Available online at www.sciencedirect.com
SCIENCE DIRECT®
Journal of Hydrology 282 (2003) 95–103
Journal of Hydrology
www.elsevier.com/locate/jhydrol

Redistribution of snow and glacier mass balance from a hydrometeorological model

M. Kuhn

Institute of Meteorology and Geophysics, University of Innsbruck, Innsbruck 6020, Austria

Abstract

This paper reports on the extension of a hydrometeorological model of glacierized basins to incorporate a detailed account of glacier mass balance. In this mass balance module a constant fraction of snow fall is redistributed within the basin from ice free areas to the glacier areas. In the particular basin discussed here solid precipitation on the glacier surface is increased by a factor of 2.15 over the basin mean of each 100 m altitude interval.

The resulting mass balance parameters like mean annual specific balance, $b(h)$, balance gradients, equilibrium line altitude and accumulation area ratio are within the range of records from other glaciers.

The sensitivity of these quantities to meteorological input and model tuning parameters is evaluated, e.g. the sensitivity of mean specific mass balance to mean basin temperature is of the order of 800 mm/K. Emphasis is placed on the net accumulation outside the glacier areas in years with positive mass balance.

Keywords: Basin; Hydrometeorological; Hydrology

1. Introduction

The mass balance of a glacier can be determined or approximated in various ways that need to be adapted to the climate and hydrology of a particular glacier area. An up to date account of mass balance methods has been edited by Jansson et al. (1999), a sample of older references includes Meier et al. (1971), Hoinkes (1971) and Oestrem and Brugman (1991).

Among the methods most frequently applied, the direct glaciological method, which integrates direct measurements at strategically located points over the entire glacier, serves the purpose best in short term, small area investigations. The geodetic method, essentially the difference of two glacier maps or surface elevation models taken at different times is most useful for large areas and long term comparisons (Finsterwalder and Rentsch, 1993; Lang and Patzelt, 1971; Zwally and Brenner, 1999).

The hydrological method in its pure form tries to determine the mass balance of a glacier from the water balance of an entire glacierized basin

$$P(\text{precipitation}) = Q(\text{runoff}) + E(\text{evaporation}) + S(\text{storage}) \quad (1.1)$$

where long term storage is taken as changes in snow and ice, while groundwater longterm storage changes can be regarded as negligible. Compared to precipitation and runoff, snow and ice storage changes may be an order of magnitude smaller and of the same order as the error of these terms. The situation is worse or hopeless when the basin precipitation cannot

E-mail address: michael.kuhn@uibk.ac.at (M. Kuhn).

0022-1694/$ - see front matter © 2003 Elsevier B.V. All rights reserved.
doi:10.1016/S0022-1694(03)00256-7

be determined directly as is often the case in alpine basins.

The present study originated from the need to determine the basin precipitation of Rofenache, which drains Hintereisferner, Kesselwandferner and Vernagtferner, three of the best studied glaciers of the Alps. From these glaciers, the observed course of specific mass balance with altitude $b(h)$ was taken as representative for the whole basin. Evaporation was modelled from simple assumptions as function of altitude and snow cover duration. Runoff was recorded at a gauging station for a basin of nearly 100 km^2 and 40% ice cover.

Once annual basin precipitation was established, a hydrometeorological model was constructed that used thirty years mean monthly values of runoff, precipitation and temperature at reference valley stations and monthly vertical gradients of precipitation and temperature from the regional network of weather stations. On this basis, monthly values of $P(h, \mathrm{mo})$, $Q(h, \mathrm{mo})$, $E(h, \mathrm{mo})$ and $S(h, \mathrm{mo})$ were determined in four approximations.

The snow cover was modelled from the runoff and meteorological records, using the fraction of solid precipitation, a degree day factor, and assumptions on the fraction of ground outside glaciers being covered by snow for model tuning. The glacier mass balance enters the model only as annual basin value in the first approximation and is subsequently built up like the snow cover allowing, however, for ice ablation. As long as monthly balances for the whole basin are concerned, no further adjustment of the storage term is required. This was the state of mass balance treatment in previous applications of the model.

It is obvious, however, that the redistribution of snow in the basin by wind and avalanches is decisive for a distinction between glacier and ice free areas and in particular for the reproduction of the altitudinal mass balance profile $b(h)$. It is the purpose of this paper to show that this redistribution can be incorporated into the model and that the resulting mass balance quantities are glaciologically plausible.

