Published in Theoretical and Applied Climatology 76: 125-140, 2003

Estimates of snow accumulation and volume in the Swiss Alps under changing climatic conditions

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Abstract

Snow is a key feature of mountain environments in terms of the controls it exerts on hydrology, vegetation, and in terms of its economic significance (e.g., for the ski industry). Its quantification in a changing climate is thus important for various environmental and economic impact assessments. Based on observational analysis, surface energy balance modeling, and the latest data from high-resolution regional climate models, this paper investigates the possible changes in snow volume and seasonality in the Swiss Alps. An average warming of 4°C as projected for the period 2071-2100 with respect to current climate suggests that snow volume in the Alps may respond by reductions of at least 90% at altitudes close to 1000 m, by 50% at 2000 m, and 35% at 3000 m. In addition, the duration of snow cover is sharply reduced in the warmer climate, with a termination of the season 50-60 days earlier at high elevations above 2000-2500 m and 110-130 days earlier at medium elevation sites close to the 1000 m altitude. The shortening of the snow season concerns more the end (spring) rather than the beginning (autumn), so that it should be expected that snow melt will intervene much earlier in the season than under current conditions. The results of this study are of relevance to the estimations of the impacts that the projected warming may have on the amount and timing of water in hydrological basins, on the start of the vegetation season, and on the financial status of many mountain resorts.

1. Introduction

The importance of snow in terms of environmental (e.g., hydrology, vegetation) and economic systems (e.g., tourism, water management) has been stressed in numerous studies (Beniston, 2000; Dettinger et al., 2003; Haeberli and Beniston, 1998; etc.). Furthermore, Beniston et al. (2003) have shown that current numerical models have difficulty in representing the spatial and temporal variability of snow, even at the fine resolutions of certain regional climate models (20-50 km). As a result, the study and prediction of an important component of the natural environment is constrained by the limits imposed by the spatial resolution of current climate models. Observational data can help overcome such problems, by linking changes in snow pack behavior that are systematically related to specific shifts in climatic conditions, e.g., mild and moist winters, cold and dry winters, etc. Satellite remote sensing provides an estimate of the areal cover of snow and its changes over time, as shown for example by Wunderle et al. (2002), Ranzi and Grossi (1999), amongst others. However, satellite observations at the resolution required for studies of snow span only the last 20 years or so. There have been a number of studies that relate snow to climate more at the hemispheric scales (e.g., Frei and Robinson, 1999), or to large areas of North America or Russia (e.g., Brown, 1998; Clark et al., 1999; Hughes and Robinson, 1996; Ye, 2000), than to mountain-specific conditions. Some studies have addressed issues of snow pack at the regional scales (e.g., Beniston, 1997; Cayan, 1996; Hantel et al., 2000; Harrison, 1993; Whetton et al., 1996), but these investigations have focused more upon the duration rather than upon the seasonal accumulation of snow during the winter season.

A quantification of the amount of snow in the mountains and the changes that occur with shifts in climate is, however, crucial for assessing the amount of water that will ultimately runoff and be routed into the numerous river systems originating in the Alps in the spring and early summer. The Alps in general, and Switzerland in particular, have in the past been referred to as “the water tower of Europe” (Mountain Agenda, 1998). Any substantial changes in the mountain snow pack would have a significant impact on the flow of many major river

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basins, not only because of changes in the amount and timing of runoff, but also because of the potential for enhanced flooding, erosion, and associated natural hazards. Such issues will in time require new strategies in water resource management as recommended *inter alia* by the IPCC (2001), particularly in a country like Switzerland that relies on hydro-power for over 60% of its energy supply. The timing of snow melt is also a major determinant for initiating the vegetation cycle of many alpine plant species, and hence its quantification is necessary when assessing the response of vegetation to climatic change (Keller and Körner, 2003; Myneni et al., 1997; Prock and Körner, 1996).

This paper will thus address the issue of snow amount under various climatic conditions as experienced in Switzerland during the 20th century, and simulations of future snow seasons at the local level, using a physically-based surface energy balance model. Approaches aimed at overcoming these problems include the interpretation of observational data, as discussed by Beniston et al. (2003), and the use of a detailed surface energy balance model that can be applied to the study of snow under current or future climatic conditions, as shown by Keller and Goyette (2003). Estimates of possible future snow amounts are based on the latest assessments of climatic change for the latter part of the 21st century (Christensen et al., 2002).

2. Snow amount in the Alps for different winter types

2.1 Data sets and study sites

As in the Beniston et al. (2003) study, hereafter referred to as BKG03, the identical set of 18 climatological sites has been used in this study, that are distributed over the Swiss territory both horizontally and with height. The altitudes of the observation stations range from 317 m in Basel to 2500 m at Säntis; at all these locations, with daily data available in digital form since at least 1931, and for certain reference sites such as Säntis, Neuchâtel, or Zürich, digital data is available from 1901. The temporal extent of the series allows useful statistical interpretations to be made. The reader is referred to Figure 1 in BKG03 for further details.

In order to concentrate on the behavior of the snow pack in the main part of the Swiss alpine domain and in the lowlands to the north, the present study has not taken into consideration data from the observational sites located on the southern slopes of the Swiss Alps. The climatic regimes in this part of the Alpine chain are dominated by Mediterranean influences, with significantly different, and often asynchronous, snow precipitation characteristics from those in the Alps and to the north.

2.2 The surface energy balance model GRENBLS

The physical processes governing the state of a snow pack and its evolution are well described using equations that describe the exchange of energy and water between the land surface and the atmosphere (Coughlan and Running, 1997; Gustafsson et al. 2001). The energy aspects associated with snow melt processes require an accurate treatment of the thermodynamics of the snow pack. Such processes, expressed in terms of water and energy budgets are computed in a physically-based surface energy balance model (SEBM) called GRENBLS (GRound ENergy Balance for naturaL Surfaces). A description of the model equations, and their application to alpine snow cover was first presented by Keller and Goyette (2003), with examples of simulations at specific alpine locations.

The GRENBLS surface energy balance model used for part of this investigation resembles the formulation of land-surface schemes of intermediate complexity that are used in 3-D numerical climate models. The model is a physically based model that explicitly computes surface variables but requires appropriate atmospheric input variables. It is thus driven by hourly input data of air temperature, T_{air} (K), dew point temperature, T_d (K), anemometer-level wind velocity, V_a (m s^{-1}), precipitation, P (mm s^{-1}), and surface pressure, p_{sfc} (hPa). The model computes the radiative fluxes (incoming solar radiation, $K\downarrow$), the surface turbulent sensible and latent fluxes, as well as the heat flux in the ground and in the snow pack. Surface temperature, soil moisture and snow mass are computed as prognostic variables in the model.

The surface radiation budget, Q^*, may be expressed following the notation of Oke (1987) as:

$$Q^* = (1-\alpha)K\downarrow + L\downarrow - (1-\varepsilon)L\downarrow - \varepsilon\sigma T_{sfc}^4 \qquad (1)$$

where Q^* (W m^{-2}) is the all-wave radiation budget, $K^* = (1-\alpha)K\downarrow$ is the short-wave radiation budget. $L\downarrow$ is the incoming long-wave radiation, $\varepsilon\sigma T_{sfc}^4$ is the outgoing long-wave radiation from the surface, α is the surface albedo, ε is the surface emissivity, and σ the Stefan-Boltzmann constant. The last two terms in Equation 1 represent the long-wave radiation budget (L^*). T_{sfc} is the computed snow temperature, T_{snow}, if the snow covers entirely the surface, and the top ground temperature, T_g, if there is no snow. Otherwise, the snow pack and the ground each have their own temperature. The snow fraction at the surface, δ_{snow}, is parameterised according to Roesch et al. (2001):

$$\delta_{snow} = \min\left[1, \tanh\left(\frac{100 M_s}{\rho_{snow}}\right)\right] \qquad (2)$$

where M_s represents the snow mass in kg/m^2 and ρ_{snow} the snow density in kg/m^3. Its value is set to 1 when M_s exceeds 10 kg/m^2 and to 0 when snow is absent.

