

AUTOMATED DRY-SNOWFALL LEVEL CALCULATION FOR IMPROVED ASSESSMENT OF MELT-FREEZE CRUSTS AND ASSOCIATED FACETED WEAK LAYERS

Marie Hofmann^{1,*}, Jil Lehnert^{1,2}, Christoph Mitterer¹

¹*Avalanche Warning Service Tyrol, Innsbruck, Austria*

²*University of Innsbruck, Innsbruck, Austria*

ABSTRACT:

Faceted weak layers relevant for avalanche triggering often develop around melt-freeze crusts. These can - besides a formation through shortwave radiation or sensible heat - originate from moisture input into the snowpack during rain events. Traditionally, avalanche forecasters track such melt-freeze crusts with the help of snow profiles or by examining forecasted and observed snowfall limits. However, these approaches have limitations in spatial coverage and accuracy. This study introduces an approach to facilitate the tracking of melt-freeze crusts. We calculated dry-snowfall levels (DSLs) from automated weather station (AWS) data located in Tyrol (Austria). Unlike conventional snowfall limits, which denote the elevation where 50% of precipitation is snow, DSLs define the elevation where 100% of precipitation is dry snow, thus indicating crust-free zones above. To obtain this information, in a first step, we calculated wet-bulb temperature from air temperature and relative humidity data provided by AWSs. Assuming a threshold wet-bulb temperature of 0.3 °C for the elevation of the DSL we then calculated the respective DSL for the given locations of the AWSs. A first validation of calculated DSLs against observed snowfall limits from the 2023-2024 winter season shows promising alignment. We generally expect observers to report a snowfall limit which is lower (i.e. wetter) than the calculated DSL. In line with this, DSLs were usually higher in elevation than reported snowfall limits. Besides these results, we implemented our work into the operational system of the Avalanche Warning Service Tyrol, including a visual representation of daily maximum DSLs on a map interface. With this, forecasters can now easily assess spatial DSL distributions during specific precipitation events. Future work will focus on refining DSL calculations and enhancing observational practices to further strengthen forecasting accuracy and reliability.

Keywords: Melt-freeze crust, snowfall level, avalanche forecasting, automated weather stations.

1. INTRODUCTION

Dry-snow slab avalanches are the type of avalanche that are most often associated with avalanche fatalities and are thus of major importance for avalanche forecasting (Schweizer and Jamieson, 2001). They are often triggered on a persistent weak layer of faceted crystals with large grain sizes. These soft weak layers frequently form in the vicinity of melt-freeze crusts (Jamieson, 2006), where the freezing of wet snow releases latent heat that enhances the temperature gradient and hence favors enhanced grain growth. Furthermore, the higher thermal conductivity of the crust - especially when compared to the lowered thermal conductivity of the faceted layer - will increase the temperature of the layer above the crust and therefore further enhance the temperature gradient close to the crust thereby accelerating grain growth (Colbeck and Jamieson, 2001).

Detecting and tracking melt-freeze crusts is thus crucial for accurately assessing snowpack stability and avalanche forecasting. Melt-freeze crusts usually form at the snow surface, where the snow becomes wet through either rain or increased energy input at the snow-atmosphere interface (mostly shortwave radiation or air temperatures above the freezing) and then refreezes later on. In this study, we focus on the formation of melt-freeze crusts by rain events and do not deal with melt-freeze crusts that are caused by sensible heat or shortwave radiation. Melt-freeze crusts caused by rain-on-snow events most likely form during early or late winter season (Jamieson, 2006). However with climate change and rising air temperatures, rain up to high elevations is becoming more common also during mid-winter periods and hence melt-freeze crusts could become a problem more frequently during entire seasons in future decades (Mayer et al., 2023).

To get a picture of when and where melt-freeze crusts are forming, avalanche forecasters so far have several tools at hand. The most common approaches are looking at modelled and observed profiles, as well as checking forecasted

