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Estimation of the Heated Rain Gauge Error at a High-Elevation Site in the Italian Alps thanks to the Modelling of the Snow Cover Water Equivalent

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Abstract

Measuring the amount of solid precipitation by means of a heated rain gauge leads to systematic under-measurement because of under-catch problems. The magnitude of the measurement error can be estimated by comparing the amount of precipitation measured by the heated rain gauge with the increase in measured snow cover water equivalent (SWE). The data used are the ones recorded at the observing site of Lago Pilone (Piedmont, Italy) during the winter 2008/2009. Before being used, SWE data are checked for consistency towards manual observations. A model (SNOWPACK) is used to confirm the rain gauge under-catch, and the modelled precipitation is used as a reference to estimate the measurement error. Threshold values for wind speed and air temperature are used to explain the occurrence of measurement error. Quite good relationships are found between the error magnitude, wind speed and air temperature. By adding the error term to the measured value, the precipitation data are reconstructed quite well.

KEY WORDS : heated rain gauge ; measurement error ; SWE ; correction of precipitation data

List of Symbols

Abbreviation	Signification
LW	Long-wave
SW	Short-wave
SWE	Snow water equivalent

Symbol	Description	Units
E	Heating rain gauge error	mm
H	Snow height	m
k	Conversion factor	
$LW_{in/out}$	Incoming or outgoing long-wave radiation	$W.m^{-2}$
m_{sn}	Mass of snow	kg
P_{corr}	Corrected precipitation	mm
P_g	Amount of precipitation recorded by the precipitation gauge	mm
P_{meas}	Precipitation measured by the heating rain gauge	mm
P_{sb}	Sum of new snow SWE obtained from measurements on a snowboard	mm
Q_{gr}	Ground heat conduction	$W.m^{-2}$
Q_{int}	Changes in snowpack sensible and latent heat storage	$W.m^{-2}$
Q_{lat}	Convective exchange of latent heat with the atmosphere	$W.m^{-2}$
Q_{melt}	Loss of latent heat of fusion due to melt water leaving the snowpack	$W.m^{-2}$
Q_{rad}	Net radiant energy exchange	$W.m^{-2}$
Q_{rain}	Rainfall sensible and latent heat	$W.m^{-2}$
Q_{sens}	Convective exchange of sensible heat with the atmosphere	$W.m^{-2}$
R_{net}	Net radiation at the snow-surface interface	$W.m^{-2}$
ρ_s	Water density	$kg.m^{-3}$
$SW_{in/out}$	Incoming or outgoing short-wave radiation	$W.m^{-2}$
SWE	Snow water equivalent of snow	mm
T	Air temperature	°C
V_{sn}	Volume of snow	m^3
V_w	Volume of water represented by snow	m^3
w	Wind speed	$m.s^{-1}$

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Introduction

The Politecnico di Torino (Turin Polytechnic), founded in 1906, is one of the most important school of engineering in Italy. It consists of 6 schools, 4 of engineering and 2 of architecture, which spread over different campuses in Piedmont, and it has 18 research departments. There are around 28.000 students studying on 116 courses, there is a staff of over 900 lecturers and researchers, and around 875 administration staff. Besides, in the 2009 forecast balance, the income is 300 million euros, 129 million of which are from the Ministero dell'Istruzione Università e Ricerca (Ministry for Universities Education and Research). So, not only the Politecnico is a teaching university with a broad range of studies, including space, environment and land, telecommunications, information, energy, mechanics, electronics, chemistry, automation, electrical engineering, industrial design, and architecture and building, but it is also a research university interested in the development of both theoretical and applied research.

Among those 18 departments, there is the DITIC, Department of Hydraulics, Transport, and Civil Infrastructures, which is divided in 4 laboratories : one for theoretical and applied research in hydraulics and related disciplines, a centre for meteorology and hydrology, a roads laboratory and a transport laboratory. So, it is concerned with all areas of hydraulics and transport, which include fluid mechanics, fluvial, environmental, and maritime hydraulics, hydrology and water resources, hydraulic and maritime constructions and layouts, roads, railways, transport of goods and passengers, safety and environmental impact of transport systems. National and international partnerships in research have led to many exchanges among institutions for teaching staff, researchers and post-graduate students. Moreover, the department undertakes many projects in collaboration with regional, provincial and town councils.

The following study is part of a research project developed by the regional council of Piedmont (Direzione Pianificazione delle Risorse Idriche, i.e. Direction of Water Resources Planning), in collaboration with the DITIC, which began two years ago. Its objective is to determine the amount of water stored in the snow cover thanks to the activation of two experimental sites, and to develop indicators of the quantitative state of this water resource (Conoscenza della riserva idrica nevosa anche attraverso l’attivazione di siti sperimentali di misura e messa a punto di indicatori dello stato quantitativo delle risorse idriche).

Knowing the snow cover water equivalent (SWE) is important for snow and water management. For example, it helps to predict the amount of water which would be available for agriculture during spring and summer. It can also help to predict the intensity of possible floods during the melting period.

This amount of water stored in the snow cover comes from the different precipitation events which occur during a winter. Each of these events adds a new quantity of water to the snow cover. This quantity can be deduced from the height and the density of the new snow layer. However, the amount of water stored in the snow cover can be directly measured thanks to an instrument called snow pillow which consists of a pillow filled with an anti freeze fluid, so that the amount of snow laying on it can be converted into pressure variation. Nevertheless, this kind of instrument is quite expensive (around 4000 $) and its installation is not easy. On the contrary, instruments which measure air temperature, relative humidity, wind speed, etc. are much more widespread. Therefore, relating new snow density to some meteorological parameters would permit the determination of snow cover water equivalent in similar places where no snow pillow is present.

On the other hand, the water equivalent of winter precipitations is frequently measured thanks to a rain gauge which is heated, so that the solid precipitation can be melted. However, such an instrument provides data with a quite important uncertainty. So, comparing these data to those from the snow pillow - or later to those derived from some relationship between new snow density and meteorological parameter - would permit to better estimate measurement errors from the heated rain gauges. This last objective is the one retained for the present study.

Nevertheless, some other points will be studied as they can also be of interest for this project. Diurnal variations observed in measured snow depth will be looked more in details since the evolution of this parameter allows the determination of the new snow layer height. Besides, a method has been proposed to select periods without measurement errors, but an approximation has to be done. So, the validity of this approximation will be verified.

The two experimental sites which were chosen are Lago Pilone (2320 m) and Limone Piemonte (2020 m), both situated in the region of Piedmont (Italy). At Limone Piemonte, because of early snowfall during the winter 2007/2008, installing the snow pillow was not possible. Moreover, during the winter 2008/2009, the snow pillow emptied of its content, so that no SWE data are available. For those reasons, this site will not be studied. Regarding the site of Lago Pilone, the rain gauge was not heated during the winter 2007/2008, thus leading to inaccurate precipitation values. So, the data studied are mainly the ones measured at Lago Pilone during the winter 2008/2009.

I. Generalities about Snow

I.1. Snow Crystals : Formation and Variation

The snow-formation process requires three conditions : cold air, moisture and airborne particles. Subfreezing clouds, where liquid, vapour and solid water are present, are at the origin of snow, and for them to exist, a mean for cooling is required.

a) Cooling Air Mass :
(Doesken and Judson, 1997)

The principal means to cool air to saturation is to lift it. There are several lifting processes, operating independently or in combination.

The first mechanism is called orographic lifting : air is forced upward over elevated terrain. This phenomenon explains why sometimes one side of the mountains is covered by snow, whereas the second one doesn’t record any precipitation.

The second process is due to warm fronts or cold fronts (*Figures 1 and 2, Appendix A*). Warm fronts consist of warm air moving horizontally and being lift over colder air, while cold fronts consist of cool air pushing under warm air. Since the boundary between cold and warm air is steeper with cold fronts, lifting is more rapid, leading to more intense snowfall, generally over a smaller area and for a shorter duration than with warm fronts.

The last important process is called convergence, accompanied by a phenomenon of divergence aloft : air mass converges towards regions of low pressure at the surface, and excess flows are forced upward. It often results in widespread snowfalls.

Nevertheless, air motions in the atmosphere are complex and other processes can occur, for example vertical temperature gradients may produce upward motions. Besides, by adding moisture to the air, it is possible to reach the saturation without lifting processes. This phenomenon, called lake-effect, takes place when warm water from lakes or oceans evaporates into a cold air mass above.

b) The Snow-Formation Process :
(Doesken and Judson, 1997)

Thanks to evaporation and transpiration from oceans, lakes, soil or plants, water vapour is transported into the atmosphere. As air mass rises, it cools and its capacity to contain water vapour decreases. Once saturation conditions are reached, water vapour condensates onto small particles forming droplets. When the air keeps on cooling, most of these droplets remain as liquid even if the temperature is well below zero, thus obtaining super-cooled droplets. At the same time, water vapour changes directly into ice, which one is deposited onto tiny particles called freezing nuclei. For a specific temperature, vapour pressure over water droplets is greater than the vapour pressure over ice. Thanks to this, water molecules are transferred from droplets to ice crystals, and as water droplets get smaller, ice crystals continue to grow, thus forming snow. Then, when the mass of a crystal is enough, it begins to fall through the clouds. Sometimes it can collide with a super-cooled droplet which instantly freezes on impact, thus creating rimed crystals. However, many snow crystals melt or evaporate before they reach the ground.

c) Variation in Snow Crystal Type and Size :

The snow-formation process is always the same, but depending on meteorological conditions, the type of crystals that form changes (*Figure 3, Appendix A*). They can be classified in four general types : needle, hexagonal plate, dendrite, hexagonal column. Combinations of these types are common and their frequency of occurrence depends on meteorological conditions.

Experiments revealed that the type of crystal formed depends on air temperature : needles and irregular crystals formed between -5 and -8 °C, hexagonal plates and columns near -9 °C, dendrites at -14 to -17 °C, and hexagonal plates and columns below -20 °C (Gold and Power, 1954). In fact, the type of crystal formed depends on the number of water molecules which condense per unit of area per second (aufm Kampe et al., 1951). This quantity is proportional to the difference between the vapour pressure over water and the vapour pressure over ice, which one varies with temperature.

Furthermore, it was noted that the degree of supersaturation with respect to ice influence the shape of the crystal (aufm Kampe et al., 1951). It increases as air temperature decreases, and leads to crystals which look smaller and much simpler in shape. It was also found that the particle used as a nucleus doesn’t influence the principal shape of the mature crystal.

I.2. Density of Fresh Snow

One of the most important and useful characteristics of new snow is its density. Indeed, it links its volume and its mass, equal to its water mass, which are generally expressed in term of heights (*Equation 1*). It has to be noted that equation (1) is used for both a fresh snow layer and a complete snow cover. So, knowing the density and the height of a snow deposition, its water equivalent can be deduced. Typically, new snow density varies from 10 to 350 kg.m^{-3} (Judson and Doesken, 2000). Besides, it depends on many different parameters.

$$(1)\quad \rho\ \left[kg.m^{-3}\right] = \frac{m_{sn}}{V_{sn}} = \rho_w \frac{V_w}{V_{sn}} = \frac{SWE\ \ [mm]}{H\ \ [m]}$$

With
- m_{sn} : Mass of snow deposition
- V_{sn} : Volume of snow deposition
- V_w : Volume of water represented by the snow deposition
- ρ_s : Water density
- H : Snow deposition height
- SWE : Snow water equivalent, i.e. snow deposition height after it has been melted

a) <u>Influence of Snow Crystal Shape and Size :</u>

Grant and Rhea (1974), and Power et al. (1964) explained variations in new snow density thanks to some physical causes. Crystal shape influences density as it controls the amount of air entrapped between the individual snow crystals. For this reason, dendrites, which have spatial extensions, produce the lowest densities, whereas needles, plates, and irregular crystals are associated with higher density values. For a specific shape, variations are explained by the random fitting together of the snow crystals as they settle on the snow cover. Riming, by reducing inter-crystal space, frequently causes higher densities, and can increase it by 30 to 100 %. Likewise crystal size influences density, with smaller crystals producing higher densities as they allow for better packing.

b) <u>Influence of Meteorological Parameters :</u>

The first studies which were done concern the relationship between air temperature and new snow density. In general, new snow density seems to have a parabolic dependency on surface air temperatures, with a minimum near -11 °C, a maximum from -20 to -25 °C, and a considerable scatter between -5 and -7 °C (Bossolasco, 1954). McGurk et al. (1988) found a linear relationship between site air temperature and fresh snow density (r=0.52), with density increasing when air temperature increases. However the air temperatures used are greater than -10 °C, therefore the part where density increases with decreasing air temperature can not be seen. Besides, regarding wind speed and relative humidity influence on new snow density, weak correlations have been found.

Moreover, Diamond and Lowry (1954) studied the relation of new snow density to the upper air temperatures. They found a linear relationship (r=0.64) when using the 700-mb temperature. A good correlation was also found with the surface air temperature (r=0.50) as it is slightly below the 700-mb level, but no relationship was found between new snow density and the 500-mb temperature. Judson and Doesken (2000) also found a linear relationship (r=0.52), with a large variability in densities at warmer temperatures. Grant and Rhea (1974) also studied the relationship between the 700-mb temperature and new snow density, but a parabolic dependency was found with higher densities at both the lowest and highest temperatures.

Furthermore, instead of taking into account air temperature only, Stashko (1976) also took into account crystal size. He assumed that both crystal structure, related to air temperature, and relative humidity can be represented by the size of the crystal. So, classifying snowflakes by size, he found for each category an hyperbolic relationship between new snow density and surface air temperature with quite good correlation coefficients, generally higher than 0.80.

c) Elevation and Geographical Controls :

Grant and Rhea (1974) found a variation of density with elevation for a specific mountain pass. Greater densities are associated with lower elevations, which could be explained by a higher frequency and degree of riming of the snow crystals falling through the lower and warmer portions of clouds. On the contrary, Judson and Doesken (2000) examined new snow density distribution from 6 sites in Colorado and Wyoming, and found the lowest mean density at the lowest elevation whereas the highest mean density was found at a higher elevation. Simeral (2005) also studied the relationship between density and elevation, but no clear pattern was found.

Judson and Doesken (2000) investigated the spatial variation of fresh snow density, that is to say the relationship between density at one site and density at a second site. It was found that the correlation coefficient decreases with increasing distance between the two sites as snow climate is more likely to be different. Grant and Rhea (1974) also explained variations in new snow density by orographic influences and differences in the storm origin. So, the importance of considering the scale in the analysis of new snow density was underlined.

In summary, it appears that new snow density is affected by geographical and elevation controls. However, new snow density is mainly related to snow crystal types and size, and also to their degree of riming. Moreover, different studies has been done to correlate fresh snow density with surface and upper air temperatures with different results : either a parabolic relationship or a non linear decrease in density with a decrease in temperature. Besides, a relationship between new snow density, air temperature, relative humidity, wind speed and snow surface temperature have been proposed by Lehning et al. (2002) with a quite good coefficient of determination (r^2=0.83), but actually, very few studies have tried to correlate new snow density with all meteorological parameters.

I.3. Snowpack Energy Balance

To explain better snow cover evolution, it is necessary to understand its energy balance which is represented by equation (2) (DeWalle and Rango, 2008). It appears that the ground heat flux Q_{gr} is small in comparison with the other fluxes, and in absence of melt water in the snowpack, there is no loss of latent heat of fusion Q_{melt}. So, the snowpack energy balance is mainly governed by the energy exchanges at the surface of the snow cover, i.e. Q_{rad}, Q_{sens}, Q_{lat}, and Q_{rain}. Furthermore, even if rain on snow has an important influence on water movements in the snowpack and on the water retention characteristics of snow, the amount of energy brought by rain is of minor importance. For those reasons, we will focus on radiation and turbulent exchanges.

$$(2)\quad Q_{\text{int}} = Q_{rad} + Q_{sens} + Q_{lat} + Q_{rain} + Q_{gr} + Q_{melt}$$

With
Q_{rad} : Net radiant energy exchange
Q_{sens} : Convective exchange of sensible heat with the atmosphere
Q_{lat} : Convective exchange of latent heat with the atmosphere
Q_{rain} : Rainfall sensible and latent heat
Q_{gr} : Ground heat conduction
Q_{melt} : Loss of latent heat of fusion due to melt water leaving the snowpack
Q_{int} : Changes in snowpack sensible and latent heat storage

a) <u>Radiation Exchange :</u>

The radiation exchange is the most important component of the energy balance, especially during the day. For example, on prairies without vegetation cover, it can represent from 84 to 100 % of the daytime surface energy input. Besides, radiation transfers have to be separated into two energy balances, one for short-wave radiation, one for long-wave radiation (*Equation 3*). Short-wave (SW) radiation originates from the sun so that it is present during the day only. It consists of a direct beam and a diffuse component. The long-wave (LW) radiation originates from the sky and the surrounded terrain and is present day and night. The other wavelengths represent less than 5 % of the total incoming radiation, so that they are not taken into account (Male and Granger, 1981).

$$(3)\quad R_{net} = SW_{in} - SW_{out} + LW_{in} - LW_{out}$$

With R_{net} : Net radiation at the snow-surface interface
$SW_{in/out}$: Incoming or outgoing short-wave radiation
$LW_{in/out}$: Incoming or outgoing long-wave radiation

Regarding incoming SW radiation, the part which reflects off the snow surface is the outgoing component. The other part is absorbed by the snow surface and penetrates the snow cover, thus influencing the snow cover energy balance. The percentage of incoming SW radiation which is reflected is determined by the snow surface albedo. This parameter can present a large variability and mainly depends on snow surface characteristics : it decreases with wet and old snow, and it increases with increasing grain size. A diurnal variation is also observed because of a variation in snow reflectance with solar angle (higher at low angles of incidence). Besides, snow surface albedo increases with cloud cover because of the multiple reflection process between snow and clouds. Finally, it has to be noted that because of the high albedo of snow, the radiation balance is mainly governed by the long-wave fluxes (Male and Granger, 1981).

The incoming LW radiation coming from the sky is either transmitted through the atmosphere or emitted by the atmosphere, principally by water vapour, carbon dioxide and ozone (Male and Granger, 1981). The other component, which comes from the surrounded terrain, is an important part of the snow surface energy balance when inclined slopes are considered or when other emitting surfaces are close (Plüss and Ohmura, 1996). Moreover, for this range of wavelengths, the reflectance of snow is nearly zero. Therefore, the outgoing LW radiation is linked to snow surface temperature only, that is to say to the Stefan-Boltzmann law applied to a grey corps. Furthermore, the net LW radiation balance can be either positive or negative.

b) Turbulent Exchange :

Turbulent exchange is also an important component in the snow surface energy balance, and consists of both sensible and latent heat convection (DeWalle and Rango, 2008). Convective transfer of sensible heat depends on the magnitude and the sign of the temperature difference between air and snow surface, but also on wind speed, surface roughness and stability of the air. It causes large energy gains when air temperature is much higher than snow surface temperature, for example in late spring, whereas it causes convective energy losses when the snowpack is warmer than the air, i.e. during winter or at night during the melt period. Regarding latent heat convection, it can also be positive or negative. Transfer of vapour from the snowpack to the atmosphere causes loss of latent heat, due to evaporation when liquid water is present, or due to sublimation when subfreezing temperatures prevail. On the contrary, when vapour contained in the atmosphere condensates or sublimates onto the snowpack, energy is added to the snow cover. Energy losses generally occur during winter, whereas energy gains are more likely to occur during melting periods. Latent heat convection depends on the difference in vapour pressure between snow cover and atmosphere, wind speed, surface roughness and stability of the air.

I.4. Variation in Snow Cover Distribution

Snowfall variability is both a spatial and a temporal phenomenon. From place to place, from year to year, the amount of snow and snow cover duration can vary a lot. Nevertheless, spatial and temporal variations also appear at a small scale. According to whether the area is exposed to wind or not, according to whether a forest cover is present or not, snow cover characteristics and evolution are not the same.

In windblown areas, the phenomenon of blowing snow strongly influences the snow surface distribution. It consists of snow particles removed from the snow cover by wind and carried away (*Figure 4, Appendix A*). When vegetation or rocks protrude through the snowpack, it leads to localized drift formation. Such a phenomenon is called drifting snow, and results in an uneven snow surface. Moreover, blowing snow often leads to diminution in snowpack mass. Typically, sublimation of wind transported snow particles can remove from 15 to 45 % of winter snowfall. Blowing snow is enhanced by high amounts of snowfall, high wind speeds and cold air temperatures as they slow snow surface metamorphism and melt (DeWalle and Rango, 2008).

On the contrary, in protected areas like forested areas, variations in snow cover mainly result from differences in canopy interception and wind drift around trees. Moreover, the presence of a canopy has different consequences on the snow surface energy balance. A smaller part of the incoming SW strikes the snow surface, but an important part of the LW radiation is trapped beneath the canopy, so that the radiation balance at the snow surface is nearly zero when a canopy is present. Regarding turbulent heat fluxes, they are generally smaller because of a higher atmospheric stability. Those differences often result in longer snow cover duration in forested areas (DeWalle and Rango, 2008).

II. Measuring Snow : Instruments and Problems

II.1. Measuring Snow

To study new snow density or snowpack snow water equivalent, data must be available, which implies that snow measurements have to be done. For more than 100 years, snow data have been recorded manually, and more recently automated measurements have been developed. Manual observations are still used to verify automated measured data.

a) <u>Manual Measurements :</u>
(Doesken and Judson, 1997)

Snow depth, defined as the total amount of old and new snow on the ground, is manually measured thanks to a simple ruler or to a graduated post.

To record the amount of new snow, a snowboard can be used. It consists of a piece of plywood or other material, painted white to reduce the influence of solar radiation. After new snow depth has been measured, snow is removed and the snowboard is set flush with the snow surface. Nevertheless, the amount of fresh snow is often measured just as rain, that is to say thanks to a rain gauge.

Regarding snow precipitation, there are two ways to obtain its water equivalent. The first is to melt the contents of the rain gauge. Sometimes a heated rain gauge is used, which melts solid precipitation as it falls. The second way is to melt a snow core taken from the snowboard with a calibrated tube. Then, either the volume or the weight of the melted snow is measured, and using the section of the rain gage/calibrated tube, it can be converted into a water depth. So, knowing new snow depth and its water equivalent, new snow density can be deduced.

To determine the snow cover water equivalent, a snow core is taken from the snowpack and as previously, it is melted and converted into a water depth.

Nevertheless, to allow more frequent observations, even in some remote areas which can't be accessed easily, the automation of snow data collection has begun in the 1970's, and during the 1990's, almost all human observers have been replaced by electronic sensors.

b) Automated Measurements :

(Doesken and Judson, 1997)

Different instruments exist but only the most used ones will be presented.

To record snow depth at a specific site, acoustic or optical sensors are widely used (*Figure 1*). This kind of instrument consists of a transceiver module mounted on a horizontal arm directly above snow surface. It sends an acoustic or optical signal toward an area on snow surface, and measures the time required for it to do the round trip. Knowing the wave velocity, the distance between the sensor and the snow surface can be determined, thus obtaining snow cover height.

