int_1^{x_{\text{tot}}} \int_1^{y_{\text{tot}}} \int_{z_{\text{surf}}}^{z_{\text{top}}} \int_{t_1}^{t_2} X \, dt \, dz \, dy \, dx
\]  
(C.1)

where \( X \) stands for any of the terms considered, \( z_{\text{top}} \) is 22000 m, \( z_{\text{surf}} \) is 489 m and \( x_{\text{tot}} \) and \( y_{\text{tot}} \) are the zonal and meridional extent of the computational domain, respectively.

To convert heating rates into heat fluxes they are multiplied by the specific heat of dry air \( c_p = 1005 \text{ J kg}^{-1} \text{ K}^{-1} \), the density of moist air \( \rho_v \) and integrated vertically. \( \rho_v \) is calculated using

\[
\rho_v = \frac{P}{R_d \cdot T \cdot (1 + \left( \frac{R_v}{R_d} - 1 \right) \cdot q_v - q_c)}
\]  
(C.2)

where \( P \) is pressure, \( T \) temperature, \( q_c \) cloud water and \( R_d \) and \( R_v \) the gas constant for dry air respectively water vapor.

The budget of heat is conserved satisfactory with a maximal error of 11 % of the solar heating in rad\_cst\_nosgs with a decreases of the error if the SGS cloud scheme is utilized.

Interesting to note is that the simulations are more or less in a radiative equilibrium with the radiative cooling balancing the radiative heating. The relaxation on the other hand nearly offset the latent heating of convection. The sponge yields a considerable portion of the heating. It even exceeds surface sensible heating.

For the budget of water, soil and atmosphere are considered separately. The relaxation in the atmosphere acts on specific humidity \( q_v \). Relaxation terms \( \frac{\partial q_v}{\partial t} \) are converted into precipitation units \( \frac{\partial W_{\text{relax}}}{\partial t} \) multiplying by the density of moist air \( \rho_v \) and integrating over height:

\[
\frac{\partial W_{\text{relax}}}{\partial t} = \int_{z_{\text{surf}}}^{z_{\text{top}}} \rho_v \cdot \frac{\partial q_v}{\partial t} \, dz
\]  
(C.3)

The relaxation acts to dry and is strongest at a height of 700 hPa, the height at which the maximum amount of clouds is present (see Fig. C.1). All terms of the atmospheric budget (ET, precipitation and relaxation) are given in Table C.3. The relaxation in the sponge is negligible and thus not considered. The conservation of moisture is with 8 % of the ET satisfactory for rad\_cst but with a relative error of 36 % in
Figure C.5: Vertical profiles of time mean horizontally-averaged heating rates for (a) \text{rad\_cst}, (b) \text{rad\_cst\_lexpcor}, (c) \text{rad\_cst\_nosgs} and (d) \text{rad\_cst\_lexpcor\_nosgs}. The individual components are solar (orange), thermal (red), latent (blue), relaxation (black), sponge (grey), advection (green), turbulence (pink), \text{dpdt+adv\_fw} (red) and the residual (grey line) heating rates (K s\(^{-1}\)).

\text{rad\_cst\_lexpcor\_nosgs} considerable. Precipitation is too large by a factor of approximately 1.5 in the simulation \text{rad\_cst\_lexpcor\_nosgs}. The conservation of the moisture budget is considerably improved setting \text{lexpcor} to .FALSE. The difference is moreover smaller if the SGS cloud scheme is utilized.

The relaxation of soil water dries the soil in the uppermost soil layer, where drainage of rain water occurs. At the depth where roots extract water for transpiration the relaxation acts to refill the water. At a depth of \approx 0.7 m the relaxation dries the soil again and at the lowest active layer, where runoff from the ground occurs it refills again (not shown). Values for the different components are given in Table C.4. ET and runoff are moisture sinks, whereas precipitation constitutes a source of water. The budget of the soil is
C.1 Heat and moisture budgets and explicit moisture correction

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<tr>
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<th>solar</th>
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<th>sponge</th>
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Table C.2: Time and domain mean heating rates for the different components of the heat budget given in heating rates ($hr, 10^{-5}$ K s$^{-1}$) and in heat fluxes ($hf, Wm^{-2}$).

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</table>

Table C.3: Average values of the different components of the atmospheric moisture budget (mm h$^{-1}$).

nearly closed in all simulations.

In summary we find that the heat and moisture budget of the soil is conserved well for the simulations without diurnal cycle of incoming radiation. However by performing an explicit moisture correction in the turbulence routine the moisture budget of the atmo-

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</table>

Table C.4: Average values of the different components of the soil moisture budget (mm h$^{-1}$).
C Model sensitivity studies

sphere is strongly violated. Artificial moisture is added to the boundary layer where it produces additional low-level clouds leading to increased precipitation amounts. The constant increment in PBL moisture furthermore damps the organization of convection into superclusters over the domain that causes quasi-periodicities in convective activity.

C.1.2 80_STABLE, including diurnal cycle

The diurnal cycle of incoming radiation is taken into account in the current study to investigate the budgets of water and heat as well as the influence of the switch lexpcor. The naming of the simulations is as follows:

80_STABLE: lexpcor=.TRUE., SGS=.FALSE., identical to the simulation shown as 80_STABLE in chapter 3

80_STABLE_lexpcor: lexpcor=.TRUE., SGS=.TRUE.

80_STABLE_nosgs: lexpcor=.FALSE., SGS=.FALSE.

80_STABLE_lexpcor_nosgs: lexpcor=.TRUE., SGS=.FALSE.

Average values are derived over day 16-30 of the simulations and spatially over the domain.

Fig. C.7 shows simulated values of relative humidity. As already seen in the simulations neglecting the diurnal cycle the moisture correction leads to a strong artificial increase of low-level moisture. The SGS cloud scheme produces a low-level peak of relative humidity around 900 hPa corresponding to the top of the PBL.

The mean diurnal cycle of cloud water, cloud ice and surface rain rate for the simulations is shown in Fig. C.6. Stunning is the almost complete removal of all mid-level clouds in

<table>
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</tbody>
</table>

Table C.5: Average values of the different components of the heat budget given in heating rates (hr, 10^{-5} K s^{-1}) and in heat fluxes (hf, W m^{-2}).
C.1 Heat and moisture budgets and explicit moisture correction

<table>
<thead>
<tr>
<th></th>
<th>ET</th>
<th>precipitation</th>
<th>relaxation</th>
<th>residual</th>
</tr>
</thead>
<tbody>
<tr>
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<td>-0.139</td>
<td>-0.0393</td>
<td>0.0131</td>
</tr>
<tr>
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<td>-0.150</td>
<td>-0.0305</td>
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<tr>
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<td>-0.0535</td>
<td>0.0242</td>
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<tr>
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<td>0.148</td>
<td>-0.166</td>
<td>-0.0449</td>
<td>-0.0629</td>
</tr>
</tbody>
</table>

**Table C.6:** Average values of the different components of the atmospheric moisture budget (mm h\(^{-1}\)).

<table>
<thead>
<tr>
<th></th>
<th>ET</th>
<th>prec.</th>
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<th>runoff surf.</th>
<th>relaxation</th>
<th>residual</th>
</tr>
</thead>
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<tr>
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<td>-0.0911</td>
<td>8.47</td>
<td>0.00173</td>
</tr>
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<td>80_STABLE_lexpcor</td>
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<td>0.150</td>
<td>-8.33</td>
<td>-0.0995</td>
<td>8.43</td>
<td>0.0012</td>
</tr>
<tr>
<td>80_STABLE_nosgs</td>
<td>-0.225</td>
<td>0.148</td>
<td>-8.32</td>
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<td>8.50</td>
<td>0.000406</td>
</tr>
<tr>
<td>80_STABLE_lexpcor_nosgs</td>
<td>-0.148</td>
<td>0.166</td>
<td>-8.33</td>
<td>-0.107</td>
<td>8.42</td>
<td>0.0119</td>
</tr>
</tbody>
</table>

**Table C.7:** Average values of the different components of the soil moisture budget (mm h\(^{-1}\)).

The morning hours between 80_STABLE_lexpcor_nosgs and 80_STABLE_nosgs and the strong reduction of daily mean precipitation amounts. The character of the diurnal cycle of convection is completely changed by the explicit moisture correction. If set to .TRUE. convection is strongly damped as compared to 80_STABLE_nosgs, where deep, active convection occurs. The effect is not as drastic if the SGS cloud scheme is used, but still considerable.

Budgets for heat and water are calculated as described in the previous section. Values are given in Table C.5 for heat and in Table C.6 and C.7 for atmospheric and soil moisture respectively.

Mean vertical heating rates averaged over the diurnal cycle are shown in Fig. C.8. The diurnal cycle of radiation induces a larger variability in time and height than if a constant radiation is used. The averaged profiles show thus more details. If lexpcor is set to .TRUE. a strong low-level peak in latent heating appears. The low-level peak is shifted to a higher altitude using the SGS cloud scheme.

