

SNOWMELT SIMULATION OF SHORT LIVING SNOWPACKS

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ABSTRACT

In view of the recent evolution of increased interest for flood regulation by means of real time controlled flood reservoirs, a snowmelt model has been developed which meets the requirements for real time simulation and forecasts.

The considered snowmelt model is a point model which is based on the energy budget method. The snowpack is considered to be isothermal. The different physically based relations which describe the energy exchange processes are discussed. The snowmelt model uses measured rainfall, air temperature, air and vapor pressure, wind velocity, global radiation and an index to take account of the cloudiness as input. The output consists of the temperature of the pack, the calculated heat budget components, the snow thickness and the runoff from the pack. Since medium sized river basins are aimed at, the timebase of the simulation is kept to one hour, although this timebase has to be reduced for reasons of numerical stability whenever the water equivalent of the pack drops too low.

The snowmelt model is subsequently linked to the real time rainfall-runoff model developed by the Laboratory of Hydrology of the Free University of Brussels. Simulations of the snowmelt runoff on historical records on the Dijle river illustrate the adequacy of the model. The results of these simulations are discussed.

INTRODUCTION

Snowmelt flooding is not frequent in the Northern part of Belgium with its mild and rainy climate. During the past 12 years, the snowcover exceeded 20 cm at only three occasions and it never exceeded 35 cm during this period.

When looking for historical floods in the Dijle basin, in view of the real time management of a storm reservoir on that river, however, we established that nearly all the important floods were linked to snowmelt or rain on frozen soil.

Flooding of major rivers in that part of Belgium in general, and of the Dijle in particular, may occur whenever the rainfall exceeds about 30 mm in 24 hours, which corresponds to a storm with a return period of between five and ten years. When a snowpack of sufficient height is built up and

subsequently melted, due to the passage of a warm front accompanied with abundant rainfall, multiples of this water amount may be liberated in a very short time period. The river then faces water amounts which correspond to equivalent rainstorms with a return period of a few hundreds of years.

It was thus decided that a snowmelt routine, which meets the requirements for real time simulations, should complete the rainfall-runoff model of the Dijle river.

Although some index methods have been tested, we believed it advisable to work out a deterministic snowmelt model, completed with the necessary semiempirical relations, especially in view of the lack of data to calibrate any model.

The paper describes the model worked out so far and shows the results of a simulation.

THE SNOWMELT MODEL

The local, one-dimensional form of the equation of heat flow in a snowcover, where only the liquid and the solid component are considered, can be expressed as :

$$\begin{aligned}
 & (c_{pw} \rho_w \theta_w(x,t) + c_{pi} \rho_i \theta_i(x,t)) \frac{\partial T(x,t)}{\partial t} = \\
 & c_{pw} \rho_w v(x,t) \frac{\partial T(x,t)}{\partial x} + \frac{\partial}{\partial x} (\lambda(x,t) \frac{\partial T(x,t)}{\partial x}) + \Delta Q Q(x,t) \\
 & + L_f \rho_i \frac{\partial \theta_i(x,t)}{\partial t} + \rho_i (c_{pw} - c_{pi}) (T(x,t) - T_o) \frac{\partial \theta_i(x,t)}{\partial t} \quad (1)
 \end{aligned}$$

where $\theta(x,t)$ corresponds to the volumetric water (index w) or ice (index i) content at height x and time t; $T(x,t)$ is the temperature and $v(x,t)$ is the mass flux. Other variables and constants are : ρ , the density; C_p , the specific heat; $\lambda(x,t)$ the thermal conductivity; L_f , the latent heat of fusion and T_o , a reference temperature. $\Delta Q Q(x,t)$ stands for all external heat inputs.

When the equation of mass flow is added, one becomes a set of equations which may be solved simultaneously, taking the proper initial and boundary conditions into account. This method of coupling the mass and heat transport in a snowpack has been worked out in view of the SHE-model (Morris, 1982). With the present knowledge of snow characteristics however, a number of phenomenological relations which are needed to solve the problem remain open to doubt. Some of the major drawbacks concern the flow equation to be used, the characteristic curve, the link between the temperature and the capillary pressure and the assessment of the initial conditions to newly fallen snow. Also should the snow metamorphism be modelled as the hydraulic conductivity is, among others, a function of the grain size.

Obléd and Rosse (1977) solved the heat flow equation in a snowpack and treated the mass flow in a pragmatic way. With this procedure, the main drawback of the isothermal energy budget method is avoided, but the problem of surface melt or rain on a snowpack and subsequent refreezing

does not find a rigorous solution. The authors found out that the results of their isothermal and their non-isothermal model compared well, except in the presence of prolonged frost.

In the isothermal energy budget approach, an isothermal snowpack at some global temperature $T_S(t)$ is assumed as a working hypothesis. Integrating (1) over the height of the pack with this assumption leads to :

$$L(t)H(t) \frac{\partial T_S(t)}{\partial t} = QQ(t) + L_f \rho_i L \frac{\partial \theta_i(t)}{\partial t} \quad (2)$$

where

$$H(t) = c_{pw} \rho_w \theta_w(t) + c_{pi} \rho_i \theta_i(t) \quad (3)$$

and L is the height of the snowpack.

In an explicit form, (2) may be written as :

$$T_S(t+\Delta t) = T_S(t) + \frac{QQ(t)}{L(t)H(t)} + \frac{L_f \rho_i (\theta_i(t+\Delta t) - \theta_i(t))}{H(t)} \quad (4)$$

which may be easily solved. Indeed, when $T_S(t+\Delta t)$ is beyond the freezing point, no melt occurs and thus :

$$\theta_i(t+\Delta t) = \theta_i(t)$$

When the computed temperature at time $(t+\Delta t)$ exceeds the freezing point, snow will be melted as to keep $T_S(t+\Delta t)$ at the freezing point.

