

RAMMS::Extended Application Guide Wet-Snow Avalanches

Scenario-Based Parameter Recommendations

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Wet Snow Avalanche Dynamics Calculations

Short Recommendation for Practice

1. Introduction

Wet snow avalanches present one of the most difficult challenges for avalanche modelling. Unlike dry, fluidized avalanches — which are relatively well understood and widely measured — **wet avalanches lack robust field data, particularly on snowpack water content in release and entrainment zones**. In practice, only qualitative estimates of "wetness" are possible, and the processes that govern transitions from dry to wet flow regimes remain highly uncertain.

The presence of liquid water on snow grains fundamentally alters **avalanche mechanics**. Frictional shear heating, rainfall, or entrainment of isothermal snowpacks can all introduce meltwater into the flowing mass. At first, thin films of water raise capillary pressures, causing particles to stick together in agglomerates. This increases internal resistance and suppresses the dispersion of snow grains. At the same time, basal sliding friction is reduced by lubrication, **allowing avalanches to run long distances at relatively low velocities**.

At the center of RAMMS::Extended is a single governing statement: an internal heat-energy balance for the avalanche core,

$$(\hat{h}_\Phi E_\Phi)_t + \text{div}(\hat{h}_\Phi E_\Phi \vec{u}_\Phi) = \dot{Q}_{in} - \varrho_w L \dot{S}_m - \dot{Q}_{out}$$

developed fully in Section 4. Every parameter in this manual exists to populate one term of this equation, or to describe a process — **friction, fluidization, entrainment** — that draws energy from it or feeds mass into it. Liquid water content, air temperature, granule-scale melting, friction, and random kinetic energy are not independent knobs; they are different faces of the same energy and mass balance, which is why they must be set consistently rather than tuned one at a time.

This physical core has one further, distinctive feature. In RAMMS::Extended, snow granules are not assumed to melt uniformly: because frictional and collisional heating is generated at grain contacts, granule surfaces warm well before their interiors. The model represents this unresolved internal gradient with an effective onset temperature, $T_m = -1^\circ\text{C}$, rather than the physical melting point — a small but consequential departure from a purely thermodynamic model, developed in Sections 5 and 6.

A further complication is that "wet snow avalanches" are not a single regime but encompass a spectrum of flow types, depending on how water enters the avalanche:

Frictional-melt avalanches: Initiate dry but at warm mean snow temperatures (e.g., -3°C). Meltwater arises primarily from shear heating during flow.

Mixed entrainment avalanches: Start dry but entrain warm, isothermal snowpacks at lower elevations. Water content increases rapidly as the flow evolves.

Fully wet avalanches: Release directly from isothermal, water-saturated snowpacks. These flows are highly lubricated at initiation, perhaps bypassing the transitional capillary regime.

Each type stresses a different part of the governing equation — frictional-melt avalanches are dominated by the shear-work term, mixed-entrainment avalanches by the entrainment terms, and fully wet avalanches by the initial liquid-water state itself. The same governing framework applies to all three regimes, with differences in initial state and parameter values shifting the balance among the underlying physical processes. For this reason, the case studies in Section 11 deliberately span all three flow types..

This manual is organized to build that physical picture from the ground up. **Sections 2–3** establish the avalanche's initial thermal and hydrological state: snowcover temperature, water content, and the effective liquid-water input supplied by rain, sleet, or snowmelt. **Section 4** presents the governing heat-energy equation itself. **Sections 5–6** develop the model's melting physics in detail — why $T_m = -1^\circ\text{C}$ is used, and how the resulting complementarity condition determines local melt production. **Section 7** completes the thermal boundary conditions with air temperature. **Sections 8–10** turn to the mechanical and dynamic processes that draw on this thermal state as the avalanche moves: friction, the production and decay of random kinetic energy, and entrainment.

Section 11 grounds all of the preceding parameter guidance in six documented events — the VdIS (1999), Gatschiefer (2008), Grünbödli (2008 and 2023), Salezer Tobel (2008), and Rigopiano (2017) avalanches — spanning the frictional-melt, mixed-entrainment, and fully-wet types described above. The recommended parameter ranges given throughout this manual are calibrated against these case studies, not derived from theory alone.

Finally, **Sections 12 and 13** distill everything into two ready-to-use step-by-step recipes: how to set up a wet-snow avalanche simulation, and how to set up a transitional dry-to-wet simulation. Readers who need a working parameter set quickly can go directly to those sections and use Sections 2–11 as supporting reference for *why* the recommended values are what they are.

2. Snowcover Temperature and Wetness

The **Snowcover Tab** defines the thermal state, liquid water content, and density of the snowpack — the first of the initial-state inputs described in this manual (see Section 3 for the estimation of liquid-water input from rain, sleet and snowmelt, and Section 4 for air temperature). These inputs determine how the avalanche core evolves as it entrains snow along its path. Because wet avalanche behavior is sensitive to all three, careful specification of snowcover conditions is essential, and errors here propagate directly into the friction evolution (Section 5) and the internal heat-energy balance (Sections 8–10).

Defining Water Content

The input for **water content** m_{Σ} is given directly as a **depth of liquid water** (mm) at the reference altitude. This is the same quantity as the effective liquid-water input, W_{eff} — developed in Section 3 — use the estimation methods and engineering ranges given there (Tables 1–3) to arrive at a defensible value here.

Example: A snowcover with $m_{\Sigma} = 12\text{mm}$ at a reference altitude of 2000 m represents roughly the water content expected from a short sleet event or several hours of intense melt (see Section 3, Table 3).

Because water content is specified as an absolute depth rather than a fraction of snowcover height, the value entered applies directly at the reference altitude and does not need to be reinterpreted against local snow depth.

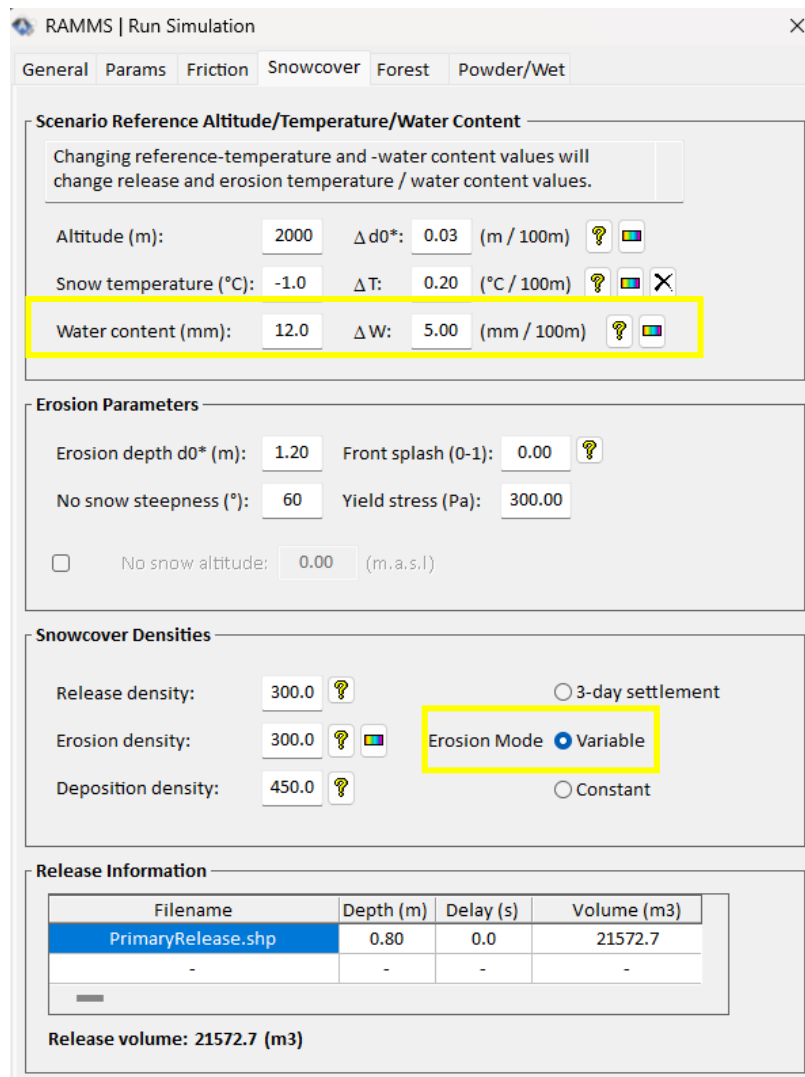
The **water content gradient**, Δm_{Σ} , is specified in **mm per 100 m elevation difference** — the same gradient, G_W , introduced in Section 3. This gradient controls how the liquid water depth changes as the avalanche descends through the calculation domain; use the recommended ranges in Section 3, Table 4, directly.

A positive gradient increases water content toward lower elevations, mimicking the effect of rain-on-snow events or isothermal snowpacks that are wetter in valleys than in high alpine starting zones.

A negative gradient is rarely used but could in principle represent situations where lower elevations are shaded or colder, resulting in reduced liquid content downslope.

Care should be taken when selecting the gradient value. As shown in Section 3 (Table 4), gradients beyond roughly **10 mm/100 m** should be reserved for exceptional events and require event-specific evidence. For example, starting with $m_{\Sigma} = 12\text{mm}$ at 2000 m and applying a gradient of $\Delta m_{\Sigma} = 20\text{mm}/100\text{ m}$, the snowcover at 1000 m elevation would contain 212 mm of liquid water — a value well beyond most field observations, likely to produce unrealistic runout behavior dominated by lubrication effects.

As a practical guideline, gradients in the range **0.5–2.5 mm/100 m** (Section 3, Table 4) are usually sufficient to represent wetter snowpacks at lower elevations without exaggerating the liquid content. Larger gradients should be reserved for extreme cases, such as heavy rain-on-snow events, and should be used cautiously, as they can overwhelm the rheological feedback mechanisms in the model.



RAMMS | Run Simulation

General Params Friction Snowcover Forest Powder/Wet

Scenario Reference Altitude/Temperature/Water Content

Changing reference-temperature and -water content values will change release and erosion temperature / water content values.

Altitude (m): 2000 $\Delta d0^*$: 0.03 (m / 100m) ?

Snow temperature (°C): -1.0 ΔT : 0.20 (°C / 100m) ?

Water content (mm): 12.0 ΔW : 5.00 (mm / 100m) ?

Erosion Parameters

Erosion depth $d0^*$ (m): 1.20 Front splash (0-1): 0.00 ?

No snow steepness (°): 60 Yield stress (Pa): 300.00

☐ No snow altitude: 0.00 (m.a.s.l.)

Snowcover Densities

Release density: 300.0 ? ☐ 3-day settlement

Erosion density: 300.0 ? **Erosion Mode** ☒ Variable

Deposition density: 450.0 ? ☐ Constant

Release Information

Filename	Depth (m)	Delay (s)	Volume (m3)
PrimaryRelease.shp	0.80	0.0	21572.7
-	-	-	-

Release volume: 21572.7 (m3)

Figure 1: Example of the input specification of a wet snowcover. In this case, 12 mm of water is defined at an elevation of 2000 m. Because water content is specified as an absolute depth, this value applies directly at the reference altitude. Typically, with wet snow avalanches, higher densities are applied — in this case $\rho_{\Sigma}=300 \text{ kg/m}^3$. We define the older snowcover using the Variable method.

Independence of Temperature and Wetness

Snow temperature (T_{Σ}) and water content (m_{Σ}) are defined **independently** in **RAMMS::Extended**. It is therefore possible to specify combinations that at first appear contradictory, for example:

A sub-zero snowpack (e.g. $-2 \text{ }^{\circ}\text{C}$) with a nonzero water content. This represents ice granules colder than melting, but with thin water films already present on their surfaces.

It is not possible, however, to assign water temperatures above $0 \text{ }^{\circ}\text{C}$. By definition, the liquid phase is always fixed at the melting point.

When such a snowpack is entrained into the avalanche:

The **ice component** enters the flow at the specified temperature (which may be below 0 °C).

The **water component** is always assumed to be at 0 °C.

This distinction is important: entrainment can cool or warm the avalanche core, depending on whether the incoming snow is colder or warmer than the flow, but the entrained liquid fraction never exceeds 0 °C.

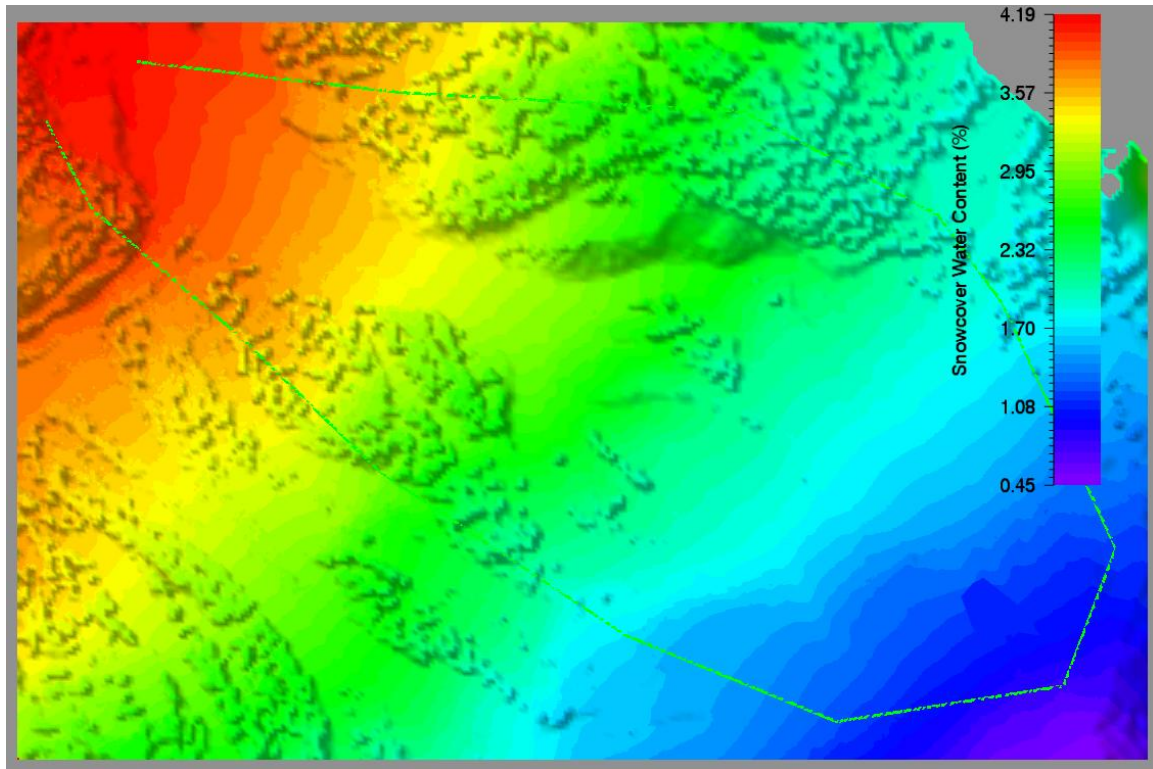


Figure 2: Snowcover water content, expressed as a depth of liquid water. In this example, we defined 12 mm of water at 2000 m (release zone) and a positive gradient of 1 mm/100 m. This leads to roughly 32 mm of water content in the runout zone. This depiction was made by clicking the colorbar next to the water content in the Snowcover input tab.

Density Specification

Another key input is snow density. The **three-day settlement option** is restricted to **new, dry snow** and should not be used for wet avalanche conditions.

If a density above 200 kg/m³ is entered, RAMMS::Extended automatically resets the value to the “new snow” limit, preventing unrealistic compaction for fresh snow.

For wet snow scenarios—where densities are often much higher—users must instead select the Variable or Constant options. These allow specification of denser, water-laden snowpacks typical of isothermal conditions.

Visualization in the GUI

The Snowcover Tab provides immediate feedback through color bar visualizations next to the water content field. These color bars show the spatial distribution of liquid water across the terrain, after applying the specified reference value and vertical

gradient. They offer a quick check to ensure that inputs are realistic and consistent with the terrain and snowcover conditions.

3. Effective Liquid-Water Input from Rain, Sleet and Snowmelt

This section develops the **effective liquid-water input**, W_{eff} — the quantity entered directly as the water content m_{Σ} in the Snowcover Tab (Section 2), together with its elevation gradient, G_W , entered as Δm_{Σ} . Where Section 2 describes how these values are used by the model, this section develops how to *estimate* them from rain, sleet, and snowmelt.

Effective liquid-water input, W_{eff}

Definition, sources, and modelling interpretation

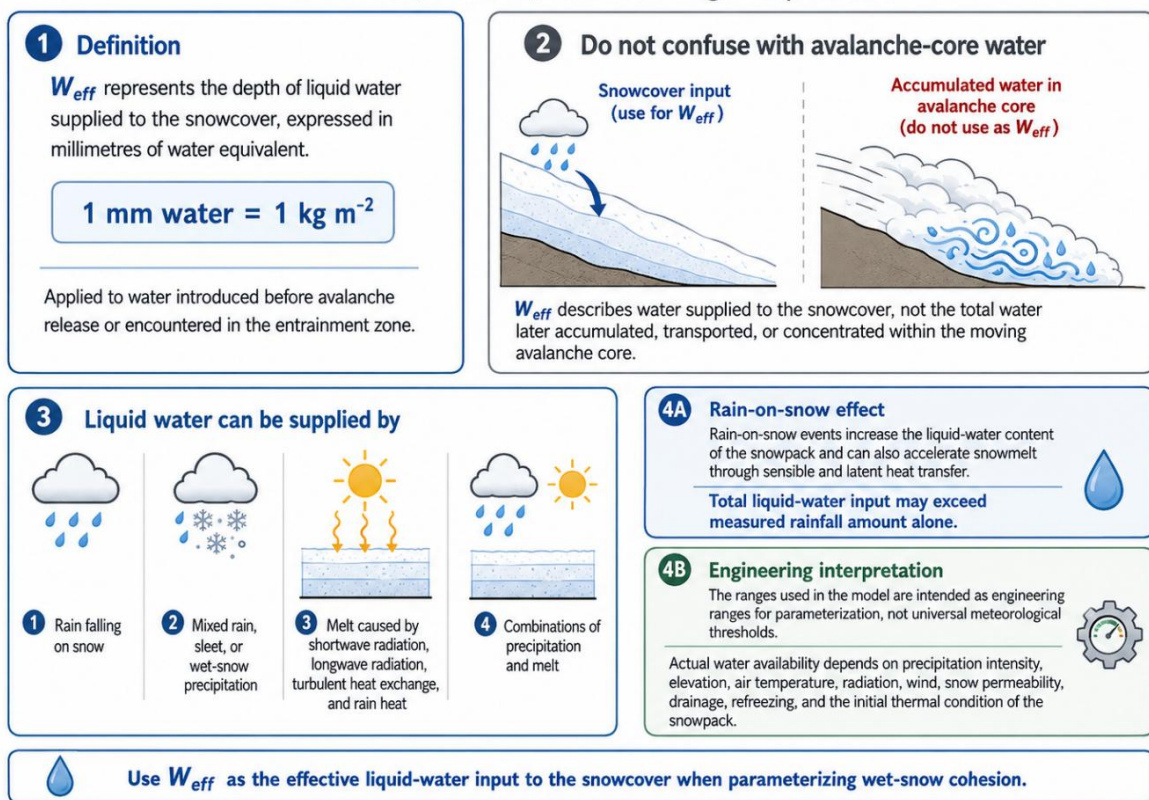


Figure 3: Definition of effective liquid-water input.

The model parameter **effective liquid-water input**, W_{eff} , represents the depth of liquid water supplied to the snowcover, expressed in millimetres of water equivalent:

$$1 \text{ mm water} = 1 \text{ kg m}^{-2}.$$

This quantity should describe water introduced into the snowcover before avalanche release or water encountered in the entrainment zone. It should not be confused with the total water subsequently accumulated, transported or concentrated within the moving avalanche core.

Liquid water can be supplied to a snowcover by:

rain falling on snow;

mixed rain, sleet or wet-snow precipitation;

melt caused by shortwave radiation, longwave radiation, turbulent heat exchange and rain heat;

combinations of precipitation and melt.

Rain-on-snow events increase the liquid-water content of the snowpack and can also accelerate snowmelt through sensible and latent heat transfer. Consequently, the total liquid-water input during such an event may exceed the measured rainfall amount alone.

The values presented below are intended as **engineering ranges for model parameterization**. They are not universal meteorological thresholds. Actual water availability depends on precipitation intensity, elevation, air temperature, radiation, wind, snow permeability, drainage, refreezing and the initial thermal condition of the snowpack.

Rain and mixed precipitation

For precipitation-driven events, the water supplied to the snow surface can be estimated from

$$W_{\text{precip}} = \int_0^T I(t) dt,$$

where:

$I(t)$ is the precipitation rate in mm h^{-1} ;

T is the duration of the event in hours;

W_{precip} is the total precipitation water equivalent in millimetres.

If a constant average intensity is assumed,

$$W_{\text{precip}} = I T.$$

Not all precipitation necessarily remains within the mechanically relevant part of the snowcover. A retention factor may therefore be introduced:

$$W_{\text{eff}} = f_{\text{ret}} W_{\text{precip}},$$

where f_{ret} is the fraction retained in the snow layer before release. Drainage, lateral flow and refreezing reduce f_{ret} , whereas an initially wet, nearly isothermal snowpack may retain or transmit water rapidly.

For mixed precipitation, only the liquid-water-equivalent contribution should be used:

$$W_{\text{eff}} = f_{\text{ret}}(W_{\text{rain}} + W_{\text{sleet}} + W_{\text{wet snow}}).$$

The **World Meteorological Organization** distinguishes precipitation according to both type and intensity and explicitly includes mixed precipitation composed of rain and snow particles.

Meltwater production

Snowmelt is governed by the surface energy balance. The principal energy contributions include net shortwave radiation, net longwave radiation, sensible heat, latent heat, ground heat and heat supplied by rain. Energy-balance snowmelt formulations are described in the **USDA–NRCS snowmelt guidance**.

When the snow is already isothermal at the melting point, the available melt energy can be converted directly into meltwater:

$$M = \frac{Q_m t}{\rho_w L_f},$$

where:

M is the meltwater depth in metres;

Q_m is the net energy available for melting in W m^{-2} ;

t is the duration in seconds;

$\rho_w \approx 1000 \text{ kg m}^{-3}$ is water density;

$L_f \approx 334\,000 \text{ J kg}^{-1}$ is the latent heat of fusion of ice.

Expressed directly in millimetres per hour,

$$\dot{M} = \frac{Q_m(3600)}{\rho_w L_f} \times 1000,$$

which gives approximately

$$\boxed{\dot{M} \approx 0.0108 Q_m}$$

with $\dot{M}$ in mm h^{-1} and Q_m in W m^{-2} . For example:

$$Q_m = 250 \text{ W m}^{-2}$$

produces

$$\dot{M} \approx 0.0108(250) \approx 2.7 \text{ mm h}^{-1}.$$

The assumption that the snow begins at 0°C is important. If the snow is initially colder, part of the supplied energy must first remove the snowpack cold content. Immediate meltwater production will then be smaller. A snowpack becomes wet when the positive radiative and turbulent energy fluxes exceed its remaining cold content.

Estimated liquid-water inputs

Table 1 provides representative ranges suitable for initial model parameterization.

The distinction between **total water supplied** and **effective liquid-water input** accounts for drainage, refreezing and spatial redistribution. The effective value should therefore normally be smaller than the measured precipitation total, unless additional snowmelt occurs during the event.

Table 1: Suggested effective liquid-water input W_{eff} .

Process or event	Representative assumptions	Total water supplied	Suggested effective liquid-water input
Light winter rain on snow	Approximately 1–2 mm/h for 4–8 h	5–15 mm	5–10 mm
Moderate winter rainstorm	Approximately 2–5 mm/h for 6–12 h	15–60 mm	10–40 mm
Strong winter rainstorm	Approximately 5–10 mm/h for 8–12 h	40–120 mm	30–80 mm
Short sleet-on-snow event	Mixed rain, ice particles and wet snow over 2–4 h	3–12 mm w.e.	3–8 mm
Prolonged sleet-on-snow event	Mixed precipitation over 6–12 h	10–35 mm w.e.	5–25 mm
Intense melt period, 2 h	Snow initially at 0°C; $Q_m = 150\text{--}350 \text{ W m}^{-2}$	3.2–7.5 mm	3–8 mm
Intense melt period, 4 h	Same energy range and initial condition	6.5–15.1 mm	6–15 mm
Intense melt period, 6 h	Same energy range and initial condition	9.7–22.6 mm	10–20 mm
Combined rain and melt event	Moderate rain combined with positive energy balance	20–80 mm or more	20–60 mm
Exceptional prolonged rain-on-snow event	Strong or multiday rainfall with limited drainage	80–200 mm or more	60–150 mm

Meltwater estimates for intense radiation

For an isothermal snowcover, the calculated meltwater production is given in Table 2.

Table 2: Meltwater production from intense radiation.

