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ESOLIP – estimate of solid and liquid precipitation at sub-daily time resolution by combining snow height and rain gauge measurements

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Abstract

Measuring precipitation in mountain areas is a demanding task, but essential for hydrological and environmental themes. Especially in small Alpine catchments with short hydrological response, precipitation data with high temporal resolution are required for a better understanding of the hydrological cycle. Since most climate/meteorological stations are situated at the easily accessible bottom of valleys, and the few heated rain gauges installed at higher elevation sites are problematic in winter conditions, an accurate quantification of winter (snow) precipitation at high elevations remains difficult. However, there are an increasing number of micro-meteorological stations and snow height sensors at high elevation locations in Alpine catchments. To benefit from data of such stations, an improved approach to estimate solid and liquid precipitation (ESOLIP) is proposed. ESOLIP allows gathering hourly precipitation data throughout the year by using unheated rain gauge data, careful filtering of snow height sensors as well as standard meteorological data (air temperature, relative humidity, global shortwave radiation, wind speed). ESOLIP was validated at a well-equipped test site in Stubai Valley (Tyrol, Austria), comparing results to winter precipitation measured with a snow pillow and a heated rain gauge. The snow height filtering routine and indicators for possible precipitation were tested at a field site in Matsch Valley (South Tyrol, Italy). Results show a good match with measured data because variable snow density is taken into account, which is important when working with freshly fallen snow. Furthermore, the results show the need for accurate filtering of the noise of the snow height signal and they confirm the unreliability of heated rain gauges for estimating winter precipitation. The described improved precipitation estimate ESOLIP at sub-daily time resolution is helpful for precipitation analysis and for several hydrological applications like monitoring systems and rainfall-runoff models.

1 Introduction

Quantification of precipitation is still a major source of uncertainty within the water budget of Alpine catchments (see e.g. Egli et al., 2009; Gottardi et al., 2012; Sevruk et al., 2009). Most of the climate/meteorological stations are located at the valley bottom, while, at higher elevations, the number of measurement devices and their accuracy is limited by accessibility, energy supply, financial costs and wind influence.

At higher elevations, where a significant part of annual precipitation falls as snow, various methods are known to measure or estimate precipitation. For accumulated precipitation measurement totalisators work well (Scherrer, 2010). Advanced instruments, such as snow pillows, snowpowers, parsivels or heated weighing rain gauges are used for a reliable measurement at short timescales (Egli et al., 2009). However, their number and use is limited by maintenance and monetary constraints. The influence of wind (e.g. snow wind drift, surrounding wind fields, etc.) causes problems in getting correct precipitation information, especially for measurements with rain gauges (Jordan et al., 1999; Sevruk et al., 2009; Sevruk, 1986) usually installed at a height of 2 m, where wind speed is higher than closer to the ground (Häckel, 2005). This makes it necessary to correct winter precipitation with empirical factors, often of more than 40 % (Carturan et al., 2012). Hence, wind influence is more critical for relatively simple heated rain gauges than for near-surface measurements (i.e. snow pillow).

In recent years automatically measured snow height and meteorological data (e.g. air temperature, relative humidity, global shortwave radiation, wind speed...) have increasingly become available. While snow height as a single parameter does not provide useful information on snow amount/volume (Armstrong and Brun, 2008), it allows an estimate of snow water equivalent (SWE) when combined with snow density. Considering the undercatch problem of heated rain gauges during snowfall (Judson and Doesken, 2000; Sevruk, 1986), the time required for manual measurements (and the related less detailed temporal resolution) and the financial and logistical effort to set up

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equipment such as snow pillows, alternative approaches to obtaining reliable precipitation data at high temporal resolution are needed.

In small catchments with a correspondingly short hydrological response (Symader, 2004) datasets with sub-daily or hourly time resolution are necessary to get a more detailed understanding of runoff production and critical flood conditions. Especially in high elevation catchments, response times have a dimension of hours or less, and the elevation of the snow line controls flood magnitudes (Allamano et al., 2009). Moreover, especially for catchments with dry climatic conditions, it is important (especially in winter and spring) to consider every single precipitation event in order to quantify the water budget. While several methods are mentioned in literature to calculate snow water equivalent (SWE) and therefore derive precipitation from snow height measurements at seasonal (Jonas et al., 2009) or daily (Egli et al., 2010; Sevruk, 1986; Sturm et al., 2010) time resolution, these approaches are not validated for hourly data of freshly fallen snow and across the whole hydrological year yet. In fact, most of the existing methods to get SWE from snow height measurements aim at total sums for the winter period instead of identifying single precipitation events.

In order to calculate SWE from snow depth measurements, an accurate estimate of snow density is necessary. The average density of freshly fallen snow is assumed to be 100 kg m^{-3} (Goodison et al., 1998; Sevruk, 1986). However, several studies (Egli et al., 2009; Hedstrom and Pomeroy, 1998; Jordan et al., 1999; Sevruk, 1986) have shown that the density of new snow may vary from 30 kg m^{-3} to more than 450 kg m^{-3} in different meteorological conditions. The most important meteorological factors influencing snow density are air temperature and wind speed. Both factors are included in the calculation of snow density proposed by Jordan et al. (1999), while Brazenec (2005) tested four methods to determine snow density from temperature only and got the best results using the method developed by Hedstrom and Pomeroy (1998).

When identifying precipitation events, a differentiating factor between rainfall and snowfall is needed. An air temperature of 0 °C is no sharp trigger point between snow or rain (Häckel, 2005). As described by US Army Corps of Engineers (1956) and Rohregger (2008), there is a transition zone from rainfall to sleet and finally snow. Two better possibilities to estimate the snow line are wet bulb and dew point temperature, where relative humidity is integrated in the calculation (Steinacker, 1983). Because of changing meteorological conditions in the course of the day the differentiation between rainfall and snowfall is easier at sub-daily time resolution. This is important because the elevation of the snowline can significantly change during one day/precipitation event in Alpine areas (Armstrong and Brun, 2008) with dramatic consequences for runoff production. Additionally, the frequency of snow depth measurements is important since sub-daily data allow a better consideration of melting, settling and redistribution processes (Brazenec, 2005). This is especially relevant for snow depth measurements at wind exposed sites, for small events, and in spring, when snow accumulation and snowmelt might occur during the same day.

The most common approach to measuring snow depth automatically is using an ultrasonic snow depth sensor. The fluctuations of the ultrasonic snow height sensor due to temperature are known as a source of error (Terzago et al., 2012). Manufacturer calibration does not fully remove the fluctuation errors (Ryan et al., 2008), thus, when working at high temporal resolution, these can also be misinterpreted as small snowfall events or vice versa. Therefore the help of filtering procedures is needed to minimize this source of error in order to calculate real snow height increments. A moving average is an easy and effective way of handling this problem (Brazenec, 2005; Terzago et al., 2012).

