Numerical study of the impact of the changes in the 
tropospheric temperature profile on the microphysics, 
dynamics and precipitation of mid-latitude summer 
continental convective clouds

Angelina Brandiyska¹, Rumjana Mitzeva²*, Boryana Tsenova³, and John 
Latham⁴,⁵

¹ Geophysical Institute, Acad. Bonchev str, Bl. 3, 1113 Sofia, Bulgaria 
imdnme@gmail.com
² Faculty of Physics, University of Sofia, James Bourchier blv, 1164 Sofia, Bulgaria 
rumypm@phys.uni-sofia.bg
³ National Institute of Meteorology and Hydrology, 
66 Tsarigradsko chausse, 1784 Sofia, Bulgaria, 
boryana.tsenova@meteo.bg
⁴ MMM Division, NCAR, 3450 Mitchell Lane, Boulder, 80301 Colorado, USA, 
⁵ SEAES, University of Manchester, Manchester, M13 9PL, UK 
latham@ucar.edu

*Corresponding author

(Manuscript received in final form October 25, 2012)

Abstract—This paper investigates the effect of the expected changes of tropospheric 
temperature profile on the dynamical and microphysical characteristics of individual 
summitime convective storms and on the processes of precipitation development in these 
storms. Two dynamically different clouds (a ‘big’ and a ‘small’ one) were simulated with the 
Regional Atmospheric Modeling System (RAMS v6.0). The differences between simulations 
of the clouds, developed in a present-day and in a modified environment are discussed. 
Macro- and microphysical evolution is examined in detail, and the changes in precipitation 
intensity and total rainfall volume are explained physically as a consequence of the 
temperature increase in the upper troposphere. Results show that the warming leads to a 
decrease of precipitation in the ‘small’ cloud case, while in the ‘big’ cloud case, warming 
leads to the increase of precipitation. The detailed analysis reveals that the main reason for the 
opposite direction of the impact of the projected tropospheric changes on different sized 
clouds lies in the ice phase evolution.

Key-words: climate warming, convective cloud, numerical model, RAMS
1. Introduction

In the scientific community, it is expected that the projected global warming will give rise to greater frequency and severity of extreme precipitation events, as stated in the Report of the Intergovernmental Panel on Climate Change (IPCC, 2007), based on studies such as Semenov and Bengtsson, 2002; Kharin and Zwiers, 2005; Meehl et al., 2006 and others. Trenberth et al., 2003 studied different methods of assessing the impact of various thermodynamical factors on precipitation, and concluded that the increased moisture content as a result of climate warming would have a significant impact on precipitation amount and intensity. As extreme precipitation relates to increases in moisture content and due to the nonlinearities involved with the Clausius-Clapeyron relationship, for a given increase of temperature, the projected increase of extreme precipitation is more pronounced than the mean precipitation increase (Allen and Ingram, 2002). In mid-latitude continental regions, summertime precipitation falls mainly from convective clouds with a great vertical extent, often supplemented by lightning, hail, floods. Obviously, changes in extreme local weather events are connected to the airmass properties – changes occurring in temperature and humidity, their vertical profiles or concentration and size of aerosols. McCaul et al. (2005), using RAMS model, examined the sensitivity of supercell storms to environmental temperature by changing only the temperature at the lifting condensation level with convective available potential temperature and other parameters being fixed. Their simulations indicate that in the limit of their assumptions, the updraft speed and precipitation efficiency are higher at a colder environment, while the peak precipitation rate in a warmer environment is comparable to that in colder environment. Numerical simulations of Takemi (2010), by Advanced Research WRF model reveal the dependence of the precipitation intensity in mesoscale convective systems on the temperature lapse rate. In the frame of their model simulations they found that with the increase of the lapse rate the mean precipitation intensity increases and the maximum precipitation intensity decreases. They stress on the need for diagnosis of stability in climate simulations and the need to investigate further the effects of cloud microphysics on the production of precipitation in a given temperature environment. To clarify this question, preliminary investigations on the impact of change in temperature and humidity profiles on microphysical characteristics of convective storms and on the processes of precipitation development would be of use. In Mitzeva et al. (2008) the influence of global warming on convective cloud dynamics and microphysics is investigated with a 1.5 D cloud model and compared with the influence of the pollution represented by CCN concentration. Their results show that in warmer profile temperature more vigorous cloud form, with larger cloud, rain water, and ice content.

In the present study, the impact of changes in tropospheric temperature and humidity profiles on individual summertime convective continental cloud
development and precipitation is investigated. Numerical simulations are performed with RAMS (regional atmospheric modeling system) and changes of temperature and humidity are imposed on a reference sounding. Two clouds were simulated which are formed with the same meteorological conditions, but triggered by different sized thermal bubbles. The simulated clouds dynamical evolution, hydrometeor mixing ratios, and resulting precipitation are evaluated in detail. The analysis focuses on clarifying the changes in the process of formation of precipitation, with emphasis on the ice phase process. The present study is a first step for testing whether and how temperature profile changes would affect convective clouds development without pretending to be sufficient for the drawing of general conclusions.

2. Cloud model

The model used in the present study is the Regional Atmospheric Model System (RAMS v.6.0), which is developed by the Colorado State University and is widely used as a research tool for numerical studies of thunderstorms (Pielke et al., 1992; Cotton et al., 2003). It combines the capabilities of different types of atmospheric models, from mesoscale to large eddy simulations. RAMS is a 3-dimensional non-hydrostatic cloud resolving atmospheric model. It includes equations and parameterizations for a wide range of physical processes, as it is designed as a comprehensive meteorological modeling system: advection, diffusion, turbulence, radiation, and large-scale precipitation; it also includes a land surface model and a bulk microphysics scheme for resolved clouds and precipitation. Spatial resolution can vary from hundreds of meters to hundreds of kilometers and the time-step can be fixed or varying, user-defined, or automatically calculated by the model.