Net energy available for melt	Melt rate	Melt after 2 h	Melt after 4 h	Melt after 6 h
150 W/m²	1.62 mm/h	3.2 mm	6.5 mm	9.7 mm
250 W/m²	2.69 mm/h	5.4 mm	10.8 mm	16.2 mm
350 W/m²	3.77 mm/h	7.5 mm	15.1 mm	22.6 mm

These values should be interpreted as potential meltwater production under sustained positive energy input. Actual shortwave radiation at the snow surface is not identical to net melt energy because part of the incoming radiation is reflected by the snow and the remaining energy-balance terms may either add or remove energy. The NRCS snowmelt framework treats radiation, turbulent exchange, rain heat and other fluxes as separate contributions to the total energy available for melt.

Recommended parameter values

For systematic numerical experiments, the following discrete values provide a useful parameter series:

Table 3: Representative effective liquid-water input values.

Effective liquid-water input	Representative interpretation
0 mm	Dry reference condition
5 mm	Short sleet event, weak rain or short melt period
10 mm	Light rain-on-snow or several hours of intense melt
20 mm	Prolonged sleet, strong daytime melt or moderate rain
40 mm	Moderate rain-on-snow storm
60 mm	Strong rain-on-snow event
80 mm	Very strong or prolonged winter rainfall
100 mm	Severe rain-on-snow input
150 mm	Exceptional prolonged precipitation input
200 mm	Extreme multiday input; not representative of short-duration solar melt

Inputs between approximately 5 and 20 mm are therefore plausible for short melt or mixed-precipitation events. Values between approximately 40 and 100 mm are more naturally associated with rain-on-snow storms. Inputs approaching 150–200 mm should be reserved for exceptional, prolonged precipitation scenarios and should not be interpreted as the product of a two- to six-hour solar melting period.

Elevation-dependent liquid-water input

The effective liquid-water input may be spatially uniform or vary with elevation. An elevation gradient can be specified as

$$G_W = -\frac{\partial W_{\text{eff}}}{\partial z},$$

expressed in millimetres of additional liquid-water input per 100 m decrease in elevation. Positive values therefore indicate wetter snowcover at lower elevations.

The local water input is calculated from

$$W_{\text{eff}}(z) = W_{\text{ref}} + G_W \frac{z_{\text{ref}} - z}{100}.$$

A temperature-driven melt gradient is generally small over short periods. For snow initially at the melting point, two to six hours of enhanced melt commonly produces an elevation-related difference of less than approximately 1 mm/100 m, unless radiation, aspect or turbulent heat exchange also vary strongly.

Larger gradients may occur during rain-on-snow or mixed-phase precipitation events. Near the rain–snow transition, the fraction of precipitation falling as rain can change rapidly with elevation, producing gradients of several millimetres and occasionally approximately 10 mm/100 m. Such strong gradients should only be used when supported by meteorological or field evidence.

For initial calibration, a spatially uniform value is recommended. Suggested sensitivity values are:

$$G_W = 0, 1, 2.5, 5 \text{ mm/100 m.}$$

Values near 10 mm/100 m represent a strong rain–snow transition or exceptional precipitation gradient.

Recommended parameter combinations

Liquid-water inputs and elevation gradients for RAMMS::Extended calculations

Suggested combinations for systematic numerical experiments and sensitivity studies of wet-snow conditions.

	Effective liquid-water input	Recommended gradient G_W	Representative interpretation	Likely combination / example	Use in practice
DRY	 0 mm	0 mm/100 m	Dry reference condition	Dry snowcover with no downslope wetting	Baseline reference case
	 5 mm	0–0.5 mm/100 m	Short sleet event, weak rain or short melt period	Nearly uniform wetting; only a slight increase toward lower elevations	Good first wet-snow sensitivity test
	 10 mm	0.5–1 mm/100 m	Light rain-on-snow or several hours of intense melt	Weak temperature-driven gradient; lower elevations slightly wetter	Suitable for mild wet-snow conditions
	 20 mm	1–2.5 mm/100 m	Prolonged sleet, strong daytime melt or moderate rain	Moderate downslope wetting during a transition event	Useful for moderate wet-snow scenarios
	 40 mm	2.5–5 mm/100 m	Moderate rain-on-snow storm	Strong melt or moderate rain-snow transition; noticeably wetter runoff zone	Good calibration range for rain-on-snow events
	 60 mm	2.5–5 mm/100 m	Strong rain-on-snow event	Substantial wetting with clear downslope increase, but not extreme	Appropriate for strong wet-avalanche scenarios
	 80–100 mm	5–10 mm/100 m	Very strong or prolonged winter rainfall / severe rain-on-snow input	Very strong event near the rain-snow transition; markedly wetter lower elevations	Use as an upper sensitivity scenario
EXTREME WET	 150–200 mm	>10 mm/100 m	Exceptional prolonged precipitation input / extreme multiday input	Exceptional gradient; should only be used with event-specific evidence	Extreme scenario, not a standard calibration case

Recommended gradients are expressed as additional effective liquid-water input per 100 m decrease in elevation.
Strong gradients should be used only when supported by meteorological or field evidence.

Figure 4: Recommended liquid-water inputs and liquid-water gradients for RAMMS::Extended calculations. The higher the liquid-water, the higher the gradient.

With W_{eff} and G_W established, **Section 4** turns to ambient air temperature — the second thermal boundary condition, entered in the Powder/Wet Tab, that together with the snowcover inputs from Sections 2–3 determines the avalanche's initial thermal state.

Table 4: Recommended liquid-water gradients G_W .

Gradient G_W	Interpretation
0 mm/100 m	Uniform snowcover input
0.5–1 mm/100 m	Weak melt or temperature-driven gradient
1–2.5 mm/100 m	Moderate downslope wetting
2.5–5 mm/100 m	Strong melt or moderate rain-snow transition
5–10 mm/100 m	Very strong event near the rain-snow transition
>10 mm/100 m	Exceptional gradient; should require event-specific evidence

4. Temperature and Melting

Sections 2 and 3 established the avalanche's initial thermal and hydrological state — snow temperature and water content (T_Σ , m_Σ) and the effective liquid-water input (W_{eff}) supplied by rain, sleet or melt. This section presents the governing equation that tracks how that initial state evolves once the avalanche is moving: the internal heat-energy balance. Each source term in this equation is a physical process with its own dedicated parameterization, developed in the sections that follow — air temperature (Section 7), friction (Section 8), random-energy production and decay (Section 9), and entrainment (Section 10). This section shows how those processes combine; the sections after it show how to set each one individually.

Role in regime transition

Liquid water is the main driver of the **dry-to-wet transition**. Small amounts form thin films at grain contacts, which reduce Coulomb friction, raise viscous resistance, damp dispersion, and encourage plug-like motion. As water content grows, basal resistance drops and wetting accelerates, pushing the flow fully into the wet regime.

In **RAMMS::Extended**, the core's thermal state is tracked by the internal heat energy E_ϕ (temperature is its scale). E_ϕ is advected with the core and gains energy from shear work, entrainment, and the decay of random kinetic energy; it also exchanges heat with the powder cloud. Melting adds one more term: a latent-heat sink. This term links heat input to the local meltwater production rate—and that rate is what ultimately governs the evolving rheology.

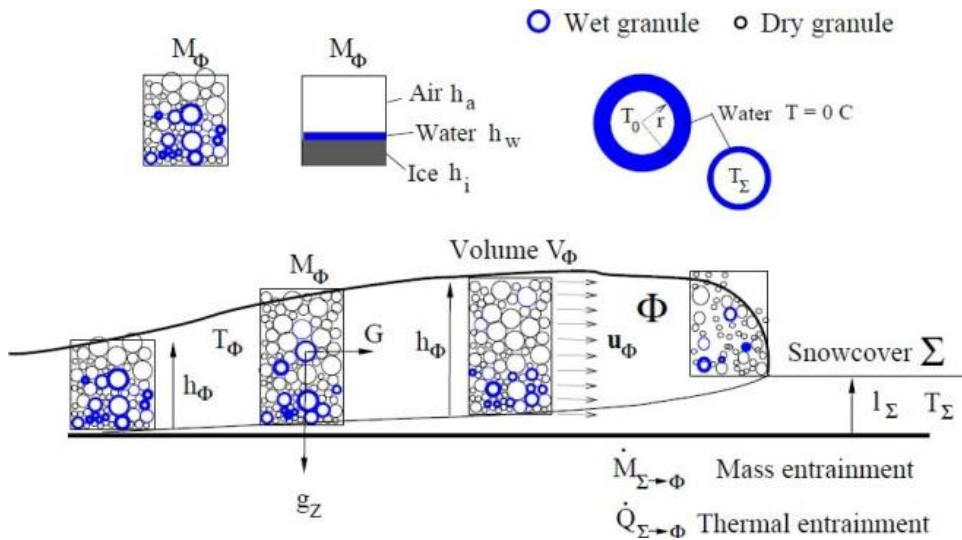


Figure 5: The “temperature” of an avalanche is hard to define. Because frictional/collisional heating raises granule surface temperatures before the mean, RAMMS::Extended starts melting slightly below the physical melting point by using a model constant $T_m = -1^\circ\text{C}$.

The internal heat energy balance for the avalanche core is

$$(\hat{h}_\phi E_\phi)_t + \text{div}(\hat{h}_\phi E_\phi \vec{u}_\phi) = \dot{Q}_{\text{in}} - \dot{Q}_{\text{out}}$$

with the thermal input

$$\dot{Q}_{in} = [1 - \alpha_\phi] \dot{W}_\phi + \rho_\Sigma [1 - \varepsilon_\phi] \dot{L}_{\Sigma \rightarrow \phi} + \beta_\phi \hat{h}_\phi R_\phi + \varrho_\Sigma c_\Sigma T_\Sigma \dot{M}_{\Sigma \rightarrow \phi} + H_\phi A_{s,\pi} (T_\Lambda - T_\phi)$$

The first four terms were already part of the model: **shear work** (reduced by the fraction diverted into random kinetic energy production, α_ϕ ; Section 9), **entrainment heat** (reduced by the fraction diverted into dispersive random energy, ε_ϕ ; Section 10), the **decay of random kinetic energy** back into heat ($\beta_\phi \hat{h}_\phi R_\phi$; Section 9), and the **thermal contribution of entrained mass** at snow temperature T_Σ (Section 2). The fifth term, $H_\phi A_{s,\pi} (T_\Lambda - T_\phi)$, is the **convective exchange with the ambient air**: positive (a gain) when the air is warmer than the core, negative (a loss) when it is colder. Its coefficients — the heat transfer coefficient H_ϕ and the granule surface area $A_{s,\pi}$ — are derived in Section 7 from standard Nusselt-, Reynolds- and Prandtl-number correlations.

The loss of internal heat energy to the cloud is

$$\dot{Q}_{out} = \dot{M}_{\phi \rightarrow \Pi} E_\phi.$$

Melting is introduced by adding a latent-heat sink to the right-hand side and a consistent moisture equation. Let m_ϕ be the **meltwater thickness** (m) in the core. Then the transport of meltwater is

$$(m_\phi)_t + \text{div}(m_\phi \vec{u}_\phi) = \dot{S}_m + \frac{\varrho_w}{\hat{\rho}_\phi} \phi_{\Sigma w} \dot{M}_{\Sigma \rightarrow \phi}$$

where $\dot{S}_m$ is the **local melt production rate**. The internal-energy equation becomes

$$(\hat{h}_\phi E_\phi)_t + \text{div}(\hat{h}_\phi E_\phi \vec{u}_\phi) = \dot{Q}_{in} - \varrho_w L \dot{S}_m - \dot{Q}_{out}$$

The **latent-heat sink** is $-\varrho_w L \dot{S}_m$. Two conditions govern $\dot{S}_m$: the effective onset temperature at which melting begins, and the complementarity condition that determines the melt rate once that threshold is reached. Both are developed next, in Sections 5 and 6.

5. Effective onset temperature for surface melting

Physical motivation

As introduced in Section 4, RAMMS::Extended applies a latent-heat sink governed by the effective onset temperature T_m , rather than the physical melting point $T_{\text{fus}} = 0^\circ\text{C}$. This section develops why.

The temperature variable tracked by the model, T_ϕ , is the *mean* temperature of the avalanche core — it does not resolve the temperature distribution within individual snow granules.

During avalanche motion, **frictional and collisional heat is generated primarily at grain contacts and within strongly sheared regions**. This heat is therefore initially concentrated near the granule surfaces. Because thermal diffusion into the granule interior requires a finite amount of time, the surface temperature can rise much faster than the temperature in the centre of the granule.

A granule may consequently begin to melt at its surface even though its volume-averaged temperature remains below 0°C . To represent this unresolved surface-core temperature difference, RAMMS::Extended introduces the effective melting-onset temperature

$$T_m = -1^{\circ}\text{C}.$$

This value must not be interpreted as a depression of the physical melting point. Instead, it is an empirical threshold applied to the mean avalanche-core temperature. It represents the condition $T_{\text{surf}} \approx 0^{\circ}\text{C}$, while $T_{\Phi} \approx -1^{\circ}\text{C}$. The corresponding temperature difference is

$$T_{\text{surf}} - T_{\Phi} \approx 1^{\circ}\text{C}.$$

This difference provides a compact representation of the unresolved thermal gradient within the snow granules.

Relation to transient heat diffusion

The physical basis of this approximation can be illustrated by considering a spherical snow granule with radius r_{π} . During rapid surface heating, the internal granule temperature can be approximated by a parabolic radial profile:

$$T(r, t) = T_0 + [\bar{T}(t) - T_0] \left[A(Fo) + B(Fo) \left(\frac{r}{r_{\pi}} \right)^2 \right].$$

Here,

$T(r, t)$ is the temperature at radial position r and time t .

the initial uniform granule temperature is T_0 ,

the volume-averaged granule temperature is $\bar{T}(t)$.

The **Fourier number** is defined as $Fo(t) = \alpha t / r_{\pi}^2$,

α is the thermal diffusivity of snow.

The Fourier number describes the extent to which heat has diffused through the granule. Small Fourier numbers correspond to short heating times, low thermal diffusivity, or relatively large granules. Large Fourier numbers indicate that heat has penetrated farther into the granule interior.

The coefficients $A(Fo)$ and $B(Fo)$ define the shape of the radial temperature profile. Conservation of energy over the spherical granule requires

$$A(Fo) + \frac{3}{5}B(Fo) = 1.$$

It follows that

$$A(Fo) = 1 - \frac{3}{5}B(Fo).$$

At the granule surface, $r = r_{\pi}$, and the surface temperature becomes

$$T_{\text{surf}} = \bar{T} + \frac{2}{5}B(Fo)(\bar{T} - T_0).$$

At the centre of the granule, $r = 0$, and the core temperature becomes

$$T_{\text{core}} = \bar{T} - \frac{3}{5}B(Fo)(\bar{T} - T_0).$$

For a positive curvature coefficient, $B(Fo) > 0$, the temperatures satisfy

$$T_{\text{surf}} > \bar{T} > T_{\text{core}}.$$

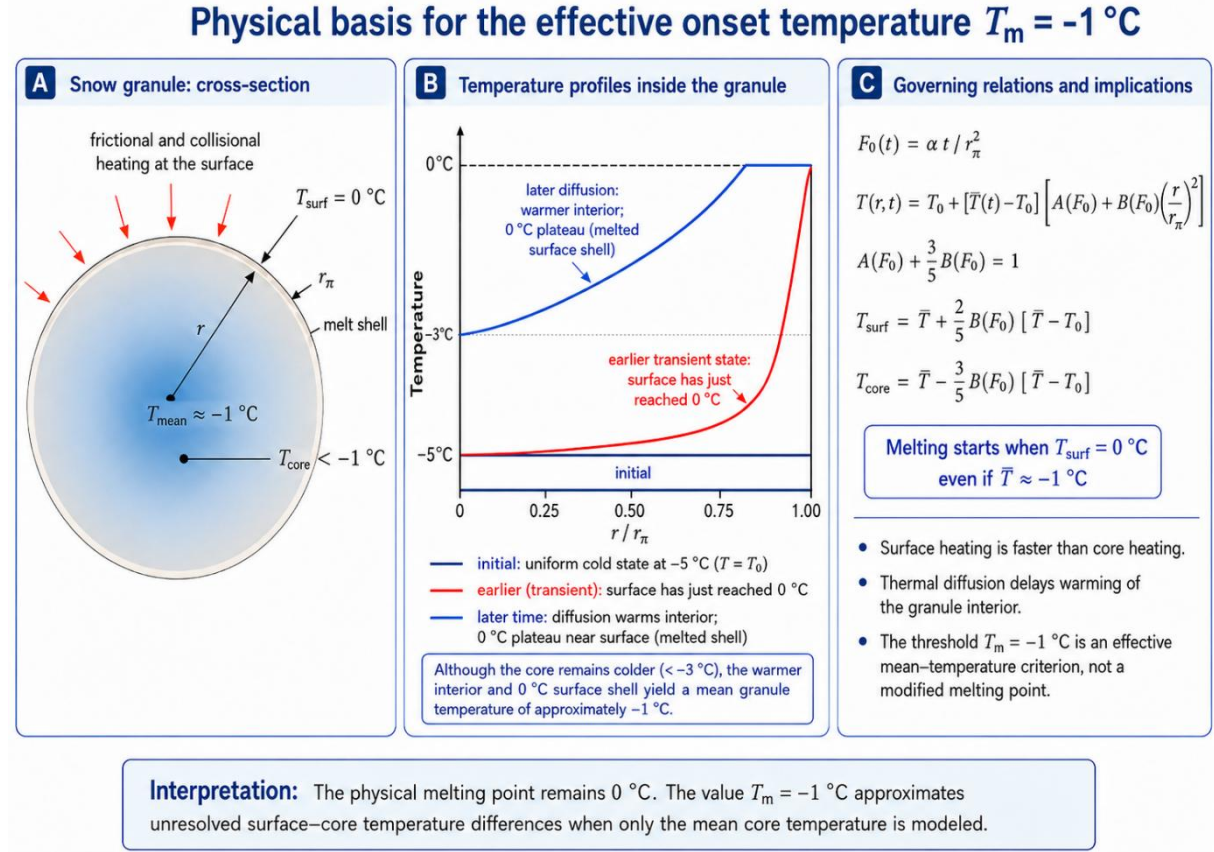


Figure 6: Model of granule surface melting.

The surface is therefore warmer than the volume-averaged temperature, while the centre remains colder than the average. At small Fourier numbers, heat remains concentrated near the granule surface. The coefficient $B(Fo)$ is then comparatively large, and the surface-core temperature gradient is strong.

The granule surface may already have reached the physical melting temperature, $T_{\text{surf}} = 0^\circ\text{C}$, while the volume-averaged temperature remains approximately $\bar{T} \approx -1^\circ\text{C}$, and the centre remains colder: $T_{\text{core}} < -1^\circ\text{C}$. The fixed threshold $T_m = -1^\circ\text{C}$ is therefore a **computationally inexpensive surrogate for this unresolved radial temperature distribution**.

Illustrative temperature evolution

Consider a snow granule with an initially uniform temperature $T_0 = -5^\circ\text{C}$. Initially, the temperature is uniform throughout the granule: $T(r, 0) = T_0 = -5^\circ\text{C}$. During the early transient stage, frictional and collisional heating raises the surface temperature rapidly, while very little heat has penetrated into the granule interior. The core temperature therefore remains approximately $T_{\text{core}} \approx -5^\circ\text{C}$, while the surface has just reached $T_{\text{surf}} = 0^\circ\text{C}$.

At a later time, heat has diffused farther into the granule. The centre may have warmed to approximately $T_{\text{core}} \approx -3^\circ\text{C}$, while an outer surface shell has reached the physical melting temperature: $T(r, t) = 0^\circ\text{C}$. For example, the isothermal surface shell may occupy the radial interval

$$0.8 \leq \frac{r}{r_\pi} \leq 1.$$

The inner boundary of this shell is therefore located at $r = 0.8r_\pi$. The volume of the complete granule is $V_\pi = \frac{4}{3}\pi r_\pi^3$. The volume enclosed by the inner radius is $V_{\text{inner}} = \frac{4}{3}\pi(0.8r_\pi)^3$. The volume of the outer shell is consequently $V_{\text{shell}} = V_\pi - V_{\text{inner}}$. Substituting the spherical volumes gives

$$V_{\text{shell}} = \frac{4}{3}\pi r_\pi^3 - \frac{4}{3}\pi(0.8r_\pi)^3.$$

The volume fraction contained within the outer shell is

$$f_{\text{shell}} = \frac{V_{\text{shell}}}{V_\pi}.$$

Therefore,

$$f_{\text{shell}} = 1 - (0.8)^3 = 0.488.$$

Thus, almost half of the granule volume lies within the outermost twenty percent of its radius. The volume-averaged temperature of a spherical granule is defined by

$$\bar{T}(t) = \frac{1}{V_\pi} \int_{V_\pi} T(r, t) dV.$$

For spherical symmetry, $dV = 4\pi r^2 dr$. The volume-averaged temperature can therefore be written as

$$\bar{T}(t) = \frac{\int_0^{r_\pi} T(r, t) 4\pi r^2 dr}{\frac{4}{3}\pi r_\pi^3}.$$

After cancelling the common geometric factors,

$$\bar{T}(t) = \frac{3}{r_\pi^3} \int_0^{r_\pi} T(r, t) r^2 dr.$$

An internal temperature distribution consisting of a colder core and a warm outer shell can consequently produce a mean temperature near $\bar{T} \approx -1^\circ\text{C}$, even though much of the granule interior remains considerably colder.

This provides the physical interpretation of the effective onset criterion $T_m = -1^\circ\text{C}$. The value does not imply that ice physically melts at -1°C . Rather, it means that when the calculated mean avalanche-core temperature reaches approximately $T_\Phi \approx -1^\circ\text{C}$, the unresolved surfaces of the snow granules may already have reached the physical melting temperature $T_{\text{surf}} = 0^\circ\text{C}$.

With the physical basis for T_m established, **Section 6** turns to how this threshold is actually enforced in the model — the complementarity condition that determines the local melt production rate $\dot{S}_m$.

6. Melting Condition in the Current Model

Section 5 established the physical basis for the effective onset temperature, $T_m = -1^\circ\text{C}$. This section shows how that threshold is enforced in the governing heat-energy equation from Section 4, and how it determines the local melt production rate, $\dot{S}_m$.

The current RAMMS::Extended formulation applies the complementarity conditions

$$T_\Phi \leq T_m, \quad \dot{S}_m \geq 0,$$

and

$$(T_\Phi - T_m)\dot{S}_m = 0.$$

The three conditions may also be written together as

$$T_\Phi \leq T_m, \quad \dot{S}_m \geq 0, \quad (T_\Phi - T_m)\dot{S}_m = 0.$$

These conditions define two possible thermal states. As long as $T_\Phi < T_m$, melting is inactive:

$$\dot{S}_m = 0.$$

When the mean avalanche-core temperature reaches the effective threshold, $T_\Phi = T_m$, the temperature is held at $T_m = -1^\circ\text{C}$. Any additional heat supplied to the avalanche core is then converted into latent heat of melting.

Let T_Φ^* denote the temperature that would be obtained from the energy balance in section 4 before the melting constraint is applied. The constrained avalanche-core temperature is

$$T_\Phi = \min(T_\Phi^*, T_m).$$

If $T_\Phi^* \leq T_m$, then $T_\Phi = T_\Phi^*$, and

$$\dot{S}_m = 0.$$

If $T_\Phi^* > T_m$, then $T_\Phi = T_m$, and the temperature excess $T_\Phi^* - T_m$ is converted into latent heat. The excess heat supplied over a computational time interval Δt is

$$Q_m = \int_t^{t+\Delta t} \dot{Q}_m dt.$$

For a computational element with flow depth $\hat{h}_\Phi$, the excess heat may be related to the unconstrained temperature overshoot by

$$\int_t^{t+\Delta t} \dot{Q}_m dt = \hat{\rho}_\Phi c_\Phi \hat{h}_\Phi (T_\Phi^* - T_m).$$

Here,

$\hat{\rho}_\Phi$ is the avalanche-core density,

c_Φ is the specific heat capacity of the avalanche core, and

$\hat{h}_\Phi$ is the avalanche flow depth.

The mean excess heat-production rate over the time interval is

$$\dot{Q}_m = \frac{Q_m}{\Delta t}.$$

The corresponding meltwater production rate is

$$\dot{S}_m = \frac{\dot{Q}_m}{\rho_w L_f},$$

where

ρ_w is the density of water, and

L_f is the latent heat of fusion.

Substituting the excess heat-production rate gives

$$\dot{S}_m = \frac{\hat{\rho}_\Phi c_\Phi \hat{h}_\Phi (T_\Phi^* - T_m)}{\rho_w L_f \Delta t}.$$

This expression applies when $T_\Phi^* > T_m$. When $T_\Phi^* \leq T_m$, the meltwater production rate remains $\dot{S}_m = 0$. The melting law may therefore be summarized as

$$\dot{S}_m = \begin{cases} 0, & T_\Phi^* \leq T_m, \\ \frac{\hat{\rho}_\Phi c_\Phi \hat{h}_\Phi (T_\Phi^* - T_m)}{\rho_w L_f \Delta t}, & T_\Phi^* > T_m. \end{cases}$$

Interpretation and limitations

The value $T_m = -1^\circ\text{C}$ provides a **practical first-order correction for thermal heterogeneity** at the granule scale. It allows surface melting to begin before the mean avalanche-core temperature reaches 0°C , as expected during rapid frictional and collisional heating (Section 5).

