Modelling of the snow-water equivalent in the mountain environment

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ABSTRACT The water equivalent (SWE) of the seasonal snow cover can be an important component of the water cycle in mountainous areas, and the knowledge of this temporary storage term may for example be very valuable for predicting seasonal discharge, for making short-range discharge forecasts and also for assessing water quality aspects. Direct measurements of the SWE usually refer to index points and are seldom part of the standard meteorological networks. Therefore it may be advantageous to simulate this storage term based on the available meteorological data. The choice of the appropriate type of model will largely depend on the purpose of the simulation and the data availability. For operational forecasting and prediction practices, various conceptual models describing snow accumulation, melt and internal processes as well as runoff processes in a rather general way are usually sufficient if a minimal data requirement is met. To answer more complex questions such as the hydrological consequences of climatic or land-use changes, conceptual models may serve as an initial step in a "scale analysis" of the most important variables such as precipitation and air temperature or of basin-specific parameters such as forest cover. However, any modelled scenario will always be an artifact of the particular model used, and caution is required when interpreting the results. With the introduction of new measurement systems for meteorological variables with a high temporal resolution new possibilities arise for the application of physically-based approaches in modelling SWE and for the understanding of individual processes in the immediate surroundings of such a station. However, unresolved problems are encountered when trying to apply detailed simulation models on a catchment scale. Possible key areas of further research are improved interpolation techniques for meteorological data, taking into account local and mesoscale climatic conditions, the coupling of meteorological and hydrological models, and the application of remote sensing techniques to validate intermediary results of these spatially distributed models such as snow covered area or, preferably, SWE.

INTRODUCTION: WHY MODEL THE WATER EQUIVALENT OF THE SNOW COVER? A SHORT HISTORICAL PERSPECTIVE

One early attempt to model the snow-water equivalent (SWE) in its own right is given by Hoeck (1952). His primary aim was to correct SWE measurements (gravimetric method) for standard timings such as the 1st and the 15th day of the month. These SWE values themselves were used as a predictor of seasonal discharge (Hoeck, 1951). In a first step, melt due to radiation and heat flow was calculated for a standard elevation of 1600 m a.s.l., the slope and aspect of the location in question, and "normal" meteorological
conditions. In a second step, corrections were applied for elevations, snow temperatures and atmospheric conditions other than "normal". Apart from some foehn periods, acceptable results were obtained for calculating the temporal variation of the SWE during the ablation period between the time of measurement and the disappearance of the snow cover.

The well-known summary report of the U.S. Army Corps of Engineers (1956) treats the SWE at individual points and over watersheds and gives a detailed treatise of the various methods of snow measurement and of the individual processes such as snow accumulation, ablation and runoff generation.

Rockwood (1972) was one of the pioneers who calculated the SWE as a storage term in a precipitation-runoff model. He treated in particular the problem of achieving hydrologically sound yet operationally practical solutions in the runoff modelling technique: for assessments of design floods he suggested the "thermal budget" approach, while for streamflow forecasting, index methods need to be used due to the limitations of data availability. The state of the art at the end of the 1970s in understanding the individual processes such as snow accumulation, distribution, melt and percolation for the purpose of runoff modelling is presented in Colbeck and Ray (1979). An overview as well as a discussion of problem areas of the various methods used in hydrological forecasting involving snow and ice melt is given by Morris (1985) and Lang (1986). Latest developments in assessing snow cover as a water balance component and aspects of hydrological modelling in regions of rugged relief are reported in Lang and Musy (1990).

Martinec (1977) used the calculation of the SWE to assess expected snow loads on building structures based on the generally available snow depth data. Applying the settling curves of individual snow-fall events he achieved good estimates of SWE for the point of maximum snow accumulation of a winter.

For many environmental assessments it is often imperative to know more about other variables besides just the SWE. A summary volume of major contributions relating physical and chemical aspects of the snow cover is given by Jones and Orville-Thomas (1987). In respect to the influence of forest covers on snow accumulation, Calder (1990) explicitly models snow storage in tree canopies. He shows that losses due to interception and evaporation in forested regions have generally been underestimated in the past.