We hypothesize that the combined use of unheated rain gauge data for rainfall events and carefully filtered snow height data, in combination with temperature- and wind-dependent density of freshly fallen snow, to calculate SWE of snowfall events is a simple and cost-effective solution to estimate sub-daily precipitation data throughout the year. In order to test this hypothesis, we validated a procedure to Estimate SOlid and LIquid Precipitation (ESOLIP), not only for seasonal totals but also for single precipitation events at sub-daily timescales. Precipitation possibility was estimated with meteorological data and meteorological information. The signal of the snow height sensor

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was filtered with moving averages to identify single snow height increments caused by snowfall events. To investigate the importance of variable snow density, different methods to calculate the density of freshly fallen snow were tested. We applied ESOLIP at one study site and the tested snow height filtering routine and indicators for possible precipitation at another test site in the Alps. Finally we discuss major limitations and advantages of the method.

2 Data and test sites

ESOLIP was validated at the study area Kaserstattalm (47°7′44′′ N, 11°18′20′′ E, LTER site Stubai) at an elevation of 1830 m a.s.l., situated in the Stubai Valley, Austria. In addition to a meteorological station, including an ultrasonic snow height sensor, a heated and an unheated rain gauge, air temperature, global shortwave radiation, wind and humidity sensors (Campbell Scientific, Logan, UT, USA), a snow pillow (Snow Pillow 3 m × 3 m, Sommer GmbH & Co KG, Austria) is installed at the site, which has provided data for three winter periods (November 2009–May 2012). The snow height filtering routine and the indicators for possible precipitation were tested at the study site Matsch Valley (46°41′10′′ N, 10°34′46′′ E), Italy, at 1500 m a.s.l. This site is equipped with a meteorological station, including a snow height sensor, unheated rain gauges, air temperature, global shortwave radiation, wind and humidity sensors (Campbell Scientific, Logan, UT, USA) providing hourly data series since installation in 2009. At the same elevation, in a distance of 3 km (46°41′40′′ N, 10°37′3′′ E), manual observations are available. Every day at 9 a.m. precipitation records (heated rain gauge, snow height) are collected. Additionally, metadata from a daily meteorological bulletin called "climareport" (Hydrographic Office of the Autonomous Province Bozen, 2013) were considered. Both stations use the same sensor type, mounted at the same height above the soil surface.

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3 Method

ESOLIP aims at estimating snow and rainfall precipitation at hourly time-scale using a snow height sensor in combination with an unheated rain gauge and auxiliary micrometeorological data illustrated in the scheme of Fig. 1. The method can be divided into three basic steps:

1. Pre-processing and a quality check of rain gauge and snow height sensor data is required to ensure data reliability. This includes a careful filtering of the noise in the snow height signal to identify snow height increments related to snowfall events.

2. In a second step the possibility of precipitation is controlled with the help of measured meteorological data and, where available, verified with metadata (such as meteorological reports and observer statements). Precipitation recorded in the unheated rain gauge, but highly improbable according to additional data, is identified as melting snow and not considered for the resulting precipitation data series.

3. The third step distinguishes between snowfall and rainfall using a threshold on wet bulb temperature. For rainfall events, rain gauge data is used directly. In case of a snowfall, snow density and SWE are calculated using meteorological data.

3.1 Data pre-processing and filtering (see 1. in Fig. 1)

Data quality control is essential when dealing with meteorological records (Zahumenskỳ, 2004). First, data from unheated rain gauges were checked for outliers and incorrect measurements (e.g. negative values) were removed from the data series.

In addition, the noise in the signal of the ultrasonic snow height sensor was accurately removed to identify snow height increments at an hourly time-scale, which might be related to snowfall events. The signal of ultrasonic snow sensors is affected by changing air temperature. In particular during sunny days with high global shortwave

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radiation, snow height measurements tend to decrease by a couple of centimetres and rise back to the initial value with decreasing temperature and global shortwave radiation (Terzago et al., 2012). Manufacturer calibration compensates ultrasonic sensor readings with respect to air temperature, but does not account for the described decrease and rise in snow height measurement, thus further filtering is needed. The filter should be an optimal compromise between the need to remove oscillations caused by temperature fluctuations and to preserve snow height increments related to snowfall. For the datasets from Matsch and Stubai, a moving average of 5 h delivered the best results, since it smoothed small oscillations but did not falsify peaks (Figs. 2, 3). A more detailed analysis on the choice of the moving average interval for smoothing is shown in the results section.

3.2 Possibility of precipitation

The next step to estimate the correct precipitation amount is to distinguish between real precipitation events (either solid or liquid) and erroneous sensor readings (see 2. in Fig. 1) which could i.e. be caused by melting snow (which was accumulated in an unheated rain gauge). This can be done with the help of additional, automatically recorded, meteorological data. If available, metadata (such as meteorological or synoptic reports, human observations) can help to exclude periods with no precipitation from further analysis. For the test site in Matsch, checking metadata allowed excluding half of the dataset from further analysis. Additional meteorological data at hourly basis describe the general weather conditions and allow further reduction of periods to investigate. Especially relative humidity and global shortwave radiation can indicate precipitation events. The objective was not to lose more than 2.5 % of precipitation when filtering data only by relative humidity or global shortwave radiation alone. If both relative humidity and global shortwave radiation are used as filters, not more than 1 % of precipitation should be lost.

Relative humidity (RH) measured at the weather station increases with precipitation (Häckel, 2005). At the Matsch test site, the average RH during precipitation events

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was 93 %, but some very low RH values were also measured (Table 1). Such low values can be found also in the literature. Rohregger (2008) and Lehning et al. (2002) used a threshold of RH > 70 % to separate between snowfall or no snowfall. Based on our analysis (Table 1), a threshold of RH > 50 % which misses only 1.4 % of real precipitation is suggested.

Global shortwave radiation (R_s) reflects cloudiness and is therefore an indicator for precipitation during daytime. High R_s indicates sunny weather, while precipitation is far more likely at low R_s values. The model snowpack (Lehning et al., 2002) includes a threshold of $250\,\mathrm{W\,m^{-2}}$ global shortwave radiation for snowfall events. At the Matsch test site (Table 1) the average R_s measured during precipitation was $124\,\mathrm{W\,m^{-2}}$, while the following maximum R_s were measured during precipitation events: $909\,\mathrm{W\,m^{-2}}$ during rainfall and $808\,\mathrm{W\,m^{-2}}$ during snowfall. The higher maxima for rainfall occur especially at the beginning of thunderstorms, when the sky is only partially covered by clouds. Based on the analysis of the dataset (Table 1), a threshold of $400\,\mathrm{W\,m^{-2}}$ is suggested as it misses only 2.2 % of precipitation.

Besides the relation to cloudiness, with increasing R_s air temperature also rises, which influences the signal of the ultrasonic snow height sensor. Excluding time periods with high R_s can thus help to reduce the problem of the oscillations of the ultrasonic snow height sensor at least during daytime.