New snow depth can not be automatically measured, it has to be deduced from measured snow depth evolution.

To measure precipitation water content, a rain gauge is still used. Often if consists of a tall standpipe charged with an anti-freeze mixture, whose volumetric contents is converted into a calibrated depth. Changes in this depth reveal the amount of precipitation. The instrument used can also be a heated rain gauge completed with a mechanical process. The liquid precipitation is collected in a container mounted on a bascule. When the weight is enough high, the container falls over, sending an electronic signal. The number of recorded signals provides the amount of precipitation.

Regarding the snow cover water equivalent, it is principally measured thanks to an instrument called snow pillow (*Figure 1*). It consists of a pillow filled with an anti-freeze fluid. Thanks to a pressure transducer or thanks to the fluid level in a standpipe, the weight of the snow lying on the pillow can be converted into an electrical reading of the SWE. Moreover, increases in SWE combined to increases in snow depth permit to deduce new snow density.

However, whether it is done manually or automatically, measuring snow is not so easy and various problems can occur.

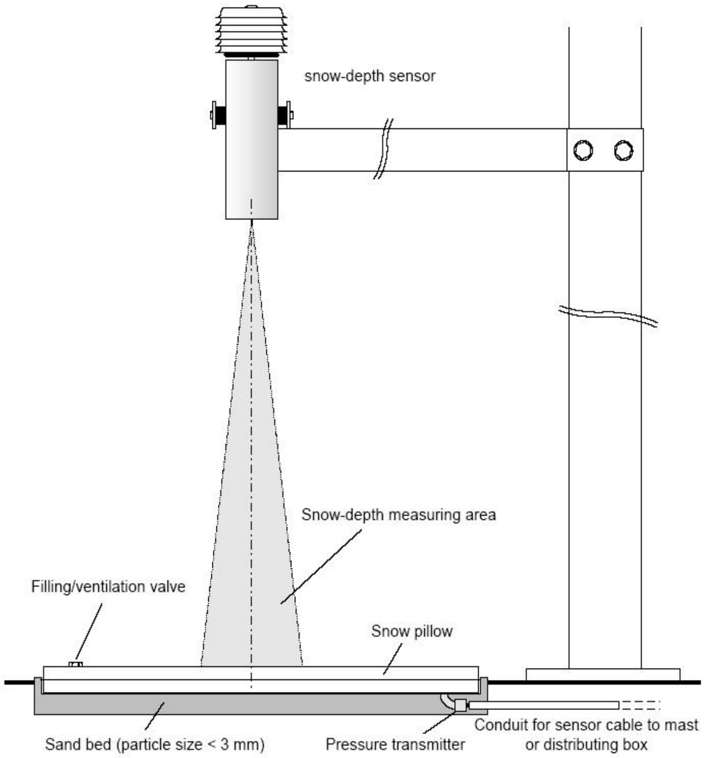

Figure 1 :
Snow Depth Sensor and
Snow Pillow, Working Scheme

II.2. Problems in Measuring Snow

a) <u>General Problems in Measuring Snow :</u>
(Doesken and Judson, 1997)

The fact that variations in snow cover distribution can occur has been underlined previously in part II.4. Besides, the snow cover is a dynamic structure, thus modifying itself with time. Because of that, measuring snow accurately can be difficult.

As just said, snow is a dynamic structure which settles as it lies on the ground. Depending on meteorological conditions and on new snow density, this phenomenon, also called compaction, would occur either rapidly or gradually. New snow is particularly prone to compaction and can undergo very fast changes. Moreover, snow often melts as it lands because of warm soil or because of warm air, wind or sunshine. These properties of snow result in problems exclusively related to manual measurements, which concern the time of the day on which observations are taken, and the frequency with which they are taken. Between a snow event and the time of observation, several hours can have passed, so that the amount of new snow is often under-measured. However, data about the whole snowpack are not affected and they are considered to be quite reliable. Besides, the influence of the snow observer has been reported as quite negligible.

Another problem which concerns both manual and automated measurements is caused by snow cover spatial variability. Indeed, snow can easily be blown and redistributed by wind, sometimes forming drifts or being completely removed from the ground or the snowboard. So, to avoid this problem and obtain accurate data, protected areas have to be preferred. Another consequence of blowing/drifting snow is the creation of an uneven snow surface, making more difficult snow depth measurement. Variations in snow melt intensity can also lead to an uneven snow surface. Therefore, when snow depth is manually measured, it has to be read on several points. When it is measured thanks to an acoustic or optical sensor, it can lead to the presence of spikes and unexplained variations in the data.

Furthermore, problems in precipitation measurements are reported. When the precipitation is liquid, rain gauge accuracy is quite good. However, when precipitation is solid, under-catch problems are frequent. Indeed, if the rain gauge is not heated, a snowcap or a bridge can form over the gauge orifice, thus reducing the amount of precipitation collected. Therefore, heated rain gauges have to be preferred, even if under-catch problems are not totally eliminated.

So, to reduce measurement problem, the observing site has to be carefully chosen. Besides, manual measurements are considered to be quite reliable, and so do automated measurements under certain conditions. For the instruments used at our observing site, measurement errors are looked more in details.

b) Errors in Precipitation Measurements :

Previously, the fact that rain gauges have to be heated has been underlined. However, under-measurements can still occur.

One of the most important problems encountered is related to wind. Indeed, it can deflect snow particles trajectory so that they are not catch by the gauge. The catch of total snowfall can be reduced by 50 to 80 % because of wind (DeWalle and Rango, 2008).

This problem can also be explained by the fact that heating the gauge causes some convection of the warmer surrounded air, which can prevent some snow crystals to be catch and may increase evaporative loss (DeWalle and Rango, 2008). Losses due to heating represent only a few percentage of measured precipitation, and losses due to convective currents are mainly expected for light snowfall during low air temperatures (Sevruk, 1983). Indeed such temperatures lead to smaller crystals, more affected by heating and blowing out.

Sevruk (1983) proposed a method to correct the series of measured precipitation which consists of using a conversion factor k (*Equation 4*). The precipitation obtained from the snowboard is supposed to be the actual precipitation, whereas the one measured by the rain gauge is the erroneous value. Using linear regressions, he tried to relate this factor k to the average wind speed on days with snowfall for a particular month, and to the portion N of snowfall which occurs on days with mean air temperature less than -8 °C (expressed in % of the precipitation gauge measurement). However, for heated rain gauges, no relationship have been found between k and the average wind speed, and only a weak correlation has been found to exist between k and N with r=0.45.

$$(4)\quad k = \frac{P_{sb}}{P_g}$$

With k : Conversion factor
P_g : Monthly or seasonal amount of precipitation recorded by the precipitation gauge
P_{sb} : Sum of new snow SWE obtained from measurements on a snowboard

Besides, such a multiplicative factor which compensates for the precipitation gauge under-catch has also been used by Rohrer and Braun (1994). However, they explained that it might be advantageous to use an additive correction term determined thanks to the comparison between the rain gauge precipitation and the water equivalent of new snow measured on a snowboard.

Actually, it is also possible to correct the data without using a conversion factor, that is to say by adding a term of error to the measured values (*Equation 5*). The term of error can be calculated thanks to some relationship found between its magnitude and meteorological parameters.

$$(5)\quad P_{corr} = P_{meas} + E$$

With P_{corr} : Corrected precipitation
P_{meas} : Measured precipitation
E : Heated rain gauge error

c) <u>Errors in Automated Snow Depth Measurements :</u>

We will focus on the errors which can occur when using an ultrasonic snow depth sensor as it is the one installed at the observing sites, and it appears that there are many different causes of errors when using such a sensor (Brazenec, 2005). The first one is related to how the sensor has been designed and installed. Indeed, if the sensor is not perpendicular to the target surface, or if this target is too small and reflects little sound, the measurement will be inaccurate. Besides, if the mounting structure is not rigidly installed, high winds can make it shake, so that the sensor is not able to accurately send or receive the signal.

Other problems can be due to climate factors. If the transducer is obstructed by ice or snow, the sensor can not work. It the wind is too strong, it may blow the echo out from under the sensor. Moreover, problems often occur during snowfall or when blowing snow is present. Instead of being reflected off the snow surface, the pulse may be reflected off the falling snowflakes, so that the signal is scattered and unable to provide accurate data. It has also been suggested that snowfall can attenuate the sound pulse.

However, the most important errors are related to the snow surface state. Depending on the surface structure, loose powder vs. hard packed crust for example, snow depth can be underestimated because the surface is a poor reflector of sound or because the signal has penetrated the snowpack. Finally, a rough or uneven snow surface can result in inaccurate measurements. Indeed, it implies that over the target area snow depth changes, and since the pulse is not always reflected off the same part of the area, the measured snow depth can be different between two consecutive measurements.

When using an ultrasonic snow depth sensor, it appears that measurement errors lead to both small and large amplitude variability in the sensor data (*Figure 2*). The first one is inherent in the data and can appear even with snow free conditions. The second one consists of occasional large data spikes which are easily detected.

Large amplitude variability is mainly caused by snow crystal type (i.e. snow surface structure), intense snowfall, presence of blowing/drifting snow and uneven snow surface. Besides, it especially occurs during snowfall events since most of the factors attributed to it are frequent during these periods. It can also be noted that effect of blowing/drifting snow can be minimized if the observing site is sheltered from wind (Brazenec, 2005).

Regarding small amplitude variability, it is mainly explained thanks to variations in air temperature. As said previously, an ultrasonic snow depth sensor sends a series of ultrasonic pulses toward an area on snow surface, measures the time required for it to do the round trip, and deduces the snow cover height thanks to wave velocity. Actually, wave velocity is not constant and principally depends on air temperature. So, measured air temperature is used to correct the ultrasonic wave velocity. However, this correction doesn't seem to be enough since small amplitude variability in snow depth can be related to air temperature variations. In fact, it has been reported that measured snow depth varies inversely with changes in air temperature, and that problems occur in the hottest part of the day (Bergman, 1989). This is probably due to an over-measurement from the air temperature thermistor located in the transceiver module, because of solar heating without the benefit of ventilation (Huwald et al., 2009). This over-measurement would lead to a higher calculated sound velocity, leading to a higher distance between the sensor and the snow surface, finally leading to a smaller measured snow depth. Besides, scattering in air temperature between the transceiver module and the snow surface can also affected snow depth sensor accuracy.

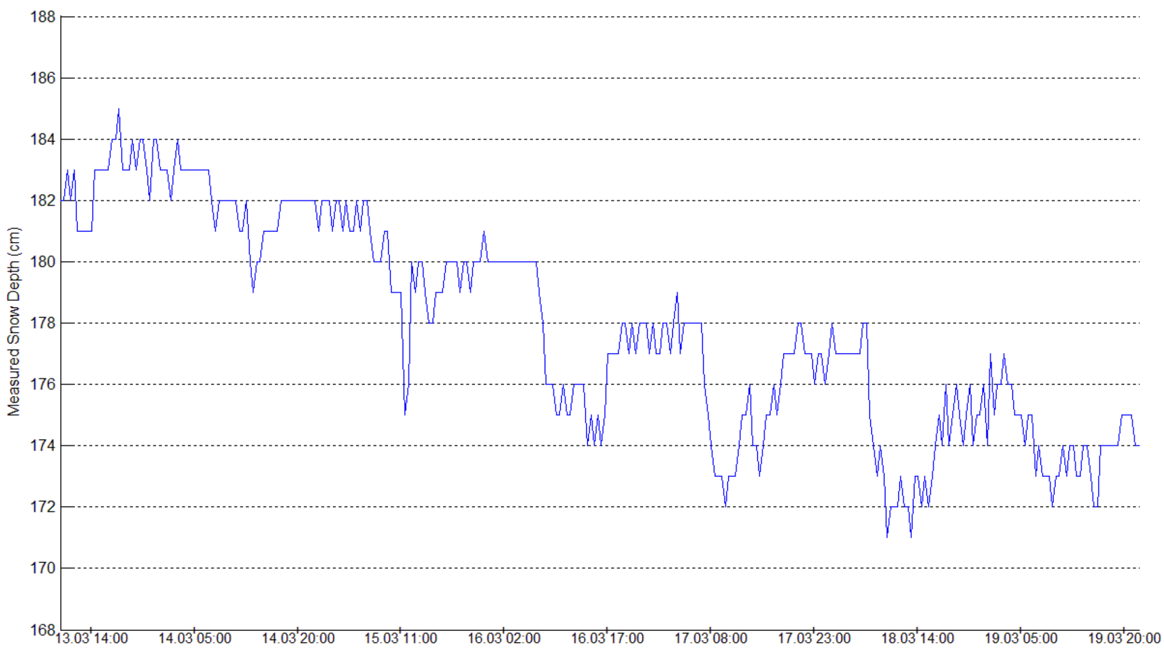

Figure 2 : Lago Pilone, Winter 2008/2009 : Zoom
Small Amplitude Variability in Measured Snow Depth

d) <u>Errors in Automated SWE Measurements (Snow Pillow) :</u>

The primary cause of errors is related to change in temperature. Indeed, when the temperature within the snowpack varies around 0 °C, melting and refreezing occur, leading to formation of ice- or crust layers. They act like a bridge, so that a part of the snow on the pillow is supported by the surrounded snow, thus reducing the pressure on the pillow and causing SWE underestimation. This phenomenon is known as snow bridging (Sorteberg et al., 2001).

Moreover, differences between the thermal properties of the pillow and those of the surrounding ground can lead to differences between the rate of snowmelt on the sensor and the one on the ground surface. If the rate of snowmelt on the sensor is the highest one, SWE under-measurement can occur (snow bridging). Conversely, if the rate of snowmelt on the ground is the highest one, the surrounded snow load is partially transferred onto the pillow, thus resulting in SWE over-measurements. Such errors can reach 40 %, even 200 % in the extreme.

Furthermore, it appears that snow pillow accuracy is greater for high SWE values. Sorteberg et al. (2001) reported deviation of ±50 % between SWE from manual observations and snow pillows in areas where the maximum SWE recorded is below 300 mm, whereas this deviation is only ±20 % in areas where the maximum SWE is above 300 mm. They also noted that there is a greater difference between SWE from the pillow and manual surveys during the melting period than during the mid-winter. However, it would rather be due to local variations in melting intensity, or to difficulties in taking good snow samples when snow is very wet. Finally, presence of blowing/drifting snow at the pillows can produce errors in measured SWE, first because it doesn't represent the actual snowpack SWE, and also because it can introduce variations which shouldn't occur.

Johnson and Marks (2004) proposed a method to detect periods during which snow pillows are supposed to give erroneous data, so that periods without measurement errors can be selected and new snow density will be studied only during these periods.

This method consists of observing simultaneously changes in snowpack SWE, snow depth and snowpack density. Snow pillow under-measurement occurs when a decrease in SWE is recorded while snow depth is constant or increasing. In this case, measured snowpack density decreases ($\rho = SWE/H$). Decreases in snowpack density are of course possible, for example when a new snow layer is added to the snow cover, nevertheless they should correspond to an increase in SWE. Likewise, over-measurement is generally characterised by an increase in SWE while snow depth decreases. Such changes can be explained by a high densification rate of the snow cover. However, too high densification rate are not conceivable.

Besides, since measurement errors are found when the base of the snowpack is at the melting point, another condition to detect them is that the snow temperature slightly above the soil must be superior to a threshold. This temperature is supposed to be a better indicator of snow cover thermal conditions than the temperature at snow-soil interface since it remains at or close to 0 °C for long periods. However, the snow temperature at a few centimeters above the ground is rarely available, so that it is approximated by a temperature index, which one is equal to the average air temperature for the prior 24 h. Furthermore, it is assumed that under- or over-measurements occur only when the snow cover density is superior to a threshold value, that is to say when it is enough high for snow to support shear stresses and bridge the sensor or transfer load to it. Table 1 summarises the parameters used to detect snow pillow under- and over-measurements.

Undermeasurement indicator	Overmeasurement indicator	Definition
$T_{index} > T_{thres}$	$T_{index} > T_{thres}$	T_{index} Snow cover temperature index (°C) T_{thres} Temperature index threshold
$\overline{\rho} > \overline{\rho}_{thres}$	$\overline{\rho} > \overline{\rho}_{thres}$	$\overline{\rho}$ Average snow cover density (kg m^{-3}) $\overline{\rho}_{thres}$ Average density threshold value
$\frac{dD_{swe}}{dt} < 0$	$\frac{dD_{swe}}{dt} > 0$	D_{SWE} SWE (mm of water) $\frac{dD_{swe}}{dt}$ SWE rate of change
$\frac{dD_{snow}}{dt} > 0$	$\frac{dD_{snow}}{dt} < 0$	D_{snow} Snow depth (mm) $\frac{dD_{snow}}{dt}$ Snow depth rate of change
$\frac{d\overline{\rho}}{dt} < 0$	$\frac{d\overline{\rho}}{dt} > \left(\frac{d\overline{\rho}}{dt}\right)_{thres}$	$\frac{d\overline{\rho}}{dt}$ Average density rate of change $\left(\frac{d\overline{\rho}}{dt}\right)_{thres}$ Average density rate of change threshold

Table 1 : Parameters used to detect snow pillow under- and over-measurements according to the method of Johnson and Marks (2004)

III. Snow Cover Modelling : SNOWPACK

Before estimating heated rain gauge under-catch thanks to SWE, data must be checked for consistency. Therefore, some manual measurements were done at our observing site, three times in a winter. It appears that the snow depth is quite well measured by the snow depth sensor. On the contrary, SWE and snowpack density from manual readings are much higher than the ones recorded, which would mean that the snow pillow under-measures snow weight (*Figures 4, 5, 6*). Such an evidence was already found for the season 2007/2008 at Lago Pilone (*Figure 3*, *Appendix E*). Hence, if the SWE data from the snow pillow don't represent the actual snow cover water equivalent, estimating heated rain gauge under-catch from this parameter would be incorrect. Moreover, for the winter 2008/2009, when comparing SWE data with those from the heated rain gauge, it appears that the SWE measured from the snow pillow and the cumulated precipitation are in good agreement (*Figure 5*). The discrepancy is mainly due to very intense snowfalls which occurred around 14/12/08. This would confirm that both SWE and precipitation are under-estimated, unless one demonstrates the manual samplings to be unreliable. So, a software able to simulate the snow cover evolution will be used, and if the comparison between our data and those from the simulation confirms the rain gauge under-catch, it would be possible to estimate its magnitude thanks to modelled data.

The software chosen is called SNOWPACK. After an extreme avalanche period on February 1999 in Switzerland, this software has been developed (Bartelt and Lehning, 2002; [1]Lehning et al., 2002; [2]Lehning et al., 2002). The aim was to provide avalanche forecasters information about the snow cover, when direct human observations are too dangerous or too time-consuming. Using meteorological data and some other parameters, this software is able to simulate the snow cover evolution, that is to say snow depth, SWE, and snowpack microstructure evolution throughout the winter. So, not only this software helps avalanche forecasters by providing them information on snow microstructure, but it is also useful to estimate the snow cover SWE, parameter directly related to the amount of precipitation.

III.1. Generalities about the Model

a) Numerical Model :

(Bartelt and Lehning, 2002; Lehning et al., 2002)

The model is a one-dimensional model, which assumes that motions parallel with the slope and lateral gradients (temperature and vapour pressure) are zero. The snow cover evolution is modelled thanks to four differential equations which govern mass (air and water), energy, and momemtum conservation. Phase changes like snow melting or refreezing are taken into account, and the boundary conditions at the snow-soil and at the snow-surface interfaces can be chosen (either Neumann or Dirichlet).

Besides, snow metamorphism routines, essentially governed by water vapour gradients, are used to determine the microstructure of each snow layer. Furthermore, snow viscosity and thermal conductivity formulations are linked with both macroscopic and microscopic snow properties (principally grain radius and temperature).

Regarding surface mass and energy exchanges, they are calculated thanks to existing or improved formulations which use air temperature, snow surface temperature and wind speed. Other properties like snow surface albedo or new snow density are determined thanks to statistical relationships derived from measurements at some sites in the Swiss Alps.

Furthermore, the amount of fresh snow is determined either thanks to snow depth evolution or to measured precipitation data and calculated new snow density. In the second case, new snowfall water equivalent (new SWE) is equal to the amount of precipitation, whereas in the first case it is deduced thanks to new snow density. Besides, it has to be noted that when snow depth evolution is used to drive the model, the new snow height is calculated as the difference between measured snow depth and snow depth expected by the model if no snow event was occurred.

Examples of SNOWPACK interfaces are shown in Appendix B.

b) Remarks about the Model Outputs and Model Sensitivity :

First, it appears that there are problems when precipitation data are used as input. The most important one is that modelled snow depth is always underestimated, even after the precipitation data have been corrected. It may be due an inappropriate threshold air temperature used to determine whether if precipitation is solid or liquid (Rasmus et al., 2007).

On the other hand, when measured snow depth is used to run the model, snow depth is quite well simulated for most of the simulations. However, this observation is fallacious since measured snow depth is used to determine increases in modelled snow depth. Furthermore, discrepancies are more important during melting periods where snow depth is often overestimated. This may be due to an underestimation of snow surface heat fluxes, so that the melting is too slow (Rasmus et al., 2007). Another reason may be due to the fact that the snowpack settling in the model is too fast, i.e. snowpack layers densify too quickly, so that too much mass is added to the snowpack (Bartelt and Lehning, 2002; Rasmus et al., 2007). This would lead to too high modelled SWE values, which is frequently observed.

Moreover, the agreement between measured and modelled snowpack structure varies. Actually, only a few comparisons have been done. Besides, it is possible that an observer and the model don't have the same description of the grain types, and this description can also be different from one observer to another.

Finally, it has been reported that manual observations and modelled data are more likely to show good agreement when snow depth is quite homogeneous, snow is relatively soft, and weather is cold and dry. On the contrary, when the snow is isothermal, dense, hard and wet or melting, and when snow depth is not homogeneous, the agreement is generally poor (Rasmus et al., 2007).

Regarding the model sensitivity, it has been studied by Rasmus et al. (2007). One by one, several inputs have been modified, and the consequences on the model outputs have been observed. Modelled SWE and snow depth seem to be the most sensitive quantities, followed by snowpack structural characteristics (temperature, grain form and size) and snow cover evolution characteristics (melt, duration, wetting).

Furthermore, the inputs whose modification leads to the most important changes include snow depth, boundary conditions and radiation parameterisation. On the contrary, using a constant albedo or changing relative humidity, ground temperature, new snow parameterisation, or residual water content products minor effects on the studied quantities.

Finally, it has been found that the model sensitivity becomes more important towards the melting period, and so does the choice of the input data and boundary conditions.

III.2. SNOWPACK Input Parameters

a) Required and Facultative Inputs :

To use SNOWPACK, different parameters have to be provided. Some parameters are absolutely required, others are facultative (*Table 2*). The required ones are : air temperature, relative humidity and wind speed. Moreover, at least incoming or reflected SW radiation has to be provided. Likewise, at least precipitation or snow depth is required. The facultative parameters are : snow surface temperature, incoming LW radiation, snow-soil interface temperature (called bottom temperature) and wind direction. Facultative unavailable parameters must be set to zero, for the model to understand that they can not be used.