It is the aim of this contribution to show some recent developments in the assessment of the spatial and temporal variations of the snow cover in alpine regions. It tries to show how measured values such as the SWE (point and areal values) or snow covered area can be used as verification criteria to calibrate and validate models of snow accumulation and ablation. The choice of the appropriate model to be used will depend on the purpose of the simulation and the availability of data.

MEASUREMENTS OF THE SWE AS A BASIS FOR VERIFICATION OF MODELLED VALUES

Temporal variability at a point

The most common way to measure the temporal variability of snow cover storage is by means of daily readings of snow depth at a stake. These readings are usually taken on a routine basis at meteorological stations, and therefore constitute a minimal information that is generally available. If snow density measurements are taken periodically (e.g. every 2 weeks by means of the gravimetric method), these snow depths can be converted to water equivalent values. Suggestions for statistical analyses of point snow depth and SWE are given by Risser and Martin (1984).

A recent analysis of some 50 measurement locations with up to 40 years of record is given by Rohrer (1991) for Swiss alpine conditions. Figure 1 shows snow density as a
function of season at three locations of different elevations. It can be shown that there is a steady increase of snow density with progressing season within a band width of about 150 to 220 kg/m$^3$ at high elevations (2535 m a.s.l.), while in lower elevations this trend is less pronounced and the band width is much larger. It is obvious from these temporal variations that snow depth alone is a rather poor measure of total snow cover storage, in particular in the lower regions. For North American conditions, seasonal variations in average snow density of various regions is presented in Gray (1970), however, no indication is given of the year-to-year variation.

![Fig. 1 Measured snow density of the total snow cover at three locations of different elevation (Rohrer, 1991). Data taken between 1945/46 and 1987/88; number of years observed: 38, 39 and 40 (for Weissfluhjoch, 2536 m, Andermatt, 1440 m and Klontal, 855 m a.s.l, respectively). Densities at the lowest station vary greatly due to frequent melting out of the snow cover.]

It is of course desirable to have continuous direct measurements of SWE, e.g. by means of snow pillows or gamma-ray measurement systems. We cannot consider here details referring to the measurement techniques. One can say, however, that after initial problems, successful operation of these systems has been reported (Schälder and Koch, 1981; Kirnbauer and Blöschl, 1990; so-called SNOWTEL sites: Schaefer and Johnson, 1988).

Areal variability

The weight of point measurements greatly increases if their representativeness can be assessed in the surrounding areas. The question can be raised over which minimal area the variability of snow cover storage should be measured to allow the verification of modelled SWE as intermediary result of a rainfall-runoff model. In this context, the Representative Elementary Area (REA) concept as suggested by Wood et al. (1988) may be helpful. If a catchment is treated as being composed of numerous points where the runoff processes form the local water balance fluxes, the REA is the smallest averaging area at each of the points that can be taken as representative of the continuum. The average hydrologic response of such a subcatchment is expected to be invariant or to vary only slowly with increasing subcatchment area. Figure 2 shows an example of mean runoff volume as a function of subcatchment area for five different rainfall realizations as reported by Wood et al. (1988). The authors found the size of REA to be strongly dependent on topography, lying in the range of 2 to 5 ha. Despite the fact that these results refer to the particular simulation exercises of the authors, they can give an indication that the area to be sampled around an index point should be in the order of 200 m by 200 m and not say 10 m by 10 m.

There are numerous contributions in the literature describing the assessment of the areal variability of the SWE. The conventional way is to take a great number of snow depth
measurements along certain profiles and only a few snow density measurements. After several intensive measurement periods the number of points are usually reduced to an essential few that are easily accessible. One recent contribution is given by Golding and Swanson (1986) for the Marmot Creek Watershed (area $A = 9 \text{ km}^2$, elevation range 1600 to 2800 m a.s.l.) in Alberta, Canada, treating the differences between forests and clearings. For the same watershed Bernier (1986) investigated the information content of continuous SWE values measured with snow pillows, individual snow courses (monthly values) and measurements taken once during the season at a great number of grid points. He showed that it is feasible to get winter-long estimates of basin SWE values by a once-a-year intensive sampling effort linked to a once-a-month sampling at a few snow courses, and that snow courses are better estimators of grid data than snow pillows.

