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Rock mass instabilities in periglacial environments induced by climate change

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Abstract

This study examines how permafrost degradation destabilizes high-mountain rock walls through the coupling of thermodynamics and fracture mechanics. The water-ice phase transition generates cryostatic pressure—governed by the Clausius-Clapeyron equation—which, alongside hydrostatic pressure, increases the stress intensity factor (K_I) at crack tips. As temperatures rise, the rheological evolution of ice leads to a loss of its "cryo-anchoring" capacity. Furthermore, repeated freeze-thaw cycles induce fatigue in rock bridges, which are further weakened by stress corrosion cracking via silicate hydrolysis; this degrades internal cohesion even during thaw periods. Using the Mont Blanc massif as a focal point, the case of Les Drus illustrates the transition from subcritical crack propagation to catastrophic failure. Conversely, the case of Aiguille du Midi demonstrates risk mitigation strategies, specifically the use of pre-stressed anchors to mechanically reduce K_I and maintain the structural integrity of the rock mass.

Keywords

Permafrost degradation, Fracture mechanics, Cryostatic pressure, Rock bridge fatigue, Mont Blanc.

1 Introduction

Climate change, manifested by rising air temperatures in high-mountain periglacial environments, is driving the degradation of mountain permafrost, resulting in an increasing frequency of large-scale rock mass instabilities. This phenomenon is particularly evident in the French Alps, where high granitic peaks are frequently affected by major rockfall and rock avalanche events during heatwave summers (Gruber and Haeberli 2007). Although these processes rarely cause human casualties, their economic and infrastructural consequences can be significant, especially in areas with dense tourism and transportation facilities.

In Alpine massifs, such as the Mont-Blanc massif, permafrost affects lithologies with negligible matrix porosity (less than 1%), primarily granites and quartzites. These formations are structured by multi-scale discontinuity networks, ranging from micro-fractures to joints. During the winter period,

under a stable cold-thermal regime, ice occupying these interstices performs a crucial geotechnical function, acting as a cohesive agent that ensures the structural stability of the rock mass.

Seasonal warming (spring and summer) initiates a critical rheological transition. Although the ice does not thaw instantaneously, its warming toward 0°C reduces its viscosity and triggers accelerated creep within the fractures. This deterioration is amplified by the infiltration of meltwater, which lubricates rock joints. Furthermore, water trapped within these fissures generates hydrostatic pressure, exerting a mechanical thrust that further destabilizes the rock walls. During the subsequent winter cycle, the transition of pore water to ice induces volumetric expansion. Within saturated, partially closed fissure networks constrained by rock bridges, this phase change generates significant cryostatic pressure. The resulting stresses are concentrated at fracture tips; once these exceed the tensile strength of the rock, they induce fracture propagation (Walder and Hallet1985). Repeated annual freeze-thaw cycles lead to cumulative mechanical fatigue, progressively increasing discontinuity extensions. Simultaneously, stress corrosion cracking (SCC), driven by silicate hydrolysis, weakens the rock bridges. Ultimately, this process leads to a critical loss of cohesion; the mechanical strength at the rock-ice interface drops abruptly, rendering the system unstable. A collapse then occurs due to a loss of equilibrium; no longer supported by ice adhesion, the block slides along a meltwater film or thermally softened ice.

2 Thermodynamic behavior of the Ice-Rock mixture

2.1 Water-Ice transition and cryostatic pressure

Water exhibits anomalous thermodynamic behavior during its transition to the solid phase, deviating from standard thermal expansion models. Upon solidification, its specific volume increases sharply as density drops. This density evolution, shown in Figure 1, occurs in four distinct phases:

a)Liquid Contraction Phase (+20°C to +4°C): As temperature decreases, reduced thermal agitation allows for closer molecular packing, resulting in a decrease in volume and an increase in density.

b) Dilatometric Anomaly (+4°C to 0°C): Approaching the freezing point, water begins a slight expansion. This anomalous behavior is attributed to the pre-organization of molecules into clusters that resemble the less dense hexagonal lattice of ice.

c) Phase Change (0°C): During the transition to the solid state, volume increases abruptly by approximately 9%. This rapid expansion generates the cryostatic pressure within rock fissures.

d) Solid Phase Contraction (T<0°C): Below the freezing point, ice behaves as a standard solid; its volume decreases proportionally with falling temperatures. As thermal energy dissipates, the vibration amplitude of the molecules within the lattice decreases, leading to lattice contraction.

