Meteorologische Zeitschrift, Vol. **16**, No. 1, 071-085 (February 2007)

Article

Numerical simulations of precipitation in the complex terrain of Iceland – Comparison with glaciological and hydrological data

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(Manuscript received June 26, 2006; in revised form October 9, 2006; accepted October 27, 2006)

Abstract

Atmospheric flow over Iceland has been simulated for the period September 1987 through June 2003, using the PSU/NCAR MM5 mesoscale model driven by initial and boundary data from the European Centre for Medium-range Weather Forecasts (ECMWF). The simulated precipitation is compared with two types of indirect precipitation observations. Firstly, snow accumulation on two large ice caps in SE-Iceland and on two large glaciers in central Iceland. Secondly, model output is used as input to the WaSiM-ETH hydrological model to calculate and compare the runoff with observed runoff from six watersheds in Iceland for the water years 1987–2002. Model precipitation compares favourably with both types of validation data. The seasonal and inter-annual variability of precipitation is investigated at low as well as high altitudes. The simulations reveal a negative trend in the winter precipitation in W-Iceland, but a positive trend in the ratio of lowland precipitation to mountain precipitation in E-Iceland. There is in general a substantial inter-annual variability in the ratio of lowland precipitation to precipitation in the mountains, especially in E-Iceland, emphasizing the limitation of precipitation observations in the lowlands as a proxy for precipitation in the mountains. In order to assess the impact of orography on the precipitation climate of Iceland, precipitation is simulated with flat Iceland and compared to a simulation with true orography. It is found that the mountains contribute to a total increase of precipitation in Iceland of the order of 40 %.

Zusammenfassung

Die atmosphärische Strömung über Island wurde für den Zeitraum von September 1987 bis Ende Juni 2003 mit Hilfe des mesoscaligen PSU/NCAR MM5-Modells und unter Benutzung von Anfangs- und Randwerten aus dem European Centre for Medium-range Weather Forecasts (ECMWF) simuliert. Der simulierte Niederschlag wird mit zwei Arten indirekter Niederschlagsbeobachtungen verglichen. Zum einen mit der Schneeansammlung auf je zwei großen Gletschern in SO-Island und in Zentralisland. Zum anderen werden die Modellergebnisse des MM5 als Ausgangsdaten für das hydrologische Modell WaSiM-ETH verwendet, um die anfallende Wassermenge zu berechnen. Diese wird dann mit der angefallenen Wassermenge von sechs Einzugsgebieten in Island für die Wasserjahre 1987–2002 verglichen. Der im Modell ermittelte Niederschlag ist mit beiden Arten der Vergleichsdaten im Einklang. Die jahreszeitliche und interannuelle Variabilität von Niederschlag wird für niedere und hohe Höhenlagen untersucht. Die Simulationen zeigen einen negativen Trend im Winterniederschlag in Westisland, jedoch einen positiven Trend im Verhältnis von Flachlandniederschlag zu Bergniederschlag in Ostisland. Es gibt im allgemeinen eine grundlegende interannuelle Variabilität im Verhältnis von Flachlandniederschlag zu Niederschlag in den Bergen, besonders in Ostisland, was die eingeschränkte Übertragbarkeit von Niederschlagsbeobachtungen in den Niederungen auf den Niederschlag in den Bergen hervorhebt. Um die Auswirkungen der Orographie auf das Niederschlagsklima von Island zu beurteilen, wird der Niederschlag für das flache Island simuliert und mit einer Simulation für die wahre Orographie verglichen. Es stellt sich hierbei heraus, dass die Berge um 40 % zu einer Gesamtzunahme des Niederschlags in Island beitragen.

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DOI: 10.1127/0941-2948/2007/0174
0941-2948/2007/0174 $ 6.75

1 Introduction

The idea of using limited area models (LAMs) for regional climate simulations was introduced by DICKINSON et al. (1989) and refined by GIORGI (1990). One of the benefits of such an approach is that it is relatively inexpensive in terms of necessary computer resources used for simulations of the atmospheric flow at relatively high spatial and temporal resolutions. As resolution is increased, processes governed by the interaction of the large scale flow and topography become better resolved by the models. One drawback of this approach which is not present in global climate models is that the simulations are dependent on the lateral boundary conditions. These can constrain the model dynamics and hence affect the results (e.g. WARNER et al., 1997). To minimize the constraining effects of the boundary conditions, QIAN et al. (2003) suggested consecutive short term integration, overlapping in time as to minimize the effects of spin-up, instead of a single long term integration. Other investigators (e.g. GIORGI and MEARNS, 1999) opt for longer integration times, emphasizing the importance of the model to be free to develop its own internal circulations. LIANG et al. (2004) used this approach when simulating precipitation over the U.S. during 1982-2002 using the MM5-based regional climate model CMM5. Several case studies investigating orographic forcing of precipitation have been made in recent years. CHIAO et al. (2004) used the MM5 model at a 5 km horizontal resolution to simulate a heavy precipitation event during MAP IOP-2B. The precipitation was satisfactorily reproduced by the model although the total amount of precipitation was slightly higher than measured by rain-gauges. BUZZI et al. (1998) simulated a 1994 flooding event in northwestern Italy. The role of orography was found to be crucial in determining the precipitation distribution and amount. Orographic precipitation has also been investigated by use of linear theory models (e.g. BARSTAD and SMITH, 2005; SMITH et al., 2005). By using a relatively simple model they identified the cloud delay time (i.e. the rate of conversion of cloud water to hydro-meteors and the rate of evaporation) as a primary unknown parameter.

The climate of Iceland is largely governed by the interaction of orography and extra-tropical cyclones, both of which can be described quite accurately by present day atmospheric models. As a result, dynamical downscaling of the climate, using limited area models, gives valuable information about precipitation distribution, especially in the data-sparse highlands.

The impact of orography on precipitation and precipitation in the mountains have an economic aspect, since hydraulic power is generated only by water that has fallen as precipitation in the mountains, and not in the lowland. However, most precipitation observations, including long time series, are from the lowland.

RÖGNVALDSSON et al. (2004) simulated precipitation in Iceland over a 10-year period using the PSU/NCAR MM5 model (GRELL et al., 1994). Simulations were compared to conventional precipitation measurements (i.e. rain-gauge data) and to precipitation estimated by a statistical model based on observed rain-gauge data and a number of topographic and geographic predictors. It was found that the simulated precipitation was in general greater than observed precipitation. However, the magnitude and the seasonal and geographic distribution of the overestimation indicated that it was to some extent associated with observation errors due to wind loss of solid precipitation and with limitations in the representativeness of the observations as well. BROMWICH et al. (2005) simulated the same 10-year period (1991–2000) using the Polar MM5 model (BROMWICH et al., 2001; CASSANO et al., 2001) and with the same horizontal resolution as in RÖGNVALDSSON et al. (2004). They concluded that simulations of the time-averaged near-surface temperature, moisture, wind and precipitation were in relatively good agreement with observations. Trends in simulated precipitation were linked to changes in the NAO index for the region.

BENOIT et al. (2000) reported some of the advantages of using one-way coupling of atmospheric and hydrological models, calibrated with observed discharge data, for validation of precipitation calculated by the atmospheric models. They conclude that stream flow record gives a better estimate of the precipitation that has fallen over a region than point measurements, and even though there were uncertainties related to their hydrological model (WATFLOOD), it was sufficiently sensitive to help improve atmospheric models. HAY et al. (2002) used output from the RegCM2 model (GIORGI et al., 1996) as input to a distributed hydrological model for four basins in the USA. Their research indicated that precipitation averaged over a large area could have the daily variations necessary for basin scale modeling. Studies focussing on one-way coupling between atmospheric models and the WaSiM-ETH watershed model in alpine landscapes have earlier been reported by JASPER et al. (2002), JASPER and KAUFMANN (2003) and by KUNSTMANN and STADLER (2005). The WaSiM-ETH model has further been integrated with a glacier sub model (KLOK et al., 2001) to simulate the discharge of a heavily glaciated drainage basin. JASPER et al. (2002) compared WaSiM-ETH simulations that were driven by observed meteorological data, with simulations driven by data from high-resolution numerical weather prediction (NWP) models. JASPER and KAUFMANN (2003) compared results from WaSiM-ETH watershed models that were on one hand driven by meteorological observations and on the other hand driven by data from atmospheric models. They concluded that the hydrological model was sufficiently sensitive to provide substantial information for the validation of atmospheric models. KUNSTMANN and STADLER (2005) were able to reproduce observed stream flow reasonably well in an alpine and orographically complex basin in Germany by driving the WaSiM-ETH watershed model with MM5 output data.