The assumptions made in the isothermal energy budget method lead to inaccuracies both in the mass flow and in the heat balance of the snowpack. As will be seen in the next chapter, the external heat input term QQ consists of shortwave and longwave radiation, turbulent fluxes, advection and heat conduction at the snow soil interface. Except for the shortwave radiation and the advection, these heat fluxes depend on the temperature of the surface layer they are acting on, and not on the global or mean temperature of the snowpack. Heating and cooling of the pack are induced from these sublayers but, due to the low thermal conductivity of snow, strong thermal gradients may exist in the pack. Treating the pack as isothermal will lead to an underestimation of the surface temperature when the pack is warmed up and an overestimation when the pack cools off, thus leading to inaccuracies with respect to the superficial energy exchanges.

Surface melting, when the global cold content is still negative, may not be modelled by this type of model, though this phenomenon plays an important part in the redistribution of water in the pack and in the snow metamorphism. Finally, the mass flow in the pack is treated in an empirical way.

THE EXTERNAL HEAT INPUTS

Shortwave radiation

When data of both albedo and global radiation are available, the net heat gained by the pack through solar radiation is simply calculated as :

$$Q_k = (1-a)R_k$$

where a is the albedo and R_k is the global radiation on a horizontal plane.

It is thus implicitly assumed that the snowpack is sufficiently deep as to absorb all the shortwave radiation. Literature data about the extinction of radiation in a snowpack are quite divergent (Yin Chao Yen, 1969). Realistic values for the minimal depth required are 5 to 15 cm for densities of 0.2 to 0.4 g/cm³ respectively.

Longwave radiation

A snowpack may be considered to behave as a black body with reference to the longwave radiation it emits. The emitted radiation is given by the Stefan law :

$$R_S = \sigma T_S^4 \quad (6)$$

where σ is the Stefan-Boltzmann constant and T_S is the temperature of the snow surface layer. Since this temperature is not evaluated in an isothermal snowmelt model, the global temperature of the pack will be used instead.

A similar equation may be applied for the incident atmospheric radiation, provided the use of an effective emissivity :

$$R_a = \epsilon \sigma T_a^4 \quad (7)$$

where T_a is the atmospheric temperature. Theoretically, a temperature which is representative for the lower atmosphere up to 1 or 2 km should be used (Kondratyev, 1969). Practically however, temperature profiles over this height are not available for real time simulations and the temperature near the ground may be used as a good approximation (Charbonneau et al., 1974)

The effective emissivity of the atmosphere is often defined as the product of the emissivity of a cloudless sky, ϵ_0 , and a correction factor ϵ_c which takes the influence of the cloudiness into account.

A number of semiempirical relations commonly used in snowmelt models have been reviewed by Male and Granger (1981). A statistical analysis was applied to confirm their and other authors' conclusions. The analysis is based on daily measurements from Aase and Idso (1978) and on data on a 10' timebase from Swinbank (1963). The global dataset consists of 37 measurements, covering temperatures between -30 and +10°C.

Table 1 shows the main statistical parameters obtained by comparison of measured and calculated clear sky emissivities. It may be concluded from this table that neither of these relations yields excellent estimates of the emissivity. The best estimates are obtained by the equations of Satterlund and Idso.

A residual analysis points out that the equations of Brunt and Swinbank show similar trends. Both equations underestimate the emissivity systematically by about 10% for positive temperatures and by 15% for temperatures lower than -20°C. In between, the deviations may reach as high as 30%. Concerning the Brunt equation, one should keep in mind Kondratyev's remark about its strict applicational limits as to temperature field and water vapor content. Special care should be taken when choosing the parameters of this formula. As, to our knowledge, no parameters are available for application of the Brunt formula over a snowfield, the parameters defined by Monteith (1961) were used in the analysis.

The Brutsaert equation also underestimates the emissivity by about 5% at the high temperatures, to 25% at the low extremes. These deviations should be explained by the author's hypotheses of "standard" temperature and water vapor profiles, which do not hold at extreme temperatures.

The equation of Idso and Jackson yields fair results for temperatures higher than -15°C , but overestimations of 40% are observed at temperatures below this value. The experimental data show clear evidence that the author's assumption of minimal emissivity near 0°C and increasing emissivity below this temperature does not hold.

The residuals as a function of temperature for the Satterlund equation are given in Fig.1. The equation gives the emissivity mainly as a function of vapor pressure, the influence of the temperature being very weak :

$$\epsilon_o = 1.08(1 - \exp(-e_a^{T_a/2016}))$$

where e_a is the vapor pressure (in mb) and T_a is the air temperature (in $^{\circ}\text{K}$). In view of the data published by Idso (1981) it moreover seems that the functional relation between ϵ_o and T_a is not correct, since the equation predicts ϵ_o to increase with temperature. Once more, it seems that the relative success of this equation is due to the correlation between e_a and T_a . It should also be stated that Satterlund optimized his equation partly on the same data as those used in this statistical analysis. This might explain the relative success of the equation according to the analysis.

Figure 2 shows the residuals as a function of the temperature for the Idso relation :

$$\epsilon_o = 0.7 + 5.95 \cdot 10^{-5} e_a \exp(1500/T_a)$$

As for his previous equation, one remarks a systematic overestimation for extremely low temperatures. It appears from the data that the minimal emissivity of 0.7 used in the equation is too high.

As a conclusion of this analysis it seems that most of the equations are not adapted to low temperature circumstances. This may be explained partly by the use of the ground air temperature, which may not always be representative for the atmospheric layer of interest, but also by the unfitness of the functional relations used. Taking the above mentioned restrictions into account, Satterlund's formula gave the best result, so this equation was used in the model.

A semiempirical correction factor for the cloudiness can be written in a general form as :

$$\epsilon_c = 1 + \sum_i c_i C_i$$

where c_i are coefficients which depend on cloud type and C_i are the cloud covers at various levels i . When only a global cloud cover report is available, the equation reduces to :

$$\epsilon_c = 1 + cC$$

with C the global cloud cover and with $0.18 \leq c \leq 0.31$ (Kondratyev, 1969). A value of 0.25 is used here.

The net longwave radiation is expressed as the balance of incoming and outgoing radiation :

$$Q = R_a - R_s$$

Turbulent mass transfer

The complex phenomena involved with turbulent transfers can be described approximately by :

$$E = f(u_a)(e_s - e_a)$$

where E corresponds to the mass transfer, e_s is the vapor pressure at the snow surface and e_a is the vapor pressure at some level above the snowpack.