*Corresponding author address:
Marie Hofmann, Avalanche Warning Service Tyrol
6020 Innsbruck, Austria;
tel: +43 67762072639
email: marie@hofmanns-mail.de

and observed snowfall limits. However, all these approaches come with their own problems and limitations. Both, observed and modelled profiles are only valid for a certain location, elevation and aspect. It is therefore hard to extrapolate the information of a melt-freeze crust from one or multiple profiles to a general rule describing where crusts might have formed. Modelled profiles additionally depend on the snowfall limit predicted by numerical weather predictions (NWP) models. Especially snowfall limits in mountainous terrain are subject to variability and are known to be challenging to predict (Fehlmann et al., 2018). Another common approach to keep track of crusts forming from rain events is to directly check forecasted and observed snowfall limits. As mentioned above, forecasted snowfall limits come with their own uncertainties. At the moment, observed snowfall limits seem to be the most reliable source for tracking where and when rain crusts form. However, the definition of the snowfall limit itself can vary strongly from observer to observer and snowfall limits are sometimes observed from a distance, meaning that the feedback is both subjective and of varying quality. Another drawback of this approach is that avalanche forecasters depend on the feedback of observers, which can be rare, especially when bad conditions prevail or during off-peak season in late spring and early autumn. This poses a problem, particularly because those are the times when rain crusts are most often observed (Jamieson, 2006). While these approaches give forecasters a rough overview on where and when melt-freeze crusts develop, none of them provides an objective and spatially continuous data set.

The present study aims to fill this gap introducing an approach for detecting elevation thresholds above which the development of melt-freeze crust due to rain-on-snow events is highly unlikely. We present a weather station data driven approach to calculate the snowfall limit in hindsight of a precipitation event. Similar approaches have been used before to distinguish between snow and rain (e.g. Steinacker, 1983). However this is - to our knowledge - the first attempt to implement the approach into the field of operational avalanche forecasting.

The snowfall limit is usually defined as the elevation, where 50% of the precipitation falls as rain and 50% as snow. However, since we aim to find the elevation above which melt-freeze crusts are highly unlikely to form from moisture input through rain, this definition is not suitable for our purpose. Instead, we intend to calculate the elevation, above which 100% of the precipitation falls as snow and is dry (the elevation, above which there is no relevant moisture input

into the snowpack and hence no melt-freeze crusts will develop from that specific precipitation event). We will refer to this definition from now on as the dry-snowfall level (DSL). Figure 1 shows the differences between the conventional snowfall limit and the DSL.

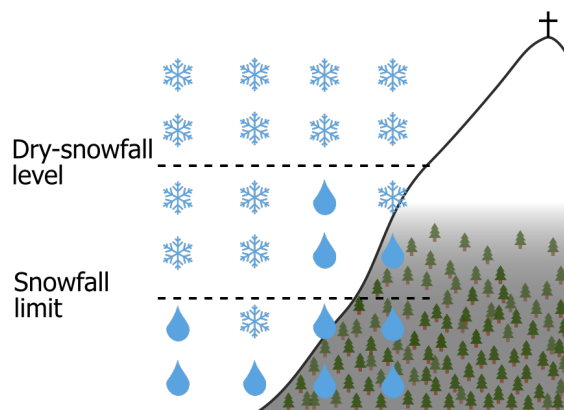


Figure 1: Comparison of dry-snowfall level (= 100% solid precipitation) and snowfall limit (defined as elevation with 50% snow, 50% rain). Dry-snowfall levels are consequently higher in elevation than snowfall limits.

2. DATA

The data set used to calculate DSL values consists of 10-minute recordings of relative humidity and air temperature from 139 automated weather stations (AWS) of the Avalanche Warning Service Tyrol (Austria). The elevation of the AWS sites ranges from 1025 m to 3454 m. For each micro-region of Tyrol (as defined by the Avalanche Warning Service Tyrol for forecasting purposes), we additionally chose a representative weather station that provides precipitation measurements, mostly using weather stations of the Hydrographic Service Tyrol. Figure 2 shows the distribution of both AWSs and precipitation measurements throughout Tyrol.

For a first qualitative validation of the calculated DSL, we used the feedback on observed snowfall limits from Tyrolean observers. The dataset consists of 424 snowfall limits reported throughout the winter season 2023-2024. The data includes official reports from observers working for the Avalanche Warning Service Tyrol, avalanche commissions reports as well as voluntary feedback from back country skiers. Sometimes observers report snowfall limits not during the event but on the next clear day when visibility allows them to report limits from a distance. Comparing these observations to calculated DSL at the time of the observation is not meaningful. Therefore, we excluded these temporally shifted observations from the direct comparison with calculated DSLs. We only used observations where the precipitation measured at the rain gauge of the cor-