Compulsory Parameters	
Air temperature	
Relative humidity	
Wind speed	
Incoming and/or Reflected shortwave radiation	
Precipitation rate and/or Snow depth	
Facultative Parameters	
Parameter	*What SNOWPACK does if the parameter is not available*
Snow surface temperature	Deduced thanks to the differential equations governing snowpack mass, energy and momentum conservation
Incoming longwave radiation	Estimated thanks to air temperature and relative humidity
Snow-soil interface temperature	Assumed to be equal to -0.1 °C
Wind direction	?

Table 2 : Required and Facultative SNOWPACK Input Data

b) <u>When an Input Parameter is not Provided :</u>

Without air temperature, relative humidity or wind speed SNOWPACK can not work. Indeed those meteorological are often used during the simulation. They are used to calculate new snow density, to determine sensible and latent heat fluxes at the snow surface, and to estimate snow surface albedo if necessary. They are also used to detect rain events and calculate the amount of energy they add to the snowpack.

Besides, it is possible to use either incoming or reflected SW radiation or both. When both parameters are available, snow surface albedo is directly calculated, but when only one of these parameters is provided, the statistical model is used to determine snow surface albedo so that the second parameter can be deduced.

Moreover, SNOWPACK can use measured precipitation and/or snow depth. If snow depth is not provided, the amount of new snow is determined thanks to precipitation (when air temperature is below 1.2 °C) and modelled new snow density. Regarding the amount of rain, it is determined thanks to precipitation (when air temperature is greater than 1.2 °C). If precipitation data are not provided, rain never occurs in the model, but the amount of new snow can still be known thanks to snow depth evolution. When both precipitation and snow depth are available, the first one is used to determine the amount of rain and the second one to determine the amount of new snow.

Regarding snow surface temperature, if it is not available, it is considered as an unknown parameter and is determined by solving the differential equations governing snowpack mass, energy and momentum conservation. For incoming LW radiation, when it is not measured, it is estimated thanks to air temperature and relative humidity. Besides, if bottom temperature is not available, it is set to -0.1 °C. Finally, regarding wind direction, the use done by SNOWPACK is not clearly understood. But we can think that if it is not provided, it is simply not used.

c) <u>Problem of Missing Data :</u>

As said before SNOWPACK requires several measured parameters to work, but unfortunately measurement instruments can have problems, and some data may be missing. Nevertheless, when replacing the missing data by -99.9, SNOWPACK can be run and keeps on doing the computation. Depending on the parameter, a linear interpolation is done, the last measured data is used, or -99.9 is used as a valid data (*Table 3*).

Parameter	*What SNOWPACK does when some data are missing*
Air temperature	Linear Interpolation
Relative humidity	Linear Interpolation
Wind speed	Last measured value used
Incoming and Reflected shortwave radiation	Linear Interpolation
Incoming longwave radiation	Linear Interpolation
Snow surface temperature	Linear Interpolation
Bottom temperature	Linear Interpolation
Precipitation	-99.9 used as a valid value
Snow depth	Linear Interpolation

Table 3 : What SNOWPACK does when some Data are Missing

d) SNOWAPCK Initialisation :

Before doing a simulation, some site characteristics must be precised, which ones include the site latitude, longitude, and altitude, its slope angle, and the canopy height. An example of the interface used to initialise the model is displayed on Figure 1, Appendix B.

Regarding the snow cover, if there is snow on the ground at the date at which the simulation begins, all its characteristics must be given. This means that for each snow layer, we have to know its date of formation, its temperature, its volumetric water, ice and void content, but also the radius, the sphericity and the dendricity of the grains.

Furthermore, one or more soil layers can be set. In this case, the density, the thermal conductivity and the specific heat of each soil layer must be defined. However, those parameters are required only if a Neumann boundary condition is used at the snow-soil interface, in order to calculate the heat flux through the ground. When no soil data is provided, a Dirichlet boundary condition is automatically used, which means that the snow-soil interface temperature is used as a boundary condition.

e) Other Operating Parameters :

After having done the initialisation, other parameters have to be chosen. An example of the interface used for this step is displayed on Figure 2, Appendix B.

As said before, the boundary condition at the snow-surface interface can be chosen, i.e. Neumann or Dirichlet. If the first one is chosen, the model automatically switches to a Dirichlet boundary condition when the snow surface temperature is close to the melting point : a threshold temperature is used and can be modified by the user. Likewise, the boundary condition at the snow-soil interface has to be selected if some soil layers have been described. In this case, it is possible to adjust the value of the geothermal heat flux.

Furthermore, it is possible to choose which stratification of the atmospheric layer will be assumed for the determination of the snow surface mass and energy exchanges. For the same reason, the model needs the height of the wind sensor, the height at which other meteorological data are measured, and the roughness length over snow. Besides, the model can test for blowing snow conditions (real erosion). Otherwise, only virtual erosion will be performed. A canopy module, which takes into account the effect of the surrounded vegetation, is also available.

Finally, the computation step length has to be chosen. Its value depends on the data measurement time step and the accuracy wanted. Typically, if the data are measured every 30 min, the computation step length is 15 min.

IV. Results

IV.1. Choice of the Input Parameters

Before doing a simulation, model inputs must be chosen. However, this choice depends on the parameters measured at the observing site but also on the data which are missing. The following work was also done for the season 2007/2008 (*Appendix E*).

a) Measured Parameters at Lago Pilone during the Winter 2008/2009 :

Table 4 shows the parameters measured at the site Lago Pilone during the winter 2008/2009. They include a heated rain gauge and the standard sensors for air temperature, relative humidity, and wind speed. Moreover, as one of the objectives of this project is the determination of new snow density, a snow depth sensor and a snow pillow are present. Other instruments are also installed in order to measure incoming and reflected SW radiations, snow surface temperature, snow temperature at the base of the snowpack, and snow temperature at 30, 60 and 90 cm from the ground surface.

Time Step	Measured Parameters
10 min	Precipitation Wind speed Snow surface temperature Incoming and Reflected SW radiations
30 min	Snow depth Air temperature Relative humidity
4 h	Bottom temperature Temperatures at 30, 60 and 90 cm from the ground

Table 4 : Lago Pilone, Winter 2008/2009
Measured Parameters and their Frequency of Measurement

First it can be noted that the required data previously listed (III.2.a)) are all measured at our observing site, and so do some facultative parameters. Moreover, it appears that all the measured parameters are not recorded with the same time step, so that the data have to be grouped together in a single time step. Using a time step of 4 h would lead to approximations which can be avoided by using a smaller time step. Moreover, using a time step as small as 10 min doesn't seem of interest. Therefore, a time step of 30 min has been chosen. This time step seems to be commonly used (Bartelt and Lehning, 2002) and is also the model default parameter. The rain rate is measured every 10 min, so to have the value for a 30-min time step, a sum is done. For wind speed, solar radiations, and snow surface temperature, a mean is used. Regarding bottom temperature and snow temperature at 30, 60 and 90 cm from the ground surface, which are measured every 4 hours, their value is considered to be constant during the prior 4 hours. This approximation seems correct as those parameters are quite constant during the whole winter, that is to say almost unmodified during a period of a few hours long.

b) <u>Missing Data :</u>

As said before, for the model to be run, missing data can be set to -99.9. Letting their value to -99.9 is certainly the best solution if only a few data are missing. However, if a series of data is missing, doing a linear interpolation or considering the data as constant is inappropriate and another solution has to be found. So, to determine whether missing data can be replaced by -99.9 or not, a threshold number of consecutive missing data has to be chosen. As input data are provided with a 30-min step time, the threshold of 10 values has been chosen, which represents a period of 5 h. Figure 3 represents, for each measured parameter, the repartition of the missing data for the winter

2008/2009 at Lago Pilone. The total number of missing data and the number of missing data which can be replaced by -99.9 are shown in Table 5.

No data are missing for wind direction and incoming long-wave radiation as they are not measured and set to zero. Besides, no data is missing for air temperature and relative humidity, and except for snow surface temperature and snow depth, all other missing data can be replaced by -99.9 without introducing important errors. Furthermore, the region of Piedmont has precised that getting back the missing data is not possible.

For snow surface temperature, missing data which can not be replaced by -99.9 represent 4 periods which last 14 hours, 17 hours, 3.5 days and 20 days. Even if it is not very accurate, it is conceivable to interpolate snow surface temperature for the periods smaller than one day, even for the period of 3.5 days. However, it is really impossible to do that for the period of 20 days. Another solution could be to stop SNOWPACK, and to restart it without using snow surface temperature, which means that this temperature would be calculated by the model. At the end of the missing period, SNOWPACK can be stopped again and restarted with snow-surface temperature set as an available parameter. However, SNOWPACK has to be initialized before each run. This means that information about the composition of the each snow layer at the beginning of the simulation must be manually given. But the number of snow layers can be greater than a hundred, and providing all the data manually seems to be a waste of time.

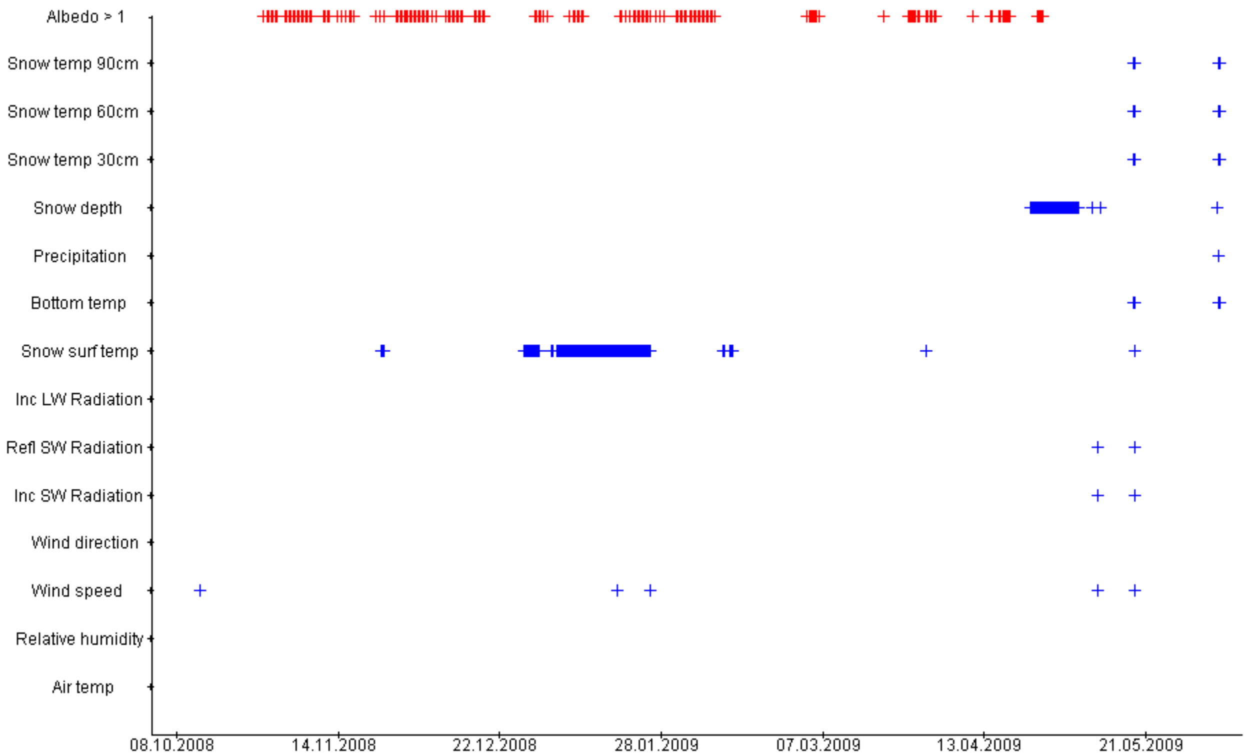

Figure 3 : Lago Pilone, Winter 2008/2009
Repartition of Missing Data between 08/10/08 and 08/06/09, Albedo Problem

Parameter	Total Number of Missing Data	Number of Missing Data which can be Replaced by -99.9
Air temperature	0	0
Relative humidity	0	0
Wind speed	7	7
Wind direction	0	0
Incoming SW radiation	2	2
Reflected SW radiation	2	2
Incoming LW radiation	0	0
Snow surface temperature	**1297**	18
Bottom temperature	15	15
Precipitation	1	1
Snow depth	**541**	3
Temperature at 30 cm from the ground	15	15
Temperature at 60 cm from the ground	15	15
Temperature at 90 cm from the ground	15	15

Table 5 : Lago Pilone, Winter 2008/2009
Number of Missing Data between 08/10/08 and 08/06/09

Regarding snow depth, missing data which can not be replaced by -99.9 represent a unique period which lasts around 11 days, at the end of April 2009, that is to say at the end of the winter when snow events are supposed to be sparse. If those missing data are replaced by -99.9, it is obvious that it will introduce errors in the simulation but actually, the simulation is not affected for the prior months, that is to say for the period we are interested in. Besides, if the simulation is stopped before a period of missing data, and restarted with precipitation data used to drive the model instead of snow depth data, there is still the same problem with SNOWPACK initialisation. Therefore, missing snow depth data are set to -99.9 and the model will do a linear interpolation.

c) Aberrations :

When observing measured data, some aberrations are found, and removing them before doing a simulation may be necessary. Nevertheless, all measured data are not checked in details. For example, too fast changes in temperatures or relative humidity can occur. However, correcting this kind of error is not easy, so that they are not taken into account. In fact, only the most important aberrations will be removed, that is to say those which are quite obvious. For the winter 2008/2009 at Lago Pilone, they concern three parameters : snow depth, incoming and reflected SW radiations.

Regarding measured snow depth, negative values are sometimes recorded at the beginning and at the end of the winter. Those values are simply replaced by zero.

Besides, it is also possible for measured reflected SW radiation to be greater than the incoming one, which leads to snow surface albedo values superior to 1 (*Figure 3*). As it is physically impossible, measured SW radiation data are deleted when this problem occurs, and considered to be missing data (i.e set to -99.9). In fact, when doing two simulations, one with those data set to -99.9 and another one with the original values, the percentages of change in modelled SWE and snow depth are smaller than 1 %, which means that those measurement problems are not very important.

d) Chosen Inputs :

The required parameters, like air temperature, relative humidity and wind speed, are given as input. Moreover, as the objective is to determine heated rain gauge under-catch, measured precipitation can not be used to drive the model and snow depth must be used to determine the amount of new snow. Besides, the effects of rain on snow cover characteristics are not negligible, therefore it is chosen to provide both measured precipitation and snow depth. Furthermore, when doing a simulation with only reflected or incoming SW radiation used, it clearly appears that snow depth evolution is not well simulated, and that using both incoming and reflected SW radiation is better (*Figure 3, Appendix B*). So, this last solution is chosen.

Regarding facultative parameters, incoming LW radiations and wind direction are not measured, therefore they are set to zero. Besides, snow-soil interface temperature is measured so that it has been chosen to use it. Actually, its value is always -0.1/-0.2 °C during the mid winter, and if it were not set as available, it would be assumed to be equal to -0.1 °C. This reveals that the availability of this parameter is not very important. About snow surface temperature, to make up for the problem of missing data, the model will be run without using this parameter. Moreover, simulations have been done with the data from the winter 2007/2008 at Lago Pilone during which there is no problem of missing snow surface temperature data (*Figure 4, Appendix B*). When this parameter is provided, it appears that until the beginning of the melt period, snow depth is better modelled when using a Neumann boundary condition. Besides, it appears that for such a boundary condition, having the snow surface temperature or not leads to very small changes in modelled SWE and snow depth (less than 2 %). It implies that obtaining modelled data with a relatively good accuracy is possible even without snow surface temperature. Table 6 summarises the parameters used to do the simulation for the winter 2008/2009 at Lago Pilone, more details can be found on Figures 1 and 2, Appendix B.

e) SNOWPACK Initialisation :

Site characteristics which are unknown are assumed to be equal to the default values. Latitude, longitude and altitude are known. As the site is quite flat, it is assumed that the slope angle is zero. Moreover there is no vegetation right nearby the observing site, therefore the canopy height and the leaf area index are set to 0.0, and the direct throughfall is set to 1.0. Regarding the snow cover, if there is snow on the ground at the date on which the simulation begins, all its characteristics must be given. Actually, having all these parameters is not possible. That is why it has been decided to start the simulation when there is no snow on the ground (i.e. 08/10/08). Furthermore, one or more soil layers can be set, but soil composition and soil properties are actually unknown, so that no soil layer is described.

f) Other Operating Parameters :

After having initialised the model and chosen which measured data will be used, other choices have to be made.

First, it appears that when the snow-surface temperature is not available, modelled SWE and snow depth are the same whatever the snow-surface interface boundary condition is. Indeed, in both cases, the snow-surface temperature used is the one calculated thanks to the differential equations, so that using a Neumann or Dirichlet boundary condition leads to the same outputs. However, since a choice has to be made, a Neumann boundary condition is arbitrarily chosen. The threshold temperature required to switch to a Dirichlet boundary condition has no consequences in this case and is therefore let to its default value.

Since no soil layer is described, the Dirichlet boundary condition at the snow-soil interface is automatically selected and the geothermal heat flux can not be modified.

Regarding the stratification of the atmospheric layer, because of a lack of knowledge in this field, the default value is used. Besides, the height of the wind sensor is known (6.8 m). For the height at which other meteorological data are measured, the height of the air temperature sensor is chosen as it is one of the most useful parameter (5.9 m).

Moreover, it is recommended not to use the blowing snow module when no snow depth data are available, but since snow depth data are provided it has been chosen to use this module. On the contrary, the canopy module will not be used. Indeed, it is of no interest at a site without canopy.

Finally, since the meteorological data are measured every 30 min, a calculation step length of 15 min was chosen.

<table>
<tr><td colspan="2">Measured Parameters Used for the Simulation</td><td>Air Temperature
Relative Humidity
Wind Speed
Incoming and Reflected SW Radiations
Bottom Temperature
Precipitation per Meteo Step
Snow Depth</td></tr>
<tr><td rowspan="2">Time Step</td><td>Measurement</td><td>30 min</td></tr>
<tr><td>Calculation</td><td>15 min</td></tr>
<tr><td rowspan="4">Initialisation</td><td>Slope Angle</td><td>0</td></tr>
<tr><td>Canopy Height</td><td>0</td></tr>
<tr><td>Number of Snow Layers</td><td>0</td></tr>
<tr><td>Number of Soil Layers</td><td>0</td></tr>
<tr><td colspan="2">Determination of New Snow Amounts</td><td>Thanks to Snow Depth Evolution</td></tr>
<tr><td colspan="2">Boundary Condition Used at the Snow-Surface Interface</td><td>Neumann (Choice without consequences)</td></tr>
<tr><td colspan="2">Blowing Snow Module</td><td>ON</td></tr>
<tr><td colspan="2">Canopy Module</td><td>OFF</td></tr>
</table>

Table 6 : Input Data and Parameters Used to Run the Model

IV.2. Comparison between Measured, Modelled, and Manual Data

a) <u>Snow Depth :</u>

Model results are reported in Figure 4. By looking at the graph, it emerges that during the accumulation periods, the modelled curve merges with the measured one since this last one is used as a reference to determine new snowfall. On the contrary, some discrepancies are observed during melting periods. This condition is very frequent, as said in part III.1.b), and prevents the model from detecting some new snow events. Moreover, during the period of missing data, measured snow depth is assumed to decrease linearly, so that new snowfall can not be detected by the model. Because of this, modelled SWE is maybe a bit smaller than it should have been, but as said in part IV.1.b), since this problem occurs at the end of the winter, the modelled data for the prior months are unaffected.

Regarding manual measurements, they show a quite good agreement with modelled and measured snow depths, even if manual data show a small under-measurement. This can be due to variations in snow cover height, leading to differences between data from the sensor and from manual observations. Besides, it has to be noted that for a specific date of observation, the three readings give three distinct values. These under-measurements might also be due to snow compaction during snow cores sampling.

b) <u>SWE :</u>

Regarding the snow cover water equivalent, the modelled one is much higher than the measured one (around 2.5 times higher), which would support the assumption of both a snow pillow and heated rain gauge under-measurement (*Figure 5*). Indeed, even if it has been reported that SNOWPACK often overestimates SWE, such a difference can not be due to this fact only. Moreover, the measured and modelled curves have more or less the same shape, which implies that even if they have a different magnitude, changes in SWE are quite well modelled. Nevertheless, the new snow events, which are not detected because of modelled snow depth evolution, are therefore not present in modelled SWE. Besides, the difference between the two parameters is detected right from the beginning of the simulation and is almost constant during the whole winter.

Furthermore, it appears that not only modelled SWE is much higher than measured SWE, but it is also higher than manual values. Nevertheless, in relation to those manual SWE, the difference is quite reasonable, and might be attributed to the model systematic overestimation.