Elder et al. (1989) used a digital elevation model of 5 m resolution in the Emerald Lake basin ($A = 1.2 \text{ km}^2$, elevation range 2780 - 3416 m a.s.l.) in the Sierra Nevada, California, to calculate basin SWE based on an extensive snow measurement programme.

Redistribution of snow in the vicinity of ridges is treated by Föhnl and Meister (1983). They were able to simulate snow depth as the superposition of a plume model of drifting snow and a potential flow model of falling snow and concluded that ridge areas as a whole accumulate comparable amounts of snow as flat regions at that elevation, in other words, that redistribution of snow is confined to a relatively small region. The influence of avalanche activity and site conditions such as elevation, slope and aspect on the disappearance of the snow cover is treated by Rychetnik (1987). Snow cover duration may be an important factor when treating the problem of reforestation for example. To assess the variability of snow cover in forested areas, Woo and Steer (1986) suggest Monte Carlo simulations to assist in finding an optimal sampling scheme.

For assessing the areal variability of the snow cover, remote sensing methods have been used widely. A recent overview is given by Gurnell (1990), and a realistic assessment of capabilities and limitations is reported by Rott (1987). As far as mapping snow covered area is concerned, Landsat Thematic Mapper (TM) and SPOT data in the visible and near infrared bands are very valuable due to their high spatial resolution. Their low repetition rate and the presence of cloud cover as observed in many mountainous regions put severe limitations on the use of this type of satellite information as a source of input variables for operational discharge models; as basis for the verification of modelled snow cover, however, they proved to be valuable (Leavesley and Stannard, 1990).
recent contribution describing the use of mapping snow cover patterns based on aerial photographs for the evaluation of various snowmelt models is given by Kirnbauer et al. (1991) and Böschl et al. (1991) for the alpine Längental catchment (A = 9.4 km², elevation range 1900 m to 3050 m a.s.l.), Tyrolian Alps.

The spatial resolution of passive microwave sensor systems is too low to be of great use in rugged terrain. However, active microwave (radar) sensor systems have a spatial resolution in the range of decameters, and mapping of wet snow covers is possible. The long-expected areal determination of the SWE is still not possible.

Bergström and Brandt (1985) are discussing possibilities and limitations of SWE estimates based on measurements of natural gamma radiation along certain flight lines. When used as an input to a conceptual runoff model, SWE estimates of this kind may prove to be valuable in basins larger than about 100 km², if no or only very sparse meteorological data are available. The effect of forest biomass on SWE estimates based on this method is discussed by Carroll and Carroll (1989).

MODELLING OF THE SWE NEAR A METEOROLOGICAL STATION USING DATA OF HIGH TEMPORAL RESOLUTION

Some general remarks referring to physically-based point snow models

Here, some aspects concerning the simulation of the accumulation and ablation of the snow cover near a meteorological station are discussed using operationally available data as input variables. Continuous simulations at a point can be verified easily, provided that measured values of the SWE representative for that location are available. With the introduction of new measurement systems for meteorological variables with a high temporal resolution (minutes to hours) new possibilities arise for the application of physically-based approaches. These detailed models themselves lead to the study of possible hydrological consequences of climatic or land-use changes and are therefore of great topicality. An example of a sensitivity study of such a process-oriented snow model is given by Morris (1982). Rohrer and Lang (1990) present such a model for the use at the new automatic meteorological stations in Switzerland, and an application in the field of agricultural land-use (liquid manure on snow and consequent runoff) is given by Braun (1990).

Snow accumulation process We find a wide recognition of the need to correct systematic precipitation measurement errors. For data of a temporal resolution of 1 month or longer numerous procedures exist (see for example Sevruk, 1989), but with respect to a higher resolution of say 1 day or 1 hour no practical solutions are known. We cannot consider here the strong research effort still needed in this field. Instead, we will treat some aspects of the transition air temperature from solid to liquid precipitation further below which are also linked to the correction problem.