2. A summary of the model

The hydrometeorological model used in this study has been developed at the Institute of Meteorology and Geophysics of the University of Innsbruck, various aspects and applications of it have been published (Kuhn et al., 1982; Kuhn and Pellet 1987; Kuhn and Batlogg, 1998,1999; Kuhn, 2000). As its primary purpose was to determine the basin precipitation as residual from the basin's water balance it uses monthly values and 100 m altitude intervals. The basic steps of the model are as follows.

In a first step the annual basin precipitation is determined from measured runoff, a first approximation of evaporation and basin storage (Eq. (1.1)). Thirty years averages are used for this exercise. Basin storage is assumed to consist of ice and snow only, it is computed from a measured reference glacier that may be outside the basin, and the ice area/altitude distribution of the basin.

The resulting basin precipitation P is distributed over the months as $P(\mathrm{mo})$ proportional to records from one or two reference stations

$$P(\mathrm{mo})/P = \mathrm{Pref}(\mathrm{mo})/\mathrm{Pref} \tag{2.1}$$

as a first approximation. Reference values are corrected for measuring errors according to Sevruk (1983). Monthly values of the increase of precipitation $P(\mathrm{mo})$ with altitude need to be determined from the records of representative stations in the region and subjectively extrapolated to higher altitude. They are first expressed as a matrix of relative values $P'(h, \mathrm{mo})/\mathrm{Pref}$ and then converted to a matrix of $P(h, \mathrm{mo})$ such that the annual basin precipitation is conserved. Since the altitudinal gradients $\mathrm{d}P(\mathrm{mo})/dh$ have a seasonal course, $P(h, \mathrm{mo})$ has to be recalculated in a second approximation.

Monthly mean air temperatures are taken from two reference stations and adapted to the proper altitude by use of altitudinal temperature gradients $\mathrm{d}T(h, \mathrm{mo})/\mathrm{d}h$, again determined from the surrounding station network and extrapolated to the upper levels of the basin. The temperatures $T(h, \mathrm{mo})$ need to be known for two applications: the decision whether precipitation is rain or snow, and the production of a factor that determines ice melt from monthly mean air temperatures.

In both applications the use of monthly means is unsatisfactory: 1. Actual temperatures at the time of precipitation may be higher in winter and lower in summer than monthly means. This needs to be compensated when calibrating an equation for the solid fraction K of total monthly precipitation

$$K(h, \mathrm{mo}) = K1 - K2^{*}T \qquad (2.2)$$

2. A positive monthly degree day sum does not reflect the negative daily means of air temperature that may have occurred in that month (e.g. Braithwaite and Zhang 2000), nor does an instantaneous positive temperature necessarily mean melting conditions (Kuhn, 1987). Nevertheless, monthly means of positive degree days appear to be an economic compromise in this model.

Degree days are converted into potential melting by way of the degree day factor DDF (mm per positive degree day), an important tuning factor in the calibration of the model. Potential melting on ice free areas is limited to the amount of snow presently on the ground and proceeds at the potential rate on glaciers. Even if snow cover is calculated for an altitude interval in ice free areas, it will not cover it completely, i.e. the potential melt rate is not applicable to the entire area but to the snow covered fraction FS which has to be introduced by trial and error in the final tuning.

Evaporation was determined only according to the state of the ground (snow, bare ground, grass and forest), time of the year and altitude, ranging from 0.5 mm per day from snow to 2 mm per day from vegetation in summer. Values of $E(h, \mathrm{mo})$ were immediately subtracted from $P(h, \mathrm{mo})$ as in the long term averages each month had sufficient precipitation.