Figure 4 : Lago Pilone, Winter 2008/2009
Comparison between Measured and Modelled Snow Depth

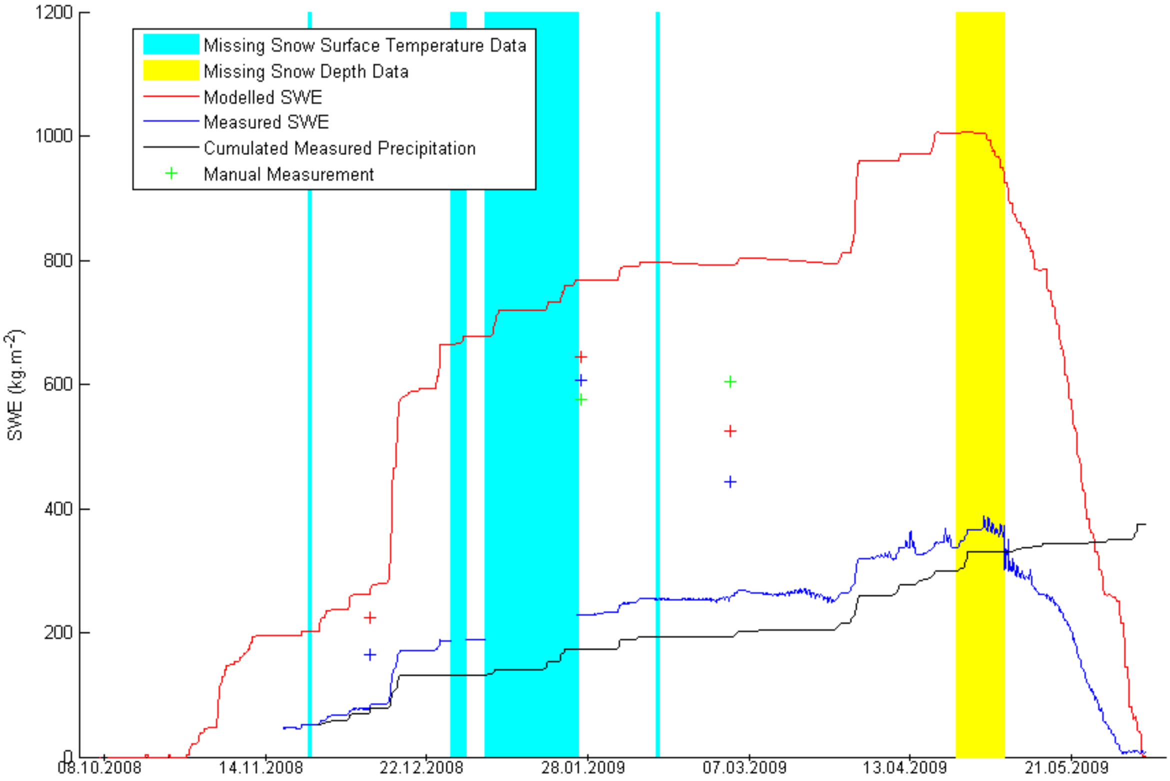

Figure 5 : Lago Pilone, Winter 2008/2009
Comparison between Measured and Modelled SWE

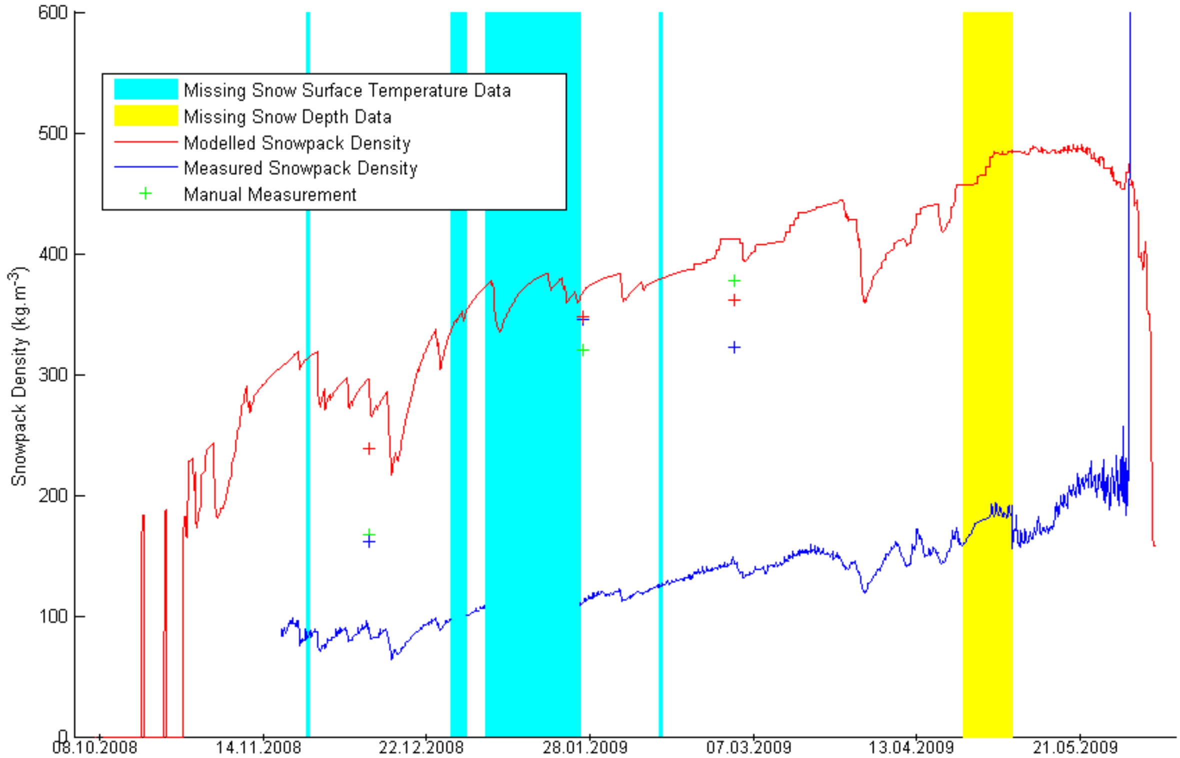

Figure 6 : Lago Pilone, Winter 2008/2009
Comparison between Measured and Modelled Snowpack Density

c) Snowpack Density :

First it has to be precised that modelled snowpack density is calculated thanks to modelled SWE and snow depth. Besides, remarks about the snow cover density are quite similar to those about SWE : measured and modelled snowpack density have the same evolution but the one modelled is around three times higher than the one measured (*Figure 6*). This kind of result was expected since for a specific snow depth, snowpack density is directly proportional to snowpack SWE ($\rho = SWE/H$).

Besides, the agreement between manual measurements and modelled snowpack density is good. Indeed, modelled SWE is higher than the one from manual observations, but so does modelled snow depth, so that snowpack densities are almost the same. The discrepancy is more important for the first observation of the winter : modelled SWE is still higher than the one from the readings but snow depth readings are much closer to the measured data.

d) Remarks about the Last Manual Observation :

Regarding SWE, the last manual observation (in March) must be looked more in details. Indeed, the two first readings follow quite well modelled SWE, whereas the last one shows a decrease in the snow cover SWE which doesn't occur according to the model. Besides, this decrease doesn't appear in the data from the snow pillow. But it can be noted that the higher value of the third sounding is almost equal to the smaller value of the second sounding, and even slightly superior. Such an observation is in good agreement with evolution of both modelled and measured SWE. So, this unexpected decrease in SWE showed by manual data might also be explained by difficulties in snow sampling because of variation in snow cover distribution.

IV.3. Comparison between Measured and Modelled Precipitation

So, the simulation done with SNOWPACK confirms that SWE and precipitation are really underestimated by the measurement instruments. On these premises, it is now possible to focus on the heated rain gauge error. But before studying its magnitude, we have to look more in details the way the model detects precipitation events. On this point, even if it has been said that changes in SWE are quite well modelled, several problems are observed when comparing the measured precipitation with the one from the model (*Figure 7*). The most evident is that measured and modelled new SWE have quite different values. Moreover, it appears that the model records new SWE whereas no precipitation is measured. In the following of this study, the data referred as input data correspond to those provided to the model, whereas the output data correspond to those modelled by SNOWPACK. Moreover, since there are still some doubts on the manual measurements accuracy, they will not be used as a reference. Therefore, even if modelled SWE is certainly over-estimated, we assumed that each value of modelled new SWE represents the actual amount of precipitation.

Figure 7 : Lago Pilone, Winter 2008/2009
Comparison between Measured and Modelled Precipitation

a) <u>Rain with Air Temperature below 1.2 °C :</u>

Previously, it has been said that the amount of new snow is determined thanks to snow depth evolution, and the amount of rain thanks to input precipitation. It would mean that precipitation data are used only when air temperature is greater than 1.2 °C, and that when air temperature is below 1.2 °C, the model doesn't take into account the input precipitation. To verify this assumption, measured precipitation data are replaced by filtered data, i.e. precipitation values are set to 0 when air temperature is below 1.2 °C. If the assumption is correct, doing a simulation with the filtered precipitation should lead to the same outputs as when the original input precipitation is used.

In fact, by doing so, even if they are small (less than 0.4 %), changes in snow depth and SWE are found. This is explained by the fact that when using the original precipitation, 7 output rain events are found with air temperature below 1.2 °C, events which have been removed by the filtering. Air temperatures range from 0.4 to 1.1 °C when such events occur. More details can be found in Table 7. The filtered is applied again with a threshold temperature of 0.4 °C, so that the prior events are not preserved. It appears that modelled snow depth and SWE are exactly the same as the ones obtained thanks to the original precipitation. Moreover, it can be noted that the amount of rain measured and the one detected by SNOWPACK are exactly the same, except for the prior events where the modelled rain rate is twice smaller than the measured value.

No explanation has been found but since there are only a few events, it seems correct to say that when air temperature is below 1.2 °C, measured precipitation is useless. It implies that the measured precipitation doesn't influence the modelled new SWE, that is to say the study of the heated rain gauge error. This fact is confirmed by the following observation, that is to say the presence of modelled new snow events even if no precipitation is the measured by the rain gauge.

Modelled Rain Events with Air Temperature below 1,2 °C							
Date	01/11/08	02/11/08	07/11/08	08/11/08	09/11/08	28/03/09	06/06/09
Hour	13.30	10.00	14.30	15.00	15.30	13.30	15.30
Air Temperature (°C)	1,1	1,1	0,9	0,4	0,9	1,1	0,8
Measured Precipitation (mm)	0,8	0,6	0,2	0,4	0,6	0,4	2,8
Modelled Rain (mm)	0,4	0,3	0,1	0,2	0,3	0,2	1,4
Modelled SWE from New Snow (mm)	0,0	0,0	0,0	0,0	0,0	0,0	0,0
Wind Speed ($m.s^{-1}$)	2,7	4,2	0,0	0,3	0,9	1,6	3,1
Relative Humidity (%)	98	100	85	68	76	80	96

Table 7 : Lago Pilone, Winter 2008/2009
Modelled Rain Events with Air Temperature below 1.2 °C : Details

b) Snow Events Detected by the Model while No Precipitation is Measured :

As just said, it appears that even if there is no precipitation measured by the heated rain gauge, new snow can be detected by SNOWPACK. Actually, snow depth data are used to drive the model. So, an increase in snow depth can be interpreted as a new snow event. However, we can wonder why the heated rain gauge doesn't record any precipitation. Before the 22/11/08, the rain gauge was not heated. Therefore, to eliminate under-catch problems due to this fact, events which occur before this date are not taken into account.

Such a problem occurs 83 times during the winter 2008/2009. A first explanation for this problem could be found thanks to the rain gauge sensibility which is equal to 0.2 mm. So, if the new SWE is smaller than 0.2 mm, a snow event occurs but no precipitation is recorded. However, it appears that when the problem occurs, the new SWE determined by SNOWPACK is always higher than 1 mm and can even reach 15.5 mm. So, the problem is not due to the rain gauge sensibility.

However, it is known that wind can reduce the catch of total snowfall. So, if the wind is strong, it is possible that the heated rain gauge doesn't record anything. Among the events where new SWE is detected by the model but no precipitation is measured, the ones with wind speed superior to 4 $m.s^{-1}$ represent 29 % of the events. For those events, wind speed is between 4 $m.s^{-1}$ and 8 $m.s^{-1}$, and the average value is 5.9 $m.s^{-1}$ (std dev 1.3 $m.s^{-1}$).

Furthermore, in order to melt solid precipitation, the rain gauge is heated. A resistance is used but it is not directly on contact with the gauge. Actually, the surrounded air is heated, which can enhance the evaporation loss and especially the phenomenon of blowing snow out due to convective currents. Besides, when air temperature is very low, it is possible that the surrounded air is not enough heated, and that the solid precipitation is not totally melted, which can lead to a partial or total obstruction of the heated rain gauge orifice and cause under-catch problems. The rain gauge constructor indicates that the instrument works for temperatures superior to -30 °C (http://www.cae.it/it/stazioni.php), but the region of Piedmont has reported that dysfunctions appeared for air temperatures around -10 °C. So, it is assumed that problems due to low air temperature can occur when air temperature is below -5 °C. The events selected in this way represent 73 % of the total events, for which air temperature is between -14.5 and -5.2 °C, with an average value of -8.7 °C (std dev 2.4 °C).

In summary, 8 % of the events are only caused by wind speed superior to 4 $m.s^{-1}$, 53 % are only caused by air temperature below -5 °C, and 20 % are caused by both strong wind and low air temperature (*Table 8*). So, when the model detects a snow event but the heated rain gauge doesn't record any precipitation, 81 % of the events can be explained by high wind speed, low air temperature or both. Besides, it seems that errors are more likely to occur because of low air temperatures, and that they are more important when air temperature and wind speed effects are combined (*Figures 1 and 2, Appendix C*).

Thresholds values	
Wind speed	4 m.s-1
Air temperature	-5 ℃
Events explained	
By high wind speed only	8 %
By low air temperature only	53 %
By both high wind speed and low air temperature	20 %
Total	81 %

Table 8 : Lago Pilone, Winter 2008/2009
Modelled Snow Events while No Precipitation is Measured : Explanation

IV.4. Heated Rain Gauge Error

a) <u>When No Precipitation is Measured by the Heated Rain Gauge :</u>

For the prior events, the relationship between wind speed/air temperature and the magnitude of the measurement error can be studied (*Figures 3 and 4, Appendix C*). The relation proposed by Sevruk (1983) can not be used as the measured precipitation is zero. Therefore the difference between modelled and measured precipitation, referred as heated rain gauge error, is used. Linear regressions lead to weak correlations between the magnitude of the error and wind speed or air temperature, and so does the multiple regression. Besides, when using a logarithmic regression, the correlations with wind speed or air temperature are still weak (r^2=0.28 and r^2=0.02), but when doing a multiple regression, a quite good relationship is found with r^2=0.43 (*Equation 6*). The p-value for each coefficient, which ones are all significant at a 5 % level, can be found in Table 1, Appendix C.

$$(6)\ \ln(E) = 0.136\ w + 0.049\ T + 0.890$$

With E : Heated rain gauge error, i.e. difference between modelled and measured precipitation (mm)
w : Wind speed ($m.s^{-1}$)
T : Air temperature (°C)

Nevertheless, when no precipitation is measured, knowing if precipitation has occurred or not is rather difficult. Therefore, the transposition of the prior relation to less instrumental sites (i.e. only rain gauge) is not straightforward.

b) <u>When Some Precipitation is Measured by the Heated Rain Gauge :</u>

When precipitation is measured, correcting the data on the basis of meteorological parameters is possible. Therefore, the measurement error can be confronted to wind speed and air temperature (*Figures 5 and 6, Appendix C*). If a relationship is found, the measurement error could be estimated, and the measured data could be corrected thanks to equation (5). Besides, as previously, the precipitation events which occur before the rain gauge is heated are not taken into account.

First, linear regressions were done, but it appears that correlations are better when using logarithmic regressions. One negative value, causing problem with the logarithmic function, was removed. Only weak correlations are found with air temperature (r^2<0.16), whereas the relationship between the logarithm of the error and wind speed is quite good (r^2=0.50). Besides, it is even better doing a multiple regression with both wind speed and air temperature (r^2=0.56). The regression can also be done with other meteorological parameters like relative humidity. When doing a regression with this parameter only, no result is found. Besides, multiple regressions which use air temperature and relative humidity at the same time lead to coefficients of correlation a bit higher, but also to not significant p-values. Actually, it is due to a problem of multi-collinearity since a good linear relationship can be found between these two parameters (air temperature and relative humidity). Nevertheless, doing a multiple regression with relative humidity and wind speed leads a better correlation than when using wind speed only (r^2= 0.52). Finally, the multiple regression between wind speed, air temperature and the logarithm of the error is the one retained since it shows the higher r^2 value and the smaller p-values (*Equation 7*). The p-value for each coefficient can be found in Table 2, Appendix C.

$$(7)\quad \ln(E) = 0.231\, w \ +\ 0.056\, T +\ 0.463$$

With	E :	Heated rain gauge error, i.e. difference between modelled and measured precipitation (mm)
	w :	Wind speed ($m.s^{-1}$)
	T :	Air temperature (°C)

c) <u>Data Correction :</u>

Since quite good relations were found, correcting the precipitation data is possible. The snow events, which are modelled and also measured, are corrected using equations (5) and (7), while the ones which are modelled but not measured are corrected using equations (5) and (6). Modelled data, which are assumed to represent the actual precipitation, and corrected data are confronted on Figure 8. Figure 7, Appendix C shows the measured, modelled and corrected precipitations during the whole winter. It has to be underlined that the snow events which occur before the rain gauge is heated (20/11/08) can not be corrected since they are not taken into account in the correlations.

By looking at the graphs, it appears that for the events with modelled precipitation superior to 8 mm, the corrected precipitation is quite smaller than the modelled one, but it is also much higher the measured one (*Figure 7, Appendix C*). Actually, those events correspond to the very intense snowfalls which occurred around 14/12/08 and led to the discrepancy between data from the rain gauge and data from the snow pillow. Otherwise, the precipitation is quite well corrected.

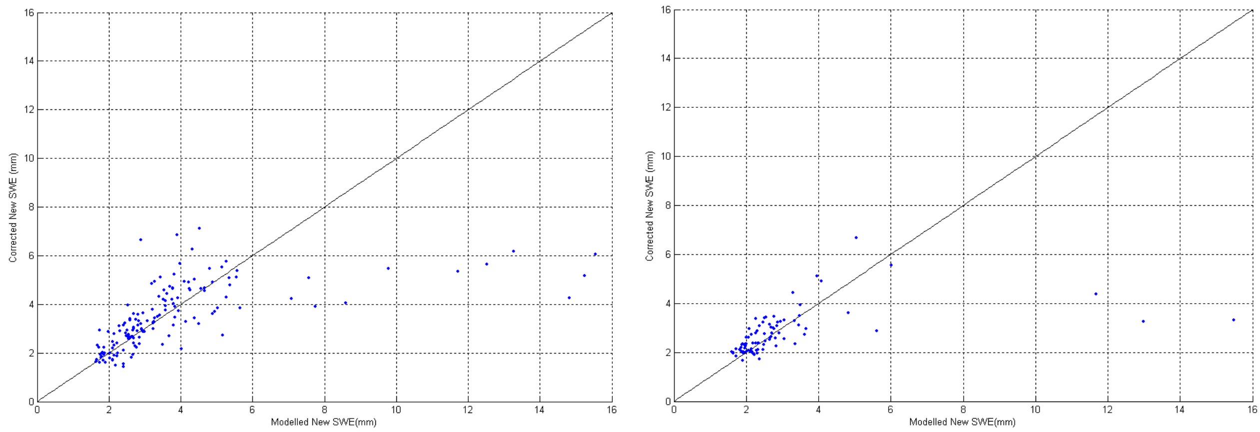

Figure 8 : Lago Pilone, Winter 2008/2009
Comparison between Modelled and Corrected Precipitation when Some Precipitation is Measured by the Rain Gauge (left) and when No Precipitation is Measured (right)

IV.5. Additional Work

a) <u>Diurnal Variations in Measured Snow Depth :</u>

When observing the curve representing measured snow depth, it appears that oscillations are frequent. They appear at a small temporal scale, even when no snow event occurs. The technical characteristics of the snow depth sensor are : precision = ± 1.5 cm max, sensitivity = 0.1 cm. So, oscillations of 1 or 2 cm each 30 min can be explained. However, changes more important can be recorded : increases or decreases of 5 cm in 30 min are quite frequent, and they can even reach more than 10 cm. Therefore, those changes have to be studied more in details. Previously, it has been said that such a phenomenon is often observed and some causes have been proposed : because of solar heating, temperature would be over-measured during the day, leading to decreases in measured snow depth. So, the objective is to verify if the explanations already found are valid for our data. The period chosen is included between 07/03/09 and 26/03/09. During these two weeks, both measured and modelled precipitations are zero, so that they can not be the cause of the observed variations.

First, when comparing snow depth variation to changes in snow surface energy balance, it appears that they are inversely related : when the energy balance is positive, that is to say when energy is added to the snow cover (i.e. during the day), snow depth shows a decrease (*Figure 1, Appendix D*). Moreover, observing both snow depth and air temperature variations, it appears that they are also inversely related (*Figure 2, Appendix D*). So, it confirms that oscillation in measured snow depth is a diurnal phenomenon which may be related to solar heating. In this case, the sound pulse velocity is corrected with a temperature which doesn’t represent actual air temperature. This would support the explanation of a snow depth under-measurement because of a too high air temperature measured at the snow depth sensor.

Besides, the same observation is done when doing the comparison with changes in snow surface temperature as it is directly linked with snow surface energy balance (*Figure 3, Appendix D*). It can be noted that this temperature is often close to 0 °C during the days for the period studied, so that snow surface characteristics vary, especially its albedo which decreases with wet snow. Therefore, the amount of reflected SW radiation might be a bit smaller, thus reducing the sensor heating, but the pulse is more prone to penetrate the snowpack, thus leading to snow depth under-measurement.

So, variations in snow depth are quite well explained. The next step could be to look more in details the relationship between changes in air temperature and changes in measured snow depth, so that the oscillations can be removed from the data. In this case, the detection of new snow events thanks snow depth evolution would probably be more accurate. Therefore, without using a numerical model and when no precipitation is measured by the rain gauge, it would be possible to know if precipitation should have been measured. Its amount could be estimated by equation (6) for example, so that the data from the rain gauge could be better corrected. Regarding the modelled data, they may not be very affected by this noise since the model already includes a control routine, which one eliminates spikes and uses maximum changes in snow depth.

b) <u>Selection of Events without Measurement Errors :</u>

The method proposed by Johnson and Marks (2004) to detect periods during which measurement errors can occur has been applied to our data (Allamano, 2009). This method requires that the snow temperature a few centimeters above the ground, if not available, is approximated by the average air temperature during the prior 24 h. At our observing site, this temperature is not measured so that the approximation has been done, but the bottom temperature and the snow temperature at 30 cm from the ground are available. Therefore, to try to valid this approximation, 24h-average air temperature will be compared to bottom temperature in a first time, and to snow temperature at 30 cm from the ground in a second time.

Comparison : 24h-average air temperature / bottom temperature

First, it has to be noted that when there is no snow on the ground (October and June), the temperature measured at the ground level almost corresponds to air temperature. But those data are not of interest since snow-pillow measurement errors can not occur when there is no snow layer.

Then, as expected, it appears that bottom temperature is almost constant during mid-winter, from December to May (*Figure 4, Appendix D*). Its value is either equal to -0.2 or to -0.1 °C, which means that even if 24h-average air temperature changes a lot, bottom temperature doesn't change. Besides, snow depth is always higher than 50 cm. Since snow is a good insulator, it can explain why bottom temperature is not influenced by air temperature. So, when snow depth is high enough, no relationship is found between the 24h-average air temperature and the temperature at the base of the snowpack.

During the period of snow cover formation (November), bottom temperature range from -0.8 °C to 0 °C, for a mean value of -0.2 °C. In fact, as soon as snow begins to accumulate, bottom temperature is almost constant. So, it is possible to consider that as soon as there is a fresh snow layer on the ground, bottom temperature is not influenced by air temperature.

Finally, during the end of the melting period (end of May), bottom temperature is not constant any more. It varies a lot, and can reach values higher than 10 °C, even up to 23 °C. However, snow depth is still around 40 cm when bottom temperature begins to vary. It may be explained by the fact that snowpack has begun to melt, so that it contains more liquid water and becomes less insulating : bottom temperature is more easily influenced by air temperature. Nevertheless, no relation between bottom temperature and 24h-average air temperature can be found (*Figure 5, Appendix D*).

Comparison : 24h-average air temperature / snow temperature at 30 cm from the ground

As previously, the data are not taken into account when snowpack height is below 30 cm since the measured snow temperature at 30 cm from the ground almost corresponds to air temperature.

For the other data, four remarkable periods are observed : one where snow temperature varies below 0 °C, another one where this temperature gradually increases and comes closer to 0 °C, a third one where it is constant, and a last one where it varies a lot (*Figure 6, Appendix D*).