The approximation nevertheless has **several limitations**. The surface-mean temperature difference $T_{\text{surf}} - \bar{T}$ is prescribed indirectly rather than calculated explicitly. The threshold does not explicitly account for the granule radius r_π , the heating duration t , or the thermal diffusivity α — all of which appear in the Fourier-based derivation in Section 5. The same threshold $T_m = -1^\circ\text{C}$ is used even though the internal temperature profile evolves during avalanche motion. The method also cannot calculate the thickness of the melted surface layer. If the inner radius of the melted shell is denoted by r_m , the melt-shell thickness would be $\delta_m = r_\pi - r_m$ — an analogue of the outer-shell fraction $f_{\text{shell}} = 0.488$ worked out in Section 5, but one the current formulation does not compute directly.

A more physical formulation replaces the fixed threshold with a Fourier-based radial temperature profile. In that approach, the physical melting temperature remains $T_{\text{fus}} = 0^\circ\text{C}$, and melting begins when the calculated surface temperature satisfies $T_{\text{surf}} \geq T_{\text{fus}}$. Equivalently,

$$T_{\text{surf}} \geq 0^\circ\text{C}.$$

The mean temperature may nevertheless remain below the melting point: $\bar{T} < 0^\circ\text{C}$.

The core temperature may also remain below the physical melting point: $T_{\text{core}} < 0^\circ\text{C}$.

Once the surface reaches $T_{\text{surf}} = 0^\circ\text{C}$, additional heat is used to melt an outer shell of the granule rather than raising the surface temperature above the physical melting point.

The present formulation using $T_m = -1^\circ\text{C}$ can therefore be regarded as a simplified and computationally efficient approximation of the more complete surface-heating and surface-melting process.

With the governing equation, the onset threshold, and the melting condition now established, **Section 7** returns to the individual boundary conditions that feed into $\dot{Q}_{\text{in}}$ —beginning with air temperature, the atmospheric thermal input introduced there.

7. Air Temperature

The **Powder/Wet Tab** contains the input for ambient air temperature (T_Λ). At first glance this parameter may appear secondary, but it provides an important thermal boundary condition for the avalanche as it travels across large elevation gradients — entering directly as the fifth term, $H_\Phi A_{s,\pi}(T_\Lambda - T_\Phi)$, in the internal heat-energy balance $\dot{Q}_{\text{in}}$ developed in Section 4. The surrounding air can either **retard melting** (in cold conditions) or **accelerate melting** (in warm conditions).

Cold air (below freezing): Suppresses additional meltwater generation at the surface of snow granules, prolonging the persistence of dry or transitional regimes.

Warm air (near or above freezing): Promotes enhanced melting, particularly in the runout zones, encouraging the transition to wet, lubricated flow.

Unless direct measurements are available, it is common practice to set air temperature equal to the snowcover temperature ($T_\Lambda \approx T_\Sigma$, Section 2), assuming a near-equilibrium between air and snowpack.

Convective Heat Exchange

RAMMS::Extended includes a convective heat-transfer formulation originally developed in chemical engineering during the heat exchange with spherical particles in turbulent flow (Ranz and Marshall, 1952). The air temperature T_Λ enters the model through the heat flux term:

$$q_{\Lambda \rightarrow \Phi} = H_\Phi A_{s,\pi}(T_\Lambda - T_\Phi)$$

where:

T_Φ = avalanche core (or phase) temperature,

T_{Λ} = ambient air temperature,

$A_{s,\pi}$ = total surface area of snow granules exposed to air,

H_{Φ} = convective heat transfer coefficient.

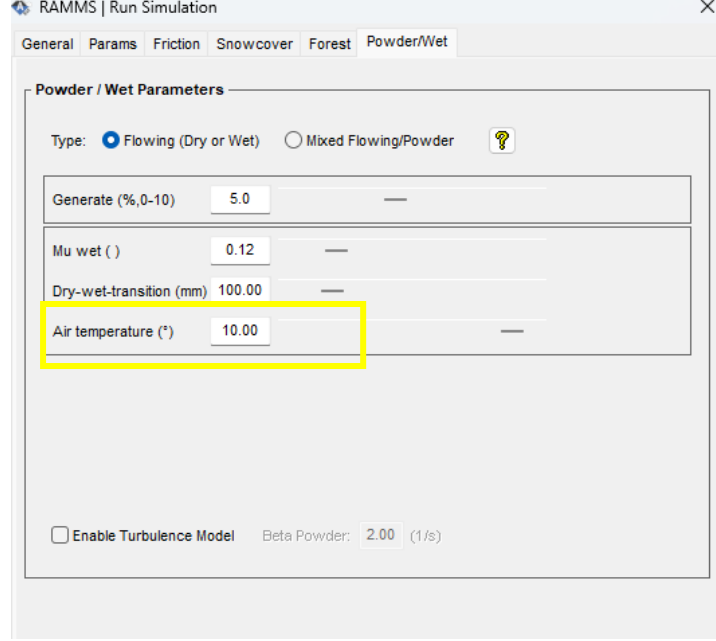


Figure 7: Input of the air temperature is found in the Powder/Wet tab. Temperatures can vary between -30°C and $+30^{\circ}\text{C}$.

The heat transfer coefficient is obtained from standard dimensionless correlations:

$$H_{\Phi} = \frac{Nu \cdot k_a}{2r_{\pi}}$$

with

k_a = thermal conductivity of air ($\approx 0.0257 \text{ W}\cdot\text{m}^{-1}\cdot\text{K}^{-1}$),

r_{π} = mean particle radius,

Nu = Nusselt number, defined as:

$$Nu = 2 + 0.6Re^{\frac{1}{2}}Pr^{\frac{1}{3}}$$

The constant term, $Nu = 2$, is the pure-conduction limit for a sphere in a stagnant fluid; the correction term, $0.6 Re^{1/2} Pr^{1/3}$, is the empirical forced-convection enhancement of Ranz & Marshall (1952), valid for the moderate Reynolds numbers expected for snow granules moving through air — not for high-Re turbulent regimes.

Here:

The Reynolds number depends on the speed of the avalanche:

$$Re = \frac{2\|\mathbf{u}_{\Phi}\|r_{\pi}}{\nu_a}$$

where:

$\|\mathbf{u}_\Phi\|$: flow velocity of the avalanche core (slope parallel),

ν_a : kinematic viscosity of air.

The Prandtl number of air is:

$$Pr = \frac{c_a \nu_a}{k_a}$$

with c_a the specific heat capacity of air.

The total surface area exposed to air is computed assuming spherical snow granules:

$$n_p = \frac{3 \hat{\rho}_\Phi \hat{h}_\Phi A}{4 \rho_\pi \pi r_\pi^3}$$

with

$\hat{\rho}_\Phi$ = co-volume density of the avalanche core,

$\hat{h}_\Phi$ = co-volume height of phase p,

A = computational cell area,

ρ_π = material density of snow granules (300–500 kg/m³ typical).

The corresponding total granule surface area follows directly, since each granule contributes a surface area $4\pi r_\pi^2$:

$$A_{s,\pi} = n_p \cdot 4\pi r_\pi^2 = \frac{3 \hat{\rho}_\Phi \hat{h}_\Phi A}{\rho_\pi r_\pi}$$

Practical Implications

Because snow granules are small, their specific surface area is high, and even modest temperature differences with the atmosphere can have a measurable effect. Still, in most simulations the effect of air temperature is **secondary** compared to frictional heating and entrainment of warm snow.

In **very cold air masses** (e.g., $T_A < -10$ °C), convective cooling can delay or even prevent wet transitions, leading to shorter runout distances than expected.

In **warm air masses** ($T_A > 0$ °C), convective fluxes can enhance meltwater production at the particle surface, particularly in valley bottoms, promoting lubricated flow.

Thus, **while air temperature is not usually the dominant factor in wet avalanche dynamics**, it provides an important correction that can shift the timing and extent of dry–wet transitions, especially for avalanches spanning large elevation ranges.

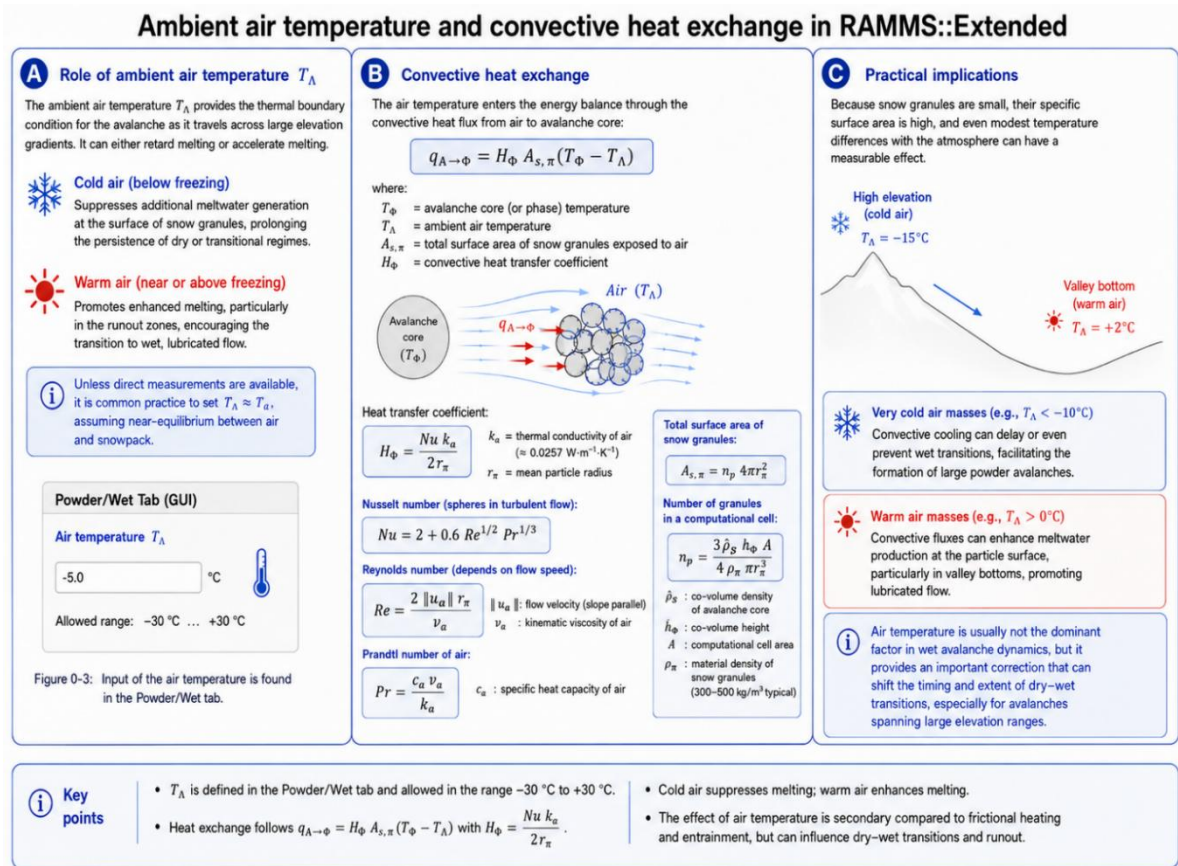


Figure 8: Ambient air temperature and convective heat exchange in RAMMS::Extended. The air temperature T_A , specified in the Powder/Wet Tab, provides the atmospheric thermal boundary condition for the avalanche core. Cold air suppresses surface melting and can preserve dry, fluidized flow, whereas warm air promotes meltwater production and the transition toward wet, lubricated motion. Convective heat exchange is calculated from the temperature difference between the avalanche core and the surrounding air using standard Nusselt-, Reynolds-, and Prandtl-number correlations for spherical particles. Because snow granules have a large specific surface area, air temperature can influence the timing of dry–wet transitions, although its effect is generally secondary to frictional heating and entrainment.

With the thermal boundary conditions now complete, **Section 8** turns to friction — the mechanical process most directly responsible for wet-avalanche runout behavior.

8. Flow Friction in Wet Snow Avalanches

Historical basis and modeling philosophy

Switzerland has historically relied on the classical two-parameter **Voellmy model** (Voellmy, 1955) for avalanche dynamics:

$$S_{\Phi} = \mu N_{\Phi} + \frac{\rho g V_{\Phi}^2}{\xi}$$

where N_{Φ} is the normal stress and V_{Φ} the avalanche velocity. Coulomb friction μ governs runout distance, while ξ controls terminal velocity — a clean separation of roles that has made the model durable in practice. In the standard RAMMS::Avalanche module, (μ, ξ) are fixed before the simulation begins.

RAMMS::Extended instead treats flow resistance as a dynamic process, evolving with the avalanche's changing internal state:

$$S_{\Phi} = \mu(R_{\Phi}, T_{\Phi}, m_{\Phi}) N_{\Phi} + \frac{\rho g V_{\Phi}^2}{\xi(R_{\Phi}, T_{\Phi}, m_{\Phi})}$$

Here R_{Φ} is the random kinetic energy (fluctuation energy, or "granular temperature") governing fluidization, T_{Φ} the mean core temperature (Sections 4–6), and m_{Φ} the meltwater content. S_{Φ} is precisely the friction term appearing in the momentum equation that balances gravitational driving stress in the RAMMS::Extended core.

This builds on **grain flow theory** (Haff, 1983), which treats the avalanche as a granular ensemble of flowing clumps and clods rather than a continuum fluid. A volume of densely packed dry snow particles — the **co-volume** (Section 4) — defines the reference state at which friction takes its baseline dry values (μ_0, ξ_0) . Two physically distinct processes then reduce friction below this baseline: **lubrication**, where meltwater softens basal sliding, and **fluidization**, where shear-induced dispersion increases particle spacing. Because wet flows develop cohesive forces that resist dispersion, these two mechanisms are largely mutually exclusive — a wet avalanche is either dense and lubricated, or fluidized and dry, rarely strongly both at once. This is a direct expression of Voellmy's original insight that dry, fluidized avalanches and dense, wet avalanches are not distinct phenomena, but limiting cases along a continuum.

The sections below follow the model's three-step calculation procedure: **lubrication**, **fluidization**, and **cohesion**.

Step 1 — Lubrication (wet snow)

As meltwater accumulates, the baseline dry coefficients evolve toward wet limits:

$$\begin{aligned}\mu' &= \mu_w + (\mu_0 - \mu_w)e^{\left(\frac{m_{\Phi}}{m_t}\right)} \\ \xi' &= \xi_w + (\xi_0 - \xi_w)e^{\left(\frac{m_{\Phi}}{m_t}\right)}\end{aligned}$$

where:

μ_0, ξ_0 = baseline dry friction coefficients,

μ_w, ξ_w = limiting wet friction coefficients,

m_t = transition parameter controlling the sensitivity to water content,

m_ϕ = meltwater content in the avalanche core.

These equations ensure that:

At **zero water content** $m_\phi = 0$, friction values remain at their dry limits $(\mu', \xi') = (\mu_0, \xi_0)$.

As water content increases, both μ_0 and ξ_0 decay exponentially toward their wet limits (μ_w, ξ_w) .

Physically, a thin water film — typically produced by frictional shear heating — reduces basal Coulomb resistance (μ' falls). Simultaneously, lubricated grains sinter into agglomerates, forming a semi-cohesive structure that inhibits rearrangement and stiffens the flow internally, raising the effective velocity-dependent resistance (ξ' falls, corresponding to greater turbulent drag). This dual softening/stiffening is the physical basis for the capillary-wetting picture already established below.

Transitional flows (initially dry, becoming wet).

For flows that begin dry or cold and transition toward wet, lubricated behavior:

$$\mu_w = 0.10\text{--}0.12, m_t = 1\text{--}10 \text{ mm}$$

The small m_t range reflects the high sensitivity of snow to initial wetting — even a thin meltwater film can trigger sintering and aggregation, sharply increasing resistance.

This recommendation is drawn directly from granulometric investigations of avalanche deposits (Bartelt & McArdell, 2009).

Fully saturated flows (wet from onset).

For flows where significant pore water is present from the start — where grains cannot form stable agglomerates and cohesive bonding is largely absent:

$$\mu_w = 0.08\text{--}0.15, m_t \approx 100\text{--}1000 \text{ mm}$$

Larger m_t reflects the diminished ability of already-saturated snow to generate cohesive structure; adjusting m_t upward in this regime prevents artificial stiffening in flows with little internal shear resistance to begin with.

For the turbulent term, RAMMS::Extended presently applies $\xi_w = 800 \text{ m s}^{-2}$ automatically for all wet-snow scenarios (Salm et al., 1990).

RAMMS | Run Simulation

General Params Friction Snowcover Forest Powder/Wet

Powder / Wet Parameters

Type: ☒ Flowing (Dry or Wet) ☐ Mixed Flowing/Powder ?

Generate (% 0-10)	5.0	—
Mu wet ()	0.12	—
Dry-wet-transition (mm)	100.00	—
Air temperature (°)	10.00	—

☐ Enable Turbulence Model Beta Powder: 2.00 (1/s)

Figure 9: Input tab for μ_w and dry-wet transition m_t .

Capillary wetting and friction evolution

The transition from dry to wet behavior is governed by a progressive capillary-wetting process. At the **onset of wetting**, thin water films and small liquid bridges form at grain contacts. These reduce basal sliding resistance while capillary bridges bind neighboring grains together, increasing internal stiffness — the capillary or transitional regime. As meltwater increases further, liquid bridges enlarge and water films extend across more of the granule surface: basal lubrication strengthens further, while the ability to sustain discrete aggregates weakens as the capillary structure becomes overextended — the **expanded capillary regime**. The model does not assume the grains become suspended in a continuous water phase or that the avalanche becomes debris-flow-like.

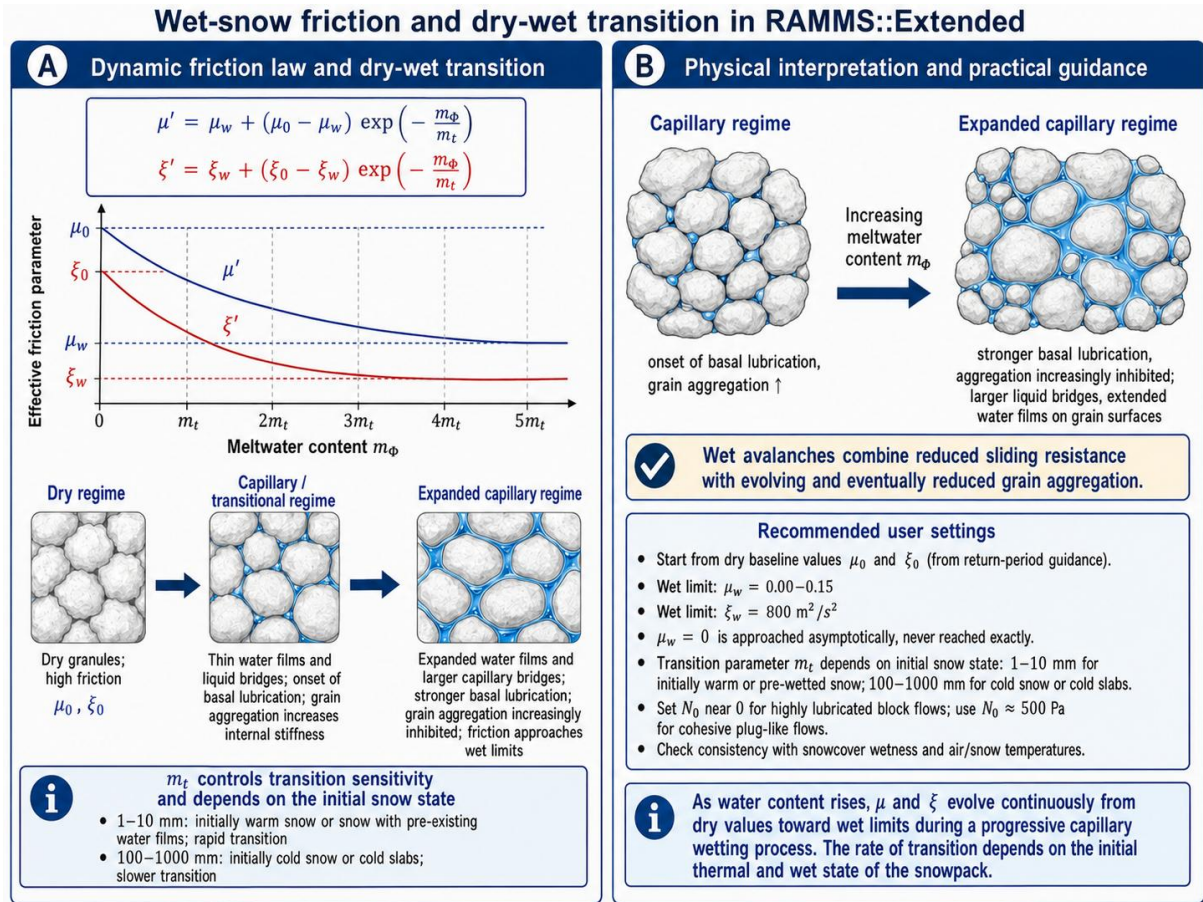


Figure 10: Wet-snow friction and dry-wet transition in RAMMS::Extended. The effective friction parameters μ' and ξ' decrease from dry toward wet limits as meltwater content increases. The transition parameter m_t controls how rapidly this occurs: 1–10 mm for initially warm or pre-wetted snow, and 100–1000 mm for cold snow or cold slabs. Increasing water content enlarges capillary bridges and surface films, strengthening basal lubrication while progressively reducing grain aggregation.

Step 2 — Fluidization

Once lubrication has set the residual friction values (μ' , ξ'), fluidization further reduces friction as granular fluctuation energy R_Φ increases and grains disperse:

$$\mu_f = \mu' e^{-R_\Phi/R_0} \quad \xi_f = \xi' e^{R_\Phi/R_0}$$

RAMMS::Extended uses a **fixed activation energy**, $R_0 = 2 \text{ kJ/m}^3$, **applied uniformly across all snow types**. Increasing R_Φ opens spaces between granules, reducing collisions and interactions, and thus lowering friction — the same fluidization process that governs powder-cloud formation in dry avalanches. Because wet flows develop strong cohesive resistance to dispersion, this mechanism is typically weak once meltwater content is significant; fluidization and lubrication are largely mutually exclusive regimes, consistent with Voellmy's original borderline-case framing.

Unlike R_0 , which is held fixed, the **production and decay of R_Φ itself** — governed by the parameters α_Φ and β_Φ — vary with snow conditions and avalanche regime. This formulation is supported by snow-chute experiments (Platzer et al., 2007) that revealed hysteresis in the shear–normal stress relationship: for a given normal stress, two distinct shear stresses could occur depending on flow history — a behavior the

classical constant-parameter Voellmy law cannot capture, but which a fluctuation-energy-dependent friction law reproduces naturally.

Step 3 — Cohesion

At low overburden stress, snow flow deviates from pure Coulomb behavior and exhibits genuine cohesion (Platzter et al., 2007; Bartelt et al., 2015). The Coulomb coefficient μ , measured as the secant slope of shear stress S against normal stress N , shows a marked change in curvature at a characteristic stress N_0 , separating cohesive and friction-dominated regimes:

$$S_{\Phi} = \mu_f N_{\Phi} - (1 - \mu_f) N_0 e^{N_{\Phi}/N_0} + (1 - \mu_f) N_0$$

As $N_{\Phi} \rightarrow 0$, the effective friction coefficient approaches its maximum ($\mu \rightarrow 1$) as cohesive forces dominate; as $N_{\Phi} \rightarrow \infty$, it asymptotes to the fluidized value μ_f . This formulation ensures shear stress vanishes as normal stress approaches zero — avoiding the unphysical prediction of nonzero shear at zero overburden — and, because N_0 primarily affects low flow heights, it suppresses the formation of unrealistic secondary flow arms in thin parts of the flow.

N_0 is set in the Friction Tab. For dry avalanches, minimal slip resistance applies ($0 \leq N_0 < 200$ Pa), characteristic of granular, non-cohesive (mixed-flowing or powder) snow. For wet avalanches, N_0 is not a single fixed choice — it scales with how much liquid water the flowing avalanche actually contains. Calibration against the Gatschiefer and Salezer Tobel case studies (Section 11) establishes the following relationship between the liquid water equivalent carried in the flowing avalanche core and the effective cohesion required to reproduce observed runout and deposition:

Table 5: Recommended effective cohesion as a function of liquid water equivalent in the flowing avalanche. Higher values are needed when there is a strong liquid-water gradient in the snowpack.

Liquid water equivalent in flowing avalanche (mm)	Effective cohesion (Pa)
0–10	200–400
10–20	400–750
20–35	750–1000
35–50	1000–1300
50–70	1300–1600
70–100	1500–2000

Higher values within each range are generally needed when there is a strong liquid-water gradient in the snowpack (Section 3). Two limiting cases fall outside this table: a fully lubricated **block-flow regime**, with no internal shearing, uses $N_0 = 0$ Pa; a transitional flow where meltwater is generated dynamically by friction rather than prescribed in the snowcover (Section 3) may require N_0 well below the 0–10 mm

bucket, since the liquid water equivalent available to the flow at any instant is small even though melting is ongoing — the VdIS case study (Section 11) is calibrated this way.

