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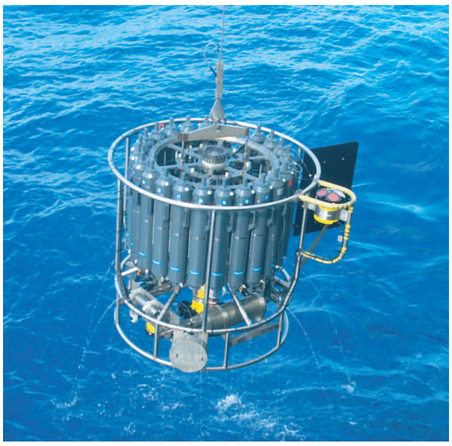
# The Hydrological Cycle: How observational data are able to improve climate models

Stefan Hagemann



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# The Hydrological Cycle: How observational data are able to improve climate models

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## Abstract

The hydrological cycle plays a prominent role within the Earth system and is crucially important to life on Earth including the human society. Thus, the current state of the hydrological cycle and its future development are key issues in environmental research. In studies of global and regional climate change, climate models are the current operational tools. Although the quality of climate models has considerably improved within the past decades, gaps or large uncertainties in the representation of some specific processes still exist. Consequently there is a lot of room for improvement. In order to improve climate models the use of observational data is inherently necessary, and various possibilities are at hand how observations may contribute to this task. This review presents an overview of these possibilities and considers several of them in more detail from a hydrological perspective.

# 1. Introduction

The climate of the Earth is influenced by increasing greenhouse gas (GHG) concentrations, changing aerosol compositions and loads as well as by land surface changes. Global climate models are used to investigate possible trends in the past and future global climate. For the future, this is done through the development of climate change scenarios. These follow specific assumptions for the evolution of greenhouse gases and aerosols, several of which have been defined by the Intergovernmental Panel on Climate Change (IPCC; *Houghton et al*., 2001) and are described in the IPCC Special Report on Emission Scenarios (SRES, *Nakićenović et al*., 2000). A classical overview on the general circulation of the atmosphere is given by *Lorenz* (1967). Further comprehensive overviews on the climate system are provided, e.g. by *Peixoto and Oort* (1992), *Hantel* (2005) and *Bengtsson* (1999). The latter gives insights into the numerical modelling of the Earth's climate.

The hydrological cycle (Figure 1) is crucially important to life on Earth as water is essential nourishment for all organisms as life on Earth is based on water. Humans and animals require water to survive as well as plants as no photosynthesis would be possible without water. Water occurs in all three states of aggregation, i.e. vapour, water and ice. The general circulation of the atmosphere is driven largely by the release of latent heat due to rain and snow formation. The hydrological cycle strongly affects the global energy cycle, and it plays also a central role in its interactions with the carbon, nutrient and sediment cycles. There are strong large-scale interconnections as, e.g. the tropical rain systems drive the mid-latitude circulations and North Eurasian snow cover modulates the South Asian monsoon. At longer time-scales, the hydrological cycle affects the groundwater storage, the thermohaline circulation in the ocean and the evolution of glaciers and ice sheets. Hydrological regimes vary dependent on local and regional climate variations. Looking towards future climate, the projected climate change in the mean and in the variability will in turn produce changes in hydrological conditions. Thus, an adequate representation of the hydrological cycle, its future development and associated uncertainties are key issues in studies of global and regional climate change (e.g. *Cubasch et al.*, 2000). In this context, it must be noted that hydrological processes cover a wide spectrum of spatial scales. Many hydrological fluxes (except from atmospheric water vapour transport) depend on processes that are generally several orders of magnitude smaller than the typical grid-size used in current general circulation models (GCMs) and in current regional climate models (RCMs). The formation of precipitation, for example, is controlled by a multitude of processes such as cloud microphysics and particle growth, radiative transfer, atmospheric dynamics on a variety of space and time-scales, and inhomogeneities of the Earth's surface, all of which have to be properly represented in a GCM or RCM. Consequently the importance of the hydrological cycle is highlighted by the Global Energy and Water Cycle Experiment (GEWEX; e.g. *Sorooshian et al*., 2005). The implications of changes in the hydrological cycle induced by climate change may affect the society more than any other changes, e.g. with regard to flood risks, water availability and water quality.

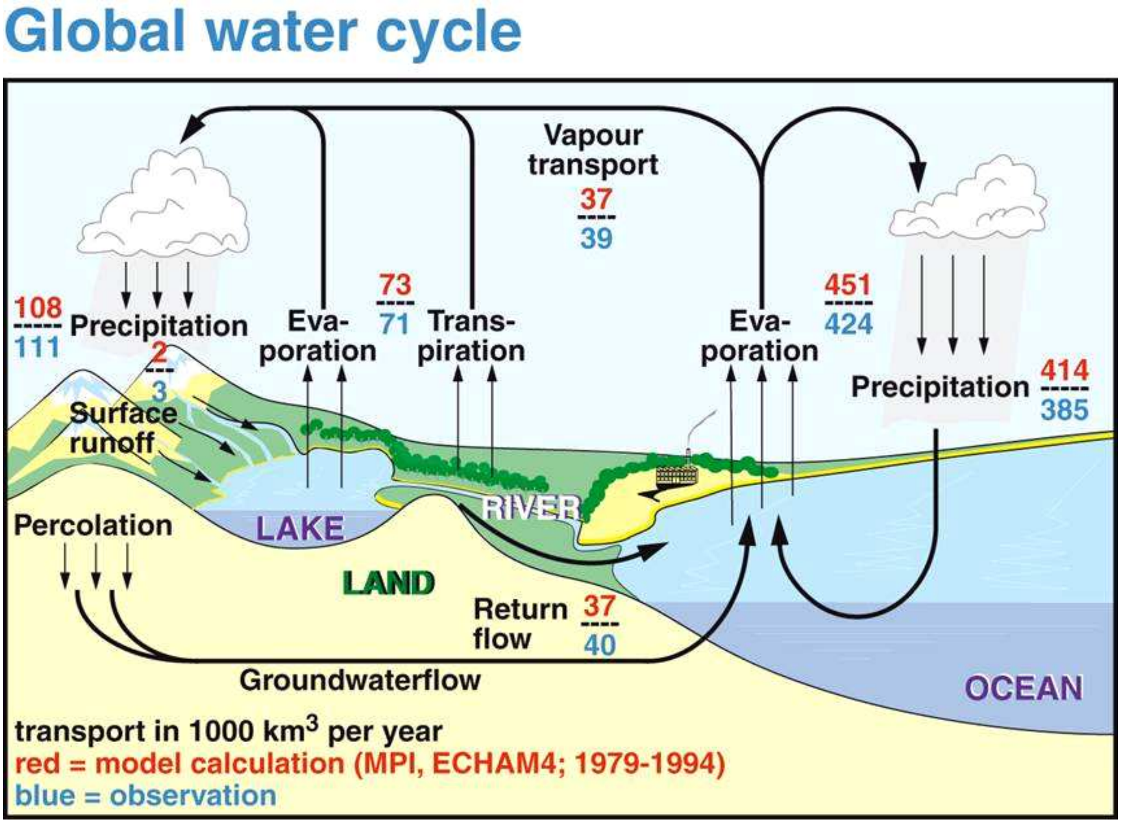


**Figure 1** Global water cycle with flux estimates based on the GCM ECHAM4 (*Roeckner et al.* 1996) for the period 1979-1994 and climatological estimates of *Baumgartner and Reichel* (1975). Unit: 1000 $km^3$/a.

Increasing $CO_2$ levels and temperatures are intensifying the global hydrological cycle, with an overall net increase of rainfall, runoff and evapotranspiration, and will increasingly do so (*Huntington*, 2006). Increasing $CO_2$ levels affect plant physiology, thereby likely reducing evaporation, and *Gedney et al*. (2006) found some evidence that recent increases in river runoff globally are due to this effect. The intensification of the hydrological cycle will likely cause an increase in extremes, i.e. floods and droughts (*Arnell et al*., 2001). There are suggestions that inter-annual variability will increase in some regions – with an intensification of the El Niño and North Atlantic Oscillation (NAO) cycles – leading to more droughts and large-scale flooding events.

Due to the lack of computer power, global climate models are generally still not able to explicitly represent surface heterogeneities on scales less than about 100 km grid length. However, global climate change has an influence on these local and regional scales, which will be experienced by man-kind directly (*Christensen et al.*, 2007a). Improved knowledge on regional climate change can be achieved with the use of different regionalization techniques, including high-resolution and variable resolution GCMs (*Cubasch et al.*, 1995, *Déqué and Piedelievre***,** 1995), nested RCMs (*Giorgi and Mearns*, 1999), and statistical downscaling (*Wilby et al.*, 1998). The accuracy of all downscaling techniques largely depends on the ability of a GCM to represent the current weather patterns. This is especially true for statistical downscaling techniques as their methods are based on current weather correlation patterns. For statistical downscaling applications in climate change studies, it is inherently assumed that these correlation patterns will not change in a different climate. This assumption is considered as a weakness of this method (*Houghton et al.*, 2001), and it will not necessarily hold true in each region of the world (*González-Rouco et al*., 2000).

RCMs are used for the dynamical downscaling of the global scale GCM simulations to regional scales (e.g. *Giorgi*, 2006a). Climate simulations performed with GCMs provide a consistent representation of the large-scale global circulation in both the atmosphere and the ocean, while RCMs introduce more details to the atmospheric simulations due to regional features such as topography and inland seas (*Rummukainen et al.*, 2001). In both cases, simulations are usually produced for a control climate representing present-day climate conditions and for future climates representing various emission scenarios. *Giorgi* (2006a) gives a concise review of regional climate modelling, from its ensuing stages in the late 1980s to the most recent developments.

In order to improve climate models the use of observational data is inherently necessary. Observations may contribute to model improvement in a variety of different ways. The aim of this review is to consider several of these ways in more detail. Here, given the importance of the hydrological cycle within Earth's climate, the possibilities of climate model improvement by the use of observational data are considered from a hydrological perspective within the framework of climate research.

Form a modeller's perspective, the most obvious application of observational data is their use for validation in model evaluation studies. Section 2 discusses some of the problems connected to climate model validation and includes also example studies for the validation of GCMs and RCMs over hydrological regimes. The second application presented in Sect. 3 is the improvement of model results by the direct use of observations at several different places within a climate model. Section 4 considers improvements yielded by the enhancement of model parameterizations and evaluation methods and their usage of observational data. Finally, Sect. 5 will shed some light on challenges that arise when future climate changes are considered. Here, a few burning topics of climate research will be tackled, such as the enhanced representation of hydrological processes within a comprehensive Earth system model, the uncertainty in climate simulations, changes in hydrological extremes and the effect of land use changes on the climate. Please note that extreme events won't be considered within this review except from a short look on them in Sect. 5.3.

## 2. Model evaluation using data for validation

One of the main applications (many researchers claim it is the most important application) of observations in climate modelling is model validation. It is a necessary process each model developer has to conduct. Consequently a lot of studies exist where this is done in more or less comprehensive ways for the different GCMs and RCMs. Integrative studies of the more comprehensive kind comprise studies that involve an ensemble of climate model simulations, either of the same model or of different models. Such ensembles can be used to analyse common model problems, to investigate climate variability issues over certain regions or to tackle questions of uncertainty in the simulated climate. *Koster et al.* (2004), e.g. investigated the land-atmosphere coupling strength of soil moisture on precipitation in the boreal summer using a suite of specific ensemble simulations from 14 atmospheric GCMs whose setup was coordinated by the Global Land-Atmosphere Coupling Experiment (GLACE; *Koster et al.*, 2006). Several model intercomparison projects (MIPs) have been launched to conduct integrative studies whereas some of them are mentioned below.

In order to validate climate models, some basic principles should be followed (see, e.g. *Schlünzen*, 1997). In this respect, the validation of the climate model is considered as the validation of the whole system of dynamical and physical processes. It is expected that the internal representation of each of these sub-processes has been tested and validated in offline simulations and case studies beforehand. In order to avoid the introduction of systematic errors via the initial or boundary conditions (often denoted as forcing of the climate model), a simulation used for validations studies should be forced by observations. Such simulations are often designated as baseline simulations. For an atmospheric GCM, this baseline simulation has to be forced by observed sea surface temperature (SST) and sea ice, such as, e.g. for an AMIP-style experiment (Atmospheric Model Intercomparison Project; *Gates et al.*, 1999) using observed monthly sea surface temperatures and sea-ice cover for the time period 1978-1999. For ocean models, Coordinated Ocean-ice Reference Experiments (COREs) are proposed as a tool to explore the behaviour of global ocean-ice models under common surface boundary forcing *(Griffies et al.*, 2007, 2008).

In order to minimize the influence of errors in the prescribed SSTs and the lateral boundary conditions (see also Sect. 3.3) used for the baseline simulation of a RCM these should be determined from re-analysis data, which are often designated as “perfect boundary conditions” (cf. *Machenhauer et al*., 1998; *Jacob*, 2001). As no comprehensive three-dimensional atmospheric observational dataset exists, re-analyses are closest to such an ideal dataset as they comprise many observations that have entered the re-analyses via data assimilation (see Sect. 3.4 for more details). In addition, a validation of coupled model systems for a control climate period must also be conducted as some model problems may arise only in a coupled model simulation. This applies to coupled atmosphere-ocean GCM simulations as well as to RCM simulation where these control climate GCM simulations are dynamically downscaled over a specific region. With regard to coupled atmosphere-ocean GCMs, three different climate model intercomparison projects (CMIP) were launched: CMIP-1 (*Meehl et al.*, 2000), the first project of its kind organized in the mid-1990s; the follow-up project CMIP-2 (*Covey et al.*, 2003, *Meehl et al.*, 2005); and CMIP-3 (*PCMDI*, 2007) representing today’s state of the art in climate models that were also used for the fourth assessment report (4AR) of the *IPCC* (2007). For RCM control simulations, e.g. *Jacob et al.* (2007) performed a model intercomparison over Europe. For North America, a regional climate model intercomparison is currently being conducted in the North American Regional

Climate Change Assessment Program (NARCCAP; http://www.narccap.ucar.edu/).

On a first glance the process of model validation looks rather trivial: one has 'just' to compare the simulated climate data with corresponding observed data. But a more thorough investigation of the climate model validation task reveals that this task is connected with a lot of problems and possible traps that may easily misled the conclusions drawn from the validation results. Especially the validation of components of the hydrological cycle is a general problem in global climate modelling. The comparison of climate model output to observational data is only useful if the errors in the data are not too large. But particularly for components of the hydrological cycle measurement errors may be large, so that the quality of the simulation of the hydrological cycle is often difficult to judge. In Sect. 2.1, the main problems of climate model validation are summarized. Sect. 2.2 considers several examples where the simulated hydrological cycle of climate models is evaluated over hydrological regimes so that some of the problems mentioned in Sect. 2.1 are solved or reduced.

## 2.1. Common problems in climate model validation studies

Climate models are generally simulating their data on a model grid while many observations exist only at point locations. A comparison of gridded climate model data to point data is connected with several difficulties. As the least well resolved wave on a numerical grid has a wavelength of 2 $\Delta x$ (*Durran*, 1998), where $\Delta x$ is the mesh size of the model grid, gridded climate model data have an accuracy of 2 $\Delta x$. Therefore, one single point measurement from an observing station must be compared to at least 4-9 model grid boxes. In order to pay regard to the exact location of the point observation within the model grid, a weighted averaging should be applied to the gridded data to calculate a horizontally interpolated model value that is representative for the point location. Here, for many variables a linearly weighted averaging is not sufficient so that more sophisticated methods should be used, such as, e.g. the *Barnes* (1964) method. A more detailed overview of this method is given in, e.g. **Hagemann et al. (2003a)**. If the area under consideration has an orographically strongly varying structure, even more complex averaging methods are needed (e.g. the amount of precipitation depends on the wind conditions of the mountains slope). *Ahrens* (2005), e.g. compared a statistical distance method to the standard use of geographical distances in the interpolation of an available coarse rain gauge network and yielded more robust interpolation results at sites of a denser network with actually lacking observations, where the performance is especially enhanced in or close to mountainous terrain.

For some variables gridded observational datasets exists (e.g. temperature, precipitation) that are either interpolated from station measurements or derived from satellite measurements. But these gridded observations also have some errors and uncertainties. For station based datasets these uncertainties are connected with general measurement errors, inhomogeneities of the measuring instruments, inhomogeneous station density, errors in the station documentation and allocation, and the accuracy of the interpolation algorithm. For satellite derived datasets uncertainties depend on general measurement errors and the accuracy of the model algorithms that are used to derive a specific quantity from the measured radiances. A more detailed look onto these problems with respect to precipitation is given by, e.g. *Rudolf and Rubel* (2005).

Trivially, the time period of the observations should be the same as the time period of the model results, especially for shorter time periods, this is absolutely necessary. But this requirement can often not be fulfilled in data sparse regions. For longer time periods, comparisons of long-term climatological values covering different time periods are also feasible (**Hagemann und Dümenil, 1999**).

A further problem encountered in the comparison of simulated and observed data is caused by elevation differences between the model grid and the observational grid/station location. This is especially important for temperature where the temperature observations (or the model data) need to be corrected depending on the elevation difference between the model grid box and the observation station/grid box. Commonly a lapse rate of 0.65 K/100 m is used to convert a temperature from one elevation to another elevation. This lapse rate is typical for wet adiabatic conditions. But also for precipitation a height correction may be appropriate if precipitation over mountainous terrain is considered. But, here no common approach is used. *Daly et al.* (1994) used a statistical-topographic model based on regression of precipitation with orography to map climatological precipitation over mountainous terrain. *Adam et al.* (2006) describe a correction method for gridded precipitation in mountainous regions that requires good discharge observations and is based on a combination of catchment water balances and variations of evaporation estimates.

A general point of concern in climate model validation, which does not only apply to many components of the hydrological cycle, is that many climate variables are not adequately measured at larger scales. In order to evaluate these variables on a global scale alternative datasets have to be chosen. For many variables, re-analysis datasets (see Sect. 3.4) of past global numerical weather forecasts are good choices as many observations are entering the numerical weather forecast system via data assimilation. But regarding the hydrological cycle the re-analysis data show a lot of problems, such as shown by **Hagemann and Dümenil Gates** (2001) and **Hagemann et al.** (2005). For some more information on re-analysis datasets, see Sect. 3.4.

Precipitation is the central component of the terrestrial hydrological cycle as it is the main water supply for all land based creatures and plants. Thus, precipitation is an important subject for climate model validation studies but particularly for precipitation data (*Legates and Willmott*, 1990) measurement errors can be large. The most important uncertainty in precipitation measurements is related to the common underestimation of the amount of precipitation due to undercatch of the measurement gauge. The precipitation measurements may have an error of up to 10-50%, which depends on wind speed, air temperature and state of aggregation (see, e.g. *Rudolf and Rubel*, 2005). Thus, in principle, the precipitation measurements must be corrected for this undercatch, but this correction is usually not done for most of the available precipitation datasets. A more detailed overview on available global precipitation datasets and their problems is given in Sect. 2.1.1.

Apart from precipitation and the re-analysis datasets, several global datasets exist that may be used for the validation of the components of the hydrological cycle. But many of them are afflicted with large uncertainties. Thus, Sect. 2.1.2 considers some of the more common global hydrological datasets.