In a recent study by JÓNSDÓTTIR and ÞÓRARINSSON (2004) the HBV watershed model (SÆLTHUN, 1996) was calibrated and driven both with observed and simulated data from the MM5 model. The main results were that the correlation between daily values of measured discharge and discharge calculated by the MM5 data was fairly good. The correlation was somewhat higher when data from nearby weather stations were used. Using the MM5 data, however, improved the water balance for each water year. TÓMASSON et al. (2005) simulated a short winter flood in the Þjórsá-Tungnaá river basin in S-Iceland, using precipitation as simulated by the MM5 model and the HEC-HMS (HYDROLOGICAL MODELING SYSTEM, 2000) runoff model. They concluded that the runoff model showed results that were in good agreement with observed discharge in the river basin. The MM5 model output has also been used as input to the University of Washington Distributed-Hydrology-Soil-Vegetation Model (DHSVM) to form an automated riverflow forecasting system (WESTRICK et al., 2002).

A atmospheric flow over Iceland has been simulated for the period September 1987 through June 2003 using version 3-5-3 of the MM5 model and initial and boundary data from the European Centre for Medium-range Weather Forecasts (ECMWF). The results are compared with two types of indirect precipitation observations. Firstly, snow accumulation on two large glaciers in SE-Iceland and on two large ice caps in central Iceland. Secondly, model output was used as input to the WaSiM-ETH hydrological model (JASPER et al., 2002; JASPER and KAUFMANN, 2003) to calculate the runoff from six Icelandic watersheds for the water years 1987–2001. The hydrological model is calibrated against measured discharge from six watersheds in different parts of Iceland where neither glaciers nor groundwater play an important role in the hydrological cycle. Hence, the hydrological model output gives a fully independent evaluation of the simulated precipitation in addition to the glaciological data.

The seasonal and inter-annual variability of precipitation is investigated at low as well as high altitudes. In order to assess the impact of orography on the precipitation climate of Iceland, precipitation is simulated with flat Iceland and compared to a simulation with true orography.

The remainder of this paper is organized as follows: In section 2 we discuss the hydrological and atmospheric model configurations. Section 3 gives a description of the validation data. Results are presented in section 4 and discussed in section 5 followed by summary and conclusions.

2 Model configurations

2.1 Atmospheric model

The PSU/NCAR MM5 model (GRELL et al., 1994) is a state of the art non-hydrostatic limited area model. It has been used to simulate the atmospheric flow over Iceland over a more than 15-year period from September 1987 through June 2003. The domain used is 123 x 95 points, centered at 64°N and 19.5°W, with a horizontal resolution of 8 km. There are 23 vertical levels with the model top at 100 hPa. A more detailed description of the model configuration can be found in RÖGNVALDSSON et al. (2004).

2.2 Modeling approach

The MM5 model was used with initial and lateral boundaries from the ERA40 re-analysis project as to 1999. After that date, operational analysis, from the ECMWF were used. The ERA40 data were interpolated from a horizontal grid of 1.125° to 0.5° prior to being applied to the MM5 modeling system. The modeling approach differs from that used by BROMWICH et al. (2005). Instead of applying many short term (i.e. of the order of days) simulations and frequently updating the initial conditions, the model was run over a period of approximately six months with only lateral boundary conditions updated every six hours. This was made possible by taking advantage of the OSU land surface model (CHEN and DUDHIA, 2001).

The period from September 2001 through August 2002 was further simulated with the orography of Iceland being reduced down to one meter.

2.3 Hydrological model

The WaSiM-ETH hydrological model is a fully distributed catchment model using physically based algorithms and parameters for the description of hydrological processes (JASPER et al., 2002; JASPER and KAUFMANN, 2003). The model offers various methods of calculating the different water balance elements depending on the availability of input data. The input data from the MM5 model used in the hydrological model were precipitation, temperature at 2 metres above ground and wind speed at 10 metres above ground. The Penman-Monteith estimate of actual evaporation requires definition of vegetation parameters that were not available, and also data on humidity and radiation that could not be used directly from the MM5 model. An attempt to use Penman-Monteith with the limited data available therefore proved unsuccessful. The Hamon approach (FEDERER and LASH, 1983) was therefore used to calculate evaporation. A temperature-wind index method

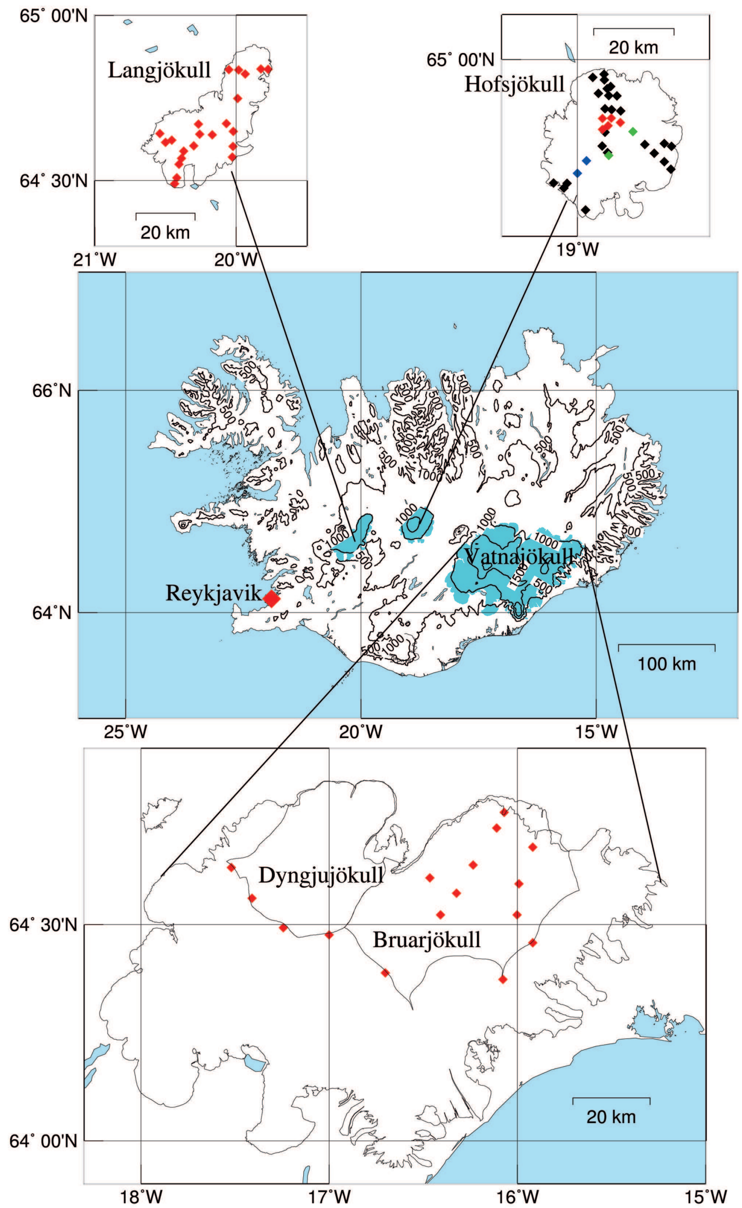

Figure 1: Overview of the four ice caps used for validation purposes, dots indicate typical location of observation sites. Red dots on Hofsjökull glacier are along profile HN (Npart), blue dots along profile HSV (SWpart) and green dots along profile HSA (SEpart), observations at locations shown in black have not been used in this study.

was used to account for higher melting when wind speed is high. The soil model used Richards equation (RICHARDS, 1931; PHILIP, 1969) for the unsaturated zone, but no groundwater model was applied.