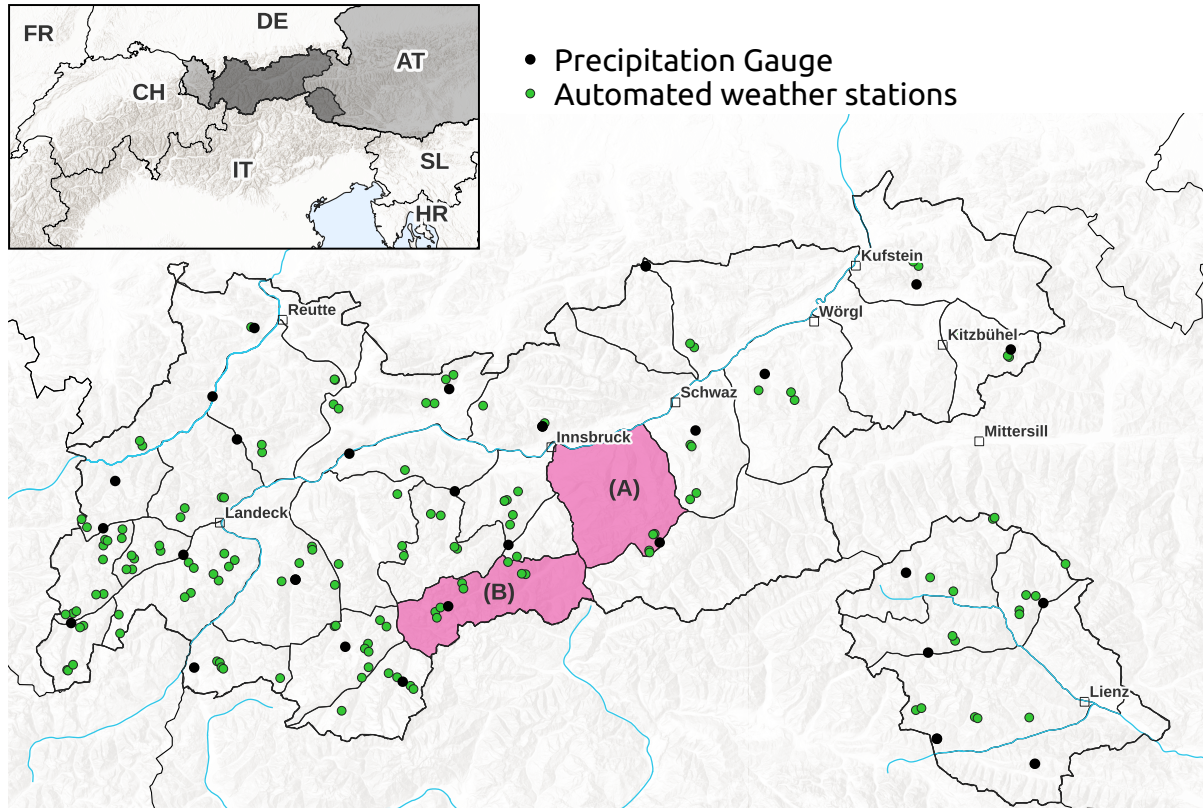


Figure 2: The study region Tyrol is located in western Austria (inlet map, dark grey). Green dots indicate automated weather stations, for which DSLs were calculated. Black dots represent precipitation gauges. For each micro-region, we chose one representative precipitation gauge. Pink shadings represent the two micro-regions (A) Western Tuxer Alps and (B) Central Stubai Alps, for which exemplary DSL plots were included in our study (Figure 3).

responding micro-region was ≥ 0 mm at the time of the observation.

3. METHODS

3.1 Calculation of dry-snowfall level (DSL)

In general, it has been shown that relative humidity (besides air temperature) highly impacts the precipitation phase of hydro-meteors close to the freezing point (e.g. Matsuo and Sasyo, 1981; Froidurot et al., 2014). The latent heat that is released when water vapor evaporates and sublimates from the snowflake surface cools the snowflake and therefore slows the melting process of snowflakes. Matsuo and Sasyo (1981) demonstrated that snowfall can occur with temperatures up to $+4$ °C in relatively unsaturated air ($RH = 50\%$). To distinguish between solid and liquid precipitation it is therefore common to use the wet-bulb temperature (T_W) instead of dry-air temperature measurements. The wet-bulb temperature measures the lowest temperature that can be achieved through the evaporation of water from an object and thus combines information on air temperature as well as humidity.

For this study, we calculated wet-bulb temperatures using the air temperature and relative humidity

data recorded at our AWSs. The data was originally provided with a 10-minute time resolution. We resampled it to an hourly resolution by first calculating the wet-bulb temperatures at 10-minute intervals and then selecting the maximum wet-bulb temperature for each hour. This approach aligns with our goal of providing forecasters with the highest possible DSL, allowing them to monitor the highest elevation where melt-freeze crust formation through moisture input may occur.