This table should be treated as a provisional engineering guideline, not a fixed physical law: it is derived from two events, and the physical connection between water content and cohesion — capillary bridging and progressive granule agglomeration — is qualitative rather than mechanistically derived from first principles. Departures from the table (as seen between individual case studies in Section 11) should be expected and are not necessarily errors.

Practical Guidance for Users

When defining wet-avalanche friction in the Friction Tab and Powder/Wet Tab:

Begin with baseline dry values μ_0 and ξ_0 appropriate for the selected return period.

For **transitional** flows (initially dry, wetting during motion): $\mu_w = 0.10\text{--}0.12$, $m_t = 1\text{--}10$ mm.

For **fully saturated** flows (wet from release): $\mu_w = 0.08\text{--}0.15$, $m_t \approx 100\text{--}1000$ mm.

Use the automatically assigned wet turbulent limit $\xi_w = 800 \text{ m s}^{-2}$.

For highly lubricated block-like flows, select N_0 close to zero. For cohesive, plug-like flows, values around $N_0 \approx 500$ Pa may be appropriate.

Check that the selected friction parameters are consistent with the prescribed snowcover water content (Section 2), snow temperature, and air temperature (Section 7).

With friction now fully specified, **Section 9** turns to the production and decay of the random kinetic energy R_ϕ itself — the quantity that drives the fluidization term derived above.

9. Generate Parameter (α_ϕ) and Random Energy Decay (β_ϕ)

Random kinetic energy production and decay

Section 8 showed how random kinetic energy R_ϕ reduces friction through fluidization ($\mu_f = \mu' e^{-R_\phi/R_0}$, $\xi_f = \xi' e^{R_\phi/R_0}$). This section turns to R_ϕ itself — specifically, the parameters α_ϕ and β_ϕ that govern its production and decay (Buser & Bartelt, 2009).

A central feature of RAMMS::Extended is its treatment of random kinetic energy, also called *fluidization energy*. This energy represents the random motions of snow granules that allow an avalanche to fluidize, disperse, and, in dry cases, generate a powder cloud. Wet avalanches behave differently because the presence of liquid water changes both the **production** and the **dissipation** of this random motion.

The production of random kinetic energy is controlled by the **Generate** parameter α_ϕ . Physically, α_ϕ describes how efficiently bulk shearing and particle collisions convert mean-flow energy into random granular motion. In dry avalanches, particle collisions

remain sufficiently elastic to sustain a significant level of fluctuation energy. Typical values are therefore

$$\alpha_{\Phi} = 0.07\text{--}0.08$$

for dry-snow avalanches.

In **wet avalanches**, however, liquid water forms capillary bridges between snow grains. These bridges make collisions sticky and inelastic, so that snow granules tend to agglomerate rather than disperse. As a consequence, only a small fraction of the shear work is converted into random motion. For this reason, the Generate parameter must be reduced, typically to

$$\alpha_{\Phi} = 0.03\text{--}0.05$$

for wet-snow avalanches.

Thus, wet avalanches are characterized not only by enhanced lubrication at the bed, but also by strongly reduced production of fluctuation energy within the flowing mass.

RAMMS | Run Simulation

General Params Friction Snowcover Forest Powder/Wet

Powder / Wet Parameters

Type: ☒ Flowing (Dry or Wet) ☐ Mixed Flowing/Powder ?

Generate (% 0-10)	5.0	—
Mu wet ()	0.12	—
Dry-wet-transition (mm)	100.00	—
Air temperature (°)	10.00	—

☐ Enable Turbulence Model Beta Powder: 2.00 (1/s)

Figure 11: Generate parameter α_{Φ} in the Powder/Wet Tab. The Generate parameter controls the production of random kinetic energy by shearing and particle collisions. Dry avalanches typically require $\alpha_{\Phi} = 0.07\text{--}0.08$, whereas wet avalanches require lower values, $\alpha_{\Phi} = 0.03\text{--}0.05$, because liquid water suppresses granular fluidization.

Decay of random kinetic energy

The persistence of random kinetic energy is controlled by the decay coefficient β_Φ . This parameter governs how rapidly fluctuation energy is dissipated once it has been generated.

In cold, dry avalanches, granular collisions are relatively elastic and random motion can persist for longer times. Typical values are therefore of the order

$$\beta_\Phi \approx 1 \text{ s}^{-1}.$$

In wet avalanches, by contrast, each collision is accompanied by energy loss associated with capillary bonding, rupture and re-formation of liquid bridges, and other dissipative contact processes. Random motion therefore decays much more rapidly, and characteristic values increase to about

$$\beta_\Phi \approx 2 \text{ s}^{-1}.$$

This means that wet avalanches not only produce less fluctuation energy, but also dissipate it much faster. In a practical sense, this is one of the defining characteristics of a wet avalanche: fluidization never becomes dominant.

Temperature dependence and physical motivation

The increase of β_Φ with warming avalanche temperature is not arbitrary. It is motivated by laboratory measurements of the temperature-dependent viscoelastic behaviour of snow (Scapozza, 2004), which show that the short-term mechanical relaxation response of snow follows an Arrhenius-type temperature shift,

$$\tau_i(T) = \tau_{i,\text{ref}} \exp \left[\frac{Q(T)}{R} \left(\frac{1}{T} - \frac{1}{T_{\text{ref}}} \right) \right],$$

where

$$R = 8.314 \times 10^{-3} \text{ kJ mol}^{-1} \text{ K}^{-1},$$

and

$$T_{\text{ref}} = 261.7 \text{ K}.$$

The activation energy $Q(T)$, determined from Scapozza's (2004) triaxial relaxation tests on snow, is defined piecewise as

$$Q(T) = \begin{cases} 10.37 T - 2747, & T > -10^\circ\text{C}, \\ 76, & T \leq -10^\circ\text{C}. \end{cases}$$

As temperature approaches the melting point, the activation energy decreases and the short-term relaxation time becomes much smaller. Equivalently, the inverse relaxation rate $1/\tau_i(T)$ increases sharply. This behavior reflects the growing importance of fast, dissipative grain-scale processes in warm snow.

RAMMS::Extended captures this same trend by defining the decay coefficient β_Φ as an increasing function of avalanche-core temperature T_Φ (Section 4),

$$\beta_\Phi = 1.40 + \frac{1.7}{\pi} \arctan [1.6 (T_\Phi - 272.5)].$$

This formulation produces a rapid increase in decay over the temperature interval where snow transitions from cold and dry to warm and nearly isothermal. In other

words, the RAMMS::Extended parameterization of β_ϕ is a practical surrogate for Scapozza's (2004) experimentally observed temperature dependence of short-term dissipation in snow.

Interpretation for wet avalanches

The combined effect of reduced production and accelerated decay of random kinetic energy explains the characteristic behavior of wet-snow avalanches.

Because α_ϕ is small, only limited fluctuation energy is produced.

Because β_ϕ is large, whatever fluctuation energy is generated **is rapidly dissipated**.

As a result, wet avalanches **do not sustain granular fluidization** (Section 8).

The flow therefore remains dense and plug-like, with little or no powder cloud.

Deposition tends to occur abruptly, especially at the avalanche tail, and runout is governed primarily by basal lubrication (Section 8) and entrainment effects (Section 10) rather than by dispersive fluidization.

Random kinetic energy production and decay in RAMMS::Extended

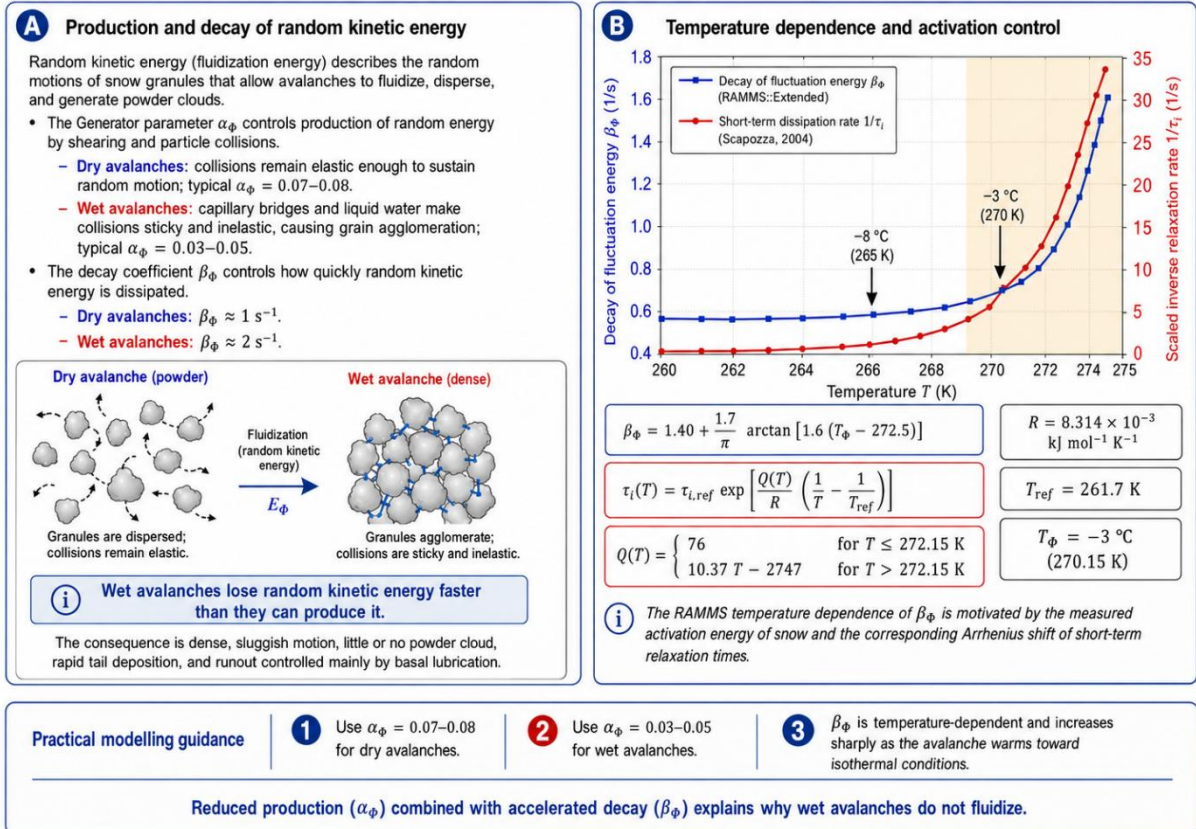


Figure 12: Temperature dependence of the decay coefficient β_ϕ . The RAMMS::Extended decay coefficient β_ϕ increases sharply as avalanche temperature approaches isothermal conditions.

The chosen parameterization is motivated by the experimentally observed Arrhenius-type temperature dependence of the short-term relaxation behavior of snow. The figure compares the RAMMS::Extended function $\beta_\phi(T_\phi)$ with the scaled inverse relaxation rate $1/\tau_i(T)$ derived from the measured activation energy $Q(T)$. Both curves show a strong increase as the avalanche warms from cold, dry conditions toward wet-snow conditions.

Practical modeling guidance

For practical applications, the following guidelines are recommended:

Dry avalanches: use $\alpha_\phi = 0.07\text{--}0.08$ with characteristic decay values near $\beta_\phi \approx 1 \text{ s}^{-1}$.

Wet avalanches: use $\alpha_\phi = 0.03\text{--}0.05$ with characteristic decay values approaching $\beta_\phi \approx 2 \text{ s}^{-1}$.

The parameter β_ϕ is not set manually in the same way as α_ϕ , but is defined automatically through its temperature dependence in RAMMS::Extended.

In summary, wet avalanches are characterized by a simple but important imbalance: they lose random kinetic energy faster than they can produce it (Buser & Bartelt, 2009). This is the physical reason why they remain dense, sluggish, and non-fluidized. With production and decay established, **Section 10** turns to entrainment — the process by which this same imbalance is fed by newly incorporated snow along the avalanche path.

10. Entrainment in Wet Snow Avalanches

Section 9 established that wet avalanches lose random kinetic energy faster than they produce it. This section shows how entrainment — the incorporation of new snow along the avalanche path — feeds directly into that imbalance, adding mass while further suppressing fluidization.

The entrainment routines in RAMMS::Extended are based on the elastic–plastic collision framework described in the *RAMMS::Extended User Manual* (Bartelt & Christen, 2026, Chapter 4). Entrainment increases the avalanche mass and simultaneously modifies the internal energy balance (Section 4). In particular, the entrainment process governs how impact energy is partitioned into heat, random kinetic energy, and irreversible dissipation.

For dry and powder-snow avalanches, frontal entrainment can generate substantial random kinetic energy. This promotes particle dispersion, fluidization of the avalanche front, and the formation of a powder cloud.

Wet-snow entrainment behaves differently. Wet snow covers are generally dense, warm and often close to isothermal conditions, with typical densities of

$$\rho_\Sigma \geq 200\text{--}300 \text{ kg m}^{-3}.$$

Their entrainment produces comparatively little random kinetic energy. The entrained snow is incorporated into the dense avalanche core, the temperatures of the two materials rapidly equilibrate, and the avalanche increasingly develops a dense, plug-like flow structure.

Key differences from dry- and powder-snow entrainment

Dense, warm snow

Wet snow is heavier and more compact than fresh snow. Its erosion therefore requires more energy, while the accelerated mass is more dissipative. **This slows the avalanche relative to powder entrainment.**

Immediate mixing and temperature balance

Entrainment of wet snow is modeled as **immediate mixing into the avalanche core.**

Both snow mass and liquid water are instantly redistributed, causing a rapid adjustment of core temperature. This can accelerate the transition toward a wet flow regime.

Suppressed fluctuation energy (ε_Φ). The production of random kinetic energy during entrainment is strongly damped in warm, wet snow. Instead of dispersive fronts and powder clouds, the avalanche tends toward dense, plug-like flow. The entrainment partitioning factor is

$$\varepsilon_\Phi = f(T_\Sigma) g(r) = \left[\exp\left(-\frac{(T_\Sigma - 267)^2}{10}\right) \right] \left[\frac{r^2 \exp\left(-\frac{r^2}{2a^2}\right)}{a^3} \right]$$

where $r = h_\Phi/h_\Sigma$ is the ratio of avalanche flow height to snowcover height, and $a = 1.1$. The Gaussian temperature term peaks near $T_\Sigma \approx -6^\circ\text{C}$ (267 K) and falls toward zero as $T_\Sigma \rightarrow 0^\circ\text{C}$ —reflecting the near-total absence of fluidization energy generated when entraining warm, wet snow (Figure 11).

Thermal feedback

Entrainment of isothermal, near-melting snow introduces both mass **and heat**. This increases avalanche core temperature, promoting meltwater generation and further reducing fluidization. The thermal feedback is therefore much stronger than in cold snow entrainment.

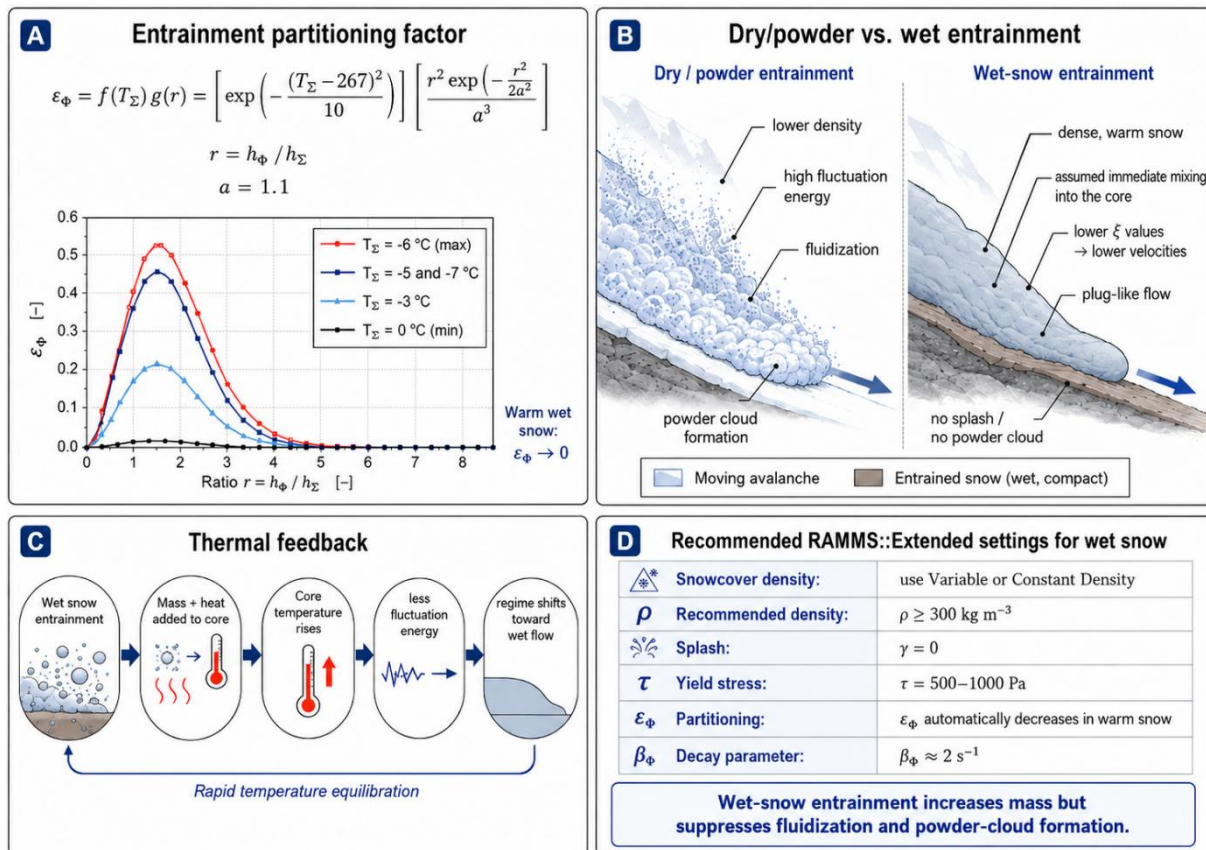


Figure 13: The entrainment partitioning factor ε_ϕ as a function of temperature and impact geometry. Entrainment of wet snow is highly dissipative, hindering the formation of powder clouds.

Parameterization in RAMMS::Extended

To capture these behaviors, users should adjust the snowcover input fields in the **Snowcover Tab** as follows:

- Release and erosion densities:**
 Use the **Variable** or **Constant Density** options (not the 3-day settlement method). Densities $\geq 300 \text{ kg/m}^3$ are recommended for wet, compact snowcovers.
- Splash (γ):**
 Set Splash = 0. Wet snow behaves plastically; no splashed front develops.
- Yield stress (τ):**
 Increase from the default 300 Pa to 500–1000 Pa. This limits entrainment to locations where the avalanche has sufficient mass and velocity and suppresses lateral flow arms.
- Partitioning parameter (ε_ϕ):**
 The parameter ε_ϕ is automatically reduced by snow temperature and avalanche geometry (

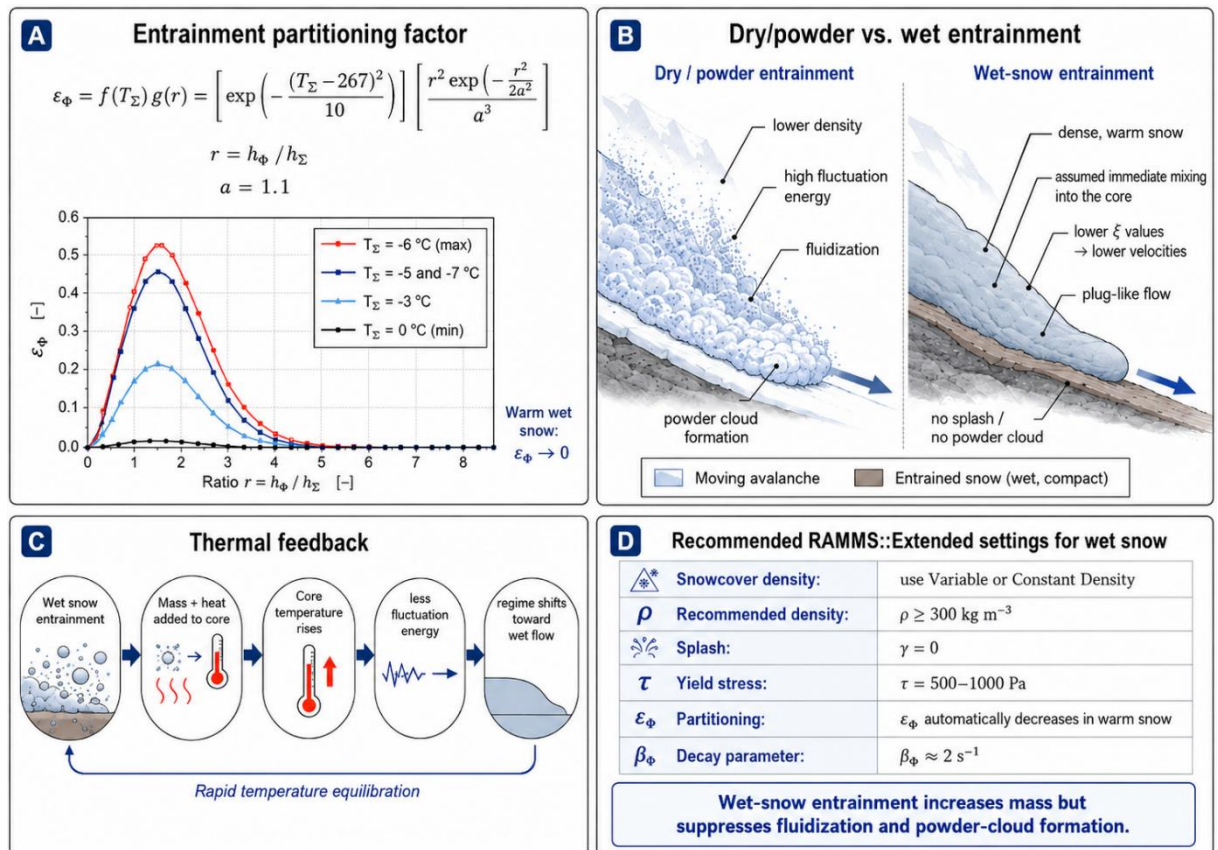


Figure 13). In wet snow conditions, $\varepsilon \rightarrow 0$, reflecting the near absence of fluidization energy.

- **Decay parameter (β_ϕ):**

Wet snow entrainment accelerates the decay of random kinetic energy. Typical values are $\beta_\phi \approx 2 \text{ s}^{-1}$ (vs. $\approx 1 \text{ s}^{-1}$ for dry snow).

Practical Recommendations

Use high-density snowcovers ($\rho \geq 300 \text{ kg/m}^3$) when simulating entrainment into old, wet snow.

Suppress splash and raise yield stress to mimic the strongly plastic and dissipative character of wet snow entrainment.

Be aware that entrainment in warm conditions not only increases avalanche mass but also heats the core, **which can trigger or reinforce a regime transition from fluidized to wet flow.**

Remember that entrainment of wet snow inherently reduces ε and accelerates β_ϕ , producing avalanches that are dense, slower, and lacking in powder clouds.

Wet-snow entrainment therefore **increases avalanche mass while simultaneously suppressing fluidization**. Through the combined effects of high snow-cover density, low entrainment partitioning, rapid temperature equilibration, faster decay of random kinetic energy and increased flow resistance, the avalanche evolves toward dense, slower and predominantly plug-like motion without a significant powder cloud.

With the full physical framework now established — snowcover state (Sections 2–3), governing thermodynamics (Sections 4–6), and the boundary conditions and dynamic processes that feed it (Sections 7–10) — **Section 11** turns to six documented case studies that ground these parameter recommendations in real, calibrated events.

11. Case Studies

VdIS 25.02.1999

The avalanche of 25 February 1999 (estimated volume: $V_0 = 316,000 \text{ m}^3$) is an outstanding example of a **dry-wet transitional avalanche**. Information about this avalanche is documented in the 1999 VdIS report.

In the week preceding the event, between 16 and 23 February, approximately 190 cm of new snow accumulated in the VdIS release zone. During this period, air temperatures at 2,400 m (the approximate elevation of the starting zones) rose to about -4°C between 18 and 22 February. This warming phase densified the snowpack considerably, with mean release zone densities of roughly 300 kg/m^3 (range: $260\text{--}340 \text{ kg/m}^3$, measured). It was followed by a colder period, with air temperatures dropping to -15°C at 2,400 m. At the time of release, the snowpack can therefore be described as relatively warm in its interior (around -4°C in the release zone), overlain by colder near-surface layers (around -10°C).