The build up of the snow cover $\mathrm{SC}(h, \mathrm{mo})$ is not identical with the storage term $S(h, \mathrm{mo})$ since the latter must include liquid water in the ground and in the snow cover. The calculation of the snow cover includes snow of the previous month, solid precipitation Ps, potential melt PM, and evaporation E

$$\mathrm{SC}(h, \mathrm{mo}) = \mathrm{SC}(h, \mathrm{mo} - 1) + \mathrm{Ps}(h, \mathrm{mo}) - \mathrm{PM}(h >, \mathrm{mo}) - E(h, \mathrm{mo}) \qquad (2.3)$$

Once snow cover is determined in Eq. (2.3), evaporation which depends on snow cover among other things, has to be recalculated in a third approximation of the balance. At that stage total storage change $S(h, \mathrm{mo})$ results from the water balance,

$$S(h, \mathrm{mo}) = P(h, \mathrm{mo}) - Q(h, \mathrm{mo}) - E(h, \mathrm{mo}) \qquad (2.4)$$

Runoff is accounted for separately for rain and melt water. In order to avoid routing procedures in this model and in view of the long time step of one month, in the first three approximations runoff is assumed to leave the basin in the same month as rain is falling or melt water is produced. It is, however, obvious that some liquid water must be stored in the basin, partly because runoff is observed in the winter and partly because the spring snowpack is infiltrated by melt water to reach field capacity before runoff can be generated. In addition, the hydraulic characteristics of the infra- and subglacial drainage system are subject to a strong seasonal variation (e.g. Roethlisberger and Lang, 1987)

This discrepancy is tackled in the fourth approximation where calculated runoff is compared to measured values. First the annual sums are brought to agreement by tuning the values of the degree day factor DDF, the fraction of solid precipitation $K(h, \mathrm{mo})$ by means of $K1$ in Eq. (2.2), and the fraction covered by snow FS. Since these display different seasonal courses, tuning then generally needs to be done with all three of them in order to reproduce $Q(\mathrm{mo})$ properly. The approximation is accepted when calculated and measured values of $Q(h, \mathrm{mo})$ agree to better than 20 mm.

3. Refinement of the glacier module: redistribution of snow in the basin

When the model as it was described above is inspected for mass balance distribution, there are large amounts of accumulation outside the glacier area, while the glaciers themselves may display unrealistic negative balances. It is true that in years with positive glacier mass balance there is also net accumulation (positive balance) in normally ice free areas, but on the long term average, or under stationary conditions, glaciers are where they are because of topographic effects. To start with, precipitation is unevenly distributed within the basin while it is falling, but once the snow is on the ground it is redistributed at the 100–1000 m scale by wind and avalanches, both of which tend to erode snow from ridges and steep slopes and deposit it in valleys and cirques.

Table 1
Model output for Paznaun basin

	Month	10	11	12	1	2	3	4	5	6	7	8	9	Year
First approximation (observed, mm)	Precipitation	74	104	120	126	108	123	116	108	151	182	179	110	1502
	Runoff	65	31	23	17	12	14	26	141	258	287	226	129	1228
	Evaporation	16	15	16	16	14	16	15	18	31	38	38	33	265
	Storage	−7	58	82	93	81	94	76	−50	−138	−143	−85	−52	9
Precipitation in parts per thousand		49	70	80	84	72	82	77	72	101	121	119	73	1000
Relative increase of P per 100 m altitude		0.028	0.038	0.038	0.043	0.051	0.057	0.053	0.027	0.009	0.009	0.010	0.015	
Reference temperature (°C)		3.7	−1.7	−5.3	−6.3	−5.7	−3.0	0.5	5.1	8.4	10.6	10.3	7.5	2.0
Temperature gradient K per 100 m altitude (°C)		−0.4	−0.4	−0.3	−0.3	−0.4	−0.5	−0.6	−0.6	−0.6	−0.5	−0.5	−0.5	−0.5
Degree day factor	5 mm/positive degree day													
Ratio of solid to total P	$0.6-0.055*T$													
Snow covered fraction of ice free area		1.00	1.00	1.00	1.00	1.00	1.00	1.00	0.80	0.60	0.60	0.60	0.60	
Redistribution factor	2.15													
Final approximation (modelled, mm)	Precipitation	72	110	126	138	125	150	137	104	128	154	153	97	1495
	Runoff	76	40	22	16	11	16	32	142	239	275	208	133	1211
	Evaporation	16	15	16	16	14	16	15	16	18	32	44	43	259
	Storage	−19	55	89	106	100	118	90	−53	−129	−153	−99	−78	26
Modelled-observed runoff (mm)		11	8	−1	−1	−1	3	7	1	−18	−1	−16	5	−2
Melt water runoff (mm)														829
Rain water runoff (mm)														382
Total storage (mm)		−19	55	89	106	100	118	90	−53	−129	−153	−99	−78	26
Liquid water storage (mm)		−30	−25	−20	−15	−10	−10	−10	30	110	50	−25	−45	0
Snow and ice storage (mm)		11	80	109	121	110	128	100	−83	−239	−203	−74	−33	26