Snowmelt processes When looking at snowmelt at a point, the key areas "energy availability" and "internal processes" need to be discussed. Kuusisto (1986) summarizes the relative importance of the individual energy balance terms over melting snow in different environments. In the mountainous regions considered here, net radiation is the determining component. This balance term represents the relatively small differences between the large energy fluxes consisting of incoming and outgoing shortwave radiation as well as atmospheric (incoming) and terrestrial (outgoing) longwave radiation. It is obvious that each of the individual components needs to be known with a high degree of accuracy before the balance term can be reliable. Major efforts to collect and evaluate direct measurements of the individual components are under way (Ohmura, 1990).

More attention has been given to the albedo of snow and ice surfaces which is very crucial and seldom measured directly. Very few practical solutions for modelling its
temporal variation have been found. A recent parameterization has been suggested by Rohrer (1991). He employs two separate albedo recession parameters for air temperatures above and below 0 °C and a threshold value of 3 mm precipitation over 3 days to recognize "real" snowfall events (with a resulting increase of the albedo to its maximum value).

A review of the turbulent transfer of heat and water vapour over snow and ice is given by Morris (1989). Adequate parameterizations for an individual point do exist where the variables wind speed, air temperature and water vapour pressure are known with sufficient temporal resolution and where local advection can be excluded. A few selected topics on basin-wide applications will follow further below.

**Internal processes** A recent application of a distributed point snowmelt model based on Siemer (1988) at an alpine site at 1930 m a.s.l. is reported by Blöschl (1990). Satisfactory simulations of "cold content" (in terms of water required to warm up the snowpack to 0 °C) and snow cover outflow are achieved, but profiles of liquid water content are poorly simulated for a variety of reasons (e.g. because of not considering ice grain size). For comparison purposes a bulk model as typically used in models of snowmelt runoff was also investigated. The bulk model failed to simulate the temporal variation of cold content and bulk liquid water content; however, the timing of reaching the isothermal state after a cold period was correctly modelled. For practical purposes such as runoff forecasting, therefore, the bulk model approach may suffice if calibration is possible.

**On the problem of the transition air temperature from solid to liquid precipitation**

The correct classification for the aggregational state of precipitation has widely been recognised as being a key task in precipitation-runoff modelling (see for example Braun and Lang, 1986; Obled, 1990). A recent study concerning statistical aspects as well as case studies is given by Rohrer (1989, 1991) for various meteorological stations in Switzerland. At about 70 locations, automatic measurement stations (ANETZ) have been introduced since the late 1970s by the Swiss Meteorological Institute, and interesting differences between the conventional and automatic stations can be observed. At conventional stations the air temperature is measured 3 times a day (term readings) with a mercury thermometer in a Stevenson screen. The air temperature at which the transition from solid to liquid precipitation occurs is spread over quite a wide range (-1 to + 7 °C for example in Arosa, Switzerland, 1847 m a.s.l., as shown in Figure 3) and can be quite different for summer and winter periods (up to 2 °C for Arosa). At automatic weather stations (ANETZ), air temperature is measured at 10 minute intervals using a ventilated thermometer with radiation protection (for details concerning the measurement system see SMA, 1985). Here, the transition between solid and liquid precipitation can be observed within much narrower limits at temperatures about 1 °C lower than at the conventional station as shown in Figure 4 for the station Davos, Switzerland at 1590 m a.s.l. In 1976, the station was moved from Davos-Platz (conventional) to Davos-Dorf (automatic), whereby both locations are situated about 30 m above the main valley floor representing for the most part the same local climate.