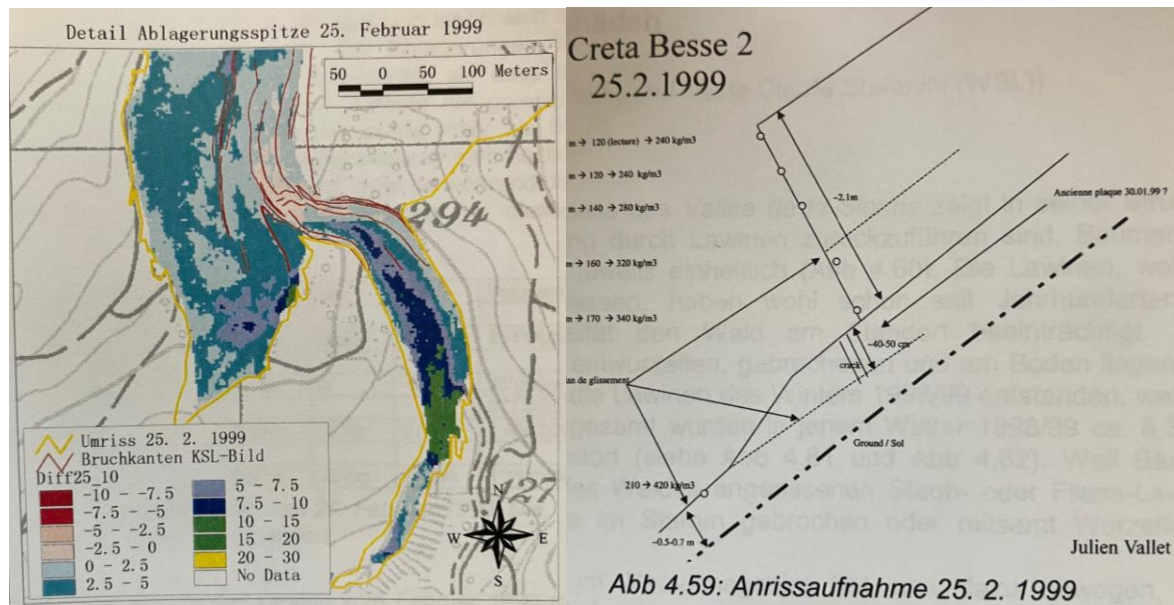


Figure 14: Measured depositions at the end of the 25.02.1999 VdIS avalanche in the region of Seillon (left). Measured snow densities near the fracture crown of the avalanche (right). Depth below surface → estimated density: 0.2 m → $\approx 260 \text{ kg m}^{-3}$; 0.5 m → $\approx 260\text{--}280 \text{ kg m}^{-3}$; 0.9 m → $\approx 280\text{--}300 \text{ kg m}^{-3}$; 1.4 m → $\approx 320 \text{ kg m}^{-3}$; 2.0 m → $\approx 340 \text{ kg m}^{-3}$; At/near the sliding plane ("Plan de glissement") → a note indicates $\approx 420 \text{ kg m}^{-3}$. Both documents are taken from the 1999 VdIS report.

The avalanche displayed a combination of **dry-snow behavior** and **wet-snow dynamics**. Its powder cloud rose dramatically during descent, destroying large sections of forest on the opposite slope of Berguges, while its wet-snow runout left massive depositions stretching the entire length of the Berguges valley down to 1,300 m. The avalanche front continued more than one kilometer beyond the VdIS bunker, following the course of the Sionne river. A further portion of the flow advanced into the

Seillon area, where deposition depths reached up to 5 m. These extensive deposits are particularly valuable for constraining wet-snow transitional parameters μ_w and m_t .

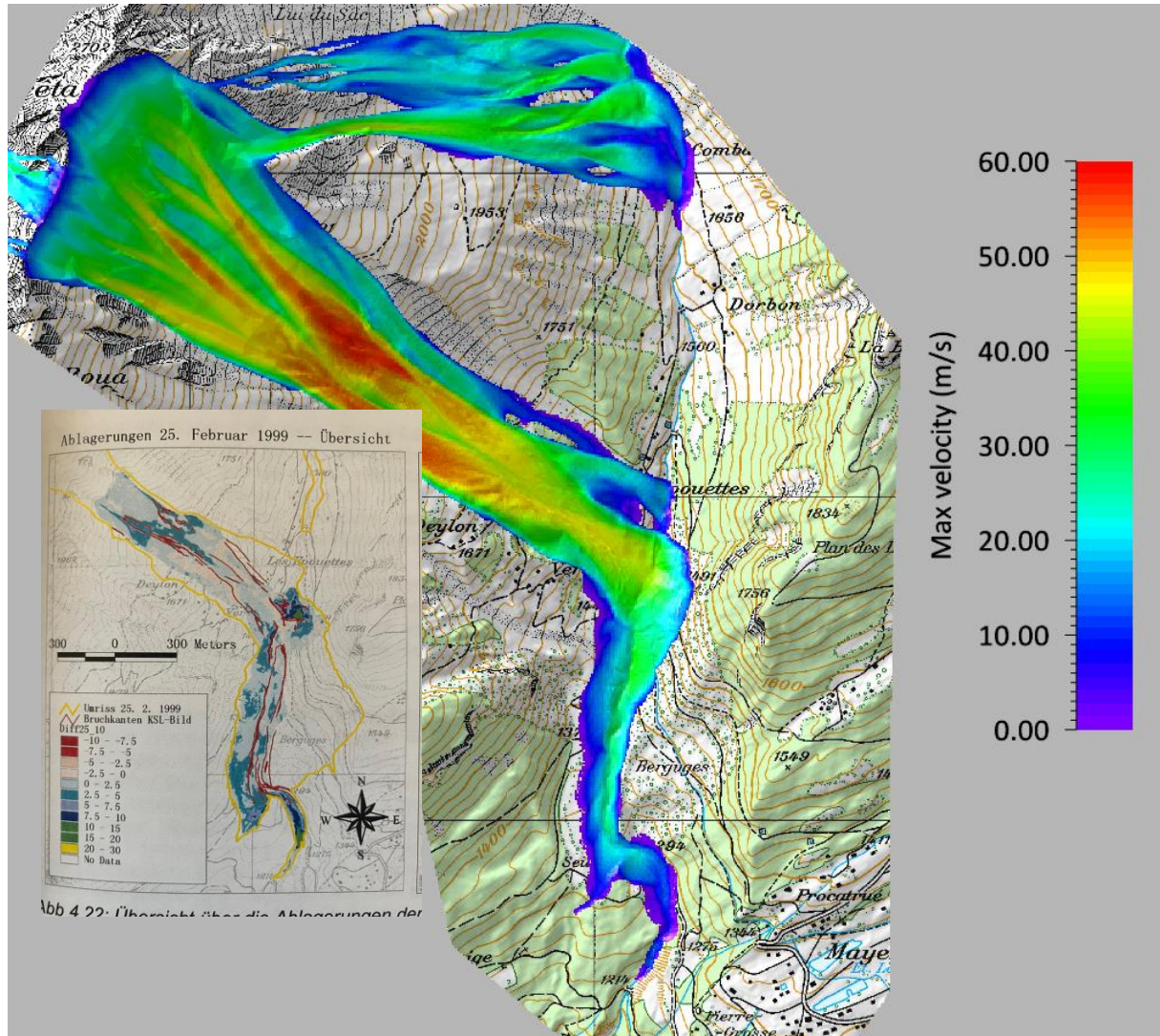


Figure 15: Simulated maximum flow velocity and runout of the 25.02.1999 VdIS avalanche. The avalanche reaches a peak velocity of 60m/s. Maximum powder cloud velocities varied between 60m/s and 70m/s. Note the long simulated and observed runout to the region of Seillon.

To simulate the VdIS avalanche of 25 February 1999, we relied extensively on information from the official avalanche report, including the mapped extent of the release zone. The release area exhibited highly variable crown heights, ranging from 0.5 m to 3.0 m (report). For the simulation, we adopted the mean crown height of 1.5 m (report), yielding a release volume of approximately 315,800 m³ (in close agreement with the reported value of 316,000 m³).

The snowpack in the release zone (at ~2600 m) was specified as relatively warm, with an assumed temperature of $T_0 \approx T_\Sigma = -5^\circ\text{C}$, reflecting the preceding warming phase (T_Λ up to -4°C) with colder near-surface layers. The snow was treated as dry ($m_\Sigma = 0$), with a generation parameter $\alpha_\phi = -8\%$ the upper end of the dry-avalanche range (Section 9), consistent with elastic collisions sustaining fluctuation energy in cold, unwetted snow. Curvature was set to ON.

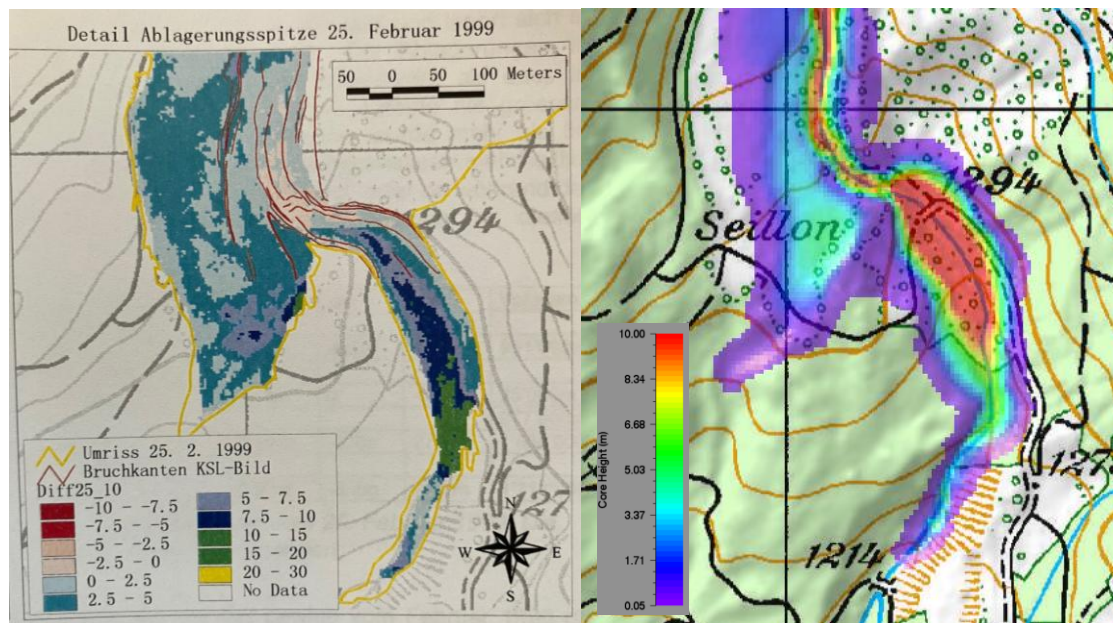


Figure 16: Measured and calculated runout VdIS avalanche 25.02.1999. Calculated deposition heights in the left arm were over 5m (observed) and in the Sionne torrent over 10m (observed). The long runout could only be reached with transitional wet snow parameters of $\mu_w = 0.15$ and $m_t = 2\text{mm}$.

A deep snow cover of $h_\Sigma = 150\text{cm}$ was prescribed, with a bulk density of $\rho_\Sigma = 300\text{kg/m}^3$, slightly lower than the reported 190 cm snowfall accumulation to account for settlement. Standard vertical gradients were imposed for snow depth ($\Delta h_\Sigma = 30\text{ cm/100 m}$) and snow temperature ($\Delta T_\Sigma = 0.5^\circ\text{C/100 m}$), using the **variable** density option. The resulting entrainment gave a calculated end volume of 1.0 million m^3 (measured: $\sim 980,000\text{ m}^3$).

The avalanche was simulated on a $5\text{m} \times 5\text{m}$ grid as a **100Y large avalanche (100L)** scenario, reaching calculated peak velocities of 65 m/s, in agreement with photogrammetric front-tracking (report).

For the wet-snow runout, we selected $\mu_w = 0.15$ and a small transition value $m_t = 2\text{mm}$, reflecting the warm, compacted interior of the snowcover, where capillary effects raise sliding friction quickly with even modest wetting. Because VdIS remains a cold avalanche overall — meltwater is generated dynamically by friction rather than prescribed in the snowcover ($m_\Sigma = 0$) — cohesion was taken from the dry-avalanche category ($0 \leq N_0 < 200\text{ Pa}$, Section 8) rather than the wet-snow relationship in Table 7, giving $N_0 = 75\text{Pa}$.

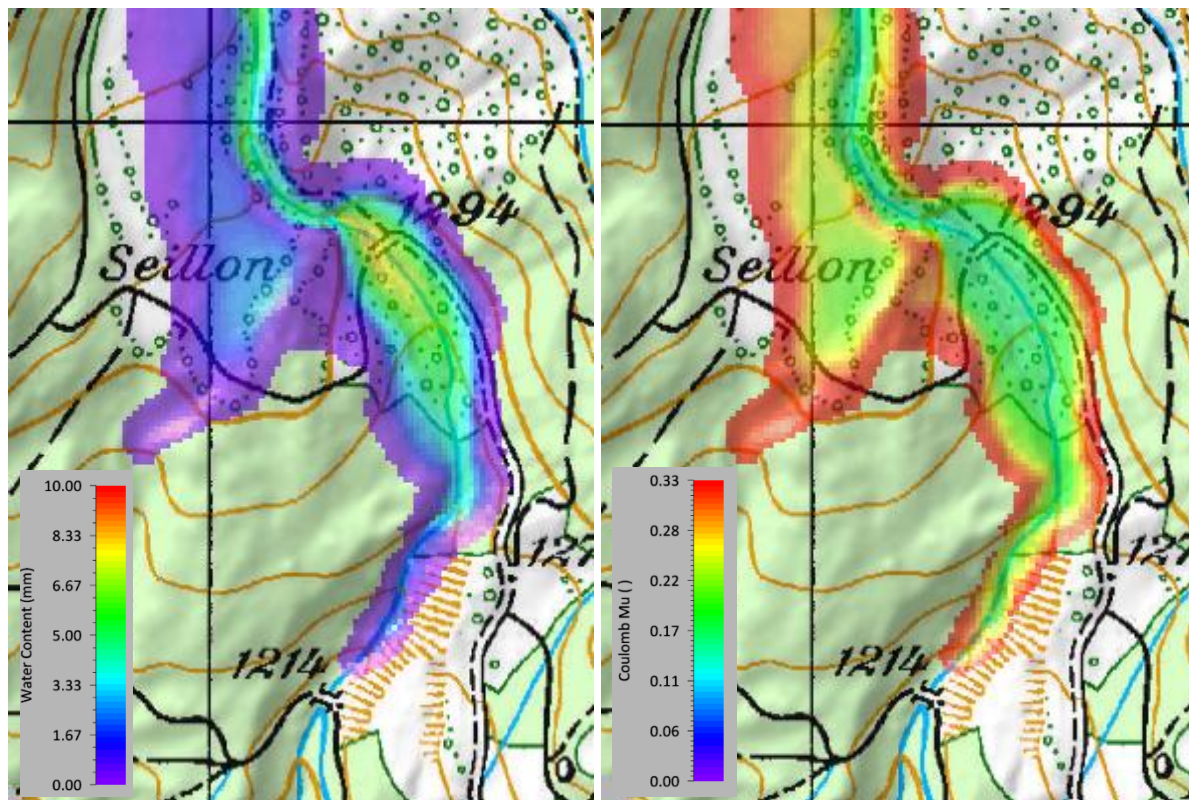


Figure 17: (Left) Calculated water content in the avalanche deposits. Only 10mm of water is found in the deposits, indicating thin water films on the granule surfaces. (Right) Calculated Coulomb friction μ_w . Higher friction values are found on the lateral edges of the flow.

In this case study, **air temperature** proved to be a key factor in controlling the dry-wet transition. When colder values of $T_A = -10^\circ\text{C}$ were prescribed, runout distances were underestimated, as melting and liquid water production were largely suppressed. Realistic runout behavior required warmer air temperatures of about $T_A = -6$ to -7°C , which allowed the formation of a wetting layer and a corresponding reduction in basal friction through increased water content.

Table 6: VdIS simulation parameters. Transitional dry-wet avalanche.

Case Study: Dry-Wet Transitional Snow Avalanche			
Case study name	VdIS 200 25.02.1999	Grid resolution (m)	5mx5m
Avalanche size / Return period	100L	Forest present (yes/no)	Yes
Starting volume (m ³)	315'800	End volume (m ³)	1 mio
Snowcover Information			
Reference altitude (m a.s.l.)	2600	Snow height gradient Δh_z (m/100 m)	0.03

Snow temperature T_{Σ} (°C)	-5	Snow temp. gradient ΔT_{Σ} (°C/100 m)	0.5
Liquid-water m_{Σ} (mm)	0	Water gradient Δm_{Σ} (mm/100 m)	0
Erosion depth h_{Σ} or d_0^* (m)	1.50	Density method (3-day/variable/constant)	Variable
Erosion density ρ_{Σ} (kg/m³)	300	Air temperature T_{Λ} (°C)	-7
Release density ρ_0 (kg/m³)	300	Release depth d_0 (m)	1.50
Powder Parameters			
Generate α_{ϕ} (%)	8	Cloud drag (-)	3
Cliff factor (-)	2	Air entrainment ψ (m/s scale)	2
Turbulence decay β_{II} (s⁻¹)	2.00		
Wet Snow Friction			
μ_w (wet sliding friction) (-)	0.15	m_t (transition thickness, mm)	2
Cohesion N_0 (Pa)	75	Stopping %	5
Comments			
A deep snow cover of $h_{\Sigma}=150\text{cm}$ with a bulk density of $\rho_{\Sigma}=300\text{ kg/m}^3$ was prescribed, slightly lower than the reported 190 cm snowfall to account for settlement. Standard vertical gradients were applied for snow depth ($\Delta h_{\Sigma}=30\text{cm}/100\text{ m}$) and snow temperature ($\Delta T_{\Sigma}=0.5\text{C}/100\text{ m}$). Under these entrainment conditions, the avalanche reached a simulated end volume of 1.0 million m^3, closely matching the measured estimate of 980,000 m^3. The event was modeled on a $5 \times 5\text{ m}$ computational grid as a 100Y large avalanche (100L) scenario. To reproduce the wet-snow runout, we selected a sliding friction coefficient of $\mu_w=0.15$ together with a small transition length $m_t=2\text{ mm}$, representing a large avalanche with high flow depths traveling over a compacted, somewhat warm snow cover where capillary effects likely decreased basal friction.			

Gatschiefer 23.04.2008

On **23 April 2008**, a large naturally released wet-snow avalanche occurred in the Gatschiefer avalanche path near Klosters, Switzerland. The avalanche released from two slab-like starting zones between approximately 1900 and 2300 m a.s.l. The upper release zone exhibited a measured fracture depth of approximately 2.2 m, while the lower release zone released approximately 2.0 m of the seasonal snow cover.

Airborne laser-scanning measurements performed immediately after the event confirmed that nearly the entire snowpack was eroded to the ground over both release areas. The resulting release volumes were estimated to be approximately 308,000 m³ in the upper release zone and 42,000 m³ in the lower release zone, giving a total initial release volume of approximately 350,000 m³.



Figure 18: Gatschiefer track and avalanche deposits. Photo taken shortly after the event in 2008 (SLF).

The avalanche occurred following several days of spring warming accompanied by rainfall and wet snowfall. Air temperatures near the release elevations were close to the melting point, and the seasonal snowpack had undergone substantial warming prior to release. However, no direct measurements of the liquid water content of the release snowpack were available. Consequently, the initial moisture content must be regarded as uncertain and was treated as a calibration parameter in the numerical simulations.

The avalanche descended through two confined gullies before entering a broad runout zone where it impacted a deflecting dam constructed after the avalanche winter of 1999. A detailed airborne laser-scanning survey performed three days after the event produced a high-resolution digital snow surface model with a spatial resolution of approximately 0.5 m, allowing release and deposition volumes to be reconstructed. Comparison with a summer digital terrain model indicated a final deposited snow volume of approximately 362,000 m³, reflecting densification of the released snow during motion. In addition, a mobile-phone video recorded from the opposite side of the valley documented avalanche velocities, roll-wave development and the interaction between the avalanche and the deflecting dam, making the Gatschiefer avalanche one of the most comprehensively documented wet-snow avalanches available for model validation.

Using the wet-snow module in RAMMS::Extended, the simulated final deposited volume increased to approximately 460,000 m³ owing to entrainment along the avalanche path. **Approximately 300,000 m³ of snow was deposited in the upper deposition area immediately upstream of the deflecting dam, while the remaining 160,000 m³ reached the lower deposition zone beyond the dam.** These observations provide an exceptionally well-constrained data set for calibrating both the mass balance and the rheological parameters governing wet-snow avalanche motion.

Table 7: Gatschiefer avalanche. Wet snow avalanche.

Case Study: Wet Snow Avalanche			
Case study name	Gatschiefer 23.04.2008	Grid resolution (m)	5mx5m
Avalanche size / Return period	100L	Forest present (yes/no)	Yes
Starting volume (m ³)	350'000	End volume (m ³)	462'000
Snowcover Information			
Reference altitude (m a.s.l.)	2600	Snow height gradient Δh_{Σ} (m/100 m)	0.05
Snow temperature T_{Σ} (°C)	-1	Snow temp. gradient ΔT_{Σ} (°C/100 m)	0.1
Liquid-water m_{Σ} (mm)	0-100	Water content gradient Δm_{Σ} (mm/100 m)	0-7.5
Erosion depth h_{Σ} or d_0^* (m)	2.0	Density method (3-day/variable/constant)	Variable
Erosion density ρ_{Σ} (kg/m ³)	300	Air temperature T_{Λ} (°C)	-1

Release density ρ_0 (kg/m ³)	300	Release depth d_0 (m)	1.50
Wet Snow Friction			
Generate α_ϕ (%)	5	Yield stress (Pa)	500
μ_w (wet sliding friction) (–)	0.15	m_t (transition thickness, mm)	100
Cohesion N_0 (Pa)	200–2000	Stopping %	5
Comments			
Together with Salezer Tobel, this case study established the effective-cohesion-vs-water-content relationship in Table 7 (Section 8). The two events do not fall on an identical curve — see discussion below.			

Calibration of Initial Water Content and Cohesion

Because no measurements of liquid water content were available for either the Gatschiefer or Salezer avalanches, the initial moisture conditions of the release snowpack remain unknown. To investigate the sensitivity of the wet-snow model to this uncertainty, the initial liquid water content of the release snowpack was systematically varied between 0 and 100 mm (up to roughly 5% of the 2.0 m erosion depth), consistent with the effective liquid-water input framework in Sections 2–3. For each assumed water content, the corresponding liquid water equivalent within the flowing avalanche was estimated and the simulations were repeated.

The results demonstrate that the observed avalanche behavior can be reproduced over a range of initial water contents, provided that the **effective cohesion** of the flowing snow is increased consistently with increasing liquid water content. **Physically, this reflects the formation of capillary bridges and the progressive agglomeration of snow granules as liquid water is transferred from grain surfaces into the interior of the growing agglomerates. At the same time, basal lubrication reduces the effective friction at the avalanche base. Thus, increasing liquid water simultaneously produces a more cohesive avalanche core and a more lubricated basal layer.**

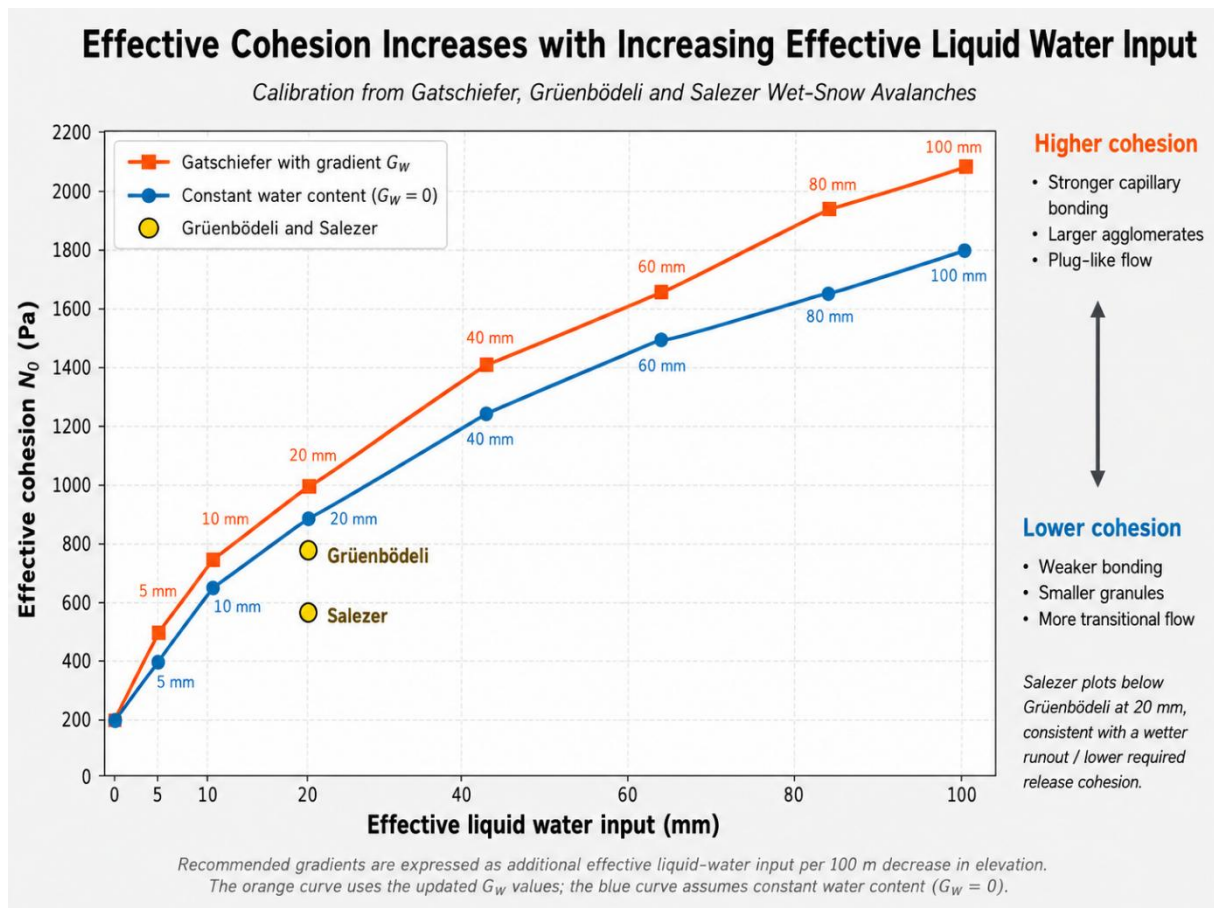


Figure 19: Increasing cohesion values for increasing liquid-water content.