The two-moment bulk microphysics scheme in RAMS predicts both mass mixing ratio and number concentration of hydrometeor species thus allowing the mean diameter of hydrometeors to evolve. (Meyers et al., 1997). CCN number concentration is defined in the beginning of the run and it is kept constant during the simulations. The size distribution of aerosols is included indirectly through the shape of the cloud droplet spectrum. Cloud droplet as well as other hydrometeor spectra are approximated by a gamma function, which depends on prognosed mean radius and number concentration, and it has a fixed shape.

Sensitivity studies (e.g., Demott et al., 1994; Van den Heever et al., 2006; Phillips et al., 2007, and others) show that nucleation processes are fundamental microphysics processes that must be routinely included in cloud schemes to capture the lifecycle of convective clouds. For the present simulations, since we focus on the impact of the temperature profile changes on cloud formation and development by simulating idealized convective clouds, CCN concentration and
gamma shape parameters for different hydrometeor species are selected in an appropriate way to represent microphysical conditions typical for mid-latitude continental climate.

In the model, seven species of hydrometeors are categorized: cloud droplets, rain, pristine ice, snow, aggregates, graupel, and hail. Cloud droplets and pristine ice nucleate from vapor and may convert into other categories after they grow. Together, pristine ice and snow categories represent a bimodal distribution of ice crystals. Graupel has higher density than snow but still consists mostly of ice and can carry only a small percentage of liquid. If the percentage of liquid is higher than 30%, the ice particle is considered as hail – the highest density hydrometeor formed by freezing of raindrops, riming or partial melting of graupel. The fall speeds of the hydrometeors vary depending on diameter and category. Note, that the definitions of graupel and hail categories in the model emphasize their composition and density rather than their method of formation. Thus, a melting graupel particle will increase its liquid fraction and density, and it will be categorized as hail.

Apart from nucleation, vapor deposition, riming, and melting, coalescence is also taken into account by using pre-calculated look-up tables that contain approximate solutions of the stochastic collection equation. The two-moment scheme also includes breakup of raindrops (formulated into the collection efficiency), diagnosis of pristine ice and snow habit dependent on temperature and saturation, evaporation and melting of each species. Details of microphysical parameterization can be found in Meyers et al., (1997).

3. Numerical simulations

In the present study, in order to lower the computational expense of the experiments, some of the schemes in the model are switched off (short-wave and long-wave radiation, soil model). They do not have significant influence on processes of such space and time scales like single convective storms. The microphysics parameters are selected in a way to represent an urban region with a mid-latitude continental climate. The hydrometeor shape-parameters as well as other initialization parameters are shown in Table 1. The time step is automatically calculated by the model, depending on the resolution.

The model domain dimensions are 30 × 30 × 15.5 km consisting of 50 × 50 × 60 points with 600 m grid spacing in a Lambert conformal projection. It covers the area of Sofia City with center point latitude/longitude of 42.65 N/23.38 E. The topography is simplified to a flat surface with elevation of 595 m above sea level. The vertical coordinate is terrain-following sigma coordinate with level thickness ΔZ increasing from 30 m at ground level to 300 m in the free troposphere with a stretch ratio of 1.25. Initial conditions (at the surface) are horizontally homogeneous and are taken from the aerological sounding
measurement in Sofia on 14 July 2006 – a day when a (moderate) thunderstorm was observed, or from the modified sounding.

Table 1. Forcing conditions of the numerical simulations

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Horizontal resolution</td>
<td>600m</td>
</tr>
<tr>
<td>Vertical resolution at z=0m</td>
<td>30m</td>
</tr>
<tr>
<td>Vertical resolution above z=2000m</td>
<td>300m</td>
</tr>
<tr>
<td>Large time step</td>
<td>4s</td>
</tr>
<tr>
<td>Small time step</td>
<td>1s</td>
</tr>
<tr>
<td>CCN number</td>
<td>$1.0 \times 10^9$ #/m$^3$</td>
</tr>
<tr>
<td>gamma shape parameter, cloud droplets</td>
<td>3.0</td>
</tr>
<tr>
<td>gamma shape parameter, raindrops</td>
<td>2.0</td>
</tr>
<tr>
<td>gamma shape parameter, pristine crystals</td>
<td>1.5</td>
</tr>
<tr>
<td>gamma shape parameter, snow</td>
<td>2.0</td>
</tr>
<tr>
<td>gamma shape parameter, aggregates</td>
<td>2.0</td>
</tr>
<tr>
<td>gamma shape parameter, graupel</td>
<td>2.0</td>
</tr>
<tr>
<td>gamma shape parameter, hail</td>
<td>2.0</td>
</tr>
<tr>
<td>Thermal bubble amplitude ($\Delta T$)</td>
<td>2K</td>
</tr>
<tr>
<td>Thermal bubble relative humidity ratio</td>
<td>1.2</td>
</tr>
<tr>
<td>Thermal bubble horizontal size</td>
<td>3km; 4.2km</td>
</tr>
<tr>
<td>Thermal bubble vertical size</td>
<td>1.1km; 1.7km</td>
</tr>
</tbody>
</table>