### 2.1.1. Global precipitation datasets

As mentioned in Sect. 2.1, global gridded precipitation data are required for the validation of a GCM. Over land, a global network of precipitation gauges exists, which can be used to construct global maps of precipitation. But the station density is largely varying from region to region so that the quality of a derived precipitation map may also have large regional differences. In order to achieve a precipitation value that is representative for the whole grid box an adequate sample of stations within the grid box is necessary. In addition the temporal availability of measurement data may also largely diverge between regions. Thus, an appropriate interpolation of data from neighbouring grid boxes has to be used to obtain values for grid boxes where little or no station data are available. Also regional differences in data quality may exist as each of the gauging instruments is afflicted with distinct systematic errors. Especially for snowfall the underestimation of its amount can be substantial.

Table 1: gives an overview of several currently available global datasets with long-term monthly mean time series of gridded precipitation over land and ocean. A more detailed discussion of global precipitation datasets is presented by *Rudolf and Rubel* (2005). In this respect, *Biemans et al.* (2008) compared seven global gridded precipitation datasets at the river basin scale and investigated the impact of precipitation uncertainty within these datasets on discharge simulations. Note that the GPCC-Vas data were constructed only for the application in studies concerning long-term aspects of climate variability. For such studies, it has to be ensured that station data used for gridding are as continuous and homogeneous as possible. Therefore, only station time series with a minimum of 90% data availability during the analysed period (1951–2000) were used for interpolation to a regular 0.5° x 0.5° grid in order to minimize the risk of generating temporal inhomogeneities in the gridded data due to varying station densities (*Beck et al.*, 2005).

**Table 1:** Global gridded monthly precipitation datasets.
CMAP (CPC Merged Analysis of Precipitation; *Xie and Arkin*, 1997), CRU = Climate Research Unit (*Mitchell and Jones*, 2005), GHCN (Global Historical Climatology Network; http://www.ncdc.noaa.gov/oa/climate/research/ghcn/ghcngrid.html), GPCC (Global Precipitation Climatology Centre – monitoring product; *Rudolf et al*., 2001), GPCC full (GPCC full dataset; *Fuchs et al*., 2007), GPCC-Vas (GPCC Variability Analysis of Surface Climate Observations; *Beck et al*., 2005) GPCP (Global Precipitation Climatology Project; *Adler et al*., 2003), HOAPS (Hamburg Ocean-Atmosphere Parameters and fluxes from Satellite data; *Jost et al.*, 2002), MW (Matsuura and Willmott; http://climate.geog.udel.edu/~climate/html_pages/download.html).

| Dataset | Resol. | Period | Database |
|---|---|---|---|
| **Land** | | | |
| CMAP | 2.5° | 1979-present | Satellite data + GPCC stations, no correction for sys. errors |
| CRU Vs. 2 | 0.5° | 1901-2002 | Up to 9000 stations with irregular coverage in time, no correction for systematic errors |
| GHCN2 | 5° | 1900-present | 20590 stations with irregular coverage in time, no correction for systematic errors |
| GPCC | 1° | 1986-present | ca. 7000 stations, no correction for systematic errors |
| GPCC full | 0.5° | 1951-2004 | Up to 43000 stations with irregular coverage in time (average: ca. 30000 stations), no correction for systematic errors |
| GPCC-Vas | 0.5° | 1951-2000 | 9343 stations, no correction for systematic errors |
| GPCP-V2 | 2.5° | 1979- present | Satellite data + GPCC stations, sys. errors corrected |
| MW | 0.5 | 1900-2006 | 4100-23300 stations including GHCN2, no correction for sys. errors |
| **Ocean** | | | |
| CMAP | 2.5° | 1979-present | Satellite data |
| GPCP-V2 | 2.5° | 1979-present | Satellite data |
| HOAPS 3 | 0.5° | 1988-2005 | Satellite data |

Even though the general patterns of precipitation over land are relative similar as all datasets more or less are using the database of the global station network, there are notable differences for specific regions (see, e.g. *Hagemann* 2002b). The main reasons for these discrepancies are the diverse ensembles of used measurement gauges, and whether and how the data were corrected for systematic measurement errors. Here, the different interpolation methods to yield gridded precipitation data from the point measurements are more of secondary importance. The availability of gauges within a grid box and, thus, the station density impacts the quality of gridded precipitation especially in data sparse regions. For example, the station density in the CRU and MW datasets is not sufficient for many regions (spatially and temporally) to justify a resolution of 0.5 degree (Rudolf, personal communication, 2001).

As archived observational data may contain metering, coding or formatting errors depending on the methods used for data retrieval, submission and archiving, an adequate quality control is necessary. Especially for GPCC data this has been extensively conducted (while a larger amount of data used for the CRU2 analysis still enclosed some errors (Rudolf, personal Communication, 2007)). The GPCP precipitation is generally larger than in the other datasets as a flat correction according to *Legates and Wilmott* (1990) was used to account for the systematic undercatch of measurement gauges. Here, it is known that this correction is too large by about a factor of 2 (*Rudolf and Rubel*, 2005).

As there are almost no measurement stations over the ocean, gridded precipitation data are usually taken from satellite observations. But these ‘observations’ are based on model algorithms used to derive precipitation amounts from measured radiances in the frequency band of the corresponding satellite. Thus, the quality of the derived precipitation is strongly dependent on the quality of the used model algorithm. Consequently, the three most commonly used datasets (see Table 1:) partially show large differences although the general patterns are similar (see, e.g. *Hagemann*, 2002b). The largest differences between the climatologies exist over the Tropics and the high latitudes (C. Klepp, S. Bakan, A. Andersson et al., in preparation). Thus, there is still a large uncertainty about the ‘true’ precipitation amounts over the ocean. Results of *Klepp et al.* (2003, 2005) indicate that HOAPS data show a more realistic distribution of extreme precipitation at the east coast of North America than CMAP and GPCP data.

### 2.1.2. Further global hydrological datasets

Table 2: gives an overview on common global hydrological observational datasets. Here, the following variables are briefly considered in the following: a) surface air temperature, b) vertically integrated water vapour (IWV) within an atmospheric column, c) evaporation, d) discharge, e) snowpack, f) soil moisture. Although the 2 m temperature is not a component of the hydrological cycle it is closely linked to hydrological processes so that it is often considered in hydrological studies, too. Thus, it has almost become a part of the cycle itself and will consequently be treated as such in the following.

**Table 2:** Global monthly observational datasets of hydrological quantities. CRU = Climate Research Unit (*Mitchell and Jones*, 2005), GHCN = Global Historical Climatology Network (http://www.ncdc.noaa.gov/oa/climate/research/ghcn/ghcngrid.html), GRDC = Global Runoff Data Centre (see, e.g., *Dümenil Gates et al.*, 2000), GSMDB = Global Soil Moisture Data Bank (*Robock et al.*, 2000), GWSP = Global Soil Wetness Project (*International GEWEX Project Office*, 2002), HOAPS = Hamburg Ocean-Atmosphere Parameters and fluxes from Satellite data (*Jost et al.*, 2002), ISCCP = International Satellite Cloud Climatology Project (*Rossow et al.*, 1996), MW = Matsuura and Willmott (http://climate.geog.udel.edu/~climate/html_pages/download.html), NVAP = National Aeronautics and Space Administration Water Vapor Project (*Randel et al.*, 1996), SDC = Snow Data Climatology (*Foster and Davy*, 1988)

| Variable | Dataset | Resolution | Period | Database |
|---|---|---|---|---|
| Temperature | CRU vs. 2 | 0.5° | 1901-2002 | ~1500-9100 stations with irregular coverage in time |
| | GHCN2 | 5° | 1880-present | 7280 stations with irregular coverage in time |
| | MW | 0.5 | 1900-2006 | 1600-12000 stations including GHCN2 |
| IWV | HOAPS 3 | 0.5° | 1988-2005 | Satellite data (over ocean only) |
| | ISCCP D2 | 280 km | 1983-2004 | Satellite data in cloud free areas |
| | NVAP | 0.5° | 1988-2001 | Satellite data and radiosondes |
| Evaporation | HOAPS 3 | 0.5° | 1988-2005 | Satellite and SST data (over ocean only) |
| Discharge | GRDC | Station | Varies | Stations for large catchments |
| Snowpack | SDC | 1° | Climatology | Stations |
| Soil Moisture | GSMDB | Station | Varies | >600 stations |
| | GSWP-2 | 1° | 1986-1995 | Multi-Model derived |

a) Surface air temperature

As for precipitation data (see Sect. 2.1.1), the CRU and MW gridded temperature datasets are based on station data so that the 0.5 degree resolution is certainly not justified in data sparse regions (spatially and temporally). But here this problem is less severe as the large-scale distribution of 2 m temperature is less heterogeneous than for precipitation. In general, e.g. in GHCN2, the best spatial coverage is evident in North America, Europe, Australia, and parts of Asia. Likewise, coverage in the northern hemisphere is better than in the southern hemisphere.

In gridded air temperature datasets, station measurements are usually height corrected for the difference between the station altitude and the mean gridbox elevation using a common lapse rate (cf. Sect. 2.1), which might introduce some uncertainty. In MW, e.g., each average-monthly station air temperature was first "brought down" to sea level (warmed) at an average environmental lapse rate (6.0 deg C/km). Traditional interpolation then was performed on the adjusted-to-sea-level average-monthly station air temperatures. Finally, the gridded sea-level air temperatures were brought up to the grid height (cooled) of a digital elevation model (DEM); once again, at the average environmental lapse rate. For some regions, the application of a constant temperature lapse rate might not be realistic and might lead to biases. Results of *Prömmel* (2008) over the Alps suggest applying a monthly varying lapse rate instead of a constant lapse rate in areas with complex orography to reduce biases caused by elevation differences. *Prömmel* (2008) also gives a good overview on problems related to the validation of gridded temperature data.

Further possible uncertainties in the data arise from the fact that different measurement instruments are used in the diverse regions of the Earth, which measure temperatures in different heights over the land surface. While, e.g., in Germany surface air temperatures are measured at 2 m height, in the US the 'Stevenson Screen' instrument is used that is measuring temperatures at a height of 1.20 m above the ground (Legates, personal communication, 1996). A height correction for the different measurement heights is generally not conducted. Even though the absolute error related to the height mismatches is difficult to quantify, it is likely small compared to other sources of uncertainty.

### b) Integrated Water Vapour (IWV)

IWV is also often referred to as precipitable water content. For atmospheric water vapour, the most widely used techniques are 1) the absorption of solar radiation, 2) the emission of microwave radiation, 3) the emission of infrared radiation, 4) the path delay of GPS radio signals due to refraction, and 5) radiosonde measurements.

The first technique allows accurate IWV measurements over land surfaces with a high spatial resolution. It is based on the absorption of solar radiation in the path sun - surface - sensor. The disadvantage of this method, though, is its high sensitivity to aerosols or thin cirrus clouds (*Albert*, 2005). Global data may be obtained from the Medium Resolution Imaging Spectrometer (MERIS) on the European Envisat platform and the Moderate Resolution Imaging Spectroradiometer (MODIS) on the U.S.-American TERRA and AQUA satellites, which became operational in the beginning of the 21$^{st}$ century.

Passive microwave measurements from polar-orbiting satellites provide the possibility to derive global gridded datasets of IWV. *Eymard* (2001), e.g., gives an overview on the retrieval of IWV from microwave radiometry. A disadvantage is that microwave retrievals are presently feasible only over oceans (*Randel et al.*, 1996). Since 1989, IWV data are commonly retrieved from the Defense Meteorological Satellite Program (DMSP) Special Sensor Microwave/Imager (SSM/I) (e.g. *Jackson and Stephens*, 1995), such as it was done for HOAPS 3 where IWV over the ocean was derived according to *Schlüssel and Emery* (1990).

IWV retrievals from infrared measurements over land and ocean can be obtained from the Television and Infrared Operational Satellite (TIROS) Operational Vertical Sounders (TOVS) (e.g. *Rossow et al*., 1991; *Wittmeyer and Vonder Haar*, 1994). But these retrievals lack general applicability as infrared satellite techniques are only applicable in the absence of significant cloud cover. The ISCCP D2 IWV is based on TOVS data, but its results are strictly valid only for relatively cloud-free locations.

Radiosonde and GPS derived data are primarily available over land with limited spatial and temporal coverage, even though the latter have the potential to provide a long-term systematic approach for monitoring atmospheric water vapour (see also Sect. 4.2). Consequently the best approach to obtain a long-term global dataset of IWV would be the combination of all available methods and data sources. This approach has been partially followed in the construction of the NVAP IWV dataset comprising a combination of radiosonde observations, TOVS and SSM/I data.

### c) Evaporation

In HOAPS 3, evaporation over the ocean was derived from SSM/I data and NODC/RSMAS (Rosenstiel School of Marine and Atmospheric Science / National Oceanographic Data

Center) Pathfinder SSTs (*Kilpatrick et al.*, 2001) according to a bulk formula using the parameterization scheme of *Fairall et al.* (1996).

Over land, a similar observational dataset currently does not exist. Direct measurements of evaporation or evapotranspiration from extended natural water or land surfaces are not practicable at present. However, several indirect methods derived from point measurements (see, e.g. *Golubev et al.*, 2001) or other calculations have been developed which provide reasonable results (*WMO*, 2006). For reservoirs or lakes, and for plots or small catchments, model-based estimates may be made by water budget, energy budget, aerodynamic, and complementary approaches. Detailed information on these methods can be found in *WMO* (1994).

There are several global model based estimates obtained by GCMs, NWP models or global hydrology models but their accuracy is highly uncertain. Results of the studies on global water resources vulnerability, available from several research groups worldwide, differ considerably, even for basic components of the global water cycle such as evaporation. This lack of knowledge has been identified by the on-going EU project WATCH (WATer and Global Change; www.eu-watch.org, see also Sect. 5.2, 5.3 and 5.4), which consequently aims at delivering a global gridded data set of observed evaporation over land and its associated uncertainties. To achieve this deliverable, evaporation data from a number of FLUXNET tower sites representing the major biomes and climatic regions will be collected, and these data will be combined with model-generated land surface evaporation.

d) Discharge

As described in Sect. 2.1.1, the validation of climate model precipitation is a problem due to the partially large errors and uncertainties in the gridded precipitation data. An alternative to the direct comparison of simulated and observed precipitation over land is the validation of discharges that are simulated using climate model data if no discharge scheme is included in the climate model (see Sect. 4.3).The discharge of most rivers can in principle be measured with comparatively small errors. For many large rivers these measurements are performed routinely, so that potentially a large global database exists (e.g. GRDC). If the global or regional distribution of lateral discharge is simulated, the validation of the simulated discharge against river gauge data therefore can provide a useful independent measure of the performance of the hydrological cycle of the climate model. If both riverflow and precipitation were given with reasonable accuracy, it would be a sufficient check of evaporation accuracy.

Some rivers must be excluded or handled carefully for validation purposes if they are heavily regulated (e.g. Nile after 1963, Volga) as the anthropogenic regulations are usually not included in discharge schemes coupled to climate models. Some hydrological models (such as WaterGAP; *Döll et al.*, 2003) include such regulations but as the corresponding formulations add a larger uncertainty, it might not be feasible to use this for the validation of the hydrological cycle of a climate model.

With regard to the lateral flow of water at the land surface, the term runoff is often used, which commonly leads to some communication problems. Sometimes it refers to the water from rain and snowmelt that is not infiltrated into the soil (surface runoff), to the whole amount of water that may be transported laterally at a certain location (total runoff), or to the amount of water that is already laterally transported as discharge by rivers (river runoff). In the long-term annual mean, total runoff equals discharge within a catchment, and in this case runoff is also equivalent to precipitation minus evaporation averaged over the catchment. To avoid these communication problems the clear specification of the term runoff is generally recommended and subsequently used in the review presented here.

e) Snowpack

*Foster and Davy* (1988) published a global climatology of snow depth where the snow depth data are commonly divided by an average snow density of 3.3333 g/cm$^3$ to yield the corresponding water equivalent, which is used to validate the snow pack simulated by climate models. Here, this calculation method is afflicted with some uncertainty as several processes affecting the snow density are neglected. The wetness of snow influences its density whereas air temperature and the availability of atmospheric moisture determine how wet or dry the snow is. The snow density increases as the snowpack becomes deeper and the lower layers are compressed. Compression has an impact on the crystalline structure of the snowpack, and density and crystalline structure affect how fast the snowpack melts and how much water it yields.

The snow depth climatology was developed based on an extensive literature research using as many data sources as possible. *Foster and Davy* (1988) classified their data over different regions into 3 quality bins (high, medium, low). Although they rated the data quality as high over Canada, United States, Scandinavia and the territory of the former Soviet Republic and medium over China and the alpine countries except for Germany (high), the quality over mountainous regions was evaluated as low. But as mountainous regions contribute the major part of the winter snowfall in many regions of the world, the resulting higher uncertainty must be regarded in model validation studies.

f) Soil Moisture

Soil moisture is an important component in the atmospheric water cycle, both on a small agricultural scale and in large-scale modelling of land-atmosphere interactions. Vegetation and crops depend at any time more on the moisture available at root level than on precipitation occurrence. Water budgeting for irrigation planning, as well as actual scheduling of irrigation action, requires local soil moisture information. Knowing the degree of soil wetness helps to forecast the risk of flash floods, or the occurrence of fog (*WMO*, 2006). There are various soil moisture measurement techniques that mainly comprise in-situ measurements at the plot scale. These are extensively described in up-to-date handbooks such as *Klute* (1986) and *Dirksen* (1999).

The GSMDB comprises soil moisture observations for over 600 stations from a large variety of global climates, including the former Soviet Union, China, Mongolia, India, and the United States. Most of the data are in situ gravimetric observations of soil moisture; all extend for at least 6 years and most for more than 15 years. But apart from this station data bank, no global gridded observational dataset exists.

The lack of global soil moisture observations (and also of global salinity information) has lead to the Soil Moisture and Ocean Salinity (SMOS) mission of the European Space Agency

(ESA) that is expected to be launched in 2009. SMOS has been designed to provide observational data on both variables from space, and this information is supposed to not only improve the understanding of the water cycle, but also to advance weather and climate prediction. In particular, soil moisture data will be important for extreme-event forecasting such as floods, landslides and droughts (*SMOS Project Team*, 2005). A limiting factor for the current applicability of SMOS data in climate modelling is that SMOS will provide soil moisture data only to a depth of few centimetres. Therefore, modelling techniques have to be developed to derive the moisture content within the root zone from time series of near surface soil moisture.

The GSWP is an ongoing environmental modelling research activity of the Global Land-Atmosphere System Study (GLASS) and the International Satellite Land-Surface Climatology Project (ISLSCP), both contributing projects of GEWEX in the World Climate Research Programme (WCRP). GSWP will provide global estimates of soil moisture, temperature, snow water equivalent, and surface fluxes by integrating one-way uncoupled land surface schemes (LSSs) using externally specified surface forcings and standardized soil and vegetation distributions. A major product of GSWP-2 will be a multi-model land surface analysis for the ISLSCP Initiative II period 1986-1995 (*International GEWEX Project Office*, 2002), which may be considered as a land surface analogue to the atmospheric re-analyses. The project will include an evaluation of the uncertainties linked to the LSSs, their parameters and the forcing variables. To obtain this land surface analysis the LSSs will be forced by near-surface meteorological data based on the NCEP-DOE re-analysis 2 (*Kanamitsu et al.*, 2002) at 3-hour intervals. For most of these fields, the re-analysis data have been hybridized with observational data, or corrected for differences in elevation between the re-analysis model topography and the ISLSCP Initiative II mean topography. As the multi-model analysis is not finished up to now, currently only data of the preceding GWSP-1 phase (*Dirmeyer et al*., 1999) for 1987-1988 are available. But as for re-analysis data (see Sect. 3.4) the data are model derived and not directly observed so that they will comprise larger uncertainties and biases.