If precipitation is recorded but highly improbable according to the applied meteorological filtering with RH and R_s, it is assumed to be melting snow in the rain gauge. An additional condition to identify such spurious snowmelt is an air temperature around the freezing point and high solar radiation to enable the melting process. These spurious precipitation data were excluded from the final data series.

3.3 Separation of snowfall and rainfall

At test sites with only snow height sensors and unheated rain gauges, liquid precipitation is measured by the rain gauge, while the SWE can be calculated from snow height increments and snow density (see 3. in Fig. 1). Rapidly changing meteorological conditions and fluctuations of the snow line can be taken into account by using sub-daily data to distinguish between time periods with possible rainfall or snowfall. Wet bulb temperature (T_w) was used as an indicator for the snow line, as Steinacker (1983) found it to work better than air temperature (T_a) and other possible methods. Dew point temperature (T_d) was considered less suitable than T_a by Feiccabrino et al. (2007). Following Steinacker (1983), a threshold of $T_w = 1\,^{\circ}\mathrm{C}$ was chosen. For periods with $T_w < 1\,^{\circ}\mathrm{C}$, the snow height sensor was checked for possible snowfall. If snow height increased by at least 0.1 cm per period, snow density was estimated to calculate the SWE.

T_w, if not measured directly, can be calculated by solving the psychrometric formula (Eq. 1) implicitly (Wittenberg, 2011)

$$e = E\,(T_w) - \gamma\,(T_a - T_w) \tag{1}$$

where e is the vapour pressure in the air, E is the saturation vapour pressure and γ is the psychrometer constant depending on air pressure (Kaspar, 2004). E and e can be derived from relative humidity, air temperature and pressure (Murray, 1967).

For periods with $T_w \geq 1\,^{\circ}\mathrm{C}$, data records from the unheated rain gauge were classified as true precipitation. For practical reasons and to keep the presented method simple, a fixed threshold temperature with no transition interval was chosen. Therefore a mixture of rainfall and snowfall in the same period cannot be represented. Such events may be underestimated, since only one record – either snow height increment or rain was considered. A sensitivity analysis of method performance with respect to the choice of the T_w threshold showed no significant variation, thus the literature value of $T_w < 1\,^{\circ}\mathrm{C}$ (Steinacker, 1983) was used.

Snow density was calculated for all snow height increments that were identified as "true" snowfall events (Sects. 3.2, 3.3). Different methods to calculate snow density were considered, e.g. the simple assumption of constant freshly fallen snow density equal to 100 $\mathrm{kg\,m^{-3}}$ (Judson and Doesken, 2000), four methods presented by Brazenec (2005) based on air temperature only and the method of Jordan et al. (1999) based on

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air temperature and wind speed. The influences of these different methods to calculate snow density on the final result and a comparison with measured snow density at the Stubai site are given in the results section.

4 Results and discussion

Below we present an evaluation of ESOLIP, using data from the field sites in Stubai and Matsch.

First, a sensitivity analysis of the method's most important parameters: (a) the choice of the averaging interval for filtering the snow height sensor measurements, both at the event- and seasonal time-scale; (b) the calculation of snow density and its impact on seasonal SWE totals. Second, results from ESOLIP are compared to SWE measured with the snow pillow and a heated rain gauge for single events and at seasonal time scale for the Stubai field site. Moreover, at the Matsch test site, the reliability of snow height filtering was checked, as well as the impact of metadata availability on the approach.

For the validation (winter 2009–2010, 2010–2011 and 2011–2012) of ESOLIP at the Stubai test site, the snow pillow was taken as reference for snowfall events. Because of a lack of metadata, the possibility of precipitation was estimated through the thresholds of relative humidity and global shortwave radiation only. If meteorological conditions allowed, snow precipitation (Sects. 3.2, 3.3), SWE increments measured by the snow pillow were taken as valid observations and compared to the performance of the calculated SWE using different freshly fallen snow densities.

As only time periods with a meteorological possibility of snowfall were taken into account, but heated rain gauges need some time to melt snow after an event, a direct comparison of event-based data is not possible. In fact, a comparison of heated rain gauge measurements and the calculated SWE for longer periods reveal an under-catch of heated rain gauges also known in literature (Goodison et al., 1998).

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4.1 Sensitivity analysis

4.1.1 Snow height filtering

In the literature different moving average intervals from 3 h up to 7 h were found (Brazenec, 2005; Terzago et al., 2012). In Figs. 2 and 3 the reduction of noise by using different moving averages compared to the original hourly data is illustrated using two typical events as examples. The first event analyzed (Fig. 2) refers to the site in Matsch and is a typical spring snowfall event, where the snow height measured by the ultrasonic sensor is disturbed by the daily oscillations in air temperature. The second event (Fig. 2) in Stubai is a typical light winter snowfall event, where the ultrasonic signal is disturbed by random noise, followed by a sunny day showing a snow height signal disturbed by daily oscillation in air temperature. The effects of applying a moving average are notably on both test sites. While the 3 h moving average tends to reproduce some noise of the original data, the 5 h and 7 h moving averages are quite similar and seem to be free of noise. However, the effect of smoothing by moving averages on the sum of snow increments is clearly evident and depends on the noise in the original signal. For the 3 h moving average reduction rates up to 40 %, for the 5 h moving average up to 50 % and for the 7 h moving average up to 60 % were found (Table 2). Due to snow crystal modifications in the snow pack, snow-compaction (settlement) occurs, especially in the first hours after the snowfall (Judson and Doesken, 2000). When comparing sub-daily automatically and daily manually measured snow height, such compaction/settlement has to be considered. A daily observation is likely to underestimate up to about 20 % the snow height compared to the sum of increments at sub-daily time resolution. At the test site in Matsch daily manual snow height data is present and used for the choice of the moving average interval. To use the 5 h moving average seems to be the best compromise, as the 5 h moving average is closer to reality than a 7 h moving average, smoothens peaks less and does not reproduce the noise of the original signal (Figs. 2 and 3).

A reduction of the sum of snow height with applied moving averages becomes evident also when applied over the whole winter season for of all snow increments > 0.1 cm in one time step. Table 3 shows the difference in snow height between the original readout and data corrected with precipitation possibility. Compared to the seasonal sum of daily observations at the test site in Matsch the sum of all snow increments is expected to be higher due to (a) snow compaction/settlement and (b) events where snowmelt started immediately after snowfall and less or even no snow height was registered by the observer. At the test site in Matsch the seasonal sum of the original hourly readout of periods where snowfall is possible has decreased by 46 % (season 2009–2010) resp. 63 % (season 2010–2011) with the use of a 5 h moving average. Differences between the two winter seasons are likely due to different meteorological conditions causing less temperature-related oscillations of the snow height sensor. The observer underestimated snowfall by 35 % in both years compared to the 5 h moving average. If metadata information is also considered, the original readout decreases by 44 % (season 2009–2010) and 61 % (season 2010–2011) respectively. The observer underestimates snowfall of 18–21 % compared to the 5 h moving average (Table 3). At the test site in Stubai the use of a 5 h moving average also lead to a reduction in the same dimension (Table 3). For those reasons, the choice of a 5 h moving average seems to be the best assumption for both our case studies.

