Technical Report

# Estimation of the Maximum Physically Possible Precipitation in Saxony Using a Mesoscale Atmospheric Model

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Leipzig, September 2008

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**Research & Development Contract:**

IP FLOODsite (EC Contract Number GOCE-CT-2004-505420)

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**To be cited as follows:**

Kunze, T., Hellmuth, O., Ch. Görner and Ch. Bernhofer, 2008: Estimation of the maximum physically possible precipitation in Saxony using a mesoscale atmospheric model. Technical University Dresden, Institute of Hydrology and Meteorology, IP FLOODsite Technical Report

## Task Setting

*Excerpt from the contract:*

Gegenstand der FuE-Arbeiten des IfT ist die Abschätzung des physikalisch maximal möglichen Niederschlags mittels eines Atmosphärenmodells für Sachsen für unterschiedliche Klimaszenarien. Ausgangspunkt hierfür bilden die Daten des Hochwasserereignisses im August 2002, welche mit über 300 mm/d dem maximal möglichen Niederschlag im Ist-Zustand des Klimas wahrscheinlich sehr nahekommen. Dieses Ereignis bildet die „Baseline“ für einen Vergleich mit der Auswirkung von verschiedenen Temperaturerhöhungen bei sonst gleichbleibenden Randbedingungen. Darauf aufbauend ist eine Anpassung von Temperatur und anderen Randbedingungen an vorliegende Klimaszenarien möglich, woraus sich Änderungen des physikalisch maximal möglichen Niederschlags in Abhängigkeit vom in den Szenarien angenommenen Klimawandel berechnen lassen.

Baseline: Niederschlag des August 2002 mit 312 mm/d (7-7 h) Maximalwert an der Station Zinnwald

Szenario 1: $\Delta T$=+2°C bei gleichbleibender relativer Feuchte (was einer Erhöhung des Dampfdruckes entspricht)

Szenario 2: $\Delta T$ =+4°C bei gleichbleibender relativer Feuchte (was einer weiteren Erhöhung des Dampfdruckes entspricht)

Die Ergebnisse der Baseline-Simulation werden mit den beobachteten Werten verglichen und die Abweichungen für unterschiedliche im Untersuchungsmonat aufgetretene Ereignisse bewertet. Als Simulationswerkzeug ist ein mesoskaliges Modell (z.B. das Lokalmodell des DWD) in einer der Fragestellung adäquaten, räumlichen Auflösung vorgesehen. Die für die beiden Szenarien erreichten Werte werden unter sonst gleichen Randbedingungen simuliert, um den durch die Änderungen des globalen Klimas bedingten zusätzlichen Niederschlag abzuschätzen.

Der Auftraggeber ist sich bewusst, dass es sich dabei um ein numerisches Experiment handelt, dass nur die veränderten Ausgangsbedingungen berücksichtigt und strenggenommen keine Übertragung auf zukünftige Klimabedingungen erlaubt.

Subject of the present research and development project is an estimation of the maximum physically possible precipitation in Saxony by means of an atmospheric model, executed for different climate scenarios. Starting point is the phenomenology of the flood of the river Elbe in August 2002. The observed rain sum of more than 300 mm/d is suspected to be close to the maximum possible rain sum, referring to the as-is state of the regional climate. This event serves as the baseline for a comparison with different scenario simulations, in which the environmental temperature is enhanced by hypothetical increments as a result of a predicted global warming, whereas other boundary conditions remain unchanged. On this base, an adjustment of the temperature and other boundary conditions to available climate scenarios is possible. This allows the calculation of changes in the maximum physically possible precipitation in dependence on prescribed climate scenarios.

| | |
|---|---|
| Baseline: | Precipitation of August 2002 with 312 mm/d (7-7 h) observed maximum precipitation at the meteorological station Zinnwald |
| Scenario 1: | $\Delta T$=+2°C at constant relative humidity (corresponding to an increase in the water vapour partial pressure) |
| Scenario 2: | $\Delta T$=+4°C at constant relative humidity (corresponding to a further Increase in the water vapour partial pressure) |

The results of the baseline simulation will be compared with observed rain sums. Detected deviations will be evaluated with respect to different meteorological events, occurring during the period of interest. As simulation tool a mesoscale model (e.g., the Local Model of the DWD) with an adequate grid resolution is envisaged. The simulations adopting optional temperature enhancements have to be performed under otherwise constant boundary conditions to estimate the hypothesised precipitation surplus in the course of a predicted change of the Earth's climate.

The principal is aware, that the intended study is a numerical experiment, which only considers changed basic conditions. Thus, strictly spoken, the results cannot be extrapolated to future climate conditions.

# Table of Content

## 1. Introduction

Owing to its socio-economical implications, the impact of climate change on the evolution of possible severe weather situations (storms, flashfloods, etc.) in the future is of high practical interest. The assessment of the spatio-temporal evolution of such singular events deserves a multiscale simulation approach based on a Global Circulation Model (GCM), which provides the necessary initial and boundary conditions to force a Numerical Weather Prediction (NWP) model on the regional scale. However, for practical and methodical reasons, conclusive statements on the frequency, location and strength of such events in a future climate are not yet possible.

Climate and weather prediction are based on finite systems of deterministic ordinary nonlinear differential equations to describe forced dissipative hydrodynamic flow. It has long been known, that among the solutions of these equations are also nonperiodic ones, which are ordinarily unstable with respect to small modifications, so that slightly differing initial states can evolve into considerably different states. This affects the feasibility of very-long-range weather prediction (known as “butterfly effect”). Apart from the extreme nonlinearity of complex atmospheric models, one of the most serious hindrances for a reliable prediction of future regional climate states is the still insufficient understanding of atmospheric phase transitions down to molecular scales (aerosol, cloud and rain formation). The question, which time and length scale mostly contributes to the evolution of a macroscopic system is hitherto not generally answered.

A comprehensive multiscale simulation approach is beyond the scope of the present paper. Nevertheless, to gain at least some tentative statements about the impact of climate change on extreme weather events, we performed an epignostic sensitivity study of the century flood of the river Elbe in August 2002.

## 2. Synoptic Characterisation of the Elbe River Flood

Owing to its disastrous consequences the flood of the river Elbe in August 2002 became an inglorious part of human commemoration, certainly not only insofar personal fates were concerned. However, with respect to the phenomenology and genesis, this meteorological event possesses textbook character.

The synoptic situation, leading to this historical event, can be categorised as a classical Vb weather situation (cf. Figs. 1-4). Such a situation can be characterised as follows: Usually, a well formed upper trough is situated over Central Europe (cf. Fig. 1, small picture). On its southwestern flank a low pressure zone develops in the Gulf of Genua (cf. Fig. 1, large picture, Fig. 2) due to the favourable orographic and baroclinic constellation. The track of the cyclone is steered by the flow at the 500 hPa pressure level, determined by the upper trough (cf. Fig. 3). During its evolution, the cyclone passes the Alps in the East, afterwards moving northward on the eastern side of the upper trough (this position is favourable for cyclogenesis) and passing the Czech Republic and Poland. While in the trough region a relatively cool air mass is present, a warm and humid air mass is advected by the propagating low from the Mediterranean Sea to the North. Such a synoptic constellation provides favourable conditions for intensive and extended lifting of a very humid air mass, which is an essential precondition for the formation of heavy rainfalls. These rainfalls can be further increased by orographic lifting. This is exactly the case in the region of interest with its special rain-promoting orographic constellation – a mountain ridge, ranging from the Ore Mountains (Erzgebirge) in the West to the Giant Mountains (Riesengebirge) in the East.

In the case of interest, occurring in August 2002, the upper trough was very intense. It developed into a cut-off upper low over the alpine region. A remnant of the old upper trough remained over the North Sea. Thus, in the upper levels we had a westerly flow, competing with an easterly flow. As a consequence, the surface low with its lifting zone only slowly moved to the North over Poland. The lifting zone stayed quasi-stationarily on the northeastern flank of the cut-off for more than one day. Due to the resulting strong northwesterly winds this position led to intensive orographic lifting at the Ore Mountains. The latent heat storage in the Mediterranean air mass, usually particularly high during this season, together with the orographic lifting led to the enormous rain amounts.

Figure 4 shows the satellite image of the cloud vortex with overlay of radar echoes. The grain texture of the radar echoes refers to embedded convective cloud cells, causing rain enhancement due to intensified vertical updrafts inside the clouds (promoting rain formation via a complex chain of interacting hydrothermodynamical and microphysical processes). The spatial distribution of the 72-h precipitation sum is depicted in Fig. 5. Wide areas of the eastern parts of Germany, Czech Republic and Austria were affected by heavy rains. One has to note, that observed rain sums above 190 mm are not explicitly resolved in Fig. 5.

Local peak values in the rain sum pattern originated from the superposition of three physical processes, mentioned above: large scale frontal lifting, regional scale orographically induced lifting and local scale convective processes embedded in the eastern unstable warm air mass. An example for the consequences of such a superposition is the extremely high accumulated rain sum measured at the meteorological station Zinnwald-Georgenfeld from 12.08.2002, 00UTC to 13.08.2002, 06UTC. With precipitation rates of up to the observed 312 mm/day in Zinnwald-Georgenfeld (vide infra), the rain sums assumed extraordinarily high values.

The precipitation observed during the century flood is suspected to be very close to the physically possible precipitation in the Saxony region. For example, the repetition interval for this flooding event at many rivers has been estimated to be in the range 200-500 years (LfUG, 2004; Arnhold, 2005). Previous reports on such high rain rates in the Saxony region in former times are not known. There is no doubt, that the simulation of such extreme peaks is one of the most challenging tasks in atmospheric science.

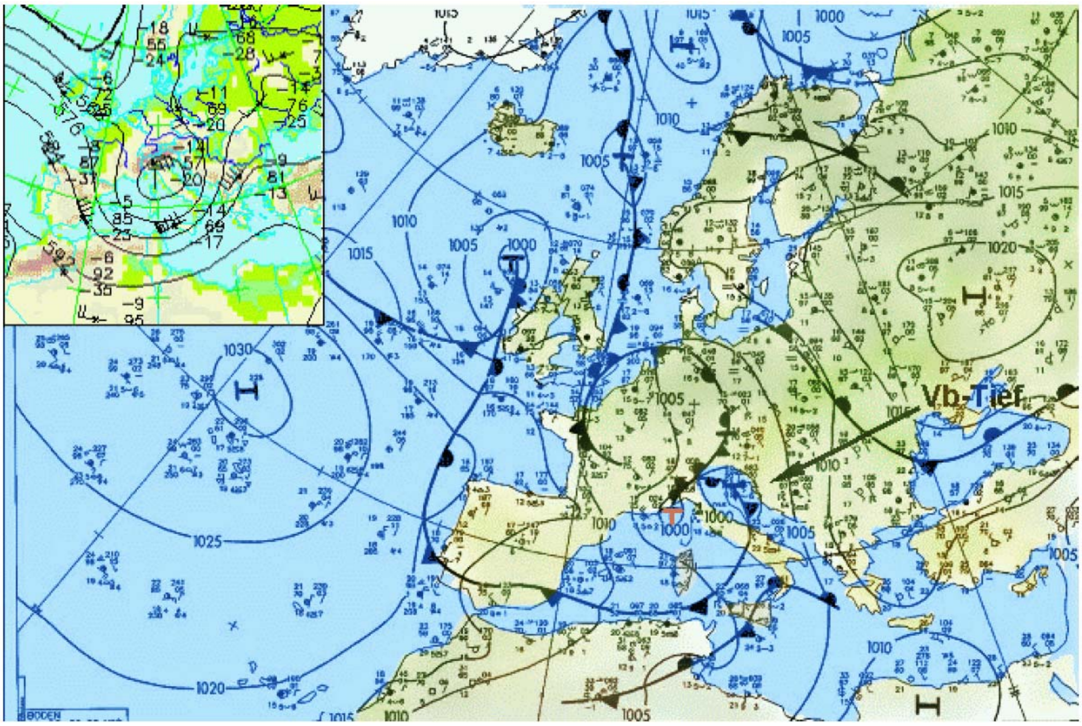

Fig. 1: DWD surface weather chart dated 11 August 2002, 00 UTC (large picture) and absolute topography 500 hPa dated 11 August 2002, 12 UTC (small picture) (DWD 2002).

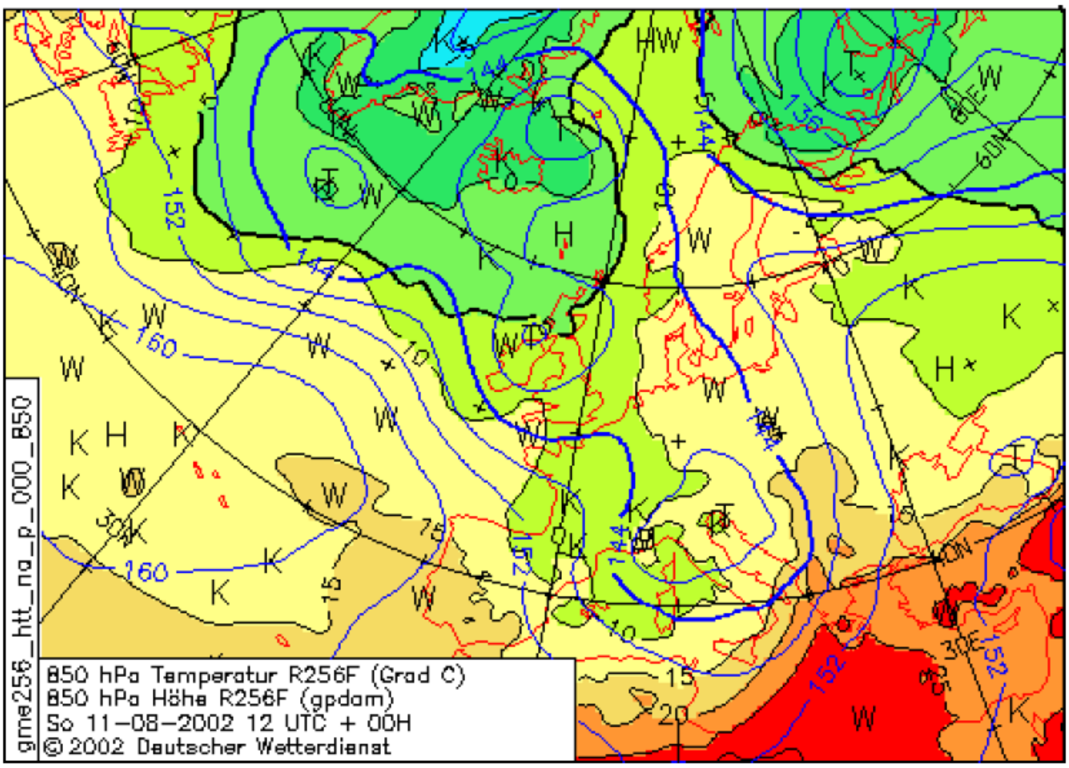

Fig. 2: Numerical analysis of the geopotential and temperature field at 850 hPa (corresponding to approximately 1.5 km height) on 11.08.2002, 12 UTC. The Vb depression forms over the northern Adriatic Sea (Fritzschner and Lux, 2002).