The use of this type of expression is impeded by the optimization of the wind speed function $f(u_a)$ for the space, time and meteorological conditions of interest.

The aerodynamic approach, based on the analogy between molecular diffusion and the dispersion of matter by turbulence, may prove to offer an alternative. Using the Prandtl theory for the turbulent shear stress and assuming logarithmic profiles of windspeed, temperature and vapor content, the heat flux associated with turbulent mass transfer may be written as (e.g. Brutsaert, 1982) :

$$Q_e = -L \frac{K_e}{K_m} \frac{k^2 u_a (q_s - q_b) \rho_a}{\ln\left(\frac{z_a - d}{z_{om}}\right) \ln\left(\frac{z_b - d}{z_{ov}}\right)}$$

where q_s (q_b) corresponds to the specific humidity at the snow surface (at level Z_b above the snow surface), u_a is the windspeed at level Z_a , ρ_a is the air density and k is the von Karman constant. The calculation of e_s is based on the saturated vapor pressure above ice at the temperature of the snowpack.

The ratio of the turbulent diffusion coefficients, K_e/K_m , tends to decrease somewhat with increasing stability of the atmospheric layer, as was reviewed by Male and Granger (1981). At neutral conditions, and for $k = 0.4$, all measurements converge to 0.16. According to Brutsaert (1982) this value may be extended to stable conditions. For unstable conditions, which seldom occur over a snowpack, a correction factor is applied :

$$\frac{K_e}{K_m} = 0.16 (1 - 16z)^{0.25}$$

where z is a stability parameter, derived from the Richardson number and approximated as :

$$z = \frac{g}{T_a} z_a \frac{T_a - T_s}{u_a^2}$$

where g is the gravitational constant.

The value of the surface roughness of a snow surface, Z_{om} , ranges from 0.1 to 0.5 cm (Konstantinov, 1963), but may reach 2 cm for soft snowpacks. Values of 0.25 cm, and 1 cm under melt circumstances are used for the simulations.

The surface roughness for vapor transport is related to Z_{om} by :

$$Z_{ov} = 7.4 Z_{om} \exp(-2.25 Re^{0.25})$$

where Re is the Reynolds number (Brutsaert, 1982). The reference level d_0 is neglected. For L, the latent heat of sublimation or the latent heat of vaporisation is used, depending on the temperature of the snowpack and on the availability of liquid water in the pack.

Turbulent heat transfer

The sensible or convective heat flux is obtained in the same way as the flux related to the turbulent mass transfer (Brutsaert, 1982) :

$$Q_c = -\rho_a \frac{K_h}{K_m} \frac{u_a C_{pl} (T_s - T_a)}{\ln\left(\frac{z-d}{z_{oh}}\right) \ln\left(\frac{z-d}{z_{om}}\right)}$$

where C_{pl} is the specific heat of dry air and Z_{oh} is the surface roughness height for heat transport. Again, according to Brutsaert :

$$Z_{oh} = 7.4 Z_{om} \exp(-2.46 Re^{0.25})$$

It is further stated that $K_e = K_h$

Heat conduction at the soil-snow interface

The heat conduction term is small in comparison to the other terms of the heat balance. Its evaluation necessitates the knowledge or, at least, a proper estimate of the temperature gradient over the interface and of the thermal conductivities at this interface. Since an isothermal snowmelt model may not provide these data, a constant heat flux of $7100 \text{ J/m}^2\text{h}$ is assumed (e.g. O'Neill, 1971), which corresponds to a melt of 0.5 mm waterequivalent a day.

Advection

Precipitation of intensity P_i and temperature T_n implies a sensible heat flux :

$$Q_p = P_i \rho_w C_i (T_n - T_s)$$

where i stands for the type of precipitation (liquid vs. solid), C_i is the specific heat of the given type of precipitation and ρ_w is the density of water. The wet bulb temperature is used as the temperature of the precipitation.

The sensible advective heat flux is small compared to the other terms of the balance. Rainfall falling on a snowpack may however strongly affect the heat balance through the release of latent heat when refreezing of rain takes place in the snowpack. Depending upon the rainfall intensity and on the water equivalent of the pack, the liberation of this heat may cause a rise of the temperature of the pack by several degrees.

The definition of the precipitation type is one of the major drawbacks for the automatic real time simulation of the snowmelt process, since it will strongly affect the formation (or not) of a new snow layer and the heat balance of an existing snowpack.

Unfortunately, there is no index which allows a unique differentiation between rainfall and snowfall. As with other variables tested, surface air temperature shows a range over which either rain or snow can occur.

Fig. 3 shows the frequency of occurrence of rain and snow as a function of air temperature. The data for the Royal Meteorological Institute at Ukkel, Belgium were obtained from the analysis of hourly precipitation and temperature data over the period 1972-1977. This curve shows no mixed area since the form of the precipitation is given on a daily timebase only. The analysis retained only the days where a single form of precipitation occurred.

WATER MOVEMENT THROUGH THE PACK

Considering the timestep used for the runoff computations (one hour) and the concentration time of the target river basin (16 hours), no effort was put in the modelling of the routing of the liquid water through the shallow snowpacks considered here. It is thus simply assumed that the water available from melting during an hour reaches the soil during the same hour.

A certain lag time is however obtained through the modelling of the retention of liquid water in the snowpack as a function of the snow density, according to the data of Laramie and Schaake (1972).

THE RAINFALL-RUNOFF MODEL

The melt calculated by the snowmelt model is converted to a runoff hydrograph by means of the conceptual real time rainfall-runoff model elaborated at the Hydrology Department of the Free University of Brussels. This model has been described in detail elsewhere (Marivoet and Vandewiele, 1980; Bauwens et al., 1983), so only a brief description of its main features is given here.