The third period (April and May) is not of interest as the snow temperature is constant. Likewise, during the second period (from the end of December to the beginning of April), the snow temperature changes very slowly, so that it is possible to consider it as constant for periods of a few days long. Besides, the comparison between snow temperature and 24h-average air temperature doesn't show any correlation. So, it can be assumed that during the mid-winter, the snow temperature at 30 cm from the ground is not influenced by changes in 24h-average air temperature. It can also be underlined that snow temperature always remains below 0 °C during this period.

During the first period (November and December), the snow temperature varies and seems to be influenced by 24h-average air temperature. Either a linear or a parabolic relationship can be found with similar coefficients of correlation, respectively $r=0.69$ and $r=0.66$ (*Figure 7, Appendix D*). Besides, snow height is smaller than 1 m, which may explain why the snow temperature at 30 cm from the ground can be influenced during this period.

During the last period (end of May), at the end of the melting period, so as previously, snow temperature changes a lot, but no relationship with 24h-average air temperature can be found (*Figure 8, Appendix D*).

However, to study new snow density, only periods where snowfall occurs are of interest. They correspond to the beginning and the middle of the winter, that is to say when bottom and snow temperature at 30 cm from the ground are always below 0 °C, and often quite constant. So, when using these data, no measurement errors associated with temperatures at the melting point can be detected. Furthermore, when bottom temperature is not constant, no relationship can be found with 24h-average air temperature. When snow temperature at 30 cm from the ground is not constant, either a linear or a parabolic relationship is found at the beginning of the winter, i.e. when snowpack depth is not very important. But later in the winter, when the snow cover is very thick, no relationship is found. Besides, at the end of the melting period, both temperatures vary a lot and are above 0 °C almost all the time, which may produce measurement errors. However, no relationship is found between their variation and changes in 24h-average air temperature. Therefore, there is no evidence to support the fact that snow temperature a few centimeters above the ground is related to the 24h-average air temperature. Nevertheless, since no information about this parameter is available, using the 24h-aveage air temperature as suggested seems to be the better solution.

Conclusions and Outlook

The heated rain gauge under-catch was investigated at Lago Pilone, a high-elevation site in the Italian Alps, during the season 2008/2009. The objective was to estimate its magnitude thanks to precipitation data from a snow pillow measuring device. Actually, the gauge under-catch is almost zero when the precipitation is rain, but it is much higher when precipitation is snow. Besides, measurement errors are more likely to occur with high wind speed and low air temperatures. High wind speed, in fact, can deflect snow particles trajectory, while low air temperatures can lead to dysfunctions and gauge orifice (partial or total) obstruction. Furthermore, to estimate the measurement error, the actual precipitation must be known. The snow water equivalent values (SWE) measured by the snow pillow can not be considered as reference values since discrepancies are observed between these data and manual samplings.

Since our observing site is equipped with a fully instrumented meteorological station, enough data are available to use the numerical model SNOWPACK, which can simulate the snow cover evolution and characteristics. The comparison between measured and modelled SWE confirms the snow pillow under-measurement, and allows us to study heated rain gauge error. So, some problems related with the detection of rain events are observed, however they are few and have scarce influence on modelled new SWE. Moreover, some new snow events are modelled even if no precipitation is measured, which confirms that modelled new SWE are not influenced by measured precipitation. Those measurement errors can be related to meteorological conditions. As expected, they are more likely to occur with low air temperatures (inferior to -5 °C) and high wind speed (superior to 4 $m.s^{-1}$). Furthermore, modelled precipitation is assumed to represent the actual precipitation, so that the rain gauge error is defined as the difference between modelled and measured precipitation. The magnitude of this error (actually its logarithm) was correlated with wind speed and air temperature thanks to a multiple regression, leading to quite good relationships with r^2=0.56 or 0.43 (depending on whether some precipitation is measured of not), and coefficients all significant at a 5 % level. The measured values were then corrected according to these relations. Except for the very intense snowfall events which occurred around the 14/12/08, modelled and corrected new SWE are in good agreement. Nevertheless, the accuracy of this correction could be improved, and other studies could be done.

The modelled SWE from SNOWPACK, in fact, is often overestimated, so that the magnitude of the error may be over-estimated compared to the actual error. Therefore, a calibration of the modelled data could be necessary. Moreover, new snow depth is determined as the difference between measured snow depth and snow depth expected by the model if no snow event was occurred. Therefore, some new snow events are not modelled. Besides, when the modelled snow depth is higher than the measured one before a snow event, new snow depth might be underestimated. For example, modifying the model implementation (i.e. snow surface energy fluxes, settling rate), could improve the modelling of snow precipitation.

Furthermore, the measurement error can be studied for all modelled snow events, which means that a single correlation between measurement error and meteorological parameters could be found. This would allow to transpose the relationship to less instrumented sites. On the other hand, it could be interesting to look for a more site-specific correlation, including other variables like snow surface temperature or eventually short-wave radiations.

Variations at a small temporal scale are also observed in measured snow depth for the season 2007/2008 at Lago Pilone. Comparing them to changes in air temperature could support the explanation of an air temperature over-measurement at the snow depth sensor. Besides, the influence of snow surface characteristics like albedo could be studied more in details.

In conclusion, the present study has confirmed that at the present time, the snow pillow installed at Lago Pilone under-estimates the snow cover water equivalent. Moreover, a method was implemented to select periods without measurement errors, i.e. periods during which new snow density could be studied. These outcomes are of interest since the present study is part of a project conducted by the DITIC and the regional council of Piedmont whose aim is to know the snow cover water equivalent thanks to direct measurements and to the development of relationships between new snow density and meteorological parameters.

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APPENDICES

APPENDIX A :

Figures regarding Generalities about Snow

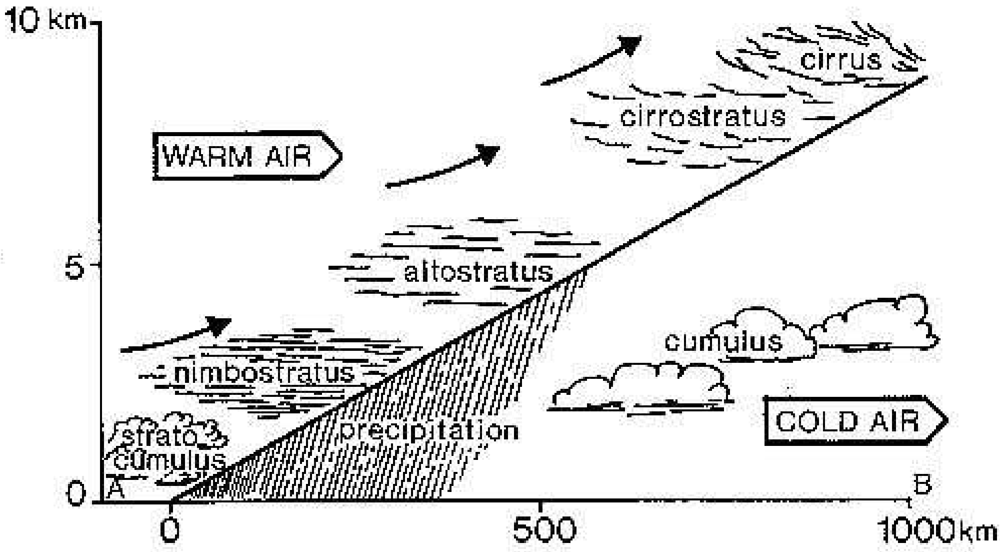

Figure 1 : Scheme of a Warm Front

[http://freedom-in-the-air.com/2007/05/30/helpful-films-and-diagrams-on-weather-cloud-fronts-wind/]

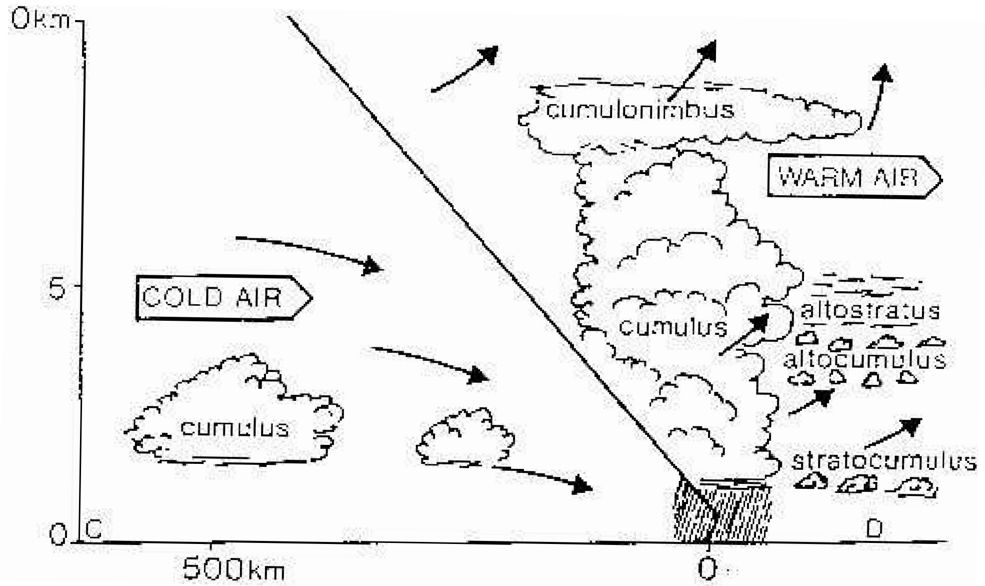

Figure 2 : Scheme of a Cold Front

[http://freedom-in-the-air.com/2007/05/30/helpful-films-and-diagrams-on-weather-cloud-fronts-wind/]

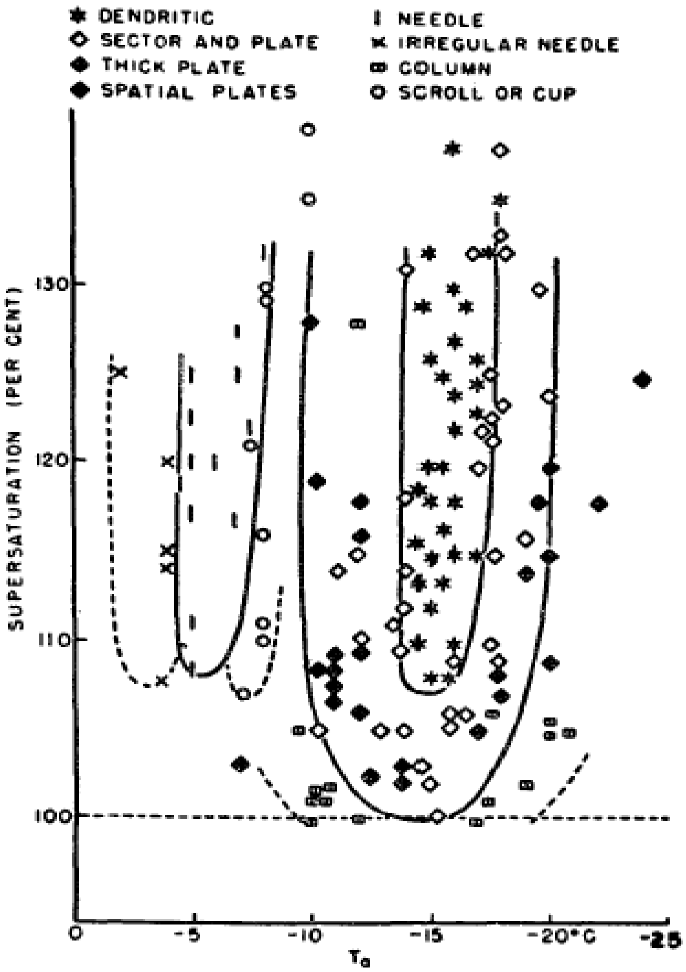

Figure 3 : Distribution of Crystal Types with Air Temperature and "Supersaturation" with Respect to Ice (%)

[Nakaya U., 1936 : Snow Crystals : Natural and Artificial. *Harvard University Press*, 510 pp.]

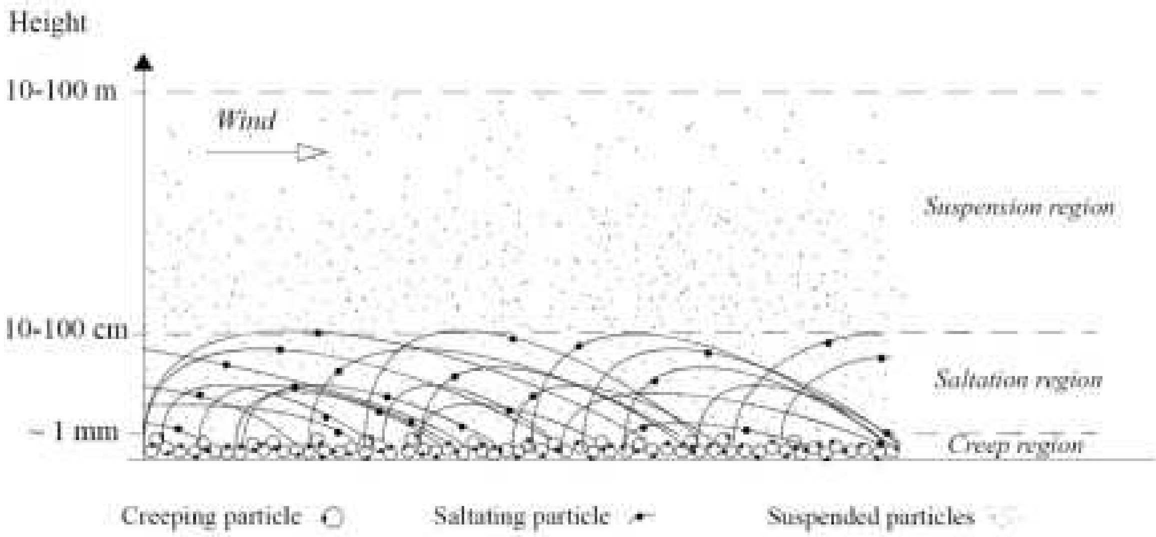

Figure 4 : Modes of Transport for Blowing Snow

[http://www.kaldtklima.net/node/234]

APPENDIX B :

SNOWPACK : Examples of Interfaces, Figures related to its Sensitivity

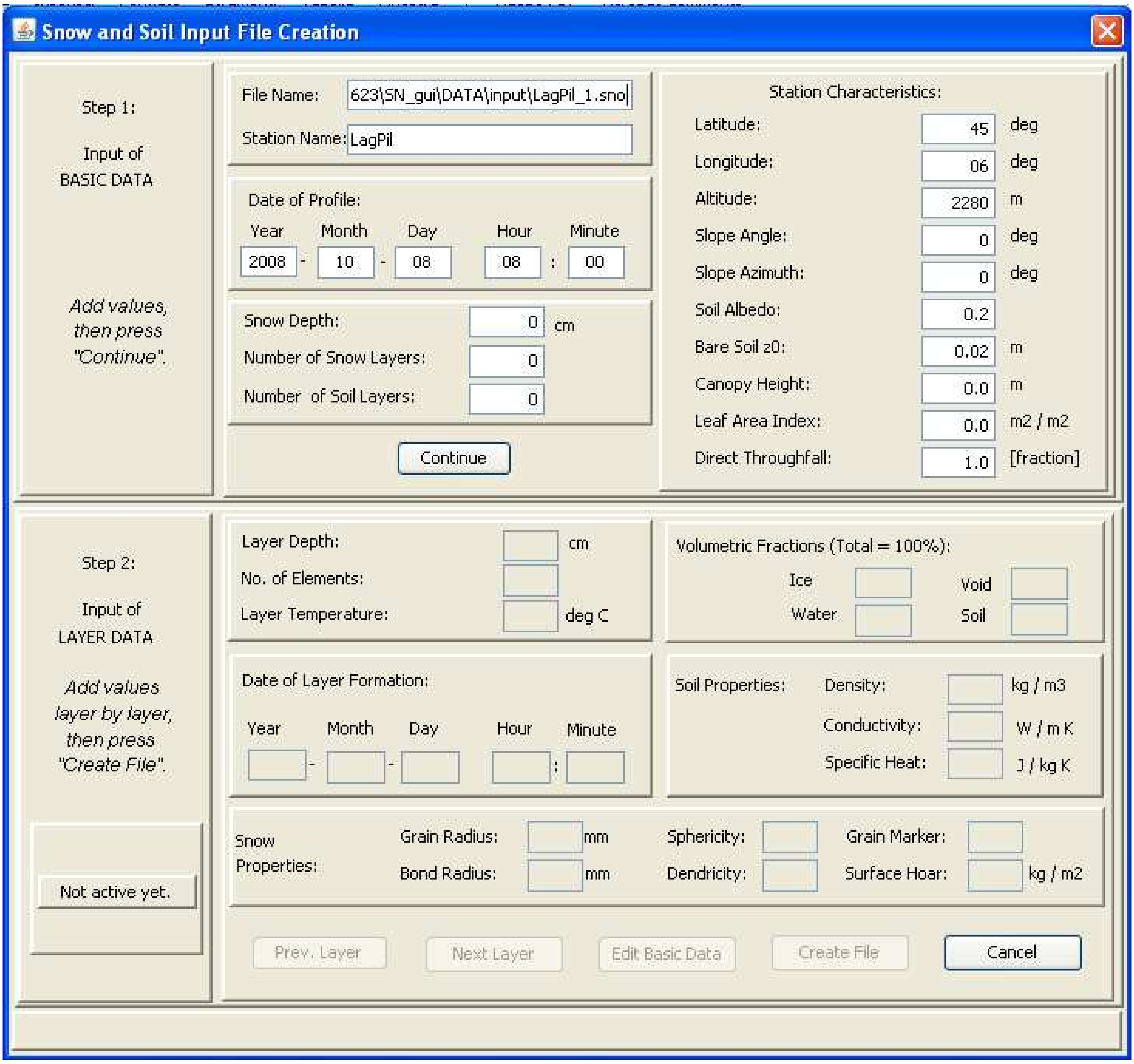

Figure 1 : SNOWPACK Initialisation
Interface for the Simulation at Lago Pilone, Winter 2008/2009

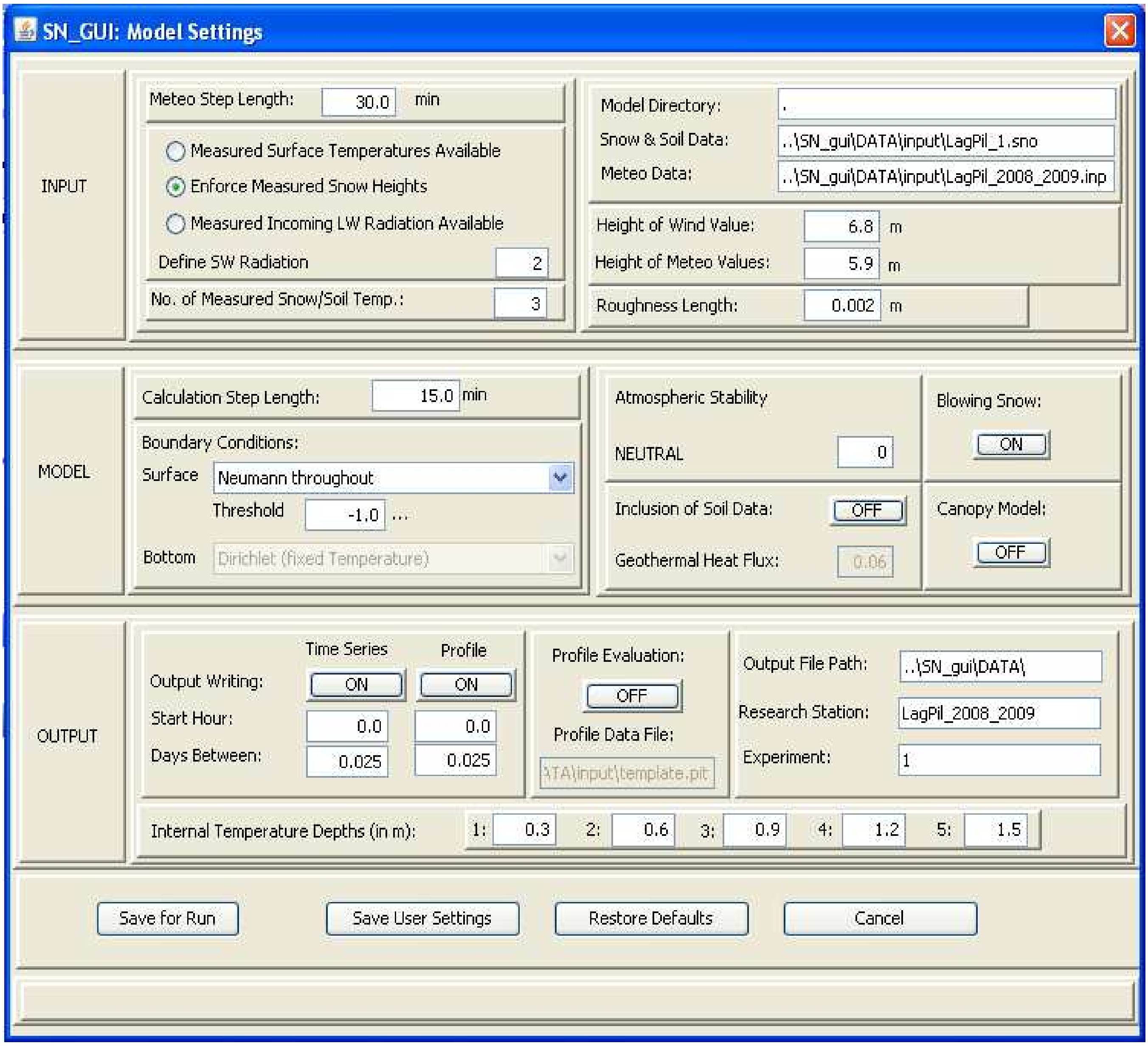

Figure 2 : SNOWPACK Operating Parameters
Interface for the Simulation at Lago Pilone, Winter 2008/2009