Boundary conditions are open. Simulations were carried out for one hour starting at 12:00 UTC. A real sounding was used in order to assure that the environmental conditions chosen for model simulation are adequate for thunderstorm formation, rather than to simulate the real thunderstorm that developed over Sofia. Clouds were initiated by introducing a warm moist bubble at the surface. The so called “thermal bubble” had 2 K temperature excess and 20% higher relative humidity compared to the environmental air. By varying the thermal bubble’s horizontal and vertical dimensions, two convective clouds were simulated – cloud A, simulated with initial bubble size of $4.2 \times 4.2 \times 1.7$ km and cloud B, simulated with initial bubble size of $3 \times 3 \times 1.7$ km. Thus, our study is an idealized modeling study. As a result of an appropriate choice of the initial conditions, two typical (for mid-latitude) single summer continental precipitating clouds were simulated. Hereafter, we will call them ‘big’ and ‘small’, because one (cloud A) is bigger than the other (cloud B). The ‘control simulation’ is the simulation carried out with the original sounding from July 14, 2006, for cloud A (simulated with the larger thermal bubble) and for cloud B (simulated with the smaller thermal bubble). The modifications of the temperature profile used in the present study (Fig. 1) are based on IPCC,
2007 and Santer et al., (2003). IPCC climate projections predict an increase of global mean surface temperature from 2 to 6 °C till the end of the 21st century. The investigations of Santer et al., (2003) reveal that temperature changes with height depend on latitude and the temperature increase in the upper troposphere for the mid-latitudes is about 1.5 times greater than the surface temperature with maximum around 300 mb. Due to the very limited studies and no clear trend in relative humidity (Elliot and Angell, 1997) in our simulations the dew-point temperature profile was modified to keep the relative humidity in accordance to the original sounding, i.e., the relative humidity was not modified. The control simulation (with the original sounding) is hereafter denoted as $\Delta T=0$ and the simulations with two modified profiles are denoted as $\Delta T=3$ and $\Delta T=5$ respectively, where the numbers 3 and 5 refer to the surface temperature increase in degrees Celsius. The original sounding is shown in Fig.1a and the modification of temperature profile is shown in Fig. 1b. The relative humidity is not modified among the runs.

*Fig. 1*. a) Aerological sounding from 14 July 2006; b) increase of temperature $\Delta T$ used for the simulations of global warming; c) atmospheric stability for the simulated cases with $\Delta T=0$, $\Delta T=3$, $\Delta T=5$, $Z$ is in km ASL.
The original sounding is characterized by alternating moist conditionally unstable and stable layers (Fig. 1c), evaluated using the temperature lapse rate. The instability increases above 5.3 km in $\Delta T=3$ and $\Delta T=5$. This tendency is most pronounced in the layers between 6.1 km and 7.2 km in $\Delta T=5$. In the original sounding, the zero isotherm is located at 3.6 km. The modifications lead to its upward displacement, at 4.1 km for $\Delta T=3$ and 4.6 km for $\Delta T=5$, respectively. The temperature interval where homogeneous freezing occurs is also shifted upward in the warmer environment. The 40 °C isotherm in the original sounding is at 9.4 km AGL above ground level, while for $\Delta T=3$ it is at 10 km AGL and for $\Delta T=5$ - at 10.3 km AGL. Due to the assumption that the relative humidity is kept constant with the warming, the condensation level is almost at the same altitude (around 1.7 km AGL). However, the temperature at cloud base (even at the same altitude) is higher with the warming modification.

4. Results

The results are presented in the following manner. First, we consider the dynamical and microphysical characteristics of cloud A (‘big’ cloud), simulated with the reference sounding. After that, the same characteristics of simulations $\Delta T=3$ and $\Delta T=5$ are analyzed, pointing to the differences between them and the control one. The analysis of cloud B (‘small’ cloud) is presented in the same way.

Cloud A simulated with the reference sounding has a maximum cloud top height of 12 km above ground level (AGL) and maximum updraft speed of 24 m/s. Its anvil is well above the level of maximum heating in warming scenarios. The smaller cloud (cloud B) simulated with the reference sounding has maximum cloud top at 10 km AGL and maximum updraft speed of 18 m/s.

4.1. Cloud A

4.1.1. Macro- and microphysical evolution of the control cloud

Condensation starts at 9 min model time (MT) at 1726 m above ground level (AGL), at temperature 14.2 °C. The cloud grows during the following 12 minutes (Fig. 2) and at 21 min MT the cloud top reaches the tropopause (10.7 km AGL). During the rapid growth stage (18–21 min), two maximums of the vertical velocity are visible in Fig. 3. At 18 min MT, these maximums can be seen at ~5 km AGL (21 m/s) and at ~7 km AGL (24 m/s), respectively. The sounding instability and high moisture (Fig. 1) provide the energy needed to accelerate the updraft. The level of the maximum water mixing ratio (Fig. 2) coincides with the level of the first updraft velocity maximum (Fig. 3). A slight decrease in velocity at 18 min MT is seen between 5 and 6 km in Fig. 3, due to the
dry and stable environment. The analyses reveal that above 6 km AGL (−14 °C),
intensive freezing occurs and the updraft speed increases again. Between 18 and
21 min MT the cloud top reaches the layer above 7 km AGL where the moisture
increases (Fig. 1) and the environment is unstable. The main reason for the
increase of updraft velocity above 9 km AGL (Fig. 3) is the latent heat released
by homogeneous freezing.

*Fig. 2.* Cloud A: sum of rain and cloud water mixing ratios (solid) and sum of pristine ice,
snow, and aggregates mixing ratios (dashed) in the control simulation (ΔT=0).