The balance between solar and thermal heating observed for rad_cst is moreover not valid. Radiative cooling becomes larger than solar heating, presumably caused by the absence of the latter during night. Radiative cooling is increased using the SGS cloud.
Figure C.6: Mean diurnal cycle of cloud water (shade, kg kg\(^{-1}\)), cloud ice (contour lines, kg kg\(^{-1}\)) and surface rain rate (mm h\(^{-1}\), black solid line and grey shaded area indicating day-to-day variability) for (a) 80_STABLE, (b) 80_STABLE_lexpcor, (c) 80_STABLE_nosgs and (d) 80_STABLE_lexpcor_nosgs. Mean diurnal precipitation amounts (mm day\(^{-1}\)) are given by the numbers in the lower left corner.

scheme which is caused by trapping of outgoing long-wave radiation. Latent heating is strongly increased if the moisture correction is switched off. This is also valid for ET (ET and latent heating in the atmosphere should be equal in the equilibrium). This stems from the decrease of near-surface humidity and the resulting increase of the saturation deficit that strongly controls bare soil evaporation and the potential evaporation (see Appendix D.1.1).
C.1 Heat and moisture budgets and explicit moisture correction

**Figure C.7**: Profiles of relative humidity (%) at (a) 0000 and (b) 1200 UTC. The shaded area indicates the 10th and 90th percentile of values over the domain starting from day 15 onwards, the solid lines show the mean value of the domain mean value averaged from day 15 onwards, separately for each hour of the day.

As stated above, the model shows a good performance in the conservation of the heat budget. Errors become smaller if the diurnal cycle is included. The good performance in keeping the soil moisture can moreover be confirmed. For the budget of the atmosphere again a strong violation of the moisture budget occurs if lexpcor=.TRUE.. In these cases ET decreases and precipitation increases at the same time. The increase of precipitation can be attributed to the artificial moisture source in the turbulence routine. The errors are about equally large as in the rad_cst runs.

Summarizing we strongly recommend to set lexpcor to .FALSE. as the moisture correction acts as an erroneous moisture source, thereby violating the moisture budget. The artificial moisture source moreover decreases ET by reducing the near-surface saturation deficit and increases at the same time cloud cover and thereby precipitation amounts. This is observed both for simulations with and without the diurnal cycle of radiation.
Figure C.8: Vertical profiles of time mean horizontally-averaged heating rates for (a) 80_STABLE, (b) 80_STABLE_lexpcor, (c) 80_STABLE_nosgs and (d) 80_STABLE_lexpcor_nosgs. The individual components are solar (orange), thermal (red), latent (blue), relaxation (black), sponge (grey), advection (green), turbulence (pink), dpdt+adv_FW (red) and the residual (grey line) heating rates (Ks⁻¹).
C.2 Relaxation

The relaxation of predicted variables towards the reference profile is a central element of the framework used in this thesis (cf. chapter 2). It represents the large-scale synoptic forcing on convection. The time constant $\tau$ determines how strongly the modeled atmosphere is relaxed towards the reference profile, or how active large-scale advection functions. For the relaxation time scale a default value of $\tau=24$ h is chosen. The purpose of the current section is to document the outcome of experiments using altered values, namely $\tau=1, 3, 6, 12$ and $48$ h.

The reference setup is the CTL experiments using the SGS cloud scheme and no explicit moisture correction in the turbulence scheme. A soil moisture of $S=60\%$ together with the intermediate stability and the intermediate atmospheric humidity are used ($dT/dz=-0.7 \text{ K (100 m)}^{-1}$ and RH=70, 40\%).

Fig. C.9 shows the mean diurnal cycle of the relaxation tendency on the specific humidity and Fig. C.10 the tendencies on temperature for the set of experiments. The tendencies mostly act to dry and cool the the atmosphere to counteract the moistening and heating of the ongoing convection. Relaxation drying is strongest in the middle troposphere ($\approx 600-700$ hPa), cooling is strongest in the upper troposphere where the

![Figure C.9](image)

**Figure C.9:** Relaxation tendency on specific humidity ($10^{-3}$ kg kg$^{-1}$ s$^{-1}$) for the experiments using different relaxation timescale $\tau$. 
Figure C.10: Relaxation tendency on temperature ($K s^{-1}$) for the experiments using different relaxation timescale $\tau$.

Figure C.11: Domain mean skew-T log-P diagram of the simulations using different relaxation time scales averaged over day 16-30 of the simulations at (a) 0000 UTC and (b) 1200 UTC. Temperature is shown by the solid lines and dew-point temperature by the dashes lines. The reference profile is shown by the grey solid line.
C.2 Relaxation

Figure C.12: Mean diurnal cycle of cloud water (shaded area, kg kg$^{-1}$), cloud ice (contour lines, kg kg$^{-1}$) and surface rain rate (black solid line and grey shaded area, mm h$^{-1}$) for the experiments using a different relaxation timescale $\tau$. Mean diurnal precipitation amounts (mm day$^{-1}$) are given by the numbers in the lower left corner.

latent heat released by condensation and freezing needs to be balanced. As to be expected, the tendencies are larger for smaller timescales.

Fig. C.11 shows a skew-T log-P diagram of the set of simulations at 0000 and 1200 UTC. It can clearly be seen that a shorter relaxation timescales confines the profile stronger to the reference profile. The design of the height-dependent relaxation in constraining the upper-tropospheric profile strongly, while the lower-tropospheric profile evolves nearly unconstrained is moreover visible.

In Fig. C.12 the resulting mean diurnal cycle of cloud water, cloud ice and surface rain rate is shown. The qualitative behavior is mostly unchanged by a change in the time scale. The mid-level cloud cover vanishes for small values of $\tau$ as the atmosphere is sufficiently dried to counteract the moistening caused by radiative cooling. Precipitation amounts are slightly decreased for shorter relaxation timescales. This is presumable caused by the removal of water from the atmosphere by the relaxation. Furthermore, a more sudden onset of cloud formation is visible for shorter timescales as it removes variability in space by constraining the profile stronger.
In Fig. C.13 the mean diurnal cycle of precipitation and 2 m temperatures are displayed. The onset of precipitation occurs at the same time in all simulations but is more sudden using shorter timescales. 2 m temperatures furthermore increase with increasing timescale. If the atmosphere is allowed to act more freely the PBL can heat up stronger and precipitation evolves over longer timescales.

The experiments confirm that the chosen timescale of $\tau = 24$ h is an intermediate value, where the model is on the one hand considerably constrained to the prescribed background profile, but where on the other hand enough variability of the variables over the domain is allowed.

Figure C.13: Mean diurnal cycle of (a) precipitation (solid line, mm h$^{-1}$) with the minimum and maximum values of domain mean precipitation over the 15 simulation days indicated by the shaded area and (b) 2 m temperature (°C) for the set of simulations using different relaxation timescales.
C.3 Sub-grid scale cloud scheme, shallow convection scheme

A grid-spacing of $\Delta x = \Delta y \approx 2.2 \text{ km}$ is sufficient to resolve the bulk properties of organized convection (e.g. Weisman et al., 1997). The resolution is however not capable of determining the details of clouds or shallow convection. The CCLM model thus includes a sub-grid-scale (SGS) cloud scheme affecting radiation and turbulence, and a shallow convection scheme. In the default setup the SGS scheme is considered but the shallow convection scheme not. In the following the effect of those two schemes is demonstrated.

C.3.1 Formulation

The SGS cloud scheme concerning radiation follows Slingo (1987) and is based on a relative humidity criterion, whereas the one affecting turbulence is a statistical cloud scheme (Sommeria and Deardorff, 1977). The schemes affect radiation and turbulence but do not produce precipitation.

Relative humidity cloud scheme

The SGS scheme for the radiation determines the SGS cloud fraction $R$ using:

$$R = \max \left[ 0, \min \left[ 1, \left( \frac{q_t}{q_{sat}} - \alpha_{sgs} \right) \left(1 - \alpha_{sgs}\right)^{-1} \right] \right]^2 \quad (C.4)$$

where $\alpha_{sgs}$ is the critical relative humidity, where SGS clouds start to form. It is determined using:

$$\alpha_{sgs} = 0.95 - 0.8\sigma(1 - \sigma) \left(1 + \sqrt{3(\sigma - 0.5)}\right) \quad (C.5)$$

where $\sigma = p/p_s$ is height, $q_t = q_v + q_l$ is the total water content,

$$q_{sat} = q_{sat,l}(1 - f_{ice}) + q_{sat,i}f_{ice}$$

the saturation mass fraction and $f_{ice}$ the fraction of ice:

$$f_{ice} = 1 - \min \left[ 1, \max \left(0, \frac{T_C - (-25^\circ C)}{-5^\circ C - (-25^\circ C)} \right) \right]$$

with $T_C$ the temperature in $^\circ C$. Figure C.14 illustrates the dependencies of $\alpha_{sgs}$ on $\sigma$ and of $R$ on $q_t$. It can be seen that for heights around 600 hpa cloud formation occurs already if roughly 75% percent of the saturation specific water content is reached.
Figure C.14: (a) Dependency of the critical relative humidity $\alpha_{sgs}$ on height $\sigma$ and (b) dependency of SGS cloud fraction $R$ on total water content $q_t$ for different heights $\sigma$ in the relative humidity SGS cloud scheme.