4.1.2 Snow density calculation

For the Stubai test site snow densities were calculated with four methods suggested by Brazenec (2005), depending only on T_a [°C] (Eqs. 2–5), and the method of Jordan

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et al. (1999), depending on T_a [K] and wind speed (U) (Eqs. 6 and 7):

$$\rho_{\text{Hedstrom–Pomeroy (1998)}} = 67.92 + 51.25e^{\left(\frac{T_a}{2.59}\right)} \tag{2}$$

$$\rho_{\text{Diamond–Lowry (1954)}} = 119 + 6.48T_a \tag{3}$$

$$\rho_{\text{LaChapelle (1962)}} = 50 + 1.7(T_a + 15)^{1.5} \tag{4}$$

$$\begin{aligned}\rho_{\text{Fassnacht–Soulis (2002)}} = 85\big[&(1 - 0.03\cos(0.33T_a + 0.418)) + 0.15\cos(0.662T_a + 0.418) \\ &- 0.029\cos(0.993T_a + 0.418) + 0.123\sin(0.331T_a + 0.418) + 0.009\sin\big(0.662T_a \\ &+ 0.418\big) - 0.026\sin(0.993T_a + 0.418)\big](-1.75) + 1\end{aligned} \tag{5}$$

$$\rho_{\text{Jordan et al. (1999)}} = 500\left[1 - 0.951e^{\left(-1.4(278.15-T_a)^{-1.15}\right)} - 0.008U^{1.7}\right] \tag{6}$$

for $260.15 < T_a \leq 275.65\,\text{K}$

$$\rho_{\text{Jordan et al. (1999)}} = 500\left[1 - 0.904e^{\left(-0.008U^{1.7}\right)}\right] \tag{7}$$

for $T_a \leq 260.15\,\text{K}$

In contrast to assuming a constant density of freshly fallen snow of $100\,\text{kg}\,\text{m}^{-3}$, applying these equations accounts for the dependence of snow density on temperature (and wind) and even includes density changes during single snowfall events. The five methods for all snowfall events in three winter seasons were tested at the test site in Stubai. Some of the density equations have intrinsic minima and maxima and some averages differ significantly from a constant density of freshly fallen snow of $100\,\text{kg}\,\text{m}^{-3}$ (Fig. 4). The methods of Jordan et al. (1999) and Hedstrom and Pomeroy (1998) can reproduce the wide range of density of freshly fallen snow, and seasonal averages are close to the constant density of $100\,\text{kg}\,\text{m}^{-3}$. The method of Diamond and Lowry (1954) showed similar average values to a constant density of $100\,\text{kg}\,\text{m}^{-3}$, but the range of maxima and minima did not satisfy the requirements for SWE calculation of single events. The methods of LaChapelle (1962) and Fassnacht and Soulis (2002) showed a higher average compared with a constant density of $100\,\text{kg}\,\text{m}^{-3}$ and the range of possible minima and maxima again did not satisfy the requirements to estimate density of freshly fallen snow for single events (Fig. 4). Methods considering only the air temperature are performing similarly to the method based on wind and air temperature, indicating that wind speed is not that relevant for the chosen test sites when calculating the density of freshly fallen snow at such short time scales. For further analysis we focused on the methods of Jordan et al. (1999) and Hedstrom and Pomeroy (1998), as these two fulfilled the requirements to estimate the density of freshly fallen snow for single events.

4.1.3 SWE measurement and calculation methods

To validate estimations of SWE, snow pillow and heated rain gauge measurements at the Stubai field site were compared to results of seasonal SWE based on three different methods to calculate snow density: (1) constant snow density of 100 $\mathrm{kg\,m^{-3}}$, (2) Hedstrom and Pomeroy (1998) and (3) Jordan et al. (1999). For all SWE calculations a 5 h moving average filter was used for the snow height sensor. Time periods in which no snowfall was possible according to meteorological data were excluded. The constant density of 100 $\mathrm{kg\,m^{-3}}$ gives a rough estimate of SWE. However, due to the large variability in the density of freshly fallen snow (Judson and Doesken, 2000), using a variable density performs better than the snow pillow (Table 4). The strong undercatch of heated rain gauge measurement is in agreement with previous studies (Sevruk et al., 2009).

4.2 Comparison of the estimate to snow pillow and heated rain gauge

4.2.1 Modeling single events

Below the performance of ESOLIP is compared to measurements with a snow pillow and a heated rain gauge for two typical precipitation events. The first event is a winter snowfall event with little snow (Fig. 5, T_w −5 °C), the second one is a mixed event with both snowfall and rainfall (Fig. 6, T_w decreasing from > 1 to 0.5 °C) characteristic for

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spring conditions. ESOLIP shows a good match with the snow pillow in both timing and sum of event precipitation for both events. Especially if compared to the heated rain gauge, the timing of modeled precipitation is significantly better for both events.

The winter snowfall event was slightly underestimated. While the snow pillow registered a total sum of 2.6 mm for the snowfall event, ESOLIP showed a total of 2.3 mm (for both tested snow densities). The heated rain gauge registered only 1.0 mm. The earlier start and end of the events for ESOLIP is related to the use of a 5 h moving average (Fig. 5).

In the second sample precipitation event, the first two hours with $T_w > 1\,^{\circ}\mathrm{C}$ showed rainfall, followed by two hours during which only the heated rain gauge reported precipitation (T_w decreasing from 1 to 0.5 °C) and then, after 23:00 LT, six hours with snowfall as indicated by the snow pillow signal. In terms of precipitation timing, the proposed approach corresponds quite well with the snow pillow. However, using a 5 h moving average smoothes the precipitation at 02:00 LT. In terms of total precipitation, the approach gives similar results to both snow pillow and heated rain gauge. The snow pillow registered a total of 8.9 mm SWE (rainfall at the start of the event was not registered as the snow pillow was not covered by snow), ESOLIP showed a total of 8.1 mm (both rain and snow, for both tested snow densities) and the heated rain gauge registered 8.4 mm of total precipitation (Fig. 6). In this example an under-catch of both heated rain gauge and approach is likely, because the initial rain was not registered by the snow pillow. However, the under-catch of the heated rain gauge was not as big as expected, which is probably due to warmer temperatures.

4.2.2 Modeling winter season

In Fig. 7 and in Table 4 the SWE and the total precipitation estimated with ESOLIP are compared on a seasonal timescale with heated rain gauge and snow pillow records for the Stubai site. While the season totals estimated with ESOLIP (with both density calculation methods) correspond very well with the snow pillow, overestimates during some periods became evident. The use of a constant density of freshly fallen snow of 100 kg m^{-3} resulted in an overestimate for the whole winter period. When comparing different measurement/calculation methods for snow precipitation to snow pillow measurements for a whole winter period (event identification according to Sects. 3.2 and 3.3) a strong underestimate of heated rain gauge measurements became evident (Fig. 7 – left, Table 4).