*Fig. 3.* Cloud A: updraft velocity (solid) and temperature (dashed) at 18 and 21 min
model times in the control simulation (ΔT=0).
When the simulated cloud reaches the tropopause (at about 21 min MT), it stops moving upwards and an anvil begins to spread (Fig. 2). At this time the cloud is in a mature stage, precipitation reaches the ground (Fig. 2 bottom panel), and the updraft core start to decay (not shown here).

When the simulated cloud reaches the tropopause (at about 24 min MT) it stops moving upwards and an anvil begins to spread (Fig. 2). At this time the cloud is in a mature stage and some of precipitation reaches the ground (Fig. 2 bottom panel).

The first ice particles form between 15 and 18 min MT at 5.5 km AGL (see Fig. 2), where the in-cloud temperature is about –12°C. Afterwards, the mixing ratio of all ice species increases (Fig. 2 and Fig. 4). Around 21 and 27 min MT, the maximum values of mixing ratio of hail and graupel, respectively, are reached (Fig. 4, Table 2a). At 27 min the updraft starts to weaken and part of graupel and hail particles fall down, melt, and transform to rain drops. As a result, graupel and hail mixing ratios in the simulated cloud A (Fig. 4) decrease. After 33 min MT the cloud starts to dissipate. The precipitation intensity simulated by the model with reference sounding maximum (in space and time) is 66.18 mm/hr, and the maximum accumulated precipitation (in space) is 10.25 l/m². The total rainfall volume yield (total volume of liquid water fallen on the ground) is $1.98 \times 10^6$ m³.

4.1.2. Effect of warming – comparison between simulations with modified soundings and the control run.

As it can be seen from Fig. 1, the increase in atmospheric temperature as a result of the projected regional climate change is expected to be the largest at 300 hPa. Below 4 km ASL (~650 hPa), in warmer environment, the lapse rate is slightly lower, and convection will be suppressed, while above 9 km (300 hPa) there is an increase of the lapse rate and stability decreases. Due to the assumption that the relative humidity is not changed, specific humidity increases where temperature increases. Increased humidity means increased reservoir of water and the associated latent energy, so there is a potential for intensification of cloud dynamics. Intensive storms (with strong updraft) usually reach the 300 hPa level which in our simulation is at 8900 m AGL.

Another important consequence of warming is the upward displacement of the zero isotherm (from 3.6 km in $\Delta T=0$, to 4.1 km in $\Delta T=3$ and 4.6 km in $\Delta T=5$), which affects the formation and development of ice particles in the simulated clouds. Additionally, the cloud base in $\Delta T=3$ is at 17.5 °C, while in $\Delta T=5$ it is at 19.7 °C, or respectively, it is about 3.3 and 5.5 °C warmer than the control cloud base. Together with the zero isotherm upward displacement, the increase of the cloud base temperature due to the warming (and the keeping of the relative humidity unchanged) leads to an extension of the region with temperature higher than 0°C in the simulated ‘warming’ clouds in $\Delta T=3$ and $\Delta T=5$. 
Table 2a. Maximum (in space and time) values for cloud, rain, pristine, snow, aggregates, graupel, hail mixing ratio [g/kg] for simulated “big” cloud

<table>
<thead>
<tr>
<th></th>
<th>Cloud</th>
<th>Rain</th>
<th>Pristine</th>
<th>Snow</th>
<th>Aggregate</th>
<th>Graupel</th>
<th>Hail</th>
</tr>
</thead>
<tbody>
<tr>
<td>ΔT=0</td>
<td>6.39</td>
<td>4.09</td>
<td>0.59</td>
<td>0.67</td>
<td>1.14</td>
<td>2.5</td>
<td>7.14</td>
</tr>
<tr>
<td>ΔT=3</td>
<td>6.55</td>
<td>4.81</td>
<td>1.04</td>
<td>0.64</td>
<td>1.15</td>
<td>2.4</td>
<td>9.35</td>
</tr>
<tr>
<td>ΔT=5</td>
<td>6.78</td>
<td>7.93</td>
<td>1.42</td>
<td>0.69</td>
<td>1.37</td>
<td>2.23</td>
<td>11.04</td>
</tr>
</tbody>
</table>

Table 2b. Integrated (space and time) values for cloud; rain, pristine, snow, and aggregates; graupel and hail mixing ratios [g/kg] for simulated “big” cloud

<table>
<thead>
<tr>
<th></th>
<th>Cloud</th>
<th>Rain</th>
<th>Pristine+Snow+Aggregates</th>
<th>Graupel+Hail</th>
</tr>
</thead>
<tbody>
<tr>
<td>ΔT=0</td>
<td>5758</td>
<td>9933</td>
<td>12766</td>
<td>33497</td>
</tr>
<tr>
<td>ΔT=3</td>
<td>6062</td>
<td>12696</td>
<td>17041</td>
<td>32558</td>
</tr>
<tr>
<td>ΔT=5</td>
<td>6163</td>
<td>14999</td>
<td>18335</td>
<td>28668</td>
</tr>
</tbody>
</table>

Fig. 4. Cloud A: hail (solid) and graupel (dashed) mixing ratios from 18 to 33 min model time in the control simulation (ΔT=0).
The results presented in Table 2a show that maximum values of cloud water and aggregates mixing ratio only slightly increase with warming, while there is a significant increase in maximum values of pristine ice, rain, and hail mixing ratios. The warming leads to the increase of space and time integrated mixing ratio of liquid hydrometeors and of small ice particles and to the decrease of the sum of graupel and hail mixing ratio (Table 2b). An idea for changes of the evolution of space integrated mixing ratios as a result of warming gives Figs. 5a – 5d. The effect of warming on space integrated cloud water mixing ratio is not well pronounced (Fig. 5a), while a significant increase of space integrated rain water mixing ratio until 30 min MT is visible in Fig. 5b for warmer cases. After 22 min MT, the sum of space integrated ice, snow and aggregates mixing ratios also increases with warming (Fig. 5c). The decrease in sum of space integrated graupel and hail mixing ratios (Fig. 5d) can be explained by melting of graupel and hail in the warmer environment, which contributes to the increase of rain water mixing ratio.