The in-cloud water respectively ice content is then determined assuming that 0.5 % of the saturation mass fraction is condensed water:

$$q_{c,sgs} = 0.005q_{sat}(1 - f_{ice})$$

$$q_{i,sgs} = 0.005q_{sat}f_{ice}$$

Statistical cloud scheme

The statistical cloud scheme for the turbulence follows the work of Sommeria and Durack (1977) and Mellor (1977) and is documented in Avgoustoglou et al. (2006).

Figure C.15: (a) Dependency of SGS cloud fraction $R$ on normalized saturation deficit $Q$ in the statistical scheme.
In the CCLM a Gaussian distribution is assumed for the saturation deficit $dq = q_{\text{sat}} - q_t$ with standard deviation $\sigma_{dqs}$. The fractional cloud fraction $R$ is then calculated utilizing:

$$R \approx \min \left[ 1, \max \left[ 0, clc_{\text{diag}} \left( 1 + \frac{Q}{q_{\text{crit}}} \right) \right] \right]$$

(C.6)

where $Q$ is the normalized saturation deficit

$$Q = \frac{dq}{\sigma_{dqs}}$$

(C.7)

The standard deviation $\sigma_{dqs}$ is determined in the turbulence routine. The run of $R$ is illustrated in Fig. C.15.

The tuning constants for the scheme follow the originally proposed values of Sommeria and Deardorff (1977):

- $clc_{\text{diag}}$ (cloud cover at saturation in statistical cloud diagnostic): 0.5
- $q_{\text{crit}}$ (critical value for normalized saturation deficit): 1.6

The liquid water content resulting from SGS clouds is then determined using:

$$ql = R \cdot \gamma \sigma_{dqs} \cdot \frac{(Q + q_{\text{crit}}) \cdot (Q + q_m)}{2 \cdot q_{\text{crit}}}$$

where $q_m = q_{\text{crit}} \cdot \left( \frac{1}{R} - 1 \right)$ and $\gamma = \frac{1}{1 + L_e/c_p q_{\text{sat}}}$.
C Model sensitivity studies

Shallow convection scheme
The shallow convection scheme is a Tiedtke massflux scheme (Tiedtke, 1989) where the part concerning the shallow convection has been extracted. The basic assumption of the scheme are (see Theunert and Seifert, 2006):

- Only temperature and moisture are affected by shallow convection, the momentum fluxes are neglected.
- Shallow convection is non-precipitating. Rain formation is neglected and the evaporation of rain below the cloud base is not considered.
- Convective downdrafts are neglected.
- Shallow convection is limited to a cloud depth of 250 hPa. For any clouds deeper than this threshold the scheme is switched off.

The scheme calculates therefore only separate tendencies on temperature and humidity but it is not coupled to the turbulence scheme. Theunert and Seifert (2006) documented that the shallow convection scheme was able to reduce the overprediction of clouds at the PBL top and to better represent the vertical moisture fluxes in the convective boundary layer.

In this study convective tendencies on temperature and humidity are updated every 2 minutes (≈ every 6th timestep). The mean entrainment rate for shallow convection is set to its default value of 3·10⁻⁴ m⁻¹ (cf. eq. 15 in Tiedtke, 1989). The simulation without the SGS cloud scheme is called “no_sgs”, the one using the shallow convection scheme and the shallow convection scheme “sh_conv”. The control simulation is using the sgs cloud scheme but not the shallow convection scheme. Furthermore, the reference atmospheric stability, a height-dependent relaxation, a soil moisture saturation of 60% and the explicit moisture correction set to .FALSE. are used.
C.3.2 Results

Cloud cover and precipitation for the simulations is shown in Fig. C.16, cloud fraction in Fig. C.17 and precipitation, net heating at the surface and 2 m temperatures in Fig. C.18. The SGS cloud scheme has a large influence on the simulations, especially in the morning hours. Additional radiative cooling resulting from SGS clouds occurs that leads to condensation and the formation of grid-scale clouds. These grid-scale clouds in turn affect incoming and outgoing radiation and lead to a reduction in the net radiation at the surface and 2 m temperatures. The decline of incoming radiation leads furthermore to a reduction of latent and sensible heat-fluxes into the atmosphere (not shown). Precipitation amounts are however smaller if the SGS scheme is switched off since the mid-level cloud cover produces precipitation resulting from the grid-scale clouds. The onset time of precipitation is the same in all simulations.

The effect of the shallow convection scheme on the other hand is of minor importance. There is hardly any difference between CTL and sh_conv visible. Fig. C.19 illustrates the difference in cloud water between CTL and sh_conv. There is an increase of mid-level clouds in the morning and in the early phase of cloud formation (1500 - 1700 UTC). In the evening the cloud amount is reduced using the shallow convection scheme. Precipitation amounts are nearly identical as compared to CTL.

It is worthwhile to take a look at the first few days of the simulations when diurnal equilibrium has not yet been established and the pre-moistening of the atmosphere from convection at previous days is absent. Fig. C.20 shows cloud water, cloud ice and surface rain rate for the first four days of the simulation. Neglecting the SGS cloud scheme results in considerably less mid-level clouds (not shown). The shallow convection scheme simulates moreover low- and mid-level clouds that are absent if the

![Figure C.16](image_url): Mean diurnal cycle of cloud water (kg kg\(^{-1}\), shaded area), cloud ice (kg kg\(^{-1}\), contour lines) and precipitation (mm h\(^{-1}\)) for (a) CTL, (b) no_sgs and (c) sh_conv.
scheme is omitted. This is most prominently visible at day 4 of the simulation. Thus the suspect is that the atmosphere is already sufficiently premoistened from convection on previous days in the state of diurnal equilibrium and the shallow convection scheme is therefore unnecessary.

To summarize, the SGS cloud scheme has a considerable influence on the diurnal cycle of convection by modifying the radiation transfer and thereby clouds. The formed additional clouds in turn affect both the radiation further and lead to additional precipitation. The shallow convection scheme shows negligible changes in the simulations performed in the diurnal equilibrium. Only for an atmosphere that requires shallow convection for a premoistening for deep convection to occur the shallow convection scheme modifies the picture. Such a situation is in our framework present during the first few days of the simulation when diurnal equilibrium is not reached.

![Figure C.17](image1.png)

**Figure C.17:** Cloudfraction for (a) CTL, (b) no_sgs and (c) sh_conv.

![Figure C.18](image2.png)

**Figure C.18:** (a) Domain mean precipitation (mm h\(^{-1}\), solid line) and the day-to-day variability of the domain mean precipitation (mm h\(^{-1}\), shade), (b) net heat at the surface (W m\(^{-2}\)) and (c) 2 m temperature (°C) for CTL (black line), no_sgs (blue line) and sh_conv (red line).
Figure C.19: Difference of specific cloud water (kg kg$^{-1}$, shade) between $sh_{\text{conv}}$ and CTL and the $10^{-6}$ kg kg$^{-1}$ contour line for CTL (black solid line) and $sh_{\text{conv}}$ (black dashed line).

Figure C.20: Domain mean cloud water (kg kg$^{-1}$, shaded area), cloud ice (kg kg$^{-1}$, contour lines) and precipitation (mm h$^{-1}$) for (a) CTL and (b) $sh_{\text{conv}}$ for the first four days of the simulation.
C.4 Microphysics, Ice parameterization

The sensitivity of the model to tuning constants in the parameterization for ice microphysics is assessed. In the first part, the autoconversion from ice to snow and the fall-speed of snow is modified. In the second part, the focus lies on the aggregation of ice onto graupel particles. The importance of tuning constants in the warm microphysics (e.g., the autoconversion from water vapor to cloud water) is not assessed here due to time constraints.