In this study, ten parameters describing both the unsaturated zone and snow accumulation and melt were fitted to each watershed. For the unsaturated zone, the following six parameters were fitted: (1) storage coefficient of direct runoff, k_d, (2) storage coefficient of inter flow, k_i, (3) drainage density, d, (4) recession constant for base flow, k_b, and (5) saturated hydrological conductivities of the uppermost aquifer and (6) the fraction of surface runoff on snow melt. The four snow model parameters that were fitted were (7) temperature limit between rain and snow, $T_{R/S}$ (8) temperature at which snow melt starts, T_0, (9) degree-day factor without wind consideration, c_1, and (10) degree-day factor with wind consideration, c_2.

A one-way coupling between the MM5 and WaSiM-ETH model was applied by using the output from the MM5 model as input to the WaSiM-ETH model. The MM5 output was on an 8×8 km horizontal grid, while the grid of the watershed model was set to 1×1 km resolution to catch more of the characteristics of the landscape. Each grid point in the MM5 model was treated as a meteorological station, and the input to each grid

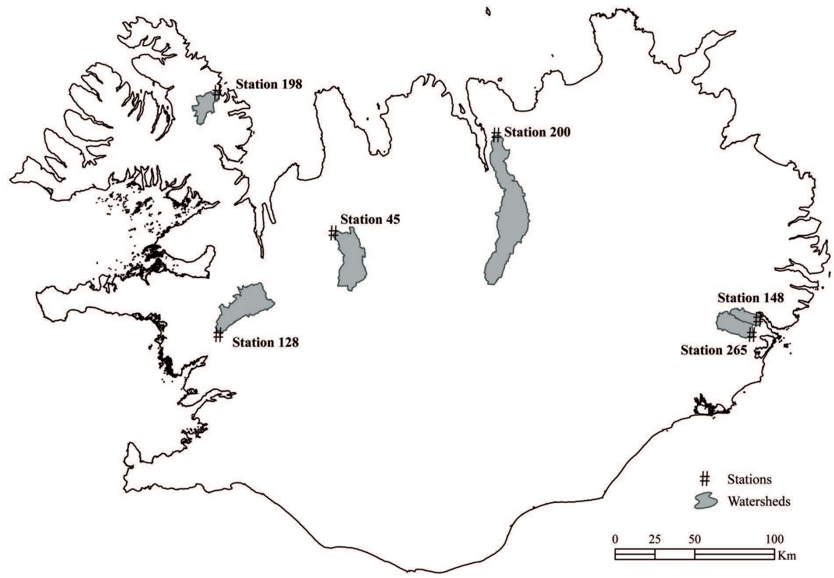

Figure 2: The location of the six watersheds and corresponding gauging stations used for validation of the MM5 precipitation data.

cell in WaSiM-ETH was evaluated by inverse distance weighting between the grid points of the MM5 model. The MM5 model output values are available for every six hours, while the watershed model was run at a daily time step because of the time resolution of observed data. The MM5 model output was therefore regridded to a daily time step, with precipitation from each of the four within-day time steps being accumulated, and with daily averages calculated for temperature and wind speed.

3 Validation data

3.1 Glaciological data

The spatial variability of the mass balance on large ice masses, such as Vatnajökull and Langjökull ice caps, can be mapped given data along several profiles extending over the elevation range of the ice caps. Since 1991 annual mass balance has been observed on parts of Vatnajökull ice cap in SE-Iceland (Björnsson et al., 1998) and from 1996 on Langjökull ice cap, central Iceland (Björnsson et al., 2002). Here, we only use measurements of accumulated wintertime snow, expressed in terms of liquid water equivalents. Björnsson et al. (1998) estimated the uncertainty of the areal integrals of the mass balance to be a minimum of 15 %. Due to surging of the Dyngjujökull glacier in 1998–2000 the uncertainty is considerably greater for this period and the foolowing winter (Pálsson et al., 2002a). The ice caps and typical locations of the mass balance stakes are depicted in Figure 1.

Precipitation on Hofsjökull ice cap has been observed at sites along profile HN (cf. Figure 1) since 1987 and along profiles HSV and HSA since 1988 (Sigurðsson et al., 2004). In our model configuration the maximum elevation of the Hofsjökull ice cap is approximately 1540 metres, i.e. more than 250 metres lower than in reality. Hence, we use area-integrated data from an elevation range of approximately 1450–1650 metres along the three profiles HN, HSV and HSA (Jóhannesson et al., 2006).

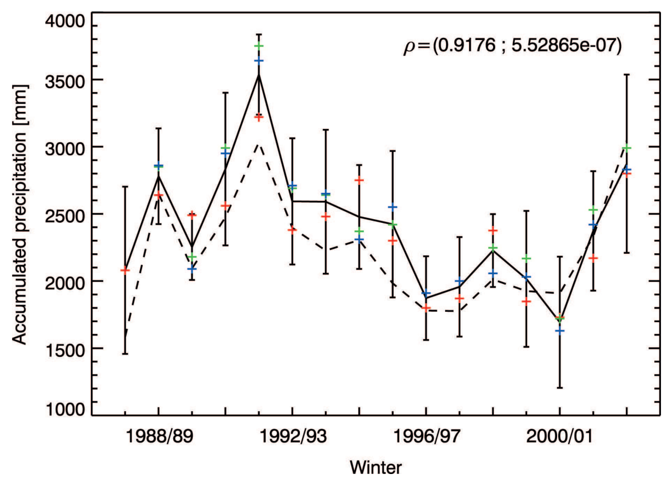

Figure 3: Estimated mean accumulated winter precipitation [mm] along profiles HN (N-part), HSA (SE-part) and HSV (SW-part) at altitudes between 1450 and 1650 metres (solid line, Jóhannesson et al., 2006). Dashed line represents simulated precipitation by MM5 from a single grid cell over Hofsjökull ice cap at altitude 1540 metres. Red, green and blue crosses represent mean values along profiles HN, HSA and HSV respectively on the altitude interval 1440–1680 metres (cf. Figure 1). Error bars indicate the standard deviation of the observations. Observed values from individual snow stakes are from Sigurðsson (1989, 1990, 1993), Sigurðsson and Sigurðsson (1998) and Sigurðsson et al. (2004).

Table 1: Comparison of observed and calculated discharge at six discharge stations and Nash-Sutcliffe coefficients of model fit.

Station	$Q_{meas}[m^3/s]$	$Q_{calc}[m^3/s]$	Difference [%]	R2	R2log
198	26.8	25.4	-5.2	0.62	0.60
265	19.6	20.8	6.1	0.70	0.74
45	12.3	13.4	8.9	0.69	0.62
128	29.4	29.4	9.7	0.61	0.64
148	9.1	10.4	10.4	0.64	0.71
200	48.4	11.4	11.4	0.53	0.53

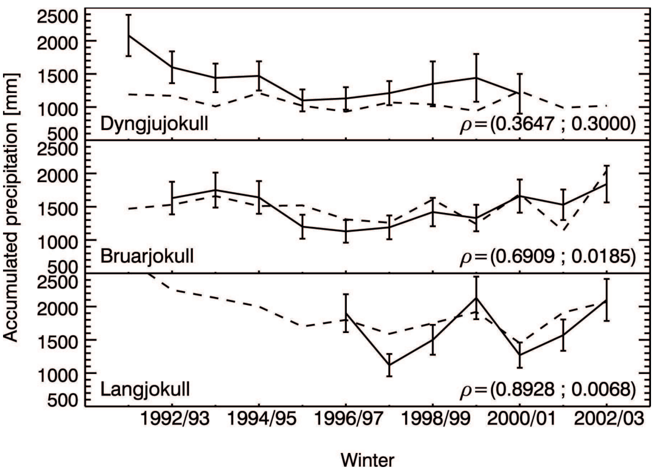

Figure 4: Estimated from observations (solid) and simulated by MM5 (dashed) accumulated winter precipitation for Dyngjujökull (top) and Brúarjökull (middle) glaciers and Langjökull (bottom) ice cap. Error bars indicate 15 % uncertainty of the observations, except for 1998–2001 at Dyngjujökull where it is 25 %. Glaciological data for Dyngjujökull and Brúarjökull are from BJÖRNSSON et al. (1998, 2002) and PÁLSSON et al. (2002a, 2002b, 2004b, 2004c). Data for Langjökull ice cap are from BJÖRNSSON et al. (2002) and PÁLSSON et al. (2004a).