For the calculation of the wet-bulb temperature T_W (given in °C) we used an empirical formula, where T_A is the air temperature and RH the relative humidity at each AWS:

$$T_W = -5,806 + 0,672 \cdot T_A - 0,006 \cdot T_A^2 + (0,061 + 0,004 \cdot T_A + 0,000099 \cdot T_A^2) \cdot RH + (-0,000033 - 0,000005 \cdot T_A - 0,000001 \cdot T_A^2) \cdot RH^2 \quad (1)$$

Rohregger (2008) tested different formulas for the calculation of wet-bulb temperature, including the one we used (Eq. 1). He compared calculated wet-bulb temperature values to psychrometric tables for different ranges of air pressure, air temperature and relative humidity and found the formula in Eq.

1 to be the most reliable, especially for air temperature recordings ranging from -7°C to $+5^{\circ}\text{C}$. As we are interested in the transition from snow to rain which happens close to freezing, this air temperature range is the most relevant to our study. In addition, Rohregger (2008) found that Eq. 1 is most reliable when the air is almost saturated (i.e. RH above 90%). Errors were smaller than 0.1°C for air temperatures ranging from -10°C to $+2^{\circ}\text{C}$ and air pressure values of 850 hPa to 990 hPa. In general, errors were positive meaning Eq. 1 overestimates wet-bulb temperature. Multiple studies derived a statistical relationship between wet-bulb temperature and precipitation phases (e.g. Froidurot et al., 2014; Fehlmann et al., 2018; Wang et al., 2019). They usually determine the phase of precipitation (i.e. snow, mixed, rain) by a threshold of the wet-bulb temperature. For example, a common accepted rule of thumb assumes that the snowfall limit is equal to an elevation where $T_W = +1^{\circ}\text{C}$ (Steinacker, 1983). Most of the studies, however, neglect the existence of a transition zone in between solid and liquid precipitation. Their proposed thresholds represent the wet-bulb temperature at which the conditional probability of solid precipitation equals the probability of liquid precipitation, both being 50% (e.g. Sims and Liu, 2015; Steinacker, 1983). Since this study seeks to find the dry-snowfall limit, where a 100% of the precipitation is solid only, these thresholds are not appropriate for our needs.

Fehlmann et al. (2018) included mixed precipitation in their statistical models. They proposed thresholds for the lower and upper boundary of mixed precipitation of $+0.3^{\circ}\text{C}$ and $+0.8^{\circ}\text{C}$, respectively.

For our calculations we used their threshold of $+0.3^{\circ}\text{C}$, assuming dry snowfall above the elevation where $T_W = +0.3^{\circ}\text{C}$ and liquid/mixed precipitation below.

Steinacker (1983) developed a relationship between snowfall limit and wet-bulb temperatures from 5-year climate observations in Innsbruck. We used his formula to compute DSLs from wet-bulb temperatures, but replaced the threshold temperature he used ($T_W = +1^{\circ}\text{C}$) with the one Fehlmann et al. (2018) proposed for dry snowfall ($T_W = +0.3^{\circ}\text{C}$). Using the stations' elevation h in meters and hourly maximum T_W values, we calculated the DSLs as follows:

$$DSL = h + \frac{(T_W - 0.3)}{0.006} \quad (2)$$

3.2 Maximum daily DSL

To provide forecasters a quick overview of the DSLs of a specific precipitation event or day, we furthermore calculated a daily maximum DSL for each AWS and plotted the results on a map interface. While we calculated hourly DSLs for the entire 24 hours of a day (in order to create a continuous plot), the daily maximum DSLs were calculated only for periods where precipitation intensity was $\geq 0.1 \text{ mm h}^{-1}$.

4. RESULTS & DISCUSSION

We present first a qualitative validation of our DSL results, followed by a proposal on how to incorporate our findings under operational constraints of an avalanche warning service.