Table 2
Model output of specific mass balance and its change with altitude, for the conditions indicated in Table 1

	Accumulation in 1000 m^3		Total basin	Specific balance		Glacier area (km^2)
	On ice free area	Glacier		$b(h)$ (mm)	db/dh (mm/100 m)	
1700–1800	0	0				
1800–1900	0	0				
1900–2000	0	0				
2000–2100	0	0				
2100–2200	0	0				
2200–2300	0	0				
2300–2400	0	0				
2400–2500	0	−467		−1845		0.3
2500–2600	0	−1014		−1242	603	0.8
2600–2700	0	−1350		−628	614	2.1
2700–2800	0	−327		−111	517	2.9
2800–2900	0	1128		359	470	3.1
2900–3000	0	2097		837	478	2.5
3000–3100	0	880		1322	485	0.7
3100–3200	391	166		1806	484	0.1
3200–3300	1109	0				
3300–3400	1179	0				
Volume(1000 m^3)	2679	1113	3792			
Area (km^2)	107	13	120			
Mean specific balance (mm)	25	88				
Referring to basin area (mm)	22	9	32			
Equilibrium line altitude	2770					
Accumulation area ratio	0.69					

In order to translate these observations into the present model, snow has to be taken from the ice free areas and added to the glacier surface, the question being how much and wherefrom to where. Having observed that the ratio of net accumulation to simultaneous precipitation as measured in glacier basins may exceed a factor of two (Kuhn, 1981) and keeping in mind that according to Bernoulli's principle basins receive more precipitation than ridges I made the first guess that glaciers receive twice as much precipitation as the basin average, regardless of altitude and time of the year. The volumes of solid precipitation in the basin VP, on the glacier VPG and in ice free areas VPU are then

$$\mathrm{VP}(h,\mathrm{mo}) = P(h,\mathrm{mo})^{*}A(h) \tag{3.1}$$

$$\mathrm{VPG}(h,\mathrm{mo}) = 2^{*}P(h,\mathrm{mo})^{*}\mathrm{AG}(h) \tag{3.2}$$

$$\mathrm{VPU}(h,\mathrm{mo}) = \mathrm{VP}(h,\mathrm{mo}) - \mathrm{VPG}(h,\mathrm{mo})$$

$$= P(h,\mathrm{mo})^{*}(A(h) - 2^{*}\mathrm{AG}(h)) \tag{3.3}$$

and the respective solid precipitation in mm is

$$\mathrm{PG}(h,\mathrm{mo}) = \mathrm{VPG}(h,\mathrm{mo})/\mathrm{AG}(h) \tag{3.4}$$

$$\mathrm{PU}(h,\mathrm{mo}) = \mathrm{VPU}(h,\mathrm{mo})/(A(h) - \mathrm{AG}(h)) \tag{3.5}$$

where A is the total area and AG the glacier area, respectively. At some altitudes where the glacier area is more than half of the total area, the value of Eq. (3.3) may become negative which can be compensated by manually redistributing the volume to other altitudes.

This redistribution procedure was applied to the Paznaun basin (46°55′N, 10°10′E, 1700–3400 m a.s.l., 11% glacier area) which had previously been well calibrated (Kuhn, 2000). In this basin it turned out that the first guess factor of two had to be changed to a value of 2.15 before it would plausibly produce the characteristic values of glacier mass balance of this basin. In general, the factor of 2 in Eqs. (3.2) and (3.3) has to be replaced by a 'redistribution

Table 3
The sensitivity of model mass balance (*b*) and equilibrium line altitude (ELA) to the positive degree day factor

Positive degree day factor (mm/pdd)	3		4		5		6		7
b (mm/a)	983		546		89		−377		−836
d*b*/d pdd		−437		−457		−466		−459	
ELA (m)	2524		2666		2773		2863		2931
d ELA/d pdd		142		107		90		68	

factor' whose value depends on basin topography and has to be inferred from tuning experiments.