C.4.1 Snow

Two important parameters are the autoconversion by collection from ice to snow $c_{iau}$ and the terminal fall velocity of snow. Autoconversion is based on the formulation of Lin et al. (1983):

$$S_{iau}^i = e_i(T) \cdot \max [c_{iau}^i(q^i - q_0^i), 0]$$  \hspace{1cm} (C.8)

with the constant conversion rate $c_{iau}^i=10^{-3}$ s$^{-1}$, the autoconversion threshold value of $q_0^i=0$ g kg$^{-1}$ and the temperature-dependent sticking efficiency

$$e_i(T) = \max [0.2, \min[\exp(0.09(T - T_0)), 1.0]]$$  \hspace{1cm} (C.9)

with $T_0=273.15$ K. The autoconversion rate $c_{iau}^i$ is increased (decreased) by a factor of 1.5 (0.5) in a first sensitivity experiment called "zciau$1.5$" ("zciau$0.5$"). The autoconversion threshold $q_0^i$ was not varied in this study since ice content is already relatively high. The high values can be understood as the bulk microphysical scheme of the CCLM model does not include a direct sedimentation of ice crystals.

The constant $v0_s$ is a factor affecting the terminal fall velocity of snow and thereby the riming and aggregation. Its default value is 20 and it is increased respectively decreased by a factor of 1.5 or 0.5 in a set of sensitivity experiments. The simulations are called "zv0s$1.5$" and "zv0s$0.5$".

<table>
<thead>
<tr>
<th></th>
<th>$c_{iau}^i$</th>
<th>$v0_s$</th>
</tr>
</thead>
<tbody>
<tr>
<td>zciau</td>
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<td>20</td>
</tr>
<tr>
<td>zciau</td>
<td>$5 \cdot 10^{-3}$</td>
<td>20</td>
</tr>
<tr>
<td>zv0s</td>
<td>$10^{-3}$</td>
<td>30</td>
</tr>
<tr>
<td>zv0s</td>
<td>$10^{-3}$</td>
<td>10</td>
</tr>
</tbody>
</table>

Table C.8: Values for the parameters varied in the sensitivity experiments concerning cold microphysics.
Table C.8 summarizes the parameters used for the experiments. Fig. C.23 shows the mean diurnal cycle of clouds and falling hydrometeors. Ice clouds and snow are affected by the changes in both constants but graupel remains nearly unchanged. A decrease (an increase) of the autoconversion rate $c_{au}^d$ leads to an increase (a decrease) of ice clouds and snow. Surface precipitation amounts are decreased in both simulations but more strongly for a reduction of the autoconversion rate. Increasing the terminal fall velocity of snow decreases the amount of ice anvils and snow. Graupel is again unchanged.

The mean diurnal cycle of surface rain rate, surface net radiation and 2 m temperature for all simulations is shown in Fig. C.22. There is hardly any difference discernible between the simulations.
Figure C.21: Upper row: Mean diurnal cycle of specific water content (kg kg$^{-1}$, shaded area), specific ice content (kg kg$^{-1}$, contour line) and minimum and maximum values of domain mean precipitation over the 15 days of the simulation (dark grey shade). Lower row: mean diurnal cycle of rain (kg kg$^{-1}$, shaded area), snow (kg kg$^{-1}$, solid lines) and graupel (kg kg$^{-1}$, dashed lines).
C.4.2 Graupel

As seen in the previous section the impact of changing the conversion rate from ice to snow and the fall-speed of snow is of minor importance to the diurnal cycle of convection in our experiments. As seen in Fig. C.21 the formation of graupel is an important component of the microphysics scheme. To investigate the effect of the aggregation of cloud ice by graupel the constant $z_{\text{agg}}$, that influences the aggregation rate is changed. The rate of change of graupel due to aggregation of cloud ice is derived as follows:

\[
\left( \frac{\partial q_g}{\partial t} \right)_{\text{aggregation}} = e_i(T) \cdot q_i \cdot agg \cdot R
\]  \hspace{1cm} (C.10)

where $e_i(T)$ again is the temperature-dependent sticking rate, $q_i$ cloud ice and

\[
R = \exp(0.94878 \cdot \log(\rho \cdot q_g))
\]  \hspace{1cm} (C.11)

The default value for $agg$ is 2.46. It is increased by a factor of 1.5 and decreased by a factor 0.5 yielding the simulations “agg 1.5” and “agg 1.5”.

Clouds and precipitation for the experiments are shown in Fig. C.23. There are slight influences in the amount of snow, whereas graupel remains unchanged. Snow amounts increase if the aggregation rate is decreased and decrease if the rate is increased. Fig. C.24 shows the mean diurnal cycle of surface rain. It decreases slightly in both sensitivity experiments.

The set of sensitivity experiments demonstrated the relative unimportance of specific constants in the parameterization of ice microphysics. Compared to the effects that the parameterization of turbulence, mainly affecting the PBL, has on the diurnal cycle of convection the tuning of ice-microphysical constants are of minor importance.

Figure C.22: Mean diurnal cycle of (a) precipitation and minimum and maximum values of domain mean precipitation over the 15 days of the simulation (mm h$^{-1}$), (b) net heat (W m$^{-2}$) and (c) 2 m temperatures ($^\circ C$) for CTL (black line), zciau (blue lines) and zv0s (red lines).
### Figure C.23: Upper row: Mean diurnal cycle of specific water content (kg kg\(^{-1}\), shaded area), specific ice content (kg kg\(^{-1}\), contour line) and minimum and maximum values of domain mean surface precipitation over the 15 days of the simulation (dark grey shade). Lower row: mean diurnal cycle of rain (kg kg\(^{-1}\), shaded area), snow (kg kg\(^{-1}\), solid lines) and graupel (kg kg\(^{-1}\), dashed lines).

### Figure C.24: Mean diurnal cycle of precipitation (solid line) and minimum and maximum values of domain mean precipitation (shaded area) over the 15 days of the simulation (mm h\(^{-1}\)) for CTL (black line), agg 0.5 (blue line) and agg 1.5 (red line).
C.5 Time step

From a numerical point of view the model should ideally be insensitive to decreasing the time step $\Delta t$ below a certain value $\Delta t_{crit}$. In order to test this convergence, four different simulations with $\Delta t = 30$, 20, 15 and 10 s are carried out. The coefficients $\alpha_i$ for horizontal diffusion are adapted such that a constant diffusion rate is applied (see Weisman et al., 1997): $\frac{\alpha_i}{\Delta t} = const$. The setup of the simulation is as follows: the atmospheric profile used is the CTL profile $(dT/dz=-0.7\, K/100\, m, RH=40, 70\%)$, a height-dependent relaxation is used, soil saturation is set to 60 % and vegetation is used. The explicit moisture correction is furthermore set to .FALSE. The simulation using a time step of 20 s is identical to the 60_CTL simulation shown in chapter 3 of this thesis.

Fig. C.25 shows the mean diurnal cycle of precipitation, 2 m temperature and 2 m dew-point depression. Small differences are observable. Precipitation amounts decrease to some extent for a smaller time step and an earlier onset of precipitation of 0.5 h in the 30 s simulation is visible. 2 m temperature is nearly identical in all simulations and near-surface humidity is unimportantly larger for smaller time steps.

Cloud water and cloud ice are shown in Fig. C.26. The overall picture is equal for the simulations. The mid-level cloud cover increases while decreasing the time step.

As some differences exist in the timing of precipitation going from 30 s to 20 s, but differences between 20 s, 15 s and 10 s are small, we decided to use a time step of 20 s for the simulations at 2.2 km in all studies.

**Figure C.25:** (a) Mean diurnal cycle (mm h$^{-1}$, solid lines) and day-to-day variability of the domain mean precipitation (mm h$^{-1}$, shaded area). Mean diurnal cycle of (b) 2 m temperature and (c) 2 m dew-point depression (°C) for 30 s (black), 20 s (blue), 15 s (red) and 10 s (green).
Figure C.26: Mean diurnal cycle of specific cloud water (kg kg$^{-1}$, shaded area), specific ice content (kg kg$^{-1}$, contour line) and minimum and maximum values of domain mean precipitation over the 15 days of the simulation (dark grey shade) for a time step of (a) 30 s, (b) 20 s and (c) 10 s.
C.6 Horizontal resolution, convergence

Together with a decrease of the grid-spacing $\Delta x$ of the model, surface fields and topography are represented more realistically on the one hand and on the other hand, the model can simulate processes as e.g. convection and turbulence in more detail. For the framework used in this thesis, only the latter point is of importance as we employ a homogeneous lower boundary condition. We thus expect the model to resolve more fine-scale structures and a better resemblance with reality increasing the resolution.