Rohrer (1989) also showed that all cases of snowfall at relatively high air temperatures observed at ANETZ stations are of low intensities, and that erroneous classifications in this domain are of only minor consequence (Figure 5a). Rainfall events occurring at air temperatures below 0 °C, however, may be quite intense at low-lying stations such as Zurich-Kloten (436 m a.s.l.) when warm moist air glides up over cold air (temperature inversions) resulting in freezing rain.
FIG. 3: Relative frequencies of mixed precipitation for each 0.5 °C air temperature class shown as hatched columns for winter and summer situations at the conventional station Arosa (1847 m a.s.l.), term readings at 13.00 hours. Figure taken from Rohrer (1989).

FIG. 4: Relative frequencies for snow, rain and mixed precipitation for each 0.5 °C air temperature class at a) Davos-Platz (conventional station) and b) Davos-Dorf (automatic station), years 1978—1987, 1590 m a.s.l., term readings at 07.00 hours. Figure taken from Rohrer (1989).

MODELLING OF THE SWE OVER A BASIN

Factors to consider when choosing the most appropriate snow model

When moving from the plot scale to a basin, one should consider the different levels of abstraction of individual processes that have been found most relevant at the various spatial scales. If one looks at snow and ice melt runoff, a distinction between dominant processes at a point (such as availability of melt energy, percolation of melt water in the
snow cover) and in various size basins (such as snow cover distribution, runoff mechanisms, presence of glaciers) was found useful (Braun and Slaymaker, 1981). Detailed process-oriented models that were found necessary at the plot scale to adequately simulate melt water production were not applicable in larger basins. The reasons were that the necessary meteorological input data were not known, the influence of melt energy availability became less dominant in controlling melt runoff, and the various runoff processes and streamflow routing became more and more important (Braun and Lang, 1986).

Obled (1990) identified some key topics that need to be addressed when applying hydrological models in regions of rugged relief:

(a) the role of the input/output variables (such as precipitation, other meteorological variables, the quality of discharge data),

(b) the role of basin discretization,

(c) the decision which processes, at what level of sophistication, should be explicitly included and applied in the various sub-units (lumped, semi-distributed, distributed models).

A discussion relating to the first topic will follow further below. With respect to the second topic, Blöschl (1990) made a comparison between a grid-based (25 m) distributed snowmelt model, a snow band model and a parametric model (Ferguson, 1986) in the alpine Längental catchment in the Tyrolian Alps. He used an a priori distribution of SWE at the onset of the melt period being considered, applied the same calculation procedures in each model, and compared modelled snow covered area and cumulative mean basin melt. In Figure 6 the results are summarized: if one compares observed and modelled snow covered area, it is striking how well all discretizations perform towards the end of the melt period. If one looks at the variations in time, however, it is obvious that the grid-based distributed model is superior to the other ones. The parametric model yields quite unrealistic snow covered area values at times, but errors compensate with the result that the overall cumulated discharge is quite similar to the one given by the grid-based model. All models, however, tend to overestimate total basin discharge due to an unrealistically assumed SWE distribution at the onset of the period considered. It appears that an equal effort in modelling snow accumulation and redistribution due to wind, etc. is needed if one wants to take full advantage of the grid-based catchment discretization.

With respect to the third topic, namely which processes should be explicitly modelled, the choice will largely depend on the purpose of the SWE simulation. If one is interested in assessing expected snow loads based on snow depth data, for example, Martinec (1977) only needs to consider settling curves of the individual snowfall events. This approach was extended by Rohrer (1991) to get continuous simulations of the SWE based on the standard meteorological measurements of total and new snow depth. This
FIG. 6 Comparisons of different basin discretizations on the basis of percent snow covered area and cumulative mean basin melt, April 24 - June 26, 1989, Längental catchment, Tyrolian Alps. It is striking how well all investigated models perform towards the end of the melt period, while only the distributed grid model yields good results through the whole period considered.

Observed values of snow covered area taken from aerial photographs. Figure taken from Blöschl (1990).

type of model can be used for running plausibility tests of measured values of SWE as shown in Figure 7.