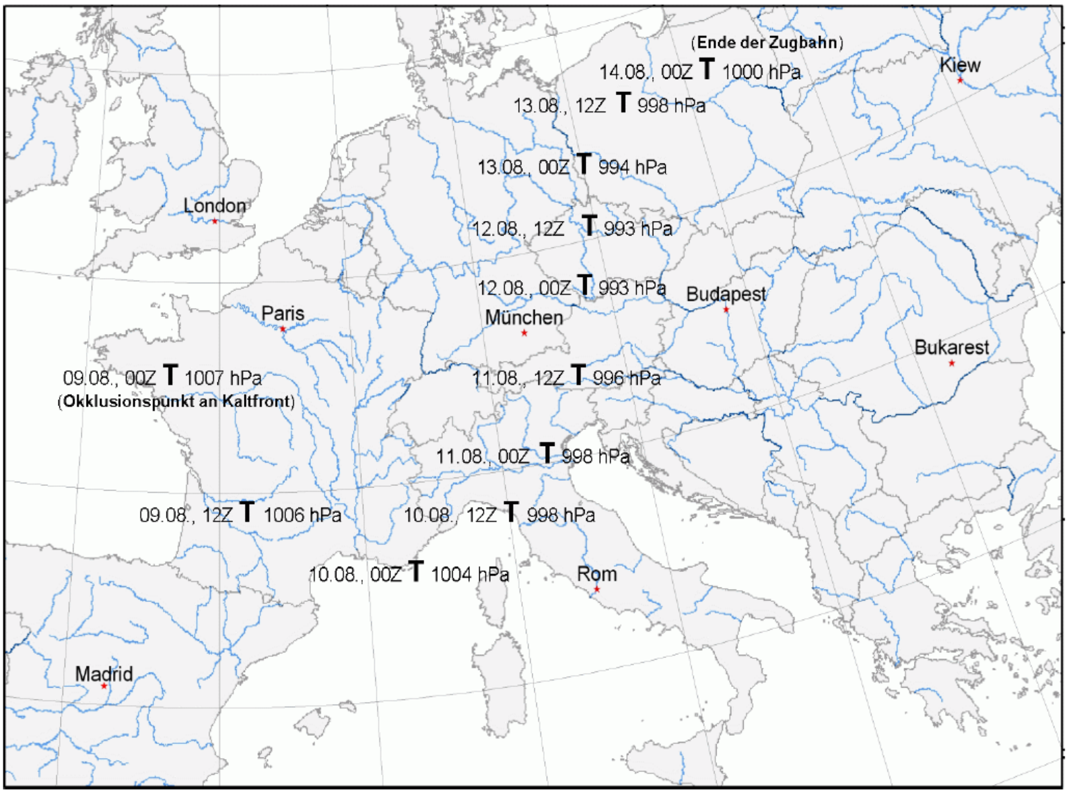

Fig. 3: Track of the “Vb-low” with indication of the core pressure (Rudolf and Rapp, 2003, map by M. Neumann, DWD).

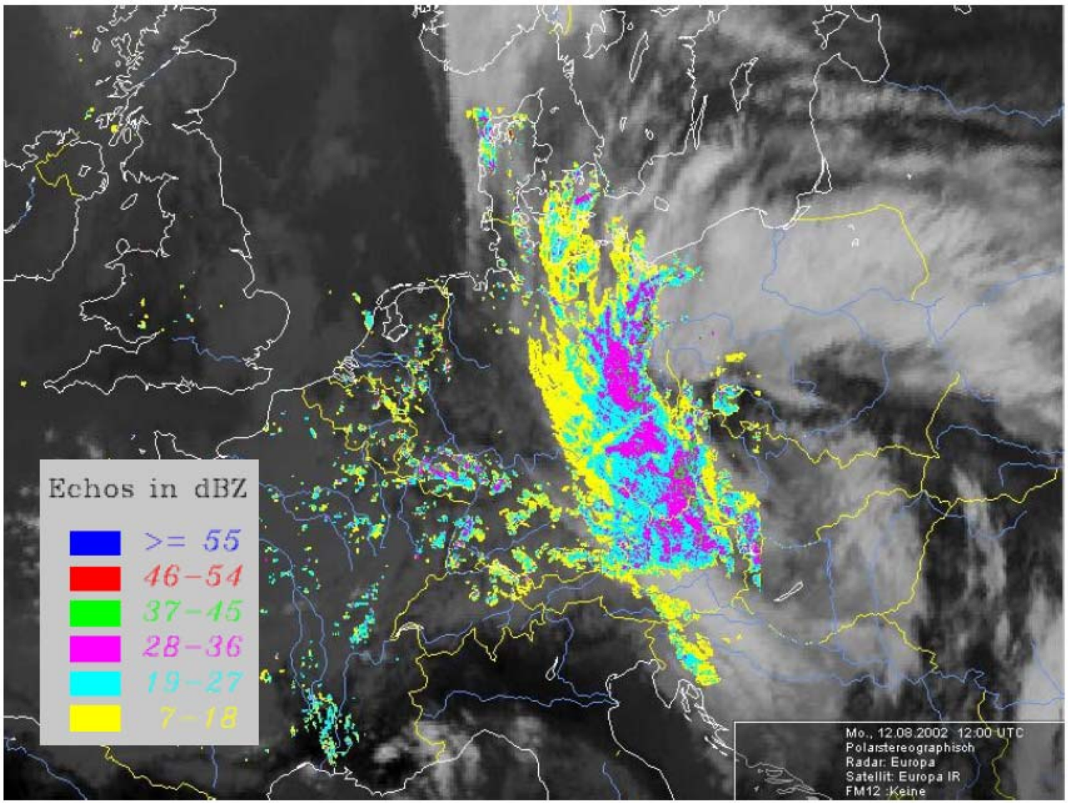

Fig. 4: Radar echoes and infrared satellite image, 12.08.2002, 12 UTC. Upper tropospheric warm air mass glides up on lower tropospheric cold air mass (DWD 2002).

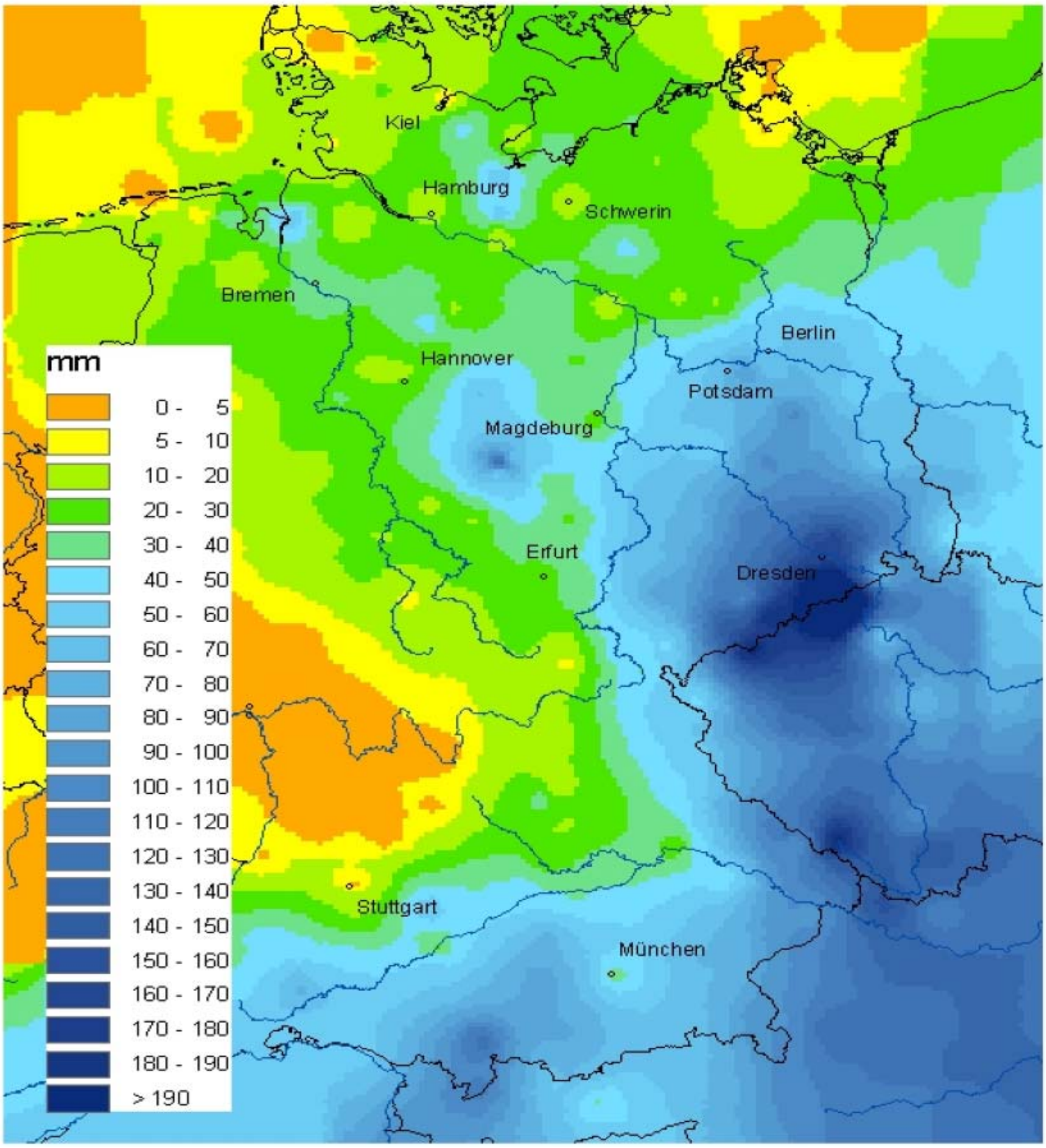

Fig. 5: Precipitation amounts from 10 August, 06 UTC to 13 August 2002, 06 UTC (Fritzschner and Lux, 2002).

## 3. Previous Investigations on the Overall Model Performance of Operational Runs

The operational rain prediction of the Deutscher Wetterdienst (DWD) using the Local Model (LM) gives some valuable hints to assess its suitability for application in the present study. For a detailed evaluation of the rain prediction capability of the model the reader is referred to the dedicated study on the Elbe flood event, performed by the DWD (DWD, 2002). From a previous verification study it was found, that the NWP model predicted alarming rain signals not before the 12 UTC run of 11.08.2002. At the beginning, the maximum rain sums were still predicted to occur over Czech Republic and Poland. In that case, the water runoff would have taken place via the rivers Neiße and Oder. Starting with the run on 12.08.2008, 00 UTC the NWP model delivered the best predictions with 24 h rain sums of more than 150 l/m² in the Ore Mountains (cf. Fig. 6). The comparison with observations reveals a qualitatively good model performance, although the observed rain sums were strongly underestimated by the model. As mentioned above, the rain amount is one of the most challenging meteorological parameters with respect to numerical prediction. The NWP performed by the DWD represents exemplarily the state-of-the-art in rain forecast. The identification of the reasons for the rain underestimation is beyond the scope of the present study. Here, we focus on the model sensitivity against changing boundary conditions.