The surface albedo, α, modulated by the snow fraction, is calculated separately for two different wavelength bands, namely the visible ($0.3 < \lambda < 0.68$ µm) and near infrared ($0.68 < \lambda < 4.0$ µm) bands through the following relationships:

$$\alpha = [\alpha_s, \alpha_{nir}] = \delta_{snow}[\alpha_{s,snow}, \alpha_{nir,snow}] + (1-\delta_{snow})[\alpha_{s,sfc}, \alpha_{nir,sfc}] \qquad (3)$$

where α_{sfc} =[$\alpha_{s,sfc}$, $\alpha_{nir,sfc}$] is the surface albedo and α_{snow} =[$\alpha_{s,snow}$, $\alpha_{nir,snow}$] is the albedo of snow. Its value is further modulated by snow ageing, by the presence of vegetation (McFarlane, 1992), and by solar radiation penetrating into the snow pack.

Snow is modelled as an evolving one-layer pack characterised by a temperature, T_{snow} (K), a mass, M_s, and a density, ρ_{snow}. The surface energy budget is computed as follows:

$$Q_{sfc} = Q^* - (Q_h + Q_e + Q_{snow} + K\downarrow_{snow}) - L_f(M_F/\partial t + M_s/\partial t) \qquad (4)$$

where Q_{sfc} is the heat storage term, Q_h the sensible heat flux, Q_e the latent heat flux, Q_{snow} the heat flux through the snow pack, and $K\downarrow_{snow}$ the solar flux penetrating the snow pack. The last term in the equation is the energy change associated with the melting rates of frozen soil moisture ($M_F/\partial t$) and snow ($M_s/\partial t$), where L_f is the latent heat of fusion. The temperature of the snow pack is computed in a deterministic manner through the heat storage using a force-restore method (e. g., Stull, 1988; McFarlane et al., 1992) as:

$$Q_{sfc} = \frac{C_*}{2}\left[\frac{\partial T_{sfc}}{\partial t} + \omega(T_{sfc} - T_o)\right] \qquad (5)$$

where T_o is an historical temperature taken as a 24-hour moving average of T_g, ω the diurnal frequency, and C_* the effective heat capacity of the surface. The effective heat capacity of snow, $C_{*,snow}$, is fixed at 9.6×10^4 J/m^2/k.

The surface sensible, latent and snow heat fluxes are computed as follows:

$$Q_h = \rho_{air}\, c_p\, C_h\, V_a\, (T_{sfc} - T_{air}) \qquad (6)$$

$$Q_e = LE = L\rho_{air} C_e V_a \beta [q_{sat,sfc}(T_{sfc}) - q_{air}] \qquad (7)$$

$$Q_{snow} = -\lambda_{snow} \frac{\Delta T_s}{\Delta z} \tag{8}$$

where $\Delta T_s = T_{snow} - T_g$, $\lambda_{snow} = 2.576 \times 10^{-6} \rho^2_{snow}$ represents the heat conductivity of snow (W/m/K) and $\Delta z = M_s \, \rho^{-1}_{snow}$ is the snow depth (m), where a gain of energy by the snow pack is represented by a negative sign of Q_e and Q_h, whereas the opposite is true for the other fluxes. The bulk heat and moisture transfer coefficients, C_h and C_e respectively, are parameterised according to Benoit (1977) on the basis of the Monin-Obukov similarity theory. ρ_{air} is the air density, and β is an evapotranspiration factor, which turns 1 if the whole surface is snow-covered (McFarlane et al., 1992). The saturation specific humidity at the surface is $q_{sat,sfc}$ and q_{air} represents the specific humidity at the screen level in kg/kg. L is the latent heat of vaporisation or sublimation (J/kg) according to the phase of water considered. c_p is the specific heat of air at constant pressure (J/kg/K).

The water budget at the surface is calculated as follows:

$$\begin{aligned} \frac{\partial W}{\partial t} &= P_L - E_g + M_s - R_{off} \\ \frac{\partial M_s}{\partial t} &= P_s - E_s - M_s \end{aligned} \tag{9}$$

where W is the total soil moisture (liquid and frozen) in kg m^{-2}, P_L and P_S are the liquid and solid precipitation rates, M_s the melting rate of snow, E_g the ground evaporation, and E_s the snow evaporation rate where the total evaporation rate E equals E_g+E_s. R_{off} is the total runoff, *i. e.,* surface saturation and bottom drainage following ideas of Noilhan and Planton (1989).

The partitioning of W into E_L and E_F depends on the T_{air} and is estimated as follows:

$$\begin{aligned} &E_L = W, E_F = 0 \quad ; T_{air} > T_f \\ &E_L = 0, E_F = E_W \ \ ; T_{air} \leq T_f \end{aligned} \tag{10}$$

where E_L is the liquid and E_F the solid part of the soil moisture. T_f is the freezing point.

Precipitation is considered as solid if T_{air} is less than that of the triple point of water. Liquid precipitation falling on the snow pack induces snowmelt. Melted snow ($M_s / \partial t < 0$ in Equation 4) enters directly into the soil as liquid moisture, latent heat is released and energy is transferred to the surface.

GRENBLS has shown genuine skill in reproducing the observed snow pack characteristics in the Swiss Alps, such as depth and duration. GRENBLS is a semi-prognostic numerical model driven by hourly input data of screen-level (2 m) air and dew-point temperatures, anemometer-level wind velocity, liquid-water equivalent precipitation flux, and surface pressure. The model computes the solar and infrared radiative fluxes, although incoming solar radiation may also be prescribed from observations, the turbulent sensible and latent heat fluxes, as well as the heat flux within the snow pack and its underlying surface. The precipitation determines the amount of snow that falls whenever the air temperature is below freezing. Snow is modeled as an evolving one-layer pack characterized by a particular temperature, mass, and density. The snow temperature is computed in a prognostic manner from the energy balance of the pack. Total runoff is the result of bottom drainage and soil saturation. The former is generated when the soil moisture exceeds the field capacity and the later is related to the level of soil moisture saturation. The melting snow directly enters the underlying surface and thus contributes to the soil moisture budgets.

2.3 Response of snow to winter type

BKG03 have shown that, according to the type of winter, the response of snow in terms of the duration of the season varies considerably with altitude. Wunderle et al. (2002) have quantified, on the basis of satellite imagery analyses, that the elevation of the snow-line below 2500-3000 m asl varies substantially according to the type of winter. Martin and Durand (1998) also ascertained through modeling techniques that changes in snow duration in the French Alps are very sensitive to temperature at low to medium elevations (below 2000 m asl), whereas at much higher elevations (above 3000 m, for example), there is almost no change under warmer conditions. This is basically due to the fact that at high altitudes, warmer conditions will not raise temperatures beyond the freezing point, so that the controlling factor on snow amount and duration is not temperature but rather the quantities of precipitation that occur during the winter season.

In order to emphasize this point, Figures 1a and 1b illustrate the observed behavior of a modest snow pack in the Swiss Alps for a low-elevation station (Château d'Oex, at 981 m asl) and a more extensive snow pack at a higher-elevation station (Arosa, at 1847 m asl). Although lower elevations than Château d'Oex could have been selected, it is interesting here to analyze the behavior of snow that has a truly seasonal, continuous character rather than an episodic character that is typically found at sites such as Bern (570 m) or Basel (317 m). At these low levels, snow will appear or disappear intermittently according to cold or warm cycles associated with frontal systems affecting the Alps during the winter.