Elder et al. (1989) suggest a more statistically oriented approach to assess a detailed distribution of SWE over a small basin as a storage term. Using a digital terrain model of 5 m grid size in the Emerald Lake watershed of 1.22 km$^2$ in the Sierra Nevada, California, mapping zones based on elevation, slope and radiation were identified and a very large number of SWE measurements were statistically related to these zones. Changes in SWE during the melt season were then simulated successfully through the use of a radiation index, but their approach is unable to effectively model the component of accumulation caused by redistribution of snow. While two-dimensional snow drift
over simple uniform barriers is well understood and easier to model (Schmidt et al., 1984), the problem of understanding this process over three-dimensional terrain remains largely unresolved. However, the question should be raised whether it is meaningful to look at such a high spatial resolution of 5 m as suggested by Elder et al. (1989) for the mentioned purpose of the investigation. In the light of the findings of Wood et al. (1988) as discussed earlier a grid size of say 25 m seems to take better note of the concept of the Representative Area (REA). Furthermore, caution is in order when applying significance tests to "real cases" such as the one presented by Elder et al. (1989) in contrast to "pure statistical" applications (Hornung, 1977).

On the use of conceptual models

The advantage of conceptual models is that they can be readily applied to "real case" basins in various physiographic regions if a minimal data requirement is met, thus constituting a valuable tool in operational hydrology. Tangborn (1988, 1990) gives examples of model applications using only a few carefully selected precipitation stations, two temperature stations lying at greatly differing elevations, and one discharge station which allows calibration over many years. As intermediary results the storage term SWE and snow cover area fraction need to be calculated only over the upper and the lower half of the basin. It is also shown that good discharge simulations can be achieved for the wrong reasons (for example when used in heavily glacierized basins: Braun and Aellen, 1990). Furthermore, no conclusive relationships between parameter values and basin characteristics could be found as demonstrated by Braun and Renner (1991) when applying a rather simple conceptual runoff model in different physiographic regions in Switzerland. These findings show the usefulness and robustness of a simple conceptual model when it can be well calibrated, but also manifest the limitations of its use as a means to assess hydrological consequences of climatic or land-use changes. In the literature we find numerous attempts at this latter end. Pertaining to changes in snowmelt runoff, Rango and Martinec (1987), for example, used a temperature-index method to show the consequences of an air temperature rise of 2 °C on the distribution of discharge through the melt season. Cooley (1990) applied the more detailed approach of the NWSRFS model to show possible hydrological consequences as a result of changes in average daily air temperature of +2, +4 and -2 °C and an increase in daily precipitation of 10 %. Specifically, the consequences for the SWE at two SNOWTEL sites as well as discharge of the Lower Willow Creek basin (A = 190 km², elevation range 1430 to 2400 m a.s.l.) in Montana were evaluated. An example relating to the influence of forest cover
on minimum and flood discharges as well as mean runoff volumes during the summer months in an alpine watershed in central Switzerland is given by Leuppi and Forster (1990). A recent assessment of the influence of forest snowpack management practices on water yield is reported by Ffolliot et al. (1989).

It appears that these approaches using conceptual models can serve as an initial step in a "sensitivity analysis" or "scale analysis" of the most important variables and basin parameters involved. The modelled scenarios, however, will always be artifacts of the particular model used, and caution is in order when interpreting the results. In particular, these scenarios usually do not consider altered seasonal temperature patterns and different distributions and frequencies of precipitation events. In respect to the temporal resolution of the data and the calculation time steps used, it appears that a one-day or shorter time step is imperative when modelling the SWE. In particular, it seems doubtful that mean monthly values of air temperature and precipitation are ever able to represent adequately the individual processes involved (see, for example, Moussavi et al., 1989).

Physically-based approaches: key problems

When attempting to apply a physically-based snow accumulation and ablation model on a catchment scale in mountainous regions as opposed to an individual point, many serious problems need to be addressed. One of them is the spatial and temporal variability of the albedo: at higher elevations of a basin there may be freshly fallen snow with high albedo values, at lower elevations old snow with a lower albedo, on valley floors one might observe rime at the snow surface due to nighttime temperature inversions resulting in a very high albedo, etc. Anderson (1972) shows the domain of snow albedo variability as a function of season as observed in Vermont, and Müller (1984) reviews the contributions relating to the albedo of snow and glacier surfaces in the Alps, addressing in particular diurnal variations and problems associated with measurement techniques. While one is aware of the complexity of the issue, no practical solution to estimate the spatial variation of albedo is at hand.