3.2 Hydrological data

Large areas of Iceland are covered with post-glacial lava. In those areas, precipitation infiltrates through the porous surface, to the groundwater aquifers and in some cases through the groundwater aquifers to the ocean. Furthermore, the temperature at high altitudes in Iceland remains below zero for some months during the winter, so that some of the autumn and winter precipitation is stored until spring and glaciers may store precipitation from one season, year or decade to the next. The complexity of the hydrological cycle therefore varies from one area to the other. In this study, six watersheds were selected where the rivers are primarily direct-runoff rivers and are therefore relatively free from the complications of groundwater components and glacier mass balance changes. The locations of the six selected watersheds are shown in Figure 2. However all the watersheds have substantial snow cover during the winter, so that the models were run on the basis of a water year, i.e. from September 1, 1987 to August 31, 2002. Average daily discharges from the database of the Hydrological Service of the National Energy Authority were used to calibrate the model. A 500-m digital elevation model (ICELANDIC METEOROLOGICAL OFFICE et al., 2004), a soil map from the Agricultural University of Iceland and a digital vegetation map from the Icelandic Institute of Natural History were used in WaSiM-ETH to describe the watersheds. The geographical data were all regridded to a 1 x 1 km spatial resolution.

4 Results

4.1 Comparison with glaciological data

The simulated wintertime precipitation at Hofsjökull ice cap is in good agreement with observations (cf. Figure 3) over the northern part of Hofsjökull (HN, red dots, cf. Figure 1), the SE-part (HSA, green dots, cf. Figure 1) and the SW-part of the ice cap (HSV, blue dots, cf. Figure 1). The solid line in Figure 3 shows the estimated wintertime precipitation, taking into account ablation due to liquid precipitation and/or melting, at altitude between 1450 and 1650 metres at locations HN, HSA and HSV. The dashed line shows the wintertime precipitation simulated by MM5 at a single grid cell over Hofsjökull ice cap at altitude 1540 metres. The simulated precipitation is within one standard deviation of snow accumulation for the whole observation period (1987–2003), observed at snow stakes between 1440 and 1680 metres altitude. The Spearman's rank correlation[1] is 0.92 with a significance value of $5.5 \cdot 10^{-7}$.

When compared with estimated areal integrals of wintertime precipitation over the Dyngjujökull (1040 km^2) and Brúarjökull (1695 km^2) glaciers and the Langjökull ice cap (925 km^2), the rank correlation decreases somewhat (see Figure 4). The model shows the least skill on Dyngjujökull (ρ =0.365;0.300) and the greatest skill on Langjökull (ρ =0.893;0.007). The correlation for Brúarjökull is 0.691 with a significance

[1]We used the *r_correlate* function within the IDL® software package. The function computes the Spearman's rank correlation of two sample populations X and Y. The result is a two-element vector containing the rank correlation coefficient and the two-sided significance of its deviation from zero. The significance is a value in the interval [0.0, 1.0]; a small value indicates a significant correlation.

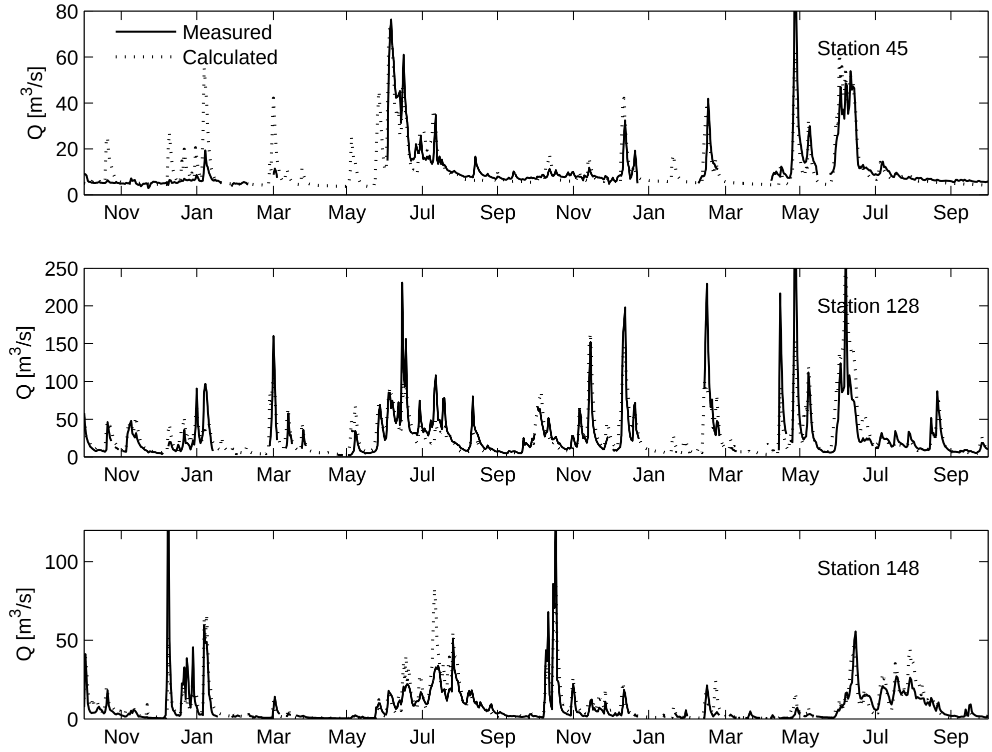

Figure 5: Comparison of measured (solid lines) and calculated (dashed lines) runoff from September 1, 1998 to August 31 2000 at stations 45, 128 and 148.

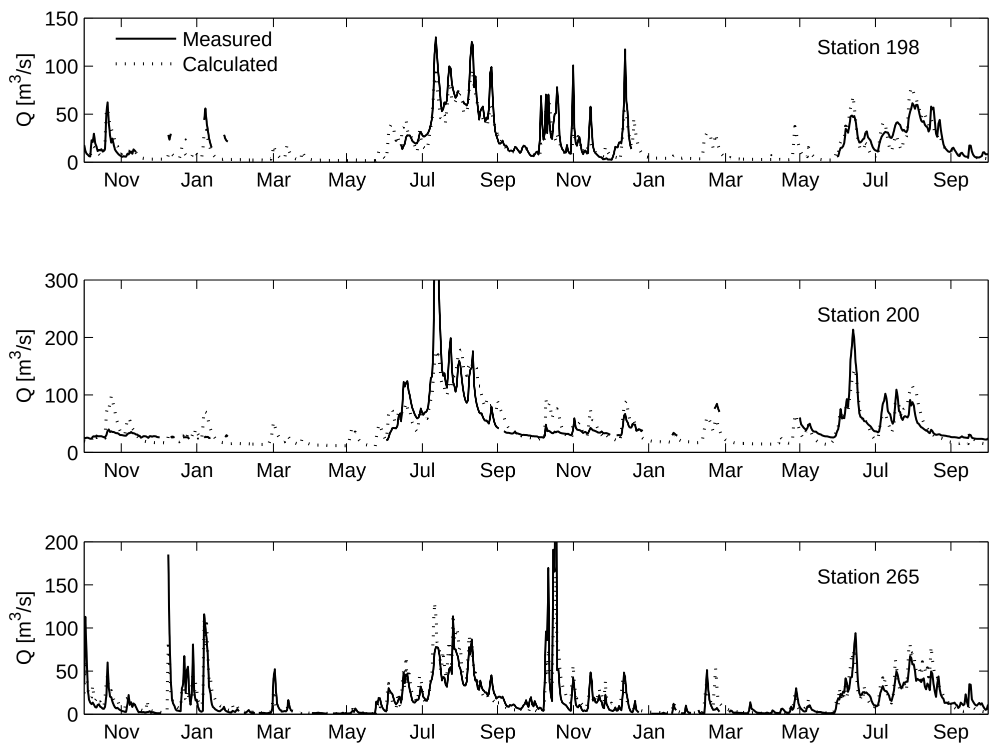

Figure 6: Comparison of measured (solid lines) and calculated (dashed lines) runoff from September 1, 1998 to August 31 2000 at stations 198, 200, 265.

value of 0.019. The simulated precipitation is within estimated observational error-margins for 5 out of 10 winters for Dyngjujökull, 9 out of 11 for Brúarjökull and 5 out of 7 for Langjökull ice cap.