The present study focuses on four climatic winter modes, namely cold/dry, cold/moist, warm/dry, and warm/moist. The selection of these modes is based on a combination of winters that are below the lower quartile (i.e., 25-percentile) or above the upper quartile (i.e., 75-percentile) of mean winter precipitation and temperature. The use of quartiles rather than finer percentiles allows to capture the four basic modes more readily. Here, winter refers to the mean of the three months of December, January, and February (DJF); it was shown in BKG03 that the DJF average minimum temperatures and precipitation provide a reasonable measure of the seasonal snow behavior, and are well correlated with duration and accumulation of snow even if the snow season at high elevations lasts longer than just the three months of winter. All curves have been smoothed to remove the noisiness of day-to-day variability in snow cover. Total (also referred to as seasonal) snow accumulation, as discussed below, is defined as the integral of daily snowfall amounts during the winter; this estimate does not take into consideration morphological transformations of the snow resulting from ambient temperature and moisture conditions during the season. The curves provided here are thus somewhat different from those that would be obtained from simply integrating daily snow accumulations, but nevertheless provide a basis for the subsequent discussion.

Insert Figure 1a and Figure 1b here

It is seen that the greatest snow abundance occurs for cold/moist winters at both sites, while the least abundance is associated with warm winters in Arosa, as could be expected intuitively. Warm and moist winters at Château d'Oex are frequently associated with DJF precipitation in the form of rain, so that the extent of the winter snow pack is rapidly reduced through surface runoff. At higher elevations, on the other hand, the warm-moist winter mode leads to greater total snow accumulation than the warm-dry mode, but does not offset the negative influence of the higher temperatures compared to cold winters. It can also be seen that the duration of the snow season is strongly modulated by the various combinations of temperature and precipitation, the longest being related to the cold-moist winter mode, the shortest to the warm modes (dry at high elevations and moist at low elevations). Although there is much variability in the behavior of the snow pack, in the sense that similar winters modes may exhibit different responses in the shape of the snow total snow accumulation curves, the general pattern discussed here is nevertheless confirmed for most stations.

In a climate that is changing as a result of enhanced greenhouse gas concentrations, the separation into different winter modes can be used as an indication of the type of snow cover that could be expected in the future, by analogy with current conditions. Table 1 lists typical statistics of DJF mean temperature and precipitation associated with each mode for the two stations illustrated in Figure 1.

Insert Table 1 here

A rise in temperature of over 5°C at Arosa and a 40% reduction in precipitation, i.e., a switch from the cold/moist to the warm/dry modes, leads to a reduction of 54 days in snow cover and almost a 50% reduction in the seasonal snow accumulation. For an equivalent rise in temperature and reduction in precipitation at the lower site of Château d'Oex, there is a reduction of over 100 days of snow cover, from 125 days to 22 days, and an 80% reduction in snow cover. As can be expected intuitively, the characteristics of the snow pack are substantially more sensitive to changing temperature and precipitation patterns at low elevations as compared to those at higher elevations. Indeed, BKG03 have shown that at the elevations of Säntis in eastern Switerland (2500 m) and above, snow duration was longer and snow amounts larger in the warmer 1990s than in the cooler 1960s. However, that particular paper also showed that beyond a certain level of temperature rise, enhanced precipitation cannot compensate for increased warming, in terms of snow amount or duration.

Insert Figure 2 here

Because each winter has its own "signature", a large dispersion of points is seen when attempting to directly relate DJF mean minimum temperature with seasonal snow accumulation, as illustrated in Figure 2 for all 18 sites. The two sets of points, and their associated linear regression lines differentiate between drier-than-average and moister-than-average winters. Although the correlation coefficients are high and significant (r=0.85), the clustering of points in the lower right segment of the graph largely contributes to this high correlation. Indeed, when considering only the data points that are associated with temperatures below 0°C and seasonal accumulations beyond 200 cm per season, the correlation drops to r=0.72 for both cases. One reason for the high dispersion of points is related to the fact that, for a same average winter temperature, the timing of the start or the end of the snow season may differ widely. Cold temperatures that occur early in the season are sometimes associated with an extension of continental high-pressure systems, that bias the overall winter mean temperature while at the same time blocking or diverting precipitation belts away from the alpine region. Cold temperatures that are encountered after the snow has begun to accumulate in significant amounts may lead to the same DJF mean temperature as in the preceding example, but much more consequent snow amounts by the end of the season. There is thus a strong link between duration and total snow amount, as will be seen later.

A first-order analysis could allow some measure of assessment of the response of total snow accumulation to changes in temperature, including the biases imposed in terms of precipitation amounts. For example, Figure 2 shows that for moister-than-average winters, a 4° rise in DJF mean minimum temperature from –4°C to 0°C reduces on average snow amount from 670 cm to 290 cm, i.e. roughly a 55% reduction. The reduction in percent falls to 30% when warming from –10°C to –6°C, even though the absolute amounts of total snow accumulation are the same. However, it is seen that because of the large spread of points, seasonal snow accumulation for a DJF mean minimum temperature of –4°C ranges from less than 300 cm to over 900 cm under current climatic conditions. When DJF mean minimum temperatures are around the freezing point, accumulation ranges currently from close to nil to about 400 cm.

Insert Figure 3 here

Figure 2 is thus too inaccurate to be applied to the problem of quantification of snow directly on the basis of temperature alone. However, a more accurate estimate for seasonal snow accumulation can be obtained when linking this parameter to snow duration. Figure 3 has been prepared using data from all available stations and is thus an aggregate of altitude and type of winter; because of the very strong correlation that exists between height and average snow duration and height and total snow accumulation, the curves illustrated here can be considered to be representative of the alpine domain of Switzerland. Figure 3 highlights the fact that the dispersion of points around the mean is much lower than when relating snow accumulation to temperature as in Figure 2. As in the previous illustration, the data for wetter-than-average and drier-than-average winters have been disaggregated, thus allowing the diagram to be interpreted also in terms of the sensitivity to precipitation. The correlation coefficient for the regression lines is high, on the order of 0.96 for both curves. The non-linear nature of the best-fit curves reflects the fact that there is a non-linear increase in seasonal snow accumulation for a linear increase in snow duration. For example, in the case of wetter-than-average winters, an increase in snow duration by 60 days, from 30 to 90 days is accompanied on average by a rise in snow accumulation of 225 cm, from 50 – 275 cm. The same increase of 60 days, but from 240 to 300 days, results in an average additional accumulation of 315 cm, from 960 – 1275 cm. The duration/accumulation relationship can also be viewed in terms of altitude, since the level at which snow falls in the mountains is the major determining factor for the subsequent behavior of the snow pack.

This type of analysis enables the possibility of assessing the average quantities of snow that accumulate at different altitudes in the Alps as a function of shifting climatic conditions. While the following is based on average quantities of snow at various elevations in the mountains, and does not enter into any particular detail on local site characteristics (south- or north-facing slopes, for example), the results yield interesting estimates that are not far removed from observed snow amounts.

In order to compute snow volume, which is the most important snow-related variable when investigating peak flow from snow-melt in river systems originating in the Alps, it is necessary to know the surface area of Switzerland that is associated with a particular height class. Dividing Swiss topography into 44 intervals of 100 m each, it is seen that 55% of the country is located above the 1000 m level, 24% above 2000 m, and 2.5% above 3000 m. Using the current values of average snow-cover duration, which as shown in BKG03 are a linear function of altitude, it is thus possible to assign the average volume of seasonal snow accumulation for each 100-m interval, using the information provided in Figure 3 for both dry and moist winter modes. This yields the bold curves of Figure 4a and 4b, respectively. The shape of the curve reaches a maximum at about 2000 m asl, and represents the combination of snow depth and surface area that yields the greatest amount of snow. Below that level, even though the surface area considered is much greater, the amount of snow that accumulates during the season is much less, and thus the snow pack volume is reduced. Conversely, above 2000 m, even if the total snow accumulation is much greater, the volume is lower because deep snow accumulates on increasingly smaller surface areas. The other curves in Figures 4a and 4b will be discussed in more detail in a subsequent paragraph.