## 2.2. Validation of climate models over hydrological regimes

A solution to overcome some of the problems mentioned in Sect. 2.1 is the performance of the validation over large areas that cover many model gridboxes. Here, the calculation of means averaged over these large areas usually compensates problems related to sparse station density, randomly distributed elevation differences between model grid and observations and horizontal interpolation problems. In GCM validation studies these means typically comprise global and zonal means. In hydrological studies, means over hydrological regimes such as river catchments or climate zones (such as, e.g., defined by *Köppen*, 1923) are well suited for this purpose. The evaluation of the hydrological component of climate models has mainly been conducted uncoupled from atmosphere/ocean GCMs (*Bowling et al.*, 2003; *Nijssen et al.*, 2003; *Boone et al.*, 2004). This is partly related to the difficulties of evaluating runoff simulations across a range of climate models due to variations in rainfall, snowmelt and net radiation (*Randall et al.*, 2007). Some attempts have, however, been made. *Arora* (2001) used the AMIP-2 framework to show that the Canadian Climate Model's simulation of the global hydrological cycle compared well to observations, but regional variations in rainfall and runoff led to differences at the basin scale.

*Milly et al.* (2005) considered an ensemble of 26 integrations of 20$^{th}$ century climate from nine GCMs and showed that at regional scales these models simulated river runoff with good qualitative skill. Further, the models demonstrated highly significant quantitative skill in identifying the regional runoff trends indicated by 165 long-term stream gauges. They concluded that the impact of changes in atmospheric composition and solar irradiance on observed discharge was, at least partially, predictable.

The validation described in **Hagemann et al. (2006)** is a further example for the hydrological validation of a GCM. This study investigates the impact of model resolution on the hydrological cycle in a suite of model simulations using a new version of the Max Planck Institute for Meteorology (MPI-M) atmospheric GCM ECHAM5 (*Roeckner et al.*, 2003). Special attention is paid to the evaluation of precipitation on the regional scale by comparing model simulations with observational data in a number of catchments representing the major river systems on Earth in different climate zones (Figure 2). It was found that a higher model resolution is generally improving the simulation of the hydrological cycle, such as shown for the annual mean precipitation in Figure 3. Remarkably, in most of the catchments (except for the Baltic Sea catchment), increasing vertical resolution is more beneficial than increasing horizontal resolution.

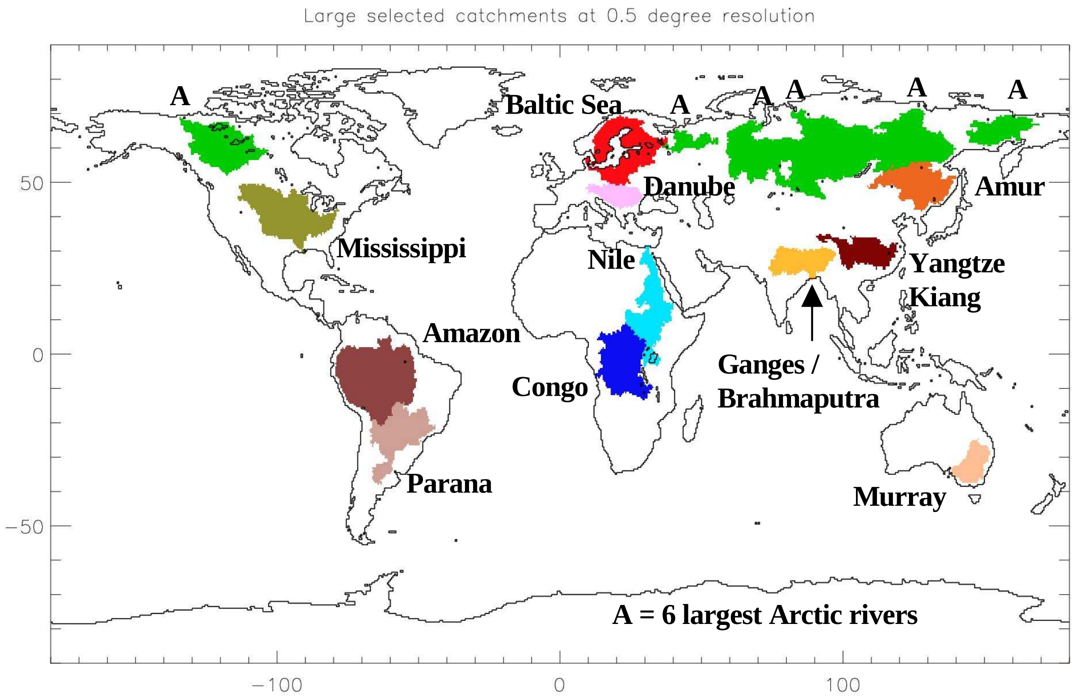


**Figure 2** Selected large catchments of the globe at 0.5 degree resolution.

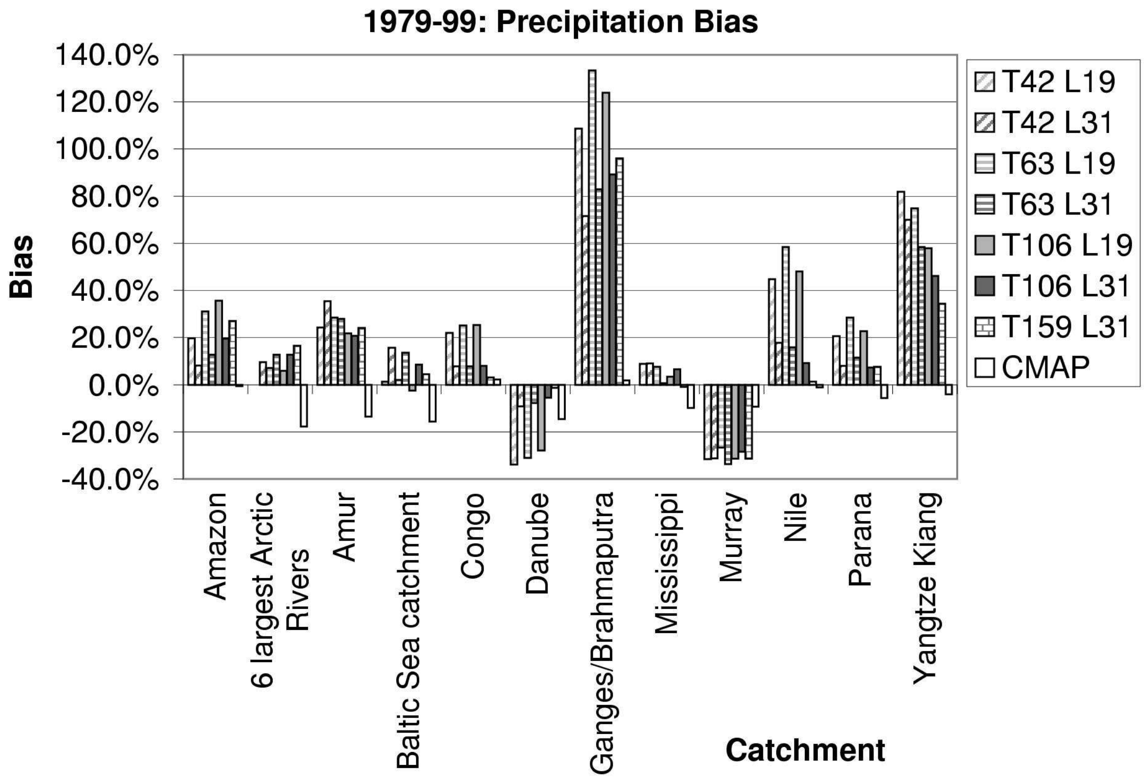


**Figure 3** Annual mean bias in simulated precipitation over several catchments. The bias was calculated from the difference of the simulated precipitation minus GPCP data. The horizontal resolutions T42, T63, T106 and T159 correspond to grid-sizes at the equator of about 313, 208, 125 and 83 km, respectively. Two vertical resolution with 19 (L19) and 31 (L31) layers are considered.

RCMs are usually applied at a much finer resolution than GCMs, currently ranging from typical resolutions of 50 km down to about 10 km. Higher resolutions are currently only used for process studies, but with increasing computer power they will soon be used for climate modelling, too. Thus, high resolution observational datasets are required for the validation of RCMs. These are currently available only for specific regions but not at larger scales. *Frei and Schär* (1998), e.g., have constructed a high resolution dataset of Alpine precipitation that they used for a RCM intercomparison and validation study with respect to daily precipitation statistics over the Alps (*Frei et al.*, 2003). However, due to the limited availability of high resolution observations at larger scales a focus on hydrological regimes is even more appropriate for the validation of RCMs than it is for GCMs.

As mentioned in Sect. 2, a baseline simulation should be performed for the evaluation of a RCM. This has been done with five RCMs within the EU project MERCURE (Modelling European Regional Climate: Understanding and Reducing Errors) that was launched to improve RCMs by understanding and reducing sources of errors, notably those arising through poor parameterization of physical processes and insufficient model resolution. Here, the 15 years re-analysis dataset (ERA15; *Gibson et al.*, 1997) of the European Centre for Medium-Range Weather Forecast (ECMWF) was used to provide the 'perfect' boundary conditions for the RCMs. **Hagemann et al. (2004)** evaluated the water and energy budgets simulated by these five RCMs and focussed especially on common model problems. A thorough budget analysis was conducted over the catchments of the Danube and the Baltic Sea (land area only). Here, a method was applied to estimate different components of the water balance which are not measured, i.e. the monthly changes in soil water storage. An alternative and spatially more widely applied approach for estimating changes in water storage was developed by *Hirschi et al*. (2005). A first comparison of the results yielded by the two methods showed a good agreement for the two catchments (Hirschi, personal communication, 2005)

For the Danube catchment, **Hagemann et al. (2004)** focused on the prominent summer drying problem. This special model feature is the too dry and too warm simulation of climate over central and south-eastern Europe during the summer (*Machenhauer et al*., 1998), which is typical for many RCMs, and to a less extent is also visible in some GCMs. Figure 4 compares the mean monthly annual cycle of precipitation of the five RCMs (ARPEGE, *Déqué et al.*, 1998; CHRM, *Lüthi et al.*, 1996; HadRM3H, *Jones et al.*, 1995; HIRHAM4, *Christensen et al.*, 1996; REMO Vs. 5.0, *Jacob*, 2001) to CRU Vs. 1 observations (*New et al.*, 2000) and ERA15. Here, it can be seen that the summer drying problem is a major feature of all models except ARPEGE. **Hagemann et al. (2004)** found two different reasons for problems in the RCM simulations. For ARPEGE and CHRM, the problems are related to deficiencies in the land surface parameterizations, while for HIRHAM, HadRM3H and REMO systematic errors in the dynamics appear to be causing the main errors in the simulations over the Danube catchment. The exact reasons for the summer drying problem are still not identified and are currently under investigation in the EU project CLAVIER (CLimate change And Variability: Impact on central and eastern EuRope, http://www.clavier-eu.org/).

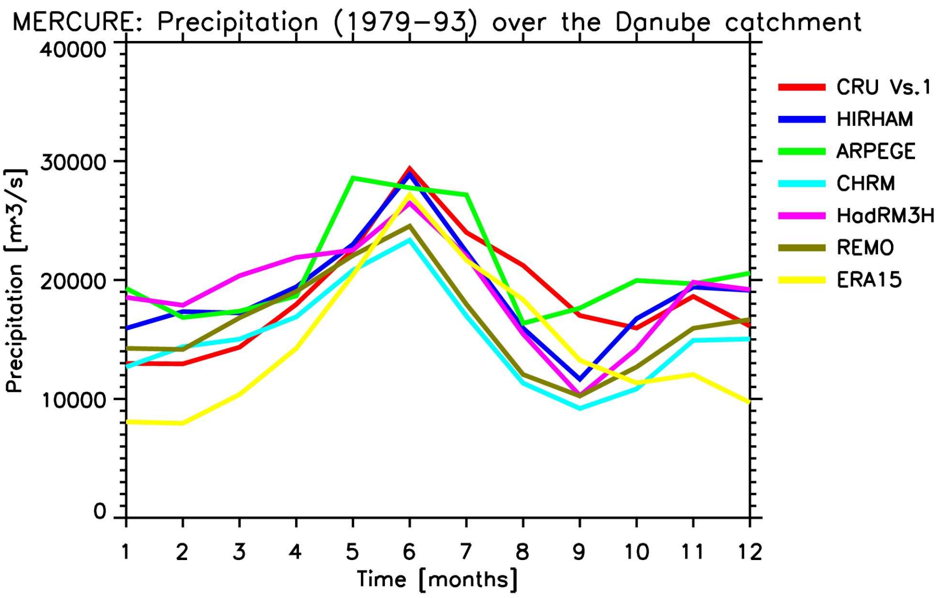


**Figure 4** Precipitation over the Danube catchment in mm/month.

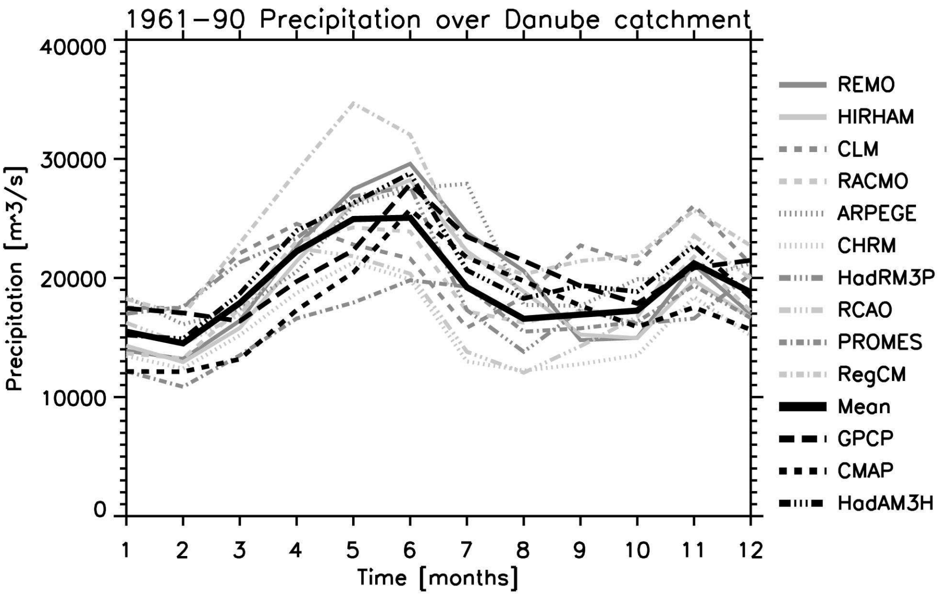


**Figure 5** Mean annual cycle of precipitation over the Danube catchment. Mean designates the multi-model ensemble mean of the 10 PRUDENCE RCMs.

Even though the evaluation of the performance of a RCM should focus on a baseline simulation to minimize the influence of errors introduced through the lateral boundary conditions, the validation of a control simulation (see Sect. 2) may also be desirable. This is important if the quality of a GCM-RCM combination shall be considered, or a qualitative intercomparison between different combinations shall be conducted. The latter has been done, e.g., in the EU project PRUDENCE (Prediction of Regional scenarios and Uncertainties for Defining EuropeaN Climate change risks and Effects; *Christensen and Christensen*, 2007), which aimed to predict uncertainties in RCM simulations over Europe (see also Sect. 5.2). Here, 10 RCMs were forced with observed SST and lateral boundary conditions provided by one GCM. **Hagemann and Jacob (2007)** evaluated the simulated hydrological cycle of the 10 RCMs and their multi-model ensemble mean over the catchments of the Baltic Sea (land area only), Danube and Rhine. Figure 5 reveals that the summer drying problem shows up again in the multi-model ensemble mean of precipitation over the Danube catchment as only two of the 10 RCMs do not have this problem (ARPEGE and RegCM3, *Pal et al.*, 2007). Please see Sect. 5.2 for further PRUDENCE studies focusing on specific catchments, which mostly include also some validation of the control simulations.

# 3. Improvement of model results by the direct use of observational data

The direct use of observational data in climate models is mainly advantageous at four places. Observations can be used for the initialization of certain model fields at the beginning of a simulation (Sect. 3.1). They can be used to prescribe boundary conditions at the Earth's surface-atmosphere interface that are not simulated within the climate model. Apart from the prescription of surface conditions over the ocean (SST, sea ice) in atmosphere-only climate simulations, this is of particular importance at the land-surface interface, especially for orography or vegetation dependent characteristics (Sect. 3.2). In limited area modelling, observational datasets are used as lateral boundary conditions when RCMs are applied for the dynamical downscaling of these datasets (Sect. 3.3). In order to improve the downscaling or to detect climate model errors, observational data can also be assimilated or nudged into the climate model system (Sect. 3.4).

## 3.1. Model initialization

Certain variables of the climate system have a long-term memory that may range from several months to several decades. As the state of these variables is usually largely influencing the general state of the climate system, their accurate initial representation is crucially important when a climate simulation is started. These variables comprise soil moisture and soil temperatures (time scale of seasons to several years), the ocean states of salinity and temperature (decadal), the distributions of snow (seasons) and sea ice (years).

The initial state of the ocean largely determines the climate development from the next season up to the next decade. This is a major focus in the currently increasing activities on seasonal to decadal predictions. *Smith et al.* (2007) presented a new modelling system that predicts both internal variability and externally forced changes of surface temperature from a global climate model, which allows them to forecast surface temperature with substantially improved skill throughout a decade, both globally and in many regions.

*Christensen* (1999) pointed out the importance of an adequate initialization of soil temperature and soil moisture in climate modelling experiments. An inadequate initialization of these fields may lead to transient signals that have to be suppressed as much as possible in modern numerical climate experiments as climate sensitivity experiments operate with quite small signals. He suggested a technique where the climate model is run to soil equilibrium in order to obtain starting conditions where transients are minimized and less random than what has been the case previously.

A theoretical study by *Walker and Houser* (2001) has illustrated that by assimilating near-surface soil moisture observations, as would be available from a remote sensing satellite, errors in forecast soil moisture profiles as a result of poor initialization may be removed and the resulting predictions of runoff and evapotranspiration improved. For future climate modelling studies, satellite data retrieved by the SMOS mission are supposed to improve the model initialization of soil moisture and ocean salinity (see Sect. 2.1.2f).

## 3.2. Model improvement by using data for boundary conditions at the land-surface interface

*Maynard and Royer* (2004) address the sensitivity to different parameter changes in African deforestation experiments and find that changes of roughness, soil depth, vegetation cover, stomatal resistance, albedo, and leaf area index all could make significant contributions. *Voldoire and Royer* (2004) find that such changes may impact temperature and precipitation extremes more than means, in particular the daytime maximum temperature and the drying and temperature responses associated with El Nino events (*IPCC*, 2007). Consequently an accurate representation of the land surface is necessary for the adequate modelling of processes at the land surface boundary to the atmosphere. This section gives a short glance on land surface parameters dependent on a) orography and b) vegetation.

a) Orography

Apart from the mean gridbox elevation DEM data are commonly used to calculate orography dependent parameters, such as the orographic variance, the orographic roughness length, and the shape parameter in the surface runoff/infiltration scheme of *Dümenil and Todini* (1992; Arno scheme). In climate models, these parameters are usually derived from a very high resolution orography such as the 30-arc-second topography dataset GTOPO30 (*Bliss and Olsen*, 1996).