4.2 Comparison with hydrological model data

Runoff from the six Icelandic watersheds used in this study is strongly influenced by snow accumulation and

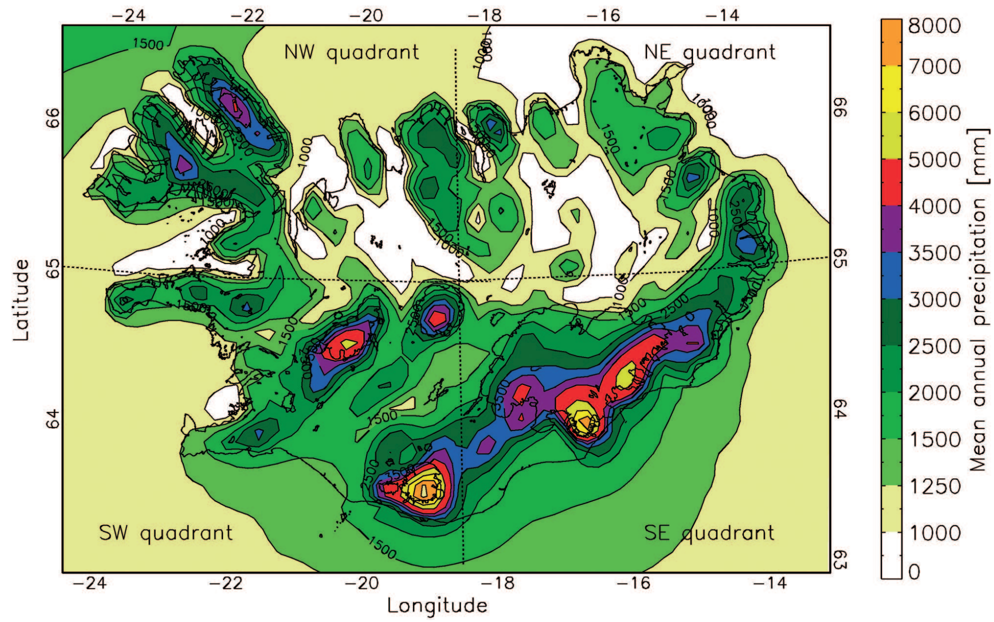

Figure 7: Mean annual precipitation from March 1988 through February 2003 as simulated by the MM5 model. Dashed lines show the definition of NW, NE, SE and SW quadrants.

snow melt. Therefore, the fit of simulated to observed runoff is highly dependent on both temperature and precipitation, while the overall water balance of water years depends primarily on precipitation data. Here, the Nash-Sutcliffe coefficient $R2$ (NASH and SUTCLIFFE, 1970) and $R2log$ is used to measure how well the simulated runoff fits the observed runoff. Both coefficients $R2$ and $R2log$ range from 1 to $-\infty$, where a perfect fit equals 1. The coefficient $R2$ emphasizes the fit for high flows and floods while $R2log$ puts greater weight on how well low flows are simulated.

Table 1 shows the $R2$ and $R2log$ coefficients as well as the fit of the water balance for the period September 1, 1988–August 31, 2002. The average simulated runoff is 7 % higher than the observed runoff, the $R2$ and $R2log$ values are higher than 0.6 except for one basin where they are both 0.53. Figures 5 and 6 also show observed and simulated hydrographs, for the period September 1, 1998–August 31, 2000, for the six watersheds.

4.3 Simulated time series

Figure 7 shows the mean annual precipitation as simulated with the MM5 model over the 15-year period from March 1988 to February 2003. It shows a realistic precipitation pattern with the greatest precipitation over the large ice caps in S- and SE-Iceland and over the three large ice caps in central and NW-Iceland.

Figure 8 shows simulated seasonal precipitation for lowland points, defined as model grid points below 100 metres (11 % of Iceland) and highland points (altitude above 100 metres) for the four quadrants shown in Figure 7 as well as the whole of Iceland. The greatest absolute difference between the lowland and highland points is during the winter months (December to February, DJF) and minimum difference is during the summer months (June to August, JJA). These two seasons show the greatest (DJF) and smallest (JJA) inter-annual variability. On average, the greatest difference between precipitation at low- and highland points is in the SW quadrant for all seasons, while the NE quadrant shows the smallest difference. There is considerable inter-annual variability for all quadrants, but least in the NE. The NE quadrant is the driest quadrant. Precipitation at lower altitudes sometimes exceeds precipitation in the mountains, most frequently so during SON in the NE quadrant and in JJA in the SE quadrant.

Figure 9 (left) shows the seasonal precipitation over a 15-year period from MAM 1988 through DJF 2002. The seasonal variability is clear in all quadrants and lowland precipitation is clearly considerable lower than precipitation in the mountains at most times. The exception is the NE and to a less extent, the SE quadrant from 1997 to 2002. The right panel of Figure 9 shows the seasonal precipitation for all quadrants for the same period. A negative trend can be seen in wintertime (DJF) precipitation in the western part of Iceland (cf. Figure 8, upper left panel).

Figure 10 shows the ratio of simulated low- and highland precipitation to total precipitation for each quadrant, as well as the sum of all quadrants. In the NE and to some extent in the SE, there is a positive trend in the relative proportion of lowland precipitation during winter and springtime but the greatest inter-annual variability in the precipitation of lowland to highland precipitation is during JJA in the SE quadrant. There appears to be an oscillation in the lowland precipitation during winter (DJF) and summer (JJA) in the southern quadrants with a period of about five years in this period. The greatest amplitude is found in the SE quadrant during JJA 1992–2002.