Insert Figure 4a and 4b here

2.4 Snow characteristics in a warmer climate

With the available empirical approaches that have been developed here, it is possible to estimate the mean changes in snow volume that may intervene with a mean rise in temperature. As was seen in BKG03, changes in snow duration are associated with combinations of shifts in temperature and precipitation (op. cit., Figure 10). Preliminary findings from the analysis of the HIRHAM4 Regional Climate Model (RCM; Christensen et al., 1998) simulation results used in the context of the EU 5th Framework Program "PRUDENCE" project (Christensen et al., 2002) suggest that mean minimum winter temperatures will rise in response to greenhouse forcing by up to 4°C and more in the Swiss Alps by the end of the 21st century. These simulations are based on the global greenhouse-gas emission scenario A2 established by the IPCC (Nakićenović et al., 2000), which lies within the upper bounds of the IPCC emission scenarios that would raise global atmospheric greenhouse-gas concentrations to about 800 ppmv by the end of the 21st century, i.e., a three-fold increase over their pre-industrial values. The driving General Circulation Model (GCM) for these simulations is the UK Hadley Centre's HADCM3 coupled ocean-atmosphere model.

The shifts in temperature and precipitation for Switzerland, averaged over the decades 2071-2080, 2081-2090, and 2091-2100 are given in Table 2. The HIRHAM simulations are of course one possible future course of climate, and cannot be considered to be a prediction; the model results do provide, however, a plausible set of scenarios upon which impacts analyses can be constructed.

Insert Table 2 here

Superimposing the future climate data in the temperature-precipitation space shows that the expected winter minimum temperature warming of over 4°C by the end of the 21st century is likely to lead to a reduction in snow duration by more than 100 days at both Säntis and at Arosa, the two sites that were used in BKG03. The increase in winter precipitation that intervenes during this period only slightly modulates the dominant effect of the +4°C warming, as seen in Figure 5. This figure is an extension of a previously-published graph that was applied to current climatic conditions. The "migration" of the Arosa and Säntis statistics can be considered to be highly significant, because the location of snow duration in the temperature-precipitation space, under future climatic conditions, is well outside the range of natural variability of current snow duration. This variability is defined by the ellipses that are centered on the average temperature-precipitation data for the two alpine sites (cf. BKG03 for more details). The standard deviations of temperature and precipitation under future climatic conditions are very close to those for current climate, so that the ellipses are essentially the same; the ellipses for future climatic conditions have thus not been plotted for the sake of clarity in this diagram.

Insert Figure 5 here

On the basis of shifts in snow duration, estimated from the chart in Figure 5 and using the snow depth/snow duration relationship illustrated in Figure 3, it is possible to generate snow-volume curves for two scenarios of 2°C and 4°C warming of winter minimum temperatures, for drier-than-average and moister-than-average conditions. As shown in Figures 4a and 4b, the overall volume of snow under warmer conditions is significantly less than under current climate, irrespective of the conditions of winter precipitation. The maximum volume of snow remains in all cases at 2000 m, but the reduction in volume corresponds to 10% per degree of warming. At the 1000 m level, the reduction in snow volume for the wetter-than-average winter season is 50% for a 2°C average winter minimum temperature rise, and over 95% for an average 4°C rise in DJF minimum temperatures, whereas at 4000 m, the respective volume changes are 8% and 14%. In other words, strong warming at low elevations will lead to the removal of almost all the snow pack, while at very high elevations, the reduction in snow amount will be modest. This is mainly because even in a warmer world, temperatures will remain cold enough to sustain snow for much of the year. Similar conclusions can be reached for the drier-than-average winters (Figure 4b), where it needs to be emphasized that for current climate, the volume of snow is reduced by 20% or more compared to wetter-than-average winters. At all elevations, the sensitivity to warming is somewhat greater than in the case of wetter-than-average winters, because there is less precipitation during dry winters that can to some extent compensate for the warming effect on total snow accumulation. Hence, at 2000 m, for example, a 2°C warming is accompanied by a 40% reduction in snow volume, while a 4°C warming results in a 60% reduction (compared to 20% and 40%, respectively, for moist winters).

In terms of hydrological systems whose source region is in the alpine domain, and therefore where the dominant factor for flow is controlled by the seasonal snow pack, similar reductions in runoff are to be expected as soon as the melting season intervenes. The integral reduction in snow volume, i.e., the change in total snow amount cumulated for all elevations, is of the order of 30% for a 2°C warming and 55% for a 4°C warming, compared to current conditions. In terms of absolute figures for snow volume, current figures for moist and dry winters are $165x10^9$ and $128x10^9$ m^3, respectively; these drop to 77 and $59x10^9$ m^3, respectively, for a 4°C rise in average winter minimum temperatures. If other climatic conditions do not change in a warmer climate in the Alpine region, then the direct consequence of the reduction of snow pack will indeed be to curtail peak stream-flow at the moment of maximum snow-melt by at least a factor of 2, if the +4°C temperature scenario is realized in coming decades. In order to compensate for such shortfalls, rainfall in the summer and autumn months would need to increase substantially compared to current climate. Analyses of the HIRHAM RCM simulations for the alpine region show that the only season with an increase in precipitation occurs during winter (25-30% for 2071-2100 over 1961-1990). Spring and autumn show little change (between –2% and + 2% for future climate compared to current climate), while summer shows a marked decrease of roughly the same amount as the wintertime increase.

There is thus no compensation for the strong reduction in snow pack through enhanced precipitation during other seasons.

The empirical approaches discussed in this paper, as well as in BKG03, allow useful estimates of changes in snow amount to be made under changing climatic conditions. However, it is difficult to assess an additional important parameter for hydrology and for vegetation, namely the *timing* of the beginning and end of the winter season. While BKG03 and this paper have shown how snow amount and duration change in response to shifts in temperature and precipitation, estimates of the change in duration alone does not allow any conclusions to be reached as to the timing of this change. In terms of hydrology, it is important to know not only the total amount of water that the snow pack will contribute to in terms of runoff, but also the period of the year when peak runoff can be expected to occur. Similarly, when investigating the vegetation cycle of many mountain species, the start of the cycle is essentially controlled by the time of the year when the snow begins to melt, and the end of the season is triggered by the decrease in day length; by the time the plants have terminated their annual cycle, the first snows accumulate at the higher elevations in the mountains. Empirical relationships linking winter mean temperature and precipitation to specific characteristics of snow seasonality cannot be established, because very dissimilar winters in terms of their beginning and end can be related to very similar mean winter temperature and/or precipitation characteristics.

Insert Figure 6a and 6b here

Simulations made with the GRENBLS model have been undertaken to address this point and the results are illustrated in Figure 6a and 6b, for the particular case of the high-elevation summit of Säntis (2500 m above sea-level). The two examples given here are for a warm/dry winter (1988/89) and a warm/moist (1998/99) winter, respectively; note the difference in scale on the ordinate between the two figures. For both winter types, the observed winter snow pack and the control simulation are provided. Two additional curves are given, where precipitation remains the same as in the control simulation, but where temperatures are modified by imposing daily temperature increments that are based on the daily averages simulated by the HIRHAM model for the period 2071-2100 (see Table 2). One curve (Case A) predicts the behavior of the snow pack if only the minimum temperatures increase compared to the control simulation (i.e., the maxima remain the same as in the control simulation); this corresponds to a situation where the diurnal temperature range decreases throughout the year. The other curve (Case B) highlights changes in snow amount and duration if both the minimum and the maximum temperatures are increased.