The model initiates from Horton's concept of direct and indirect runoff. The amount of direct runoff is calculated as the balance term of the net precipitation and the groundwater recharge. The latter is calculated as a fraction of the net precipitation and is a function of the soil moisture index. The distribution of the direct runoff according to time is achieved by means of a lognormal unit hydrograph. Baseflow runoff is considered constant and equal to the runoff at the start of the rainstorm event.

The rainfall-runoff model was calibrated on an hourly timebase upon 72 rainstorm events, covering the more intense floods of a five year period. This dataset did not include any snowmelt events.

NUMERICAL STABILITY

As stated before, the heat flow equation is solved by an explicit finite difference form (3). It may be proven that for this method to be numerically stable, Δt must be chosen such that

$$\Delta t < 1/a$$

where

$$T(t+\Delta t) = (1-a\Delta t)T(t) + c\Delta t.$$

An additional stability condition which implies positive values for c is always fulfilled for the model presented here. The coefficient a is highly variable and may even prove to be negative under specific meteorological conditions. It may however be shown that, for standard meteorological inputs, Δt should be chosen as a function of the waterequivalent of the pack. A minimal timestep of 1' is used for the

simulations when the waterequivalent is less than 1 mm. The timestep is continuously increased as a function of the waterequivalent up to 10' when the latter reaches 10 mm. For higher waterequivalents 20', 30' and one hour are used.

DESCRIPTION OF THE RIVER BASIN AND DATASET USED FOR THE SIMULATIONS

The upstream part of the Dijle river basin (Fig. 4), on which the model was tested, has an area of 750 km² and elevation extremes of 22 and 160 m above sea level. The land use in the basin is mainly agricultural, but some forested (less than 10 %) and urbanized areas are included. The soil is loamy above a sandy aquifer.

Meteorological data from the Royal Meteorological Institute at Ukkel, which lies just outside the Dijle basin, were used for the simulations. The original dataset consisted of hourly precipitation, albedo and global radiation; two-hourly data of air temperature and pressure, wind velocity and vapor pressure and three-hourly cloudiness values. The data on a timebase larger than one hour were reduced to hourly values by interpolation.

A SIMULATION EXAMPLE

The snowmelt model was applied to a ten day period (november 27 to december 6, 1973). The first three days of this episod were characterized by nearly continuous snowfall, at the end of which a snowpack with a waterequivalent of 40 mm was built up. The temperature then dropped to -5 to -10°C for the following three days, after which the pack melted, due to a gradual increase of the temperature up to +8°C and rainfall on the snowpack.

The balance of the external heat fluxes is shown on Fig. 5. The high positive peaks before the final melting of the pack (time 180) are mainly due to the absorption of shortwave radiation. It can be seen on Fig. 7 that these peaks cause midday and early afternoon melting at some occasions.

When looking at the different components of the balance, one notices that the longwave radiation balance is predominantly negative. Only at the end of the episod, when the air temperature exceeds the snowpack temperature by several degrees, this component becomes significantly positive. The latent heat flux associated with turbulent mass transfer, undergoes a diurnal variation with condensation during the night and early morning and evaporation or sublimation during daytime. Since the air temperature exceeds the snow temperature during most of the time (Fig. 6) the convective component is mainly positive. Solar radiation evidently shows a diurnal variation with a marked influence of the cloudiness. The remaining heat fluxes are small compared to the previously mentioned fluxes as can be seen on table 2, which shows the global balance of the different components over the episod.

The runoff from the snowpack, including meltwater and non-refrozen rain, is given on Fig. 7, together with the simulated and observed runoff hydrographs. The good correspondance between the rising limbs and the peaks of the hydrographs may be observed. The recession is characterized by a systematic underestimation of the runoff, what leads to an underestimation of the mass balance by about 10 %. Considering the uncertainty about the total amount of snowfall in the basin, the fact that a point snowmelt model is used and the stochastic character of the runoff model-parameters the result can be judged favourably.

CONCLUSIONS

It appears from this paper that the building, the evolution and the final melting of shallow snowpacks can be simulated quite well using the isothermal energy budget approach. Moreover, the model proposed in this paper uses only meteorological variables which can be measured automatically in real time. This makes it possible to use the model in conjunction with a runoff model, as to simulate runoff hydrographs in real time.

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REFERENCES

- ANONYMOUS, Snow hydrology, Summary report of snow investigations, Corps of Engineers, U.S. Army, North Pacific Division, Portland, Oregon, U.S., 1956.
- AASE, J.K. and IDSO, S.B., A comparison of two formula types for calculating longwave radiation from the atmosphere, Water Res. Res., Vol. 14, n°4, p. 623-625, 1978.
- BAUWENS, W, MARIEN, J., MARIEVOET, J., VANDEWIELE, G.L. and VAN DER BEKEN, A., The real time control of the reservoir on the river Dijle, Scientific Procedures Applied to the Planning, Design and Management of Water Resources Systems, Proc. Hamburg Symp., IAHS at IUGG, August 1983 (in press).
- BRUNT, D., Notes on radiation in the atmosphere, Q.J.R. Met. Soc., Vol. 58, p. 389-420, 1932.
- BRUTSAERT, W., On a derivalbe formula for longwave radiation from clear skies, Water Res. Res., Vol. 11, N°5, p. 742-744, 1975.
- BRUTSAERT, W., Evaporation into the atmosphere, D. Reidel Publishing Company, Dordrecht, Holland, 299 pp., 1982.
- CHARBONNEAU, R., LARDEAU, J.P. and OBLED, C., Problems of modelling a high mountainous drainage basin with predominant snow yields, Hydrological Sciences Bulletin, Vol. 26, N°4, p. 345-361, 1981.
- IDSO, S.B., A set of equations for full spectrum and 8-14 um and 10.5-12.5um thermal radiation from cloudless skies, Water Res. Res., Vol. 17, N°2, p. 295-304, 1981.
- IDSO, S.B. and JACKSON, R.D., Thermal radiation from the atmosphere, J. Geophys. Res., Vol. 74, p. 5397-5403, 1969.
- KONSTANTINOV, A.R., Evaporation in nature, Israel Program for Scientific Translations, 523 pp., 1966.
- LARAMIE, R.L. and SCHAAKE, J.C.jr., Simulation of the continuous snowmelt process, M.I.T., Dpt. of Civil Eng., Massachusetts, U.D., Report n°143, 167 pp., 1972.
- MALE, D.H. and GRANGER, R.J., Snow surface energy exchange, Water Res. Res., Vol. 17, n°3, p. 609-627, 1981.
- MARIVOET, J.L. and VANDEWIELE, G.L., A real time rainfall-runoff model, IAHS Publ. N°129, Proc. of the Oxford Symposium, p. 409-418, 1980.
- MONTEITH, J.L., Dew, Q.J.R. Met. Soc., Vol 83, p. 322-341, 1957.