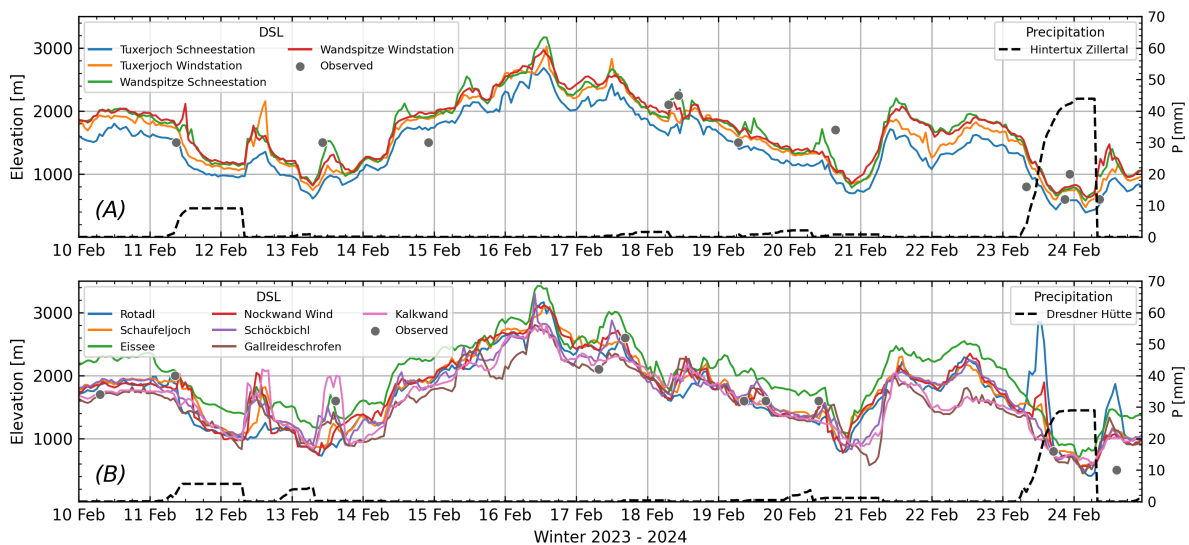


Figure 3: Calculated DSL (colored lines, in meters) and measured precipitation sums (black dashed line, in millimeters) for the two micro-regions (A) Westliche Tuxere Alpen and (B) Zentrale Stubai Alpen during multiple precipitation events in the period of February 10 to 24, 2024. Grey dots depict observed snowfall limits reported by observers.

4.1 Validation of calculated DSLs

For an initial validation of our method, we examined several precipitation events during the 2023-2024 winter season. For each event, we compared calculated DSLs for all available AWSs to the snowfall limits reported by observers. In addition, we compared all AWS-based DSLs to each other for their specific micro-region to estimate the spread.

Figure 3 illustrates an exemplary situation from February 10 to 24 for the two micro-regions Western Tuxer Alps (A) and Central Stubai Alps (B) (cp. pink shaded micro-regions in Figure 2). Figure 3 shows multiple precipitation events (black dashed line) and the corresponding calculated DSL at the AWS (colored lines) as well as observed snowfall limits (grey dots).

The presented period started with a cold front on February 11, during which Geosphere Austria (National Austrian Weather Service) forecasted snowfall limits dropping from 1500 m to about 1100 m. A low-pressure system brought weak precipitation on February 12 and 13. Starting February 14, high pressure dominated the weather in Tyrol, resulting in very mild and dry air masses. Another weak cold front passed on February 17 with snowfall limit forecasted between 1700 m and 2000 m. Finally, a significant cold front on February 23 brought more than 40 mm of precipitation in Region (A) and 30 mm in Region (B). The precipitation was accompanied by a strong drop in air temperature, causing snowfall limits to sink. Geosphere Austria forecasted snowfall limits of 800 m to 1000 m during the day and further drops during the night.

The calculated DSLs generally capture the broader weather events well. During the first cold front passage, DSLs dropped from approximately 1800 m to 1000 m to 1200 m, aligning well with Geosphere Austria's forecasts (not shown in Figure 3). In the following high-pressure phase, DSLs peaked at around 2600 m to 3000 m, reflecting the mild air masses present at that period. During the strong cold front event on February 23, DSLs dropped from about 1500 m to 700 m.

The spread of calculated DSLs between the individual AWS is generally small (≤ 400 m) during precipitation phases (e.g., Region (A) on February 11 and 23, Region (B) on February 11, 12, 20, and 23). However, in both regions, one of the AWS consistently showed an offset compared to the others. In Region (A), AWS *Tuxerjoch Schneestation* displays lower DSL values, while in Region (B), AWS *Eissee* overestimates the DSL, both throughout the entire period. These outliers are likely due to measurement errors at the respective AWS.

The observed offsets of both AWSs prevailed all winter. Ignoring these AWSs, the spread between calculated DSLs during precipitation events ranged from about 50 m to 400 m.