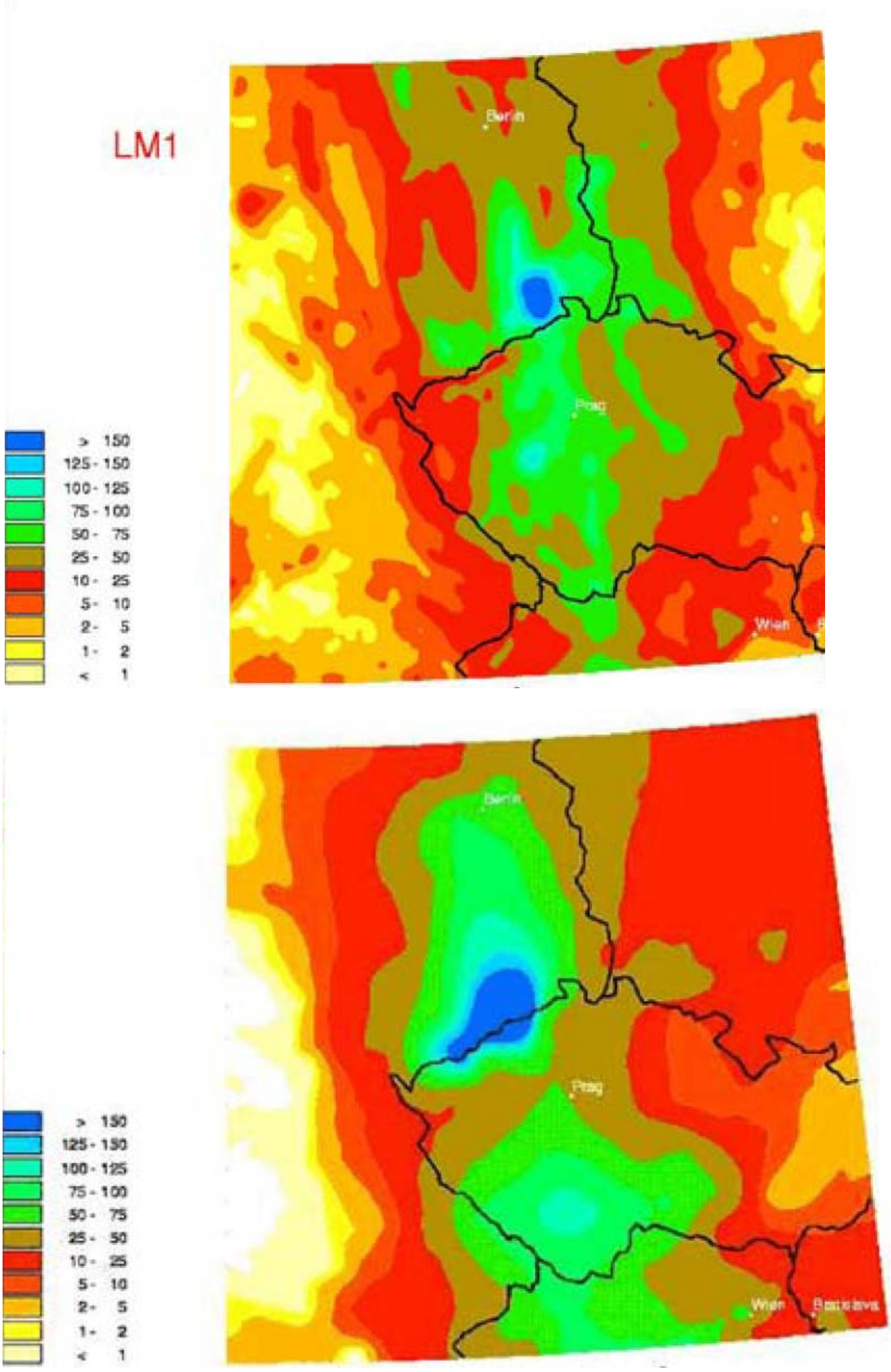

Fig. 6: Predictive capability of the Local Model. Top: Predicted 30 h precipitation sum, 12.08.2002, 00 UTC–13.08.2002, 06 UTC. Bottom: Observed 24 h precipitation sum, 12.08.2002, 06 UTC–13.08.2002, 06 UTC (Fritzschner and Lux, 2002).

## 4. Model Setup for Epignostic Studies

In the present study, the simulations were performed using the model COSMO-DE Version 4.2 (an advanced version of the former Local Model), which is a key element in the NWP model system of the DWD (Schättler and Doms, 2002). The model is executed with a horizontal grid resolution of $\Delta x \approx 2.8$ km (in geographical coordinates: $\Delta\lambda = 0.025°$). The number of vertical layers is 45. Other models of the DWD model family are the mesoscale model COSMO-EU (the same source code as COSMO-DE, but with $\Delta x \approx 7$ km horizontal grid resolution and partly other parameterisation schemata) and the global scale model GME. For an appropriate handling of convective processes, as appearing in the present synoptic situation, a high resolution is necessary. Thus, the COSMO-DE is the adequate choice.

The options for the physical parameterisations were the same, as the ones used in operative model runs by the DWD. As the COSMO-DE is a local scale model, it needs initial and boundary conditions, e.g., topography, land-use data, soil data, external forcing data from the surrounding atmosphere at its boundaries. Usually, these data are provided by corresponding pre-processors to generate the time-invariant boundary conditions, and by coarser models, covering a larger area to generate the time-dependent boundary conditions.

In the present study, the boundary conditions are taken from the LMQ data set, in which the COSMO-DE model domain is "embedded". The LMQ data are reanalysis data, purpose-generated by the DWD for simulations of the Elbe flood event for scientific re-evaluation, epignostic studies etc. This data can be considered as the best guess of the atmospheric conditions and, consequently, the most sophisticated model input for the time being.

At time intervals of one forecast hour the model values at the boundaries are fitted to the analysis data provided by the surrounding LMQ analysis field. In this way, the model is (at least at its boundaries) forced by the external, large scale evolution to remain close to the observation data.

The initial and boundary data for the COSMO-DE are calculated from the LMQ data with a special program pre-processor (called "int2lm", version 1.7). This program is also used by the DWD.

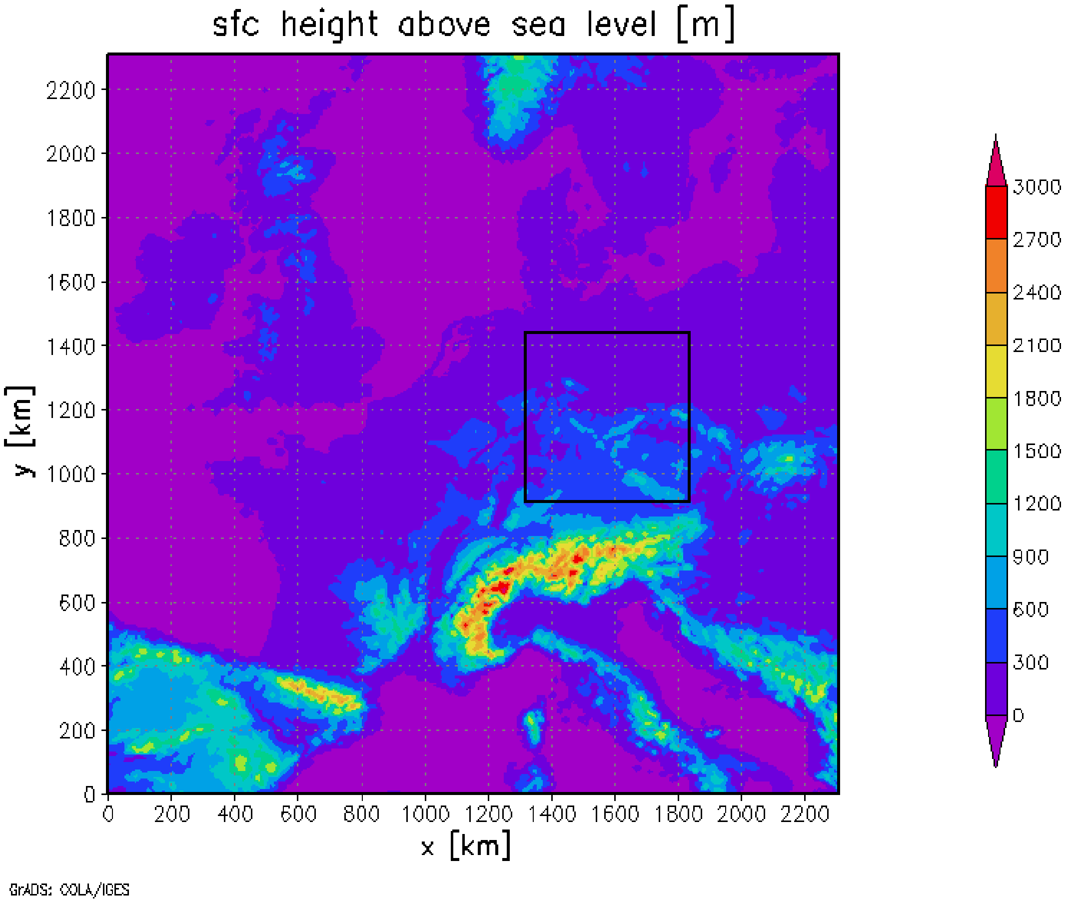

Fig. 7: Area, covered by LMQ data. The black quadrate shows the location and size of the highly resolved subdomain of 201 x 201 gridpoints, employed in the present study.

The LMQ data cover an area including most of Central and Western Europe (cf. Fig. 7). For the simulation with the COSMO-DE a model domain with a grid of 201 x 201 grid points was chosen. It covers the northern part of Central Europe with an area of approximately 565 x 565 km². The exact location of the grid is shown in Fig. 7 (black rectangle). The choice of this domain positioning is motivated by synoptic arguments. In the chosen constellation the surface low is situated at the right boarder of the model grid. Thus, the physical values at this sensible point are forced to remain at the prescribed boundary values (strict forcing). In this way one can avoid the surface low making “a life of its own” inside of the model domain more or less independent from the real large-scale evolution, which would in consequence misplace the precipitation fields. An eventual “a priori” misplacement of the precipitation field could not be “a posteriori” corrected by a regional area model, neither by a very high grid resolution nor by a sophisticated physical parameterisation. Here one has to consider, that the simulation of a Vb depression track is a multiscale phenomenon per excellance. Its spatio-temporal evolution is highly controlled by the superscale pre-conditioning of the atmosphere, impacting the cyclogenesis and energy uptake in the embryonic stage of the depression.

The starting point of the investigation is a reference run using the above mentioned model settings. No artificial data modification has been carried out. The model was initialised at August 12, 00 UTC. Model integration time was 30 hours. The predicted 24 h rain sum was evaluated in the time interval from 00+06 UTC until 00+30 UTC. This integration interval refers to the synoptic period, during which most of the precipitation was recorded. The resulting precipitation field is shown in Fig. 8.

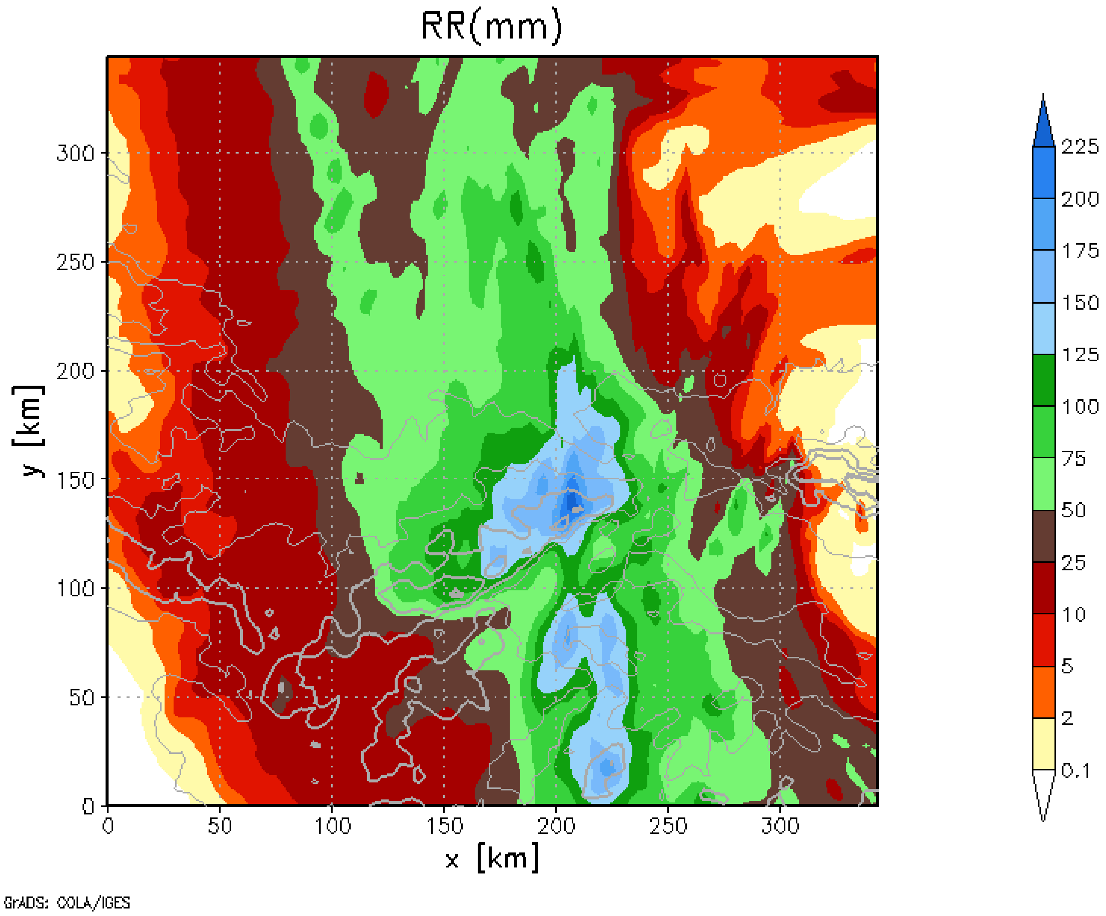

Fig. 8: Reference case (ID01): 24 h rain sum from COSMO-DE simulation, August 12, 2008, 00+06 UTC to 00+30 UTC (isolines refer to topography, Ore Mountains extending from [120,80] to [230,140]).

To get some feeling for the model sensitivity against the physical model setup, several experiments with other optional physical model parameterisations have been carried out, e.g., by activation of other turbulence schemes. However, as should have been expected the best result was obtained employing the operational model settings used in daily routine runs, which yield a precipitation maximum of 230-240 mm near the location of Zinnwald-Georgenfeld. Indeed, the highest rain sum was recorded just there. In general, the observed rainfall pattern - especially the location of the maximum - is fairly well reproduced, although the observed local maxima are underestimated by up to 75 mm (24 h rain sum in Zinnwald-Georgenfeld: 304 mm).

Figure 9 shows the accumulated rain sum measured at the meteorological station Zinnwald-Georgenfeld from 12.08.2002, 00 UTC to 13.08.2002, 06 UTC. In the first part of the considered time interval until around 12.08.2002, 15 UTC, the precipitation was mostly of convective origin (with a peak rain rate of 8.1 mm / 10 min), visible by the batch-wise accumulation of precipitation. Later on, the rainfall more and more loses the convective fractions and totally evolves into stratiform rain. The curve is getting smoother with rain rates between 1 mm / 10 min and 3 mm / 10 min. In comparison, Fig. 10 shows the predicted cumulated rain sum in Zinnwald-Georgenfeld from 12.08.2002, 00 UTC to 13.08.2002, 06 UTC, obtained by the reference run. Comparing the observed cumulated precipitation time series according to Fig. 9 with the predicted one in Fig. 10, one can see a large underestimation of the observed rain sum by the model. In such extraordinary situations deficits in the model performance become obvious. Both, stratiform as well as convective precipitation seem to be significantly underestimated.