Although the initial percentage water content is uncertain, the simulations indicate that the effective cohesion is more naturally related to the **liquid water equivalent contained within the flowing avalanche**, expressed as an equivalent water depth (mm), rather than the percentage water content of the release snowpack alone. This quantity accounts implicitly for differences in snow depth between avalanche paths and therefore provides a more transferable engineering parameter for wet-snow avalanche simulations.

The recommended calibration ranges derived from the Gatschiefer and Salezer case studies are summarized in **Table 5**. Individual events do not fall precisely on a single curve — Grünbödéli and Salezer Tobel, for example, require different cohesion at the same 20mm water content — indicating that factors beyond water content alone (entrainment geometry, elevation range, local water-content gradient) also influence the required calibration. Table 5 should be read as a central tendency, not an exact law. These values should be regarded as provisional engineering guidelines until direct measurements of liquid water content in natural wet-snow avalanches become available.

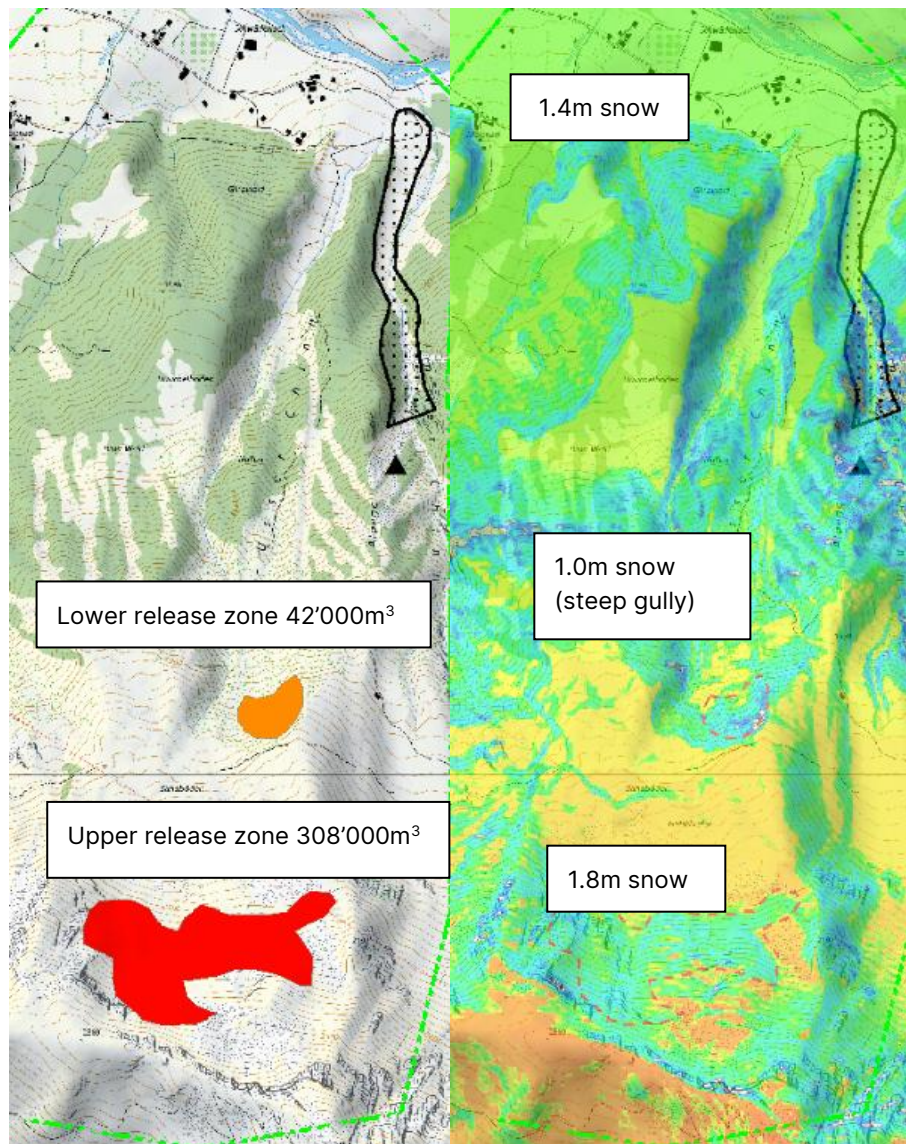


Figure 20: Release zones and initial snow distribution Gatschiefer case study.

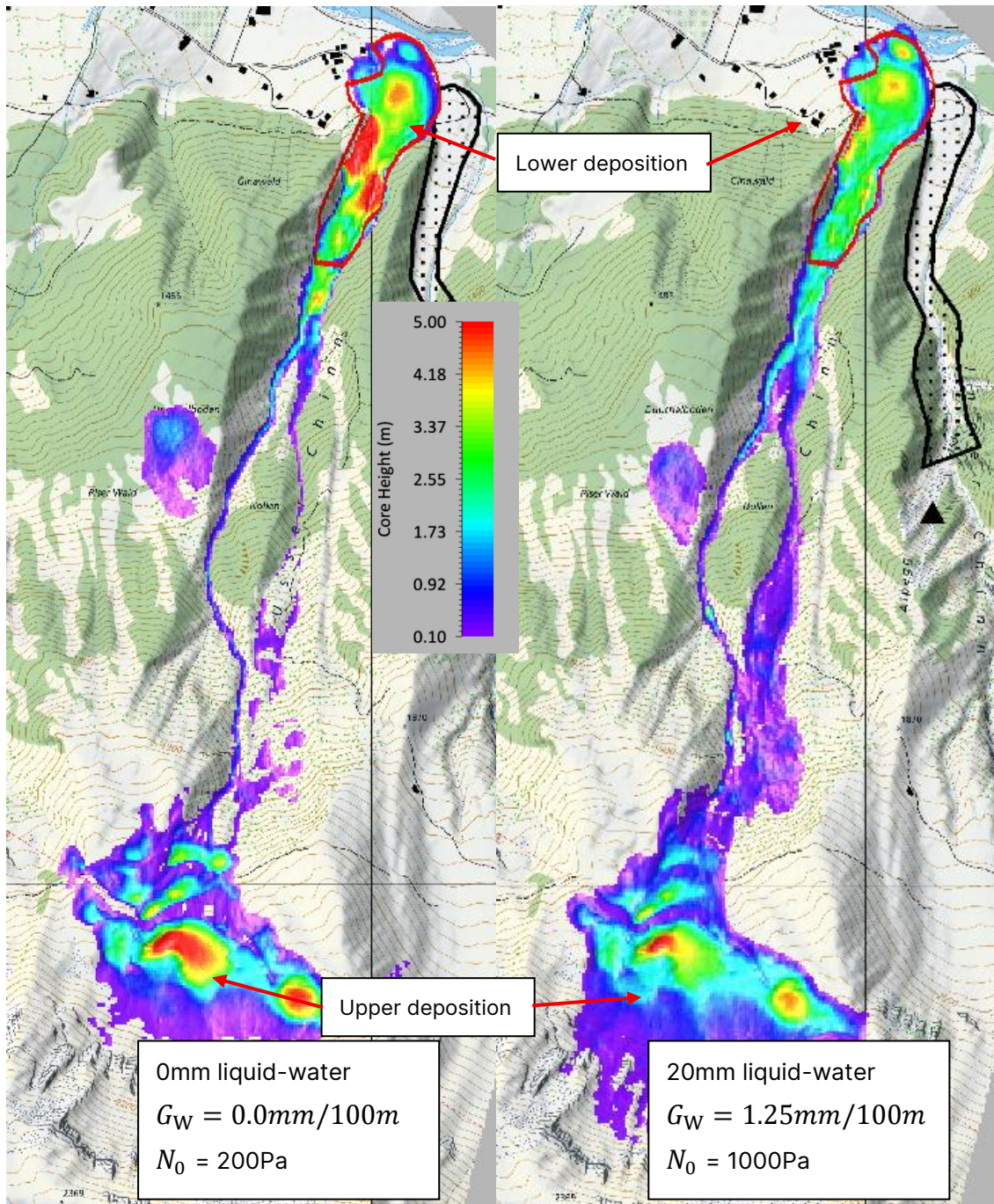


Figure 21: Two simulations of the Gatschiefer avalanche, both showing a good agreement with the final depositions (lower). Left: 0mm water content and $N_0 = 200\text{Pa}$ with $G_W = 0\text{mm}/100\text{m}$. Liquid-water comes completely from melting. More deposition in the lower runout zone. Right: 20mm water content and $N_0 = 1000\text{Pa}$ with $G_W = 1\text{mm}/100\text{m}$. More deposition in the upper zone.

Grüenbödeli 23.04.2008

The Grüenbödeli avalanche of 23 April 2008 occurred probably at approximately the same time as the avalanches on the opposite Gatschiefer slope. The release area covered about 63'000m³, although part of the snowcover outside the actual avalanche track had already been entrained by previous avalanches. Laser-scan measurements indicated an average pre-event snow depth of about 0.55m in the release zone, corresponding to an estimated release volume of approximately 35'000m³. For the simulation, the lidar-derived geometry was used directly, including a fracture height of 0.60m, the measured entrainment depths along the track, and an effective liquid-water input of 20mm in the snowcover.

Table 8: Grüenbödeli 2008 simulation parameters.

Case Study: Wet Snow Avalanche			
Case study name	Grüenbödeli 23.04.2008	Grid resolution (m)	2mx2m
Avalanche size / Return period	100M	Forest present (yes/no)	No
Starting volume (m ³)	35'000	End volume (m ³)	66'500
Snowcover Information			
Reference altitude (m a.s.l.)	2000	Snow height gradient Δh_{Σ} (m/100 m)	0.03
Snow temperature T_{Σ} (°C)	-1	Snow temp. gradient ΔT_{Σ} (°C/100 m)	0.5
Snow water content m_{Σ} (mm)	20	Water content gradient Δm_{Σ} (mm/100 m)	1.00
Erosion depth h_{Σ} or d_0^* (m)	0.60	Density method (3-day/variable/constant)	Variable
Erosion density ϱ_{Σ} (kg/m ³)	300	Air temperature T_{Λ} (°C)	-1
Release density ϱ_0 (kg/m ³)	300	Release depth d_0 (m)	0.60
Wet Snow Friction			
Generate α_{ϕ} (%)	5	Yield stress (Pa)	500
μ_w (wet sliding friction) (–)	0.15	m_t (transition thickness, mm)	100
Cohesion N_0 (Pa)	800	Stopping %	5

Comments
The simulation used a Generate value of 5%, a snowcover yield stress of 500 Pa, a wet-snow sliding friction coefficient of $\mu_w=0.15$, and a transition thickness $m_t=100$ mm. Warm, approximately isothermal snow conditions were assumed. An effective liquid-water input of 20 mm was prescribed in the snowcover, corresponding to a wet-snow cohesion of approximately $N_0 = 800Pa$. Importantly, these parameters were not adjusted independently to reproduce the Grünbödeli event; rather, they form a consistent parameter set comparable to that applied in the Gatschiefer and Salezer simulations. The good agreement obtained using the measured release and entrainment geometry therefore provides further support for the adopted wet-snow parameterization.

The final deposition volume was estimated at about 81,200 m³, indicating that both entrainment and snow densification contributed substantially to the final deposit. A cohesion value of $N_0 = 800Pa$ was used to reproduce the observed behaviour — near the upper end of Table 5's guidance for 20mm water content. This is higher than the $N_0 = 550Pa$ used for Salezer Tobel and lower than the $N_0 = 1000Pa$ used for Gatschiefer, despite comparable nominal liquid-water inputs; as noted in Section 8, individual events do not fall on a single curve, and event-specific geometry and entrainment history evidently modulate the required cohesion beyond water content alone. The Grünbödeli simulation therefore provides an additional, largely independent test of the wet-snow parameterization — using values drawn from the same calibration framework rather than tuned to match it exactly.

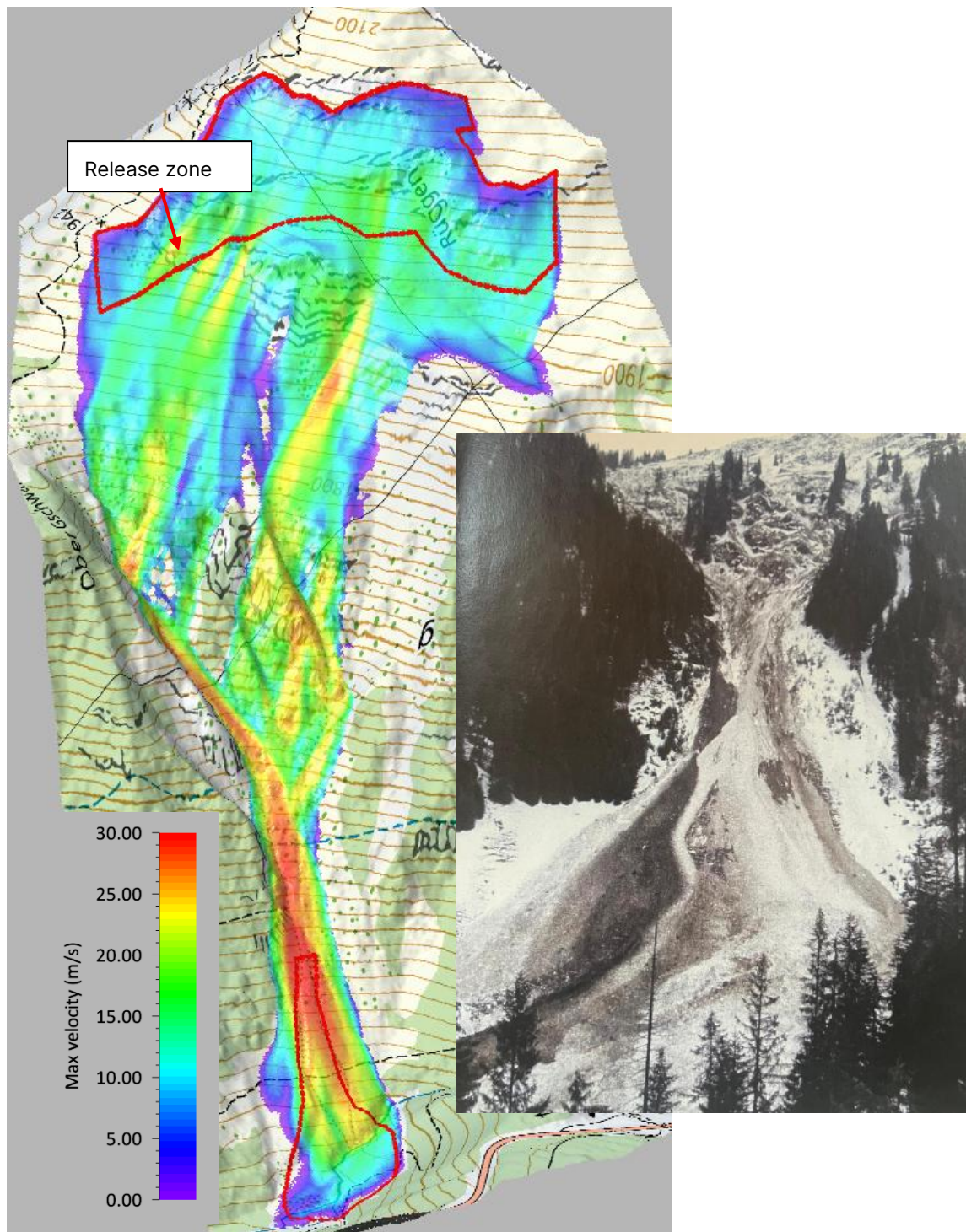


Figure 22: Calculated maximum velocities. Inset contains a post-event photograph of the event.

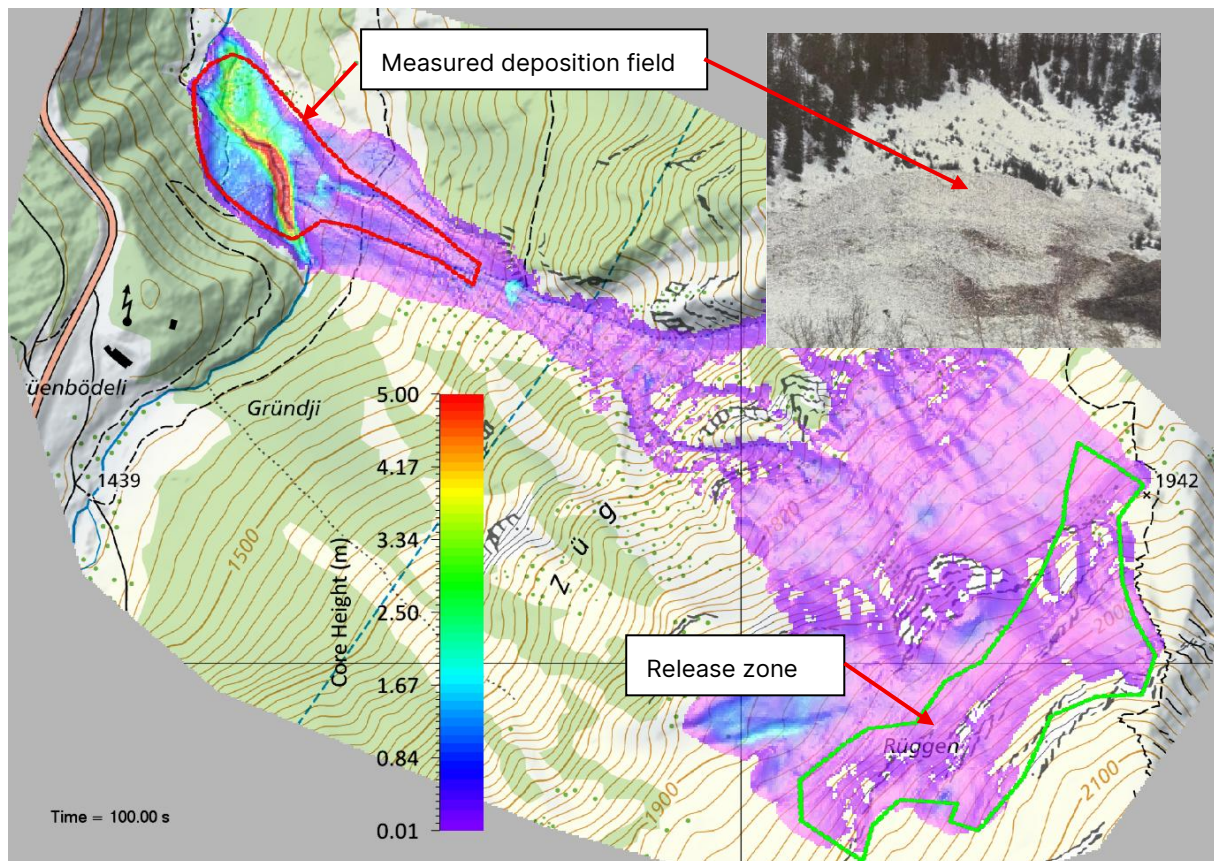


Figure 23: Calculated deposition 2008 Grünbödeli avalanche.

Salezzer Tobel Avalanche 23.04.2008

The Salezzer Tobel avalanche occurred on 23 April 2008 on the Dorfberg above Davos during a period of pronounced spring warming. Snow conditions in the region were exceptional for the season: at the Weissfluhjoch experimental site, a snow depth of 2.85 m was recorded, corresponding to approximately 135% of the long-term mean. Bühler et al. (2009) report that an abrupt warming period immediately before the event considerably increased the danger of wet-slab avalanches and produced widespread avalanche activity in the Davos region. The Salezzer avalanche formed the largest accessible avalanche deposit within the area surveyed during the subsequent airborne remote-sensing campaign.

The event was characterized by a complex sequence of releases and secondary entrainment along the avalanche track. Two principal release zones were identified. Release zone 1 covered approximately 31,200 m², while release zone 2 had an area of approximately 13,500 m². With a representative release depth of about 0.9 m, these areas correspond to initial release volumes of approximately 28,100 m³ and 12,100 m³, respectively. Additional secondary releases occurred during the descent, resulting in an estimated total released volume of approximately 67,000 m³. The final deposition volume was estimated at about 35,750 m³, although local deposition along the strongly varying avalanche path could not be quantified completely. The avalanche therefore represents a useful case study of a moderately wet, warming-driven spring avalanche with substantial redistribution of snow along the flow path.

Based on the documented rapid warming and the absence of evidence for an exceptional rain-on-snow event, the Salezer avalanche was simulated using an **effective liquid-water input of 20 mm**. To represent the increasing degree of wetting toward lower elevations, a water-content gradient of $G_W = 1.25\text{mm}/100\text{ m}$ was applied. This places Salezer within the moderate wet-snow parameter range and allows the lower sections of the avalanche path to become progressively wetter without imposing the much stronger liquid-water inputs associated with intense rain-on-snow conditions. Despite this comparatively strong gradient — which, per Table 5, would be expected to require higher rather than lower cohesion — the calibrated value ($N_0 = 550\text{ Pa}$) is lower than Grünbödeli's (800 Pa) at the same nominal water content. This likely reflects the event's greater geometric complexity: multiple release zones, secondary entrainment, and deposition that could not be fully quantified along the track (above) make Salezer a noisier calibration point than the more cleanly constrained Gatschiefer and Grünbödeli events.

Table 9: Salezer Tobel simulation parameters.

Case Study: Wet Snow Avalanche			
Case study name	Salezer 23.04.2008	Grid resolution (m)	4mx4m
Avalanche size / Return period	100L	Forest present (yes/no)	No
Starting volume (m ³)	67'000	End volume (m ³)	80'000
Snowcover Information			
Reference altitude (m a.s.l.)	2300	Snow height gradient Δh_Σ (m/100 m)	0.05
Snow temperature T_Σ (°C)	-1	Snow temp. gradient ΔT_Σ (°C/100 m)	0.1
Snow water content m_Σ (mm)	20	Water content gradient Δm_Σ (mm/100 m)	1.25
Erosion depth h_Σ or d_0^* (m)	0.90	Density method (3-day/variable/constant)	Variable
Erosion density ρ_Σ (kg/m ³)	300	Air temperature T_A (°C)	-1
Release density ρ_0 (kg/m ³)	300	Release depth d_0 (m)	0.90
Wet Snow Friction			
Generate α_ϕ (%)	5	Yield stress (Pa)	500

μ_w (wet sliding friction) (–)	0.15	m_t (transition thickness, mm)	100
Cohesion N_0 (Pa)	550	Stopping %	5
Comments			
The simulation used a Generate value of 5%, a snowcover yield stress of 500 Pa, a wet-snow sliding friction coefficient of $\mu_w=0.15$, and a transition thickness $m_t=100$ mm. Warm, approximately isothermal snow conditions were assumed. An effective liquid-water input of 20 mm was prescribed in the snowcover, corresponding to a wet-snow cohesion of approximately $N_0 = 550Pa$.			