To improve the understanding of measurement errors of heated rain gauges during winter periods, all registered events – including probable melting water – were taken into account. Rainfall recordings from the unheated rain gauge were added to SWE calculated with different snow densities (event identification according to Sects. 3.2 and 3.3) (Fig. 7 – right). Such a comparison is only possible when there is snow on the snow pillow and no rainfall water is running off it. Results were similar to findings for only snowfall events for the whole winter period. Although melting water of the heated rain gauge was included in the seasonal comparison of total precipitation, the performance of the heated rain gauge remained poor (Fig. 7, Table 4). Such results confirm once more literature findings (Sevruk et al., 2009). Only at the start and the end of the season can the heated rain gauge provide reliable precipitation information. This is probably due to the fact that higher temperatures allow an immediate start of the melting process. Considering the snow pillow as reference in all seasons, ESOLIP, with the density equations by Jordan et al. (1999) and Hedstrom and Pomeroy (1998), corresponded better to the snow pillow than using a constant density of freshly fallen snow of 100 kg m^{-3} in each season. Especially in the seasons 2010–2011 and 2011–2012, differences were small (Table 4). In contrast, the use of the constant density of freshly fallen snow of 100 kg m^{-3} tended to overestimate seasonal precipitation in all seasons up to 30 %.

5 Conclusions

In order to improve the estimate of winter precipitation, especially in small, under-equipped Alpine catchments, a new approach to estimate solid and liquid precipitation

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(ESOLIP) was developed. SWE was calculated at a sub-daily timescale from snow height measurements and the density of freshly fallen snow. Differentiation of rainfall from snowfall was easier on a sub-daily timescale, providing an improved determination of precipitation possibility and giving better information on the snow line. It was demonstrated that original readouts from the snow height sensor need filtering to give reasonable snow heights and a moving average of 5 h delivered best performance. However, this causes a loss of time resolution at the single event scale. Several methods to estimate snow density were tested. While a constant density (i.e. $100\,\mathrm{kg\,m^{-3}}$) might produce reasonable results for a rough seasonal estimate, a variable density calculation method (i.e. based on air temperature and wind speed) is recommended for single events. Compared to measurement with a snow pillow, the performance of ESOLIP, using a variable snow density, was promising and far better than the use of a heated rain gauge to measure SWE because of the strong undercatch of the heated rain gauge. Since both measurement devices are rather expensive and have quite high power requirements, ESOLIP, combining a snow height sensor, micrometeorological observations of global shortwave radiation, relative humidity and air temperature with an unheated rain gauge is a cost-effective alternative. Thus a larger number of sites might be installed in a catchment or region, improving the estimate of spatial variability of precipitation.

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References

Allamano, P., Claps, P., and Laio, F.: An analytical model to the effects of catchment elevation on the flood frequency distribution, Water Resour. Res., 45, 1–12, 2009.

Armstrong, R. L. and Brun, E. (Eds.): Snow and Climate, Cambridge University Press, Cambridge, 222 pp., 2008.

Brazenec, W. A.: Evaluation of ultrasonic snow depth sensors for automated surface observing systems (ASOS), M.S. thesis, Colorado State University Fort Collins, Colorado, 124 pp., 2005.

Carturan, L., Dalla Fontana, G., and Borga, M.: Estimation of winter precipitation in a high-altitude catchment of the eastern Italian Alps: validation by means of glacier mass balance observations, Geogr. Fis. Dinam. Quat., 35, 37–48, doi:10.4461/GFDQ.2012.35.4, 2012.

Diamond, M. and Lowry, W. P.: Correlation of density of new snow with 700-millibar temperature, J. Hydrol., 11, 512–513, 1954.

Egli, L., Jonas, T., and Meister, R.: Comparison of different automatic methods for estimating snow water equivalent, Cold Reg. Sci. Technol., 57, 107–115, doi:10.1016/j.coldregions.2009.02.008, 2009.

Egli, L., Jonas, T., and Bettems, J.: A virtual network for estimating daily new snow water equivalent and snow depth in the Swiss Alps, Ann. Glaciol., 51, 32–38, 2010.

Fassnacht, S. R. and Soulis, E. D.: Implications during transitional periods of improvements to the snow processes in the land surface scheme – hydrological model WATCLASS, Atmos. Ocean., 40, 389–403, 2002.

Feiccabrino, J. and Lundberg, A.: Precipitation phase discrimination by dew point and air temperature, Western Snow Conference, Kailua-Kona, Hawaii, 16–19 April 2007, available at: http://www.westernsnowconference.org/sites/westernsnowconference.org/PDFs/2007Feiccabrino.pdf, 2007.

Goodison, B. E., Louie, P. Y. T., and Yang, D.: WMO solid precipitation measurement intercomparison, WMO/TD-No. 872, Instruments and Observing methods report 67, 212 pp., 1998.

Gottardi, F., Obled, C., Gailhard, J., and Paquet, E.: Statistical reanalysis of precipitation fields based on ground network data and weather patterns: application over French mountains. J. Hydrol., 432–433, 154–167, doi:10.1016/j.jhydrol.2012.02.014, 2012.

Häckel, H.: Meteorologie, 6th Edn., Ulmer Verlag, Stuttgart, Germany, 447 pp., 2005.

Hedstrom, N. R. and Pomeroy, J. W.: Measurements and modelling of snow interception in the boreal forest, Hydrol. Process., 12, 1611–1625, 1998.

Hydrographic Office of the Autonomous Province Bozen: http://www.provinz.bz.it/wetter/climareport.asp (last access: 8 April 2013).

Jonas, T., Marty, C., and Magnusson, J.: Estimating the snow water equivalent from snow depth measurements in the Swiss Alps, J. Hydrol., 378, 161–167, doi:10.1016/j.jhydrol.2009.09.021, 2009.

Jordan, R. E., Andreas, E. L., and Makshtas, A. P.: Heat budget of snow-covered sea ice at North Pole 4, J. Geophys. Res., 104, 7785–7806, doi:10.1029/1999JC900011, 1999.

Judson, A. and Doesken, N.: Density of freshly fallen snow in the central Rocky Mountains, B. Am. Meteorol. Soc., 81, 1577–1587, doi:10.1175/1520-0477(2000)081<1577:DOFFSI>2.3.CO;2, 2000.

Kaspar, F.: Entwicklung und Unsicherheitsanalyse eines globalen hydrologischen Modells, Ph. D. thesis, Universität Kassel, Kassel, Germany, 129 pp., 2004.

LaChapelle, E. R.: The density distribution of new snow, USDA Forest Service, Alta Avalanche Study Center, Project F, Progress Rep. 2, Salt Lake City, UT, 13 pp., 1962.

Lehning, M., Bartelt, P., Brown, B., and Fierz, C.: A physical SNOWPACK model for the Swiss avalanche warning. Part III: Meteorological forcing, thin layer formation and evaluation, Cold Reg. Sci. Technol., 35, 169–184, 2002.