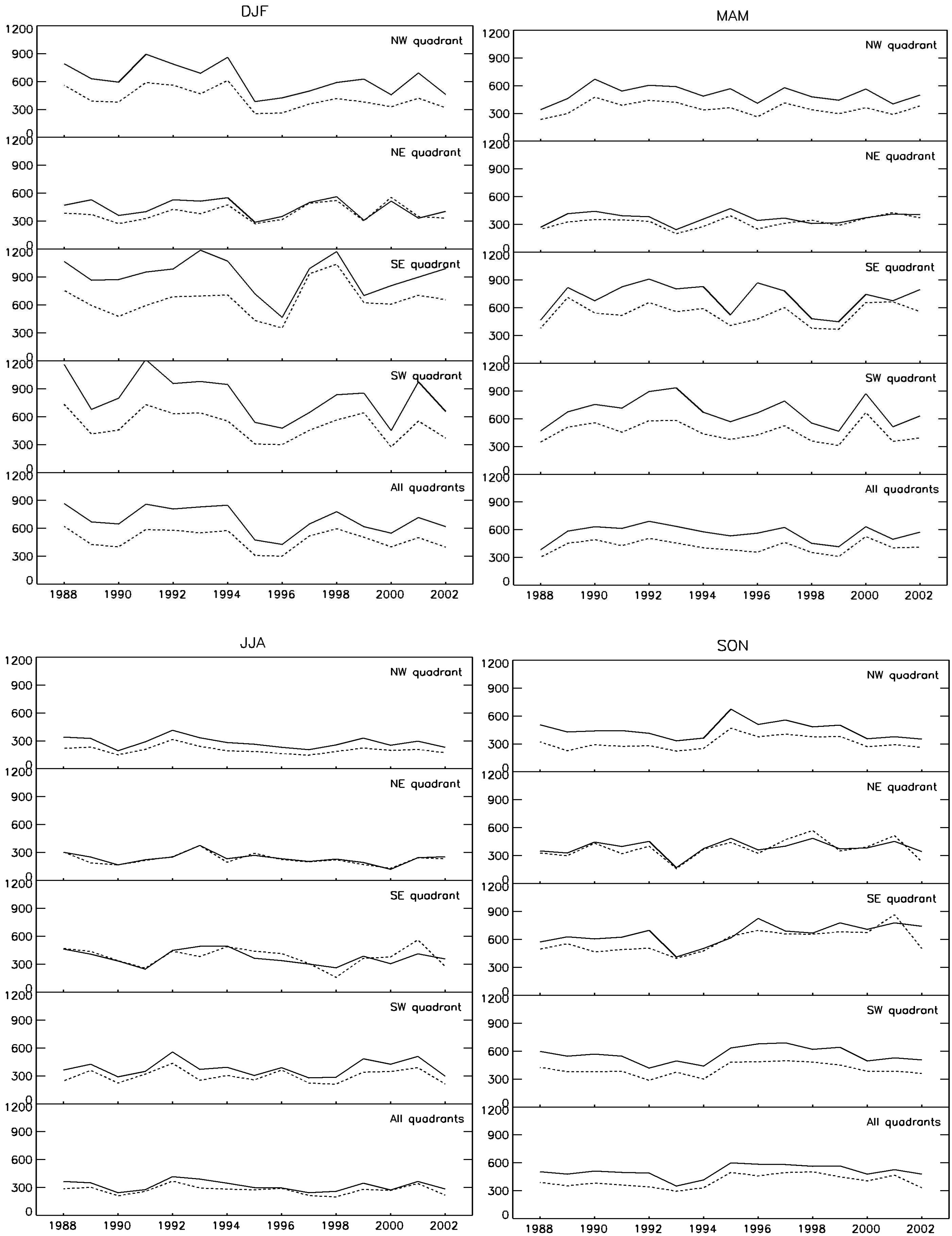

Figure 8: Simulated seasonal (DJF, MAM, JJA and SON) precipitation [mm] for the lowland (dashed lines, topography below 100m) and the highlands (solid lines, topography above 100m) from March 1988 through February 2003. The country is divided into four quadrants, NW (top panel), NE (second from top), SE (middle panel) and SW (second from bottom). The lowest panel shows the sum of all quadrants.

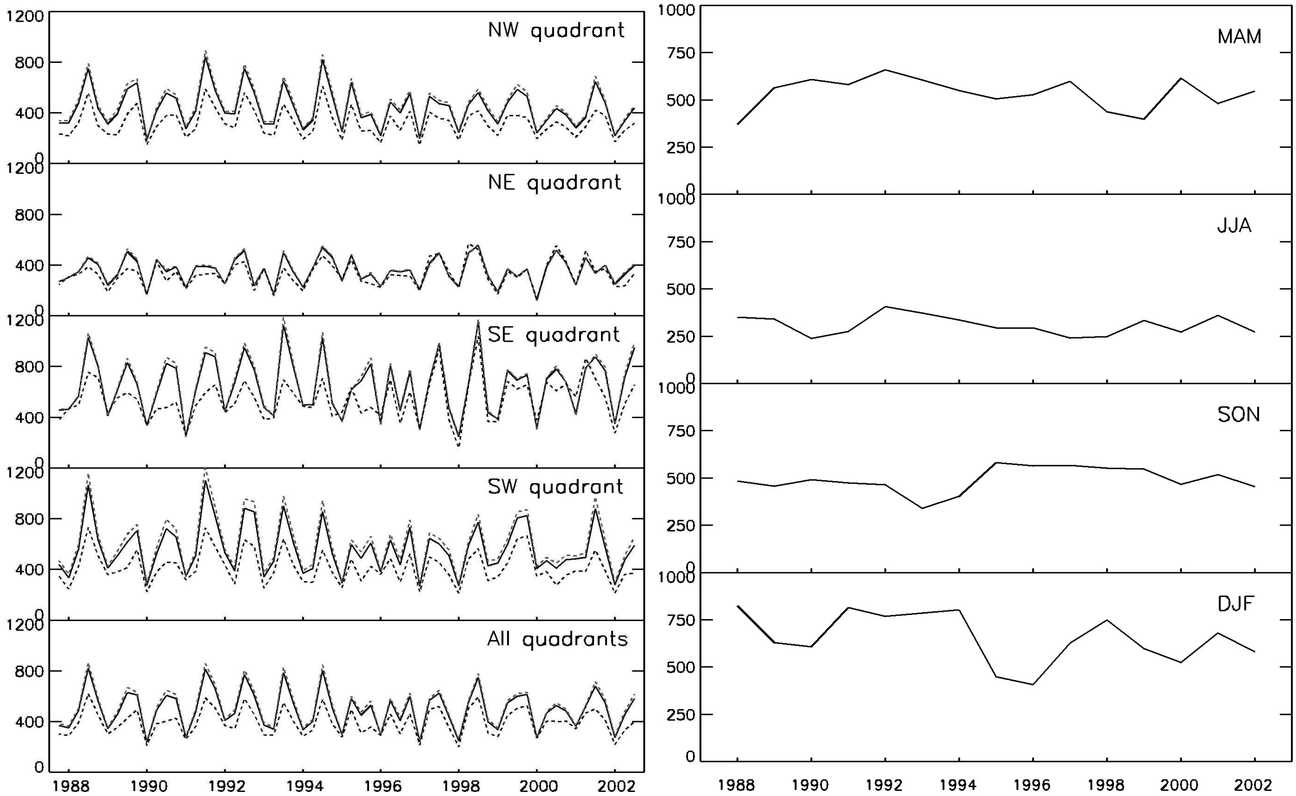

Figure 9: Simulated seasonal precipitation [mm] from MAM 1988 through DJF 2002 for the NW, NE, SE and SW quadrants and sum of all quadrants (left). Solid black line shows the total precipitation, dashed line shows the precipitation for the highlands (z>100m) and dashed black lines for the lowlands (z<100m). The right panel shows the seasonal precipitation [mm] for all quadrants for the same period.

4.4 Orographic effects

Figure 11 shows the accumulated precipitation between September 2001 and August 2002 with unmodified (CONTROL, left) and flat terrain (FLAT, right). The absolute difference between the two simulations is shown in Figure 12 (left) as well as relative difference (right). The mean monthly precipitation for both simulations is shown in Figure 13 along with the relative difference.

The mountains constitute about 40 % increase in precipitation over Iceland. The differences in monthly values range from 25 % to 55 %. The mountains cause a drying in the highlands north of the Vatnajökull ice cap and north of the two large ice caps in central Iceland. The valley areas in the central and southeast part of the NW quadrant and the two largest fjords in the northwesternmost part of Iceland are also drier when the mountains are present. The mountains cause an increase in precipitation that reaches far south of Iceland, while a decrease in precipitation is evident far to the north and northeast of Iceland.

5 Discussion

In this study, numerically simulated precipitation has been compared with unconventional observations of precipitation, i.e. runoff and snow accumulation. This type of data only provides validation on a much longer timescale than conventional rain-gauge data, and the daily error in the precipitation downscaling remains unclear. However, the comparison with the observational data shows that the climatological values of the simulated precipitation are of good quality. The correlation between observations and simulations is in fact much better than in RÖGNVALDSSON et al. (2004). The relatively poor correlation in RÖGNVALDSSON et al. (2004) is mainly because of observational errors associated with undercatchment by the rain-gauges but not because of a poor statistical model treating the rain-gauge observations or a poor quality in the numerical simulations. In this study, precipitation from the MM5 model has not been scaled in any way to fit the observed discharge. The good fit of the watershed models, particularly with regards to accumulated water balance, therefore, suggests that MM5 precipitation in these areas is close to the actual precipitation. However, no conclusions on the precision of other meteorological variables, such as temperature and wind speed, can be drawn from this study because parameters in the hydrological model involving snow melt and accumulation were adjusted to improve the fit as measured by $R2$ and $R2log$. These results do suggest that meteorological output from the MM5 models can be used with WaSiM-ETH to set up successful models of runoff in the areas of these six watersheds.