For grid-spacings smaller than a threshold value $\Delta x_{\text{conv}}$ the simulated structures as for instance convective cells should not differ between simulations, as the model has converged. This will generally be achieved if the horizontal resolution was no coarser than one quarter of the sub-cloud layer depth or less (Petch et al., 2002). Recent studies show that a grid-spacing of $\Delta x = 100$ m (Bryan et al., 2003) or $\Delta x = 200$ m (Petch, 2006) is needed to resolve clouds and to produce sensible cloud sizes. The mentioned studies however used a 3D version of the Smagorinsky-Lilly model (Smagorinsky, 1963, Lilly, 1967), which is a typical LES model, for the parameterization of turbulence. Craig and Dörrbrack (2008) stated that for large eddy simulations (LES) of clouds, the smaller of the length scale for buoyancy or the cloud size should be resolved, which is typically in the order of a few hundred meters. In our experiment we use a one-dimensional TKE closure at level 2.5 (Mellor and Yamada, 1974, Raschendorfer, 2001). There are few studies on how the numerical convergence using such a turbulence model looks like. For the CCLM it was found that the bulk total net heating is nearly independent of the grid spacing. This resulted however from a compensation of the diabatic and adiabatic processes at different resolutions (Wolfgang Langhans, personal communication).

Using a grid-spacing $\Delta x$ of 2.2 km we thus don’t expect to accurately predict some details of convection but we are nevertheless able to reproduce the basic structures of convection (Bryan et al., 2003). We thus also limit our requirement for convergence to occur and demand that integrated quantities such as area-averages or averages over time remain constant going to higher resolution.

The numerical convergence of our model is tested decreasing the grid-spacing from the default value of 0.02° (2.2 km) to 0.01° (1.1 km) and 0.005° (550 m). The domain size is kept constant at 220 km resulting in 100 x 100, 200 x 200 and 400 x 400 grid points, respectively. In addition, the time step is reduced from 20 s to 10 and 5 s. Coefficients $\alpha_i$ for horizontal diffusion of moisture and horizontal winds are adapted such that the diffusion rate is constant (see Weisman et al., 1997) $\frac{\alpha_i}{\Delta t} = \text{const}$. As it is computationally very expensive to calculate domain mean values for the calculation of the relaxation
tendencies (cf. eq. 2.1) we updated the coefficients only every 5 minutes in the 550 m simulation. This change has little impact on the diurnal cycle of convection in the CTL simulation (not shown). The simulations are performed with the CTL profile for the atmosphere, a height-dependent relaxation, a soil moisture of 60% including vegetation and without the explicit moisture correction. It is identical to the CTL simulation in chapter 2 and the 60_CTL simulation in chapter 3.

Table C.9 summarizes modeled integrated quantities. Net surface radiation (shortwave plus longwave) is constant across different resolutions. The partitioning between the fluxes is slightly different, resulting in less sensible heating for decreased grid-spacing. Both up- and downward integrated convective mass-fluxes are larger for an increased resolution and resulting precipitation amounts are smaller for higher resolutions. Fig. C.27 shows the mean diurnal cycle of clouds and surface precipitation. Using a higher resolutions the phase, where clouds are present is extended, but peak cloud amounts are reduced. This is true for both water and ice clouds. The mid-level cloud cover in the morning is reduced using a decreased grid spacing. The mean diurnal cycle of upward convective mass-fluxes is shown in Fig. C.28. A longer phase of convective activity for higher resolutions going together with reduced peak intensities can be seen. Fig. C.29a shows the mean diurnal cycle of precipitation. Consistent with the picture in cloud cover and mass-fluxes, precipitation occurrence is prolonged, whereas domain mean peak precipitation rates are smaller at higher resolutions. The day-to-day variability of the domain mean precipitation is moreover increased for decreased grid spacing. Albeit the earlier onset of precipitation at 1.1 km and 550 m, the integrated amount is reduced by 8 and 6% respectively. Fig. C.29b shows the distribution of simulated precipitation intensities normalized by the number of rainy points (≥0.5 mm h⁻¹). Precipitation intensities are highest for 2.2 km grid spacing. This is in accordance with the more intense phase of cloud cover and convective mass-fluxes.

<table>
<thead>
<tr>
<th></th>
<th>precipitation [mm h⁻¹]</th>
<th>cmf up [kg m⁻¹ s⁻¹]</th>
<th>cmf down [kg m⁻¹ s⁻¹]</th>
<th>sensible [W m⁻²]</th>
<th>latent [W m⁻²]</th>
<th>( R_{\text{net}} ) [W m⁻²]</th>
</tr>
</thead>
<tbody>
<tr>
<td>2.2 km</td>
<td>0.140</td>
<td>101</td>
<td>39.2</td>
<td>29.6</td>
<td>129</td>
<td>162</td>
</tr>
<tr>
<td>1.1 km</td>
<td>0.129</td>
<td>115</td>
<td>57.8</td>
<td>27.5</td>
<td>130</td>
<td>162</td>
</tr>
<tr>
<td>550 m</td>
<td>0.131</td>
<td>121</td>
<td>89.1</td>
<td>27.1</td>
<td>130</td>
<td>162</td>
</tr>
</tbody>
</table>

Table C.9: Mean values of precipitation, vertically integrated up- and downward convective mass fluxes (cmf), sensible and latent heat fluxes and net surface radiation averaged over day 16-30 of the simulations and spatially over the domain.
Surprising to see is, that the simulation at 550 m grid spacing does not possess more shallow convection as could be expected. This arises probably due to the used parameterization scheme for turbulence that begins to get insufficient at such resolutions. As seen in section C.3 the use of the shallow convection increased shallow clouds neither. The result from the higher resolution again indicates, that the atmosphere probably is already sufficient moist from previous convection and shallow convection is unrequired. 2 m temperature and dew-point depression are shown in Fig. C.30. The diurnal cycle of 2 m temperatures is nearly identical going to higher resolution, whereas near-surface humidity is slightly decreased for 550 m grid spacing.
C Model sensitivity studies

**Figure C.29:** (a) Mean diurnal cycle (mm h\(^{-1}\), solid lines) and day-to-day variability of the domain mean precipitation (mm h\(^{-1}\), shaded area) and (b) logarithmic histogram of hourly precipitation sums normalized by the number of rainy points collected at each grid point of the domain over the time period of day 16-30. Bins are 0.5 mm h\(^{-1}\) wide. Simulations shown are 2.2 km (black), 1.1 km (blue) and 550 m (red).

**Figure C.30:** Mean diurnal cycle of (a) 2 m temperature (°C) and (b) dew-point depression (°C) for 2.2 km (black), 1.1 km (blue) and 550 m (red).

Fig. C.31 shows spectra of vertical velocity within the PBL and above the PBL. Spectra are computed at each hour of the day calculating 1-dimensional spectral decompositions of \(w\) along the east-west direction and averaging spectra across the north-south direction of the grid and in time over the 15 evaluation days (Skamarock, 2004). All computed spectra show a drop at wavelengths around 7 \(\Delta x\), the effective resolution of a numerical model. It is evident from Fig. C.31 that a decrease of the grid spacing yields more fine-scale structures in our model. Striking is the relatively flat spectra for
large wavelengths. This results presumably from the lack of large-scale disturbances in our idealized setup. Energy is only generated at small scales and transferred to larger scales.

At 1300 UTC the 550 m simulations shows considerably more energy at all scales. This is consistent with the earlier formation of convection and clouds. Energy within the PBL is furthermore larger than above it. Within the PBL the 1.1 km simulation holds more energy than the 2.2 km simulation, again consistent with the earlier onset of convection. At 1800 UTC, the time of peak precipitation amounts, energy within and above the PBL are comparable as deep convection takes place. Only at the smallest wavelengths there is more energy in the free atmosphere. Slightly more energy is contained in the 2.2 km simulation, where most intense convection is observed.

Considering the convergence of the model we see firstly, that the mean diurnal cycle becomes smoother using higher resolutions. Secondly, the integrated mean quantities are hardly changed, which is what we demanded.
C.7 Vertical resolution

A peculiar feature of all simulations performed is the very short, nearly absent phase of shallow convection. Observations show that in the morning hours, before deep convection sets in there are some hours where shallow convection is prevalent. Cloud bases of shallow cumulus clouds coincide with the LCL, their tops grow continuously until at some point they exceed the LFC (Chaboureau et al., 2004). The most important role of shallow convection is to gradually moisten the environment until at some point the atmosphere is sufficiently humid that an ascending parcel is no longer completely diluted by dry-air entrainment but can ascent to the height of neutral buoyancy. This process is called the transition from shallow to deep convection (Siebesma, 1998, Stevens, 2005, Wu et al., 2009). Shallow convection is largely non-precipitating, some precipitation can however occur, which helps to produce cold pools and organized structures. These in their term help to promote the formation of larger cloud sizes that experience less lateral entrainment and can evolve into deep convection (Kuang and Bretherton, 2006, Khairoutdinov and Randall, 2006).

One hypothesis, why shallow convection is largely absent in our model is, that the vertical resolution of the grid is not sufficient to capture the process. In the following, we investigate if an increase of the number of vertical levels in the model can improve the representation of shallow convection and if so, what the influence of the shallow con-

![Figure C.32: (a) height difference (m) between two adjacent main levels and (b) height (m) of the main levels for the old manual 50 layers of COSMO (black line), the cosh formulation of the COSMO (green line), the new formulation following ARPS for 70 level (blue line) and for 100 layers (red line).]
convection on the mean quantities is? To this aim the following simulations are performed: **50 layer**: the control simulation, 50 vertical levels using prescribed values for the construction of the vertical coordinate $v_{\text{coord}}$. 60% soil moisture and the intermediate atmospheric stability (-0.7 K (100 m)$^{-1}$) are used.