Murray, F. W.: On the computation of saturation vapor pressure, J. Appl. Meteorol., 6, 203–204, 1967.

Rohregger, J. B.: Methoden zur Bestimmung der Schneefallgrenze, M. S. thesis, Universität Wien, Wien, Austria, 78 pp., 2008.

Ryan, W. A., Doesken, N. J., and Fassnacht, S. R.: Evaluation of ultrasonic snow depth sensors for US snow measurements, J. Atmos. Ocean. Tech., 25, 667–684, doi:10.1175/2007JTECHA947.1, 2008.

Scherrer, S.: Die Niederschlagstotalisatoren der Schweiz – eine basisklimatologische Netz-analyse und Bestimmung besonders wertvoller Stationen, Arbeitsberichte der MeteoSchweiz, Zürich, 231, 35 pp., 2010.

Sevruk, B.: Conversion of snowfall depths to water equivalents in the Swiss Alps, in: Proc. Workshop on the Correction of Precipitation Measurements, edited by: Sevruk, B., Zürcher Geographische Schriften, ETH Zürich, 23, 81–88, 1986 (see WMO/TD-No 104, 1985), Zürich, 1–3 April, 1985.

Sevruk, B., Ondrás, M., and Chvíla, B.: The WMO precipitation measurement intercomparisons, Atmos. Res., 92, 376–380, doi:10.1016/j.atmosres.2009.01.016, 2009.
Steinacker, R.: Diagnose und Prognose der Schneefallgrenze, Wetter Leben, 35, 81–90, 1983.
Sturm, M., Taras, B., Liston, G. E., Derksen, C., Jonas, T., and Lea, J.: Estimating snow water equivalent using snow depth data and climate classes, J. Hydrometeorol., 11, 1380–1394, doi:10.1175/2010JHM1202.1, 2010.
Symader, W.: Was passiert, wenn der Regen fällt?, Ulmer Verlag, Stuttgart, Germany, 256 pp., 2004.
Terzago, S., Prola, M. C., Fratianni, S., Cremonini, R., and Barbero, S.: Validazione di misure nivometriche automatiche, Neve e Valanghe, 75, 36–43, 2012.
US Army Corps of Engineers: Snow hydrology, Summary report of the snow investigations, technical report, US Army Corps of Engineers, North Pacific Division, Portland, OR, 437 pp., 1956.
Wittenberg, H.: Praktische Hydrologie: Grundlagen und Übungen, Vieweg+Teubner Verlag, Springer Fachmedien, Wiesbaden, Germany, 116 pp., 2011.
Zahumenskỳ, I.: Guidelines on Quality Control Procedures for Data from Automatic Weather Stations, World Meteorological Organisation, available at: http://www.wmo.int/pages/prog/www/IMOP/publications/IOM-82-TECO_2005/Papers/3(14)_Slovakia_2_Zahumensky.pdf (last access: 8 April 2013), 2004.

Table 1. Relative humidity (RH) and global radiation (R_s) values during precipitation events (Matsch test site).

During rainfall	lowest RH values	35 %
	highest R_s values	909 $\mathrm{W\,m^{-2}}$
During snowfall	lowest RH values	39 %
	highest R_s values	808 $\mathrm{W\,m^{-2}}$
Average values during precipitation	RH	93 %
	R_s	124 $\mathrm{W\,m^{-2}}$
Missed yearly precipitation amount with a threshold of:	RH > 70 %	5.9 %
	RH > 50 %	1.4 %
	$R_s < 250\,\mathrm{W\,m^{-2}}$	4.3 %
	$R_s < 400\,\mathrm{W\,m^{-2}}$	2.2 %
	$R_s < 250\,\mathrm{W\,m^{-2}}$ and RH > 70 %	1.8 %
	$R_s < 400\,\mathrm{W\,m^{-2}}$ and RH > 50 %	0.5 %

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Table 2. Daily sums of snow height increments [cm] during a snowfall event at the Matsch and Stubai test sites. The original (hourly) data, data filtered by moving averages and daily manual measurement of the observer, if available, are reported (see also Figs. 2 and 3).

Field site	Snow height increments [cm]	Original	Moving average			Observer
			3 h	5 h	7 h	
Matsch	9 Mar 2010 09:00–10 Mar 2010 09:00	20.0	17.6	16.7	15.0	14
Matsch	10 Mar 2010 09:00–11 Mar 2010 09:00	7.4	4.5	3.7	2.8	3
Stubai	5 Feb 2010 23:00–7 Feb 2010 12:00	28.3	16.5	13.2	11.9	–

Table 3. Sums of seasonal snow height increment [cm] at the test sites in Matsch and Stubai with or without meteorological filtering (snow height increments are considered only if $R_s < 400\,\mathrm{W\,m^{-2}}$ and RH > 50 %) and with consideration of metadata (meteorological reports, observations etc.) for the original (hourly) readout and three different moving averages, as well as the seasonal sum of daily new snow height measured by the observer.

Test site	Winter season	Meteorological filtering	Original hourly data	Moving average 3 h	5 h	7 h	Observer
Matsch	2009–2010	none	497	318	258	216	112
		RH and R_s thresholds	318	207	173	153	
		RH and R_s thresholds, metadata	252	168	142	126	
	2010–2011	none	845	402	274	215	81
		RH and R_s thresholds	336	176	125	99	
		RH and R_s thresholds, metadata	253	137	99	81	
Stubai	2009–2010	none	959	593	481	414	–
		RH and R_s thresholds	758	496	411	361	
	2010–2011	none	1089	654	509	435	
		RH and R_s thresholds	820	517	412	356	
	2011–2012	none	1170	766	646	586	
		RH and R_s thresholds	800	572	499	459	

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Table 4. Seasonal sums of only SWE [mm] and total precipitation [mm] using different measurement or calculation methods at the test site in Stubai. The differences to the snow pillow values (%) are given in brackets.

Winter season	Measurement device		ESOLIP with used density method to calculate SWE		
	Snow pillow	Heated rain gauge	Constant density of freshly fallen snow (100 kg m^{-3})	Jordan et al. (1999)	Hedstrom and Pomeroy (1998)
			SWE		
2009–2010	288	176 (61 %)	411 (142 %)	378 (131 %)	379 (131 %)
2010–2011	354	96 (27 %)	393 (111 %)	353 (100 %)	366 (103 %)
2011–2012	501	109 (22 %)	475 (95 %)	501 (100 %)	475 (95 %)
			Total precipitation		
2009–2010	349	211 (60 %)	460 (132 %)	427 (123 %)	428 (122 %)
2010–2011	471	167 (35 %)	535 (114 %)	488 (104 %)	501 (106 %)
2011–2012	619	258 (42 %)	671 (108 %)	674 (109 %)	647 (104 %)