The roughness length $z_0$ is an integration constant for the logarithmic wind profile in the surface boundary layer. In this Prandtl layer, the wind velocity becomes independent of the Reynolds number *Re* (for $Re >> 1$) and the wind speed depends logarithmically on the height above the surface (*Mason*, 1987). Formally, $z_0$ is the height at which the wind speed becomes zero when the logarithmic wind profile above the roughness sub-layer is extrapolated to zero wind speed. Thus, $z_0$ can be seen as a measure of the unevenness of the surface. In the models, the turbulent exchange of momentum, energy, and moisture between the surface and the atmosphere is calculated as a function of $z_0$. In areas of low orography, the vegetation part of the roughness length often controls this mixing (*Henderson-Sellers et al.*, 1986). According to *Schlichting* (1979), it is assumed that atmospheric drags induced by the surface roughness can be linearly combined. Thus, the roughness length $z_0$ is commonly separated into two parts: a roughness length $z_{0,oro}$ computed from the variance of orography, and a roughness length $z_{0,veg}$ of vegetation and land use. As a coarse approximation according to *Tibaldi and Geleyn* (1981), the square of $z_0$ equals the sum of the squares of $z_{0,oro}$ and $z_{0,veg}$.

For the orographic roughness length $z_{0,oro}$, different methods are available, e.g. developed by *Tibaldi and Geleyn* (1981), *Sattler* (2004). These methods comprise drag partition theory, effective $z_0$ and blending height concepts, etc. $z_{0,oro}$ is often calculated only from sub-grid scale orography variance in each grid-square, such as in the parameterization scheme of *Wood and Mason* (1993) where an 'effective roughness' length proportional to the standard deviation of orography at sub-grid scales is used to enhance the exchange coefficient for momentum. These kinds of values of $z_{0,oro}$ depend on the model's horizontal resolution and availability of high-resolution orography data, but are not strictly related to the physical processes the parameter is expected to represent (*Rontu*, 2007). Thus, improvements in the calculation of $z_{0,oro}$ may lead to a better simulation of orographic roughness effects. *Miller et al.* (1989) could improve their gravity wave drag scheme by the use of directionally-dependent sub-grid scale orographic variances. *Sattler* (2004) compared a linearized aggregation method of $z_{0,oro}$ to a non-linear method, and found advantages of the latter.

b) Vegetation dependent parameter

The amount of energy available to the climate is controlled by the global energy cycle which is largely dominated by atmospheric processes (see *Rosen*, 1999). About 46% of the energy entering the global climate system by incoming solar radiation is absorbed by the surface and about 31% is exchanged with the atmosphere as sensible and latent heat (*Rosen*, 1999). The land surface significantly influences the partitioning of energy between sensible and latent heat, and acts as significant medium to store energy on both the diurnal and seasonal time scale (*Pitman et al.*, 2004). In addition, the vegetation and the soil are major carbon stores. Therefore, the land surface is a key component of the climate system and the coupling between atmosphere and biosphere is of particular importance at the land surfaces from both the atmospheric and hydrological point of view.

The different processes at the land surface-atmosphere interface are affected by the land surface characteristics, such as the surface albedo, which determines how much of the energy that reaches the surface is reflected. A model used to simulate processes at this interface requires a proper determination of the land surface characteristics that are used in its process equations and parameterizations as boundary conditions. Therefore, the description of the land surface is a significant problem in global and regional climate modelling since deficiencies or inconsistencies in these boundary conditions may lead to errors in the climate simulations.

For an adequate modelling of climate, an appropriate representation of the land surface characteristics is required. As stated in a review by *Rowntree* (1991), numerous climate simulations have shown that anomalies in albedo and surface roughness can produce significant changes in the atmospheric circulation. *Pielke et al.* (1997) have demonstrated that the landscape, including its spatial heterogeneity, has a substantial influence on the overlying atmosphere. An adequate determination of land surface characteristics dependent on plant canopies is of particular importance because they strongly modify the evapotranspiration over large areas of the land surface which is a major component of the surface thermal and moisture balance and of the hydrological cycle (see e.g. *Kabat et al.*, 2004). Thus the assessment of new or improved land surface datasets was central to a number of programs and experiments, e.g. the International Satellite Land-Surface Climatology Program (ISLSCP) and the International Geosphere-Biosphere Program (IGBP). For an overview about these programs and experiments, see *Feddes et al.* (1998). As the nature of many land surface data is rather fractious in temporal and spatial coverage, *Dirmeyer* (2004) pointed out the importance of data consolidation for land surface data.

As mentioned by **Hagemann et al. (1999)**, several global land surface parameter datasets existed but the available (in 1999) datasets were inaccurate in some regions of the world and, generally, their spatial resolution was too coarse to fit the demands of high resolution limited area models. For example, the land surface parameter dataset of *Claussen et al.* (1994) was based on several of these datasets and showed an allocation of vegetation (visible in the albedo) in the coastal regions of the Sahara and Saudi Arabia which seems to be unrealistic according to a desertification map of *Diercke* (1988, 1992). Also several specific areas were not resolved such as the Namib desert in South Africa, the Persian highlands, the Sierra Madre in Mexico, the Great Basin and the Great Plains in North America, the Gobi desert and the desert and mountain ranges north of Tibet. Recent development in remote sensing facilitated the measurement of present land surface characteristics at a very fine spatial resolution thereby offering the possibility to create consistent land surface boundary conditions for numerical models.

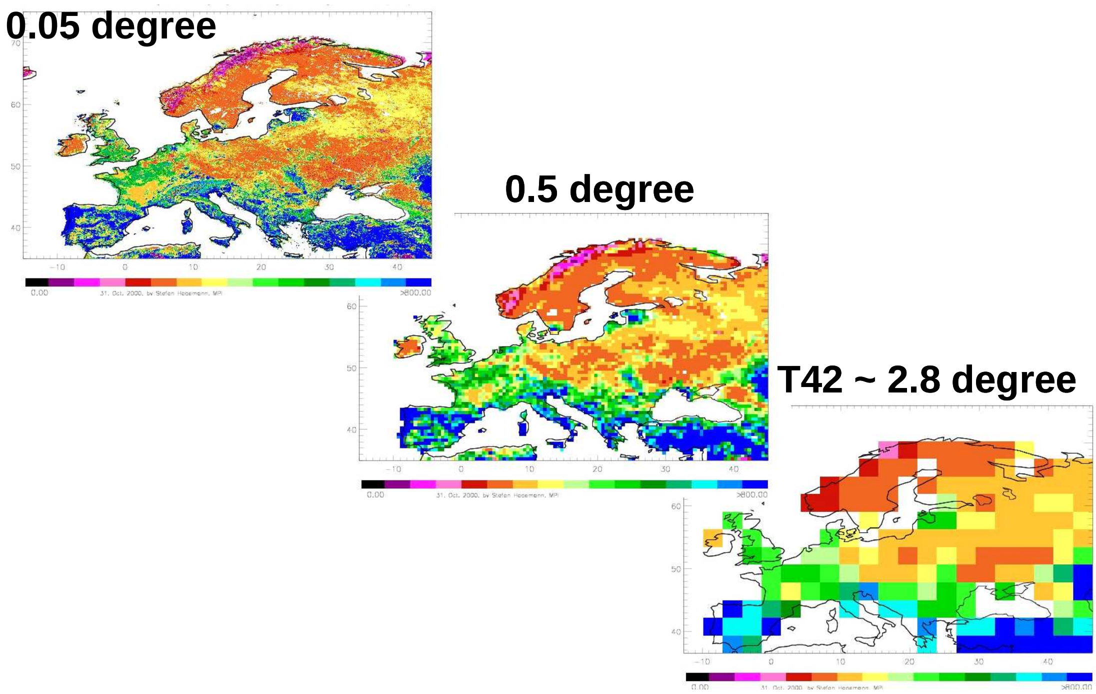


**Figure 6** Soil water holding capacities at 0.05° (about 5 km), 0.5° (about 50 km), and T42 (about 300 km) resolution over Europe according to the LSP2 dataset (**Hagemann, 2002a**). Colour steps: 50 mm.

**Hagemann et al. (1999)** have constructed a global dataset of land surface parameters (LSP) which is based on a 1 km global distribution of major ecosystem types (*Loveland et al.*, 2000) including glacial ice and open water according to the definitions given by *Olson* (1994a, 1994b). The latter was made available by the *U.S. Geological Survey* (1997). The set of the chosen parameters of the LSP dataset (background surface albedo, surface roughness length due to vegetation, fractional vegetation cover and leaf area index for the growing and dormancy season, forest ratio, plant-available and total soil water holding capacity) was defined by the parameters that are used or shall be used in the climate models of MPI-M. Later, the *U.S. Geological Survey* (2001) has distributed an updated version of their ecosystem dataset where land cover classes over 10% of the Earth‘s land area were revised. Consequently, **Hagemann (2002a)** incorporated these changes into the LSP dataset. During this implementation, several improvements were made to the LSP dataset. Over Africa, the background surface albedo of bare soil was corrected with METEOSAT albedo data. In addition, the seasonal variation of vegetation characteristics was considered and monthly mean fields of vegetation ratio, leaf area index and background albedo were developed and implemented.

From the basic resolution of 1 km of the ecosystem type dataset the LSP values can be aggregated to the respective model resolution, such as shown for soil water holding capacities over Europe in Figure 6. Due to the finest resolution of 1 km that may be obtained, the LSP dataset has been shown to be very suitable for the application in very high resolution regional climate modelling as it was done with the HIRHAM model (*Christensen et al.*, 2001; **Hagemann et al., 2001**) and the REMO model (*Rechid and Jacob*, 2006). But the implementation of the LSP dataset in the global ECHAM model has also led to improvements in the simulation of the hydrological cycle at the coarse resolution of T42 (about 2.8°) as shown in *Hagemann et al.* (2000). They have conducted several ECHAM4-T42 simulations using climatological AMIP SST where only a single parameter field was exchanged compared to the control simulation. Comparisons to observations yielded, e.g., that the new soil water holding capacities from the LSP dataset largely improve the simulation of evapotranspiration in southern and central Africa and therefore also of the 2 m temperature as shown for the Congo and Zambezi rivers in Figure 7. Thus, the global LSP dataset is available for use in regional and global climate modelling and it is implemented in the currently operational versions of the RCMs HIRHAM (*Christensen et al.*, 1996) and REMO (*Jacob*, 2001) as well as in the global ECHAM5 model (*Roeckner et al.*, 2003).

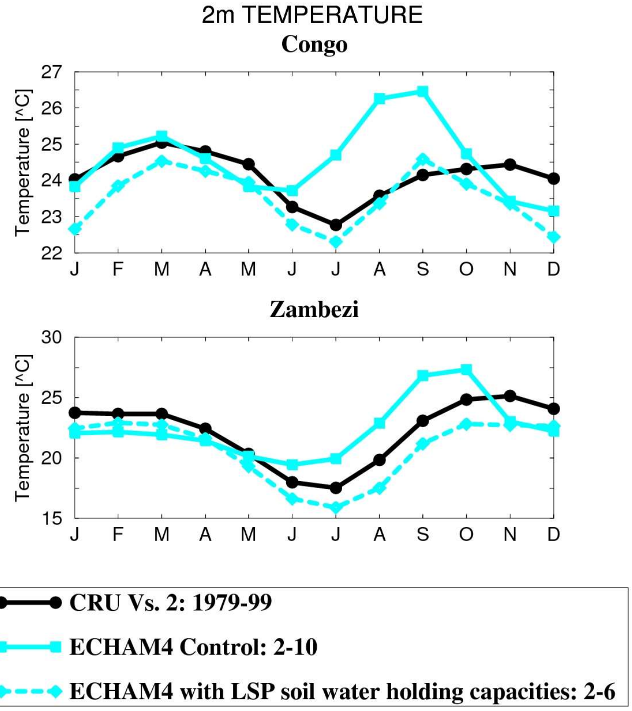


**Figure 7** Observed (CRU Vs. 2) and simulated 2 m temperatures over the Congo and Zambezi catchments.

*Rechid et al.* (2008b) have further refined the background land surface albedo and its seasonal variations using data products from MODIS. Here, they derived global fields of bare soil albedo and vegetation albedo. The total surface background albedo of a model gridbox is a composite of both albedos where the seasonal varying composition depends on the leaf area index. For sensitivity studies, this new surface albedo was integrated into the land surface schemes of the GCM ECHAM5 and the RCM REMO and the sensitivity of the climate model to the advanced surface albedo parameterization was tested (*Rechid et al.*, 2008a). This albedo parameterization has become operational in REMO since Vs. 5.7, and it has been implemented into the most recent version of ECHAM5-JSBACH (T. Raddatz, personal communication, 2008). JSBACH is the new MPI-M's land surface model comprising the ECHAM5 physics plus the interactive representation of vegetation and carbon fluxes (*Raddatz et al.*, 2007).

## 3.3. Lateral boundary conditions in limited area modelling

In order to perform RCM simulations, the RCM must be provided with initial and lateral meteorological boundary conditions (typically wind components, temperature, water vapour and cloud variables, surface pressure, chemical tracers if needed) and surface boundary conditions (SST and sea ice) (*Giorgi and Mearns*, 1999). If these are provided from a GCM simulation, the nesting technique (*Giorgi*, 2006a) is used, which is usually implemented in a one-way mode where the RCM information does not feed back into the GCM. Only recently a first two-way nesting study has been published by *Lorenz and Jacob* (2005) where the RCM solution feeds back into an atmospheric GCM. Before successful two-way nesting studies have only been performed with ocean models or RCMs alone (Lorenz, personal communication, 2008).

In the course of RCM development, regional simulations are usually carried out for a period in the past. The general model performance is then assessed through a validation procedure in which model results are compared against observational datasets on different temporal and spatial scales. In order to minimize errors and uncertainties originating from imperfect large-scale driving fields, lateral boundary conditions as well as initialisation fields are then usually provided by re-analysis or analysis products rather than by GCM control simulations (i.e. global climate simulations forced by observed atmospheric greenhouse gas concentrations). As mentioned in Sect. 2, re-analysis fields are also referred to as perfect boundary conditions since they are based on the observed state of the atmosphere and provide the best estimate of multi-decadal time series of large-scale conditions.

Several studies have dealt with the RCM sensitivity to the utilization of lateral boundary conditions, such as, e.g., *von Storch et al*. (2000), *Vidale et al.* (2003), *Marbaix et al.* (2003), and *Wu et al.* (2005), as well as to the resolution of these boundary data (e.g. *Denis et al.*, 2002).

## 3.4. Data assimilation and nudging

The assimilation of observational data is a common technique in numerical weather forecasts systems where the assimilated data are used to achieve an improved state of the atmosphere at the initial time for the next forecast. These initial fields are provided by operational analyses and comprise a data assimilation suite combining observations, previous forecasts, and model assumptions about the evolution of different meteorological variables. Since operational analyses are an estimate of the actual weather situation, long time series of these analyses should give an adequate description of the evolution of weather patterns and their average would describe the climate. However, the individual analyses are influenced by changes in the model, analysis technique, data assimilation, and the use of observations, which are an essential product of research and development at a numerical weather forecast centre. Thus, apparent changes of atmospheric conditions may occur in long time series of analysis fields that are caused only by changes in the corresponding analysis system. This led to the implementation of the re-analysis projects, in which a fixed analysis/forecast system is used to assimilate past observations over a long period of time. (Certain inconsistencies are still present, however, since the amount of available observations varies for different time periods.) For more detailed information on these topics, see, for example, *Uppala* (1997) and *Kållberg* (1997).

Because of a lack of globally distributed observations of many atmospheric variables, researchers in meteorology, climatology, or hydrology often use re-analysis data as pseudo-observations for validation, verification, initialization, or for the forcing of their regional models (see Sect. 3.3). Therefore, the validation by independent data not entered in the assimilation of the re-analysis data itself is an important issue. Especially with regard to the hydrological cycle the current re-analysis data show a lot of problems, such as shown by **Hagemann and Dümenil Gates (2001)** for the ERA15 re-analysis (*Gibson et al.*, 1997) and the re-analysis of the National Centers for Environmental Prediction (NCEP; *Kalnay et al.*, 1996), and by **Hagemann et al. (2005)** for the 40 years re-analysis (ERA40) of the ECMWF (*Uppala et al.*, 2005). Figure 8 shows an example for the Arctic Ocean catchment represented by its six largest rivers (cf. Figure 2). Here, NCEP data show a large overestimation of precipitation compared to GPCC and GPCP data, ERA15 slightly underestimates the precipitation, and ERA40 fits well within the span of the two observational datasets. For temperature, ERA15 has a severe cold bias in winter compared to CRU2 data, while the other two re-analyses show a winter warm bias that is more pronounced in ERA40. If daily time series of re-analysis precipitation and temperature are used to simulated discharge with the simplified land surface (SL) scheme (**Hagemann und Dümenil Gates, 2003**) and the Hydrological Discharge (HD) model (**Hagemann und Dümenil Gates, 2001**) (see also Sect. 4.3 and 4.4), the re-analysis biases partially accumulate in the simulated discharge. Here, ERA40 yields a more realistic discharge than the other two re-analyses.

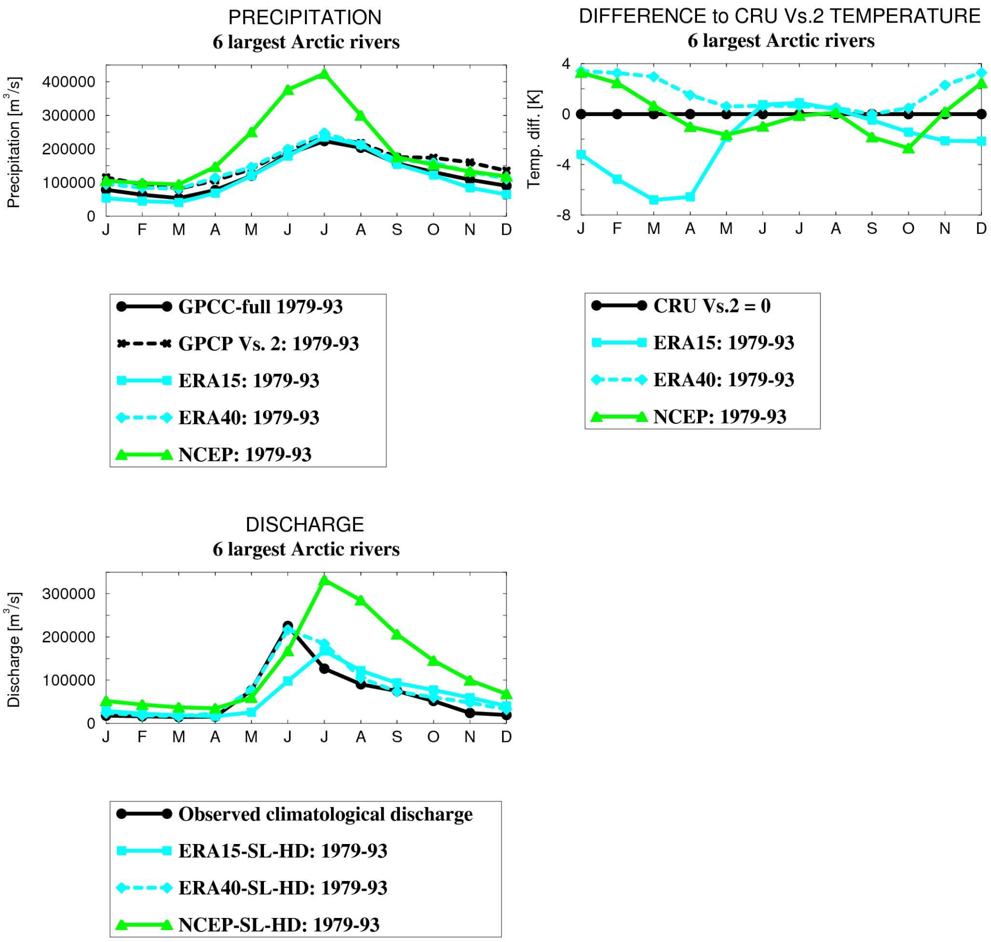


**Figure 8** Monthly mean precipitation (upper left panel), temperature differences to CRU2 data (upper right panel) and simulated discharge using SL scheme and HD model (lower panel) for the years 1979-1993 over the Arctic Ocean catchment represented by its six largest rivers (cf. Figure 2).