Possible reasons might be: (a) deficits in the physical parameterisation (e.g., soil physics, boundary layer physics, cloud microphysics, convection), (b) insufficient resolution of topography and land use, (c) deficits in the numerical model. These influencing factors refer to both, the forcing and forced model. Owing the above-mentioned nonlinearity of the hydrothermodynamic system, small inaccuracies in the initial and boundary conditions as well as in the parameterisations can evolve into large deviations of the trajectory of the system in the phase space. However, considering the extreme atmospheric conditions in the present case the overall model performance is surprisingly good.

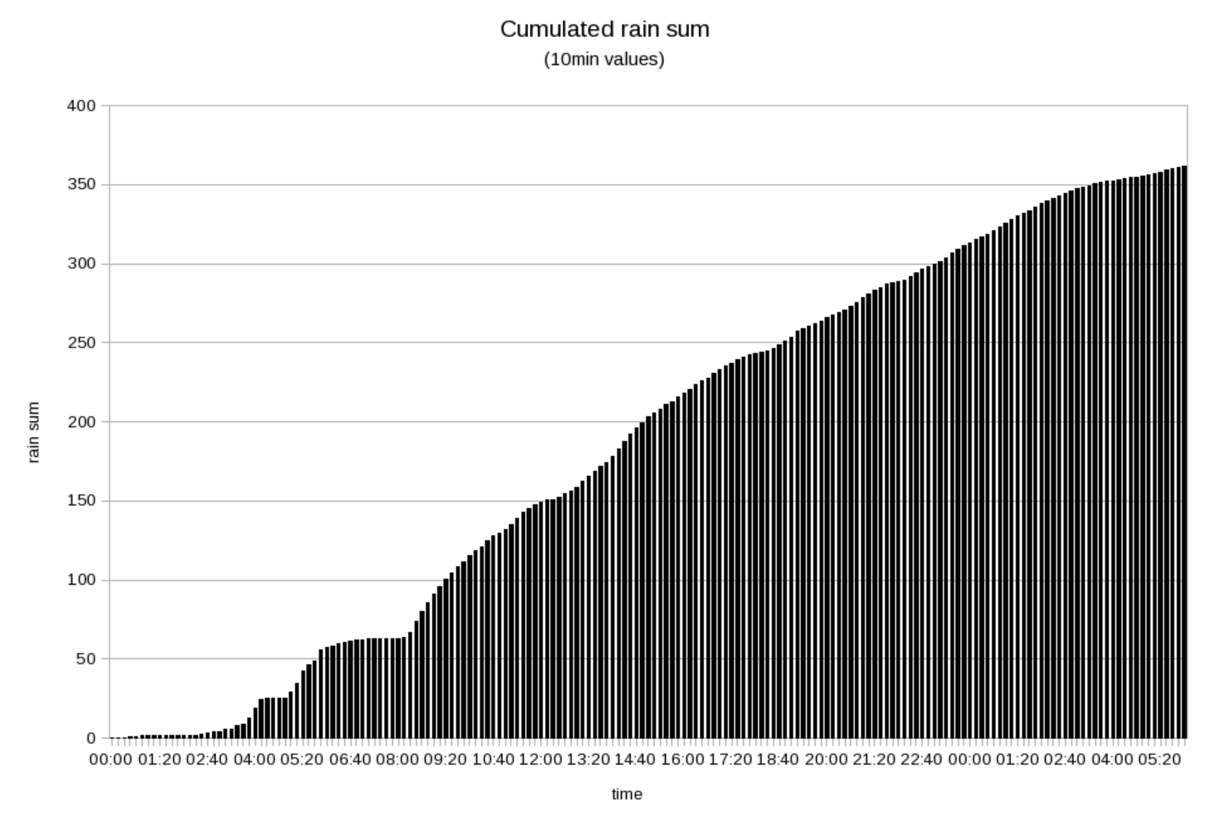

Fig. 9: Observed cumulated rain sum in Zinnwald-Georgenfeld from 12.08.2002, 00 UTC to 13.08.2002, 06 UTC.

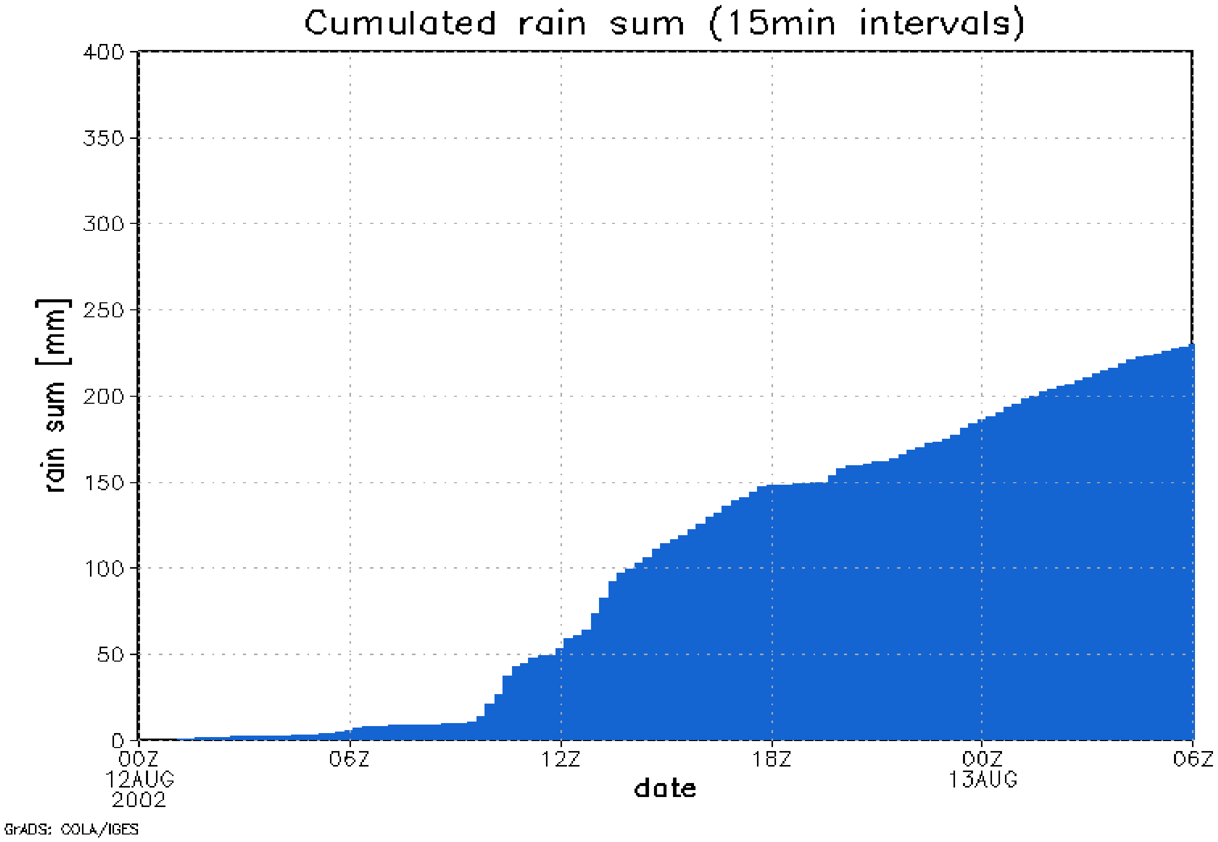

Fig. 10: Reference run: Predicted cumulated rain sum in Zinnwald-Georgenfeld from 12.08.2002, 00 UTC to 13.08.2002, 06 UTC.

# 5. Scenario Simulations

## 5.1 Scenario Definition and Setup of the Base Case

According to the directive for the simulations to the Elbe flood event under presumed future climatic conditions, offsets of temperature and humidity were added onto the initial and boundary fields. As a guideline, two scenarios with temperature offsets of $\Delta T$=+2 K and +4 K, in each case at constant relative humidity, have been prescribed. However, in the COSMO model the relevant predictive humidity parameter is the water vapour mixing ratio, denoted by $q_v$, rather than the relative humidity RH=$e/e_w(T)$ (here scaled between 0 and 1), with $e$ denoting the partial water vapour pressure and $e_w(T)$ denoting the saturation water vapour pressure, i.e., the partial water vapour pressure in equilibrium with bulk water reservoir having a flat surface. The change of the relative humidity depends on both, the change of the partial water vapour pressure and the change of the saturation vapour pressure:

$$\frac{1}{\mathrm{RH}}\frac{\mathrm{dRH}}{\mathrm{d}T}=\frac{1}{e}\frac{\mathrm{d}e}{\mathrm{d}T}-\frac{1}{e_W}\frac{\mathrm{d}e_W}{\mathrm{d}T}. \qquad (1)$$

For RH=const we obtain from Eq. (1) the following relation:

$$\frac{\mathrm{d}e}{\mathrm{d}T}=\mathrm{RH}\left(\frac{\mathrm{d}e_w}{\mathrm{d}T}\right)>0. \qquad (2)$$

As the saturation water vapour pressure increases with increasing temperature, the condition RH=const can only be realised by increasing the partial water vapour pressure according to Eq. (2). The term $\mathrm{d}e_w/\mathrm{d}T$ on the right-hand side of Eq. (2) is given by the Clausius-Clapeyron equation, which yields a well defined positive value. Hence, a temperature increase at constant relative humidity requires an increase in the partial water vapour pressure $e$, and via $q_v \approx 0.622\, e/p$ with $p$ being the air pressure, an increase in the water vapour mixing ratio. As the RH values vary along the model boundaries corresponding to the actual meteorological conditions, the fulfilment of the condition RH=const would require a water vapour pressure offset $\Delta e$ varying along the model boundary. However, small-scale variations of the absolute humidity cannot be motivated by large-scale climate change. From a meteorological point of view therefore it seems to be more reasonable to consider constant offsets of the respective partial water vapour pressure (or water vapour mixing ratio) rather than demanding a constant offset of the relative humidity. In the general case, the change of the relative humidity ΔRH depends on both the offsets of the water vapour mixing ratio $\Delta q_v$ and the temperature $\Delta T$:

$$\frac{\Delta \mathrm{RH}}{\mathrm{RH}} \approx \frac{\Delta q_V}{q_V} - \left( \frac{T}{e_W(T)} \frac{\mathrm{d}e_W}{\mathrm{d}T} \right) \frac{\Delta T}{T} . \tag{3}$$

In the present study, we vary the water vapour mixing ratio and the temperature independently with the aim to keep the changes of the relative humidity as small as possible.

Owing to the strong vertical decrease of the water vapour mixing ratio, height-independent $q_v$ offsets could easily result in unrealistically high water vapour supersaturations in the upper model layers (magnitude $q_v=10^{-1}$ g/kg there). To avoid unrealistical supersaturations, the water vapour mixing ratio offset is expressed in relative units $\Delta q_v/q_v$.

For $\Delta T$=+2K we performed runs with $\Delta q_v/q_v$=+10% and +15%, for $\Delta T$=+4K we chose $\Delta q_v/q_v$=+20% and +25%. These conditions ensure nearly constant values of the relative humidity (vide infra). The offsets are added onto the initial field and onto the boundary values in all model heights.

An example for the original (not modified by an offset value) humidity initial field is shown in Fig. 11. The field of the water vapour mixing ratio, modified by a constant offset (not shown here), reveals the same spatial structure. The vertical cross sections of $q_v$ and RH are depicted in Figs. 12 and 13. For comparison, in Fig. 14 the vertical cross section of relative humidity for $\Delta T$=+2K, $\Delta q_v/q_v$=+10% is displayed. Compared to the reference run (cf. Fig. 13), the relative humidity is slightly lower, because the relative humidity offset of 10% is below that required to yield a constant relative humidity at a temperature enhanced by 2 K.

Technically, the temperature offset is directly added to the initial and boundary values, whereas the same offset value is used over the whole area and in all levels. In this way, the temperature gradient in the atmosphere has not been changed.

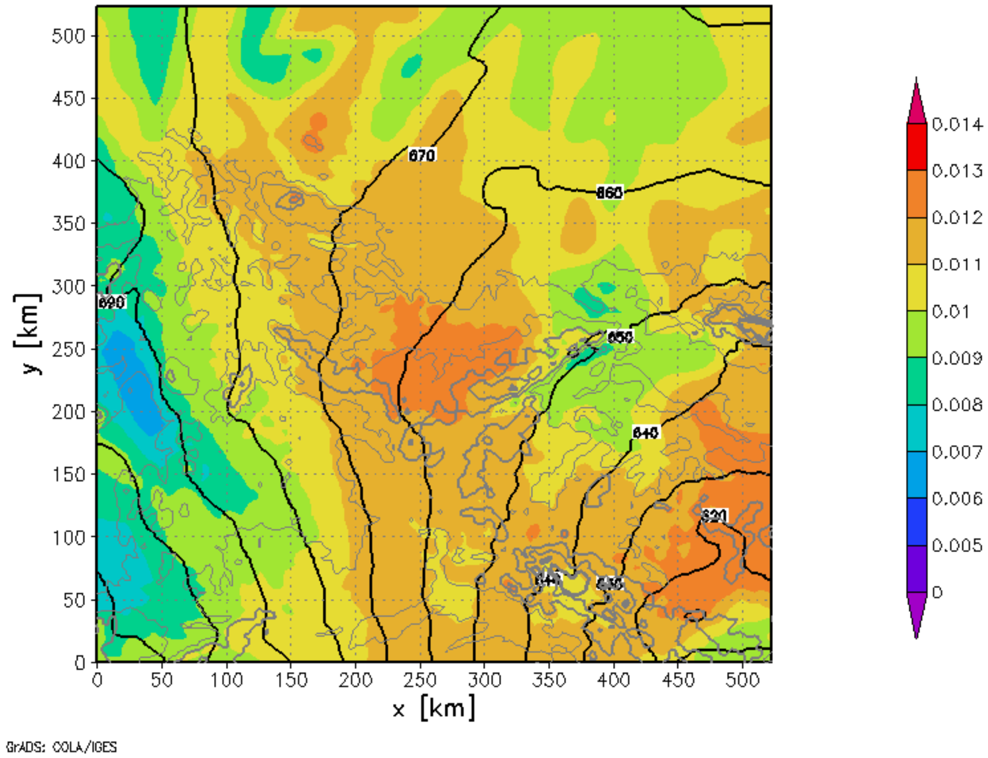

Fig. 11: Initial field of water vapour mixing ratio $q_v$ in [kg/kg] (colours) and the geopotential (isolines in gpdm) of the 925-hPa pressure level on 12.08.2008, 00 UTC (model domain: 201 x 201 grid points).