- MORRIS, E.M., Sensitivity of the European Hydrological System snow models, Hydrological Aspects of Alpine and High Mountain Areas, Proc. Exeter Symp., IAHS, Publ. n°138, p. 221-231, 1982.
- OBLED, CH. and ROSSE, B., Mathematical models of a melting snowpack at an index plot, J. of Hydrology, Vol. 32, n°1-2, p. 139-163, 1977.
- O'NEILL, A.D.J. and GRAY D.M., Energy balance and melt theories, Runoff from snow and ice, N.R.C. Associate Committee on Geodesy and Geophysics, Subcommittee on Hydrology, Symposium n°8, Quebec, p. 31-60, 1971.
- SATTERLUND, D.R., An improved equation for estimating long-wave radiation from the atmosphere, Water Res. Res., Vol. 15, n°6, p. 1649-1650, 1979.
- SWINBANK, W.C., Long-wave radiation from clear skies, Q.J.R. Met. Soc, Vol. 89, p. 339-348, 1963.
- YIN CHAO YEN, Recent studies on snow properties, Advances in Hydrosience, Vol. 5, p. 173-214, 1969.

author	standard deviation	correlation coefficient	efficiency coefficient (Nash)
Brunt (1932)	.09	.51	-.63
Swinbank (1963)	.10	.55	-.91
Idso and Jackson (1969)	.09	-.68	-1.28
Brutsaert (1975)	.12	.58	-1.98
Satterlund (1979)	.05	.59	.44
Idso (1981)	.06	.47	.29

Table 1 : Statistical results of the intercomparison between measured and calculated emissivities.

component	energy flux (10^3 J/m^2)
Solar radiation	5443.
Longwave radiation	-10740.
Turbulent mass transfer	8529.
Turbulent heat transfer	13990.
Conduction	1424.
Advection (sensible)	554.

Table 2 : Balance of the flux components over the simulated episod.

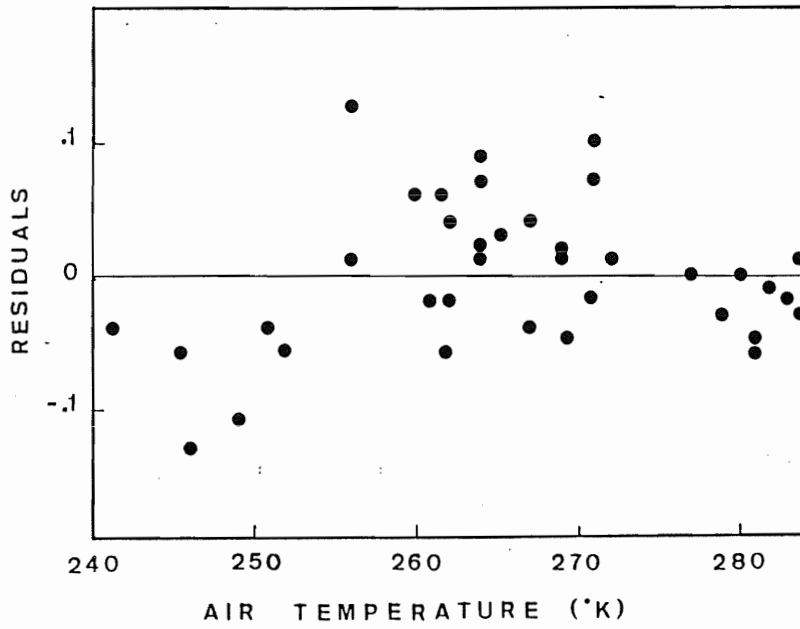


Fig. 1 : Residual analysis of the Idso equation for clear sky emissivities.

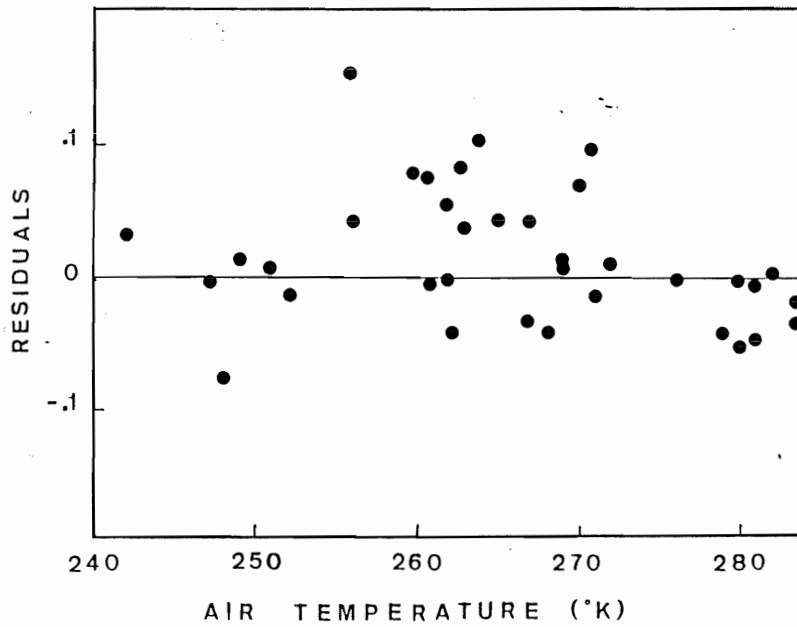


Fig. 2 : Residual analysis of the Satterlund equation for clear sky emissivities.