*Jeuken et al.* (1996) introduced the nudging technique as an alternative method of GCM validation. Nudging is the dynamical adjustment of a GCM with atmospheric fields taken from a re-analysis, such as, e.g., vorticity, divergence, temperature and surface pressure. Due to the nudging the GCM is drawn towards the re-analysis, so that usually an improved climate simulation is yielded by the use of a constrained atmospheric circulation when compared to the free GCM simulation. The nudging technique can be also considered as a second-step of data assimilation as the GCM does not directly assimilate observations but re-analysis data. An important outcome of the nudging simulation is the obtaining of tendency errors, i.e. the quantification of the tendency (nudging residuals) of the GCM to drift away from the atmospheric state imposed by the nudging. Using this outcome, a more precise estimation of the causes of climate model errors (such as revealed for RCMs in the MERCURE project; **Hagemann et al., 2004**) may be achieved by systematic initial tendency error (SITE) estimates (*Machenhauer and Kirchner,* 2000) using re-analysis data. SITE estimates can be used to assess errors in the model physics or to find missing external forcings. Hence, nudging has a variety of applications, e.g. *Kaas et al.* (1999) used the nudging technique to tune the parameterization of unresolved scale interactions, *Déqué et al*. (2000) used it as a method of GCM validation through short-range forecasts, and *Guldberg et al*. (2005) used the nudging for the reduction of systematic errors to analyse its impact on the skill of seasonal predictions.

An alternative method used with RCMs is spectral nudging in which the large-scale driving fields are allowed to force the low wave number component of the regional simulation in the higher altitudes throughout the entire domain (e.g. *Waldron et al.*, 1996; *von Storch et al*., 2000; *Radu et al.*, 2007). The advantage of this approach is the full consistency between the large-scale fields simulated by the RCM and those provided by the lateral boundary conditions. However, it can also prevent the formation of small-scale, surface-forced circulation systems which are not present in the driving field (*Giorgi and Mearns*, 1999).

Although most Numerical Weather Prediction (NWP) centres have incorporated land surface schemes in their models, errors in the NWP forcing accumulate in the surface and energy stores, leading to incorrect surface water and energy partitioning and related processes. This has motivated the NWP community to impose ad hoc corrections to the land surface states to prevent this drift. A methodology under development here is to implement a Land Data Assimilation System (LDAS), which consists of uncoupled models forced with observations, and is therefore not affected by NWP forcing biases (http://ldas.gsfc.nasa.gov). North American (NLDAS; *Mitchell et al.* 2004) and Global (GLDAS; *Rodell et al.*, 2004) LDAS systems are being developed that will lead to more accurate re-analysis and forecast simulations by numerical weather prediction (NWP) models. Specifically, these systems will reduce the errors in the stores of soil moisture and energy which are often present in NWP models and which degrade the accuracy of forecasts, and thus also the accuracy of re-analyses used in climate studies. The LDAS systems are currently forced by terrestrial (NLDAS) and space based (GLDAS) precipitation data, space-based radiation data and numerical model output. In order to create an optimal scheme, the projects involve several land surface models, many sources of data, and several institutions.

# 4. Improvements of parameterizations and evaluation methods

This section considers improvements yielded by the enhancement of model parameterizations and evaluation methods and their usage of observational data. The model parameterizations itself may be improved using knowledge obtained from new observational data (Sect. 4.1). On the other hand, observational data may also be evaluated using re-analysis data and/or independent model results (Sect. 4.2). The methods of model evaluation may be further developed in two ways. First, climate models may be extended so that more climatological variables are simulated, which then can be validated with observations that previously could not be used (Sect. 4.3). Second, by the improvement or utilization of observational datasets (such as re-analyses) to yield data that may be used for model validation studies (Sect. 4.4).

## 4.1. Model improvement by improving model parameterizations using new data

The availability of new data, especially from the increasing amount of satellite measurements, has the potential to further model improvements as the data can be used to improve climate model parameterization schemes. This is possible if the kind of data has not been available before or if already available data are produced with a much higher resolution. An example for the first kind is given by *Rechid et al*. (2008b) who parameterize the snow free surface background albedo as a function of the leaf area index using global distributions of soil albedo and vegetation albedo that they have derived from MODIS data (see Sect. 3.2). This albedo parameterization will also be used in the phenology that is currently being developed where the leaf area index will be interactively calculated within the climate model simulation (*Rechid et al*., 2008a). For some measurement programs, the improvement of model parameterizations is one of the major driving forces. For example, *Frühwald* (2000) stated that polarimetric radar data together with Doppler radar information may help to give hints for improving parameterizations of cloud micro-physical processes in coarse resolution atmospheric models. *Voyles* (2004) noted that data streams produced by the Atmospheric Radiation Measurement (ARM) Mobile Facility (AMF) are available to the atmospheric community for the use in testing and improving parameterizations in GCMs. Within the EU project CLOUDMAP2 (http://darc.nerc.ac.uk/Envisat/cloudmap2.htm), a main objective was to assess, qualitatively and quantitatively, how their cloud database could be used to improve the veracity and/or validation of NWP models, so that in the longer term this information will be used to improve the physical representation and parameterization of sub-grid scale characterisations of clouds within the typically coarse-resolution NWP models and GCMs.

In the second case, high resolution data may provide information on the sub-grid scale of a climate model that can be used to improve parameterizations representing sub-grid scale processes that are not resolved by the operational model resolutions. For example, the recent new global very high resolution datasets of land surface parameters based on satellite observations (e.g. see **Hagemann et al., 1999**) can be used to increase the applicability of existing parameterization schemes to finer resolutions, but they may also be used to improve the parameterization schemes themselves. This was done in the study of **Hagemann and Dümenil Gates (2003)** where the use of soil water capacities at a very high resolution led to the improvement of a surface runoff parameterization scheme. The improved parameterization scheme is a further development of the Arno scheme that is widely used in climate research, e.g., in the ECHAM model (*Roeckner et al.*, 1992), HIRHAM model (*Christensen et al.*, 1996), REMO model (*Jacob*, 2001), VIC model (*Liang*, 1994), and the Xinanjiang model (*Zhao*, 1977). Here, surface runoff is computed as infiltration excess from a "bucket" type reservoir which takes the sub-grid variability of soil saturation within a model gridbox into account. Instead of prescribing a distribution of sub-grid scale soil water capacities as it is done in the original Arno scheme, the array of high resolution soil water capacities (cf. Figure 6) taken from **Hagemann (2002a)** was used to obtain individual fractional saturation curves for each model gridbox. From each saturation curve, the three parameters required in the modified formulation of the scheme were derived via optimization. Figure 9 shows the fractional saturation curves for an example gridbox located in steep terrain. Here, the saturation curve yielded by the fitted shape parameter (dot-dashed curve) of the improved Arno scheme is much closer to the subgrid capacity distribution than the curve yielded by the purely orographic shape parameter from the original Arno scheme (turquoise curve). This will be the case for the majority of the gridboxes with heterogeneous subgrid capacity distributions. The improved Arno scheme has become operational in the REMO model since Vs. 5.7.

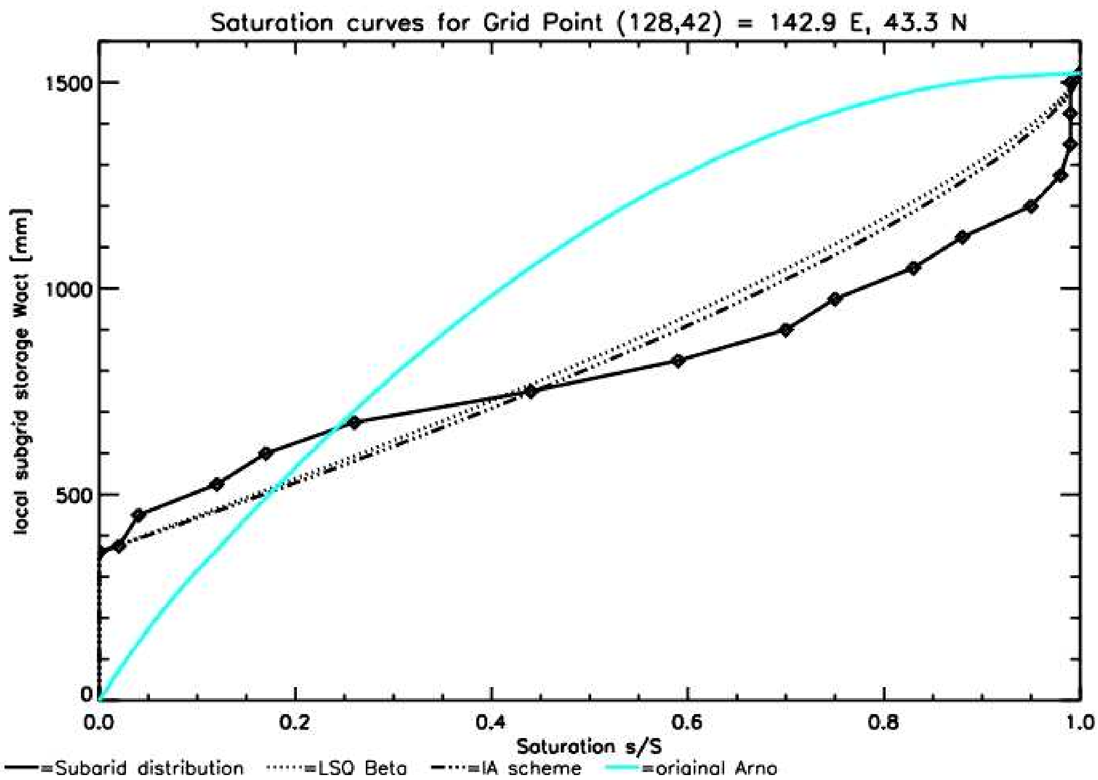


**Figure 9** Saturation curves for an example T106 gridbox at 142°E, 43.3°N (Japan).

## 4.2. Data evaluation using re-analysis data and/or independent model results

Re-analysis and climate model data can also be used for the evaluation of observational data. This evaluation may comprise a quality control of the observations as well as obtaining regional constraints for station data that determine limits in model resolution where the stations should not be used for model validation.

**Hagemann et al. (2003a)** give an example for this usage of model data. They have retrieved IWV from surface based GPS measurements of zenith path delay (*Gendt*, 1999) in order to apply a quality control to the IWV for each GPS station by comparing it with the IWV from the ECMWF operational analyses (*Hagemann et al.* ,2003b) have shown that the usage of ERA40 instead of the operational analyses yields very similar results). The zenith path delay values are converted into IWV using observed surface pressure and mean atmospheric water vapour column temperature obtained from the ECMWF operational analyses. Although the main objective of the study was to assess the usefulness of global GPS measurements for climate monitoring and model validation, **Hagemann et al. (2003a)** highlighted that also the analyzed fields can be used to identify errors in the GPS derived data and to identify areas where the GPS data are less relevant to use. They found several examples where the GPS derived data have systematic errors. For example, if the mean IWV bias between GPS derived IWV and analysed IWV for a station is larger than its standard deviation, this indicates a systematic error either in the zenith path delay measurements or in the surface pressure and its interpolation. This includes possible errors in the height that is assigned to the GPS or the pressure gauge. The approach to identify suspicious data is analogues to the methods applied in operational numerical weather prediction (e.g. *Hollingsworth et al.*, 1986). **Hagemann et al. (2003a)** also have identified areas where the numerical model has insufficient resolution to describe the water vapour profile due to sharp climate and weather boundaries. Typical cases are stations located at steep mountain slopes, or near major land ice areas such as Greenland or Antarctica. As an example, Figure 10 shows results from the station HOFN (Iceland) situated at the eastern coast near Mount Vatnajökull (2119 m). The ERA40 IWV is systematically smaller than the GPS derived IWV since the model is likely to represent conditions over the large glacier and not the conditions at the station. So in order to arrive at a representative sample of GPS station such errors or anomalies need to be identified. Using the four months considered in this study, it was possible to identify problematic stations that must be blacklisted in model validation studies at resolutions comparable to T106 or coarser. For studies of long-term changes in the IWV itself these stations can still be used.

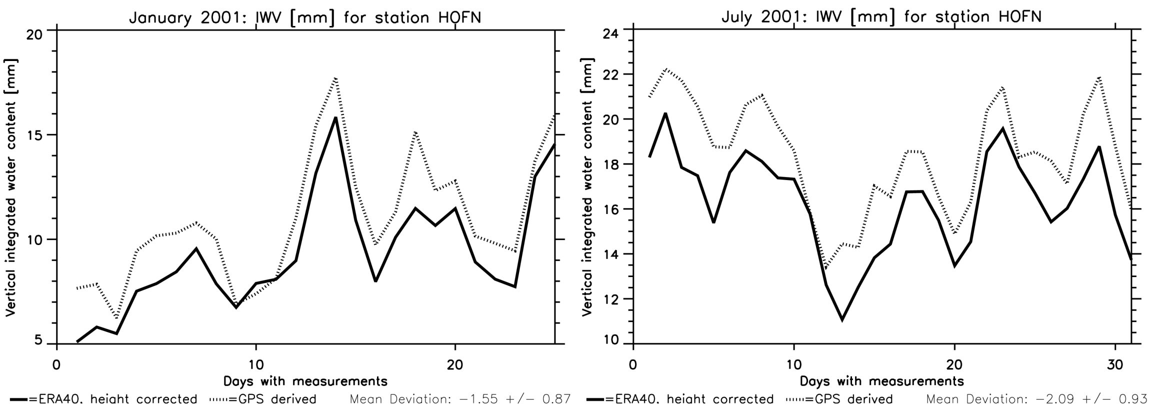


**Figure 10** GPS derived (dotted) and ERA40 (solid) IWV at station HOFN (Iceland) for a) January 2001, b) July 2001.

## 4.3. Extending a model to make use of more observations for evaluation

The extension of a climate model is twofold development. On one hand more processes are included into a climate model to represent and simulate more processes within the Earth system (see Sect. 5.1), on the other hand more observations may be used to validate the model. Although the inclusion of more processes often will raise the degrees of freedom of the climate model, especially if the newly introduced processes feed back to the model's simulated climate. The latter depends on the kind of processes, whether they are used for further calculations or only calculated diagnostically within a coupled system. Discharge, e.g., is currently a purely diagnostically resolved process in an atmosphere only climate model. In a coupled atmosphere-ocean model, the discharge is the interface between the land surface hydrology and the ocean, and thus an integral part of the coupled system.

As mentioned in Sect. 2.1.2d, climate model validation with observed discharge is an alternative to the direct validation of precipitation over land. It is possible to compare the model's total annual mean runoff to annual discharge observations for validation purposes. It is not feasible to compare the simulated runoff to observed discharge in time intervals of a season or less for big drainage basins. But this is an important validation task, especially regarding the timing of the runoff, e.g. timing of the snowmelt and the peak of snowmelt induced runoff. In order to perform an adequate validation of runoff processes, the runoff calculated by a GCM or RCM must be laterally transported over the land surface by a discharge model. The requirement for such a validation is that the climate model can be coupled to a discharge model (on- or offline). For the MPI-M climate models ECHAM and REMO this has been achieved with the HD model (**Hagemann und Dümenil, 1999**). For the ERA15 and NCEP re-analyses the direct application of the HD model was not possible so that the simplified land surface (SL) scheme was used to calculate the required input fields for the HD model from the re-analysis time series of precipitation and 2 m temperature (**Hagemann and Dümenil Gates, 2001**).

Sometimes it also makes sense to describe specific processes with a more physical exact formulation instead of using a simplified parameterization. In this case, more exact may mean temporally, spatially or numerically. This would allow the direct comparison to process-based measurement studies. In this respect also the regional downscaling of GCM data with a RCM (see Sect. 1) enhances the possibilities of model evaluation, especially if for a process under consideration the same process formulations are used in the GCM and RCM. Another example is the implementation of a cloud resolving model (CRM), such as conducted by *Khairoutdinov and Randall* (2001) who replaced their GCM's conventional convective and stratiform cloud parameterizations with a CRM, thereby allowing the explicit computation of the global cloud fraction distribution for radiation computations.

## 4.4. Using new methods to improve data or their usability for model evaluation

The improvement of existing observational data that may be used for climate model evaluation comprises two main issues. The first is the application of models or model algorithms to evaluation datasets to retrieve an improved dataset where the current data have some deficits. Here, especially re-analysis data are a candidate as their simulated hydrological cycle shows some problems and biases in its different components (**Hagemann and Dümenil**

**Gates, 2001)**. Thus, the application of hydrological or land surface models to re-analysis datasets may lead to an improved simulation of hydrological and land surface fields compared to the re-analysis data. This strategy is being extensively used in the GSWP (see Sect. 2.1.2). *Sheffield et al.* (2006) have developed a 50-years global 1.0° dataset of meteorological forcings by combining a suite of global observation-based datasets with the NCEP re-analysis that is supposed to be used to drive land surface hydrology models. **Hagemann et al. (2005)** have shown that the application of the Simplified Land surface (SL) scheme (**Hagemann and Dümenil Gates, 2003**) to the ERA40 data has lead to an improved simulation of annual evapotranspiration and runoff over many large catchments of the globe. This can be seen in Figure 11 where the bias in the runoff coefficient is shown for ERA40 and values simulated by the SL scheme using ERA40 precipitation and 2 m temperature as input.

The second is the utilization of observations by deriving quantities from the data that are also simulated by a climate model. This application comprises the increasing utilization of satellite data where the measured radiances or signals are used to calculate all sorts of different atmospheric and land surface variables, such as precipitation (cf. Table 1: in Sect. 2.1.1), evaporation, IWV (cf. Table 2: in Sect. 2.1.2), land use (cf. Sect. 3.2), albedo, FPAR (fraction of photosynthetically active radiation, e.g. from MODIS data; *Myneni et al.*, 2002), and terrestrial water storage (from GRACE data; *Lettenmaier and Famiglietti*, 2006). In this respect, **Hagemann et al. (2003a)** developed a method to retrieve IWV from surface based GPS measurements of zenith path delay (see Sect. 4.2).