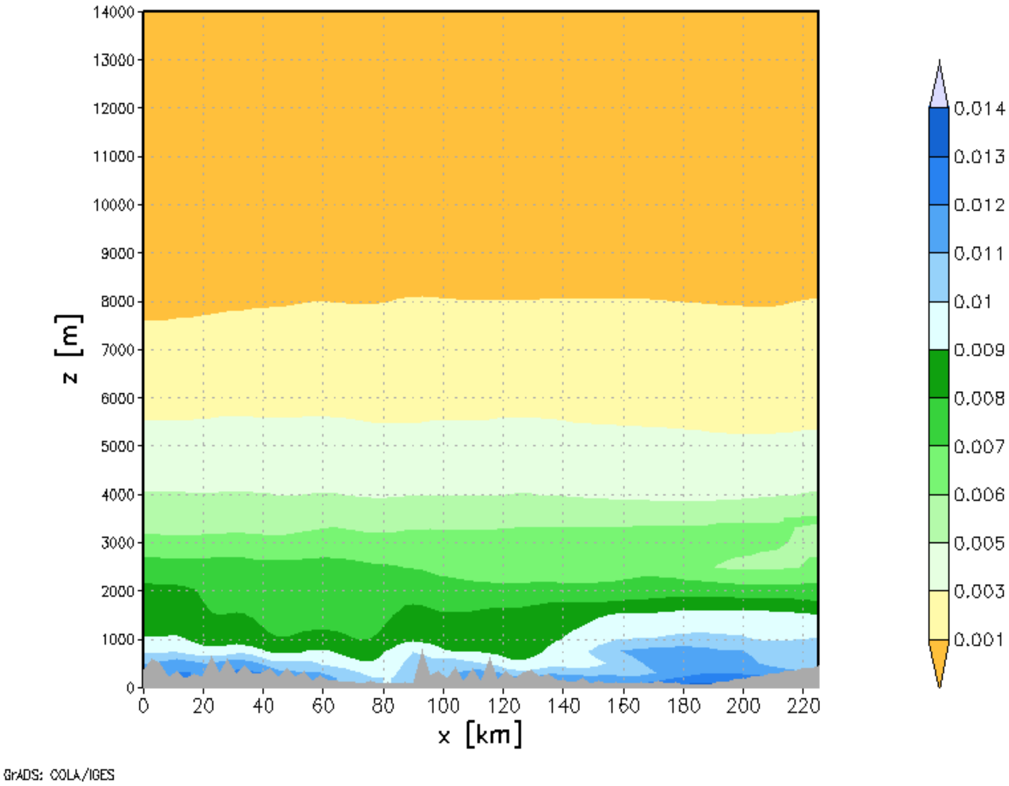

Fig. 12: Reference run (model domain: 81 x 81 grid points): Vertical cross section of $q_v$ in [kg/kg], 12.08.2008, 00 UTC.

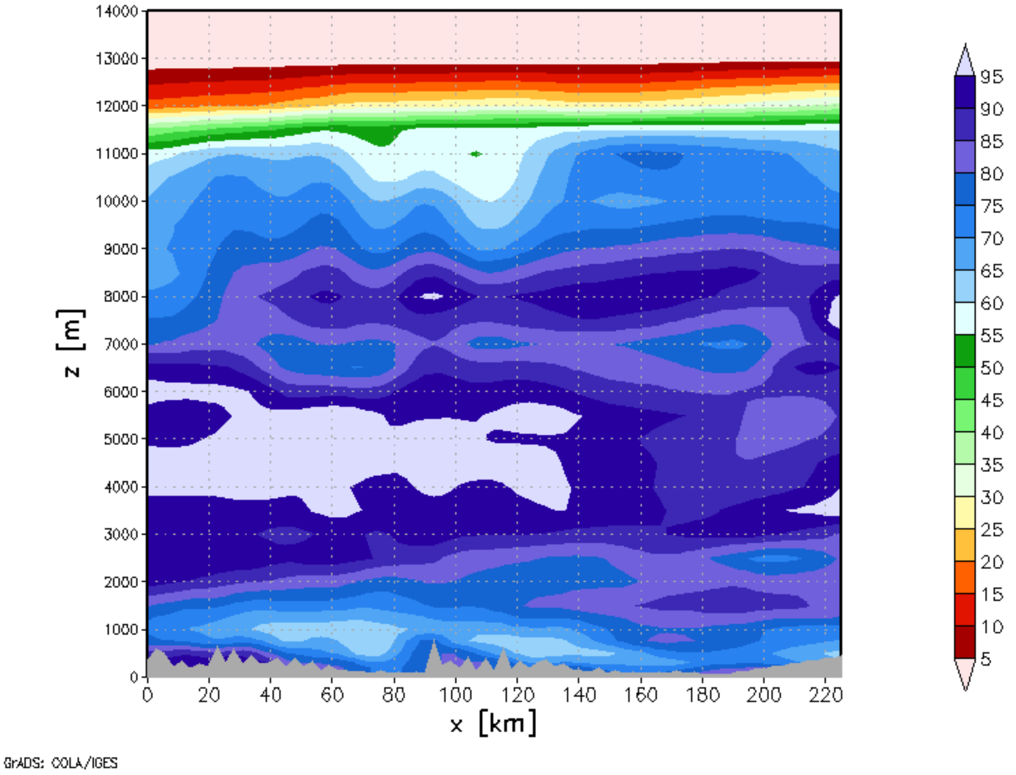

Fig. 13: Reference run (model domain: 81 x 81 grid points): Vertical cross section of relative humidity RH in [%], August 12, 2008, 00 UTC.

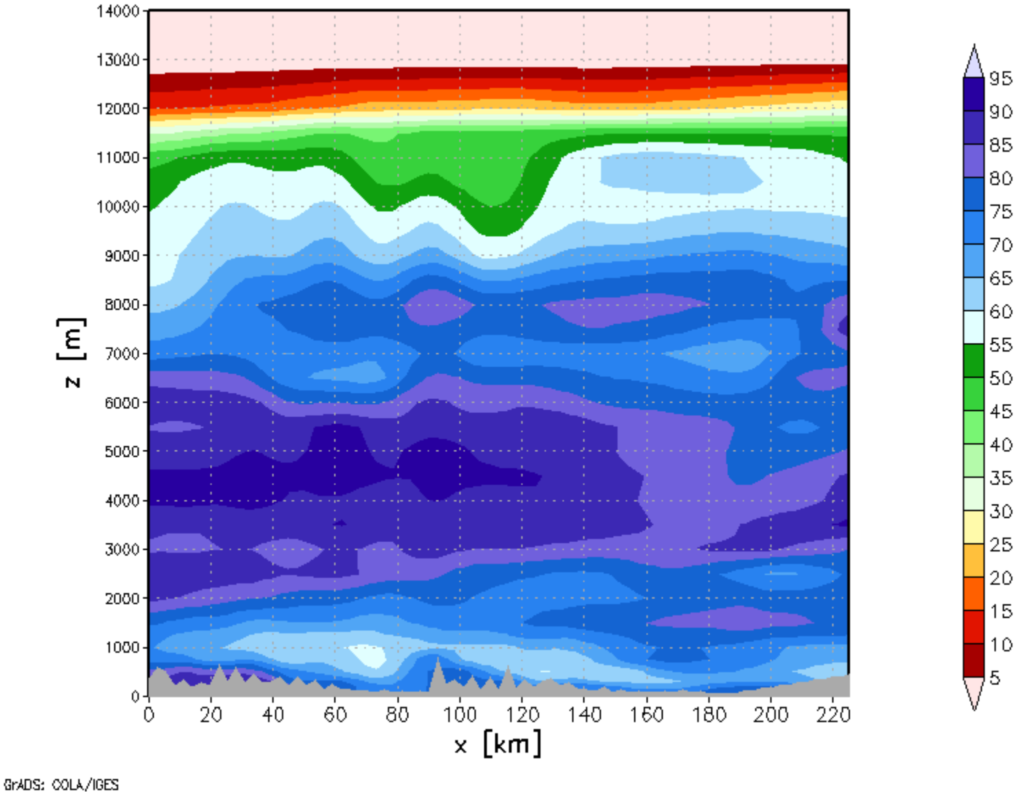

Fig. 14: Offset run with $\Delta T$=+2K, $\Delta q_v/q_v$=+10% (model domain: 81 x 81 grid points): Vertical cross section of relative humidity RH in [%], 12.08.2008, 00 UTC.

A general problem of using artificial initial and boundary conditions is, that the physical balancing of the meteorological fields at the initial state according to conservation laws is not ensured any longer. Furthermore, an artificial change of the initial and boundary conditions in the COSMO-DE model decouples the mesoscale future of the system from its large scale history. Climate change should affect the atmospheric pre-conditioning for cyclogenesis in general, i.e., the system evolution in its embryonic stage but not only in its mature stage, as considered here. Therefore, the choice of artificial initial and boundary data is justified only insofar as the model sensitivity against changes in the initial and boundary conditions is addressed. Thus, the epignostic simulation of an already happened event under artificially changed mesocale conditions allows, strictly spoken, only conclusions with respect to the *model* behaviour but not with respect to the *cyclone* behaviour in a future climate.

However, in our experiments, we use arithmetic offsets of the two predictive variables temperature and water vapour mixing ratio, while the other variables are kept constant. This should provide at least some tentative results regarding possible changes of atmospheric pre-conditioning for severe weather events.

### 5.2 Simulations for Different Thermo-Humid Conditions

The first set of calculations has been performed using the above described domain size of 201 x 201 grid points. To get an overview of the impacts of temperature and $q_v$ offsets of the initial and boundary data, simulations with different magnitudes of the offsets have been performed. In Tab. 1 the results of these simulations are shown.

Not in every run the peak value of precipitation sums is located in the Ore Mountains. Therefore, in column 4 of Tab. 1, the location of the rain sum maximum is denoted in brackets. In column 6 (change in rain sum over total model domain), the relative change of the precipitation sum integrated over all grid points of the model domain is shown.

Tab. 1: Maximum local 24h rain sums in the whole model domain and in the Ore Mountains region in dependence of constant offsets of temperature and water vapour mixing ratio.

| Run ID | Temperature offset [K] | Specific humidity offset Relative units [%] | Max. local 24h-rain sum in model domain [mm] | Max. local 24h-rain sum in Erzgebirge region [mm] | Change in rain sum over total model domain Relative units [%] |
|---|---|---|---|---|---|
| **ID01(*)** | **0** | **0** | **230-240 (Erzgebirge)** | **230-240** | **-** |
| ID02 | +2 | 0 | 180-190 (Erzgebirge) | 180-190 | -23.6 |
| **ID03(**)** | **+2** | **+10** | **190-200 (Erzgebirge)** | **190-200** | **-4.4** |
| **ID04(**)** | **+2** | **+15** | **190-200 (Erzgebirge)** | **190-200** | **+4.3** |
| ID05 | +2 | +20 | 240-250 (Brandenburg) | 190-200 | +11.7 |
| ID06 | +2 | +40 | >450 (Brandenburg) | 190-200 | +28.4 |
| **ID07(**)** | **+4** | **+20** | **210-220 (Böhmerwald)** | **170-180** | **-11.4** |
| **ID08(**)** | **+4** | **+25** | **250-270 (Erzgebirge)** | **250-270** | **-1.7** |
| ID09 | +4 | +30 | 230-240 (Brandenburg) | 160-170 | +5.6 |
| ID10 | +4 | +40 | 330-350 (Brandenburg) | 200-210 | +17.6 |

(*) This run represents the “undisturbed” reference case.
(**) These runs correspond to nearly constant relative humidity.

The results of this runs can be summarised as follows (cf. Tab. 1, Figs. 8, 15 and 16):

- The observed rainfall pattern, especially the location of the maximum is well reproduced by the reference case simulation (cf. Fig. 8). However, the observed rain amounts were underestimated in all regions up to a maximum deviation of 130 mm.
- Owing to the character of the underlying physical equations, the model system does not respond linearly to a change of the initial and boundary conditions. In dependence of the corresponding change of the storage capacity for atmospheric water vapour, constant offsets of temperature and humidity can decrease or enhance the resulting rain amount. In one case this results in a superposition, in another case in a displacement or splitting of dynamic and orographic lifting zones, the latter leading to the weakening of rain maxima and the appearance of new maxima.
- A temperature increase leads to a decrease of the total rain sum due to an enhanced storage capacity for atmospheric water vapour (increased equilibrium water vapour pressure), ensuring longer lasting sub-saturated conditions, which results in a higher lifting condensation level (cf. Tab. 1, ID02). This case corresponds to an effective decrease of the relative humidity.
- The simulations with nearly constant relative humidity provide nearly the same (cf. Tab. 1, ID03/04) or a small decreasing (ID07/08) amount of overall precipitation, respectively. The simulations insinuate a tendency of an overall

enhancement of the rain sum for an increase of the water vapour mixing ratio above a certain threshold. Therefore, the relative humidity must have exceeded the value of the reference run.