The spread between individual calculated DSLs is generally most pronounced around noon, where some AWSs showed a significant increase in DSL, while others did not (e.g., peak on February 14 in Region (B)). This discrepancy might be due to a lack of ventilation on some of the AWSs. Especially on days with little or no wind activity, air temperature sensors without ventilation might heat up and can thus be prone to measurement deviations (Georges and Kaser, 2002).

Grey dots indicate snowfall limits reported by observers. Generally, the observed snowfall limits and calculated DSLs align well. Especially during the heavy precipitation event on February 23, the differences between observations and calculations are fairly small (~ 200 m or less). One exception occurs in Region (A), where a single observation deviates significantly; however, a second observation recorded almost simultaneously aligns well with our calculations.

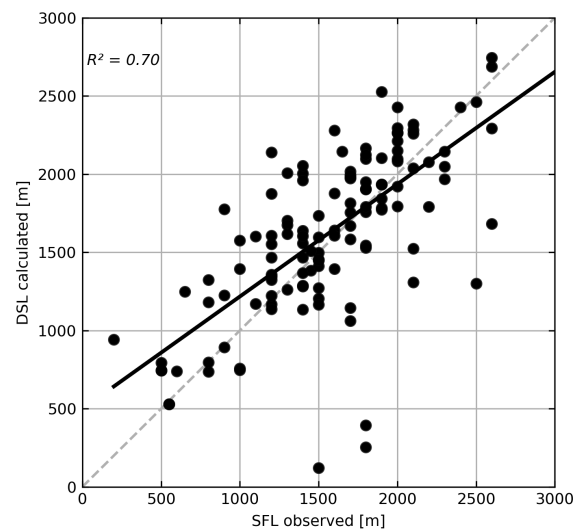


Figure 4: Comparison of calculated DSLs and observed snowfall limits ($N = 124$). Reported snowfall limits (SFL) were used only, if the precipitation measured at the respective rain gauge of the micro-region was ≥ 0 mm at the time of the observation. The black line indicates the linear regression between calculated DSLs and observed SFLs, with a Pearson correlation coefficient (R^2) of 0.7. The grey dashed line represents perfect correlation

Although observations and DSLs generally agree fairly well, there are obviously also differences and outliers. This might be due to various reasons which we try to address in the following: Observers often report snowfall limits from a distance, and sometimes not during the event (since there is usually poor or no visibility) but in hind-cast, when