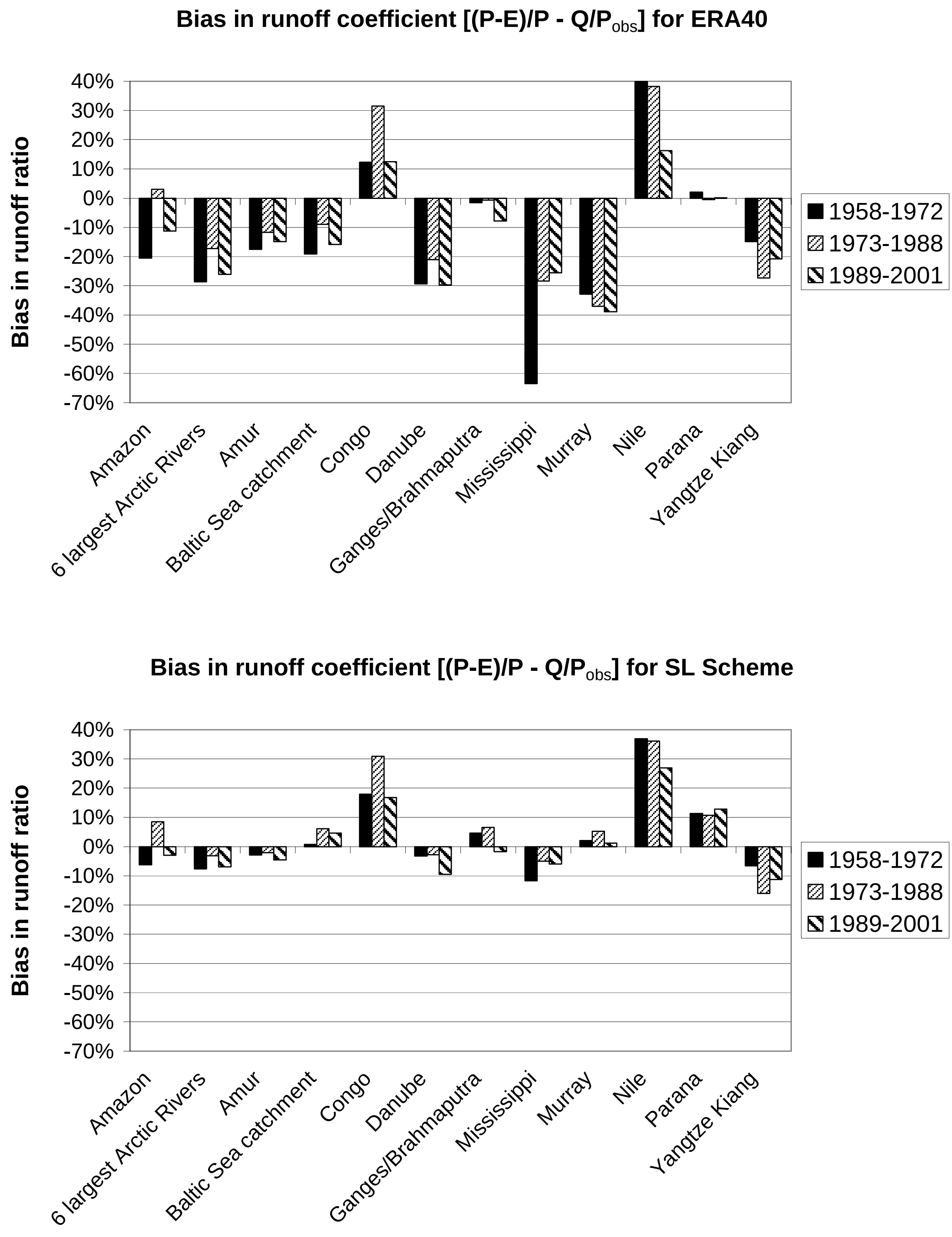


**Figure 11** Bias in runoff coefficient ([precipitation minus evapotranspiration] divided by precipitation) of ERA40 (upper panel) and simulated with the SL scheme (lower panel) for large river catchments. Observed values for runoff were taken from climatological discharges, for precipitation they comprise GPCC (1989-2001) and CRU Vs. 1 data (1958-1972, 1973-1988).

# 5. Challenges in modelling future climate changes

In this section, challenges are highlighted that climate change research is currently facing not only in general, but also with respect to the hydrological cycle. A few examples of recent or on-going projects and activities dealing with these challenges will be given in all sub-sections.

In the ensemble of climate simulations conducted for the IPCC 4AR, several processes and sources of future radiative forcing are not accounted for, including those from land use change, variations in solar and volcanic activity (*Kettleborough et al.*, 2007). Most of the current climate models only fulfil the minimum requirements for the use in long-term climate change studies (such as, e.g., stated by *Grassl*, 2000), i.e. they consists of a 3D coupled atmosphere-ocean GCM that includes sea ice dynamics and a (often comparably simple) representation of land surface processes. The introduction of further processes into a climate model system will require renewed model evaluation and may create necessities to obtain new observational datasets. Sect. 5.1 will shortly look upon these steps that will lead the current climate models into a stage where they become comprehensive Earth System Models (ESMs).

Each climate change simulation is afflicted with distinct uncertainties, which have to be taken into account and preferably reduced in the analysis of the simulation results. Sect. 5.2 tackles issues of uncertainty and gives some examples on recent studies about these issues. These studies often consider only changes in the mean climate. In this respect, changes in the variability of climate in the future are more uncertain than changes in the mean, as larger samples are required to quantify changes at the tails of the frequency distribution. But the latter could have very significant impacts on lives and livelihoods. Climate change is likely to increase the costs and impacts imposed by extreme weather, both by shifting the temperature probability distribution upwards and by intensifying the water cycle, so that severe floods, droughts and storms occur more often. Consequently research is needed to better assess the future probability of extreme events in the different regions of the Earth. Sect. 5.3 shortly looks on recent and on-going research on hydrological extremes. As mentioned above, the role of future land use change was not specified in the current climate projections, which adds also to the uncertainty in these projections. Thus, this role is considered in several on-going activities such as presented in Sect. 5.4.

## 5.1. The hydrological cycle within a comprehensive Earth system model

Feedbacks between the climate and hydrology occur (*Claussen et al.*, 2004). The snow/climate feedback, e.g., is well known and described. However, feedbacks between $CO_2$ increases, vegetation, soil moisture and climate are less well understood and are not well described in most climate and hydrological models. The investigation of these feedbacks requires the coupling of different processes and compartments of the Earth system. Thus, in order to cover all aspects of the Earth system and its future changes the current climate models (as described in Sect. 5) are gradually evolving into comprehensive ESMs. The development of complex ESMs is a major enterprise conducted at the international level, and specifically in Europe (Max Planck Society in Germany, Hadley Centre in the UK, Institut Pierre Simon Laplace in France), in the US (NSF/NCAR, NASA, DOE, NOAA) and in Japan (Frontier Program). In Germany, the COSMOS (Community Earth System Models) network was initiated to build up community ESMs at the European level (http://cosmos.enes.org). ESMs integrate our knowledge regarding the atmosphere, the ocean, the cryosphere and the biosphere as well as the anthroposphere, and account for the coupling between physical and biogeochemical processes in these components of the Earth System. ESMs are needed to understand large climate variations of the past and to predict future climate changes. International programs, including WCRP and IGBP, coordinate Earth system modelling initiatives, e.g. through their Global Analysis, Integration and Modelling (GAIM; http://gaim.unh.edu) project.

The advancement from a climate model into an ESM means that more and more processes will be implemented into the modelling system that is used to simulate Earth's climate. Previously a comparable step has been accomplished when the expansion from atmosphere only GCMs to coupled atmosphere-ocean GCMs was carried out. Currently, ESMs are utilized to include an interactive vegetation and the closed carbon cycle (e.g. *Cox et al*., 2004; *Wetzel et al*., 2006; *Raddatz et al*., 2007) as well as sophisticated aerosol models especially to adequately represent aerosol-cloud interactions (e.g. *Stier et al*., 2006, *Tost et al*., 2007). Further work is being conducted to couple atmospheric chemistry, air pollution or desert dust modules to a number of GCMs and RCMs (e.g. *Brasseur et al*., 2006; *Jöckel et al*., 2006; *Langmann*, 2000; *Zakey et al.*, 2006), and to include land use changes (see also Sect. 5.4) and vegetation dynamics, e.g., to investigate whether there will be a future dieback of Amazonian rain forest due to climate change as simulated by *Cox et al*. (2004). In the future more processes may be added that involve the biosphere or that may even come from the socio-economic side to couple the anthroposphere to the climate system. In order to obtain an integrated assessment of climate policies, *Bahn et al.* (2006) established a two-way coupling between the economic module of a well-established integrated assessment model and an ESM of intermediate complexity (EMIC; see, e.g. *Claussen et al.*, 2002). They showed that further applications of their method could include the coupling of an economic model and an advanced ESM that is able to describe the carbon cycle.

Even if some components are not fully implemented into an ESM, the impact of certain effects on hydrology may be investigated. For example, the impact of a large stratospheric sulphur loading on the hydrological cycle is currently under investigation by *Timmreck and Hagemann* (2009) who carried out a series of Mt. Pinatubo experiments with the coupled Atmosphere-Ocean GCM ECHAM5/MPIOM. Here, they have conducted a number of 10-member ensemble simulations with three different initial conditions starting in January and June with and without prescribed Pinatubo aerosol forcing. Figure 12 shows that in all cases the aerosol forcing leads to global cooling and a reduction in evaporation, which can be both attributed to the reduced incoming solar radiation reaching the surface. The reduced temperatures are leading to a reduced moisture holding capacity of the atmosphere and, thus, to a clear reduction in the integrated water vapour. These large-scale effects are superimposed to the first indirect effect of aerosols on the cloud condensation nuclei. (Note that the GCM accounts for the direct and first indirect effect of aerosols.) In total, the aerosol forcing causes a reduction of global precipitation.

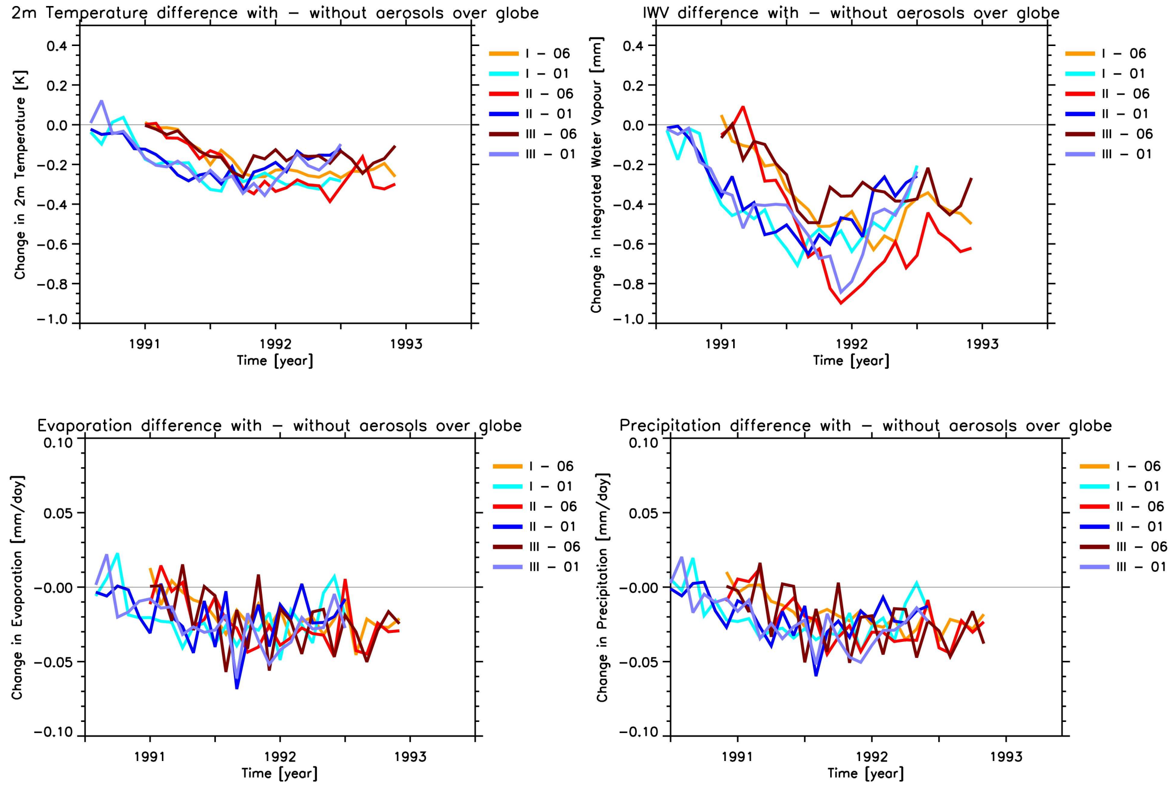


**Figure 12** Differences in the global average of 2 m temperature (upper left), integrated water vapour (upper right), evaporation (lower left) and precipitation (lower right) between the ensemble means of simulations with and without Pinatubo aerosol forcing. The bluish curves denote the three cases with initial conditions of January, the reddish curves the three cases with initial conditions of June.

With respect to the hydrological cycle, processes will be implemented that are usually not included in the current state of the art climate models. These processes comprise permafrost, wetland dynamics, irrigation, and the dynamical expansion and retreat of glaciers. With regard to the latter *Kotlarski* (2007) has implemented a dynamical glacier module into the RCM REMO and successfully applied over the Alpine region. In a second step, this module is now under application in the Himalayan region (F. Saeed, personal communication, 2008).

Permafrost and wetlands are two focal points in the coupling of hydrology to biogeochemical processes under climate change conditions. A large part (~24%) of the terrestrial land surface is underlain by permafrost that is mainly situated in high latitudes. Here, climate warming is more pronounced than elsewhere, and is very likely to continue to do in the future according to *IPCC* (2007). Permafrost soils build a globally relevant carbon reservoir as they store large amounts of deep-frozen organic material with high carbon contents. If permafrost melts under global warming conditions, the stored carbon can be decomposed and released to the atmosphere as additional greenhouse gas, which will lead to positive feedback. Consequently, relevant scientific questions are: How fast, how deep and to what temperature are permafrost soils going to thaw in the future? Thawing permafrost will also contribute to the formation of wetlands that currently cover about 6-8% of the land surface. They store water, regulate river discharge and provide a huge area for maximum evapotranspiration. The extension of wetlands determines the area where anoxic decomposition instead of oxic decomposition is taking place. While $CO_2$ is released under oxic conditions, the anoxic decomposition yields methane that is a far more active greenhouse gas than $CO_2$. Thus, an increase in wetlands area may lead to an enhanced methane production. On the other hand, a decrease will reduce moisture fluxes to the atmosphere and may lead to a reduction in precipitation. Thus, their future development is of major interest in climate change studies.

Consequently the two topics of permafrost and wetlands play also an important role within the ENIGMA project of the Max Planck Society that furthers the cooperation and common development of ESM components between MPI-M, the Max Planck Institute for Chemistry, the Max Planck Institute for Biogeochemistry, and the Potsdam Institute for Climate Impact Research as an associated partner. At MPI-M, the physical representation of permafrost (T. Blome, personal communication, 2008) and wetlands (T. Stacke, personal communication, 2008) within climate models is currently under development.

All new processes that are implemented into the climate model may change the quality of the simulated hydrological cycle and require renewed model validation and evaluation of the new model component as well as of the whole coupled model. For new model components, especially for those that directly add new hydrological processes, new or other kinds of observational data than commonly used, may be required for the model validation, initialization, and for the determination of parameters used in the numerical formulation of these processes. In this respect, the evolving availability and diversity of remote sensing data from satellite measurements enhance the probability that these future data requirements can be fulfilled.

## 5.2. Uncertainty of projected hydrological changes

If a single climate model simulation is considered, its results enclose different kinds of uncertainty. There is (1) uncertainty due to the use of one specific climate model as each GCM or RCM uses different techniques to discretize physics and dynamics and to parameterize sub-grid effects and, hence, has different model errors, (2) uncertainty in the prescribed future boundary conditions such as greenhouse gas and aerosol concentrations, which are usually based on the different IPCC SRES scenarios (*Nakićenović et al.*, 2000) and land use (see Sect. 5.4), (3) uncertainty due to natural climate variability, and (4) for RCMs the uncertainty in the GCM forcing at the lateral boundaries. In this respect, the usage of different scenarios is not causing a real uncertainty but rather spanning a range of possible futures that might become reality under the set assumptions. The importance of the sources of uncertainty varies between GCMs and RCMs (see *Déqué et al.*, 2007), and also depends on the climatological field, the region and the season. Results of *Déqué et al.* (2007) indicated that regarding uncertainty based on several models, the number of GCM forcings involved is at least as important as the number of RCMs, and that it is also necessary to consider several scenarios, at least in the case of future southern Europe summer warming.

In order to analyse these uncertainties more thoroughly, several approaches have recently been undertaken. The use of ensembles of GCMs developed at different modelling centres has become established in climate prediction/projection on both seasonal-to-interannual and centennial time scales (*Meehl et al.*, 2007). To the extent that simulation errors in different GCMs are independent, the mean of the ensemble can be expected to outperform individual ensemble members, thus providing an improved 'best estimate' forecast. Results show this to be the case, both in verification of seasonal forecasts (*Palmer et al.*, 2004; *Hagedorn et al.*, 2005) and of the present-day climate from long-term simulations (*Lambert and Boer*, 2001). However, members of a multi-model ensemble share common systematic errors (*Lambert and*

*Boer*, 2001), and cannot span the full range of possible model configurations due to resource constraints. Using a composite measure of model performance, *Reichler and Kim* (2008) found that the multi-model mean of the IPCC 4AR models usually outperforms any single model and it performs nearly as well as the NCEP re-analysis. They concluded that multi-model ensembles are a legitimate and effective means to improve the outcome of climate simulations. The reason for the superiority of the multi-model mean compared to any individual model is not clear, but a possible explanation is that the model solutions scatter more or less evenly about the truth (unless the errors are systematic), and the errors behave like random noise that can be efficiently removed by averaging. Such noise arises from the simulated internal climate variability, and probably to a much larger extent from uncertainties in the formulation of models (*Reichler and Kim*, 2008).

Within the framework of the 4AR of the IPCC (*IPCC*, 2007) and thereafter, further studies were conducted that considered the results from the GCM multi-model ensemble over several large regions of the globe focussing on the hydrological cycle, e.g.: Results of *Previdi and Liepert* (2007) suggest that future changes in the hydrologic cycle are likely to be strongly influenced by atmospheric dynamics. *Giorgi* (2006b) identified climate change hot spots, i.e. regions on the globe that are most responsive to climate change with regard to changes in the mean and interannual variability of precipitation and surface air temperature. *Nohara et al.* (2006) investigated the projections of river discharge for 24 major rivers in the world during the 21st century simulated by 19 GCMs based on the A1B scenario. To reduce model bias and uncertainty, they used a weighted ensemble mean for their multi-model projections of discharge. They found projected increases of discharge in high-latitude rivers (Amur, Lena, Mackenzie, Ob, Yenisei, and Yukon), where also the peak timing shifts earlier because of an earlier snowmelt caused by global warming. Discharge tends to decrease for the rivers in Europe to the Mediterranean region (Danube, Euphrates, and Rhine), and southern US (Rio Grande). *Milly et al.* (2005) showed that an ensemble of 12 GCMs exhibits qualitative and statistically significant skill in simulating observed regional patterns of 20th century multi-decadal changes in river runoff. These models project 10–40% increases in runoff in eastern equatorial Africa, the La Plata basin and high-latitude North America and Eurasia, and 10–30% decreases in runoff in southern Africa, southern Europe, the Middle East and mid-latitude western North America by the year 2050. Such changes in sustainable water availability would have considerable regional-scale consequences for economies as well as ecosystems.