**70 layer arccos**: 70 vertical layers using the arccos formula specified in the COSMO model to construct the levels.

**70 layer tanh**: 70 vertical layers using the tanh formula specified in the Advanced Regional Prediction System (ARPS) to construct the levels (ARPS, Version 4.0, chapter 7.3.2.).

**100 layer**: 100 vertical layers using the tanh formula specified in the ARPS model to construct the levels.

The explicit moisture correction (lexcor=.TRUE., cf. subsection C.1) is switched off.

In the default setup the height of the model layers is specified manually. If coordinates are not specified manually an arccos formula is used in the CCLM to specify the spacing between adjacent levels:

$$v_{\text{coord}}(k) = z_{z_{\text{top}}} \frac{2}{\pi} \cdot \arccos \left( \frac{k - 1}{ke} \right)^{\alpha} \quad \text{(C.12)}$$

where $z_{z_{\text{top}}}$ is set to 22 000 m, and $\alpha=3.0$. The spacing in the lower part of the atmosphere is even coarser than in the original 50 layer simulation in this simulation, where the levels are explicitly prescribed.

In a further approach the level spacing is constructed as described in the ARPS model using a tanh formula to calculate the height difference between the model levels:

$$dz(i) = dz_m + \frac{dz_{\text{min}} - dz_m}{\tanh(2\alpha)} \tanh \left[ \frac{2\alpha}{1 - a} (i - a) \right] \quad \text{for } i = 1, n_2 \quad \text{(C.13)}$$

where $n_2=70$ is the number of levels, $dz_{\text{min}}=15$ m corresponds to the grid spacing of the lowermost level, $dz_m=314.5$ m, $\alpha=1.0$ and $a = (1 + n_2)/2$. A further simulation with this setup for the vertical coordinate using $n_2=100$ levels, $dz_{\text{min}}=10$ m and $dz_m=220$ m called “100 layer” is performed. The difference between the levels and the resulting height of the levels for the four simulations is shown in Fig. C.32. For the simulation using 100 layers the time step needs to be reduced to 10 s due to arising numerical instabilities caused by the fine vertical grid. Using the results from Appendix C.5 this should have a small impact on the overall picture of the diurnal cycle of convection.
The mean diurnal cycle of clouds and precipitation is shown in Fig. C.33. There is no significant difference between the simulations using 50 levels and those using more levels. Slightly less mid-level clouds are simulated in both 70-layer simulations, whereas in the 100-layer simulation the mid-level cloud cover between 600 and 700 hPa is increased. Fig. C.34 shows the diurnal cycles of precipitation. The onset and peak of precipitation is slightly advanced for simulations using more vertical levels. Precipitation amounts are nearly unchanged. Fig. C.35 shows the mean diurnal cycle of 2 m temperature and dew-point depression for the set of simulations. Near-surface temperatures and dew-point depression decrease slightly for 70 layer tanh 100 layers but increase for 70 layer arccos.
C.7 Vertical resolution

Figure C.34: Mean diurnal cycle of surface rain rate \( \text{mm h}^{-1} \). The solid lines indicate the mean diurnal cycle, the shaded area indicates the day-to-day variability.

Figure C.35: Mean diurnal cycle of (a) 2 m temperature \( \degree \text{C} \) and (b) 2 m dew-point depression for 50 layer (black line), 70 layer arccos (green line), 70 layer tanh (blue line) and 100 layer (red line).

Zooming in on details, hardly any differences in the development of clouds between the simulations are visible. Fig. C.36 shows the cloud fraction for the boundary-layer top region. Starting from 1300 UTC shallow clouds are simulated in all simulations.

Vertical profiles for \( \theta \) for 0000 and 1200 UTC from the surface up to 650 hPa are shown in Fig. C.37. The profile is very similar for all simulations.

Vertical profiles of the relative humidity for 0000 and 1200 UTC are shown in Fig. C.38. Domain mean values as well as spatial and temporal variations are nearly identical in all simulations.

Summing up, an increase of the number of vertical levels in the atmosphere should enable the model to simulate more structures. Looking at the mean diurnal cycle of clouds, precipitation and temperature, there is hardly any change visible between the simulations.
Figure C.36: Mean diurnal cycle of cloud fraction for (a) 50 layer, (b) 70 layer arccos, (c) 70 layer tanh and (d) 100 layer.

Figure C.37: Vertical profile of domain mean $\theta$ (K) at (a) 0000 UTC and (b) 1200 UTC for 50 layer (black line), 70 layer arccos (green line), 70 layer tanh (blue line) and 100 layer tanh (red line).
Figure C.38: Vertical profile of RH (%) at (a) 0000 UTC and (b) 1200 UTC for 50 layer (black line), 70 layer arccos (green line), 70 layer tanh (blue line) and 100 layer tanh (red line). Solid lines indicate domain mean values, thin lines indicate the 10th and 90th percentile, respectively of values simulated over all points and evaluation days.
C Model sensitivity studies

C.8 Horizontal diffusion, advection scheme

The aim of this section is to investigate the sensitivity of the model to numerical issues, namely the advection scheme for moist quantities, and the strength of the numerical diffusion on the horizontal wind.

The CCLM offers different advection schemes for moist quantities, whereof an Eulerian second-order Bott scheme (Bott, 1989) and a semi-lagrangian (SL) scheme are used in the current study. Details on the implementation of the schemes can be found in Förstner et al. (2006). The Bott scheme has the advantage of being positive-definitive, meaning that the advection does not produce negative undershoots. It has the disadvantage of being a one-dimensional scheme, implying that directional splitting is necessary to receive a three-dimensional advection scheme. The Bott scheme shows furthermore stronger diffusion on the advected quantities than a SL scheme. A SL scheme determines in a first step the originating point of the trajectory for the current time step and in a second step determines the concentration of the tracer at that starting point by interpolation from the neighboring grid points (see e.g. Staniforth and Côté, 1991). In the CCLM a tri-cubic interpolation is used. The SL scheme holds the benefit of being a three-dimensional scheme and of being numerically stable (Courant numbers larger than 1 are possible), whereas it is not positive-definite. A multiplicative filling procedure is used to redistribute negative undershoots.

To avoid the built-up of numerical noise at the grid scale numerical diffusion is performed on the horizontal wind and specific humidity. No diffusion is done on temperature. The horizontal diffusion is a 4th-order filter:

\[ \frac{\partial \phi}{\partial t} = -K_\phi \nabla^4 \phi \]

where \( \phi \) is the variable filtered and

\[ K_\phi = \alpha_\phi \frac{(\Delta x)^4}{\Delta t} \]

where \( \alpha_\phi \) is the dimensionless filter coefficient. For moisture \( K_{qv} = 0.33 \cdot \frac{(\Delta x)^4}{20 \text{s}} \) in all simulations performed. For the current sensitivity study the smoothing on horizontal wind is varied with \( \alpha_u=0.15 \text{s}^{-1} \) and \( \alpha_u=0.5 \text{s}^{-1} \) (the default value).
Simulations are termed:

**Bott, 0.15** Bott advection scheme \( \alpha_u=0.15 \text{ s}^{-1} \)

**Bott, 0.5** Bott advection scheme \( \alpha_u=0.5 \text{ s}^{-1} \)

**SL, 0.15** Semi-lagrangian advection scheme \( \alpha_u=0.15 \text{ s}^{-1} \)

**SL, 0.5** Semi-lagrangian advection scheme \( \alpha_u=0.5 \text{ s}^{-1} \)

---

**Figure C.39:** Mean diurnal cycle of (a) precipitation (solid lines, mm h\(^{-1}\)) and minimum and maximum value of domain mean value simulated over the 15 evaluation days (shade), (b) 2 m temperature (°C) and (c) 2 m dew-point difference.

The mean diurnal cycle of precipitation is shown in Fig. C.39a. Striking are the considerably larger precipitation amounts if the SL scheme is used. Precipitation amounts are furthermore larger if stronger computational diffusion on wind is used with a larger difference for the SL scheme.

The mean diurnal cycle of 2 m temperature and dew-point difference is shown in Fig. C.39b and c. There is hardly any difference in these two quantities visible. Only a small increase in night-time near-surface humidity for the SL scheme is visible.

Looking at the diurnal cycle of clouds shown in Fig. C.40 there is no significant qualitative change. The mid-level cloud cover is decreased if stronger diffusion on the horizontal wind is applied and slightly more clouds exist if the SL is used.