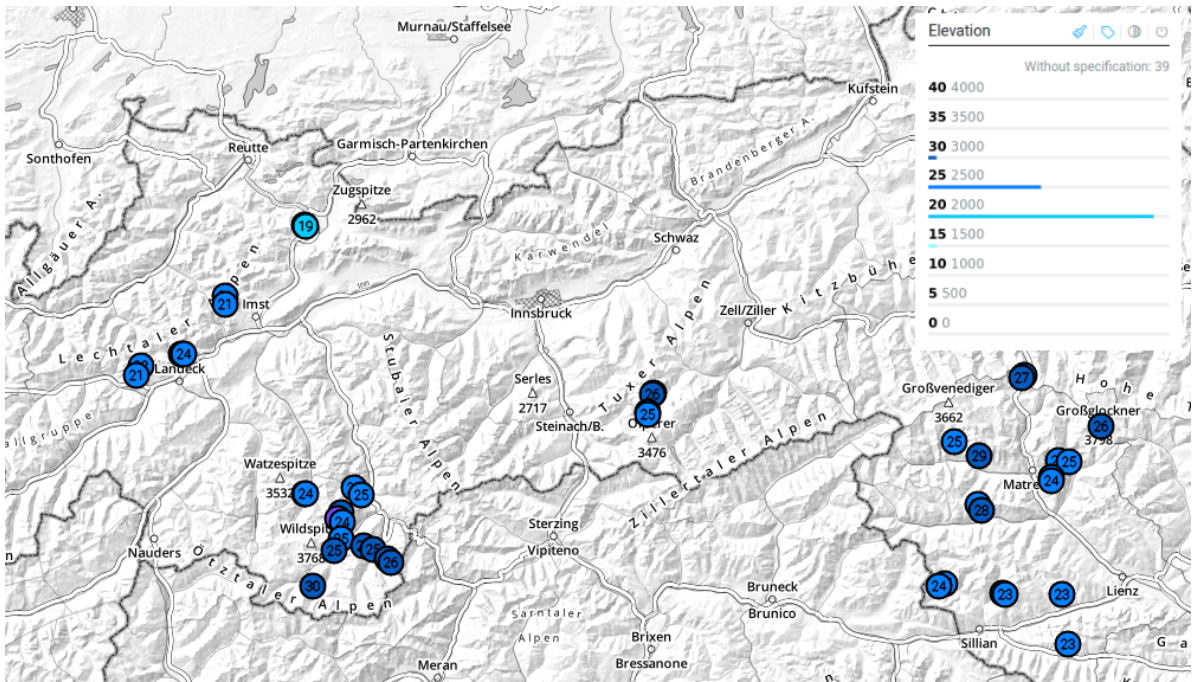


Figure 5: Calculated maximum daily DSLs of a precipitation event on May 2. Each dot represents an AWS, the darker the blue the higher the DSL. Numbers show the respective elevation of the DSL (e.g. 26 for 2600 m). AWSs are only shown, if there was a period with precipitation $\geq 0.1 \text{ mm h}^{-1}$ during that specific day.

the storm has already passed. For example, the observation in Region (B) on February 24 possibly reflects the snowfall limit from the night before when there was still precipitation. Both factors - reporting snowfall limits from a distance and in hind-cast - reduce the accuracy of observations and might lead to discrepancies between observations and calculations. Besides these challenges, it remains unclear, if observers are aware of the definition for the snowfall limit (i.e. 50% solid and 50% liquid). We generally expect observers to report an elevation that can be anything from the lower limit of the mixed snow-rain layer up to the DSL. Consequently, this also means that observed snowfall limits should always be lower than the DSL - or the DSL itself. Consistent with our expectations, reported snowfall limits in Figure 3 tend to be lower in elevation compared to DSL calculations. Nonetheless, there are cases where observations are clearly higher in elevation than calculated DSLs (e.g., Region (A) on February 14 and 20).

In a next step, we wanted to obtain a more quantitative feeling on how DSL values and observed snowfall limits correlate. Figure 4 shows the relationship between calculated DSLs and observed snowfall limits (black dots). The black line represents the linear regression of calculated DSLs and observed snowfall limits. Generally, observations and calculations align well ($R^2 = 0.7$), but there are notable outliers (for example, 1500 m observed vs. 120 m calculated).

Below 1800 m, as indicated by the black regression line, the calculated DSLs are usually higher in elevation than the respective observed snowfall limit. This is in line with what we would expect: we aim to calculate the highest (i.e., driest) snowfall limit possible, while observers typically report a lower (i.e., wetter) limit. However, above 1800 m, the opposite trend appears: the calculated DSLs are generally lower in elevation than the observed snowfall limits.

4.2 Potential operational use

To provide forecasters with an easy way to access calculated DSLs, we integrated this work into the operational system of the Avalanche Service Tyrol. Forecasters in Tyrol use a tool that displays all reported observations for a selected period on a map. This tool allows them to sort observations by type, and color and/or label them according to their values. We added the calculated DSLs to this system by showing the daily maximum DSL of each AWS on the map. Forecasters can color and label the DSL dots by their respective elevations, offering a quick overview of the DSL spatial distribution during a precipitation event. Figure 5 illustrates an exemplary map for a precipitation event on May 2. AWS points are shown only if there was a period with precipitation of 0.1 mm h^{-1} or more on that day. Therefore, in Figure 5, AWS points are not shown in the north-east of Tyrol where there was no precipitation that day. Clicking on an AWS dot opens a time-series

graph similar to Figure 3, showing the DSLs for the past week and the corresponding micro-region.

5. CONCLUSION

In this study, we applied an existing method for calculating the dry-snowfall level (DSL) to assist avalanche forecasting. By focusing on the elevation where precipitation falls exclusively as dry snow, we aim to improve tracking the formation of melt-freeze crusts and associated faceted weak layers, which are critical in avalanche danger assessment.

Our approach involves using data from automated weather stations (AWS) to calculate the wet-bulb temperature, from which we in turn determine the DSLs. We implemented this method into the operational system of the Avalanche Service Tyrol, enabling forecasters to access daily maximum DSL values on a map. The visual representation, coupled with the ability to interact with detailed time-series graphs, provides forecasters with a comprehensive overview of DSLs' spatial distribution during precipitation events.