The control simulation is shown in order to illustrate the model's capacity in reproducing the observed situation, and to test the robustness of the climatic inputs that enter into the model computations. It is seen that the overall patterns of total snow accumulation and spring-time ablation are well reproduced in the control simulation, event if in some instances there are some over- or under-estimations of snow depth, for reasons that are discussed in Keller and Goyette (2003).

In both simulations of a warmer climate, it is seen that the duration of the snow season and the total accumulation of snow is reduced. In the warm/dry mode, peak snow amount is reduced by 1 m for Case A and by 1.5 m for Case B. The snow season ends about 30 days earlier in Case A and 60 days earlier in Case B. Case A is thus seen to be an intermediate response of snow cover and duration between the current climate and a general warming of maxima and minima as given by Case B. The simulations serve to emphasize the sensitivity of the system to differential forms of warming; for simulations in which only the maximum temperatures increase (not shown here), the total snow accumulation and the duration of the season are slightly longer than if the warming concerns only minimum temperatures, presumably because cooler nocturnal temperatures partially offset the enhanced diurnal temperatures. In the warm/moist mode (Figure 6b), the reduction in peak snow amount is about 1.3 m for Case A and close to 2 m for Case B. The duration of the snow season, however, is much reduced, by over 70 days for Case A and by 100 days for Case B. The implications of the simulations illustrated in Figure 6b are that the higher precipitation associated with the simulation is not capable of compensating for the compounding effects of increased temperature and changes in the various components of the surface radiation balance.

The GRENBLS simulations all show that the melting period at the end of the snow season is the one that exhibits the greatest sensitivity to warming. According to the numerical simulations, the beginning of the snow season is likely to remain the same as for current climate, but the end of the season will occur earlier than today. At a site such as Säntis, the period of accumulation is thus likely to commence at roughly the same time as today (because temperature conditions, even though warmer than today, would be sufficiently cold to sustain a snow surface once it is established), but the termination of the season will probably occur much earlier. This is not in contradiction to what is already observed under current climatic conditions, as illustrated in Figure 1.

Relating the numerical results to the empirical approaches exemplified in Figure 5, the simulations of the warm/moist mode correspond well with the change in the duration of the snow season at Säntis. The reduction of total snow accumulation that Figure 3 allows to estimate on the basis of the reduction of snow duration is also in good agreement with the model results. For the warm/dry mode, for example, the observed annual snow accumulation is 777 cm with a snow season of 262 days; the empirical relation illustrated in Figure 3 for such a duration is 839 cm. A reduction of 60 days, implied by the model for a 4°C increase in minimum and maximum temperatures leads to an overall snow accumulation of 586 cm, while the empirical relationship suggests a figure of 584 cm.

3. Conclusions

The results from this study confirm other findings, summarized in the IPCC (2001) among others, that the snow pack in mid-latitude mountain regions of the world such as the Swiss Alps is highly sensitive to changes in climatic conditions, and in particular temperature. Under contemporary climatic conditions, dramatic shifts in snow amount and duration can already be observed when winters change from cold and moist situations to much warmer regimes, accompanied by wetter or drier conditions. The amplitude of change is damped with altitude, but at all elevations, the change is significant and corresponds roughly to a reduction in snow cover duration of 15-20 days for each degree of wintertime warming. If the projections of warmer winter temperatures throughout Europe are realized in the course of the 21st century, and in particular the possible 4°C increase in wintertime minimum temperatures compared to current climate in Switzerland, then sharp reductions in the length of the snow season and the total amount of snow are to be expected. The change in seasonality and amount will then have direct and indirect implications for numerous mountain environments, in particular rivers and ecosystems and, in the more populated parts of the Alps, also economic consequences. At altitudes up to 1000 m, reductions in snow amount will exceed 90% for an average winter minimum temperature increase of 4°C; at 2000 m, corresponding reductions will range from 45-60% according to the precipitation regime, and from 30-40% at 3000 m elevation. The spring melt during warm and moist winters is seen to intervene 60-100 days earlier at high elevations such as Säntis (2500 m) and 110-150 days earlier at medium elevation sites close to the 1000 m altitude, while the beginning of the season seems to be relatively insensitive to warmer climatic conditions.

Insert Figure 7 here

Such statistics can be summarized in graphical form as in Figure 7, where the reductions in snow cover duration as a percentage of current average conditions have been charted for Switzerland. This figure is essentially a projection in 2-D space of the information provided in Figures 4a and 4b. The location of regions with topography contrasts sharply with the low-lying areas of the country, where snow in a warmer climate will be almost totally absent below elevations of 500 m. This figure is based on an average change in climatic conditions and does not show what kind of changes may occur with strong interannual variability of climate in the future. Additional maps could be generated for the four winter modes discussed towards the beginning of this paper. Improvements to this type of representation could also be achieved by multiple simulations with the GRENBLS model for groups of sites that exhibit similar orientations and aspects. In this manner, contrasts for north and south facing slopes, for example, would be emphasized, adding a further degree of realism to the representation of snow pack behavior under future climatic conditions.

The results of this study thus have high relevance for investigations of the response of hydrological systems originating in the Alps, where the dominant control on river flow is the amount of snow and the timing of spring melt and surface runoff. In addition, various plant species may respond to these shifts in snow conditions, and in particular the earlier exposure of soils in the spring, in different manners. More robust, "opportunistic" species may compete with those that are genetically geared to current climatic conditions and may have difficulty in adapting to the rapidly changing climatic conditions that are suggested by high resolution regional climate models for Europe in the course of the 21st century. Finally, the investigations reported here can be of value to mountain resorts in their long-term planning for investments and diversification of activities in the face of winter seasons that are likely to become shorter and with a tendency towards marked reductions in snow amounts.

4. Acknowledgements

This study has been partially supported by the NCCR-Climate (Swiss climate research network funded by the Swiss National Science Foundation), the 5th Framework Program PRUDENCE project (funded by the Swiss Federal Office for Education and Science), and the Department of Geosciences of the University of Fribourg. The authors would like to thank Dr. Ole Bossing Christensen from the Danish Meteorological Institute for providing the data from the HIRHAM regional climate model simulations. Thanks also to Stefan Winter from the Department of Geosciences of the University of Fribourg for preparing Figure 7.

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Table captions

Table 1: Statistics of snow duration and total snow accumulation in the winter season for the four winter modes, namely cold/dry (C/D), cold/moist (C/M), warm/dry (W/D), and warm/moist (C/M).

Table 2: Average decadal change in winter (DJF averages) temperature and precipitation in the alpine domain in a future climate compared to the 1961-1990 mean climate, based on Regional Climate Model simulations by the HIRHAM model. The 30-year average change is also indicated in the last line of the table.

Figure captions

Figure 1a: Seasonal snow accumulation curves at Château d'Oex (981 m asl) for four different winter modes (cold/dry; cold/moist; warm/dry; warm/moist).
Figure 1b: As Figure 1a, except for Arosa (1847 m asl). Note the difference in scale for the ordinates in the two figures.

Figure 2: Relation between winter mean temperatures and total seasonal snow accumulation for wetter-than-average (closed squares) and drier-than-average (open circles) winters. Bold and dashed lines represent the linear regressions for both sets of data, respectively.

Figure 3: Relation between snow duration and total seasonal snow accumulation for wetter-than-average (closed squares) and drier-than-average (open circles) winters. Bold and dashed lines represent the regression curves for both sets of data, respectively.

Figure 4a: Altitudinal distribution of maximum snow volume for wetter-than-average winters, under current climatic conditions (bold line), and for warming scenarios of 2°C and 4°C, respectively.
Figure 4b: As Figure 4a, except for drier-than-average winters.