When considering the turbulent transfer of heat and water vapour, one problem area is related to local-scale advection as typical of late-lying snowfields. Olyphant and Isard (1988) report for the Niwot Ridge (Front Range, Colorado) that near the leading edge of snowfields the turbulent fluxes may exceed net radiation as the major source of energy available for melt. As far as the calculation of these fluxes on the catchment scale is concerned, Morris (1989) advocates improved methods to estimate the wind, temperature and humidity patterns over the whole of the basin. Three-dimensional models of boundary layer flow (as a textbook reference see Pielke, 1984) need to be coupled with distributed hydrological models, which in itself is a critical task due to feedback mechanisms. On one hand, snow cover extent controls the spatial distribution of the meteorological variables, on the other it is predicted by the snow model component of the hydrological model.

It has also been suggested to use upper air data from radiosondes for the calculation of turbulent fluxes over larger areas. The use of this kind of air mass characteristics seems to be promising in the prairie environment (Male and Granger, 1981) and in maritime regions (Moore and Owens, 1984) where the air masses tend to be in a well-established equilibrium with the underlying surface. More critical is the situation in an area such as the Alps where the air mass characteristics are highly variable depending on the large-scale synoptic situation. Lang and Rohrer (1987) were able to show how maximum snowfall events are related to air flow conditions and humidities at the 850 and 700 hPa levels, and similarly, highly variable conditions are to be expected under snowmelt situations.

In the field of interpolation techniques, Bénichou (1986) investigated the influence of the local relief on monthly totals of precipitation in the Massif Central, France. If
physically-based snow models are to be used on a catchment scale, big efforts need to be undertaken to apply this kind of interpolation methods for data of high temporal resolution. Jensen (1989) employed intrinsic random functions in three-dimensional kriging to interpolate hourly values of precipitation and air temperature. On the large catchment scale this approach may yield an adequate representation of the spatial and temporal variability of meteorological data, which are very useful for river flow forecasting purposes (Lang et al., 1987). For the application of physically-based models, however, influences of the local relief which alter the general patterns of interpolated fields may need to be considered among other aspects.

CONCLUSIONS

1.) Modelling of the SWE is a valuable tool for assessing the snow storage term in mountainous basins, and appropriate methods exist for the use of hydrological prediction and forecasting purposes.

2.) Measurements of the SWE are valuable for the verification of models of snow accumulation, redistribution and melt. With the given measurement techniques, it is only reasonable, in terms of cost and effort, to assess point values on a continuous basis, or values along selected lines (snow courses) at infrequent intervals (e.g. monthly or at the time of maximum accumulation, etc.). These measured SWE values will always retain their character as index values.

3.) In this context, the primary value of remotely sensed data today is as a means to test modelled extent of snow cover.

4.) For the study of possible hydrological consequences of climatic or land-use changes detailed physically-based approaches are needed. Although these can only be applied and verified at individual points at present, they still aid in understanding individual processes. Future "laboratory-type" applications of these physically sound models over "hypothetical" catchments are requested and will give answers regarding an integrated response.

5.) To answer SWE-related problems over "real" montainous basins in view of climatic or land-use changes, snow models as part of conceptual precipitation-discharge models may serve as a means of "scale analysis" of the influence of selected variables such as total amounts of precipitation and mean air temperature, or of selected basin-specific parameters such as forest cover.

6.) The attempt to apply physically-based models of SWE and runoff on the catchment scale in mountainous regions will necessitate an enormous research effort in the field of interpolation of meteorological data of high temporal resolution taking into account e.g. the influence of the local topography, of the regionalization of selected model parameters such as albedo and roughness length, and the coupling of meteorological mesoscale models with highly distributed hydrological models. For the validation of intermediary results such as snow covered area or preferably SWE, remote sensing techniques may be helpful.

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