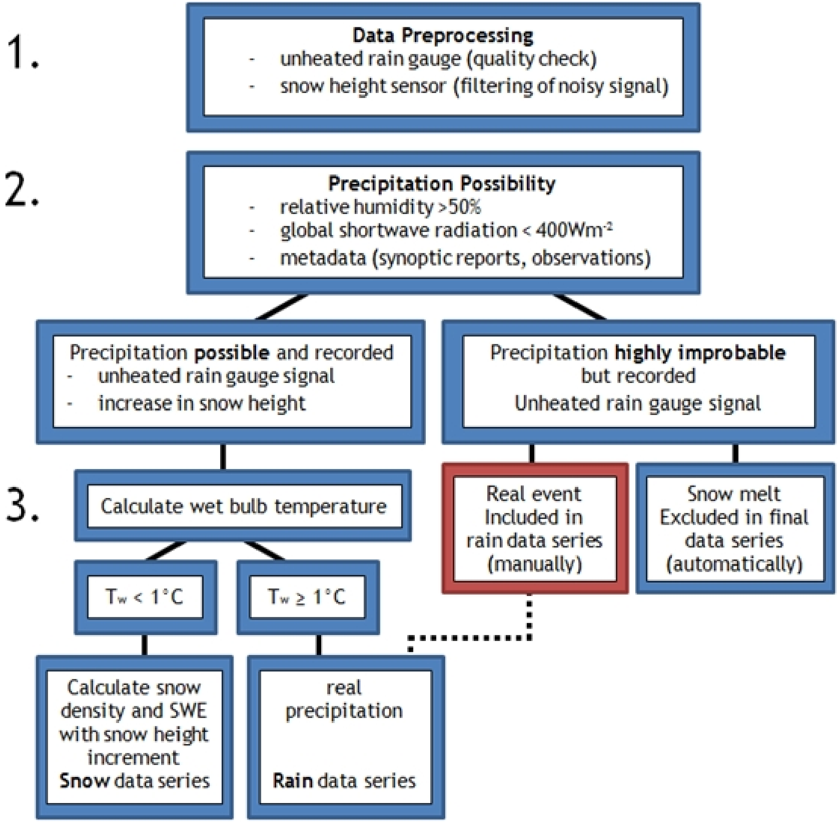

Fig. 1. Scheme of the proposed methodology to estimate solid and liquid precipitation (ESOLIP).

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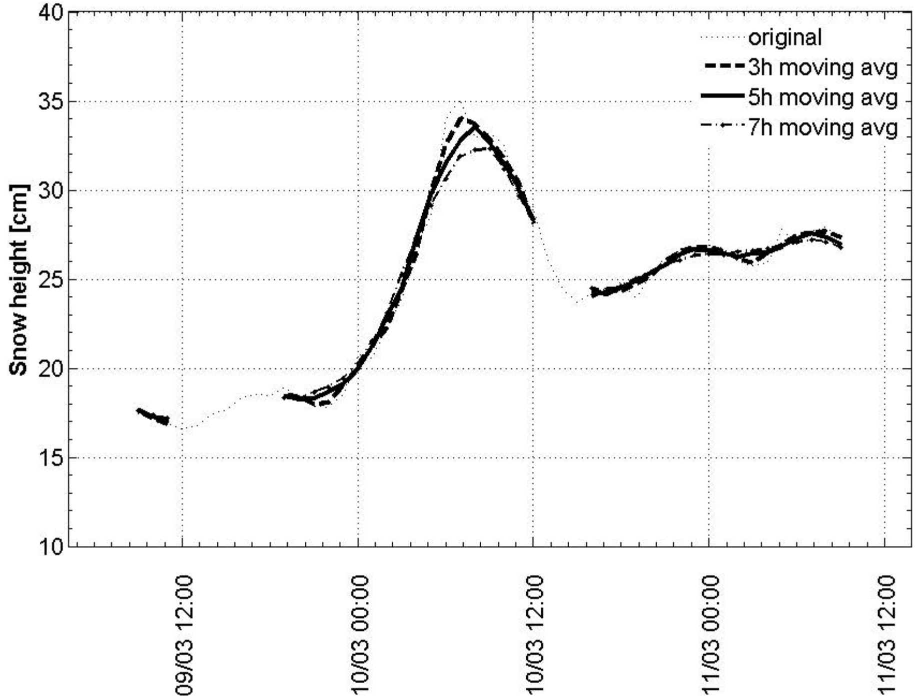

Fig. 2. Comparison of different moving average (avg) intervals for snowfall events to reduce the noise in the snow height signal for a period with snowfall at the Matsch test site (see also Table 2).

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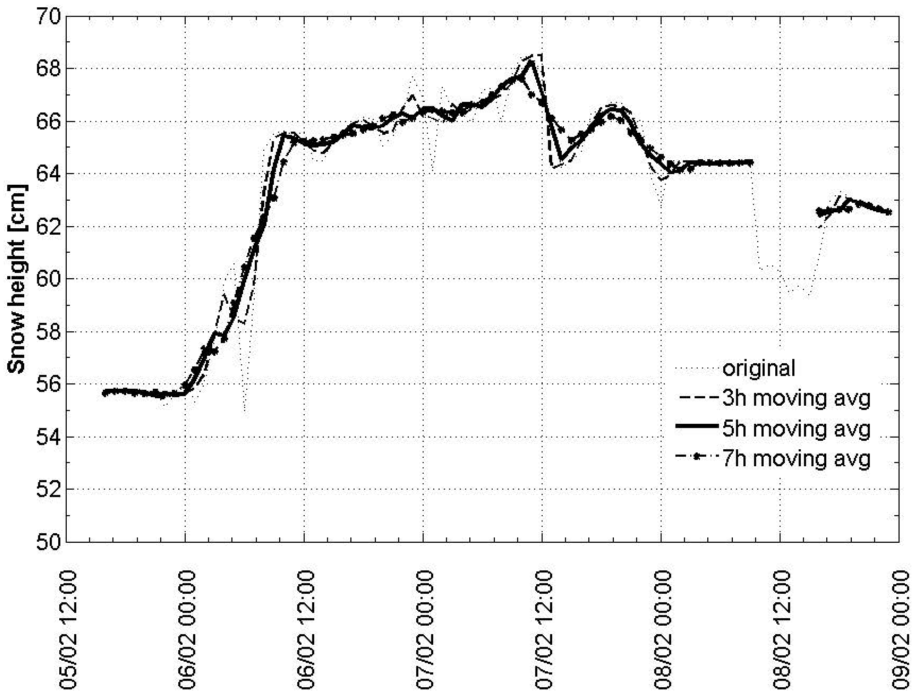

Fig. 3. Comparison of different moving average (avg) intervals to reduce the noise in the snow height signal for a snowfall period followed by sunny weather at the Stubai test site. The moving averages were applied only if weather conditions allowed snowfall (see also Table 2).