Figure 5: 2-D contour surfaces of snow-cover duration as a function of winter (DJF) minimum temperature and precipitation, based on the data extracted from all climatological sites used in this study. Superimposed on this figure is the temperature-precipitation-snow duration data for the Arosa and Säntis sites, for current climatic conditions, and for the three last decades of the 21st century (see text for details). The ellipses show the 2σ range of DJF minimum temperature and precipitation, and corresponding spread of snow-cover duration for Arosa, and Säntis. The orientation of the ellipses is related to the covariance of temperature and precipitation. The figures on the isolines identify the length of the snow season.

Figure 6a: Simulations with the GRENBLS surface energy balance model of the 1988/89 winter snow pack at Säntis, for current climate (ctrl) and for two future climate scenarios (Case A - min: increase of DJF mean minimum temperatures only, according to the scenarios simulated by the HIRHAM model; Case B – mean: increase of minimum and maximum temperatures according to the HIRHAM model-generated scenarios). The closed circles represent the observed (obs) winter snow pack evolution.
Figure 6b: As Figure 6a, except for wetter-than-average winters, using the 1998/1999 winter as the reference.

Figure 7: Map of Switzerland illustrating the reduction in snow volume that can be expected following an increase of 4°C in DJF minimum temperatures, compared to contemporary climate.

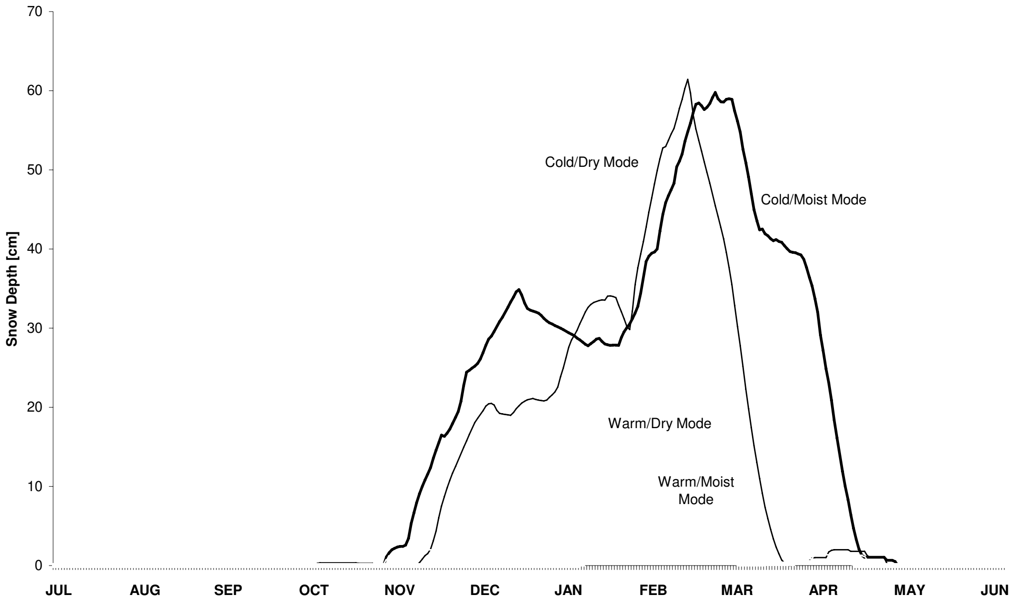
Beniston et al., 2003: Table 1

Winter mode	Average DJF temperature (°C)	Average DJF precipitation (mm/day)	Duration of snow pack (days)	Seasonal snow accumulation (cm)
Château d'Oex				
C/D	-3.8	2.5	125	322
C/M	-3.8	4.3	155	520
W/D	0.5	2.4	67	138
W/M	1.0	4.8	22	94
Arosa				
C/D	-6.4	1.9	215	959
C/M	-6.4	3.3	237	1150
W/D	-1.3	2.1	183	562
W/M	-0.7	3.8	198	552

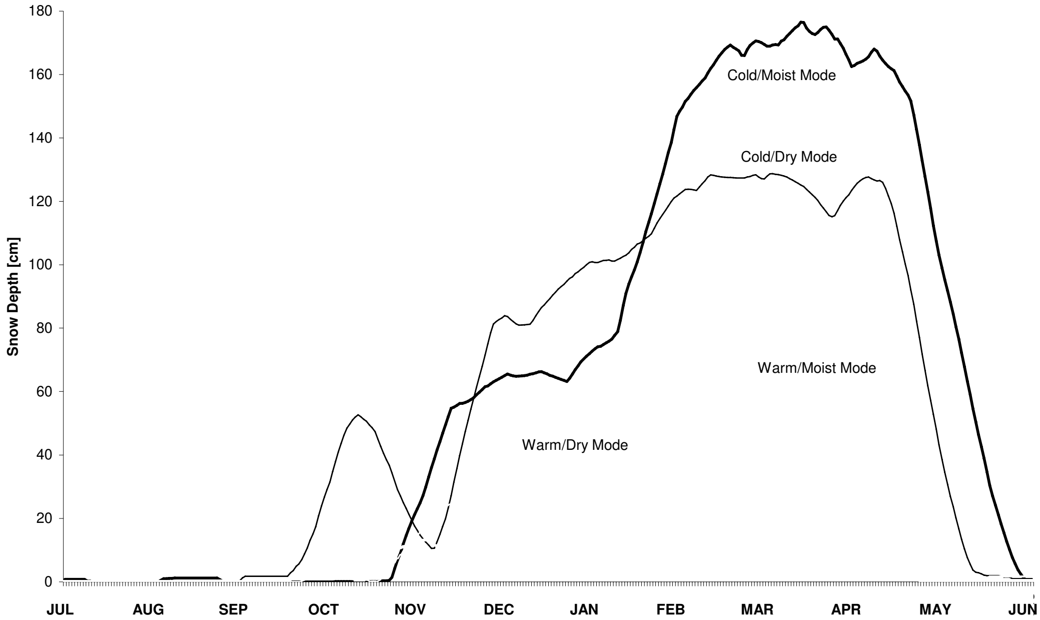
Beniston et al., 2003: Table 2

Period	Average DJF Tmin change (°C)	Average DJF precipitation change (mm/day)
2071-2080	3.6	1.1
2081-2090	4.2	0.9
2091-2100	4.6	1.0
2071-2100	4.1	1.0