- An alteration of the thermo-humid initial and boundary conditions is accompanied by a change of the flow conditions. This may result in a change of the location and intensity of hydro-thermodynamic structures such as fronts, air mass internal convergence zones, convective updrafts etc.
- The runs referring to nearly constant relative humidity conditions (cf. Tab. 1, cases (**)) result in only minor changes in the total area accumulated rain sum. It is suspected, that the local decrease of the rain sum over the Ore Mountains (due to displacement between hydro-thermodynamic and orographic lifting zones) is partly compensated by the general higher availability of atmospheric water vapour for rain formation by phase transition.

Summing up, offsets of temperature and water vapour mixing ratio did neither lead to an increasing precipitation maximum nor to an increasing total area accumulated rain sum. The rain prediction is much less affected by the initial and boundary conditions as we expected. The offsets apparently lead to a disturbance of perfectly arranged initial fields, which influences the synoptic circulation.

The water vapour mixing ratio usually decreases rapidly with height. So if we add a relative offset $\Delta q_v/q_v$, the vertical gradient of the water vapour mixing ratio and consequently, the static stability will be changed. Such a change may lead to a re-constellation of frontal, orographic and convective processes, i.e., original superposition can be changed into split-position, which would result in a fairly different precipitation pattern.

Another change may happen due to rising temperatures. As a result, the pressure decreases throughout the model domain, while the spatial features of the pressure field are not essentially affected.

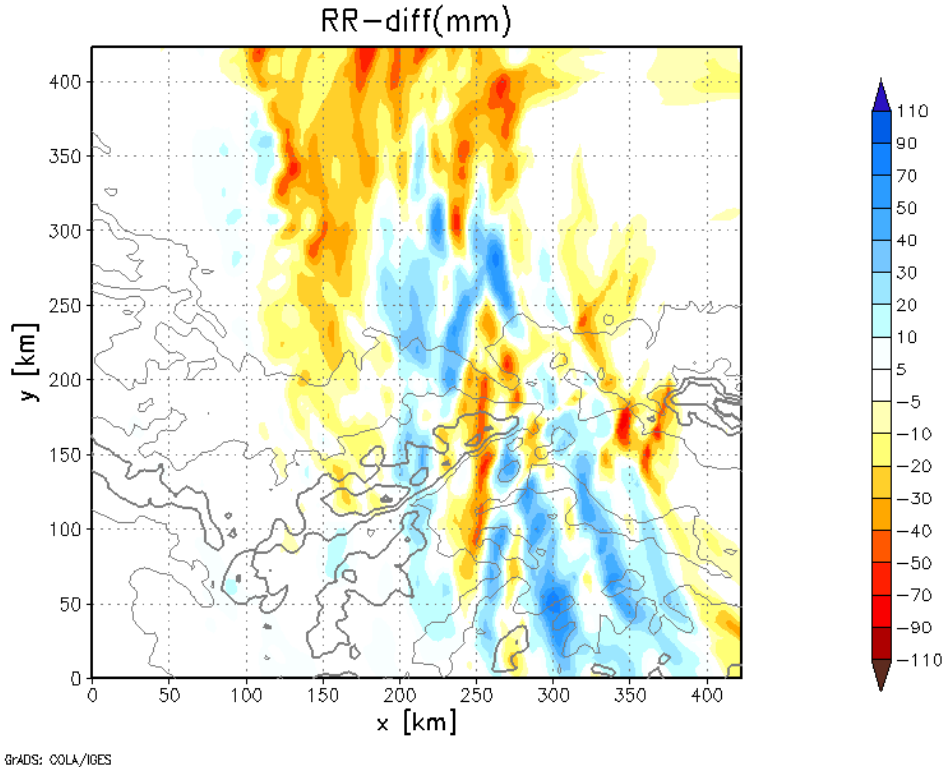

Fig. 15: Case ID03: Difference “ID03 minus ID01” of 24 h rain sum from COSMO-DE simulation, August 12, 2008, 00+06 UTC to 00+30 UTC, temperature offset $\Delta T$=+2 K.

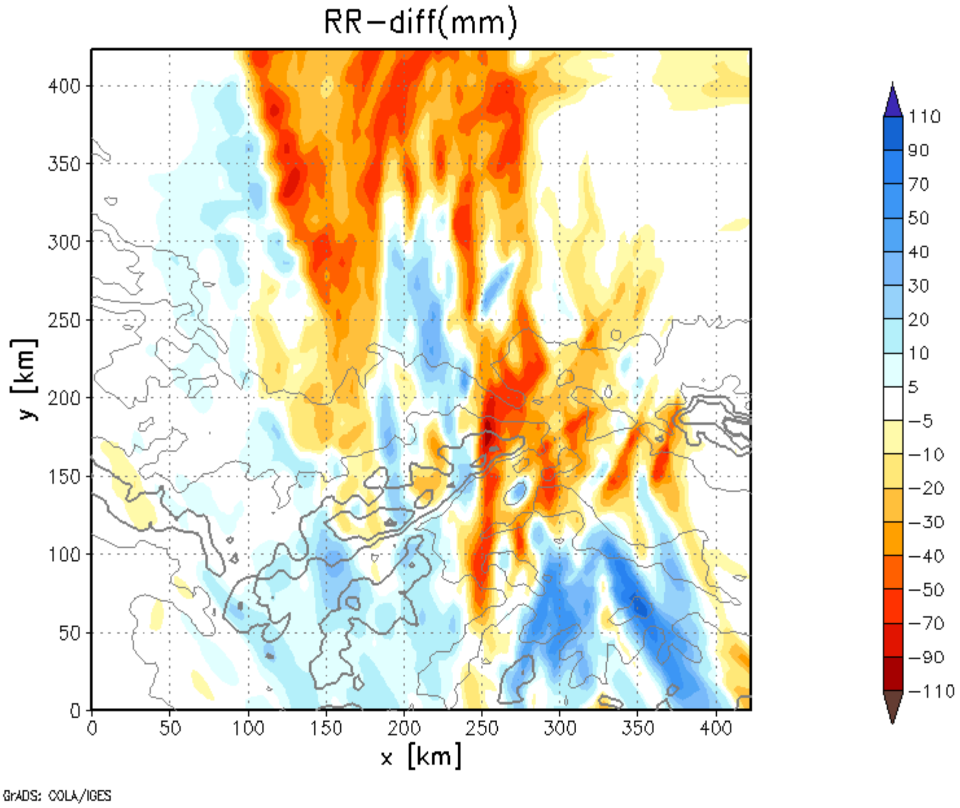

Fig. 16: Case ID07: Difference “ID07 minus ID01” of 24 h rain sum from COSMO-DE simulation, August 12, 2008, 00+06 UTC to 00+30 UTC, temperature offset $\Delta T$=+4 K.

## 5.3 Scenario Simulations Employing Different Sizes of the Model Domain

In further experiments, the size of the model domain was reduced to enhance the strength of the large scale forcing onto the cloud vortex via the boundary conditions. This measure was suspected to maintain the superposition of precipitation generating processes, which essentially led to the enormous rain sums observed in the period of interest.

As mentioned above, the boundary fields are analysis data. This forces the model predictions near the boundaries to remain closely to the observed meteorological evolution. By decreasing the size of the model domain we expect an enhancement of the strength of the external forcing onto the evolution in the interior of the domain. Thus, the observed superposition of the rain-generating processes over the Ore Mountains is supposed to be better reproduced by the model. The larger the domain size, the more the model physics inside the domain is able to “make a life of its own”, and vice versa, the smaller the domain size, the more the model is forced to “make a prescribed life”.

For the smaller domain a grid size of 81 x 81 grid points was chosen, corresponding to a domain area of 222 x 222 km². It is placed in such a way, that the frontal lifting zone is located in the centre of the domain. The other model settings remain the same as before.

The results can be summarised as follows (cf. Tab. 2, Figs. 17-19):

- Using a smaller model domain, a new conspicuous feature in the spatial precipitation field appears. In all subsequent runs, an area with high precipitation values in the region southwest of the Lusatia Mountains (Lausitzer Gebirge) evolves. An accompanying analysis of other predictive meteorological elements revealed, that this rainfall area originates from the development of a strong convergence zone in the early stage of the modelled evolution. Thus, this precipitation is mainly of convective origin, resulting from re-configuration of the model domain under otherwise unchanged conditions. Consequently, the predicted local rain maxima are nearly completely located within the convective zone. This phenomenon is a strong hint toward the “butterfly effect”.
- The mentioned phenomenon can be clearly seen in the reference run employing the reduced domain (cf. Fig. 17). The model does not reproduce

the observed rainfall pattern as good as the reference run for the larger domain (201 x 201 grid points).

- Anyway, the precipitation maximum in the Ore Mountains is located in the Zinnwald-Georgenfeld region with a peak value of around 200-210 mm, corresponding to an underestimation of up to 160 mm with respect to the observation.
- As in experiment series employing the larger model domain, a temperature increase leads to a decrease of the total rain sum due to an enhanced storage capacity of the atmosphere for water vapour, ensuring longer lasting subsaturated conditions (cf. Tab. 2, ID12). The above described convective line in the area of the Czech Republic is much less affected by the decreasing precipitation.
- For the decreased model domain large amounts of precipitation originate from the synoptic convergence. For small and moderate humidity offsets the rain sums originating from the associated convective lifting zone are higher than the ones originating from the orographic lifting zone of the Ore Mountains. However, with an increasing offset of the water vapour mixing ratio, the lifting condensation level decreases (LCL). A lower LCL leads to an earlier onset of condensation and, consequently, to more effective rain formation over the Ore Mountains under otherwise unchanged hydrothermodynamic flow conditions. Consequently, an increase of the water vapour mixing ratio leads to a higher contribution of orographically induced precipitation to the total rain sum. Convective precipitation is not affected that strongly by this fact, because it originates from thermal updrafts rather than from dynamic lifting.
- Via the boundary conditions, the frontal lifting zone is forced to remain in the correct (observed) position to the Ore Mountains. In contrast to this, the strength of the convective line south of the Ore Mountains is strongly overestimated, while most of the embedded convective parts of the precipitation in Zinnwald-Georgenfeld are missing. This is confirmed by the simulated radar animations (not presented here) and by the deficit in the cumulated rain sum in the early part of the time series for Zinnwald-Georgenfeld (cf. Fig. 20).
- In general, the tendencies for the cumulated rain sums confirm the simulation results employing the 201 x 201 grid. Thus, while the spatial distribution of the

rain within the model domain is very sensitive against the size of the model, the cumulated rain sums are not in equal measures.

Tab. 2: Maximum local 24h rain sums in the whole model domain and in the Ore Mountains region in dependence of constant offsets of temperature and water vapour mixing ratio.

| Run ID | Temperature offset [K] | Specific humidity offset Relative units [%] | Max. local 24h-rain sum in model domain [mm] | Max. local 24h-rain sum in Erzgebirge region [mm] | Change in rain sum over total model domain Relative units [%] |
|---|---|---|---|---|---|
| 81x81 | | | | | |
| **ID11(*)** | **0** | **0** | **220-230 (conv)** | **200-210** | - |
| ID12 | +2 | 0 | 200-210 (conv) | 110-120 | -46,7 |
| **ID13(**)** | **+2** | **+10** | **220-230 (conv)** | **190-200** | **-11,5** |
| **ID14(**)** | **+2** | **+15** | **250-270 (Erzgebirge)** | **250-270** | **+9.7** |
| ID15 | +2 | +20 | 290-310 (conv) | 270-290 | +36.2 |
| ID16 | +2 | +40 | >490 (northern+western saxony) | 450-470 | +133.7 |
| **ID17(**)** | **+4** | **+20** | **230-240 (conv)** | **180-190** | **-21,4** |
| **ID18(**)** | **+4** | **+25** | **250-270 (conv)** | **220-230** | **+0.6** |
| ID19 | +4 | +30 | 310-330 (conv) | 270-290 | +27.5 |
| 89x89 | | | | | |
| **ID20(*)** | **0** | **0** | **200-210 (conv+Erzgebirge)** | **200-210** | - |
| **ID21(**)** | **+2** | **+10** | **230-240 (conv)** | **220-230** | -6,9 |
| **ID22(**)** | **+4** | **+20** | **240-250 (conv)** | **180-190** | -15,6 |

(*) This runs represent the “undisturbed” reference case.
(**) These runs correspond to nearly constant relative humidity.
(conv) Meaning: “located in the convergence line”.

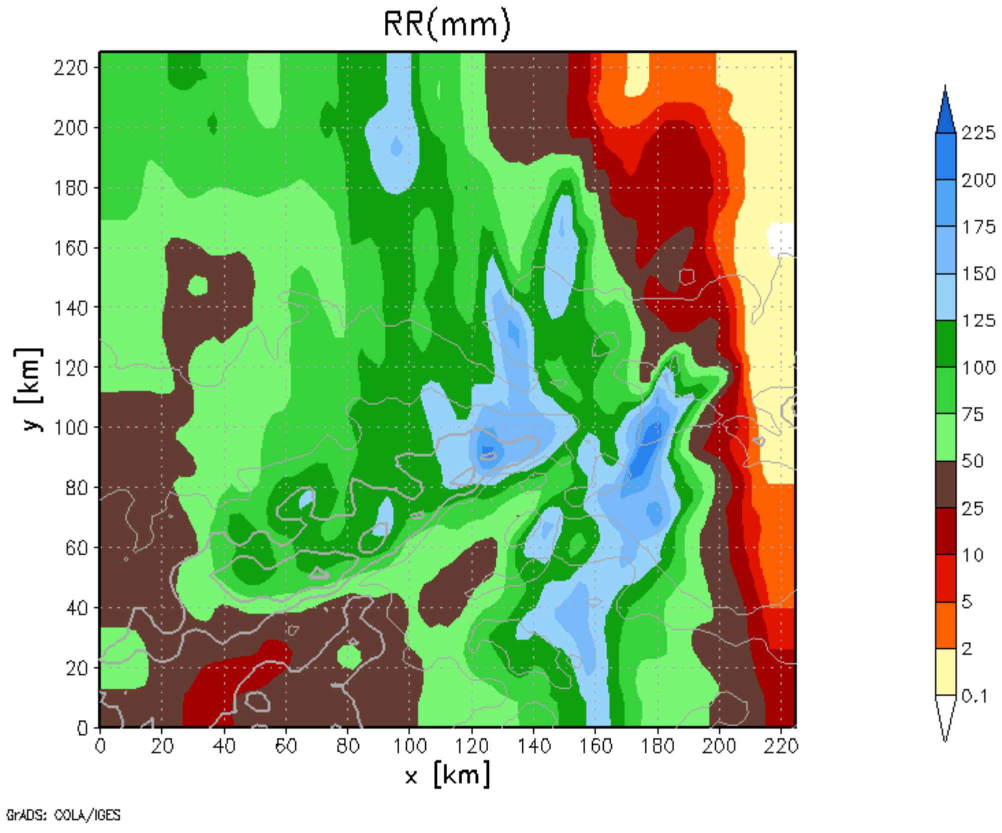

Fig. 17: Reference case (ID11) (model domain: 81 x 81 grid points): 24 h rain sum from COSMO-DE simulation, August 12, 2008, 00+06 UTC to 00+30 UTC.