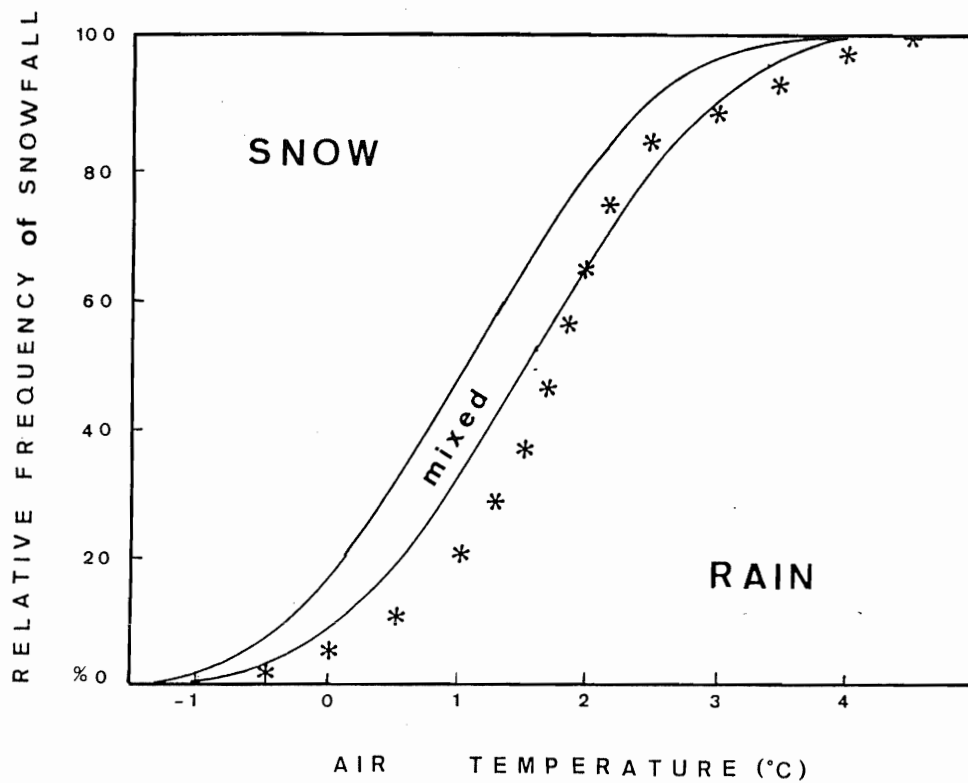


Fig. 3 : Relative frequency of snowfall as a function of temperature
 (- : U.S. Corps of Engineers (Anonymous, 1956);
 * : Ukkel, Belgium).

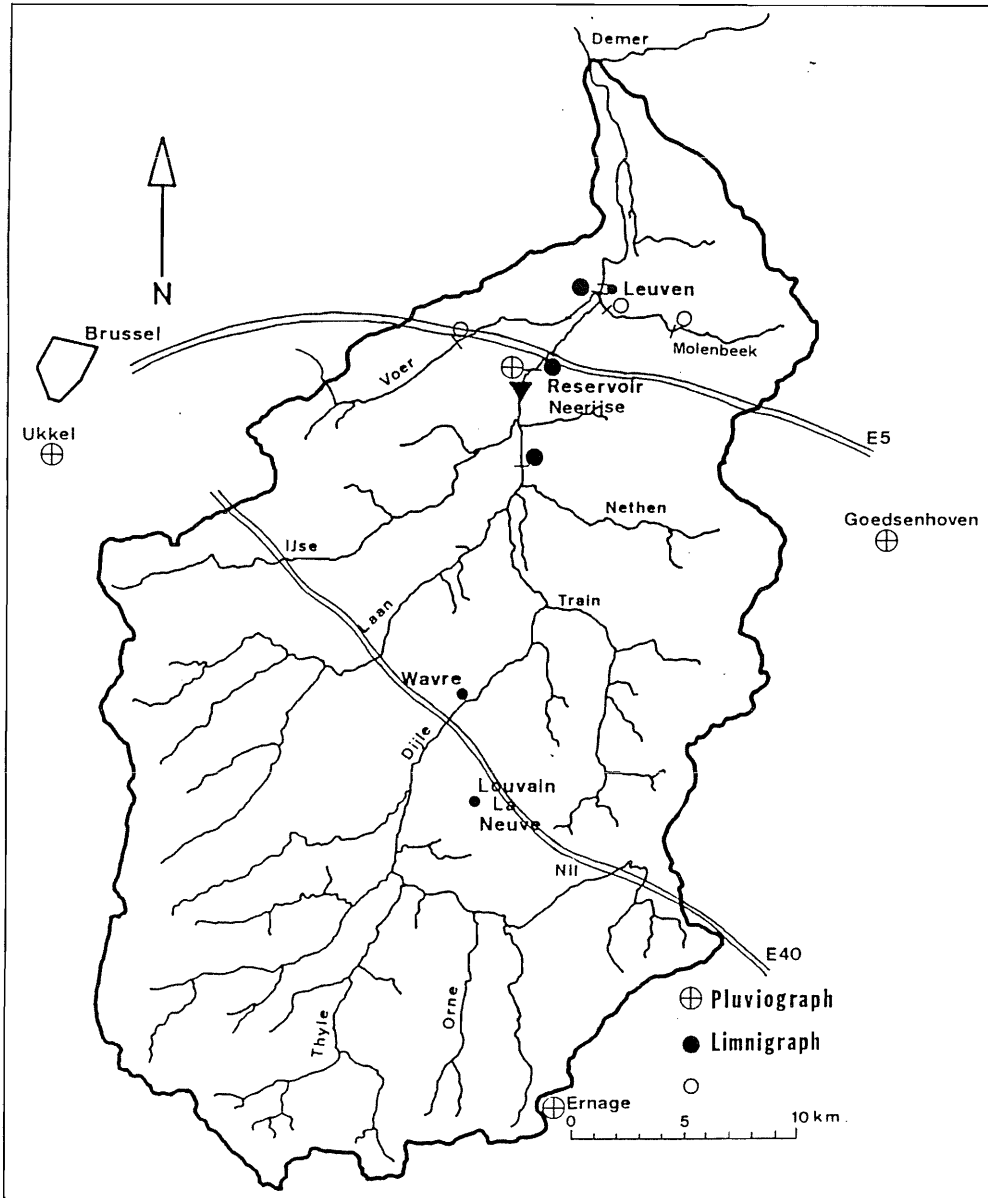


Fig. 4 : The Dijle river basin.