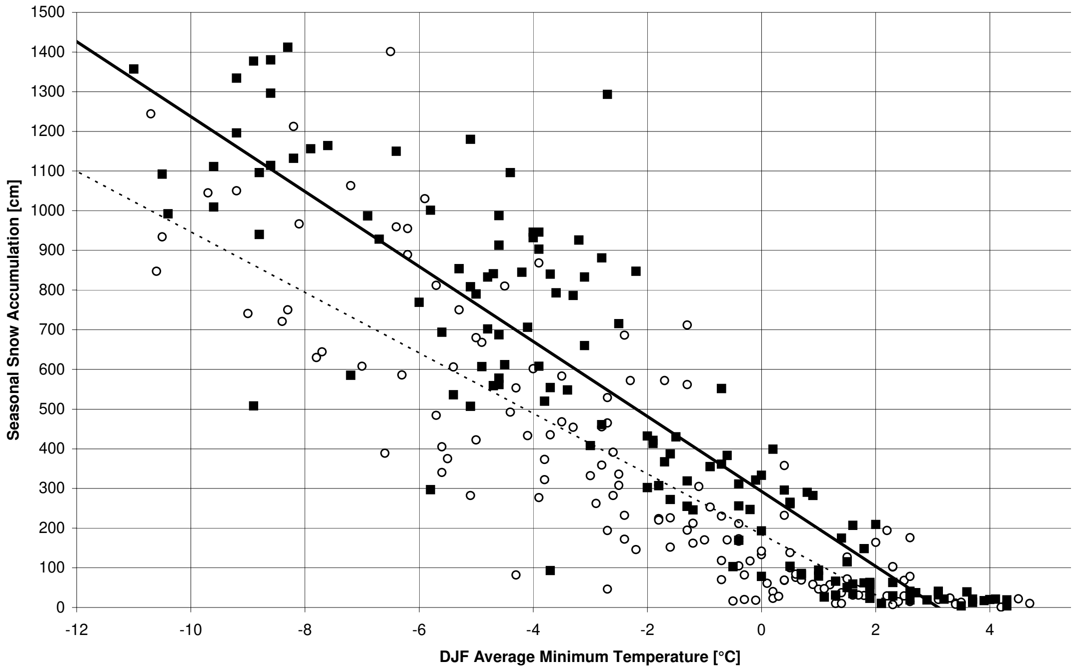
Beniston et al., 2003: Figure 1a



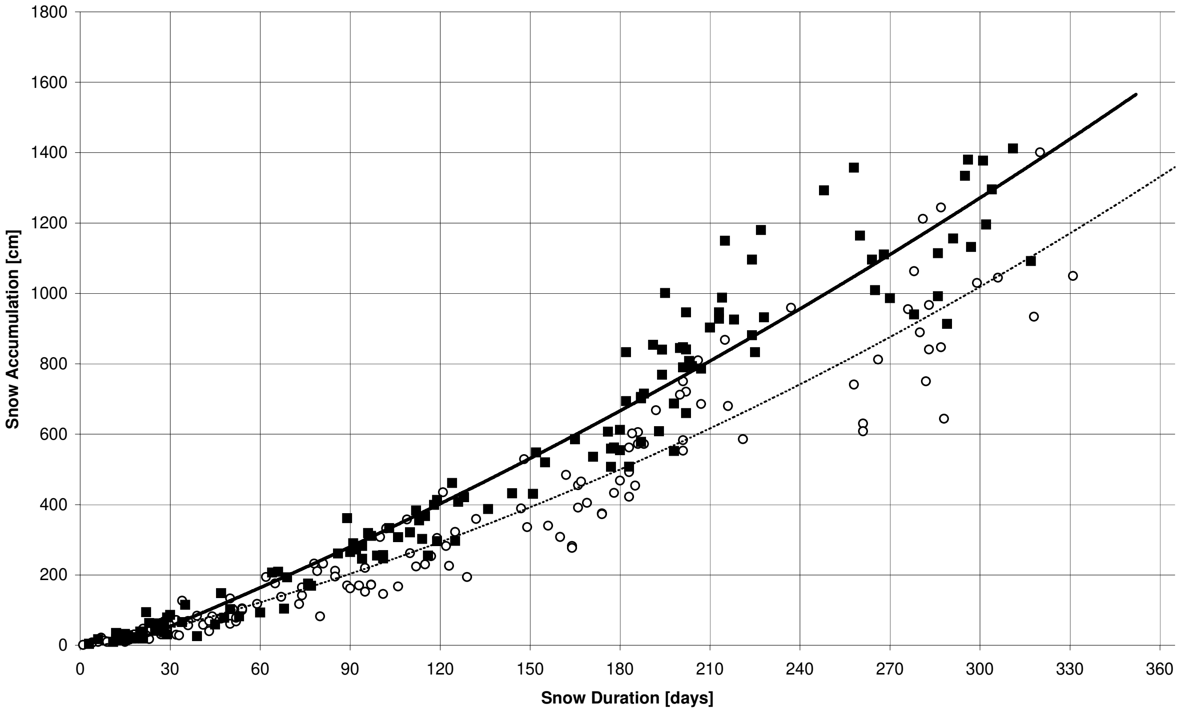
Beniston et al., 2003: Figure 1b



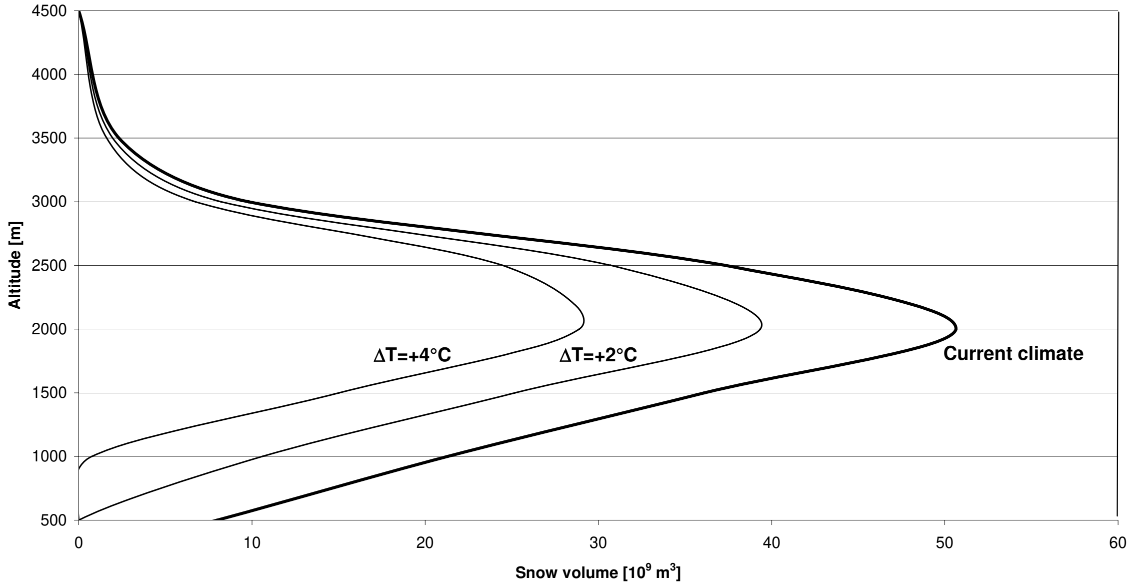
Beniston et al., 2003: Figure 2



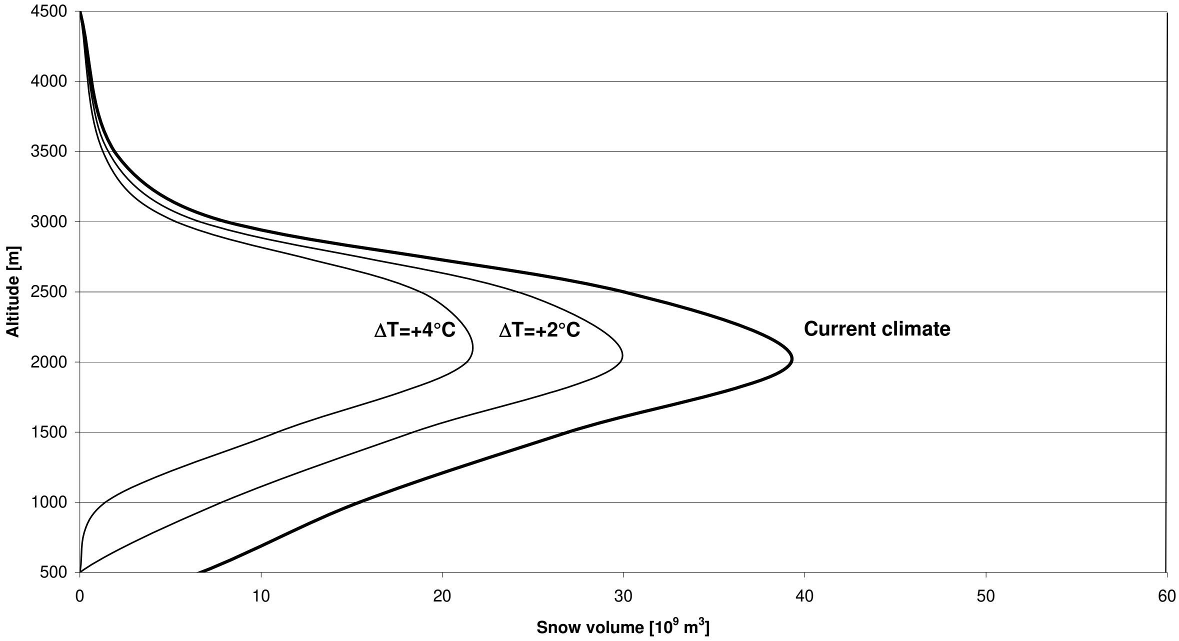
Beniston et al., 2003: Figure 3