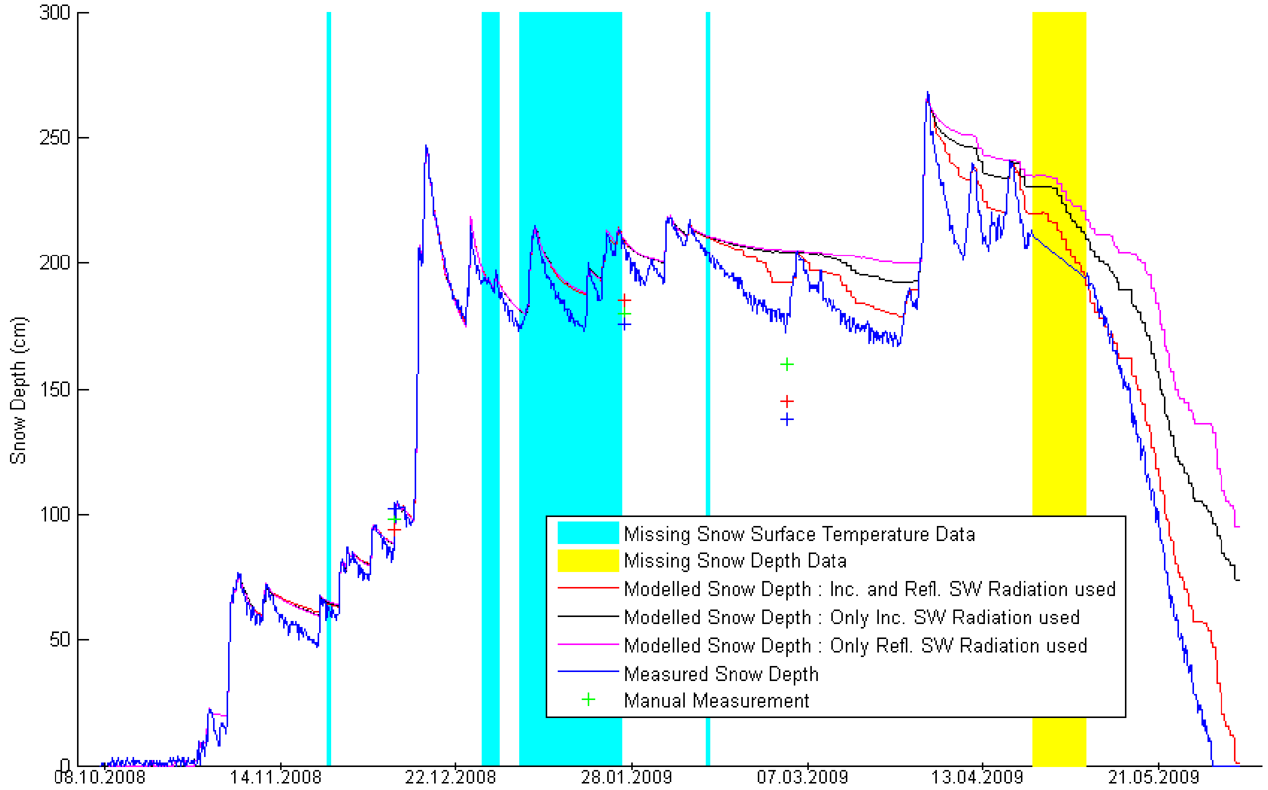

Figure 3 : Lago Pilone, Winter 2008/2009
Influence of Short-wave Radiation Used on Modelled Snow Depth,
i.e. Incoming, Reflected, or Both Short-wave Radiation

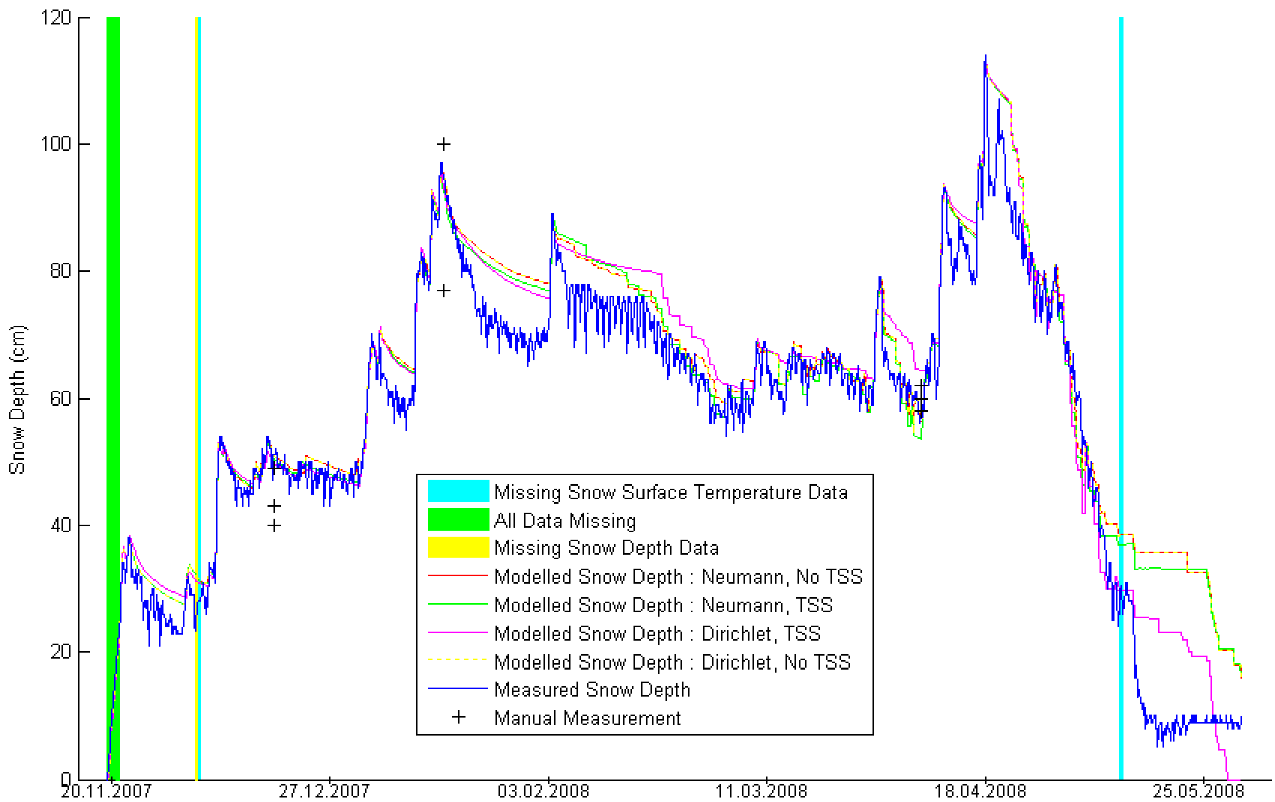

Figure 4 : Lago Pilone, Winter 2007/2008
Influence of Snow-Surface Boundary Condition (Neumann / Dirichlet) and Snow Surface Temperature Availability (TSS / No TSS) on Modelled Snow Depth

APPENDIX C :

Figures related to Modelled Precipitation and Data Correction

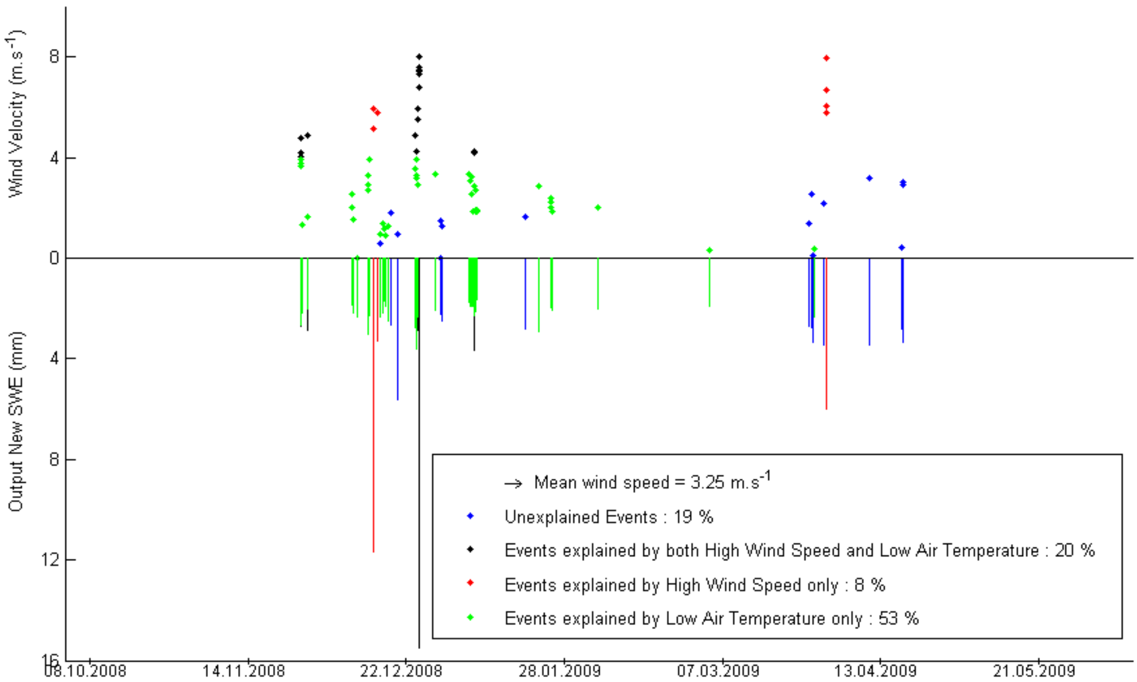

Figure 1 : Lago Pilone, Winter 2008/2009
Modelled New SWE when no Precipitation is Measured, Confrontation with Wind Speed

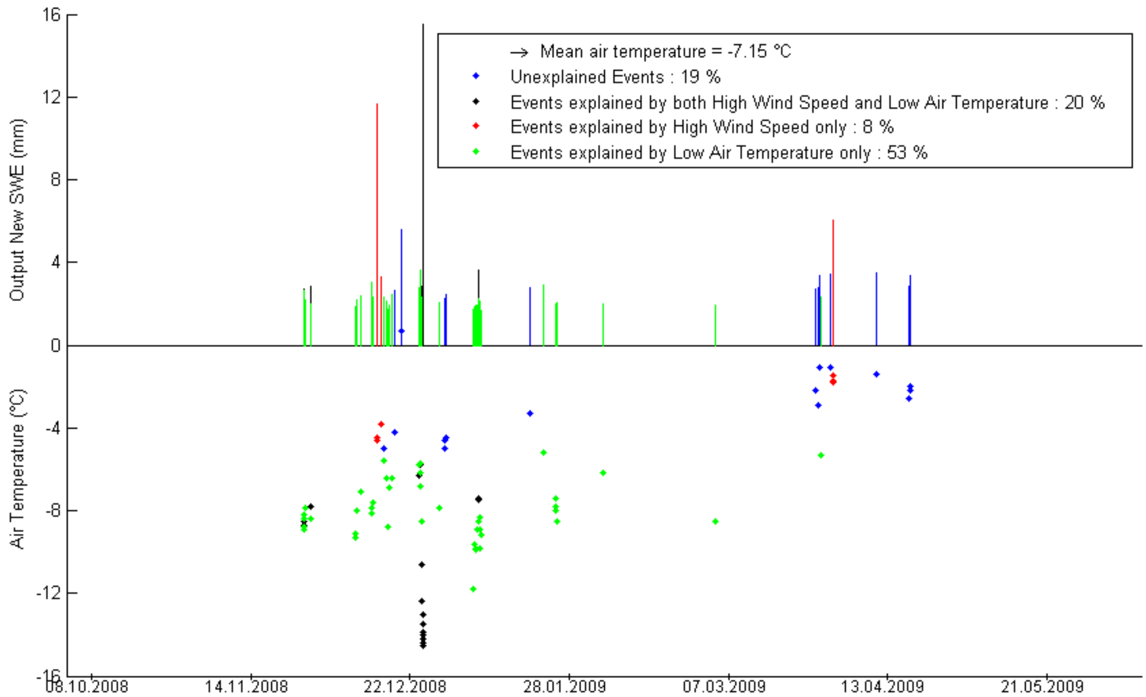

Figure 2 : Lago Pilone, Winter 2008/2009
Modelled New SWE when no Precipitation is Measured, Confrontation with Air Temperature

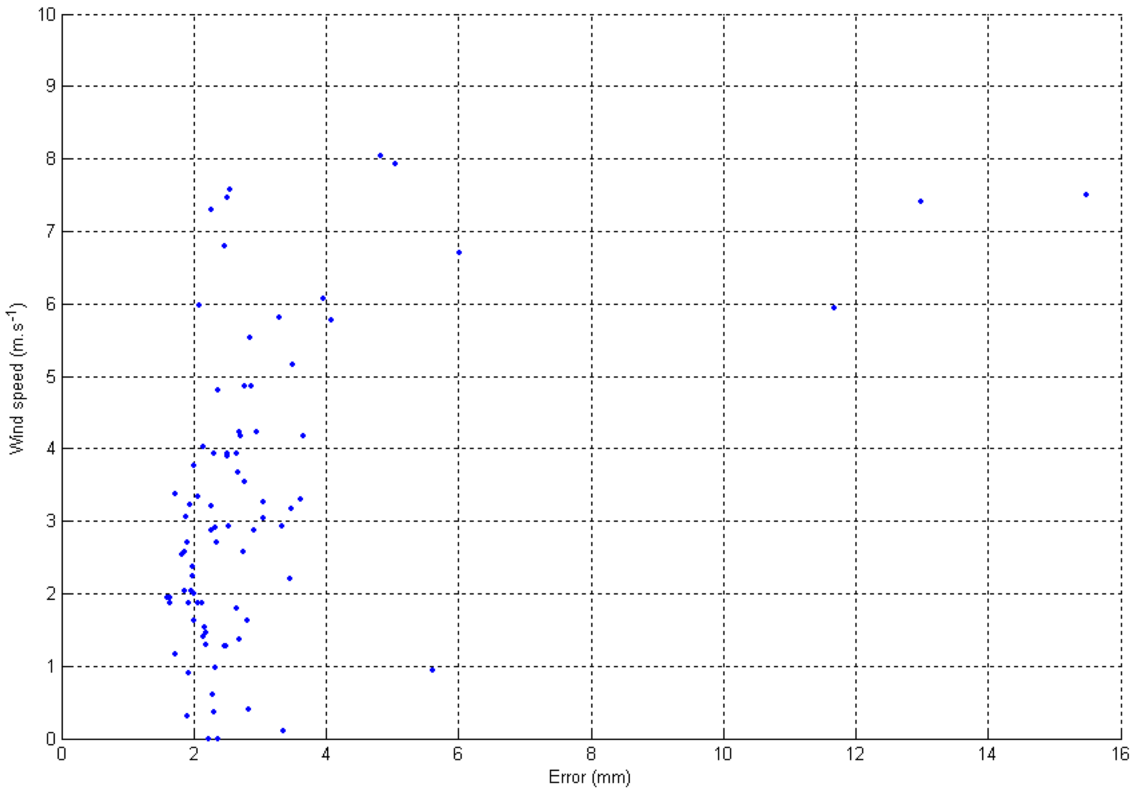

Figure 3 : Lago Pilone, Winter 2008/2009
Comparison between Rain Gauge Error and Wind Speed when No Precipitation is Measured

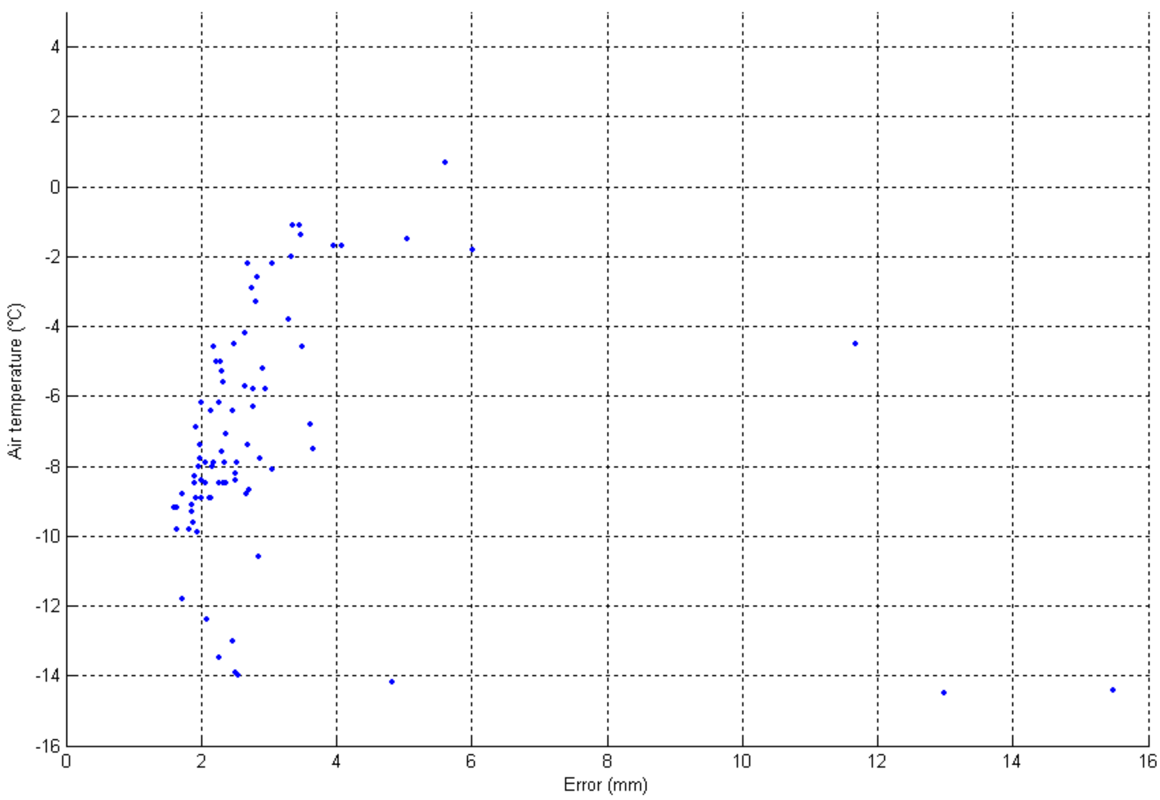

Figure 4 : Lago Pilone, Winter 2008/2009
Comparison between Rain Gauge Error and Air Temperature when No Precipitation is Measured

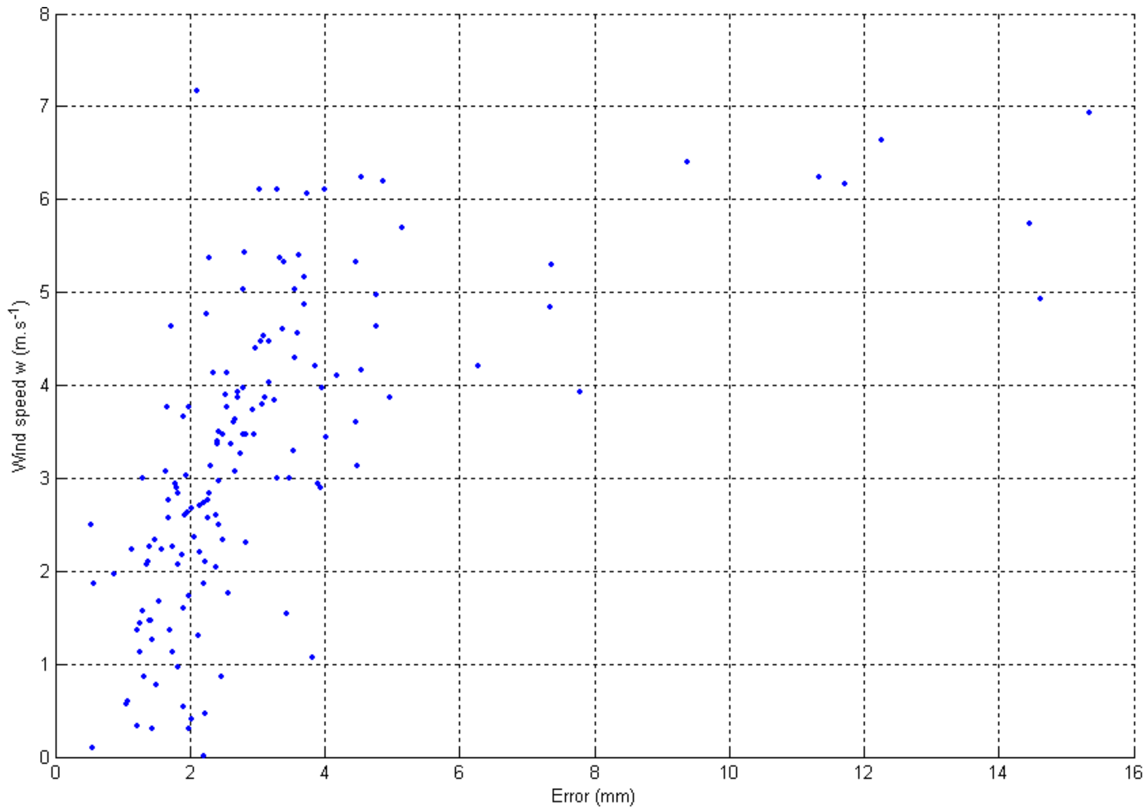

Figure 5 : Lago Pilone, Winter 2008/2009
Comparison between Rain Gauge Error and Wind Speed when Some Precipitation is Measured

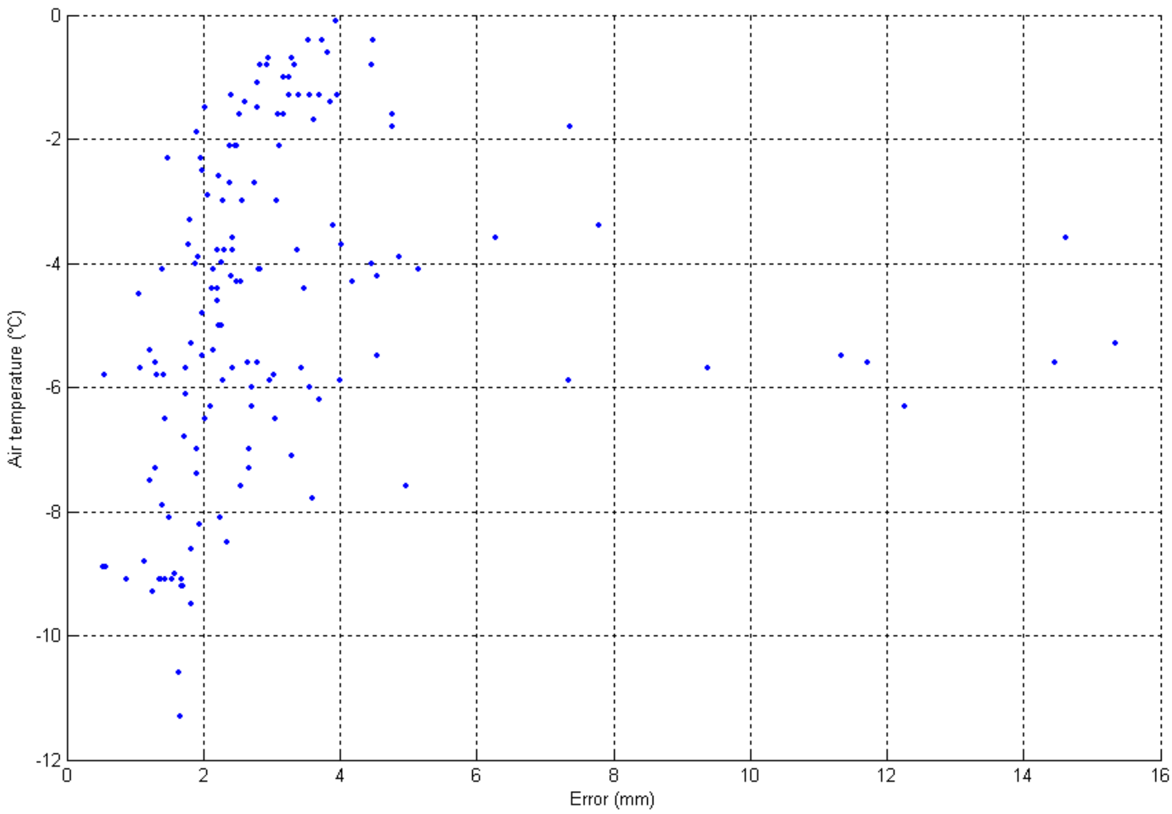

Figure 6 : Lago Pilone, Winter 2008/2009
Comparison between Rain Gauge Error and Air Temperature when Some Precipitation is Measured