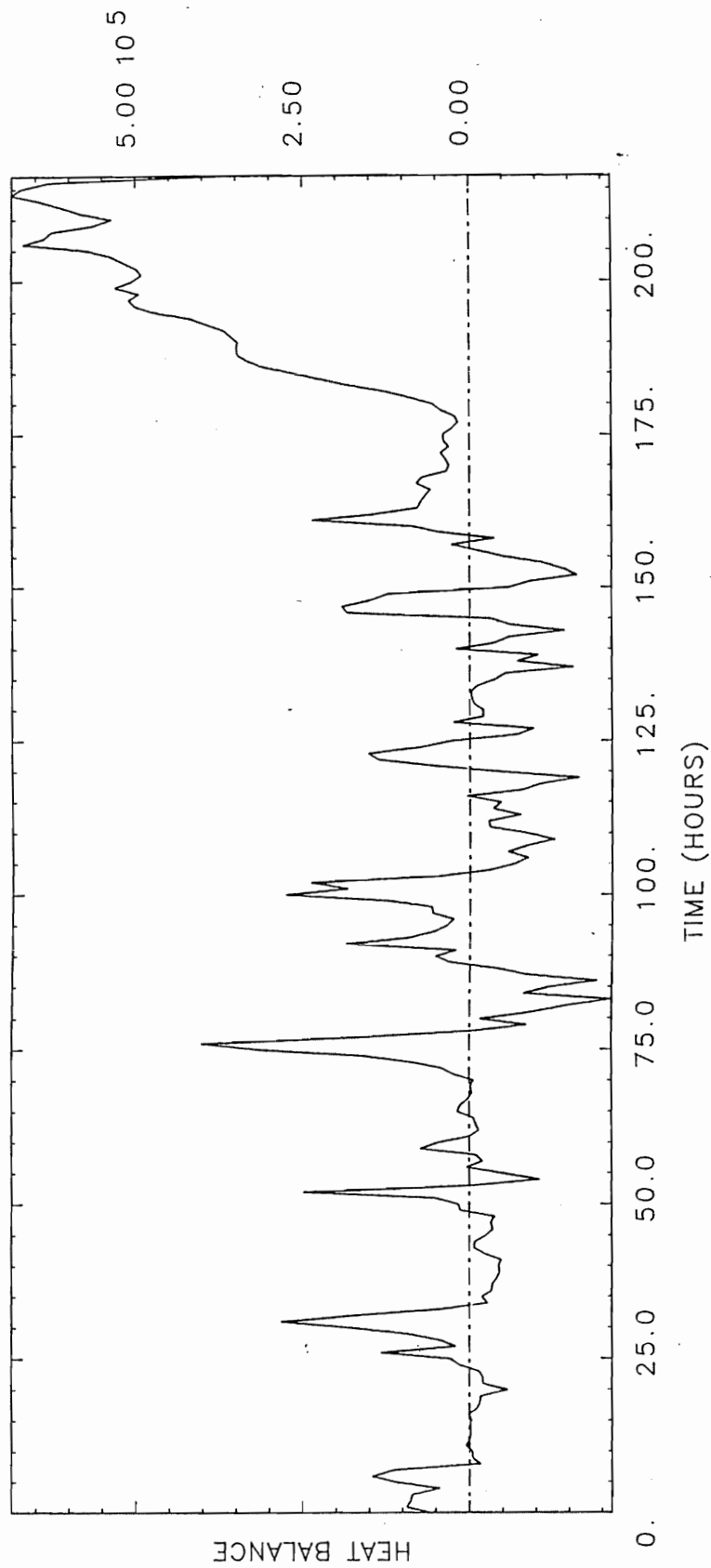


Fig. 5 : The external heat flux balance (J/m^2 hour) over the episod.