---

**Fig. 5.** Cloud A: temporal evolution of a) space integrated cloud water; b) space integrated rain water; c) sum of space integrated ice, snow and aggregates; d) sum of space integrated graupel and hail mixing ratios in g/kg as a function of model time MT in min for ΔT=0, 3, 5.

The top of the control cloud reaches 300 hPa (8.9 km AGL) at 18 min MT. After 18 min MT, the updraft velocity and the processes of formation and
growth of ice particles are intensified and the cloud top rises fast, due to the homogeneous freezing that occurs between $-30^\circ$C and $-50^\circ$C in the model. The analyses reveal that the clouds simulated with $\Delta T=3$ and $\Delta T=5$ experience similar stages of development to the control cloud ($\Delta T=0$). There are only slight differences in cloud top height (Fig. 6) and location and intensity of updraft core (Fig. 7). From Fig. 7 one can see that at 18 min MT, the updraft velocities are almost identical in $\Delta T=0$, $\Delta T=3$, and $\Delta T=5$.

Fig. 6. Cloud A: cloud top height as a function of time for $\Delta T=0, 3, 5$.

Fig. 7. Cloud A: updraft velocity (solid) and temperature (dashed) at 18 min MT (top panel) and 21 min MT (bottom panel) for $\Delta T=0, 3, 5$ (left, middle, and right panel, respectively).
In the warmer environment, the updraft velocity in the lower parts of simulated clouds has slightly lower values due to slightly lower lapse rates at heights below 4 km. Later, above 4 km, the updraft velocity in the warming cases is larger due to the released latent heat from the frozen water during the formation of hail particles. The top panel of Fig. 9 shows that the hail mixing ratio increases with warming. Ice crystal particles (Fig. 8a, top panel) form a few minutes earlier in the control run than in the clouds, simulated with $\Delta T=3$ and $\Delta T=5$. The reason for the earlier formation of ice particles at lower altitude in $\Delta T=0$ is the lower temperature in the original sounding. At 21 min MT, all three clouds have passed the $-40$ °C isotherm which is at higher altitudes (9.3, 9.9, and 10.3 km AGL, respectively for the control run, $\Delta T=3$, and $\Delta T=5$) in warming cases. The updraft velocity in the latter case is highest, showing very intensive growth (Fig. 7). This corresponds to the thermodynamic effect of the increased lapse rate above 9 km AGL in warming cases. Another reason for the increase of the updraft velocity in the “big” cloud developed in the warmer environment is the higher water mixing ratio at higher levels in these cases (Fig. 8a, top panel). Liquid water, once frozen, leads to the increase of the ice mixing ratio (Fig. 8a, bottom panel) above $-30$ °C and the release of more latent heat. A strong maximum of 33 m/s is seen (Fig. 7) in $\Delta T=5$ case (compared to 24 m/s in the control cloud and 30 m/s in $\Delta T=3$).

From Figs. 8 and 9 it is visible that the higher liquid water and higher ice mixing ratio (Fig. 8a) in the warmer environment lead to higher hail mixing ratio (Fig. 9a, top panel). At 24 min MT, the hail mixing ratio in the layer between 5 and 10 km is almost double in $\Delta T=5$ compared to $\Delta T=0$. The analyses show that in cases $\Delta T=3$ and $\Delta T=5$, the hail particles have larger liquid fraction, higher density, and larger fall velocity, so they fall earlier to lower levels in comparison with $\Delta T=0$. Precipitation for all three simulations begins at about 21 min MT and stops at 50 min MT (Fig. 10). Melted hail particles contribute to the earlier occurrence of the maximum precipitation rate in the warming cases – 30 min MT in $\Delta T=5$, compared to 33 min in $\Delta T=0$. Precipitation rate, accumulated precipitation, and total rainfall volume increase in the warm simulations (Table 3). The peak precipitation rate in $\Delta T=5$ is almost twice as high as in $\Delta T=0$.

<table>
<thead>
<tr>
<th>$\Delta T$</th>
<th>Peak precipitation rate [mm/h]</th>
<th>Peak accumulated precipitation [l/m²]</th>
<th>Total rainfall volume [m³]</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\Delta T=0$</td>
<td>66.18</td>
<td>10.25</td>
<td>$1.98 \times 10^6$</td>
</tr>
<tr>
<td>$\Delta T=3$</td>
<td>71.71</td>
<td>11.40</td>
<td>$2.23 \times 10^8$</td>
</tr>
<tr>
<td>$\Delta T=5$</td>
<td>112.24</td>
<td>12.75</td>
<td>$2.42 \times 10^8$</td>
</tr>
</tbody>
</table>

Table 3. Peaks of rainfall rate, accumulated (for 50 min MT) rainfall, and accumulated total rainfall volume yield from cloud A (“big” cloud)
Fig. 8. Cloud A: sum of rain and cloud water mixing ratios (solid) and sum of pristine ice, snow, and aggregates mixing ratios (dashed) at 18 min MT, 21 min MT, 24 min MT, and 27 min MT (from top to bottom panel) for $\Delta T=0$, 3, 5 (left, middle, and right panel, respectively).
Fig. 9. Cloud A: hail (solid) and graupel (dashed) mixing ratios at 24 min MT, 27 min MT, 30 min MT, and 33 min MT (from top to bottom panel) for $\Delta T=0$, 3, 5 (left, middle, and right panel, respectively).
4.2. Cloud B

Cloud B is the ‘small’ cumulonimbus that develops in the same environment as cloud A, but it is initiated with a smaller thermal bubble (see Table 1).