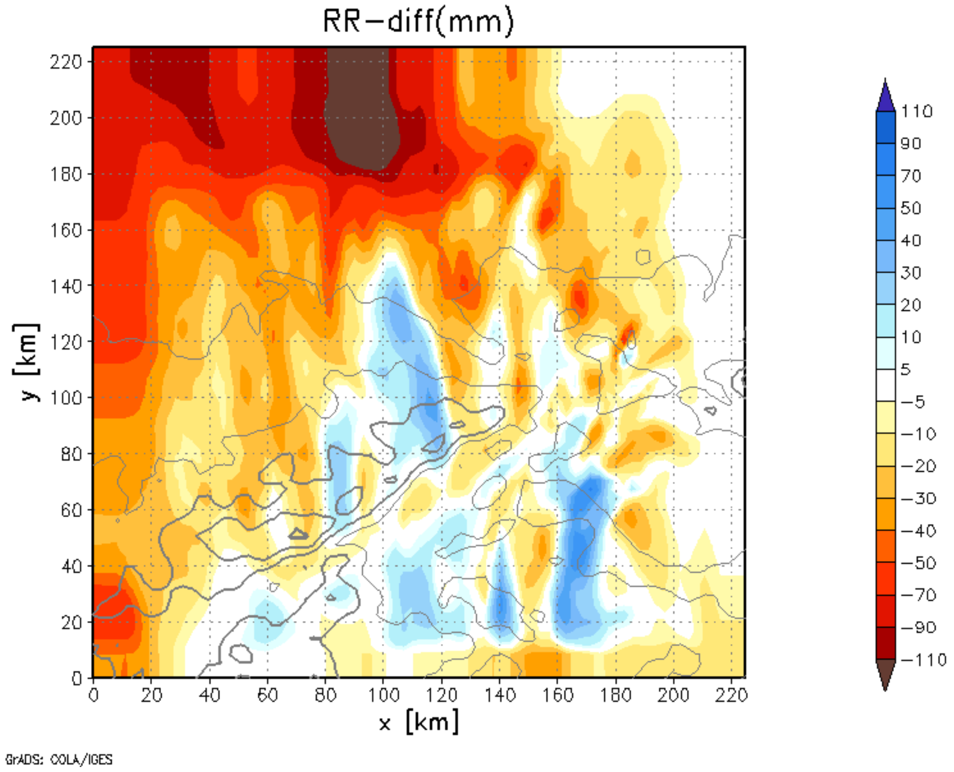

Fig. 18: Case ID13 (model domain: 81 x 81 grid points): Difference “ID13 minus ID11”of 24 h rain sum from COSMO-DE simulation, August 12, 2008, 00+06 UTC to 00+30 UTC, temperature offset ∆*T*=+2 K.

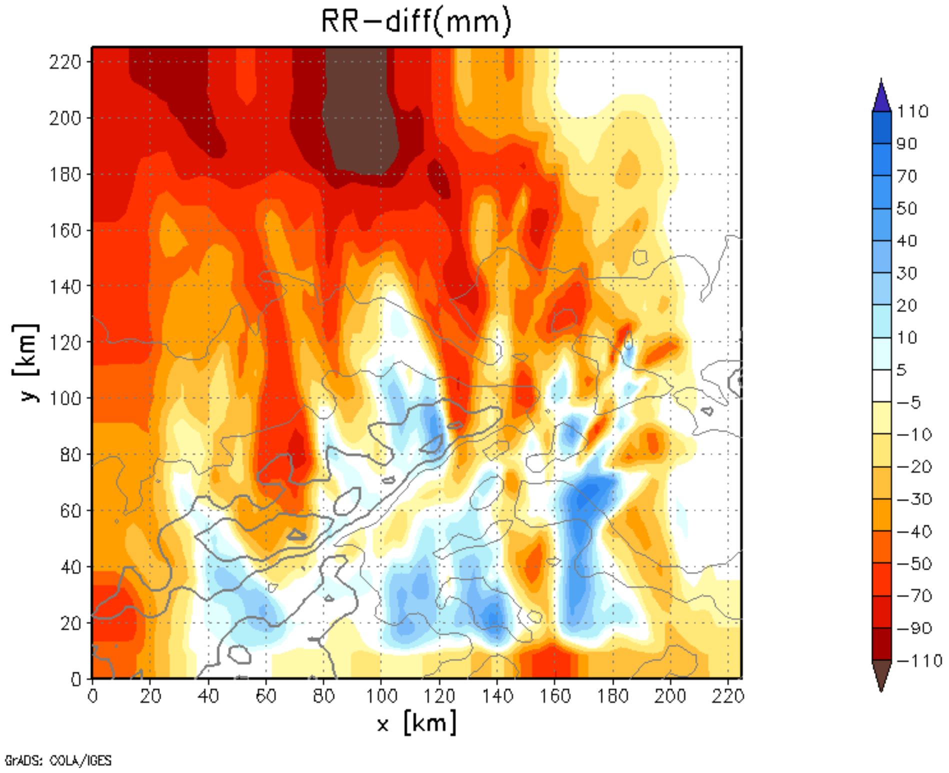

Fig. 19: Case ID17 (model domain: 81 x 81 grid points): Difference “ID17 minus ID11”of 24 h rain sum from COSMO-DE simulation, August 12, 2008, 00+06 UTC to 00+30 UTC, temperature offset ∆*T*=+4 K.

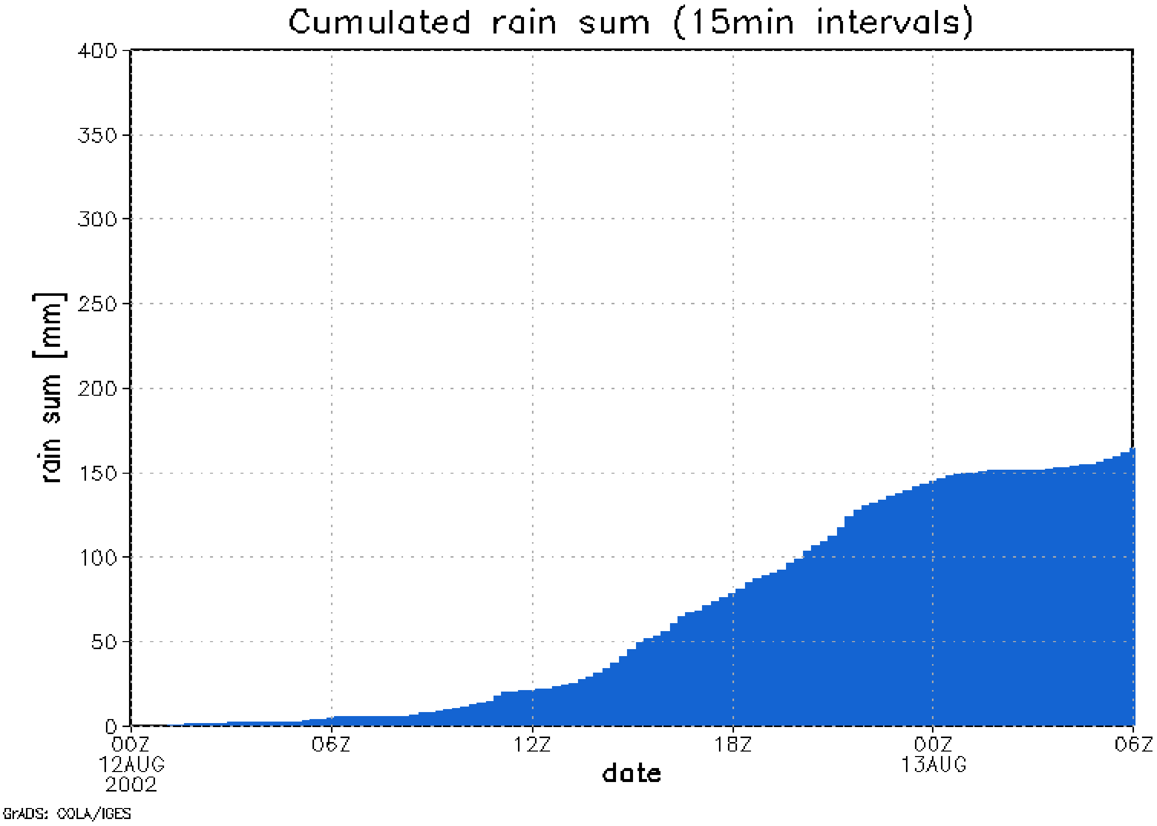

Fig. 20: Reference run (model domain 81 x 81 grid points): Cumulated rain sum in Zinnwald-Georgenfeld from 12.08.2002 00UTC to 13.08.2002 06UTC.

The use of artificial humidity and temperature offsets leads to an imbalance of meteorological fields, which can be seen, e.g., from the displacement of the geopotential field with respect to the reference field. This shift originates from an artificial thermic pressure depression evolving inside of the model domain. However, the deformation of the geopotential field is restricted to the two outermost grid points. The model domain of 201 x 201 grid points is large enough to compensate the imbalance effect, i.e., the impact of the imbalance on the rain evolution is negligible. However, in the model domain of 81 x 81 grid points, the boundary conditions may have a higher influence on the meteorological evolution in the inner model domain.

The shift in the geopotential at the boundaries influences the vertical wind velocity and, consequently, the precipitation formation. In the present case, the geopotential disturbance leads to a local decrease of the rain sum. Owing to the northern winds, this rain anomaly propagates from the northern model boundary southward. This effect can be seen in Figs. 18 and 19. Accordingly, at the northern boundary of the model domain the rain sums noticeably decrease in a broad band, while at the other boundaries the decrease occurs in smaller bands.

## 5.4 On the Impact of the Model Domain Boundaries on the Area Accumulated Rain Sum

To consider the simulated artifical rain gaps in determining the total area accumulated rain sum, as described in Section 5.3, the size of the model domain was extended to 89 x 89 grid points, while the rain integration area, embedded in the domain size, was reduced. (While the domain size corresponds to the numerical grid on which the physical equations are solved, the rain integration area denotes a subregion of the model domain, over which the rain sum is accumulated. In this way, the impact of the boundaries onto the rain sums can be reduced to some degree.) At the northern boundary a gap of 24 grid points (corresponding to 66.6 km), at the remaining boundaries gaps of 8 grid points (corresponding to 22.2 km) was excluded from the model domain, i.e., the size of the rain integration area was reduced to 49 x 65 grid points. For comparison, also runs employing a rain integration area of 89 x 89 grid points were performed. Here, the inner 49 x 65 grid points were retained at their locations in the foregoing run, while the surrounding model domain was extended northward to prolongate the rain evolution time downstream the inflow boundary, and eastward to weaken the overestimated convergence zone.

As can be seen from Tab. 2 and Figs. 21-23, the results do not differ much from the calculations employing the 81 x 81 domain size. The decrease of the integrated rain sums for the different scenarios is somewhat lower, but the tendency remains the same. Thus, the simulations employing the 89 x 89 domain size confirm the rain sensitivity against presumed climate change as found using the 201 x 201 grid size.

Even if our findings, derived from the 201 x 201 domain size simulations, are not conclusive evidences, they appear at least to be plausible.