4. Model results for the Paznaun basin

4.1. The water balance

Table 1 shows the most important stages of the model as well as all tuning parameters. The development of the first approximation has been outlined in Section 2. Annual storage enters as $(\Sigma bref(h)^{*} AG(h))$, the sum of the product of the specific balance of a reference glacier at height h and the glacier area of the basin investigated over all altitude intervals. Monthly values of storage $S(h, \text{mo})$ are calculated as residual of the water balance in the first approximation. The seasonal course of precipitation $P'(\text{mo})$ and of temperature $T(\text{mo})$ are taken from reference stations close to the basin, temperature being converted to the lowermost altitude of the basin. Changes of precipitation and temperature with altitude are averaged from the regional station network and kept constant at all altitudes.

In the final approximation precipitation and evaporation are slightly modified. Runoff is calculated from rain and meltwater and storage results as balance residual. The agreement of calculated and observed runoff $Q(\text{mo})$ must be better than 20 mm before the approximation is accepted.

In the last two lines an attempt was made to manually separate total storage into respective liquid and snow and ice components with three vague constraints: 1. winter runoff must come from liquid storage, 2. base flow must decrease asymptotically during winter and 3. negative values of snow and ice storage must not exceed calculated melt. Results are thus speculative, but have no influence on the determination of the glacier mass balance presented in the following section.

4.2. Glacier mass balance

Table 2 reproduces the volumes of water equivalent of snow and ice that have been deposited or melted for the glacier area and for the ice free areas of the basin as an average of the period 1961/62–1991/92. Mean specific glacier mass balance was positive for that period as on other glaciers of the Alps. While some glaciers were actually advancing at the terminus, large areas above 3100 m accumulated snow and ice: of the total storage change in the basin (3.79×10^6 m^3 of water equivalent), 2.68×10^6 m^3, or about 70% had accumulated outside the glacier area proper.

Specific balance $b(h)$ attains values that are comparable to the results of measurements. The altitudinal gradients $db(h)/dh$ are in the range expected for glaciers with positive mass balance, they are larger in the ablation area and are nearly constant above the equilibrium line. With linear interpolation the equilibrium line altitude is approximately 2770 m a.s.l., the accumulation area ratio is 0.7.

5. Model sensitivity to tuning parameters

The degree day factor. The role of the positive degree day factor as a tool for tuning glacier mass balance has recently been illustrated by Braithwaite and Zhang (2000) for five Swiss glaciers. Table 3 summarizes the present analysis which shows that it has a strong influence on both specific balance and equilibrium line altitude. While the model sensitivity of specific balance to the degree day factor is nearly

Table 4
The sensitivity of model mass balance (b) and mean mass balance gradient db/dh to changes in the redistribution factor RF.

RF	2.0		2.1		2.2
b (mm/a)	−90		29		148
db/dRF		1190		1190	
< db/dh > (mm/100m)	510		518		524

The mean mass balance gradient < db/dh > was determined as (b(3150 m)−b(2450 m)/7

constant, the sensitivity of the equilibrium line is strongly reduced with increasing degree day factors. While in Table 3 only whole numbers have been used, the minimum difference between observed and calculated runoff occurs at a degree day factor of 5.1 mm w.e./K.

The redistribution factor. The redistribution factor determines how much the amount of solid precipitation is increased on the glacier surface (Eq. (3.2)) compared to basin average. Table 4 shows a nearly linear increase of specific balance when this factor is increased in steps of 0.1, corresponding to 5% more solid precipitation for each altitude interval. As solid precipitation increases with altitude, an increase of the mean balance gradient follows an increase of the redistribution factor.

The transition from liquid to solid precipitation. Eq. (2.2) determines how much of the total precipitation falls in the form of snow. Changes of $K1$ will have the strongest effect at times when temperature is around freezing, and not in the winter, nor in the lower parts in summer. The reaction of mass balance therefore is not linear with $K1$ as is obvious from Table 5. It is interesting to watch the partition of rain water and melt water runoff in Table 5 while the fraction of solid precipitation increases, so does the melt water runoff; rain water runoff strongly decreases and total runoff is slightly diminishes in response to the more positive mass balance of ice and snow.