Fig. C.41 shows horizontal cross-section of the zonal wind at 1630 UTC, the time of maximum convective activity, at around 660 hPa for the set of simulations. In all of the plots some “stripes” are visible which are an indicator for numerical instability. The stripes are considerably reduced if the stronger diffusion is utilized, but still remain however. There is no directly visible major performance of the SL or the Bott scheme.
For the simulations performed we chose the Bott advection scheme because of its positive-definiteness combined with a diffusion constant of 0.5 for the horizontal wind, as the stripes could considerably be reduced.

Figure C.40: Mean diurnal cycle of cloud water (shade, kg kg\(^{-1}\)), cloud ice (contour lines, kg kg\(^{-1}\)) and surface rain rate (mm h\(^{-1}\), black solid line and gray shaded area indicating day-to-day variability) for (a) Bott, 0.15, (b) Bott, 0.5, (c) SL, 0.15 and (d) SL, 0.5. Mean diurnal precipitation amounts (mm day\(^{-1}\)) are given by the numbers in the lower left corner.
Figure C.41: Zonal wind (m s\(^{-1}\)) at layer 29 (ca 660 hPa) at 1630 UTC for (a) Bott, 0.15, (b) Bott, 0.5, (c) SL, 0.15 and (d) SL, 0.5.
C.9 Summary

Overall the model shows a reasonable performance across changes of different parameters. The numerics are robust among changes of the time step and the setup of the computational grid. Looking at integrated quantities, the model shows a sensible convergence. An increase of the number of vertical layers results in more clouds, but the influence on resulting precipitation is of minor importance.

Concerning physical parameterizations, the parameters affecting boundary-layer processes have a considerably larger influence on the resulting diurnal cycle of moist convection than parameters affecting for example ice microphysics. The use of a subgrid-scale cloud scheme furthermore proofed to have strong influences on the surface radiation. Generally, none of the parameters considered changed the overall character of the simulation completely. Only the switch lexpcor, which could be identified as a model bug affects the simulations severely.
Appendix D

The CCLM model

For this thesis the COSMO-CLM (CCLM) model version 4.0 is utilized (see http://www.clim-community.eu/). A general description of the model can be found in Böhm et al. (2006) or chapter 2. The focus of this chapter lies on specific issues related to the model that are not discussed in previous sections.

D.1 TERRA ML

D.1.1 Evapotranspiration

Fundamental for the interaction between the land surface and the atmosphere is the evapotranspiration of water. In contrast to first-generation land surface schemes (bucket models Manabe, 1969), The TERRA ML is a second-generation land-surface scheme. First generation schemes include a single soil layer with a constant water-holding capacity. Evaporation is limited by the availability of soil water, and any excess of soil water over the water-holding capacity results in runoff. Second-generation schemes include multiple soil layers with vertical transport between the layers and take into account the interaction of the soil and vegetation with the atmosphere, rather than being passive (as in the first-generation models) (see Pitman, 2003). Second-generation models furthermore differentiate between soil and vegetation at the surface. Third generation models include even more sophisticated vegetation, where the interaction of the leaves with carbon and nitrogen is represented in determining the canopy conductance (e.g. Sellers et al., 1992).

In TERRA_ML four different sources for evapotranspiration are considered: evaporation from the interception reservoir $W_i$, evaporation from the snow reservoir $E_{snow}$, bare soil evaporation from the uppermost soil layer $E_b$ and transpiration by plants $T_r$.

An bulk aerodynamic formulation for the calculation of potential evaporation $E_{pot}$ is used:

$$E_{pot}(T_{sfc}) = \rho C_d q_s \left[ Q_v(T_{sfc}) - q_v \right]$$  \hspace{1cm} (D.1)

where $T_{sfc}$ is the skin temperature, $\rho$ the density at the lowest atmospheric level, $C_d$ is the bulk-aerodynamical coefficient for turbulent moisture transfer at the surface, $q_v$ the specific humidity at the lowest atmospheric layer and $Q_v(T_{sfc})$ the saturation specific humidity at the surface. $E_{pot}(T_{sfc}) > 0$ indicates upward directed potential evaporation at soil surface temperature $T_{sfc}$ and forms the starting point for all further calculations.

Evaporation from the interception and the snow reservoir

A partial coverage of the soil surface by interception water or by snow is taken into account in the parameterization of evaporation from the interception layer. The partial
coverage by water $f_i$ is determined using:

$$f_i = \max[0.01; 1 - e^{\max(-5.0; -W_i/\delta_i)}]$$  \hspace{1cm} (D.2)

where $W_i$ is the water content of the interception store and $\delta_i$ is set to 0.001 m. The partial coverage by snow $f_{snow}$ is calculated using:

$$f_{snow} = \max[0.01; \min(1.0; W_{snow}/\delta_s)]$$  \hspace{1cm} (D.3)

where $W_{snow}$ is the snow water equivalent and $\delta_s=0.015$ m.

Evaporation of interception water $E_i$ occurs if $E_{pot}(T_{sfc}) > 0$ and $W_i > 0$ at places covered with interception water. The evaporation is limited by the availability of interception water:

$$E_i = \max[-\rho_w \Delta t W_i / f_i E_{pot}(T_{sfc})]$$  \hspace{1cm} (D.4)

The rate of evaporation from snow is calculated accordingly:

$$E_{snow} = \max[-\rho_w \Delta t W_{snow} / f_{snow} E_{pot}(T_{sfc})]$$  \hspace{1cm} (D.5)

Formation of dew is simulated if $E_{pot}(T_{sfc}) < 0$ and the temperature at the surface is above the freezing point $T_0$ following:

$$E_i = E_{pot}(T_{sfc})$$  \hspace{1cm} (D.6)

for $T < T_0$ rime is formed using:

$$E_{snow} = E_{pot}(T_{sfc})$$  \hspace{1cm} (D.7)

Bare soil evaporation

For all soil types but rock and ice, evaporation from the bare soil $E_b$ is calculated if $E_{pot}(T_{sfc}) > 0$ using the Biosphere-Atmosphere Transfer Scheme (BATS) formulation after Dickinson (1984). Evaporation is on the one hand limited by the potential evaporation (atmospheric demand) and on the other hand by the maximum moisture flux through the surface $F_m$ that the soil can sustain (the maximum supply) and occurs at whichever rate is the smallest (Desborough et al., 1996). Bare soil evaporation occurs in areas that are not covered by interception water, snow or vegetation:

$$E_b = (1 - f_i) (1 - f_{snow}) (1 - f_{plnt}) \min[E_{pot}(T_{sfc}); F_m]$$  \hspace{1cm} (D.8)

where $f_{plnt}$ is the fractional area covered by plants. The parameter $F_m$ stems from tuning a two-layer soil model that is based on average values of soil water content normalized by the volume of voids. Thus the soil type used for the study has a large influence on the bare soil evaporation through this term.
Figure D.1: Influence on the stomatal resistance of (a) PAR, (b) soil water and (c) temperature. In (b) the bold lines show the dependency for the soil types sand (red line), loam (black line) and clay (blue line). The thin dashed line indicates the field capacity, the thin solid line the turgor loss point and the dashed line the permanent wilting point for the three different soil types. The turgor loss point for this graph was calculated using $E_{pot}(T_{sfc})$ from the 60_CTL simulation averaged over the domain and the diurnal cycle (12.98 mm day$^{-1}$).

**Plant transpiration**

Plants exert a biophysical control on transpiration through their stomata. The formulation for the calculation of plant transpiration $T_r$ is based on Dickinson (1984):

$$T_r = f_{plant} \cdot (1 - f_i) \cdot (1 - f_{snow}) \cdot E_{pot}(T_{sfc}) \frac{r_a}{r_a + r_f} \quad (D.9)$$
where \( r_a \) is the atmospheric resistance given by \( r_a^{-1} = C_d |v_k| \) and \( r_f \) the foliage resistance (the resistance for water vapor transport from the foliage to the canopy air). For the foliage resistance it is assumed that the stomatal resistance \( r_s \) and the atmospheric resistance \( r_{la} \) are acting in parallel as
\[
\frac{1}{r_f^{-1}} = \frac{1}{LAI \cdot r_{la}^{-1} + r_s^{-1}}
\]
with \( r_{la}^{-1} = C' u_*^{1/2} \), \( C' = 0.05 \) and \( u_* = (u^2 + v^2) C_d q \). To arrive at the foliage resistance from the resistance of the leaf a scaling by the leaf area index \( LAI \) is applied.

The biophysical control of transpiration by plants is accounted for in the determination of the stomatal resistance using the multiplicative approach by Jarvis (1976):
\[
r_s^{-1} = r_{max}^{-1} + \left( r_{min}^{-1} - r_{max}^{-1} \right) \left[ F_{rad} F_{wat} F_{tem} F_{hum} \right]
\]
where \( r_{min} = 150 \text{ s m}^{-1} \) and \( r_{max} = 4000 \text{ s m}^{-1} \) are tuning constants and the functions \( F \) describe the dependence of \( r_s \) on radiation, soil water content, ambient temperature and ambient specific humidity, respectively. These functions vary between 0 for unfavourable conditions for transpiration and 1 for optimal conditions and are shown in Fig. D.1.