Currently there are increasing scientific activities on uncertainties in climate projections arising from natural variability, parameter uncertainty and model diversity. Initiatives like the EU project ENSEMBLES (*Hewitt*, 2005) are beginning to produce probabilistic rather than deterministic predictions of climate change. In this respect, *Murphy et al.* (2004) and *Stainforth et al.* (2005) constructed large ensembles by perturbing poorly constrained parameters in the atmospheric GCM HadAM3 (*Pope et al.*, 2000) coupled to a mixed layer ocean. This “perturbed physics ensemble” approach is using the fact that GCMs are generally built by selecting components from a pool of alternative parameterizations, each based on a given set of physical assumptions and including a number of uncertain parameters. In principle, the range of predictions consistent with these components could be quantified by constructing very large ensembles with systematic sampling of multiple options for parameterization schemes and parameter values, while avoiding combinations likely to double-count the effect of perturbing a given physical process. A similar approach using the coupled atmosphere/ocean GCM ECHAM5/MPIOM (*Jungclaus et al.*, 2006) is currently under investigation within the MPI-M’s working group on uncertainty. Here, *Haerter et al.*

(2009) have studied the uncertainty of the sulfate aerosol radiative forcing due to parametric uncertainty by perturbing seven cloud-related parameters.

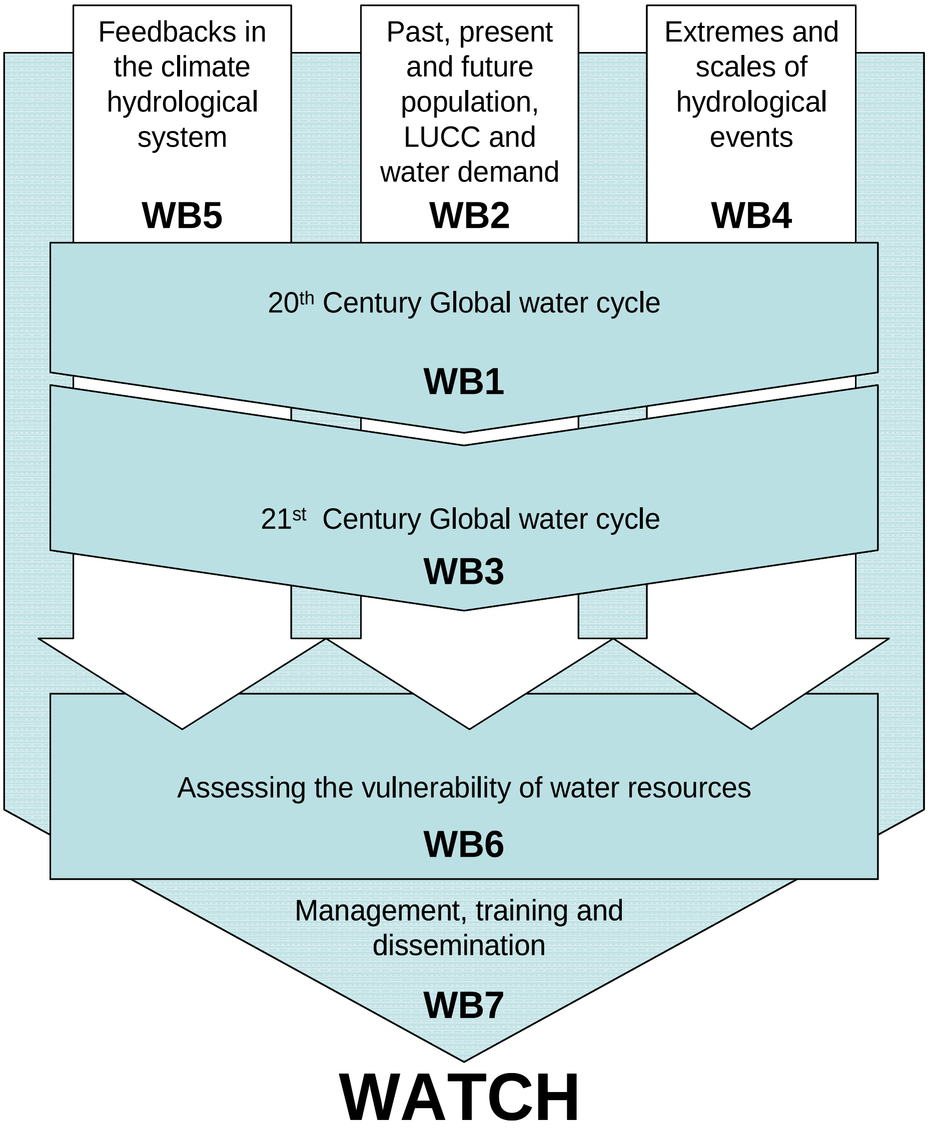


**Figure 13** Structure of WATCH (Coordinator: R. Harding, CEH): its six science work blocks consist of three main blocks (horizontal bars) providing an assessment of current (WB1: led by G. Whedon, UKMO) and future (WB3: S. **Hagemann,** MPI-M) water cycles and water resources (WB6: P. Kabat, WUR). Cross-cutting themes (vertical bars) support these with respect to the representation of feedbacks (WB5: E. Blyth, CEH), detection and attribution of extremes (WB4: van Lanen, WUR, and L. Tallaksen, UIO), and provision of dynamics of population, land-use change and water demands (WB2: D. Wiberg, IIASA).

Within the EU project WATCH (www.eu-watch.org, Figure 13), several tasks deal with uncertainty and the uncertainty transfer when climate model output is used to force hydrological models. *Schwierz et al.* (2006) reported that a hierarchy of models of varying complexity is a powerful approach to estimate and assess uncertainty, while the combination of different kinds of models of different complexity with an overlap between the model evaluations can contribute to the quantification and reduction of uncertainties from future climate model projections.

A major effort to understand uncertainties in regional climate modelling has been conducted in the EU project PRUDENCE (cf. Sect. 2.2). Here, 10 RCMs were forced with observed SST and lateral boundary conditions provided by the GCM HadAM3H (*Pope et al.*, 2000). Within PRUDENCE several studies using the output from the RCM ensemble were conducted, among those **Hagemann and Jacob (2007)** evaluated the simulated hydrological cycle of the 10 RCMs and the reduction of uncertainty by their multi-model ensemble mean over the catchments of the Baltic Sea (land area only), Danube and Rhine. They found that despite of the large differences in the control simulations of the RCMs, where the performance of the RCMs is different over the diverse catchments, the A2 climate change signal is very much confined and similar for almost all of the models. And even those RCMs who particularly disagree with regard to P and E in the control simulations (see, e.g., the annual P-E in Figure 14, upper panel), the A2 signal in the river runoff is largely constrained by each of the models (Figure 14, lower panel). This provides some confidence in the future projections even if only a few of the 10 RCMs are considered.

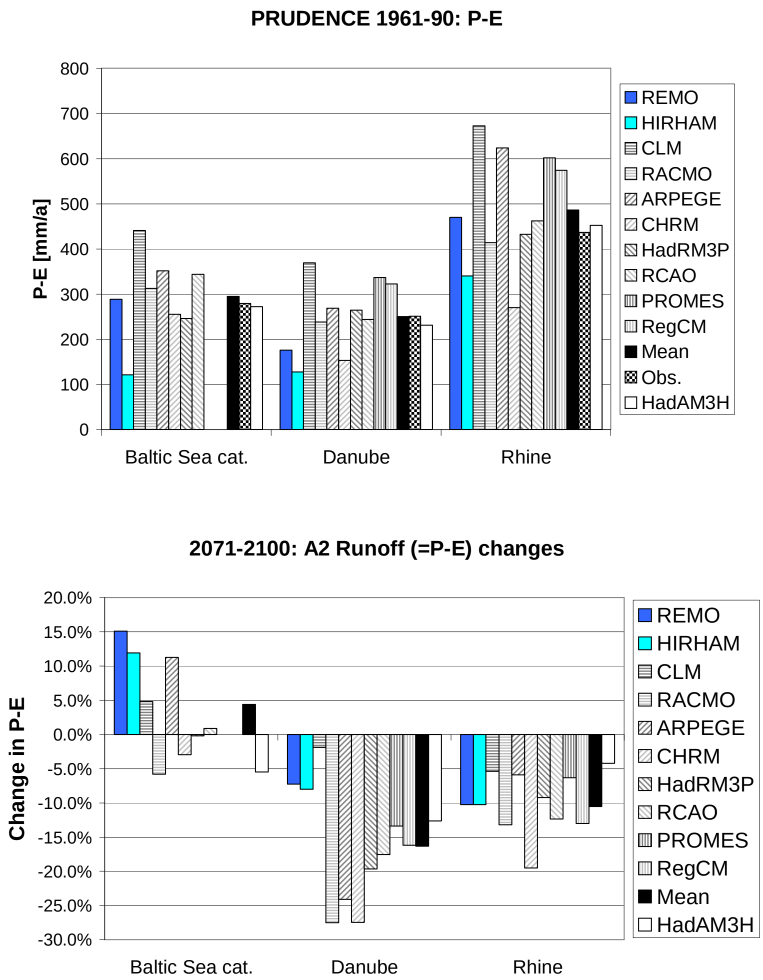


**Figure 14** Annual mean P-E (precipitation minus evapotranspiration = runoff) for the control period 1961-1990 (upper panel) and annual mean changes in P-E (lower panel) over the catchments of Baltic Sea, Danube and Rhine. In the upper panel, the observed P-E corresponds to the observed climatological discharge.

Further PRUDENCE studies focusing on the hydrological cycle over specific catchments comprise studies on the Lule river *(Graham et al.*, 2007a), the Baltic Sea, Bothnian Bay and

Rhine (*Graham et al.*, 2007b), Baltic Sea and Danube (*Hirschi et al.*, 2007; other European rivers were combined into three large-scale domains), the Baltic Sea, (*Kjellström and Ruosteenoja*, 2007), and the Rhine (*van den Hurk et al.*, 2005). In addition, *Vidale et al.* (2007) analysed the projected increases in Central European summer climate variability and found some evidence that the change in variability may be linked to the dynamics of soil-moisture storage and the associated feedbacks on the surface energy balance and precipitation.

How RCM predictions behave using different scenarios and different GCM forcing is currently being investigated within ENSEMBLES. Here, it will be of interest to determine whether using several RCMs with different GCM forcings actually results in more confidence in the overall results. First results considering two different scenarios and two different GCM forcings were obtained with the RCM RCAO (*Räisänen et al*., 2004) within the PRUDENCE project. Here, the four simulations agreed on a general increase in precipitation in northern Europe especially in winter and on a general decrease in precipitation in southern and central Europe in summer, but the magnitude and the geographical patterns of the change differ markedly between the two GCM forcings. *Rowell* (2006) made an initial attempt to estimate the uncertainty that arises from typical variations in RCM formulation, focussing on projected changes in surface air temperature and precipitation over the UK. It was found that the largest source of uncertainty, for both variables and in all seasons, is the formulation of the forcing GCM.

Within PRUDENCE, *Rowell* (2005) firstly analysed the results of a 3 member ensemble (3*control, 3*A2 scenario) of 30 year time slice simulations conducted with the GCM HadAM3P (*Pope et al*., 2000) regarding statistical significance of projected seasonal changes in temperature, precipitation and snow mass over Europe. Here, the mean precipitation anomalies in the future scenario are dominated (to first order and in all seasons) by a large-scale pattern of enhanced precipitation in the north and reduced precipitation in the south. However, the boundary between these two regimes displays a sizable annual cycle, such that it is located at about 40°N in winter, 45°N in spring, 60°N in summer and 55°N in autumn. The very recent WATCH-related study of **Hagemann et al. (2009)** used an ensemble of 12 transient coupled atmosphere-ocean GCM simulations (3*control, 3 for each of B1, A1B and A2) of ECHAM5/MPIOM and 8 RCM simulations (3*control, 3*A1B, 1*B1, 1*A2) of REMO to investigate how robust the projected changes in the hydrological cycle of the MPI-M climate models are compared to the natural climate variability as represented in these models. The study also addresses the question whether the robustness of the climate change signal differs between the GCM and the RCM forced by the GCM, thereby focusing on large European catchments. It was found that the better description of surface processes, higher resolution and non-linear scale interactions in the RCM gives a better representation of present day climate and hence a more credible climate change projection than the GCM. This is even along the lines of thoughts provided in the IPCC AR4 global and regional climate change chapters (*IPCC*, 2007). Over the Baltic Sea catchment, the RCM has an improved representation of the land sea contrast, and, hence, improved related moisture transport processes between water and land areas. Over the Danube and Rhine catchments, the better distribution of soil moisture leads to an improved representation of soil moisture feedbacks to the atmosphere. The latter is shortly illustrated in the following.

Over the Danube (see Figure 15) and Rhine catchments, noticeable differences in the robustness of the climate change signals between the GCM and RCM simulations are related to a stronger warming of about 1 K projected by the GCM in the summer. This is associated with a stronger projected summer drying in the two catchments (see Figure 15 for Danube precipitation changes). Figure 16 shows that the coupled GCM ECHAM5/MPIOM has a relatively strong summer drying problem in both catchments in the control period 1961-1990, which is consistent with the behaviour of the atmospheric GCM ECHAM5 forced by observed SST, as shown for the Danube by **Hagemann et al. (2006)**. The problem is much less pronounced in the RCM, which even shows some overestimation of summer rainfall over the Rhine catchment. Within PRUDENCE, results of **Hagemann and Jacob (2007)** indicated that the use of RCMs can overcome problems that a driving GCM might have with the representation of local scale processes or parameterizations. This supports that the RCM has the potential for an improved simulation of soil moisture feedbacks to the atmosphere, which in turn leads to the lower projected summer time warming and drying than projected by the GCM.

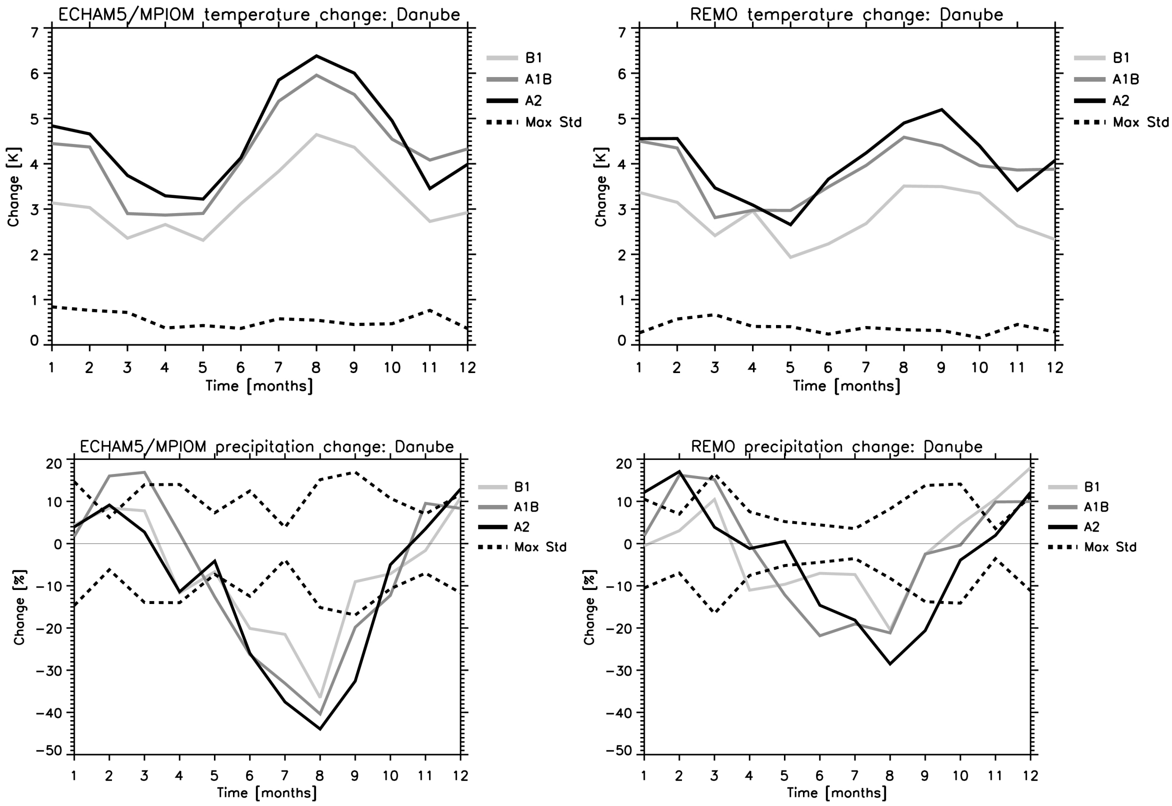


**Figure 15** Monthly mean temperature (upper panels) and precipitation (lower panels) changes (2071-2100 compared to 1961-90) over the Danube catchment as projected by the GCM ECHAM5/MPIOM (left panels) and the RCM REMO (right panels). Max Std denotes the maximum spread *S* for the corresponding ensembles.

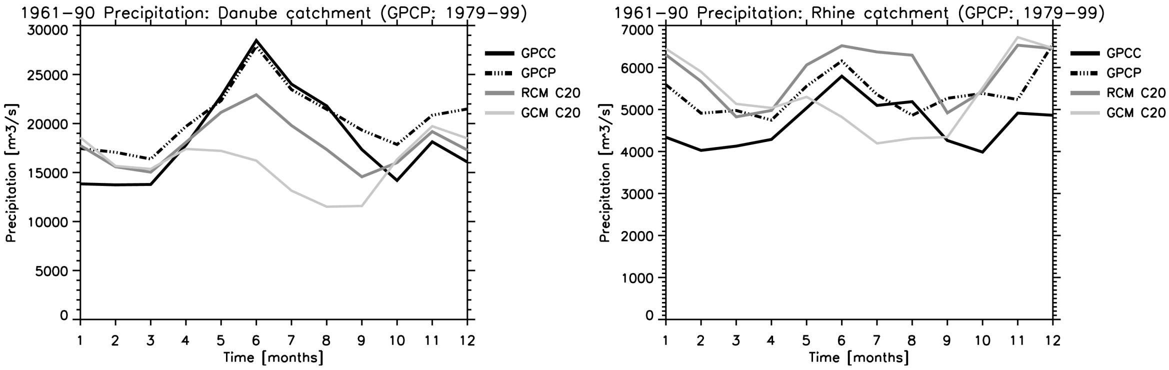


**Figure 16** Observed (GPCC and GPCP data) and simulated monthly ensemble mean precipitation over the Danube (left panel) and Rhine (right panel) catchment for the control climate 1961-90. Note that for GPCP, data for the control climate were not available so that the period 1979-99 was used. It was chosen to show both observations to reflect the uncertainty in precipitation datasets.

## 5.3. Changes in hydrological extremes

Extremes in the hydrological cycles mainly involve extreme (high or low) precipitation events. The assessment of changes in the risk of extreme precipitation events must take into account the fact that these are governed by different physical constraints than the time-averaged precipitation response (e.g. *Allen and Ingram*, 2002). As a result, a change in the risk of extreme high precipitation may emerge before a change in mean precipitation (e.g. *Hegerl et al.*, 2006). Thus, the methods to detect and quantify changes in hydrological extremes and their significance must use different techniques than the investigation of changes in the mean. Especially for the assessment of changes in hydrological extreme events, a probabilistic framework is required as these events belong to the tails of the frequency distribution where the number of occurrences is much lower than in the middle.

Research on climate extremes and their future development is still in its infancy, although there has been a large recent increase in the available analyses of changes in extremes. This allows for a more comprehensive assessment for most regions, especially concerning heat waves, heavy precipitation and droughts. Despite these advances, specific analyses of models are not available for some regions; in particular, projections concerning extreme events in the tropics remain uncertain. The difficulty in projecting the distribution of tropical cyclones adds to this uncertainty. Changes in extra-tropical cyclones are dependent on details of regional atmospheric circulation response, some of which remain uncertain (*Christensen et al.*, 2007b).