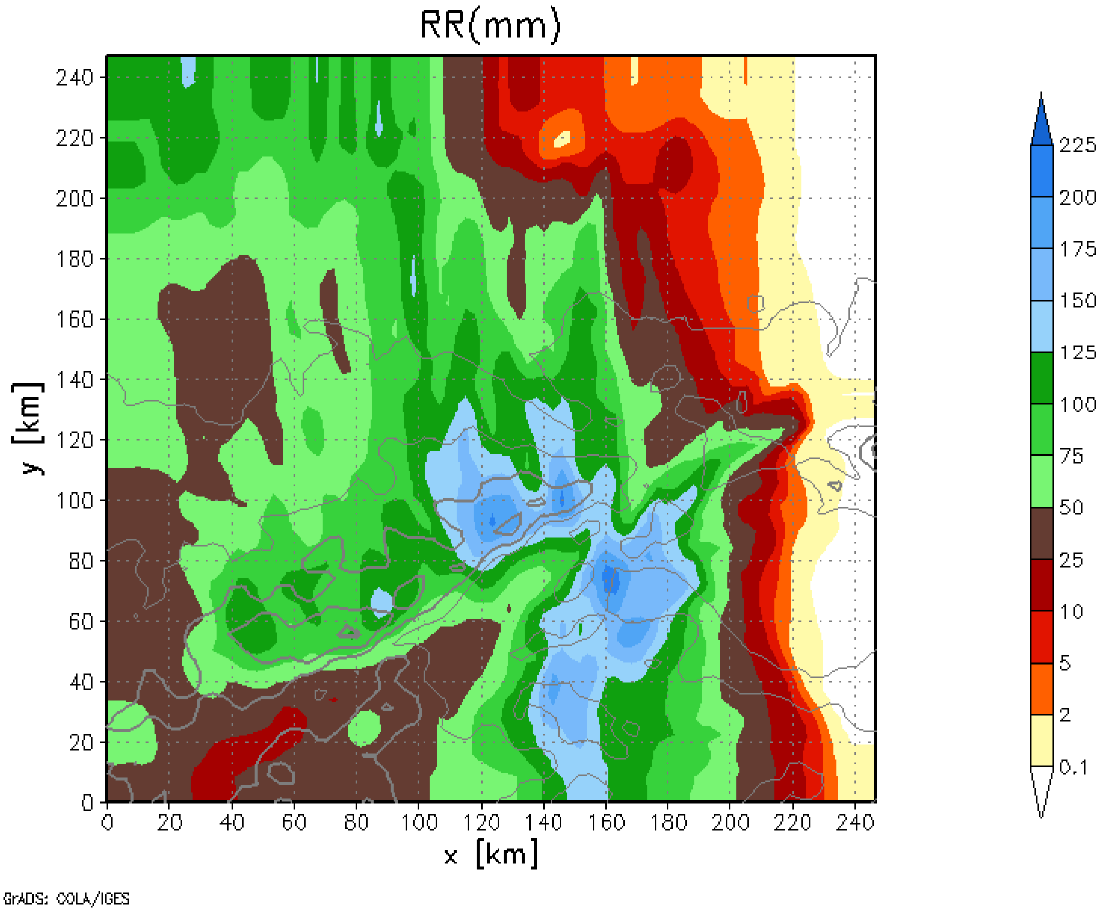

Fig. 21: Reference case (ID20) (model domain: 89 x 89 grid points): 24 h rain sum from COSMO-DE simulation, August 12, 2008, 00+06 UTC to 00+30 UTC.

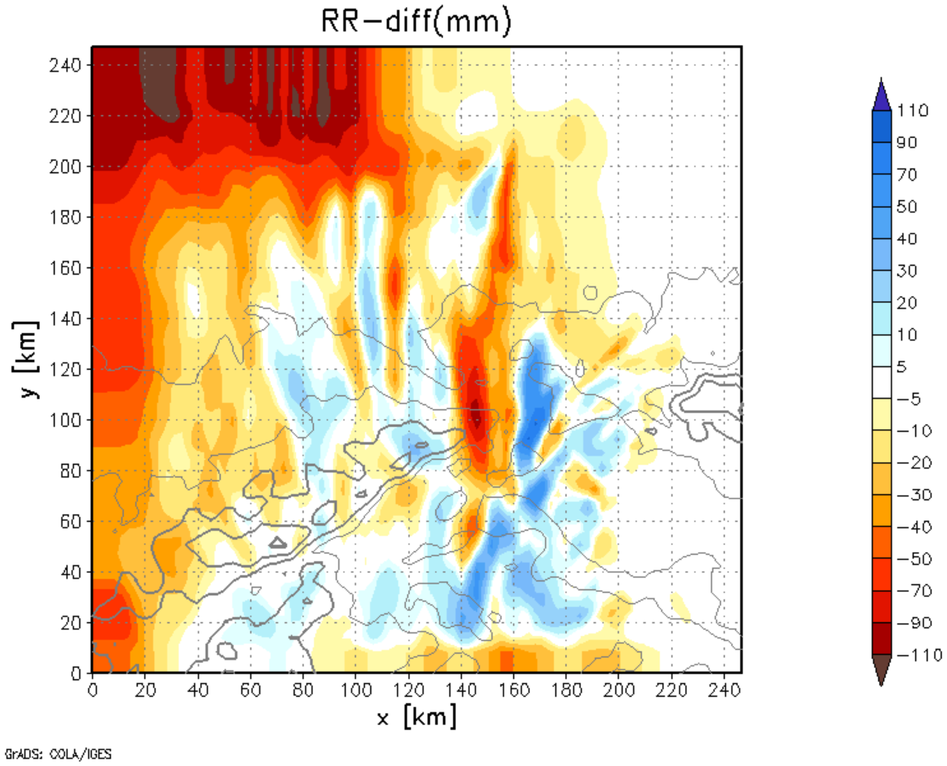

Fig. 22: Case ID21 (model domain: 89 x 89 grid points): Difference “ID21 minus ID20” of 24 h rain sum from COSMO-DE simulation, August 12, 2008, 00+06 UTC to 00+30 UTC, temperature offset $\Delta T$=+2 K.

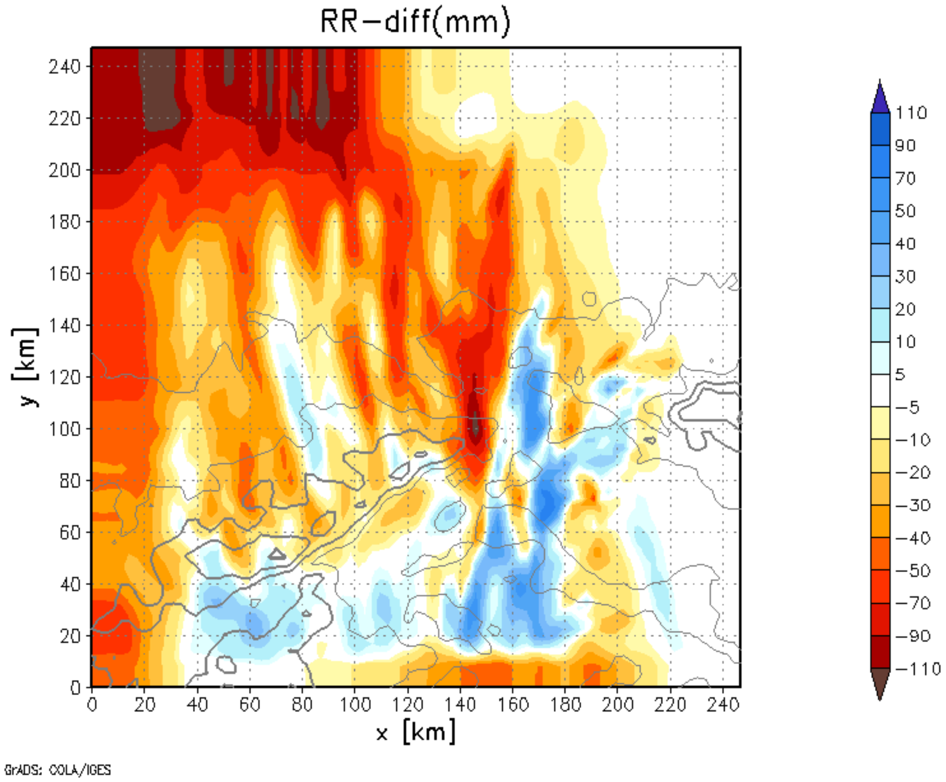

Fig. 23: Case ID22 (model domain 89 x 89 grid points): Difference “ID22 minus ID20”of 24h rain sum from COSMO-DE simulation, August 12, 2008, 00+06 UTC to 00+30 UTC, temperature offset $\Delta T$=+4 K.

## 6. Reasons for the Observed Model Sensitivity

### a) Disturbance of the predicted meteorological fields at the model boundaries

Owing to imbalances of the meteorological fields at the model boundaries, the precipitation is reduced in the vicinity of the boundaries (in a zone of up to six grid point from the domain boundary). Excluding the disturbed zone, the response of the rain sum against offsets of temperature and water vapour mixing ratio is nearly independent of the chosen domain size. In the interior model domain, impacts of the disturbed boundary conditions cannot be seen that obviously (e.g., in the pressure field). There is no evidence of a strong influence of imbalances at the boundaries on the precipitation formation in the inner model domain.

To overcome the boundary problem a more sophisticated approach to a posteriori consider large scale climate change in the COSMO model is required. However, owing to the technical challenge associated therewith it could not be realised within the framework of the present project.

### b) Change in relative humidity distribution

For different model domain sizes and $\Delta q_v/q_v$ values, that ensure nearly constant RH values (ID 03/04, ID 07/08, ID 13/14, ID 17/18) the area accumulated rain sums are close to the ones obtained for the reference runs. For example, for a domain size of 201 x 201 grid points and $\Delta T$=+4 K and $\Delta q_v/q_v$ =+25% the rain sum deviates from the reference run by only -1%. Thus, keeping the relative humidity at constant value, the rain sum remains constant, independently from the temperature offset. This is due to both, the high sensitivity of orographically induced precipitation against the lifting condensation level, and the sensitivity of the latter against the relative humidity. Vertical cross sections of the predicted relative humidity at 00+28 h (at this time, orographic lifting was the dominant rain-forming process) reveal a positive correlation between the relative humidity and rain sums (not shown). The rain sums seem to be much stronger dependent on the relative humidity than on the temperature offset.

### c) Increasing static instability due to enhancement of the water vapour mixing ratio near the surface

The enhancement of the water vapour mixing ratio decreases the air density and increases the buoyancy of an air parcel. At $\Delta q_v/q_v$=const the absolute amount of added water vapour is higher in the lower levels than aloft, i.e., the air mass becomes less stable. A changing thermal stability affects both, large-scale air mass lifting as well as convective lifting. However, as the underlying dependencies are nonlinear, one cannot derive a simple rule, how the total rain sum is affected in general. The change of the static stability disturbs somehow the synoptic configuration and leads to a rearrangement of frontal systems and convergence zones. Thus, the superposition of rain-promoting processes is affected, too.

## 7. Summary and Conclusions

A series of sensitivity studies have been performed to estimate the impact of presumed global warming on synoptic scale weather events. Within the framework of an epignostic study, mesoscale simulations of the Elbe river flood (August, 2002) employing the COSMO-DE model of the DWD have been carried out. The first set of runs was realised on a model domain size of 201 x 201 grid points with different temperature offsets at constant relative humidity. The results gave no hints on enhanced area accumulated rain sums in a warmed climate. Different temperature offsets at constant relative humidity does not result in significantly changed rain sums with respect to the reference run (climate as is). Increasing the water vapour mixing ratio so far, that the relative humidity increases, the total rain amounts increase, too. The observed rain sums originate from both thermodynamic and hydrodynamic effects (superposition of stratiform, convective and orographic rain contributions). An enhancement of the absolute water vapour mixing ratio at higher temperatures is indeed a necessary, but not a sufficient condition to enhance the rain formation. For example, a displacement of rain-promoting synoptic features might counteract rain formation, when the superposition of processes from different scale is not ensured any longer in a changed atmospheric system.

To force the relevant synoptic structures to stay in their rain-promoting positions (cf. reference run), the model domain size was reduced. Therewith, the frontal lifting zone at the Ore Mountains appeared in the observed position. However, the superposition of relevant rain generating processes could not be achieved. A convergence zone with high amounts of convective precipitation evolves in the Southeast of the Ore Mountains. In the Zinnwald-Georgenfeld region, the convective precipitation got less. The sensitivity of area accumulated rain sum against temperature change was found to be the same as in the runs on the 201 x 201 grid. Comparative runs on the model domain with 89 x 89 grid points to compensate possible imbalances at the model boundaries revealed the same model sensitivity against supposed global warming.

With respect to rain formation, the relative humidity appeared to be the most important parameter in the present study. An enhancement of the relative humidity leads to a decrease of the lifting condensation level, which favours orographically induced rain formation.

The present calculations provide some hints on the nonlinear response of the synoptic system to supposed global warming. To produce such enormous rain sums as observed during the century flood of the river Elbe, a superposition of various atmospheric processes is necessary. If an existing synoptic constellation is disturbed, e.g., by artificial offsets onto the boundary conditions, the atmospheric pre-condition for heavy rains can be removed by spatial displacement of rain-promoting processes. So the efficiency of rain formation is reduced and the area accumulated precipitation can even decrease at rising water vapour mixing ratio. An enhancement of the water vapour is a necessary, but not sufficient condition to enhance the rain sum.

We conclude, that simple offsets of temperature and water vapour mixing ratio affect the interaction of rain-promoting and rain-forming processes in a complex manner. Without a detailed scale analysis of the involved processes it is impossible to separate their contribution to the total rain sum. From a sensitivity study of the Elbe river flood we suppose, that the relative humidity itself is the most important meteorological value, which controls the rain formation. Nearly constant values of the relative humidity lead to nearly constant rain sums, despite of higher absolute water amount in a warmer atmosphere at RH=const.

To get closer insight the impact of climate change on flood events additional investigations are required:

1. Execution of global scale simulations with the GME model under changed climate conditions to avoid artificial offsets to the boundary conditions in a subdomain of the meteorological simulation. In this way, one can consider the impact of climate change throughout the whole life time of the cyclone, i.e., beginning from its embryonic stage until the dissipation of the cloud vortex. Thus, one can avoid a decoupling of the hypothesised future (due to an artificial change of the boundary conditions) from the observed past of the synoptic system. Questions of interest are:
    - How the trajectory and strength of the cyclone will evolve over a longer period?
    - Will the occurrence frequency of Vb weather situations increase or decrease in a future climate? There are indications, that Vb weather situations will occur more frequently under future climate conditions (cf. Ulbrich et al., 2002).

2. Execution of ensemble simulations employing different initial conditions and physical parameterisations. Thus, one can ensure a sophisticated weighting of different factors influencing the rain prediction.
3. Execution of conceptual studies for idealised conditions of orography (e.g., mountain ridge) and flow and thermodynamic setting. Such an approach would allow a better separation of the single contributions of the rain-promoting processes during a heavy weather event. The here employed mesoscale atmospheric model is a suitable basis for such studies.
4. Dedicated studies on the impact of cloud ice parameterisation on rain formation. Further simulations of the Elbe river flood (with a return period of up to 500 years) are subject of an ongoing investigation.

## Acknowledgement

Special thanks go to Janek Zimmer, Leipzig, for his helpful comments on the model setup and the mesoscale simulation of the Elbe river flood. We also thank the DWD for providing us with the COSMO model code and the model input data.

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