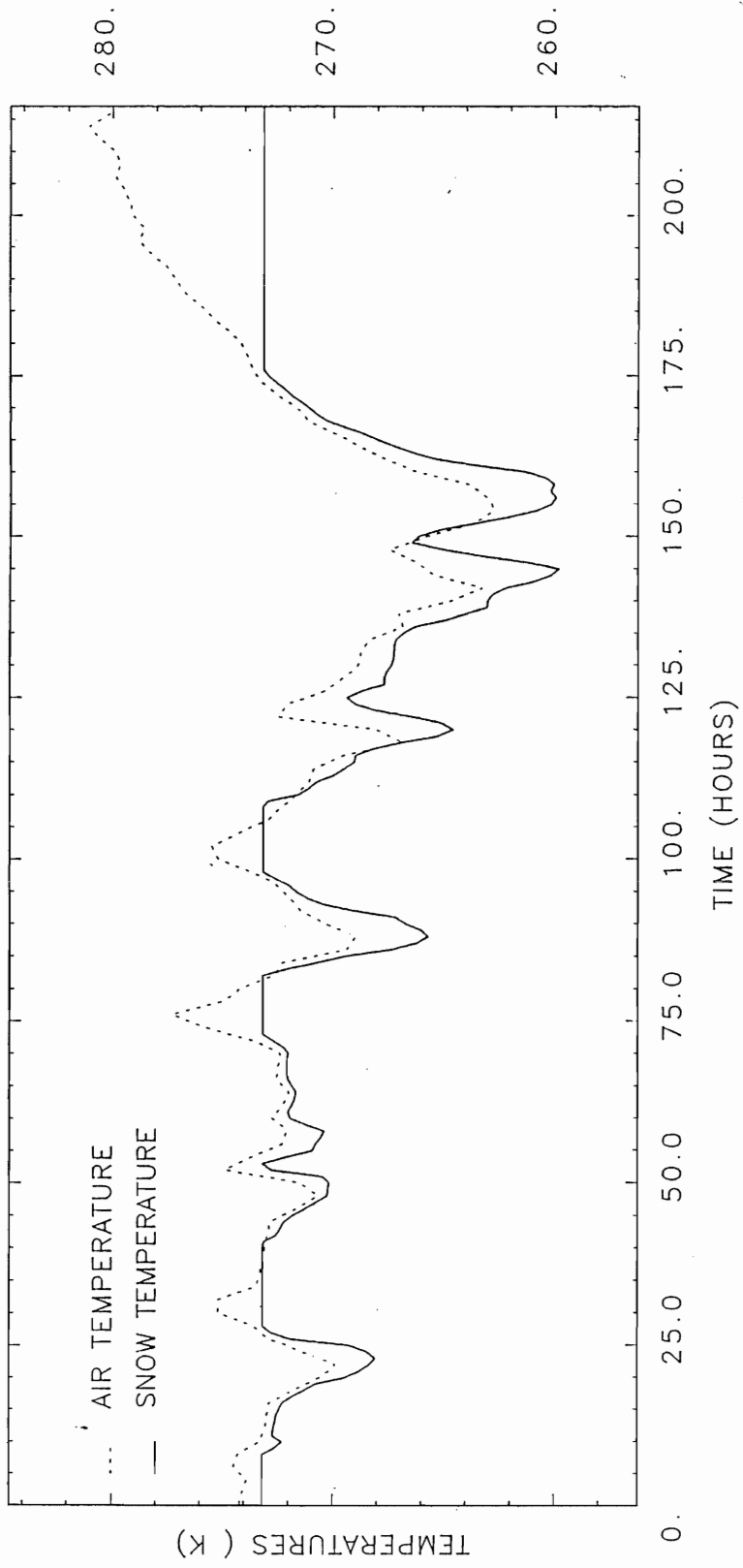


Fig. 6 : The evolution of the temperatures over the episod.

RIVER DIJLE AT ST JORIS WEERT
731127 TO 731207

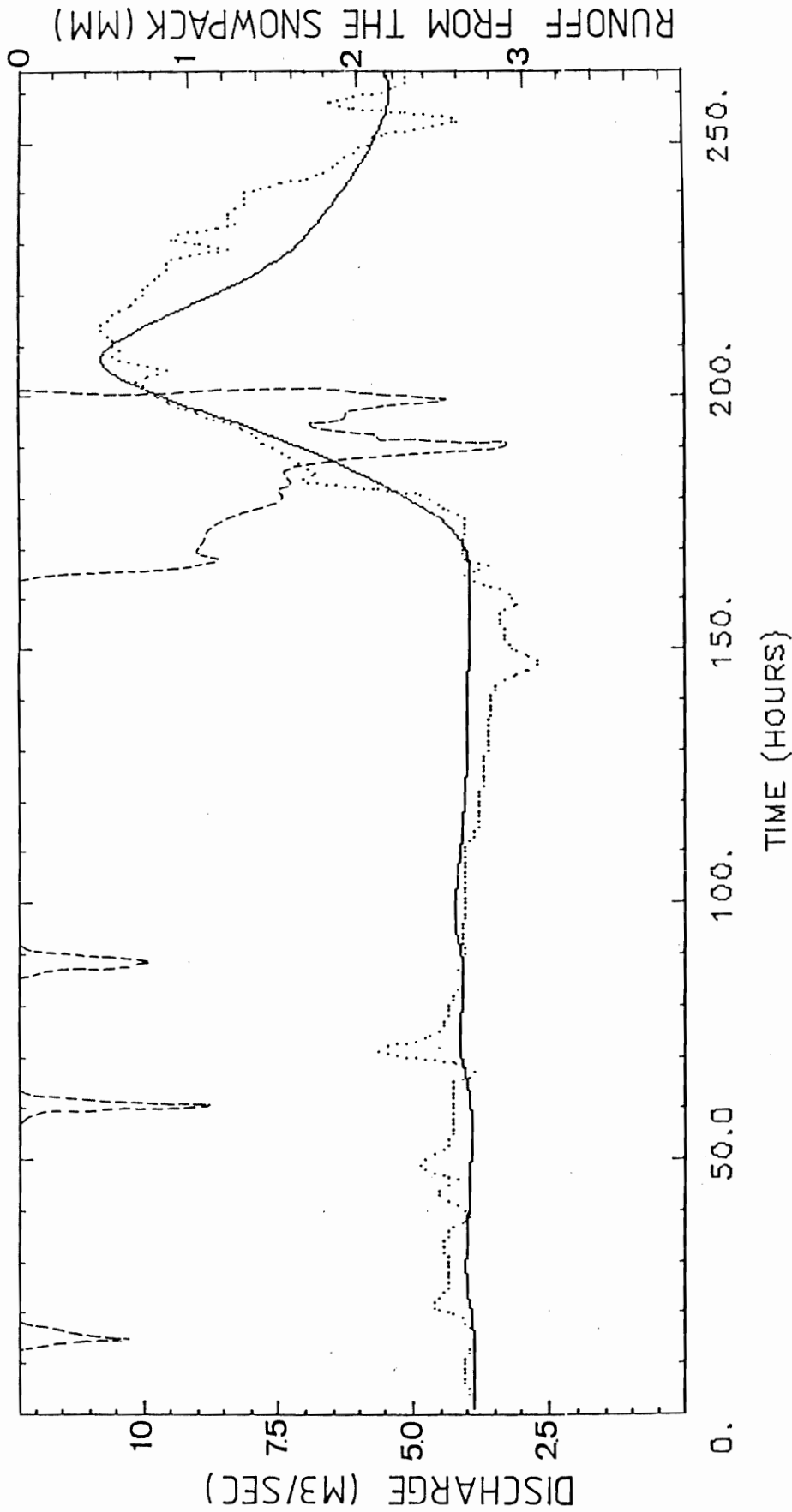


Fig. 7 : The runoff from the pack (---), the simulated (—) and the observed(...) hydrograph.