4.2.1. Macro- and microphysical evolution of the control cloud

Similarly to cloud A, cloud development begins at 9 min MT, and the cloud base height is at 1726 m AGL (14.16 °C). During the next 15 min, the cloud top rises at a rate of 10 m/s until 24 min MT and at 27 min MT it reaches its maximum height of 9.5 km AGL, where the temperature in the environment is –41 °C (Figs. 12, and 15). The maximum updraft velocity (18 m/s) is reached at ~5 km AGL at 24 min MT (Fig. 12). After 24 min MT, the updraft velocity starts to decrease and after 30 min MT, the downdrafts prevail. Then the simulated cloud B begins to dissipate, and precipitation falls on the ground.

Ice formation (Fig. 11) starts after 18 min MT at an altitude above 4 km (about –5 °C). The formation of hail and graupel starts at the same time and altitude (Fig. 13). Maximum mixing ratios of hail and graupel are reached at 21 and 24 min MT, respectively (Fig. 13 and Table 4a).
Fig. 11. Cloud B: sum of rain and cloud water mixing ratios (solid) and sum of pristine ice, snow, and aggregates mixing ratios (dashed) in the control simulation ($\Delta T=0$).

Fig. 12. Cloud B: updraft velocity (solid) and temperature (dashed) at 21 and 24 min model times in the control simulation ($\Delta T=0$).
Fig. 13. Cloud B: hail (solid) and graupel (dashed) mixing ratios from 18 to 33 min model times in the control simulation ($\Delta T=0$).

Table 4a. Maximum (in space and time) values for cloud, rain, pristine, snow, aggregates, graupel, hail mixing ratio [g/kg] for cloud B formed in different environmental conditions

<table>
<thead>
<tr>
<th></th>
<th>Cloud</th>
<th>Rain</th>
<th>Pristine</th>
<th>Snow</th>
<th>Aggregate</th>
<th>Graupel</th>
<th>Hail</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\Delta T=0$</td>
<td>5.37</td>
<td>1.51</td>
<td>0.14</td>
<td>0.4</td>
<td>0.71</td>
<td>2.32</td>
<td>2.81</td>
</tr>
<tr>
<td>$\Delta T=3$</td>
<td>5.43</td>
<td>1.47</td>
<td>0.08</td>
<td>0.51</td>
<td>0.55</td>
<td>1.62</td>
<td>3.58</td>
</tr>
<tr>
<td>$\Delta T=5$</td>
<td>5.26</td>
<td>1.81</td>
<td>0.1</td>
<td>0.37</td>
<td>0.76</td>
<td>1.17</td>
<td>3.38</td>
</tr>
</tbody>
</table>

Table 4b. Integrated (space and time) values for cloud and rain water; pristine, snow and aggregates; graupel and hail mixing ratios [g/kg] for cloud B formed in different environmental conditions

<table>
<thead>
<tr>
<th></th>
<th>Cloud</th>
<th>Rain</th>
<th>Pristine+Snow+Aggregates</th>
<th>Graupel+Hail</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\Delta T=0$</td>
<td>1743</td>
<td>1544</td>
<td>549.39</td>
<td>3895.28</td>
</tr>
<tr>
<td>$\Delta T=3$</td>
<td>1542</td>
<td>1387</td>
<td>272.97</td>
<td>1810.96</td>
</tr>
<tr>
<td>$\Delta T=5$</td>
<td>1372</td>
<td>1223</td>
<td>107.08</td>
<td>954.11</td>
</tr>
</tbody>
</table>
The lifecycle of this cloud is different from that of cloud A. There is less vigorous convection and less condensation takes place, so the cloud grows smaller and the mixing ratios of all hydrometeors are lower. Their size (not shown here) and fall velocity are also lower.

The simulated maximum (Table 5) precipitation intensity of space and time is 16.77 mm/hr and the peak accumulated precipitation is 2.25 l/m². The total rainfall volume yield is 0.253’10⁶ m³, which is ten times less than the rainfall volume yield in cloud A.

Table 5. Peaks of rainfall rate, accumulated (for 50 min MT) rainfall and accumulated total rainfall volume yield for Cloud B (“small” cloud)

<table>
<thead>
<tr>
<th>ΔT=0</th>
<th>16.77</th>
<th>2.25</th>
<th>0.25 10⁶</th>
</tr>
</thead>
<tbody>
<tr>
<td>ΔT=3</td>
<td>15.01</td>
<td>1.65</td>
<td>0.19 10⁶</td>
</tr>
<tr>
<td>ΔT=5</td>
<td>12.85</td>
<td>1.24</td>
<td>0.14 10⁶</td>
</tr>
</tbody>
</table>

4.2.2. Effect of warming – comparison between the modified simulations and the control run.

From Fig. 14 it can be seen, that cloud top heights of simulated “small” clouds in warmer environment are a few hundred meters lower than the cloud top height of control cloud B, although the maximum heights are reached at the same time, 27 min MT. However, the analyses reveals that the control run reaches temperatures around –42 °C, while the warming cases ΔT=3 and ΔT=5 reach –37 °C and –30 °C, respectively.