	Coefficients				R^2
	value	std dev	T-value	p-value	
a	0,136	0,018	7,474	8,59E-11	0,426
b	0,049	0,011	4,473	2,53E-05	
c	0,890	0,087	10,248	2,22E-16	

Table 1 : Lago Pilone, Winter 2008/2009

Statistics regarding Equation 6 : ln(E)=aw+bT+c (No Precipitation is Measured by the Rain Gauge)

	Coefficients				R^2
	value	std dev	T-value	p-value	
a	0,231	0,020	11,705	0,00E+00	0,561
b	0,056	0,012	4,629	8,01E-06	
c	0,463	0,099	4,660	7,04E-06	

Table 2 : Lago Pilone, Winter 2008/2009

Statistics regarding Equation 7 : ln(E)=aw+bT+c (Some Precipitation is Measured by the Rain Gauge)

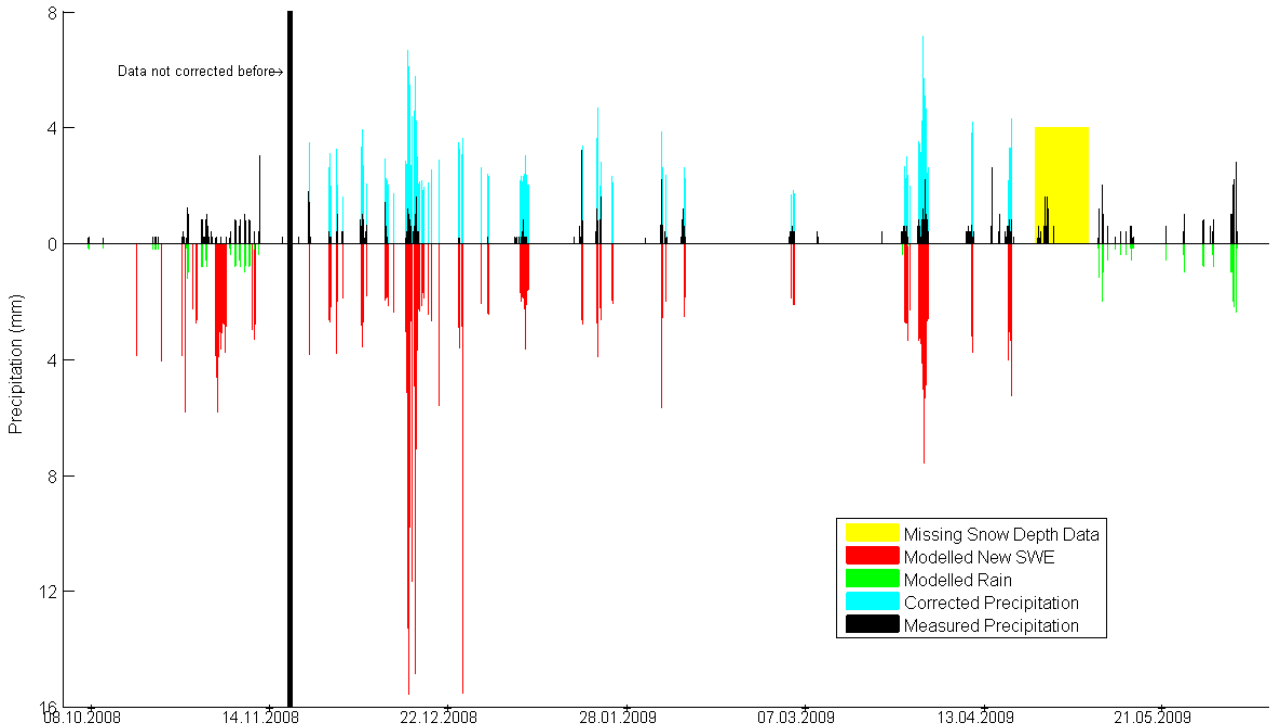

Figure 7 : Lago Pilone, Winter 2008/2009
Comparison between Measured, Modelled and Corrected Precipitation

APPENDIX D :

Additional Work

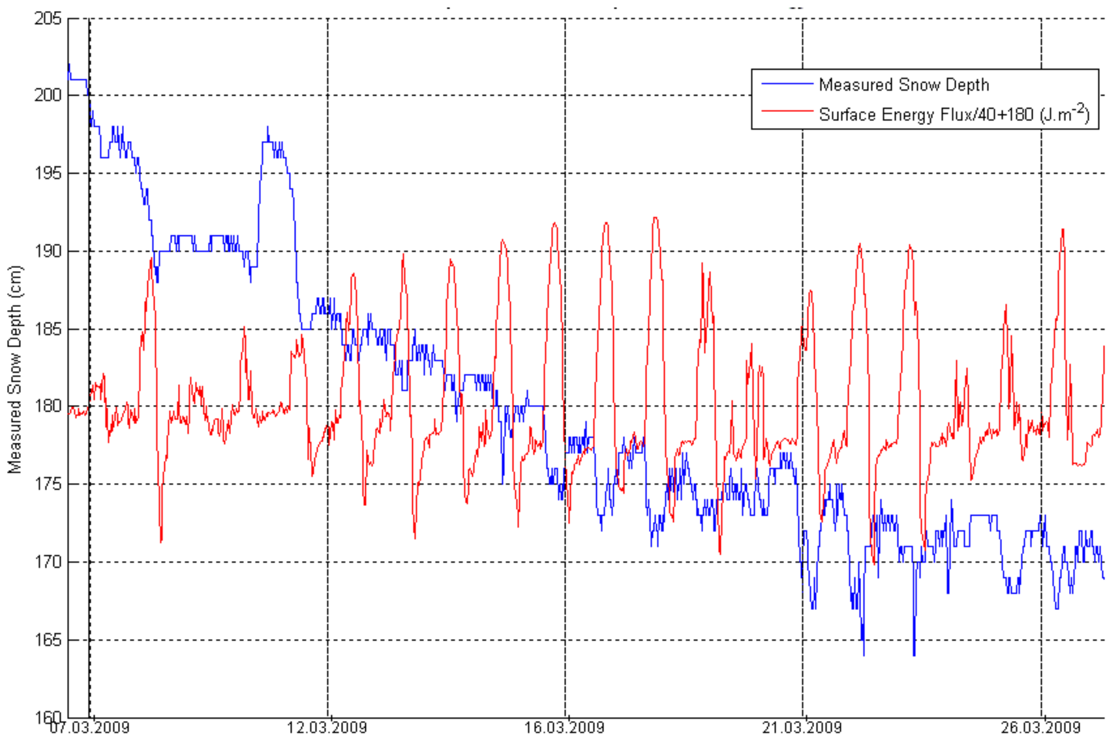

Figure 1 : Lago Pilone, Zoom : from 07/03/09 to 26/03/09
Comparison between Measured Snow Depth and Snow Surface Energy Balance

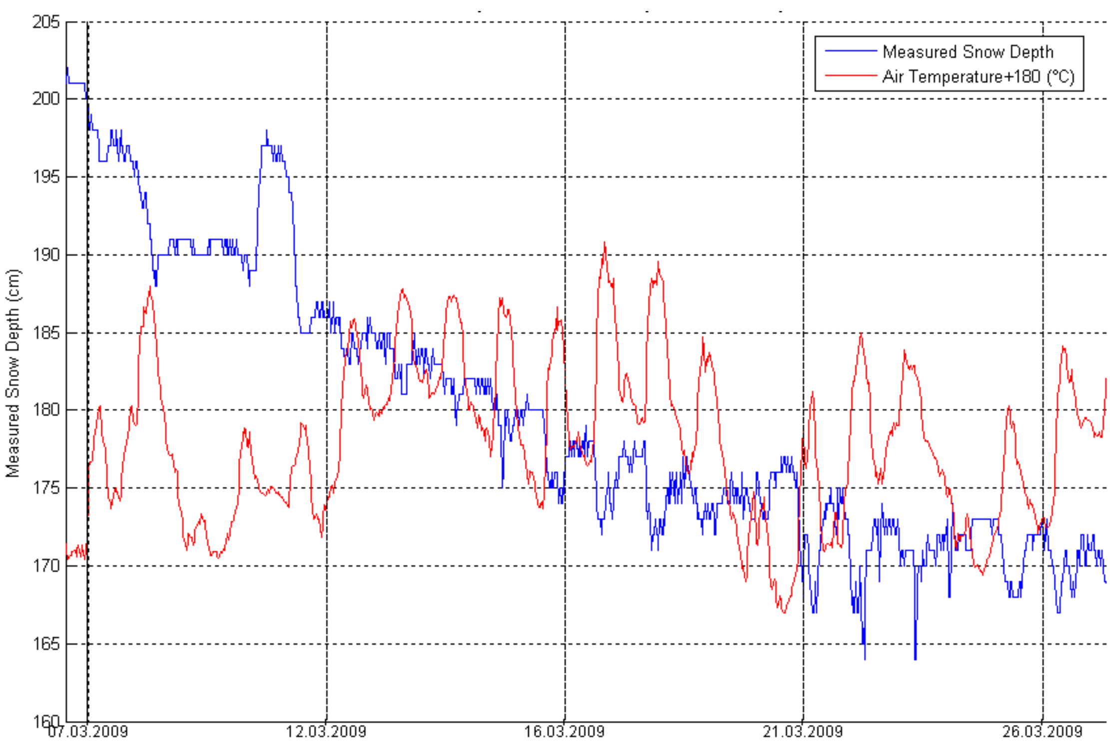

Figure 2 : Lago Pilone ,Zoom : from 07/03/09 to 26/03/09
Comparison between Measured Snow Depth and Air Temperature

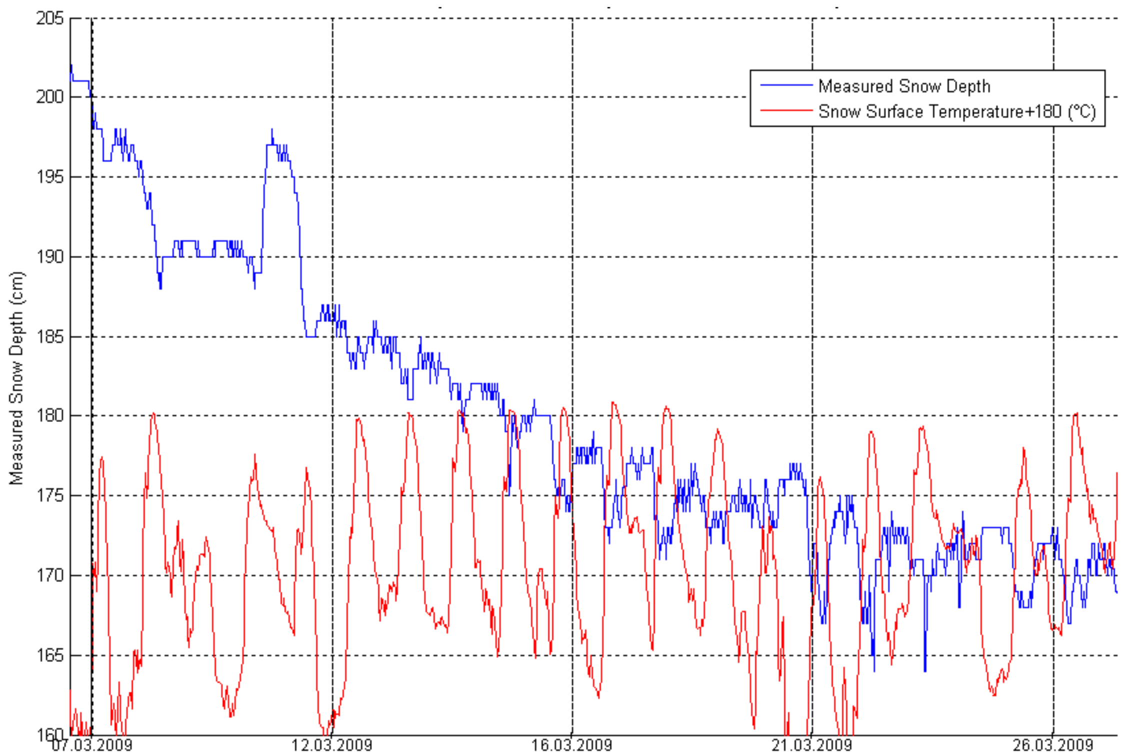

Figure 3 : Lago Pilone, Zoom : from 07/03/09 to 26/03/09
Comparison between Measured Snow Depth and Snow Surface Temperature

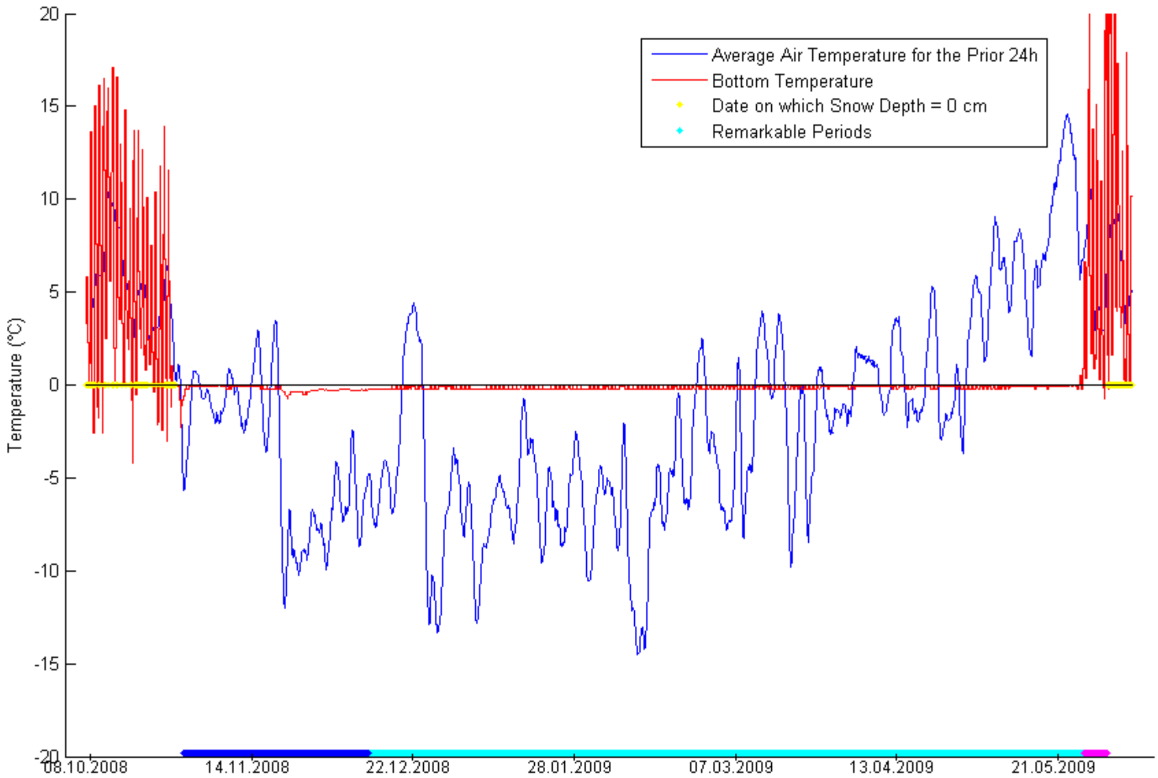

Figure 4 : Lago Pilone, Winter 2008/2009
Comparison between the Temperature at the Base of the Snowpack and
the Average Air Temperature for the Prior 24 Hours

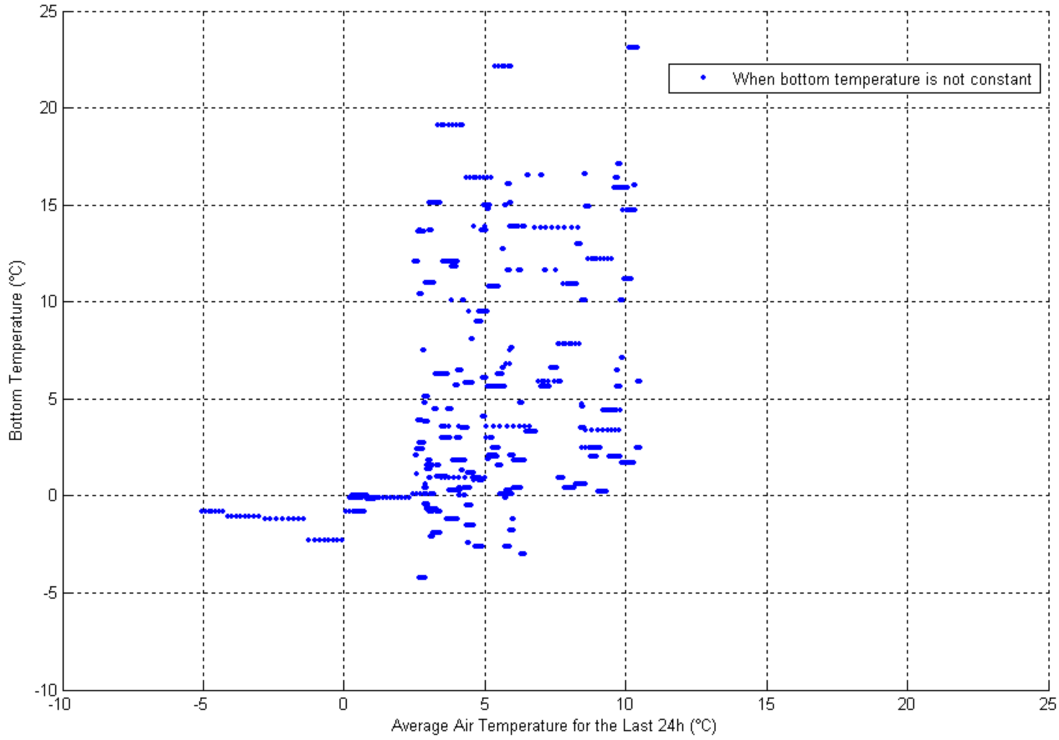

Figure 5 : Lago Pilone, Winter 2008/2009
Relatioship between the Temperature at the Base of the Snowpack and the Average Air Temperature for the Prior 24 Hours, Excluding the Period December-May where Bottom Temperature is Constant

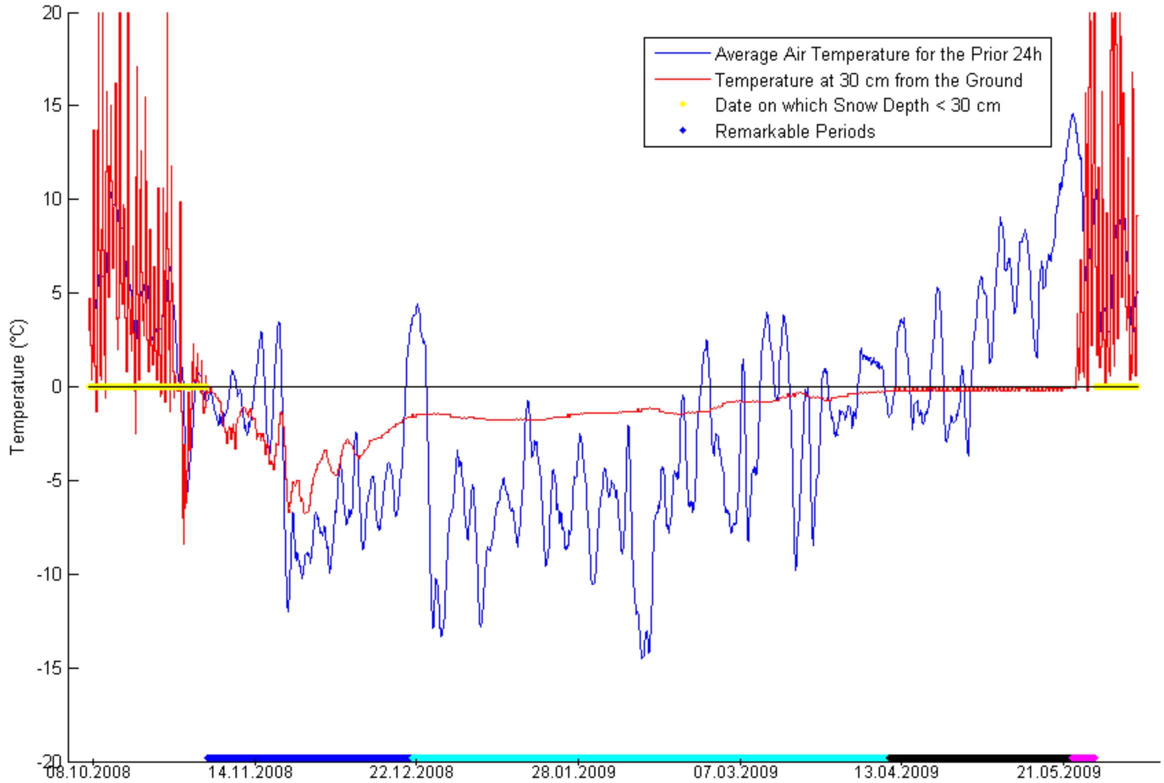

Figure 6 : Lago Pilone, Winter 2008/2009
Comparison between the Temperature at 30 cm from the Ground Surface and the Average Air Temperature for the Prior 24 Hours