Initial validations of our method show promising results, with calculated DSLs generally aligning well with observed snowfall limits. However, some discrepancies highlight the need for further refinement and validation. Differences between observations and calculations can stem from three potential sources of error or uncertainty:

- (1) Observed snowfall limits might not represent the DSL but some lower (i.e. wetter) limit. Furthermore, these limits are often reported from a distance or in hind-cast, i.e. after a storm has passed and may therefore be imprecise.
- (2) Our calculations of DSLs are dependent on AWS data of air temperature and relative humidity, meaning potential measurement errors or deviations can distort DSLs' calculations.
- (3) The formula we used is based on a relationship between T_W and the snowfall limit, that Steinacker (1983) found from analyzing 5-year climate observations in Innsbruck. As this is a very local dataset, it remains uncertain if the same conditions apply to all of Tyrol. Furthermore, we used a threshold of $T_W = +0.3$ °C proposed by Fehlmann et al. (2018) as the elevation of the DSL. It does also remain uncertain whether this threshold represents the exact elevation above which crusts no longer form.

To try to minimize these potential error sources, future work will focus on

- (1) improving observed snowfall limits: We aim to better train our observers, to report the DSL and not any limit that is lower in elevation.

- (2) Improving the accuracy of DSL calculations, by incorporating more advanced validation methods, such as using laser disdrometer data.

Overall, the automated calculation of DSL represents a further advancement in avalanche forecasting, offering a more objective and spatially continuous dataset for assessing the formation of melt-freeze crusts and thus snowpack stability.

ACKNOWLEDGEMENT

Special thanks to Marco Knoflach for providing the observer dataset, to Michael Winkler for his meteorological expertise, to Norbert Lanzanasto for implementing our work into the operational system of the Avalanche Warning Service Tyrol and to Jakob Poppeller for designing the map for Figure 2.

References

- Colbeck, S. C. and Jamieson, J.: The formation of faceted layers above crusts, *Cold Reg. Sci. Technol.*, 33, 247–252, doi:10.1016/S0165-232X(01)00045-3, 2001.
- Fehlmann, M., Gascón, E., Rohrer, M., Schwarb, M., and Stofel, M.: Estimating the snowfall limit in alpine and pre-alpine valleys: A local evaluation of operational approaches, *Atmos. Res.*, 204, 136–148, doi:10.1016/j.atmosres.2018.01.016, 2018.
- Froidurot, S., Zin, I., Hingray, B., and Gautheron, A.: Sensitivity of Precipitation Phase over the Swiss Alps to Different Meteorological Variables, *J. Hydrometeorol.*, 15, 685–696, doi:10.1175/JHM-D-13-073.1, 2014.
- Georges, C. and Kaser, G.: Ventilated and unventilated air temperature measurements for glacier-climate studies on a tropical high mountain site, *J. Geophys. Res. Atmos.*, 107, ACL 15–1–ACL 15–10, doi:10.1029/2002JD002503, 2002.
- Jamieson, B.: Formation of refrozen snowpack layers and their role in slab avalanche release, *Rev. Geophys.*, 44, doi:10.1029/2005RG000176, 2006.
- Matsuo, T. and Sasao, Y.: Melting of Snowflakes below Freezing Level in the Atmosphere, *J. Meteorol. Soc. Japan. Ser. II*, 59, 10–25, doi:10.2151/jmsj1965.59.1_10, 1981.
- Mayer, S., Hendrick, M., Michel, A., Richter, B., Schweizer, J., and van Herwijnen, A.: Projected impact of climate warning on avalanche activity at two sites in the Swiss Alps, *Int. snow Sci. Work. Proc.* 2023, pp. 572–576, 2023.
- Rohregger, J. B.: Methoden zur Bestimmung der Schneefallgrenze, Ph.D. thesis, University of Vienna, doi:10.25365/THESIS.3205, 2008.
- Schweizer, J. and Jamieson, J.: Snow cover properties for skier triggering of avalanches, *Cold Reg. Sci. Technol.*, 33, 207–221, doi:10.1016/S0165-232X(01)00039-8, 2001.
- Sims, E. M. and Liu, G.: A Parameterization of the Probability of Snow–Rain Transition, *J. Hydrometeorol.*, 16, 1466–1477, doi:10.1175/JHM-D-14-0211.1, 2015.
- Steinacker, R.: Diagnose und Prognose der Schneefallgrenze, *Wetter und Leb.*, 35, 81–90, 1983.
- Wang, Y., Broxton, P., Fang, Y., Behrangi, A., Barlage, M., Zeng, X., and Niu, G.: A Wet-Bulb Temperature-Based Rain-Snow Partitioning Scheme Improves Snowpack Prediction Over the Drier Western United States, *Geophys. Res. Lett.*, 46, 13 825–13 835, doi:10.1029/2019GL085722, 2019.