Fig. 14. Cloud B: cloud top height for cloud B as a function of time for ΔT=0, 3, 5.
Results presented in Table 4a show that there is no well defined tendency in the changes of maximum values of mixing ratios of the hydrometeors as a result of warming, while the integrated (in space and time) mixing ratios of solid hydrometeors are significantly lower (see column 3 and 4 of Table 4b) at higher environmental temperatures. The warming also leads to the decrease of both cloud (column 1 in Table 4b) and rain (column 2 in Table 4b) mixing ratios integrated in space and time. The evolution of space integrated mixing ratios (Fig. 15) reveals that the space integrated rain water mixing ratio in the clouds simulated with warmer environment is higher until 27 min MT, and after that it is lower in comparison to the space integrated rain water mixing ratio in the control cloud.

\[ \text{Fig. 15. Cloud B: temporal evolution of a) space integrated cloud water; b) space integrated rain water; c) sum of space integrated ice, snow, and aggregates; d) sum of space integrated graupel and hail mixing ratios in g/kg as a function of model time MT (in min) for } \Delta T=0, 3, 5. \]

Plots of the vertical velocity for 21 and 24 min MT (Fig. 16) show that the values of the maximum vertical velocity are almost the same, however, it is slightly decreasing and located at lower altitudes with the warming. In the three simulated clouds there is almost no difference in updraft velocities below 5 km, while above 5 km in the warming cases the updraft velocity is lower. In the upper parts of clouds in } \Delta T=3 \text{ and } \Delta T=5, \text{ compared to the control one, there is a smaller amount of freezing water due to the higher temperature. Furthermore,
the clouds denoted with $\Delta T=3$ and $\Delta T=5$ do not reach the level of $\sim 40$ °C where intensive homogeneous freezing occurs. Since freezing provides latent heat to accelerate the updraft, this leads to lower values of updraft velocity, despite the decreased stability.

**Fig. 16.** Cloud B: updraft velocity (solid) and temperature (dashed) at 21 and 24 min model times for $\Delta T=0, 3, 5$.

The analysis of mixing ratio of hydrometeors in vertical cross section at different moments of time (**Figs. 17 and 18**) shows that in clouds developed in a warmer environment the total liquid water mixing ratio is slightly higher than in the simulated $\Delta T=0$ cloud, while due to higher temperatures, the values of ice (the sum of pristine, snow, and aggregates) and graupel mixing ratio (**Fig. 18**) are lower compared to the control cloud. There is higher rain water mixing ratio in warming cases (**Fig. 17**). The more detailed analysis shows, that in $\Delta T=0$ at 21 min MT, ice starts to form at 5–7 km by freezing of cloud and rain drops, while in $\Delta T=5$ there is still no ice. Coagulation of drops with graupel and hail in $\Delta T=0$ further contributes to the decrease of the liquid water mixing ratio at these altitudes. On the other hand, in $\Delta T=5$ the mixing ratios of cloud and rain water
continue to increase. After the formation of ice in $\Delta T=5$ (see bottom panel of Fig. 17), the rain mixing ratio also starts to decrease. Hail mixing ratio increases in $\Delta T=3$ compared to $\Delta T=0$ and decreases in $\Delta T=5$ compared to $\Delta T=3$. The increase in hail mixing ratio in $\Delta T=3$ and $\Delta T=5$ is accompanied by an increase of hail liquid fraction (not shown here). The reason for that can be found in the definition of hail as a higher density hydrometeor than graupel that can carry a larger fraction of liquid water. The sources for hail particles are coalescence between solid and liquid particles (when water freezes slowly) or partial melting of graupel, and the sinks are shedding and melting into rain. In warming cases, due to the increased temperature, water droplets colliding with an ice particle freeze slower, thus increasing the ice particle density and leading to the formation of hail rather than graupel. This could be a reasonable explanation for the fact that the hail mixing ratio increases in $\Delta T=3$. Obviously, in warming cases more hail particles grow in wet regime. In $\Delta T=5$ hail, particles carry a larger fraction of liquid, have higher density and consequently a higher fall velocity, so more hail particles fall to levels having positive temperature and melt to form rain. This can be the reason for the decrease of maximum hail mixing ratio in $\Delta T=5$ compared to $\Delta T=3$. These speculations are supported by the results presented in Fig. 17, where plots of the sum of cloud and rain and the sum of pristine, snow, and aggregates are shown, and in Fig. 18, where graupel and hail mixing ratios are shown. From Fig. 17 it is evident that ice starts to form later in warming cases compared to the control run. At 27 min MT, the difference in microphysical development can be seen – the control cloud still has lots of ice in its upper part, a slightly lowered liquid water mixing ratio and some graupel and hail close to cloud base (Fig. 18), while $\Delta T=5$ has lost much of its ice mixing ratio above 5 km and has more hail close to cloud base, compared to $\Delta T=0$ and $\Delta T=3$. The melting of ice is visible in Fig. 18 where hail mixing ratios for 24 min MT are high in all three clouds, but later it decreases faster in warming cases, and there is almost no ice in $\Delta T=5$ at 33 min. From the same figure it can be seen that during the mature stage (27–33 min), the hail mixing ratio close to cloud base slightly decreases in the warm simulations, while graupel mixing ratio decreases significantly.