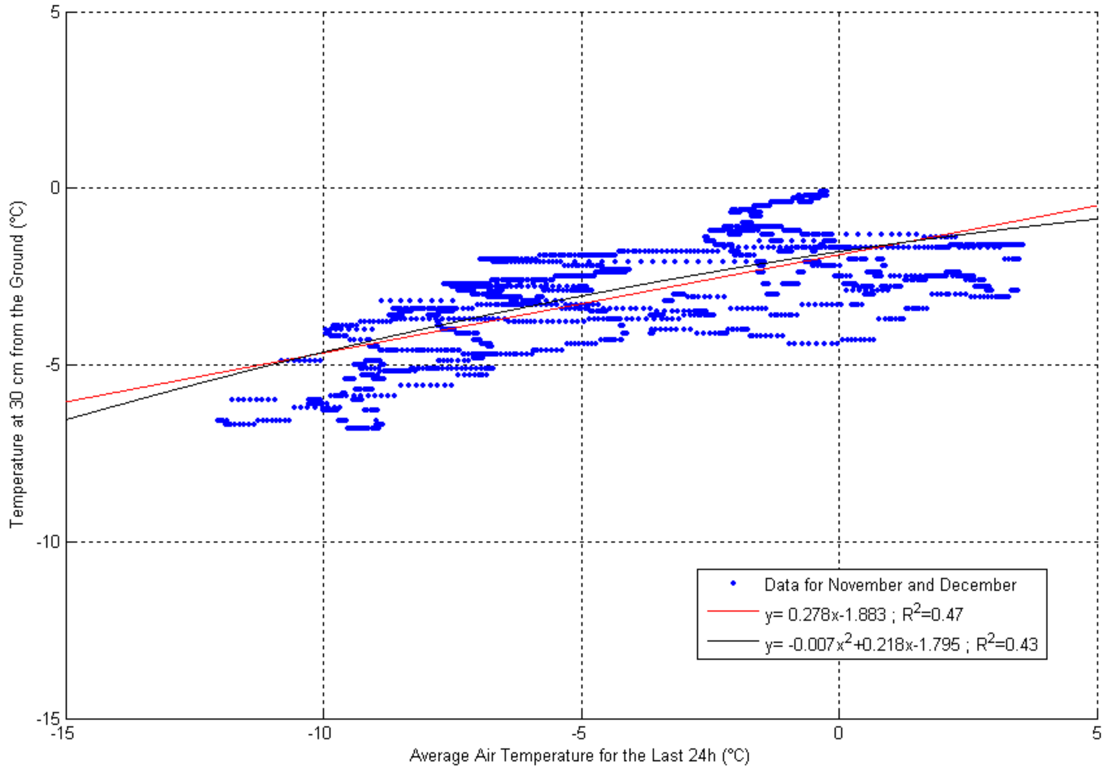

Figure 7 : Lago Pilone, Winter 2008/2009
Relationship between the Temperature at 30 cm from the Ground Surface and the Average Air Temperature for the Prior 24 Hours, for the Months of November and December

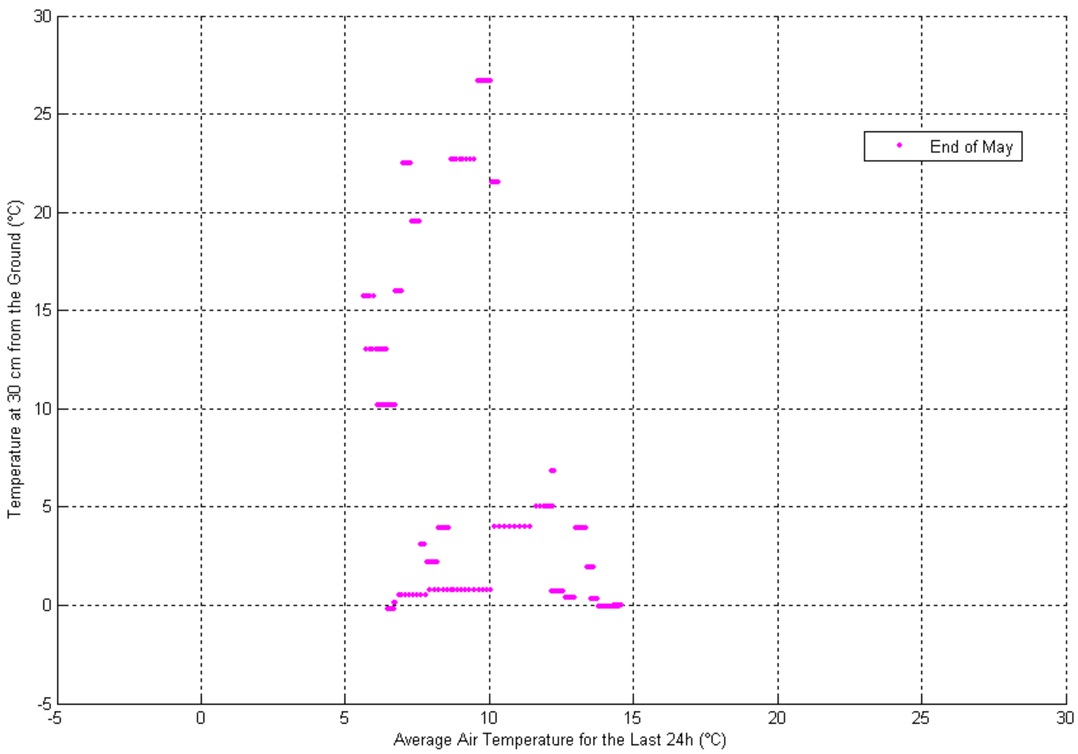

Figure 8 : Lago Pilone, Winter 2008/2009
Relationship between the Temperature at 30 cm from the Ground Surface and the Average Air Temperature for the Prior 24 Hours, at the End of May

APPENDIX E :

Work Done for the Winter 2007/2008 at Lago Pilone

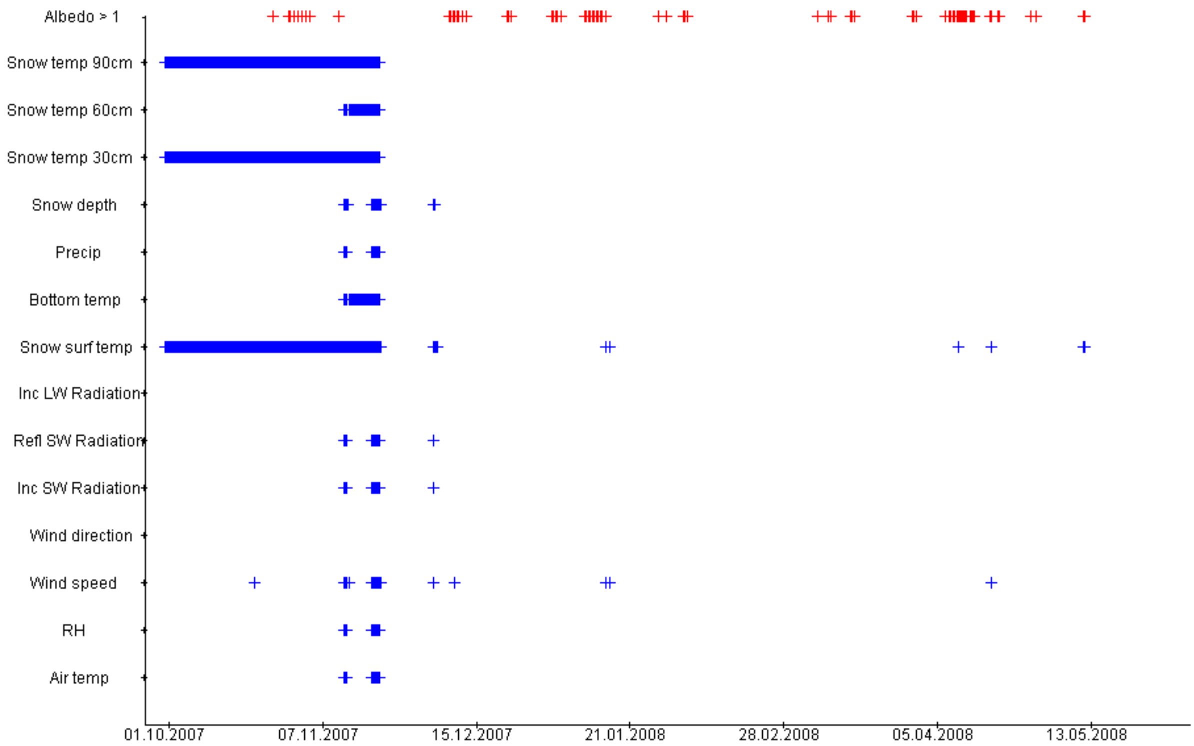

Figure 1 : Lago Pilone, Winter 2007/2008
Repartition of Missing Data between 01/10/07 and 08/06/08, Albedo Problem

Parameter	Total Number of Missing Data	Number of Missing Data which can be replaced by -99.9
Air temperature	135	0
Relative humidity	135	0
Wind speed	154	17
Wind direction	0	0
Incoming SW radiation	142	5
Reflected SW radiation	142	5
Incoming LW radiation	0	0
Snow surface temperature	2594	13
Bottom temperature	384	0
Precipitation	137	0
Snow depth	174	0
Temperature at 30 cm from the ground	2513	0
Temperature at 60 cm from the ground	384	0
Temperature at 90 cm from the ground	2513	0

Table 1 : Lago Pilone, Winter 2007/2008
Number of Missing Data between 01/10/07 and 08/06/08

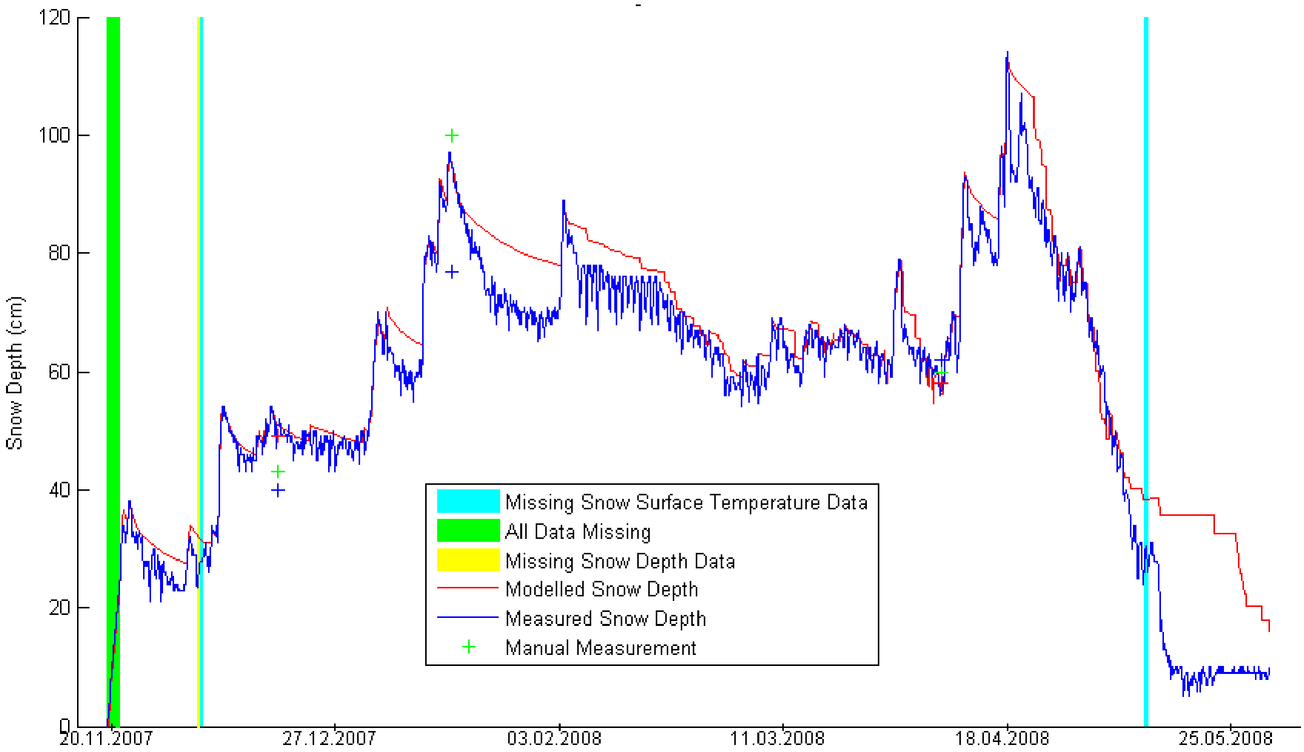

Figure 2 : Lago Pilone, Winter 2007/2008
Comparison between Measured and Modelled Snow Depth

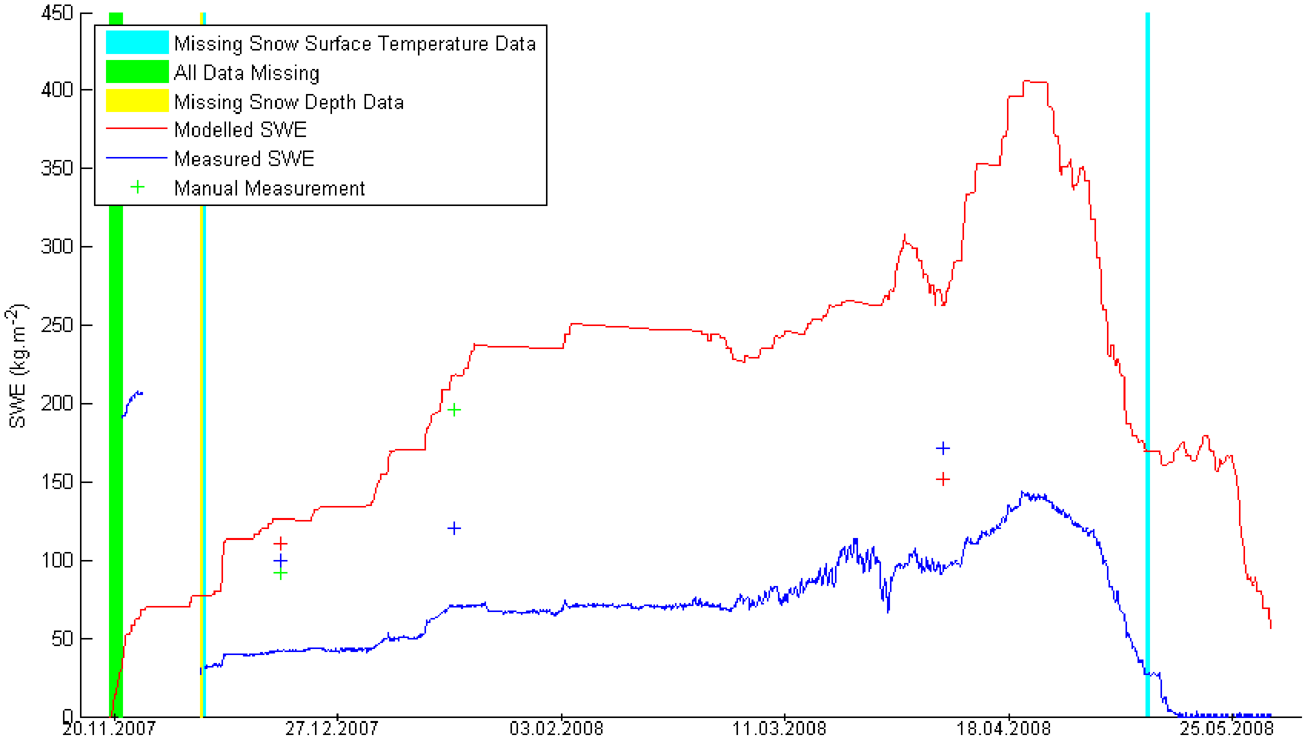

Figure 3 : Lago Pilone, Winter 2007/2008
Comparison between Measured and Modelled SWE

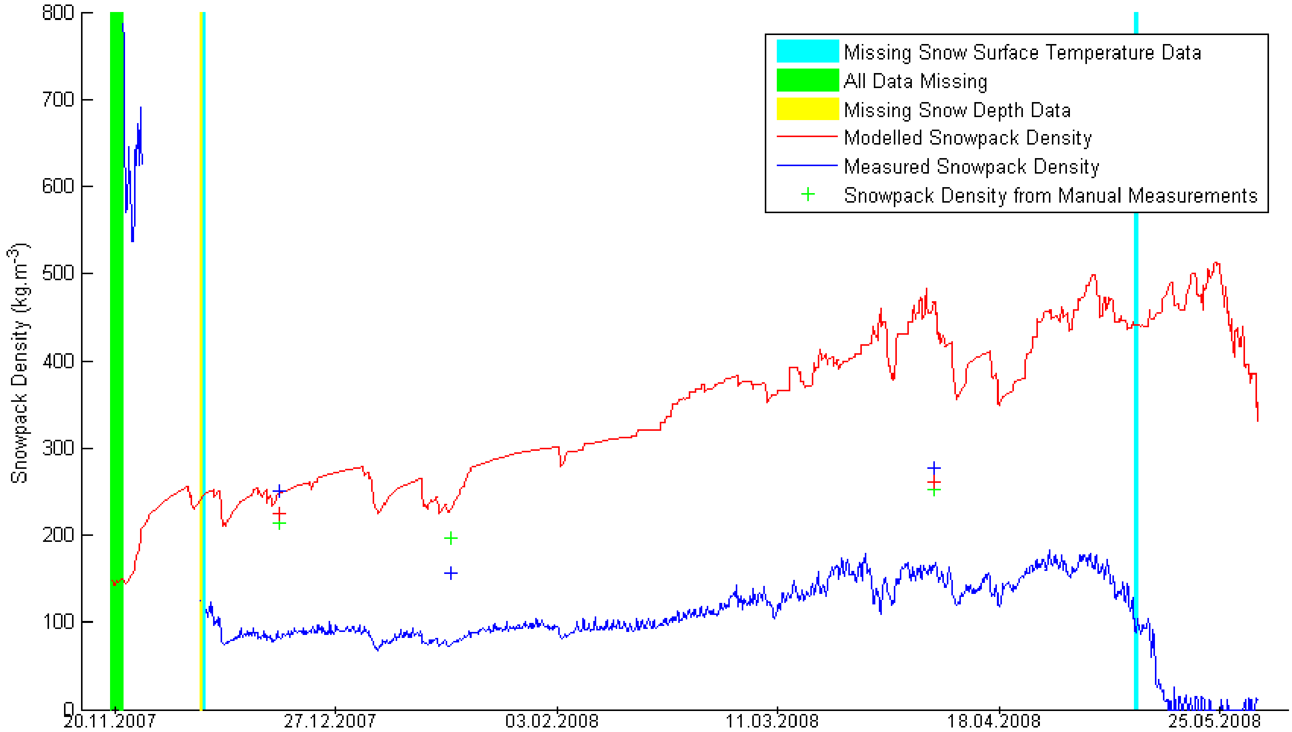

Figure 4 : Lago Pilone, Winter 2007/2008
Comparison between Measured and Modelled Snowpack Density

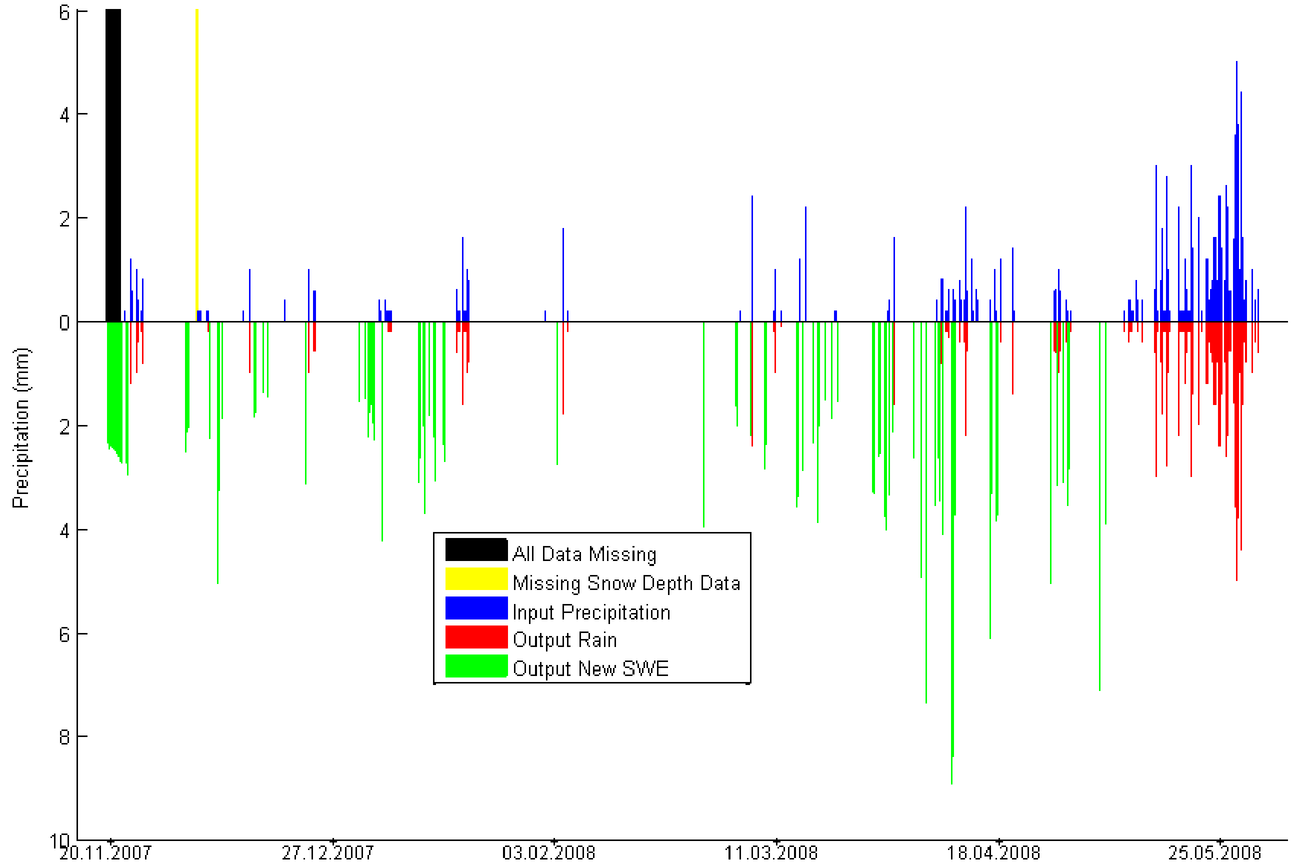

Figure 5 : Lago Pilone, Winter 2008/2009
Comparison between Measured and Modelled Precipitation

Modelled Rain Events with Air Temperature below 1,2 °C									
Date	24/11/07	19/01/08	09/04/08	09/04/08	09/04/08	10/04/08	11/04/08	30/04/08	11/05/08
Hour	13.00	3.00	10.30	15.30	17.30	16.00	12.30	12.00	0.00
Air Temperature (°C)	1,1	1,0	0,8	1,0	0,9	1,1	0,8	1,1	0,9
Measured Precipitation (mm)	0,6	0,2	0,4	0,8	0,4	0,6	0,6	0,4	0,4
Modelled Rain (mm)	0,3	0,1	0,2	0,4	0,2	0,3	0,3	0,2	0,2
Modelled SWE from New Snow (mm)	0,0	0,0	0,0	4,1	0,0	0,0	0,0	0	0
Wind Speed ($m.s^{-1}$)	0,0	1,5	0,9	0,5	0,1	1,4	2,6	1,2	3,5
Relative Humidity (%)	86	69	75	80	85	87	89	86	91

Table 2 : Lago Pilone, Winter 2007/2008
Modelled Rain Events with Air Temperature below 1.2 °C : Details

Observations :

At the beginning, many data are missing. To avoid at the maximum this problem, the simulation can be started on 20/11/07, when there is still no snow on the ground. Moreover, inputs and operating parameters are the same as for the season 2008/2009, so that results are more likely to be compared.

Regarding modelled snow depth, SWE and snowpack density, the observations are very similar. Moreover, the same problems appear with modelled precipitation, which one is quite higher than measured precipitation. There are still modelled rain events with air temperature below 1.2 °C, and even one event with both rain and snow. However, this event was not studied more in details. The problem of modelled new snow events without measured precipitation is still present and is more frequent, obviously because the rain gauge was not heated.

As we are interested in heated rain gauge error, no further work was done with the data from the season 2007/2008 at Lago Pilone.