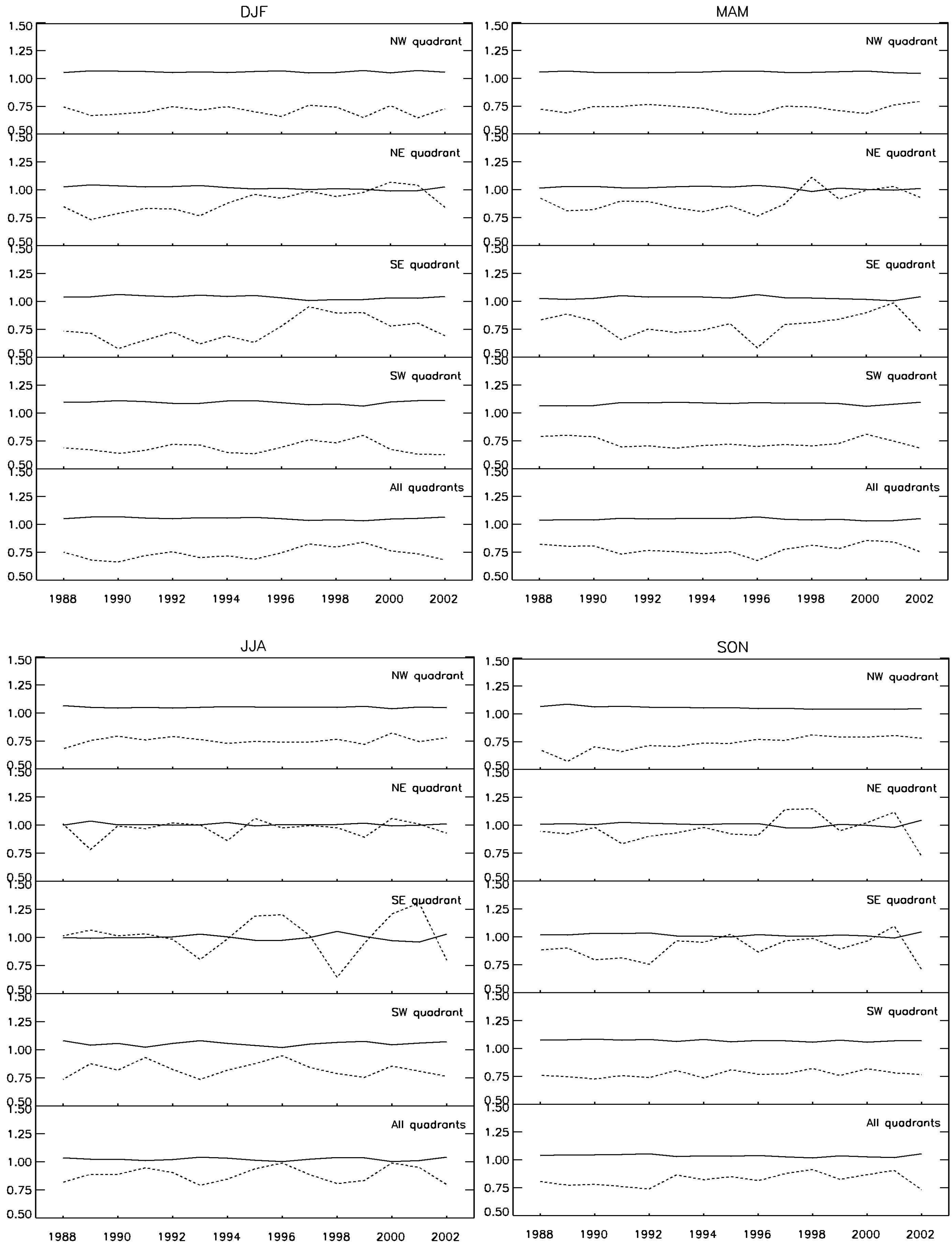

Figure 10: Ratio [%] to total simulated precipitation for lowland (dashed lines) and the highlands (solid lines) for the four seasons and individual quadrants, as well as the whole country from MAM 1988 through DJF 2002.

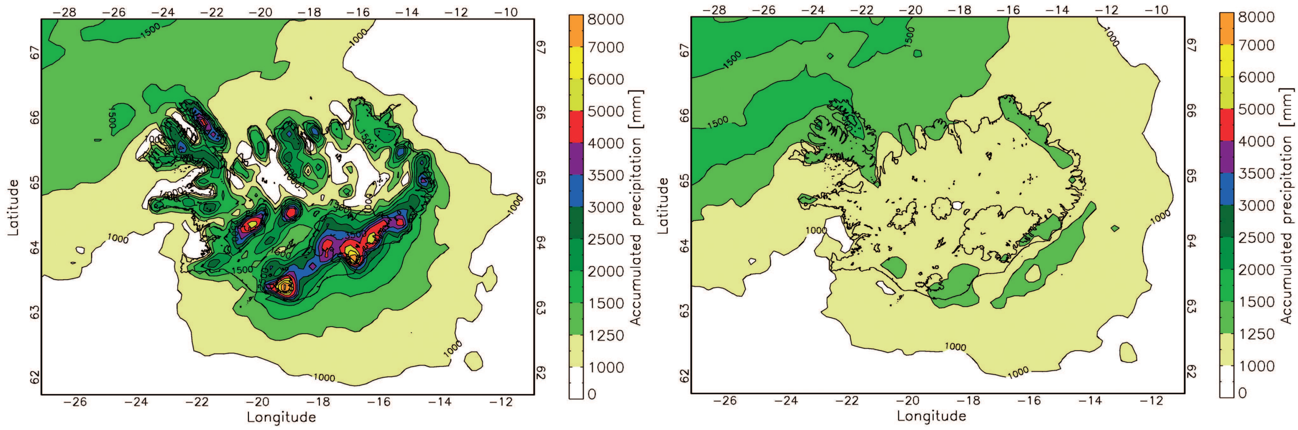

Figure 11: Simulated precipitation [mm] for 2001-02 (September through August) with unmodified terrain (CONTROL, left) and with the orography reduced to one meter (FLAT, right).

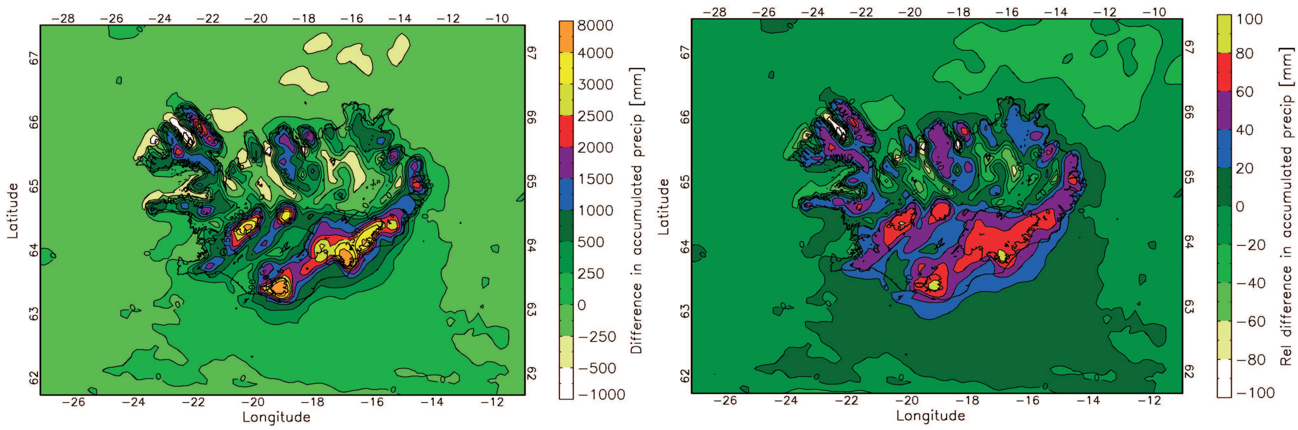

Figure 12: Absolute [mm] difference in precipitation (left) between CONTROL and FLAT and relative [%] (CONTROL-FLAT / CONTROL) difference (right).

The fit between measured and calculated discharge might be improved by adjusting the input precipitation; however this is not the purpose of the study. Also, the use of more advanced interpolation methods for the meteorological variables, with elevation dependency might improve the model. An application of more advanced evaporation schemes, according to the Penman-Monteith approach, could give a better evaluation of evapotranspiration, but as mentioned earlier the use of Penman-Monteith has been proved difficult in this study. Furthermore, the WaSiM-ETH model simulate the heat flux in the soil or subsurface, so that discharge during winter when soil is frozen might be simulated better if a different hydrological model were used. However despite these limitations, the comparison of measured and calculated discharge gives acceptable results with regard to the one-way coupling of MM5 and WaSiM-ETH.