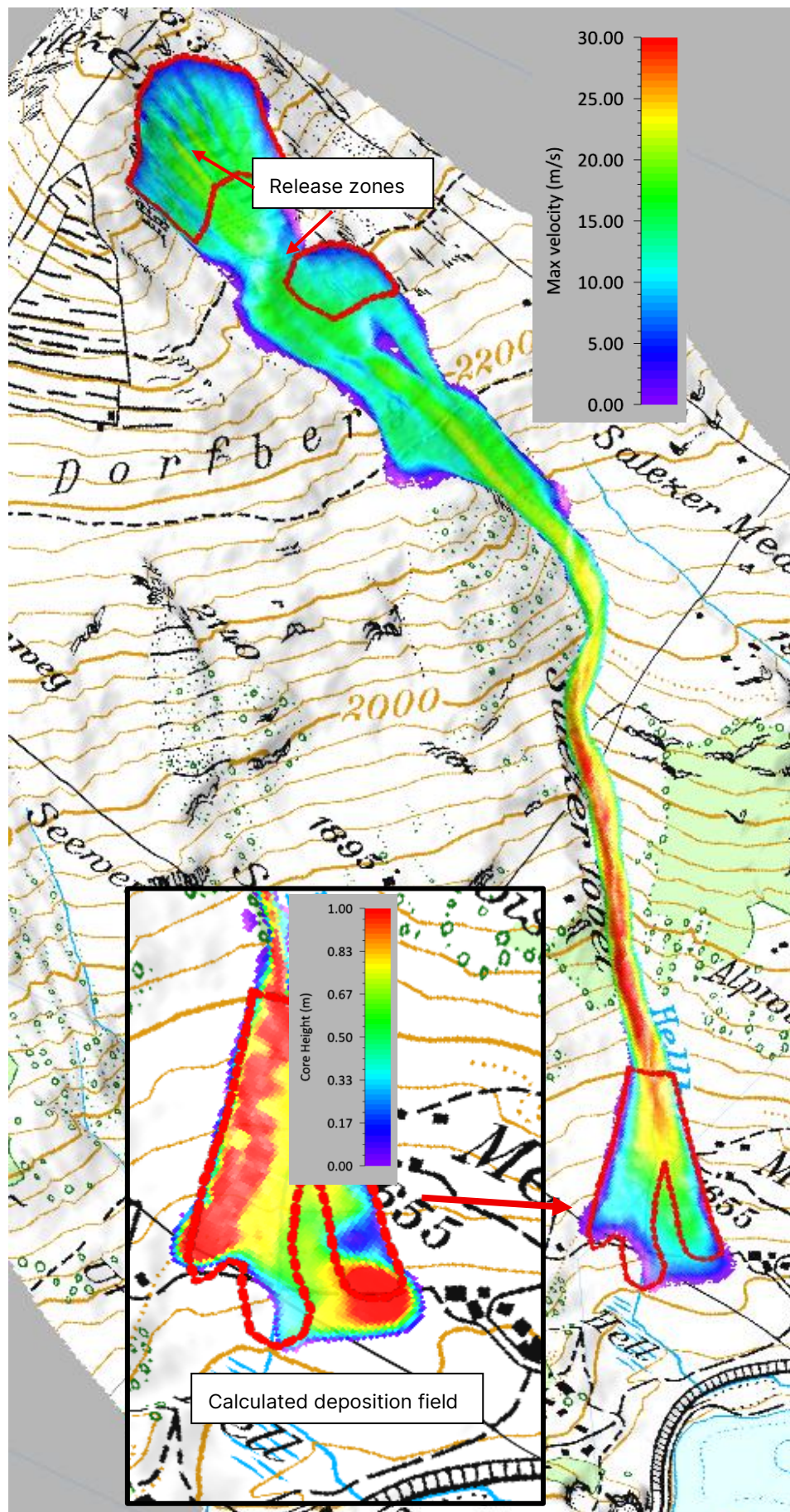


Figure 24: Simulation of the Salezer Tobel avalanche.

Rigopiano 18.01.2017

On 18 January 2017, after days of heavy snowfall and a seismic swarm, a massive avalanche swept down the eastern Gran Sasso, engulfing the Hotel Rigopiano in Farindola (Abruzzo, Italy). Twenty-nine guests and staff were killed; eleven people were rescued alive after a difficult, days-long search hindered by blocked roads and communication failures. The disaster highlighted vulnerabilities in hazard zoning and emergency coordination during compound events where weather and earthquakes interact.

The January 2017 Rigopiano avalanche is a clear example of a warm, initially dry flow that transitions to a wet regime. Snowpack measurements indicated a deep ($h_{\Sigma} \gtrsim 1.5\text{m}$), dense ($\rho_{\Sigma} = 200\text{-}250\text{kg/m}^3$), and warm snowcover ($T_0 \approx T_{\Sigma} = -2\text{C}$). A thick release slab ($\approx 76'700\text{ m}^3$) initiated high on the slope and funneled into a narrowing gully; for the simulation we prescribed $h_{\Sigma} = 1.50\text{m}$ with $\rho_{\Sigma} = 250\text{kg/m}^3$, consistent with observations.

As the confined corridor twisted, it opened onto a forested slope that was completely swept and entrained, adding mass and further warming the core. The avalanche reached $\approx 40\text{ m/s}$ before the gully and maintained this speed through the bends—consistent with weak curvature effects (curvature term disabled in the run). Grain melting began almost immediately, prior to the winding torrent, and continued downslope; the model predicts a $\sim 2\text{ mm}$ surface water film on grains in the deposition zone, explaining the long, lubricated runout and finger-like elongation documented in the field.

Table 10: Rigopiano simulation parameters.

Case Study: Dry-Wet Transitional Snow Avalanche			
Case study name	Rigopiano 18.01.2017	Grid resolution (m)	2mx2m / 5mx5m
Avalanche size / Return period	300L	Forest present (yes/no)	Yes
Starting volume (m^3)	77'150	End volume (m^3)	135'000
Snowcover Information			
Reference altitude (m a.s.l.)	1500	Snow height gradient Δh_{Σ} (m/100 m)	0.05
Snow temperature T_{Σ} ($^{\circ}\text{C}$)	-2	Snow temp. gradient ΔT_{Σ} ($^{\circ}\text{C}/100\text{ m}$)	0.1
Snow water content m_{Σ} (mm)	0	Water content gradient Δm_{Σ} (mm/100 m)	0

Erosion depth h_x or d_0^* (m)	1.50	Density method (3-day/variable/constant)	Variable
Erosion density ρ_x (kg/m ³)	250	Air temperature T_A (°C)	-2
Release density ρ_0 (kg/m ³)	250	Release depth d_0 (m)	2.0
Powder Parameters			
Generate α_ϕ (%)	8	Cloud drag (-)	4
Cliff factor (-)	2	Air entrainment ψ (m/s scale)	2
Turbulence decay β_H (s ⁻¹)	2.00		
Wet Snow Friction			
μ_w (wet sliding friction) (-)	0.00-0.05	m_t (transition thickness, mm)	2
Cohesion N_0 (Pa)	300	Stopping %	5
Comments			
Rigopiano starts as a dry avalanche (no water) at an elevation of 1500m, in warm temperatures $T_x = T_A = -2^\circ\text{C}$. No water in release zone, generate $\alpha_\phi = 8\%$. Release zone depth extreme (2m, $\rho_0 = 250 \text{ kg/m}^3$, consistent with warm snow), therefore 300L parameters for flow friction. Erosion parameters given by snow profile measurements in the runout zone $h_x = 1.5\text{m}/\rho_x = 250 \text{ kg/m}^3$, dry. After approximately 30s avalanche begins to generate meltwater. Meltwater water lubrication is responsible for long, sliding runout in deposition zone $m_t = 2\text{mm}$. Wet sliding friction set to $\mu_w = 0.00 - 0.05$ to model sliding friction on a compacted snowcover.			

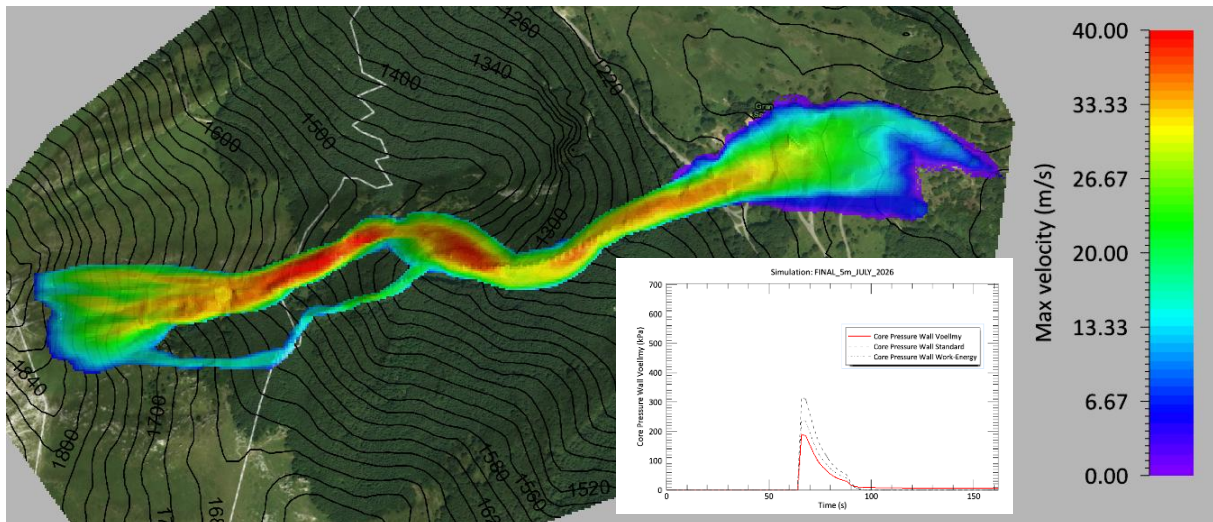


Figure 25: Calculated max velocity of the Rigopiano avalanche. The avalanche reached a maximum velocity of 42m/s. The avalanche struck the hotel with a speed of 28 m/s and exerted pressures of 320kPa (Wall), 190kPa (Voellmy) and/or 245kPa (Work-Energy). The long extend finger in the runout zone is due to lubricated motion at the end of the event.

At the hotel site, peak impact pressures were estimated between 190 and 320 kPa, depending on method: 320 kPa from the **standard wall formula** ($C_d = 2$), 190 kPa using the **Voellmy approach**, and 245 kPa from a **work-energy balance**. Simulations with colder-snow settings could not reproduce the observed runout and distinctive, fingered deposition; only warm-snow entrainment matched the event, emphasizing temperature control on flow behavior. To reflect the absence of meltwater in the release area we set Generate to $\alpha = 8\%$. A further distinctive requirement was an exceptionally low wet-friction value ($\mu_w = 0.00\text{--}0.05$) — below both ranges given in Section 8 — together with a small melt-transition thickness ($m_t = 2$ mm), indicating that even very thin water films can trigger the dry→wet regime shift under sufficiently warm conditions. Cohesion was set to $N_0 = 300\text{Pa}$, just above Section 8's dry-avalanche ceiling (0–200 Pa) and near the bottom of the wet-snow range in Table 5 (200–400 Pa at 0–10mm) — consistent with a warm release ($T_\Sigma = -2^\circ\text{C}$) that, unlike the colder VdIS event, begins generating meltwater almost immediately rather than staying dry through most of the flow.

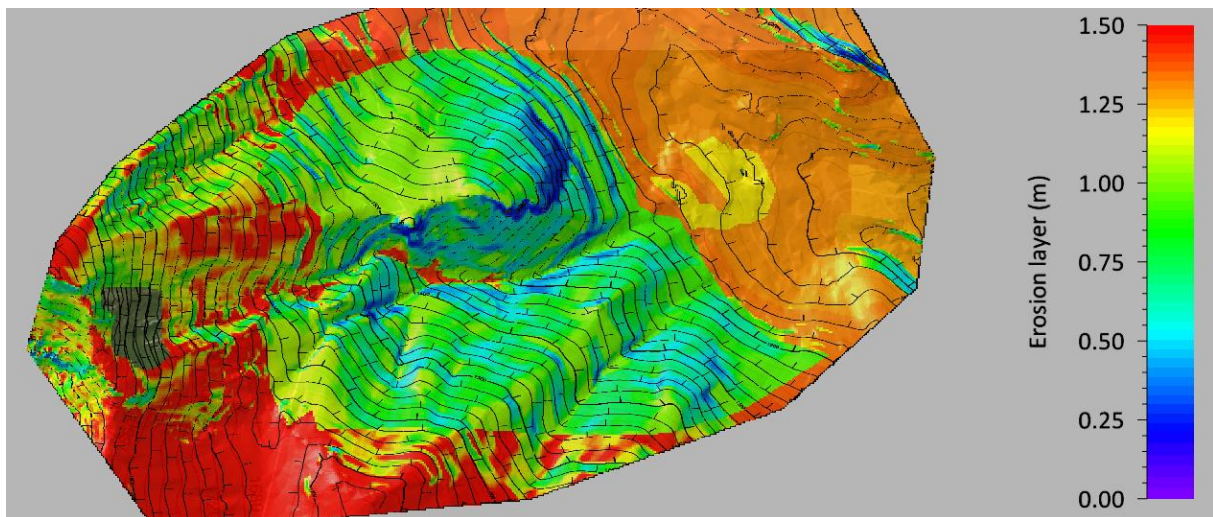


Figure 26: A deep snowcover was specified with the variable option. Forest interception was used in the forested areas surrounding the torrent.

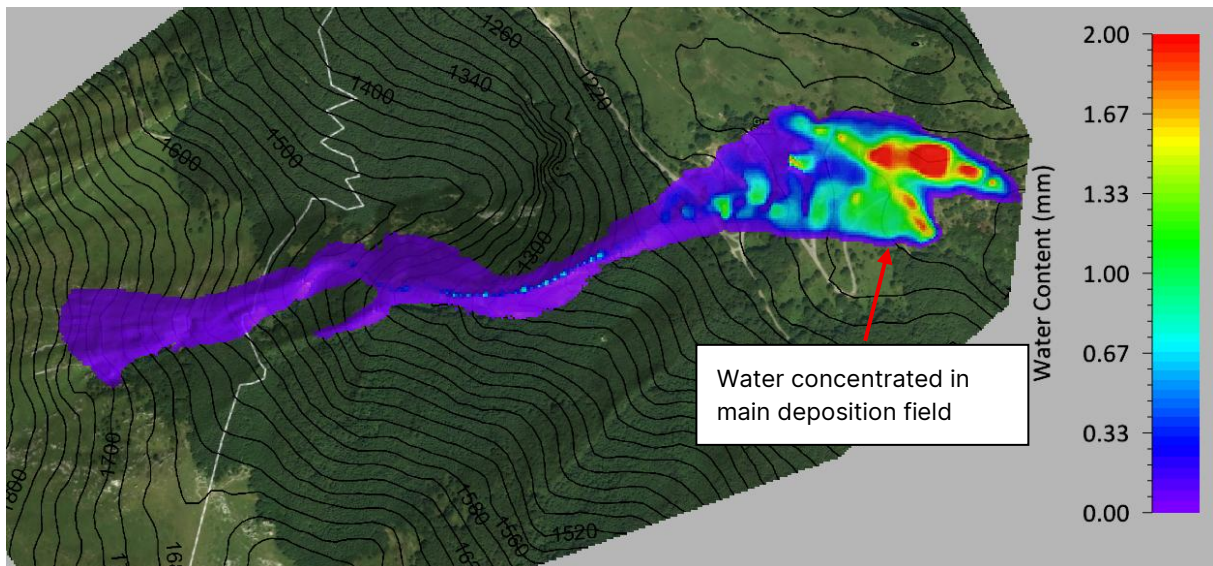


Figure 27: Calculated meltwater in the avalanche deposits. The surfaces of the granules were covered with a fine thin surface of meltwater.

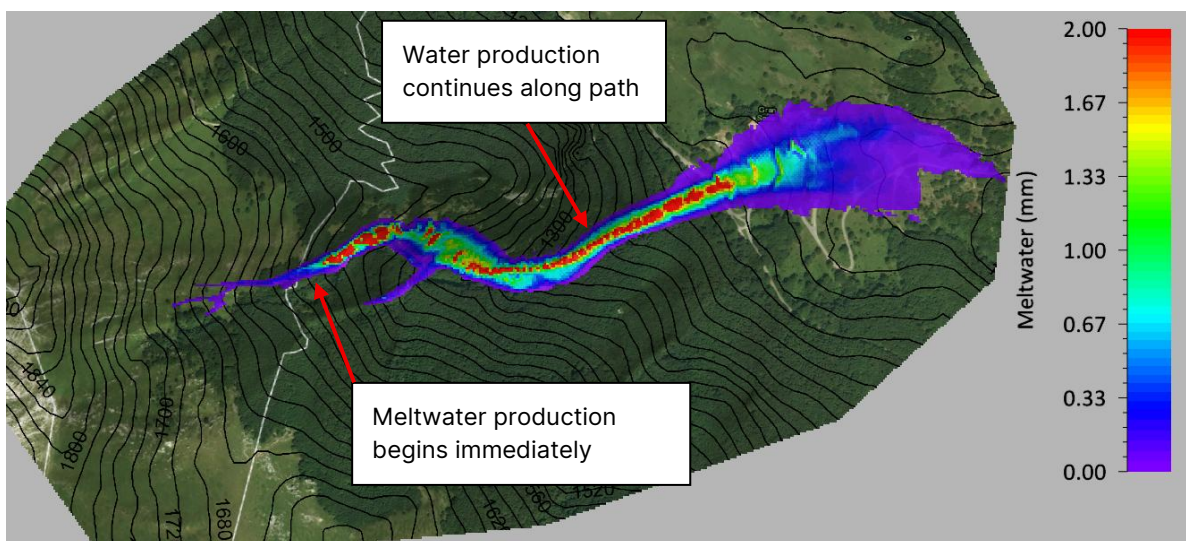


Figure 28: Calculated meltwater production. Due to the warm temperatures of the snow, meltwater was produced almost immediately after release.

Summary: Large warm ($T_0 \approx T_\Sigma = -2^\circ\text{C}$) avalanches running a moist, compacted snowcover can have low values of wet friction $\mu_w \approx 0.00\text{--}0.05$ and small transition values $m_t \approx 2\text{mm}$. The low μ_w appears to be determined by the warm snow in the runout zone.

Grüenbödeli 19.12.2023

During December 2023, several small avalanches occurred at Grüenbödeli. The largest, with an estimated volume of roughly **2,500 m³**, descended all the way to the valley floor. A drone survey successfully mapped the approximate release area as well as the runout and deposition zones (see Figures). Because of the small avalanche size (classification Tiny), we used two different grid resolutions to perform the simulations: 1m x 1m and 2m x 2m.

For modeling purposes, this avalanche was treated as a warm ($T_0 \approx T_\Sigma \approx 0^\circ\text{C}$) ground avalanche, releasing from a moist snowpack with an estimated liquid water content of about 20mm. Release and entrainment densities were taken to be 300kg/m³. We used a **variable** density distribution method (no three-day snow); **curvature ON**, due to the compact flow core. The depth of snowcover we took as 40cm-50cm with a standard elevation gradient of 3cm/100m. The temperature gradient was weak: 0.2C/100m. The liquid water gradient was set to 2.5mm/100m, providing about 25mm of liquid-water in the runout zone.

Because the avalanche began with wet snow we set generate to $\alpha_\phi = 5\%$. We modelled the avalanche using a low wet friction $\mu_w = 0.15$ and a transition parameter roughly on the scale of a single granule diameter $m_t \approx \mathbf{100\text{mm}}$. We considered the avalanche to be a 100T (100 year, tiny). This input reproduced the observed runout, flow width and deposition pattern. The avalanche was modelled without the powder cloud.

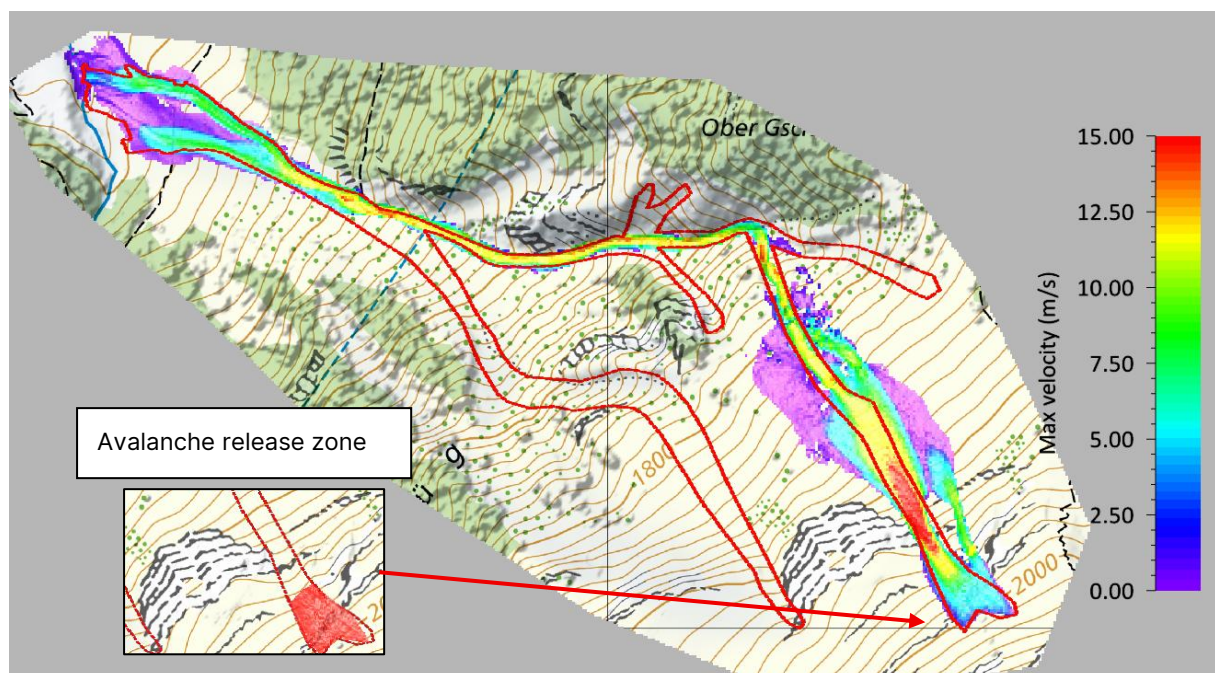


Figure 29: Calculated maximum flow velocities. The avalanche reached a maximum velocity of 17m/s. Release depth 1.0m; erosion depth 0.4m. Deposition heights at the river are approximately 1m.

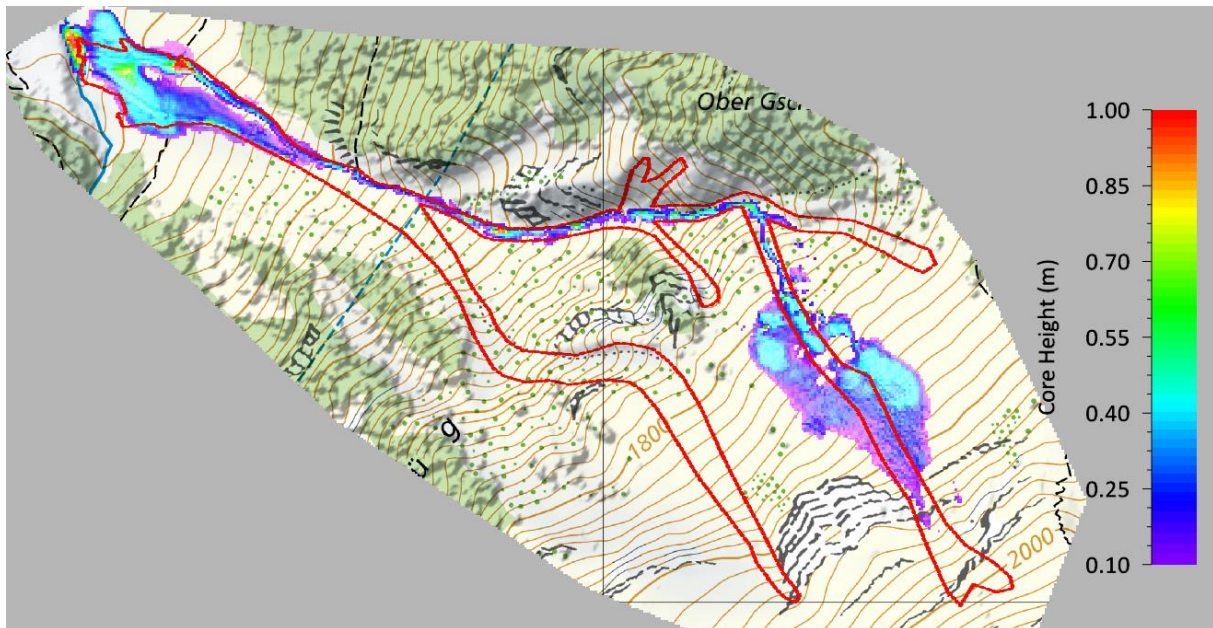


Figure 30: Calculated depositions. Maximum depositions reached 1m.

Table 11: Small wet avalanches running on the ground require higher friction values $\mu_w = 0.15$ and higher transition values $m_t \approx 100\text{mm}$. The transition values are on the scale with the granule size. Higher wet snow densities are required to simulate the event.

Case Study: Small, wet ground avalanche			
Case study name	Grüenbödeli	Grid resolution (m)	1mx1m / 2mx2m
Avalanche size / Return period	100T	Forest present (yes/no)	No
Starting volume (m ³)	2'525	End volume (m ³)	4'600
Snowcover Information			
Reference altitude (m a.s.l.)	2000	Snow height gradient Δh_{Σ} (m/100 m)	0.03
Snow temperature T_{Σ} (°C)	0	Snow temp. gradient ΔT_{Σ} (°C/100 m)	0.2
Snow water content m_{Σ} (mm)	20	Water content gradient Δm_{Σ} (mm/100 m)	2.5
Erosion depth h_{Σ} or d_0^* (m)	0.40	Density method (3-day/variable/constant)	Variable
Erosion density ρ_{Σ} (kg/m ³)	300	Air temperature T_{Λ} (°C)	0
Release density ρ_0 (kg/m ³)	300	Release depth d_0 (m)	1.0

Wet Snow Friction			
Generate α_ϕ (%)	5	Yield stress (Pa)	500
μ_w (wet sliding friction) (-)	0.15	m_t (transition thickness, mm)	100
Cohesion N_0 (Pa)	800	Stopping %	5
Comments			
This is the same avalanche path as the 2008 Grünbödéli event (above). The calibrated cohesion, $N_0 = 800\text{Pa}$ at $m_\Sigma \approx 20\text{-}25\text{ mm}$, matches the 2008 event at comparable water content — a same-site repeat measurement that supports the wet-snow parameterization independent of the cross-site scatter seen between Gatschiefer, Grünbödéli, and Salezer.			