The radiation function is determined as follows:
\[
F_{rad} = \min \left( 1, \frac{PAR}{PAR_{crit}} \right)
\]
where \( PAR \) is the photosynthetically active radiation and \( PAR_{crit} = 100 \text{ W m}^2 \) is a tuning parameter.

The influence of soil water is strongly dependent on the permanent wilting point \( w_{PWP} \), a parameter that is dependent on the soil type, and the turgor loss point of plants \( w_{TLP} \) (the point, below which the pressure potential in the plant cell’s becomes too low to support the turgidity of leaf cells):
\[
F_{wat} = \max \left[ 0, \min \left( 1, \frac{w_{l,root} - w_{PWP}}{w_{TLP} - w_{PWP}} \right) \right]
\]
where \( w_{l,root} \) is the liquid water content fraction of the soil averaged over the root depth.

The turgor loss point is parameterized after Denmead and Shaw (1962):
\[
w_{TLP} = w_{PWP} + \left( w_{FC} - w_{PWP} \right) \cdot (0.81 + 0.121 \arctan(E_{pot}(T_{sfc}) - E_{pot,norm}))
\]
where \( E_{pot,norm} = 4.75 \text{ mm day}^{-1} \). In the 60_CTL simulation \( E_{pot}(T_{sfc}) \approx 13 \text{ mm day}^{-1} \) averaged over the diurnal cycle which relatively far away from \( E_{pot,norm} \). The turgor loss point of plants is therefore close to the field capacity.
The temperature-dependency is parameterized assuming a quadratic function:

\[ F_{\text{tem}} = \max \left[ 0, \min \left( 1, 4 \frac{(T_{2m} - T_0)(T_{\text{end}} - T_{2m})}{(T_{\text{end}} - T_0)^2} \right) \right] \quad (D.14) \]

where \( T_{\text{end}} = 40^\circ \text{C} \). Optimal conditions are thus met at a temperature of \( 20^\circ \text{C} \), and both an increase and a decrease of temperature lead to an equal reduction of transpiration. At a temperature of \( 0^\circ \text{C} \) respectively \( 40^\circ \text{C} \) transpiration ceases.

The dependency of atmospheric humidity is not considered and \( F_{\text{hum}} = 1 \) for all atmospheric humidities.

### D.1.2 Vertical transport of soil water and temperature

TERRA\_ML considers only vertical, but no lateral transport of water. The formulation of the vertical transport of water and heat is summarized in the following.

#### Water transport

The transport of water is calculated using the Richardson equation:

\[ \frac{\partial}{\partial t} w_l = \frac{1}{\rho_w} \frac{\partial F}{\partial z} \quad (D.15) \]

where \( w_l = \frac{W_l}{\Delta z} \) is the liquid water fraction defined by the liquid water content \( W_l \) (m \( \text{H}_2\text{O} \)) in the layer thickness \( \Delta z \). \( \rho_w \) is the density of water and \( F \) is the soil water flux which is calculated using:

\[ F = -\rho_w \left[ -D_w(w_l) \frac{\partial w_l}{\partial z} + K_w(w_l) \right] \quad (D.16) \]

where \( D_w \) is the hydraulic diffusivity and \( K_w \) the hydraulic conductivity. \( D_w \) and \( K_w \) are calculated using the exponential formulation of Rijtema (1969):

\[ D_w(w_l) = D_0 \exp \left[ D_1(w_{PV} - \bar{w})/(w_{PV} - w_{ADP}) \right] \quad (D.17) \]

and

\[ K_w(w_l) = K_0 \exp \left[ K_1(w_{PV} - \bar{w})/(w_{PV} - w_{ADP}) \right] \quad (D.18) \]

where \( D_0, D_1, K_0 \) and \( K_1 \) are specified parameters that depend on the soiltype used. \( w_{PV} \) is the volume of voids and \( w_{ADP} \) is the air dryness point; these parameters are also specified depending on the soil type. \( \bar{w} \) is the weighted mean of the liquid water content on half levels.

At the lowest hydrologically active layer \( D_w \) is set to zero, meaning that only downward gravitational flux is considered. Ground water cannot moisten the soil by capillary rise from below. At the surface the soil water flux \( F \) in equation D.16 is replaced by the
infiltration.

Runoff from any soil layer $k$ occurs if the total water content $w_k$ of the layer exceeds the field capacity $w_{FC}$ and if the divergence of the soil water fluxes is negative. Runoff $R_k$ is then determined using:

$$R_k = \frac{w_k - w_{FC}}{w_{PV} - w_{FC}} \left( \frac{\partial F}{\partial z} \right)_k \Delta z_k$$  \hspace{1cm} (D.19)

**Heat transport**

The heat conduction equation is used for the prediction of soil temperature $T_{so}$:

$$\frac{\partial}{\partial t} T_{so} = \frac{1}{\rho c w} \frac{\partial}{\partial z} \left( \lambda \frac{\partial T_{so}}{\partial z} \right)$$ \hspace{1cm} (D.20)

where $c$ is the heat capacity and $\lambda$ the heat conductivity. The lower boundary is provided by a climatological temperature ($8.5^\circ$C in our simulations) that is constant over the simulation. At the uppermost layer the soil is coupled to the atmosphere. The heat conductivity $\lambda$ is dependent on the soil water content. In TERRA\_ML, the heat conductivity is based on a soil water content which is equal to the average between wilting point and field capacity.

**D.2 Aerosols, carbon dioxide concentrations and ozone**

Constant $CO_2$ concentrations of 360 ppm are assumed for all simulations. Greenhouse gases apart from $CO_2$ are not considered. Aerosols are specific using the Tanré et al. (1984) climatology with values valid at Munich. The vertically integrated ozone content is set to 0.06 Pa ozone and the height of maximum ozone occurs at 42 hPa.

**D.3 Periodic boundary conditions**

Periodic or cyclic boundary conditions are a set of boundary conditions that are typically employed if the domain considered is representative of a much larger horizontal area. Any object passing through one face of the domain will reappear on the opposite face with the same velocity. We employ periodic boundary conditions both in the $x$ and $y$ direction. Periodic boundary conditions have been widely used to simulate convection in a small domain of a cloud-resolving model interacting with the large-scale flow (e.g. Grabowski et al., 1996, Robe and Emanuel, 1996, Johnson et al., 2002, Guichard et al., 2004, Raymond and Zeng, 2005). Observed large-scale forcing is then applied on the CRM most often via a relaxation technique (see Randall and Cripe, 1999).
D.3 Periodic boundary conditions

Figure D.2: Implementation of periodic boundary conditions. The dark gray shaded area shows the computational domain, the light gray shaded area the halo encompassing the computational domain, black points illustrate values at grid points. The last three points of the computational domain are put in the halo at the opposite side of the domain. The procedure is shown as an example for specific points. The exchange is done for all grid points situated at the border of the domain both in the zonal and meridional direction.

In the CCLM version 4.0 periodic boundary conditions were only partly implemented and the formulation was completed as part of this PhD. The strategy chosen is to perform an exchange over the outer faces of the domain each time the exchange between neighboring processors in the interior of the domain is done. This is generally done at the end of the time step and after horizontal operations, such as advection, horizontal diffusion or after the integration of the tendencies. Technically this was done by calling the Message Passing Interface (MPI) routine “exchg_boundaries” with the argument “my_peri_neigh” in the case of a periodic exchange instead of “my_cart_neigh” in the case of an exchange between neighboring processors in the inner domain. The field my_peri_neigh respectively my_cart_neigh contains information about the location of the calling processor and the location of neighboring processors at all four sides of the calling processor.

Fig. D.2 illustrates the way information is passed from one edge of the domain to the opposite edge exemplary for the $x$ direction. The last three points of the computational domain are passed to the other side and put into the halo. Values out of the halo are then used for operations, where neighboring grid points are utilized, e.g. horizontal advection or averaging from the staggered grid of the horizontal winds to the mass points of pressure and temperature.
The performance of the periodic boundary conditions were tested by starting two simulations with a warm-air bubble at a point shifted in the $x$ direction between the simulations. The two runs were evaluated by translating one grid with the distance of the two bubbles. The difference between the two resulting fields should vanish ideally. The test was performed without the random temperature perturbation in the first time steps. The outcome of the test was satisfactory (not shown).

A further necessary modification in the code was to set the angles for the calculations of radiative forcing constant throughout the domain meaning to remove the dependence on latitude and longitude over the modeling domain.
**Bibliography**


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