Several recent European projects have partially or fully concentrated on future changes in climate extremes over Europe, three of them have formed a cluster: PRUDENCE (see *Beniston et al.*, 2007 as well as Sect. 2.2 and 5.2), MICE (Modelling the Impact of Climate Extremes; *Hanson et al.*, 2007) and STARDEX (Statistical and Regional Dynamical Downscaling of Extremes for European Regions; see, e.g., *Goodess et al.*, 2007). MICE considered the potential impacts of climate change on a range of economic sectors important to specific regions, thereby focussing on changes in temperature, precipitation and wind extremes. The research programme had three main themes – the evaluation of climate model performance, an assessment of the potential future changes in the occurrence of extremes, and an examination of the impacts of changes in extremes on six activity sectors using a blend of quantitative modelling and expert judgement techniques. STARDEX has provided a rigorous and systematic inter-comparison and evaluation of statistical, dynamical and statistical-dynamical downscaling methods for the construction of scenarios of extremes. It has identified the more robust techniques that are recommended for the use in assessment studies of future changes in extremes. The research of the project cluster on extremes is continued in ENSEMBLES.

Due to their coarse resolution, GCMs are generally less suitable to quantify extremes. They may be used to consider the future development of specific large scale indices, such as, e.g. done by *Sillmann and Roeckner* (2008). But for a detailed analysis of extremes that are often spatially highly resolved, the use of downscaling tools, such as RCMs, is necessary. However, the current generation of GCMs and RCMs is unable to reliably reproduce historical hydrological extremes (with considerable variability in the prediction of rainfall patterns, differences between climate models and between different ensemble members of the same climate model). Here, WATCH (Figure 13) will bring together advanced statistical analysis tools to handle downscaling, uncertainty, vulnerability, spatial and temporal patterning and attribution through advanced fingerprinting techniques (e.g. *Allen and Tett*, 1999). WATCH will further the development of new methods of obtaining relevant means, extremes and uncertainties from climate model output for the use in future drought and flood assessments.

Related to WATCH, *Boberg et al.* (2007, 2008) have studied a methodology for assessing the skill of a RCM in describing the full distribution of intensities of a climate variable, such as precipitation, and the projected change in the distributions for an A2 climate scenario using the PRUDENCE RCM simulations. They found that most of the RCMs perform well in describing the distribution of precipitation, with a skill of around 0.8-0.9, where one is a perfect score and zero is no skill at all. For all RCMs, the comparison of the A2 scenario precipitation distributions with the control shows a shift to more extreme intensities, and a statistical study showed the changes in the shift to be significant up to extreme precipitation intensities of about 60 mm/day. The crossing point between a reduction of the lower precipitation intensities and an increase in the higher intensities is quite constant for the different RCMs (except for one).

The understanding of droughts (terrestrial processes and associated spatial and temporal pattern development, see *Tallaksen and van Lanen*, 2004) and the skills to forecast and predict droughts have substantially lagged behind developments in flood-related areas. The propagation of dry or wet anomalies through the hydrological cycle causes a deviation between the characteristics (onset, duration, severity) of a meteorological anomaly, a soil moisture anomaly and a hydrological anomaly (groundwater, surface water). The significance of the hydrological processes may vary as a function of the climatological regime and physical catchment structure (soils, hydrogeology) and is still not well understood (e.g. *Peters et al.*, 2003). Advancing on these issues is one of the objectives of WATCH. Methods exist at regional scales to estimate the probability of a specific area to be affected by a drought of a given severity, and to derive severity-area-frequency curves to assign return periods for historical events (*Hisdal and Tallaksen*, 2003). In WATCH these will be extended from regional to larger (global) spatial and temporal scales and from droughts to also treat large-scale floods in a similar analysis.

In order to validate models with regard to their representation of extremes, datasets are required that have a high spatial and temporal resolution. Here, especially daily time series of data are important, such as provided by *Klein Tank et al.* (2002) for precipitation and surface air temperature over Europe. Even though daily time series of hydrological data often exists, their availability is currently limited in many regions of the world. Thus, it must be an aim of science management and politics to improve the accessibility of these data for scientific purposes.

## 5.4. Land use change

While the process of anthropogenic emissions due to fossil fuel burning is fairly well established in state-of-the-art climate model simulations, up to now, the possible impact of land use changes on the climate is mostly neglected in long-term climate simulation. Changes in the land surface (vegetation, soils, water) resulting from human activities can affect the regional climate through shifts in radiation, cloudiness and surface temperature. Changes in vegetation cover affect surface energy and water balances at the regional scale, so that the impact of land use change may be very significant for the regional climate over time periods of decades or longer (*Denman et al.*, 2007).

Observations and model studies in tropical forests have shown effects of changing surface energy and water balance. For example, *Marengo and Nobre* (2001) found that the removal of vegetation led to decreases in precipitation, evapotranspiration and moisture convergence in central and northern Amazonia. *Oyama and Nobre* (2004) showed that the removal of vegetation in north-east Brazil would substantially decrease precipitation. Other model studies indicated that increased boreal forest reduces the effects of snow albedo and causes regional warming (*Denman et al.*, 2007). Related to the latter, e.g., *Göttel et al.* (2008) investigated the influence of changed vegetations fields on the projected regional climate over the Barents Sea region in an off-line coupling experiment with the RCM REMO and the dynamic vegetation model LPJ-GUESS (*Sitch et al.*, 2003). They projected a forest ratio increase and a shift of the tree line to higher altitudes and latitudes caused by a warmer climate with longer snow-free periods and growing season lengths. The feedback effects to the climate of these changes were one order of magnitude lower than the effects of the greenhouse gas forcing. A further warming in spring could be attributed to the snow-albedo effect, while a cooling in summer was dedicated to changes of roughness length, enhanced transpiration and changes in surface albedo.

Especially regions located in or close to areas with strong climatic gradients may be very sensitive to land use changes. These areas comprise tropical regions vulnerable to deforestation as well as arid and semi-arid regions. In this respect, Africa is one of the hot spot areas. So far, the effect of deforestation and reduced vegetation cover in Africa has mainly been studied with coarse-grid global climate models in the form of time slice experiments and idealized forcing (see e.g. *Feddema et al*., 2005). With these coarse resolution models, effects on the local and regional climate can usually not be resolved. For this purpose, RCMs are an adequate tool, such as done by *Paeth et al.* (2008) for West Africa who conducted long-term transient climate change experiments with the RCM REMO at 50 km resolution over West Africa where they forced their simulations with increasing greenhouse-gas concentrations and land use changes until 2050. Their results indicate that significant future changes in the near-surface climate may be caused by land use changes.

The BMBF project BIOTA-South (BIOdiversity monitoring Transect analysis in Africa; http://www.biota-africa.org/) is focussing in its 3$^{rd}$ phase on South Africa and Namibia. Future land use changes over South Africa will have two main drivers: human kind and climate. The projected future drying of the area will lead to soil degradation and an enhanced desertification. Anthropogenic changes are induced by manifold causes, such as deforestation, agriculture and infrastructure development. In order to investigate the impact of future land use changes on the climate these changes shall be estimated based on scenario of socio-economic development of the considered region and on the projected local regional climate changes obtained from REMO simulations conducted at MPI-M (A. Hänsler, personal communication, 2008). *Koster et al.* (2004) identified the Sahel zone as one of the hot spot areas for the feedback of surface soil wetness to subsequent rainfall. This is also the focus of the EU funded AMMA (African Monsoon Multidisciplinary Analyses; https://www.amma-eu.org/) project, which will produce new land surface and atmospheric parameterizations for the semi-arid region. In this semi-arid region, irrigation is not a major agricultural practice, but an increase in dryland agriculture is possible which is sensitive to rainfall totals. A previous study (*Taylor et al.*, 2002) showed that future likely changes in land-cover could result in a reduction of nearly 10% in rainfall. The LUCID (Land Use Change, Impacts and Dynamics) network aims at stimulating the joint research on land use in East Africa and its implications for land degradation, biodiversity, and climate change (http://www.lucideastafrica.org).

Studies on land use change and its impact on climate require both observational land cover datasets for the past and adequate projections for the future land use. The current general global land use and land cover datasets and the reconstructions of their changes (e.g. *Klein Goldewijk*, 2001) are generally less accurate and of coarser resolution than topic-wise datasets such as: the *JRC* (2004) GLC 2000 land cover regional and global classifications, the global land cover categorisation compiled by *IFPRI* (2002), the Forest Resources Assessment of *FAO* (2001), the global inventories of irrigated land (*Siebert and Döll*, 2001), etc. Therefore, one task of the EU project WATCH is combining the respective strengths of each of these datasets to produce a single consistent high quality, high-resolution (5’) product. WATCH will also provide projected land cover and land use distributions that will be derived on the basis of aggregate land use scenarios, reflecting trajectories of the key driving forces.

Both land use datasets for the past and future will be further used in WATCH for a number of studies. The latter will be used in off-line hydrological model simulations forced by climate model input. Here, the impact of the projected anthropogenic land use changes on the hydrological cycle (e.g. river discharge) will be compared to the impact of climate change alone, so as to derive an assessment of the relative importance of these processes. This way, the sensitivity of the global hydrological cycle to specific changes at the land surface as defined will be quantified. The impact of land-cover changes on the regional climate over West Africa will be studied using the RCM RegCM3 (*Giorgi and Mearns*, 1999) with improved land-surface parameterizations taken from the AMMA project. This will be used to back up the evidence of a feedback from land-cover change to rainfall (see above). As a specific land use, the feedback of irrigation onto rainfall (*ter Maat et al.*, 2006) will be analysed by making regional and global studies. The first will concentrate on areas where changes in irrigation coincide with areas of global hotspots for land surface – atmosphere feedbacks. The impact and scale of the irrigation on the regional energy and water balances will be quantified. The second will be made using a GCM including a simplified representation of irrigation to identify any influences beyond the region due to teleconnections within the global atmospheric system.

# 6. Summary

In this review the manifold possibilities how observational data can be used to improve climate models were presented, thereby it was focused on the hydrological cycle. Each possibility was shortly discussed using references to the current scientific literature. For several of them, the accompanying personal references give detailed examples on how I myself have contributed to the progress of research in this scientific area. These are summarized in Sect. 6.1, which is followed by a short outlook in Sect. 6.2.

## 6.1. Personal contributions

Several studies contributed to the task of model evaluation using data for validation. The models investigated comprise GCMs and RCMs as well as re-analysis data.

- For many variables, re-analysis datasets of past global numerical weather forecasts are good choices for the use as validation data, since many observations are entering the numerical weather forecast system via data assimilation. But regarding the hydrological cycle the current re-analysis data show a lot of problems, such as shown by **Hagemann and Dümenil Gates (2001)** for ERA15 and NCEP re-analysis, and by **Hagemann et al. (2005)** for ERA40. These studies raised the awareness that not all re-analysis variables are suitable as validation data.

- In their catchment-oriented validation study on the GCM ECHAM5, **Hagemann et al. (2006)** investigated the impact of model resolution on the hydrological cycle in a suite of model simulations with varying horizontal and vertical resolutions. It was found that a higher model resolution is generally improving the simulation of the hydrological cycle. Remarkably, in most of Earth's major river catchments, an increasing vertical resolution turned out to be more beneficial than increasing horizontal resolution.

- A major contribution to the understanding of the summer drying problem (**Hagemann et al. 2001)** over central and south-eastern Europe was provided, which is typical for many RCMs, and to a less extent is also visible in some GCMs. Within the MERCURE project, **Hagemann et al. (2004)** considered the water and energy cycles of five RCMs over the Danube catchment where they noticed that the summer drying problem was a major feature of all models except ARPEGE. They found two different reasons for problems in the RCM simulations. For ARPEGE and CHRM, the problems were related to deficiencies in the land surface parameterizations, while for HIRHAM, HadRM3H and REMO systematic errors in the dynamics appear to be causing the main errors in the simulations over the Danube catchment. **Hagemann and Jacob (2007)** evaluated the simulated hydrological cycle of the 10 PRUDENCE RCMs and found that the summer drying problem showed up again in the multi-model ensemble mean of precipitation over the Danube catchment, as only two of the 10 RCMs do not have this problem (ARPEGE and RegCM3). **Hagemann et al. (2009)** showed that this problem is even more severe in ECHAM5 than in REMO. The exact reasons for the summer drying problem are still not identified and are currently under investigation in the EU project CLAVIER (http://www.clavier-eu.org/).

Several improvements of parameterizations and evaluation methods were achieved.

- This comprises improvements of the model simulations by using new data for boundary conditions at the land-surface interface. Here, **Hagemann et al. (1999)** and **Hagemann (2002)** utilized satellite based very high resolution data to derive a global dataset of land surface parameters (LSP2) that is available for use in regional and global climate modelling. The LSP2 dataset has been implemented in the currently operational versions of the RCMs HIRHAM (*Christensen et al.*, 1996) and REMO (*Jacob*, 2001) as well as in the global ECHAM5 model (*Roeckner et al.*, 2003).

- It also comprises the improvement of model parameterizations using new data as done in the study of **Hagemann and Dümenil Gates** (2003) where the of use the LSP2 soil water capacities at a very high resolution lead to the improvement of the Arno scheme, which is a surface runoff parameterization scheme that is widely used in climate research. The improved Arno scheme has become operational in the REMO model since Vs. 5.7.

- With respect to the data evaluation using re-analysis data and/or independent model results, **Hagemann et al. (2003a)** give an example for this usage of model data. They have retrieved IWV from surface based GPS measurements of zenith path delay (*Gendt*, 1999). Although the main objective of the study was to assess the usefulness of global GPS measurements for climate monitoring and model validation, they highlighted that also the analyzed fields from the ECMWF operational analyses can be used to identify errors in the GPS derived data and to identify areas where the GPS data are less relevant to use. In addition, areas were identified where the ECMWF numerical model used for the operational analyses has insufficient resolution to describe the water vapour profile due to sharp climate and weather boundaries.

- Model evaluation procedures may be enhanced either by extending a model to make use of more observations for the evaluation, or by using new methods to improve data or their usability for model evaluation.
  - Firstly, in order to utilize river runoff data, the requirement is that the climate model can be coupled to a discharge model (on- or offline). For the MPI-M climate models ECHAM and REMO this has been achieved with the HD model (**Hagemann und Dümenil, 1999**). For the ERA15 and NCEP re-analyses the direct application of the HD model was not possible so that the Simplified Land surface (SL) scheme was used to calculate the required input fields for the HD model from the re-analysis time series of precipitation and 2 m temperature (**Hagemann and Dümenil Gates, 2001**).

  - For the latter, this can be achieved by applying specific algorithms or even models from other disciplines to the observational data. The first means the utilization of observations by deriving quantities from the data that are also simulated by a climate model. In this respect, **Hagemann et al. (2003a)** developed a method to retrieve IWV from surface based GPS measurements of zenith path delay (see also above). As an example for the second, **Hagemann et al. (2005)** have shown that the application of the SL scheme

(**Hagemann and Dümenil Gates, 2003**) to the ERA40 data has lead to an improved simulation of annual evapotranspiration and runoff over many large catchments of the globe.

A major effort to access uncertainties in regional climate modelling was undertaken.

- Within PRUDENCE, **Hagemann and Jacob (2007)** evaluated the simulated hydrological cycle of the 10 RCMs and the reduction of uncertainty by their multi-model ensemble mean over the catchments of the Baltic Sea (land area only), Danube and Rhine. They found that despite of the large differences in the control simulations of the RCMs, where the performance of the RCMs is different over the diverse catchments, the A2 climate change signal is very much confined and similar for almost all of the models. And even those RCMs who particularly disagree with regard to P and E in the control simulations, the A2 signal in the river runoff is largely constrained by each of the models. This provides some confidence in the future projections even if only a few of the 10 RCMs are considered.

- The very recent WATCH-related study of **Hagemann et al. (2009)** investigated how robust the projected changes in the hydrological cycle of the MPI-M global and regional climate models are compared to the natural climate variability as represented in these models. The study also addresses the question whether the robustness of the climate change signal differs between the GCM and the RCM forced by the GCM, thereby focusing on large European catchments. It was found that the better description of surface processes, higher resolution and non-linear scale interactions in the RCM gives a better representation of present day climate and hence a more credible climate change projection than the GCM. This is even along the lines of thoughts provided in the IPCC AR4 global and regional climate change chapters (*IPCC*, 2007). Over the Baltic Sea catchment, the RCM has an improved representation of the land sea contrast, and, hence, improved related moisture transport processes between water and land areas. Over the Danube and Rhine catchments, the better distribution of soil moisture leads to an improved representation of soil moisture feedbacks to the atmosphere.

## 6.2. Outlook

In the final section before the summary, challenges were highlighted that climate change research is currently facing within the area of this review. These challenges make clear that the necessity to use observational data for climate model improvement has not come to an end but is rather enhanced. This is caused by the extension of climate models with further compartments of the Earth system (development of ESMs, land use) as well as setting the focus on climate characteristics beyond the mean and standard deviation (uncertainty, extremes).

Consequently the ground is set for the use of the wide range of observational data that are or will be provided by various satellite missions that recently have been or soon will be launched. An overview of these missions is given in the Earth Observation Handbook (http://www.eohandbook.com/) provided by the Committee on Earth Observation Satellites (CEOS) and the European Space Agency (ESA). But it should not be forgotten to further maintain the existing network of meteorological and hydrological stations as only the interplay between conventional and satellite measurements can provide the observational framework that is required for a comprehensive climate model improvement.

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## List of abbreviations

| | |
|---|---|
| AMIP | Atmospheric Model Intercomparison Project |
| CMAP | CPC Merged Analysis of Precipitation |
| CRU | Climate Research Unit |
| DEM | Digital Elevation Model |
| ECMWF | European Centre for Medium-range Weather Forecast |
| ECHAM5 | Atmospheric GCM operational at MPI-M |
| EMIC | Earth system Model of Intermediate Complexity |
| ERA15 | ECMWF 15-years re-analysis (1979-1993) |
| ERA40 | ECMWF 40-years re-analysis (1958-2001) |
| ESM | Earth System Model |
| GCM | General Circulation Model |
| GEWEX | Global Energy and Water Cycle Experiment |
| GHG | Green House Gas |
| GLACE | Global Land-Atmosphere Coupling Experiment |
| GPCC | Global Precipitation Climatology Centre |
| GPCP | Global Precipitation Climatology Project |
| GPS | Global Positioning System |
| GWSP | Global Soil Wetness Project |
| HD model | Hydrological Discharge model |
| IGBP | International Geosphere-Biosphere Program |
| IPCC | Intergovernmental Panel on Climate Change |
| ISLSCP | International Satellite Land-Surface Climatology Project |
| IWV | Integrated Water Vapour |
| MERCURE | Modelling European Regional Climate: Understanding and Reducing Errors, EU project |
| MODIS | Moderate Resolution Imaging Spectroradiometer |
| MPI-M | Max Planck Institute for Meteorology |
| NAO | North Atlantic Oscillation |
| NCEP | National Centers for Environmental Prediction |
| PRUDENCE | Prediction of Regional scenarios and Uncertainties for Defining EuropeaN Climate change risks and Effects, EU project |
| RCM | Regional Climate Model |
| REMO | RCM operational at MPI-M |
| SL scheme | Simplified Land surface scheme |
| SST | Sea Surface Temperature |
| WATCH | WATer and global Change, EU project |
| WCRP | World Climate Research Programme |

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