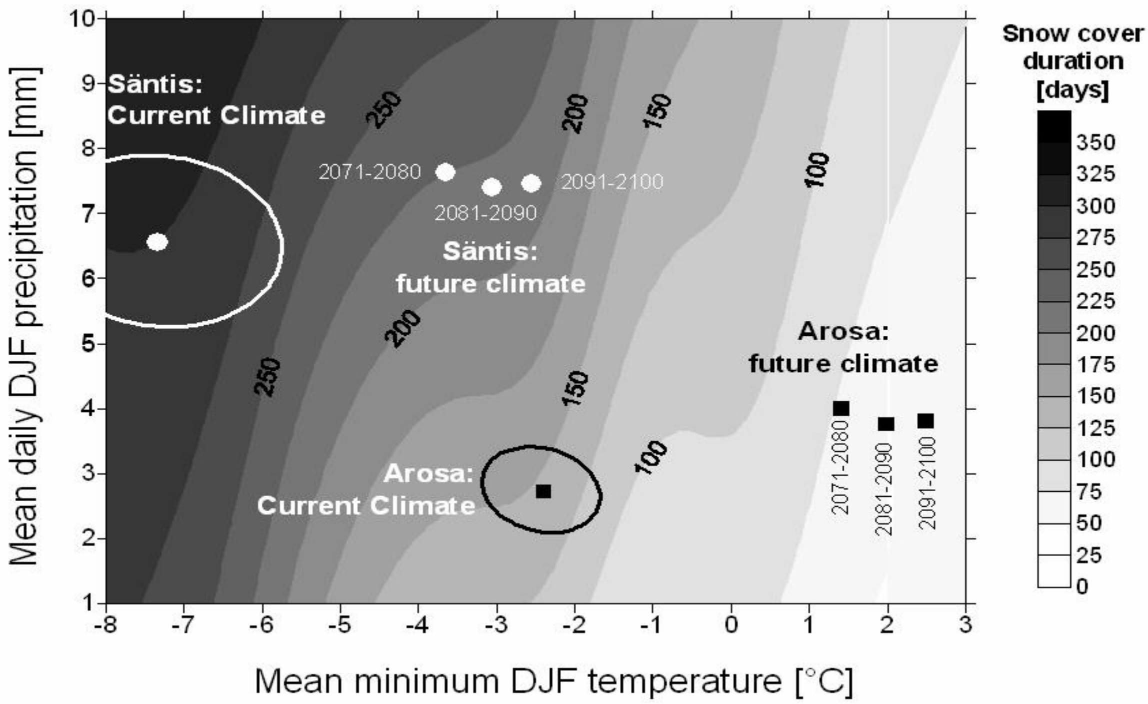
Beniston et al., 2003: Figure 4a



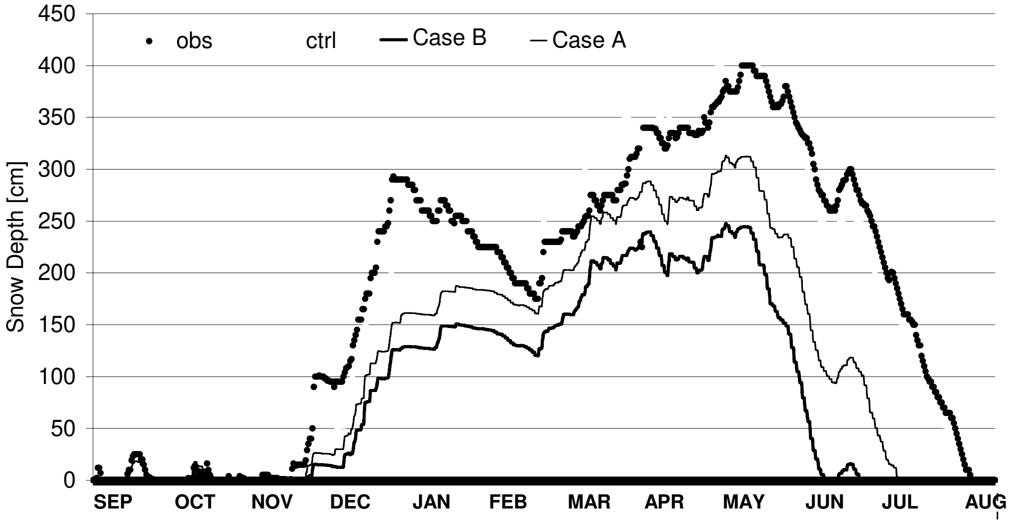
Beniston et al., 2003: Figure 4b



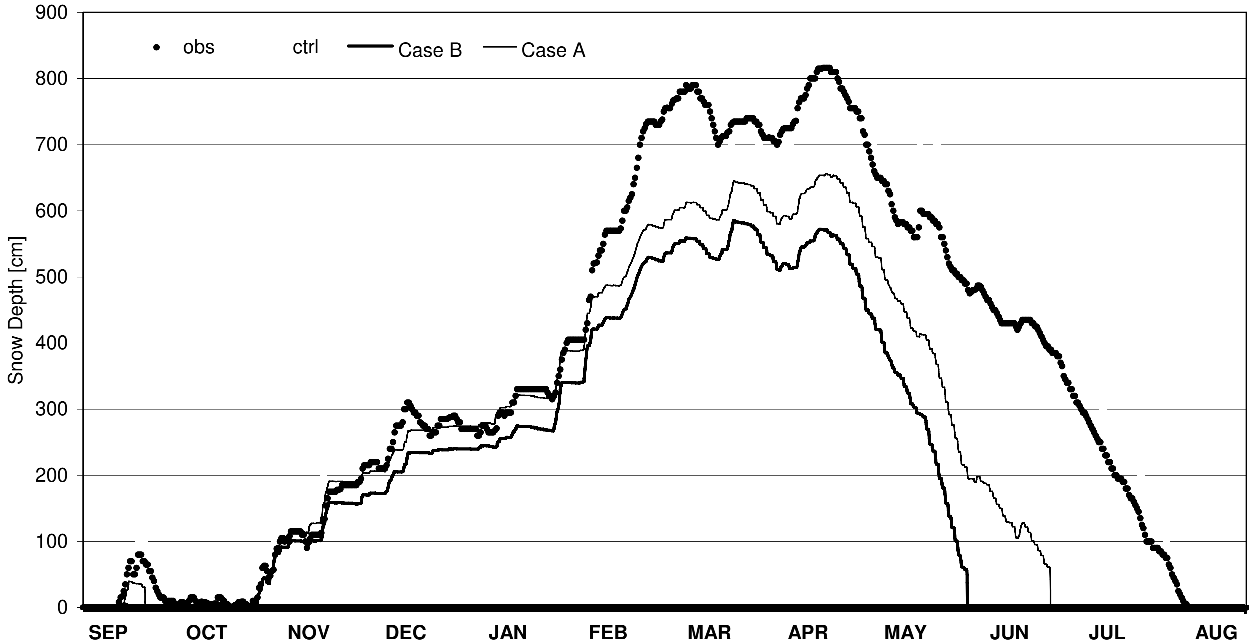
Beniston et al., 2003: Figure 5



Beniston et al., 2003: Figure 6a



Beniston et al., 2003: Figure 6b



Beniston et al., 2003: Figure 7