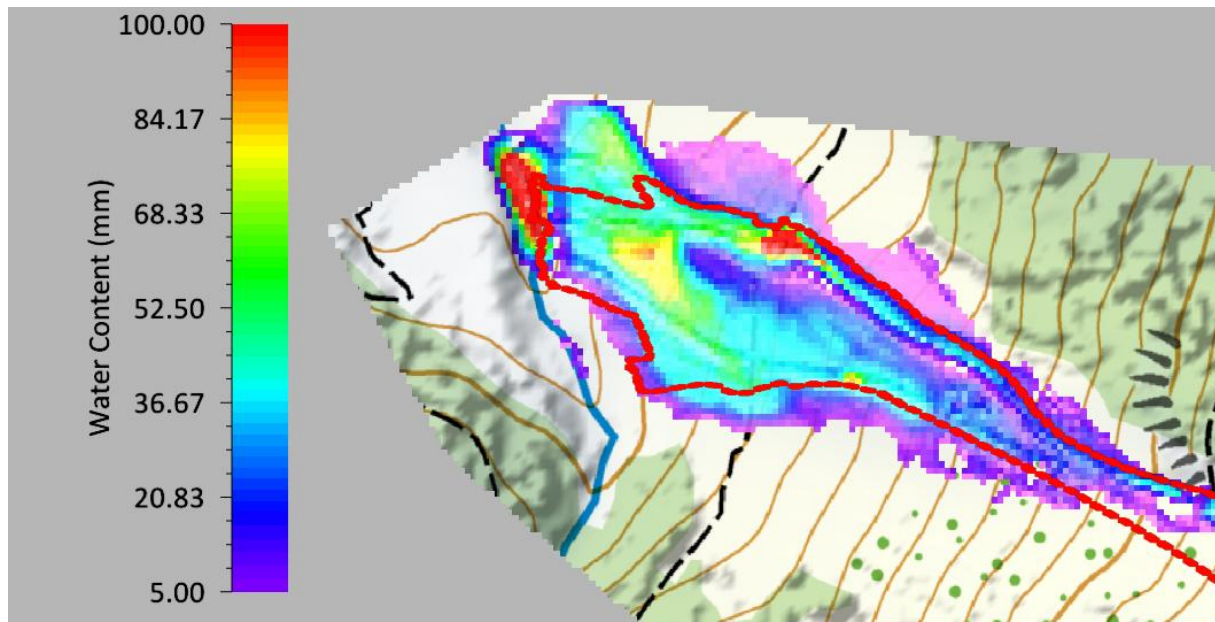


Figure 31: Calculated liquid-water content in the depositions. Maximum liquid-water content reached 200mm; in the 1m depositions liquid-water was only 120mm, about 12% by volume.

12. How to Simulate a Wet-Snow Avalanche

For wet-snow avalanche simulations, the model setup should reflect the reduced tendency for powder-cloud formation, the increased cohesion associated with wet snow, and the progressive transition toward a more plug-like flow regime.

1. **Disable powder-cloud formation.**

Select the **No Powder Cloud** option, since wet-snow avalanches are dominated by dense, cohesive flow rather than fluidized powder-cloud dynamics.

2. **Define the effective liquid-water content.**

Select the liquid-water input according to the expected degree of snowmelt or rain-on-snow forcing. Typical values for wet-snow avalanches are approximately **20–40 mm**. Substantially higher values should generally be treated as sensitivity cases, because a snowpack may become unstable and release before very large amounts of liquid water can accumulate.

3. **Select the friction category and avalanche size.**

Use the 100-year friction category, and then select the appropriate avalanche-size class:

T – tiny

S – small

M – medium

L – large

4. **Select cohesion N_0 according to liquid-water content and its elevation gradient.**

Cohesion should increase with increasing effective liquid-water content, reflecting stronger capillary bonding and the formation of larger snow agglomerates. As an approximate reference, calibrated case studies at ~20mm effective liquid-water content span $N_0 \approx 550\text{--}1000$ Pa (Section 11), with $N_0 \approx 800$ Pa a reasonable central estimate. Higher values, up to $N_0 \approx 1000$ Pa, have been used with modest downslope gradients (Gatschiefer, $\Delta m_\Sigma = 1\text{mm}/100\text{m}$); however, this is not a strict function of gradient alone — Salezer required a *lower* $N_0 \approx 550$ Pa despite a larger gradient (1.25mm/100m), reflecting event-specific factors beyond water content and gradient (Section 8, Table 5). Treat N_0 as a calibration starting point, not a precise function of G_W .

5. **Set the wet-snow transition parameters.**

Use:

Generate = 5%

Wet-transition thickness $m_t = 100\text{mm}$

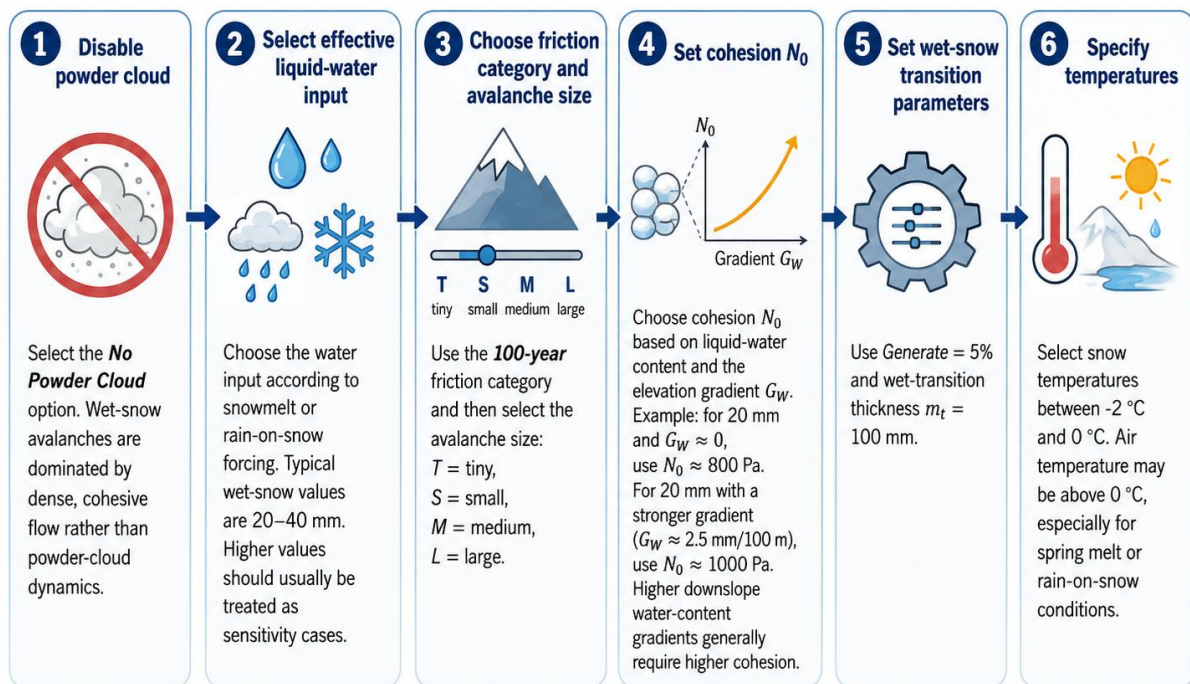
These values provide a consistent parameterization for the transition from more granular motion toward increasingly cohesive wet-snow flow.

6. **Specify snow and air temperature.**

Select snow temperatures close to the melting point, typically between approximately **-2 °C and 0 °C**. The air temperature may approach or slightly exceed **0 °C**.

How to Simulate a Wet-Snow Avalanche

Recommended setup for RAMMS::Extended wet-snow calculations



Cohesion increases with increasing effective liquid-water input and with stronger downslope wetting.

Figure 32: How to simulate a wet snow avalanche with RAMMS::Extended.

13. How to Simulate a Transitional Dry-to-Wet Snow Avalanche

Transitional dry-to-wet avalanches typically release under relatively cold, dry conditions but progressively warm during descent. In the upper part of the path, the flow may still behave like a dry avalanche and develop a powder cloud, while increasing snow temperature and meltwater formation promote a transition toward denser, more lubricated flow at lower elevations.

1. **Enable powder-cloud formation.**

The release zone and upper track are typically dry enough for powder-cloud development. Define the initial snow temperature in the release area together with an appropriate temperature gradient. Typical release-zone temperatures are in the range of approximately **-4 °C to -2 °C**. Because fracture depths are often large, the mean temperature of the released snow can be relatively high, particularly where deeper snow layers have been warmed by geothermal heat flux.

2. **Define the snow-temperature gradient.**

The initial snowpack generally contains little or no liquid water. The transition is therefore driven primarily by warming rather than by pre-existing liquid-water content. Temperature gradients can be pronounced and are often greater than approximately **0.5 °C/100 m**, allowing the avalanche snow to approach the melting point as it descends.

3. **Specify the air temperature.**

Select an air temperature representative of the event. For dry-to-wet transition avalanches, the ambient air can be warmer than the snow initially released in the starting zone, reinforcing the warming of the flowing mass during descent.

4. **Select the standard friction category and avalanche size.**

Use the same procedure as for a dry powder avalanche. Select the appropriate **return period, typically between 10 and 300 years**, and then choose the avalanche-size class: Tiny, Small, Medium, or Large. Because transitional avalanches span a wide range of thermal conditions, the wet snow parameters μ_w and N_0 vary more than for fully wet avalanches (Section 12). Colder events, where the transition is driven mainly by frictional melting under sub-zero air (e.g., $T_A \approx -6^\circ\text{C}$ to -7°C , VdIS), require $\mu_w \approx 0.15$ together with cohesion drawn from the dry-avalanche category ($N_0 = 0\text{--}200$ Pa, Section 8). Warmer events, where the release is close to the melting point and meltwater forms almost immediately (e.g., $T_A \approx -2^\circ\text{C}$, Rigopiano), require a much lower $\mu_w \approx 0.00\text{--}0.05$, with cohesion near the bottom of the wet-snow range ($N_0 \approx 300\text{Pa}$).

5. **Set the random-energy generation parameter.**

Deep fractures and relatively warm snow can result in mean release densities exceeding approximately **200 kg/m³**. To account for this comparatively dense release material, set **Generate = 8%**.

6. Set the wet-transition thickness.

As the snow approaches the melting point, thin meltwater films form on the surfaces of snow granules and begin to lubricate shearing processes. To represent the onset of this dry-to-wet transition, use a **wet-transition thickness of $m_t = 2\text{mm}$** .

How to Simulate a Transitional Dry-Wet Snow Avalanche

Recommended setup for RAMMS::Extended dry-to-wet transition calculations

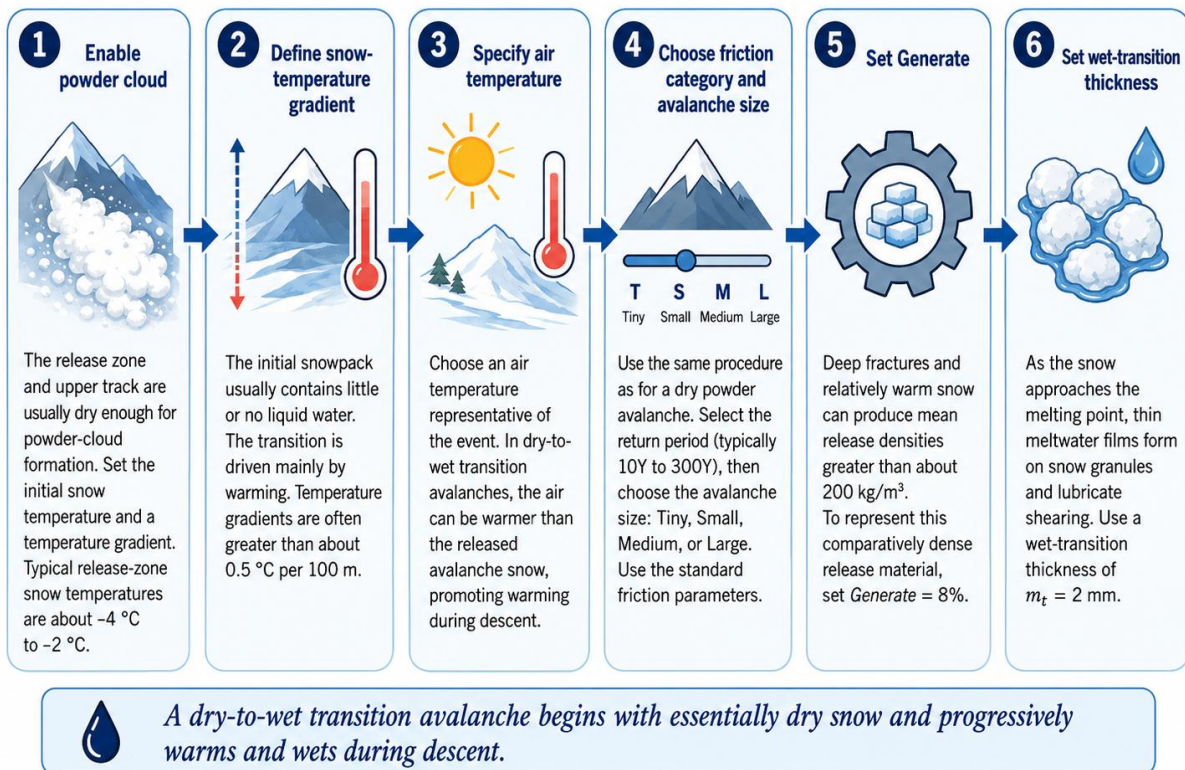


Figure 33: How to simulate a dry-wet snow transitional avalanche with RAMMS::Extended.

14. Wet and Transitional Snow Avalanches: Competing Processes and Emergent Flow Behavior

The **principal conclusion of the wet-snow case studies** is that the behavior of wet and transitional snow avalanches cannot be attributed to a single friction parameter, nor even to a single physical mechanism. Avalanche motion **emerges from the interaction of several processes acting simultaneously: some promote mobility, while others dissipate energy, increase internal resistance, or suppress particle dispersion**. The observed flow regime is therefore an emergent behavior, determined by the evolving balance between these competing processes.

Two principal routes to avalanche mobility can be distinguished. In cold, dry avalanches, shear work generates random kinetic energy, dispersing the snow granules and reducing effective friction through fluidization. If sufficient random

energy is produced and sustained, the avalanche becomes highly mobile and may develop a powder cloud. In wet avalanches, this route becomes progressively less effective. Liquid water makes grain interactions increasingly dissipative, reduces the production of fluctuation energy, accelerates its decay, and promotes grain agglomeration and cohesion. At the same time, however, water lubricates grain contacts and the basal sliding surface, providing a second and fundamentally different route to mobility. A wet avalanche may therefore travel a considerable distance despite remaining dense, relatively slow, and only weakly fluidized.

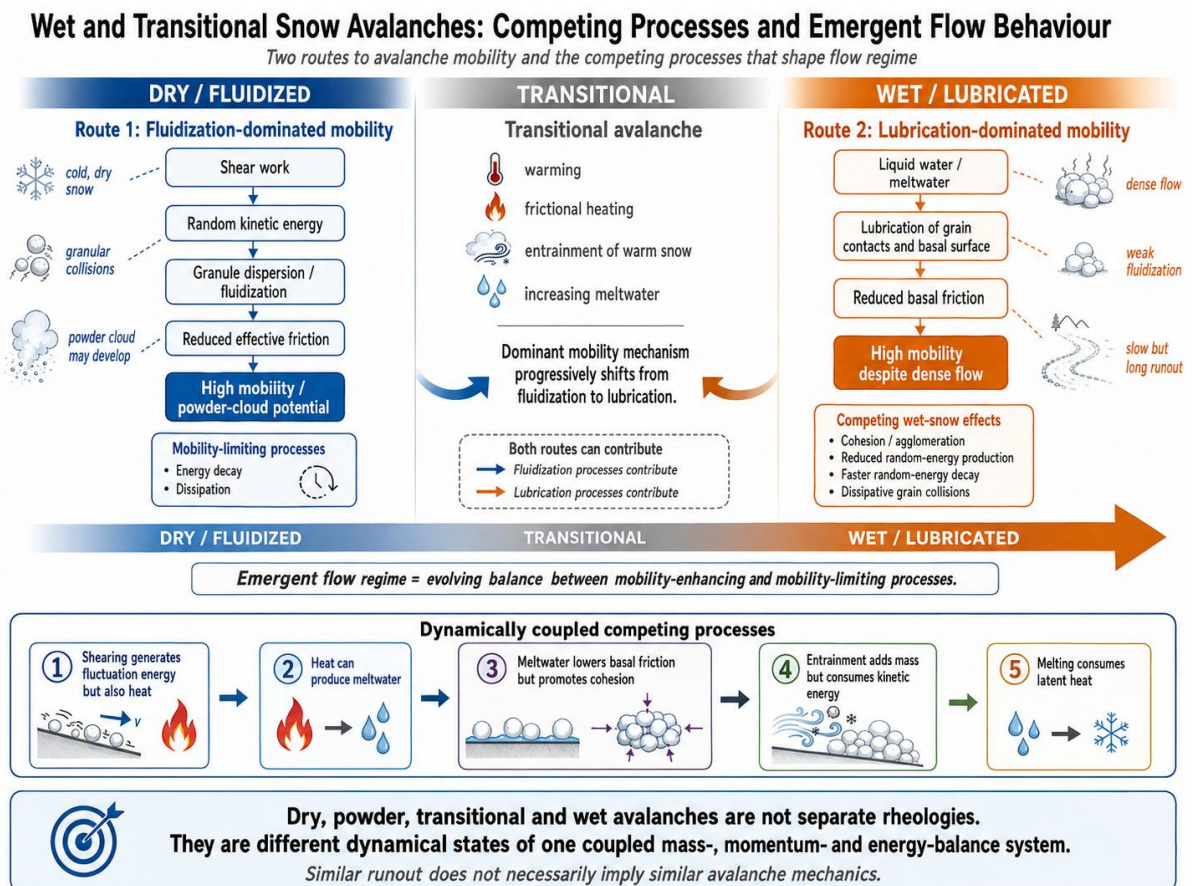


Figure 34: Conceptual framework for the emergent flow behaviour of wet and transitional snow avalanches. Two principal routes to mobility are distinguished: fluidization in cold, dry snow and lubrication in wet snow. Transitional avalanches evolve between these pathways as warming and meltwater progressively shift the dominant mobility mechanism. The resulting flow regime emerges from the balance between mobility-enhancing and mobility-limiting processes.

These processes should not be regarded as independent corrections to a friction law. They are dynamically coupled. Shearing generates fluctuation energy, but it also generates heat. That heat can produce meltwater, which reduces basal friction but simultaneously suppresses further fluidization and promotes cohesion. Entrainment adds mass and therefore increases the gravitational-energy inventory of the avalanche, yet accelerating newly entrained snow consumes kinetic energy and, in warm conditions, can strongly increase dissipation. Melting itself consumes latent heat. **Thus, processes that enhance avalanche mobility are continually**

accompanied by processes that constrain it. Avalanche behaviour emerges from this competition rather than from any individual mechanism acting alone.

This competition is particularly important in transitional avalanches. A flow may initially obtain much of its mobility through fluidization and later obtain an increasing fraction through lubrication as the avalanche warms, entrains warm snow, or generates meltwater through frictional heating. At the same time, increasing wetness suppresses fluctuation energy and promotes agglomeration and cohesion. The transition from dry to wet behavior is therefore not simply a change from high friction to low friction. It is a progressive reorganization of the mechanisms controlling mobility, in which one route to mobility weakens while another becomes increasingly important.

From this perspective, dry, powder and wet avalanche behaviour should not be regarded as fundamentally separate phenomena requiring independent rheological descriptions. They represent different dynamical states of the same coupled mass-, momentum- and energy-balance system. Different initial conditions and parameter values shift the relative importance of fluidization, lubrication, cohesion, entrainment, heating and dissipation. As these competing processes evolve in space and time, the avalanche can move continuously between dry, powder, transitional and wet flow regimes.

An important consequence is that similar runout distances do not necessarily imply similar avalanche mechanics. A dry avalanche may achieve long runout because strong fluctuation-energy production maintains a fluidized state, whereas a wet avalanche may achieve comparable runout because basal lubrication compensates for the almost complete loss of fluidization. Transitional avalanches may exploit both mechanisms at different stages of their motion. Observable mobility can therefore be similar even when the internal pathways by which that mobility is generated are fundamentally different.

The objective of RAMMS::Extended is consequently not to prescribe whether an avalanche is “dry,” “powder,” or “wet” in advance. Rather, it is to represent the principal competing processes sufficiently consistently that the resulting flow behaviour — including transitions between regimes — can emerge from the calculation itself. The wet and transitional case studies presented in this document suggest that this competition between mobility-enhancing and mobility-limiting processes provides a useful framework for understanding not only the dry-to-wet transition, but avalanche dynamics more generally.

15. References

Kojima, K. (1967). Densification of seasonal snow cover. In H. Oura (Ed.), *Physics of Snow and Ice (Vol. 1, Part 2, pp. 929–952)*. Institute of Low Temperature Science, Hokkaido University.

Jamieson, J. B., and Johnston, C. D. (1990). In-situ tensile tests of snow-pack layers. *Journal of Glaciology*, **36**(122), 102–106.

McClung, D. M. (2009). Dimensions of dry snow slab avalanches from field measurements. *Journal of Geophysical Research*, **114**, F01006.

Stoffel, L., Bartelt, P., & Margreth, S. (2024). Scenario-based avalanche simulations including snow entrainment, snow temperature and density. In K. Gislås, P. Gauer, H. Dahle, M. Eckerstorfer, A. Mannberg, & K. Müller (Eds.), *Proceedings of the International Snow Science Workshop 2024* (pp. 374–380). Norwegian Geotechnical Institute, Oslo.

Bartelt, P., & Christen, M. (2026). *RAMMS::Extended user manual*. RAMMS AG, Davos Wiesen, Switzerland.

Bühler, Y., Hüni, A., Christen, M., Meister, R., & Kellenberger, T. (2009). Automated detection and mapping of avalanche deposits using airborne optical remote sensing data. *Cold Regions Science and Technology*, **57**, 99–106.
<https://doi.org/10.1016/j.coldregions.2009.02.007>.

World Meteorological Organization. (2021). *Guide to Instruments and Methods of Observation* (WMO-No. 8), Volume I – Measurement of Meteorological Variables. Geneva, Switzerland: WMO.

U.S. Department of Agriculture, Natural Resources Conservation Service. (2004). Snowmelt. *National Engineering Handbook*, Part 630, Chapter 11. Washington, DC: USDA-NRCS.

Ranz, W.E., & Marshall, W.R. (1952). Evaporation from drops, Part I. *Chemical Engineering Progress*, **48**(3), 141–146.

Bartelt, P., Vera Valero, C., Feistl, T., Christen, M., Bühler, Y., & Buser, O. (2015). Modelling cohesion in snow avalanche flow. *Journal of Glaciology*, 61(229), 837–850. <https://doi.org/10.3189/2015JoG14J126>.

Buser, O., & Bartelt, P. (2009). Production and decay of random kinetic energy in granular snow avalanches. *Journal of Glaciology*, 55(189), 3–12. <https://doi.org/10.3189/002214309788608859>.

Platzer, K., Bartelt, P., & Kern, M. (2007). Measurements of dense snow avalanche basal shear to normal stress ratios (S/N). *Geophysical Research Letters*, 34(7), L07501. <https://doi.org/10.1029/2006GL028670>.

Bartelt, P., Buser, O., Vera Valero, C., & Bühler, Y. (2015). Configurational energy and the formation of mixed flowing/powder snow and ice avalanches. *Annals of Glaciology*, 57(71), 179–188. <https://doi.org/10.3189/2016AoG71A464>.

Salm, B., Burkard, A., & Gubler, H.U. (1990). *Berechnung von Fliesslawinen: Eine Anleitung für Praktiker mit Beispielen*. Mitteilungen des Eidgenössischen Instituts für Schnee- und Lawinenforschung, Nr. 47. Davos, Switzerland: SLF.

Haff, P.K. (1983). Grain flow as a fluid-mechanical phenomenon. *Journal of Fluid Mechanics*, 134, 401–430. <https://doi.org/10.1017/S0022112083003419>.

Bartelt, P., & McArdeil, B.W. (2009). Granulometric investigations of snow avalanches. *Journal of Glaciology*, 55(193), 829–833. <https://doi.org/10.3189/002214309790152384>.

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