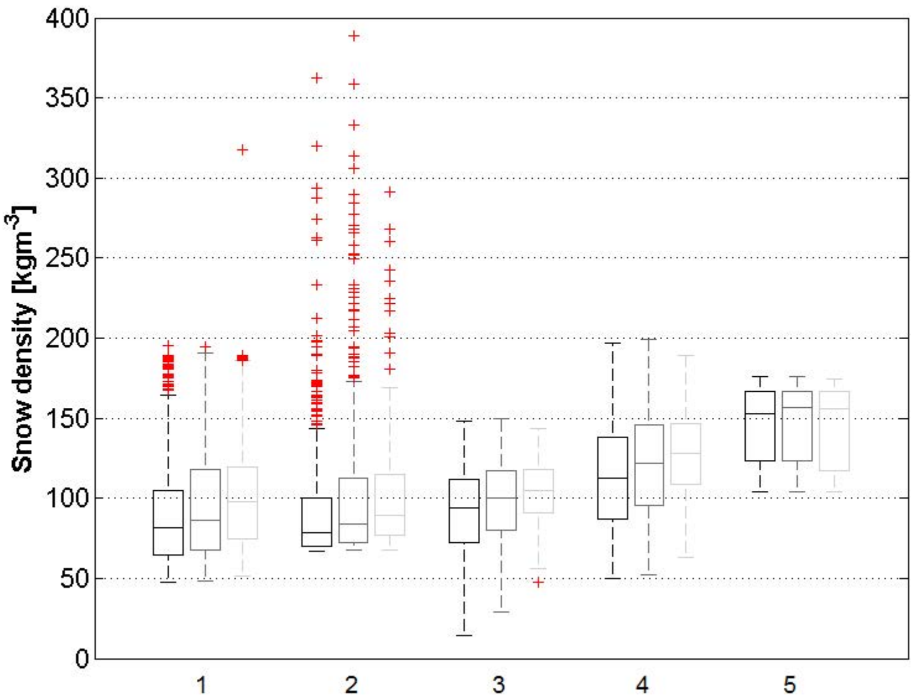

Fig. 4. Box plot of the tested snow density in the three winter seasons 2009–2010, 2010–2011, 2011–2012, represented in black, dark grey and light grey respectively. The numbers refer to the methods following (1) Jordan et al. (1999) (2) Hedstrom and Pomeroy (1998) (3) Diamond and Lowry (1954) (4) LaChapelle (1962) and (5) Fassnacht and Soulis (2002). The width of the box plots indicates 25th and 75th percentiles, the line in the middle of each box is the sample median, while dashed lines extend until 1.5 of the interquartile range. Single snowfall events outside this range are represented with red crosses.

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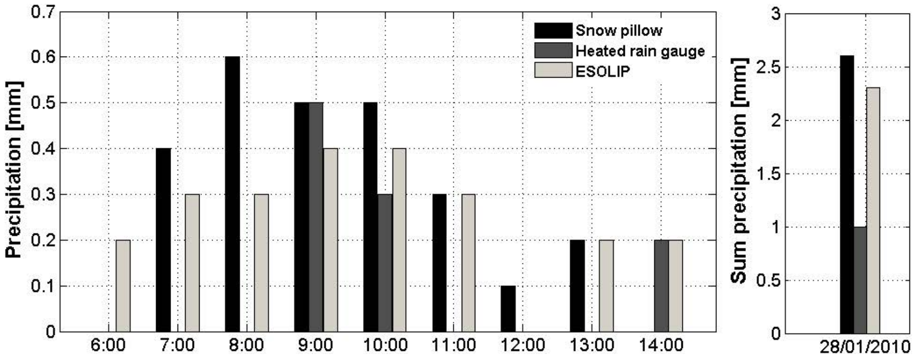

Fig. 5. Comparison of precipitation measured with a snow pillow, a heated rain gauge and ESOLIP for a single snowfall event (T_w −5 °C) at the test site in Stubai: hourly values on the left, daily sum (28 January 2010) on the right.

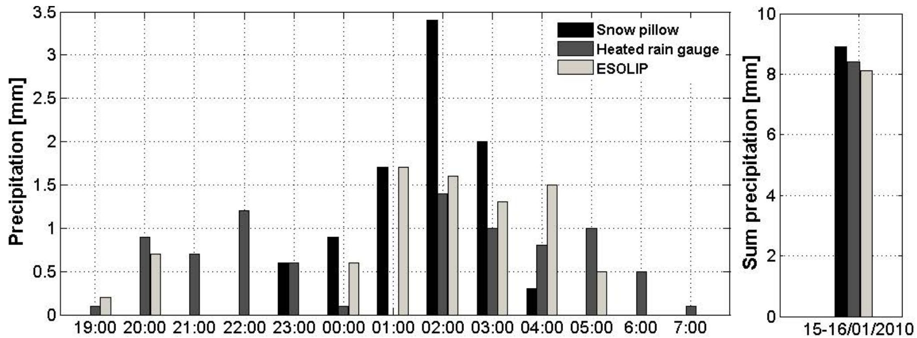

Fig. 6. Comparison of precipitation measured with snow pillow, with a heated rain gauge and ESOLIP for a single mixed snow and rainfall precipitation event (T_w decreasing from 4 to −2.5 °C) at the test site in Stubai: left hourly values, right total sum (15 May 2012).

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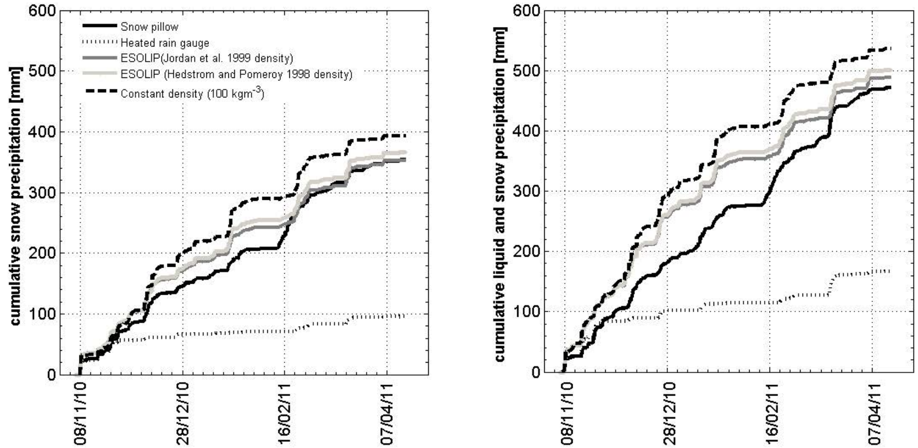

Fig. 7. (left) Comparison of accumulated SWE measured using snow pillow and heated rain gauge and calculated with ESOLIP considering snow height increments and different density calculation methods for the winter period 2010–2011 (Stubai test site). Only periods with possible snowfall are considered, delayed snowmelt in heated rain gauge is left out. (right) Comparison of accumulated rainfall and snowfall events for the winter season 2010–2011. Only the period with a continuous snow cover on the snow pillow is considered to avoid rainfall water running off it. Rainfall registered at the unheated rain gauge is added to SWE estimates. For the heated rain gauge, both rainfall and melting water are included.