Table 5
The sensitivity of model mass balance b and runoff Q to changes in the rain/snow threshold $K1$ in Eq. (2.2)

$K1$	0.5		0.6		0.7
b (mm)	−136		89		271
Difference (mm)		225		182	
Storage in ice free areas (mm)	20		22		24
Q melt (mm)	758		829		884
Difference (mm)		71		55	
Q rain (mm)	478		382		306
Difference (mm)		−96		−76	
Q total (mm)	1236		1211		1190

6. Model response to changes in temperature

The response of the water cycle of this basin to various temperature and precipitation scenarios has been the subject of previous publications (Kuhn and Batlogg, 1999; Kuhn, 2000); in the case of warming it is characterized by increased base flow in winter, an earlier rise of runoff due to earlier melting and lower peak runoff which is flatter as the contribution of the precipitation maximum stays fixed in July and August; the response to warming and cooling is not symmetric.

Table 6 gives a survey of the response of mass balance and runoff components to seven temperature scenarios. Under present conditions the sensitivity of mean specific mass balance to a 1 K change is approximately 800 mm yr^{-1} K^{-1}. The mean specific mass balance reacts very strongly to elevated temperatures and is less sensitive to cooling scenarios, a possible explanation being that there is no limit to increasing melting in warm scenarios whereas the prescribed total basin precipitation sets a limit to positive mass balance in cold scenarios independent of temperature.

The opposite is true for the storage outside glacier areas. Here, an exceptional increase happens from T-2 to T-3 when the snow line is lowered to altitudes with a very large percentage of total basin area. The area-altitude distribution of the ice free areas seems to be responsible for the uneven course of the differences between scenarios in the upper right of Table 6, and is reflected in the lower limit of storage on ice free areas.

Equilibrium line altitudes respond to temperature changes with a sensitivity of approximately 150 m K^{-1} which is in agreement with the altitudinal gradients of both accumulation and temperature (Kuhn, 1981). As they were determined by linear interpolation between 100 m levels their accuracy is less than might be expected from the four significant digits given in Table 6.

Runoff and its melt and rain water components behave in some ways different than what was seen in the discussion of the rain/snow threshold. In the six

Table 6
Sensitivity of model mass balance to temperature scenarios

	Glacier mass balance mm		Storage in ice free areas	
	mm	Difference	mm	Difference
$T+3$	−3754		0	
		1989		3
$T+2$	−1765		3	
		986		9
$T+1$	−779		12	
		868		11
T	89		23	
		717		7
$T-1$	806		30	
		664		7
$T-2$	1470		37	
		505		74
$T-3$	1975		111	

	Equilibrium line altitude		Lower limit of storage in ice free areas (m)
	m	Difference	
$T+3$	>3200		>3400
$T+2$	3067		3300
		−130	
$T+1$	2937		3200
		−164	
T	2773		3200
		−148	
$T-1$	2625		3200
		−150	
$T-2$	2475		3100
$T-3$	<2400		2600

Runoff mm	Melt water	Rain water	Total modelled
$T+3$	1038	597	1635
$T+2$	942	481	1423
$T+1$	882	429	1311
T	829	382	1211
$T-1$	784	344	1128
$T-2$	745	307	1052
$T-3$	648	277	925

Temperature has been changed by the same amount in all months.

degrees span of the temperature scenarios both components and thus total runoff decrease from the warmest to coldest scenario. Given a fixed basin precipitation and little leeway in the evaporation, the decrease of total runoff must be seen as the result of both glacier mass balance and the large storage of snow outside the glacier where ice free areas cover a total of 89% of the basin.

7. Discussion and conclusions

The model used for this study had previously been calibrated and tested in hydrological applications. It has measured values of monthly runoff as a reliable constraint and determines total basin precipitation as another boundary condition. Although the former routine produced reasonable total basin storage, details of glacier mass balance used to be unsatisfactory.

This was improved by internal redistribution of solid precipitation from ice free areas to the glacier. The redistribution was kept simple and within altitude intervals, without vertical exchange. The same redistribution factor was applied to all altitudes, its value being determined for this basin as 2.15.

Mass balance and balance gradients produced in this experiment appear plausible. Gradients lie within limits that result from direct measurements, they are, however directly affected by the redistribution algorithm used. Future model experiments will be needed to show whether any refinement in the vertical distribution of the mass transfer can make significant improvements. Equilibrium line altitudes in this model display the expected reaction to the vertical changes of temperature and accumulation.