These differences in the clouds’ microphysical evolution lead to differences in precipitation. In $\Delta T=5$ at 24 min, large amount of liquid water is “found” in the layer between 4–6 km, consisting mostly of raindrops, and it decreases sharply at 27 min (Fig. 17). At 33 min MT, there is a low amount of graupel and hail in $\Delta T=5$ compared to $\Delta T=0$, while the liquid mixing ratio in the clouds (not shown here) has similar values. Precipitation starts at the same MT for the three clouds (Fig. 19), but it is less intensive in warming cases. Despite the increased reservoir of water vapor in the warmer atmosphere, the precipitation (intensity and total volume) is lower than the precipitation from the control cloud (see Table 5, Fig. 19) The peak in the $\Delta T=5$ case is earlier than $\Delta T=0$. The decrease in total rainfall volume is 25% for “$\Delta T=3$” and another 25% for $\Delta T=5$ compared to $\Delta T=3$. 

274
Fig. 17. Cloud B: sum of rain and cloud water mixing ratios (solid) and sum of pristine ice, snow, and aggregates mixing ratios (dashed) at 18 min MT, 21 min MT, 24 min MT, and 27 min MT (from top to bottom panel) for $\Delta T=0$, 3, 5 (left, middle, and right panel, respectively).
Fig. 18. Cloud B: hail (solid) and graupel (dashed) mixing ratios at 24 min MT, 27 min MT, 30 min MT, and 33 min MT (from top to bottom panel) for $\Delta T = 0, 3, 5$ (left, middle, and right panel, respectively).
5. Conclusions

In the present study, the impact of projected changes of the temperature and humidity profiles in the mid-latitude troposphere (IPCC, 2007) on the development and precipitation from summertime convective clouds is investigated. One typical sounding from Sofia is selected, and two scenarios of projected changes, according to IPCC, 2007, are imposed on this sounding. Two clouds with different dynamics are simulated, identified here as a ‘big’ one and a ‘small’ one.

The results can be summarized in the following way for the clouds developed in a warming environment – in comparison with the “control” clouds. For both clouds (‘big’ and ‘small’):

- Sum of space and time integrated mixing ratios of graupel and hail decreases.
- Precipitation intensity has a more rapid increase in the beginning and its maximum occurs earlier.

Big cloud (A):

- Liquid water mixing ratio increases.
- Small ice particles (pristine, snow, and aggregates) mixing ratios increase.
- The updrafts are more intensive.
• Accumulated precipitation and total rainfall volume are higher.

Small cloud (B):

• Liquid water mixing ratio decreases.
• Small ice particles (pristine, snow, and aggregates) mixing ratios decrease.
• The updrafts are more intensive in lower levels and suppressed at higher levels of the cloud.
• Accumulated precipitation and total rainfall volume are significantly lower.

Results show that there are two sides of the expected pattern of change in the thermodynamic structure of the troposphere – it leads to the suppression of cloud dynamics and reduction of precipitation in the “small” cloud and to the intensification of the updraft and increase of precipitation in the “big” cloud. Our study analyzes the reasons for the contrary impact of climate warming depending on cloud intensity. For both types of cloud (“big” and “small”), the projected warming leads to higher values of liquid mixing ratio in low cloud layers and a delay in freezing, which takes place higher in the cloud. The study demonstrates that when the forcing conditions are strong enough (“big” cloud) for the cloud to reach levels with low temperatures, especially where homogeneous freezing occurs, more latent heat is released due to the freezing of a larger quantity of liquid water in warmer cases. This leads to higher content of ice particles and cloud intensification – increase of cloud updraft and cloud top. The quantity of ice precipitation particles increases and adds to the amount of liquid precipitation after melting. In these cases, the projected warming amplifies precipitation rate, accumulated precipitation, and total rainfall volume. In the cases with weaker forcing conditions (“small” cloud), when the cloud does not have enough energy to rise to higher levels (especially the level of homogeneous freezing), in a warming environment, less liquid water freezes, leading to the occurrence of a smaller quantity of ice precipitation particles inside the cloud, which in turns leads to the decrease of contribution of ice particles to the liquid precipitation. Thus, the decrease of precipitation rate, accumulated precipitation, and total rainfall volume in the “small” clouds at the increase of environmental temperature can be explained by the decrease of mixing ratios of liquid and solid precipitating particles at $\Delta T = 3$ and $\Delta T = 5$ in comparison with $\Delta T = 0$.

Results indicate the importance of the ice phase evolution in the formation of precipitation in continental mid-latitude convective clouds. Within the limitations of the model used in this study, one can conclude that in a warming environment intensive storms will have enhanced power, so they will create greater damage. On the other hand, warming can also lead to droughts by
suppressing the development of smaller cumuli. Since the present work is based on only one case study, we have to stress that different conclusions might emanate from further modeling runs encompassing wide ranges of meteorological conditions and combinations of other climate change factors. The thermodynamic forcing, investigated in the present paper, acts in the same direction as the forcing created from increasing aerosol concentrations, as explained by the conceptual model of Rosenfeld et al. (2008), so it is a challenging question how convective clouds and precipitation will change when those two factors work together.

It is also worth mentioning that the scenario tested in the present study is based on the concept of constant relative humidity. If the moisture above the continents is not enough and the relative humidity decreases, probably the effect of the thermodynamic changes in the environment air on convective clouds microphysics, dynamics, and precipitation would be reduced.

Furthermore, since the evaporation from the surface and the interactions between adjacent clouds are also important, real cases with larger domains and more than one cloud or even whole mesoscale systems should be simulated, in order to investigate the potential impact of climate warming on mid-latitude convective precipitation.

Acknowledgements — The present work is partially supported by EC through FP6 project ACCENT (GOCE-CT-2002-500337), the NATO SfP – ESP.EAP.SFPP 981393, and the Science Foundation of Sofia University (grant 153/2009).

References