The simulations reveal several interesting aspects of the precipitation pattern in time and space. Firstly, there is a negative trend in the precipitation, as pointed out by BROMWICH et al. (2005). However, this trend is small compared with the inter-annual variability, and by choosing different 15-year periods during the last 45 years, quite variable trends can be obtained (cf. Figure 14). The negative trend for 1988–2002 is primarily confined to the western part of Iceland (quadrants NW and SW) in winter. This happens at the same time as wintertime ratio of lowland precipitation to highland precipitation increases steadily in the eastern part of Iceland (quadrants SE and NE). Regional precipitation in Iceland is very dependent upon wind direction. Basically, most precipitation in each region falls when the winds are blowing from the ocean, while when winds are blowing from the central highlands, there is usually only little, if any precipitation (EINARSSON, 1984). On a day-to-day time scale, precipitation in the northeast is thus negatively correlated with precipitation in the southwest. On a longer time-scale, the correlation is not necessarily as simple and when the total precipitation falls to a bottom value in the west (winter of 2000), there is not a distinct peak in the precipitation in the northeast. On the other hand, the ratio of lowland precipitation to highland precipitation reaches a peak in the northeast this same winter. In general, strong winds favour precipitation in and immediately downstream of the upstream slopes (e.g. de VRIES and ÓLAFSSON, 2003), while in weak winds the flow is blocked and the orographic lifting may be very little, and may take place at some

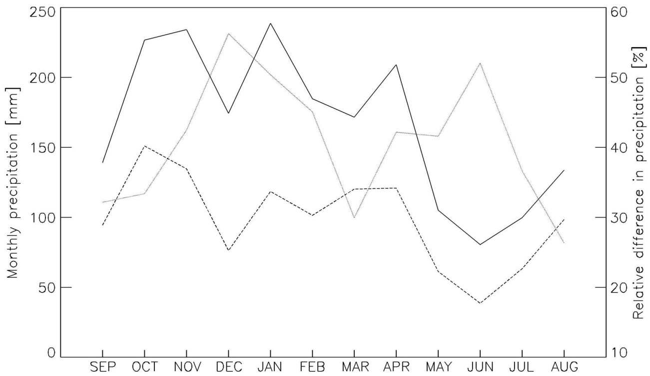

Figure 13: Accumulated monthly precipitation in Iceland [mm] as simulated in CONTROL (solid line) and FLAT (dashed line). Relative difference is shown with dotted line.

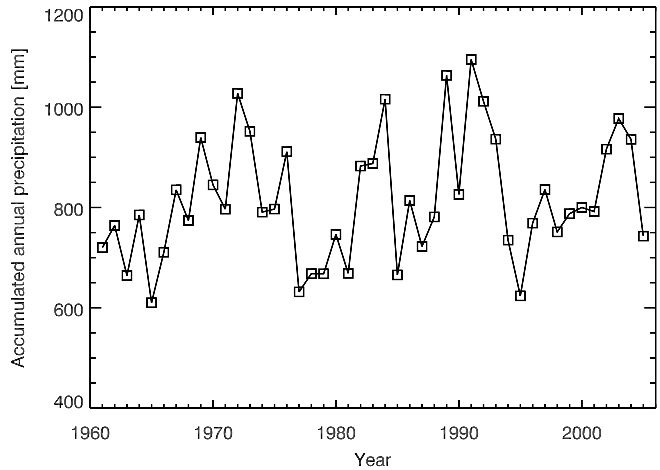

Figure 14: Observed annual precipitation [mm] at station Keflavík (WMO 4018) from 1961 to 2005.

distance upstream of the mountain. The simulated precipitation pattern indicates therefore that in the winter of 2000–2001, northeasterly, but relatively weak winds were prevailing. This was indeed the case. Investigation of observations show that in the southwest, northeasterly winds were anomalously frequent during this period, and at the northeast coast, the mean wind speed during precipitation was only 8.6 m/s, which is 1.4 m/s below the average value.

The regional variability in the proportion of precipitation falling in the lowlands can be explained by variability in the terrain. The relatively low proportion of highland precipitation in the NE is associated with the fact that there is a relatively large and dry plateau at a high elevation in the inland areas. In the NW, the lowland is sheltered and dry in northeasterly winds. Consequently, the lowland precipitation is a lower proportion of the highland precipitation than in the NE. In the SW, there is a similar sheltering of the lowlands as in the NW, but in easterly and southeasterly winds.

In general, the ratio of precipitation in the highlands to the lowlands is lowest in the summer. This is not unexpected as winds are much weaker in the summer than in the winter. This result underlines that neither summertime rain-gauge observations in the mountains nor observations of snow accumulation in the winter can be interpolated directly to the rest of the year by simple correlation with observations in the lowland, as sometimes is done.

There is substantial inter-annual variability in the proportion of precipitation in the lowlands to the highland precipitation, particularly in the summertime in the SE-part of Iceland. An investigation of weather patterns reveals that when the proportion of lowland precipitation is exceptionally low (1998), winds from the south are anomalously frequent, but winds from the east are exceptionally rare. During summers of high proportion of precipitation in the lowland (1995, 1996 and 2001), precipitation in winds from the south is less frequent than in 1998, while winds from the east are more frequent than in 1998. In winds from the east, the orographic lifting in SE-Iceland is much less than when winds are blowing from the southeast or south. In short, the large variability in the ratio of lowland precipitation to highland precipitation in SE-Iceland appears to be associated with variability in the relative frequency of wind directions.

The experiment with a flat Iceland confirms the general conception that large areas in N-Iceland are submitted to a net reduction of precipitation due to the mountains. Large parts of these areas are deserts, but that may even more a consequence of low summer temperatures, strong winds, transport of sand by wind and the volcanic nature of the soil, than due to lack of precipitation. The importance of orographic lifting for precipitation in the mountains is also confirmed by the flat Iceland experiment. This was in fact already quite clear from comparing the topography of Iceland to the simulated precipitation. In the south of Iceland, there are large areas where more than 50 % of the total precipitation is due to the impact of the mountains. In reality, this proportion may be greater, because at the current 8 km resolution, the mountains are not fully resolved (see f. inst., BROMWICH et al., 2005). The orography of Iceland contributes to some increase in precipitation as far as the domain extends to the south of Iceland, indicating that orographic lifting starts far upstream of the mountain. There is on the other hand a substantial rain shadow far offshore to the north of Iceland, indicating that it takes more than a few hundred kilometers for the precipitation systems to recover after the flow has passed a mountain range of the size of Iceland. This is in agreement with the precipitation climate of numerous regions in the world that experience rain shadow from very distant mountain ranges.

6 Summary and conclusions

A numerical weather prediction model has been shown to be very useful for mapping precipitation in complex terrain in a climate governed by extra-tropical cyclones. Snow accumulation and runoff data can be applied successfully to validate such simulations and may even be more suitable to such evaluation than traditional raingauge observations. During the period 1988–2002 there was a negative trend in wintertime precipitation in western Iceland, but a positive trend in the proportion of lowland precipitation to highland precipitation in eastern Iceland. There is substantial temporal variability in the proportion of lowland precipitation to precipitation in the mountains, and this proportion can be associated with wind speeds and prevailing wind directions. In spite of large regions in the north and in the west of Iceland being in a rain shadow, the mountains contribute to a total increase of precipitation in Iceland of the order of 40 %.

Acknowledgements

This work has been sponsored by the Icelandic Science Fund (RANNÍS), the Icelandic Energy Fund (Orkusjóður) and the National Power Company (Landsvirkjun). It is part of the Nordic and national research projects Climate and Energy (CE) and Veður og orka (VO). The authors wish to thank Tómas JÓHANNESSON and Oddur SIGURÐSSON for discussions and for providing data from the Hofsjökull ice cap. Comments from two anonymous reviewers further improved the article.

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