The model uses four tuning parameters: a positive degree day factor, a redistribution factor, the rain/-snow threshold temperature, and the snow covered fraction of the ice free area. Their impact on the model results varies with the time of the year so that it is justified and necessary to use all four of them for tuning.

The model was calibrated with a thirty years data set, parameters being adapted to average values imply that modelling of individual years may fail to give accurate results. It is rather meant to be used in climate scenarios and for the explanation of differences in the glacier and runoff behavior in different parts of the Alps. It is well suited for class work where students may learn to understand the effects of the various meteorological and hydrological variables on glacier mass balance and runoff.

Acknowledgements

This paper is a contribution of the Glaciology Group to the German Glowa–Danube Program. The original modelling work was supported by Verbundgesellschaft, Vienna.

References

Braithwaite, R.J., Zhang, Yu., 2000. Sensitivity of mass balance of five Swiss glaciers to temperature changes assessed by tuning a degree-day model. Journal of Glaciology 46 (152), 7–14.

Finsterwalder, R., Rentsch, H., 1993. Zur Hoehenaenderung von Ostalpengletschern im Zeitraum 1979–1989. Zeitschrift fuer Gletscherkunde und Glazialgeologie 27/28, 165–175.

Hoinkes, H., 1971. Methoden und Moeglichkeiten von Massenhaushaltsstudien auf Gletschern. Zeitschrift fr Gletscherkunde und Glazialgeologie VI (1–2), 37–90.

Jansson, P., Dyurgerov, M., Fountain, A.G., Kaser, G., 1999. Methods of mass balance measurements and modelling. Physical Geography, Geografiska Annaler 81 (A/4), 796.

Kuhn, M., 1981. Climate and glaciers. IAHS Publication 131, 3–20.

Kuhn, M., 1987. Micro-meteorological conditions for snow melt. Journal of Glaciology 33 (113), 24–26.

Kuhn, M., 2000. Verification of a hydrometeorological model of glacierized basins. Annals of Glaciology 31, 15–18.

Kuhn, M., Batlogg, N., 1998. Glacier runoff in Alpine headwaters in a changing climate. IAHS Publication 248, 79–88.

Kuhn, M., Batlogg, N., 1999. Modellierung der Auswirkung von Klimaaenderungen auf verschiedene Einzugsgebiete in Oesterreich. Schriftenreihe Forschung im Verbund, 98.

Kuhn, M., Pellet, F., 1987. Basin precipitation as residual in Alpine water balances. In: Sevruk, B., (Ed.), Precipitation Measurements. WMO/IAHS/ETH Workshop on precipitation measurements, St Moritz.

Kuhn, M., Nickus, U., Pellet, F., 1982. Die Niederschlagsverhaeltnisse im inneren Oetztal 17. Internat. Tagung fuer Alpine Meteorologie, 235–237.

Lang, H., Patzelt, G., 1971. Die Volumenaenderungen des Hintereisferners (Oetztaler Alpen) im Vergleich zur Massenaenderung im Zeitraum 1953–64. Zeitschrift fuer Gletscherkunde und Glazialgeologie VII (1–2), 39–55.

Meier, M., Tangborn, W., Mayo, L., Post, A., 1971. Combined ice and water balances of Gulkana and Wolverine Glaciers, Alaska and South Cascade Glacier, Washington, 1965 and 1966 hydrologic years. US Geological Survey Professional Paper 715A, 23.

Oestrem, G., Brugman, M., 1991. Glacier mass balance measurements: a manual for field and office work. National Hydrology Research Institute Science Report 4, 224.

Roethlisberger, H., Lang, H., 1987. Glacial hydrology. In: Gurnell, A.M., Clark, M.J. (Eds.), Glacio-Fluvial Sediment Transfer, Wiley, Chichester, pp. 207–284.

Sevruk, B., 1983. Correction of measured precipitation in the Alps using the water equivalent of new snow. Nordic Hydrology, 49–58.

Zwally, H.J., Brenner, A.C., 1999. The role of satellite radar altimetry in the study of ice sheet dynamics and mass balance. In: Fu, L.-L., (Ed.), Satellite Altimetry And Earth Sciences, Academic Press Inc, New York, pp. 351–369.