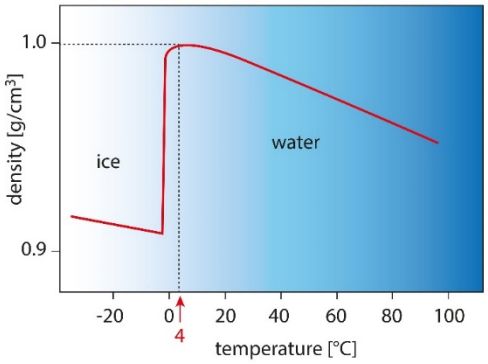


Fig. 1 Evolution of water density as a function of cooling.

Under isochoric (constant-volume) conditions, the pressure generated by the liquid-to-solid phase transition of water is quantified by the Clausius-Clapeyron relation. This equation governs the evolution of equilibrium pressure as a function of temperature:

$$\frac{dP}{dT} = \frac{L}{T(v2-v1)} \quad (1)$$

Where L Enthalpy of fusion of ice (≈333,5 kJ/kg)
T Absolute temperature in Kelvin (273.15 K at 0°C)
ν2,ν1 Specific volumes of ice and liquid water, respectively(m^3/kg)

Calculations with the specific values of water (Table 1) indicate that lowering the freezing point of water to −1°C requires an additional pressure of approximately 13.8 MPa; at −22°C, this pressure reaches a theoretical limit of approximately 210 MPa. Beyond this threshold, ice undergoes a polymorphic transition into denser crystalline structures, such as Ice II or Ice III.

Table 1. Specific volume values.

Specific volume	**Symbol**	**Value (10^{-3} m^3 / kg)**
Liquid water	v1	1,0001
Ice	v2	1,0907
Volume difference	Δv	0,0906

However, in situ observations reveal much lower values because the tensile strength of the host rock rarely exceeds 5 to 25 MPa. Consequently, mechanical failure occurs well before the theoretical Clausius-Clapeyron peak is reached. The resulting fracturing increases the available volume, causing an instantaneous pressure drop. Beyond simple volumetric expansion, this thermodynamic process induces the migration of liquid water toward the freezing front via cryogenic suction, promoting the growth of ice lenses through ice segregation.From these observations, it can be inferred that prolonged periods of intense cold result in negligible changes to ice volume; therefore, the peak pressure exerted on the rock is a transient phenomenon occurring primarily at the moment of phase transition.

2.2 Mechanical behavior of Ice

Ice is a crystalline solid with a viscoplastic response highly dependent on temperature and time (Schulsonand Duval 2009). Its stiffness decreases significantly as temperatures approach 0°C, while its deformation is highly strain-rate sensitive: rapid loading induces brittle failure, whereas slow, sustained stress promotes plastic creep. In this viscous regime, ice behaves as a high-viscosity fluid that relaxes internal stresses. This steady-state (secondary) creep is governed by Glen's flow law:

$$\dot{\varepsilon} = \dot{A}.\sigma^n \tag{2}$$

Where $\dot{\epsilon}$ Strain rate (creep rate)
σ Applied stress
n Glen's exponent (typically taken as 3)
A Fluidity parameter (viscosity)

As a non-Newtonian fluid, ice does not possess a constant viscosity; instead, its flow is temperature-dependent. The fluidity parameter A (the reciprocal of viscosity) is determined by the Arrhenius law:

$$A(T) = Ao\,.\exp\,(-\frac{Q}{R\,.T}) \tag{3}$$

Where Q Activation energy (60 kJ/mol for T > -10 °C and 139 kJ/mol for T < -10 °C)
R Universal gas constant (8.314 J/mol·K)
T Temperature (K)

It is observed that as ice warms, A increases, reflecting a decrease in viscosity. Conversely, as temperatures drop, viscosity increases because the thermal energy available for molecular dislocation and movement diminishes. Typical values of the effective viscosity are provided in Table 2.

Table 2. Effective viscosity of ice at different temperatures (liquid water viscosity at 20°C is 10^{-3}Pa·s).

Temperature (°C)	**Effective Viscosity (Pa·s)**	**State of the Ice**
0	10^{13} à 10^{14}	Flows easily (temperate glaciers)
-10	5 x 10^{14}	Cold ice, slowed flow
-25	10^{15} à 10^{16}	Very rigid
-50	> 10^{17}	Behaves as a rigid solid

The mechanical strength of ice is intrinsically linked to its crystalline structure. Between 0°C and −40°C, its compressive strength can increase fourfold. However, this strength is significantly attenuated by impurities and porosity, particularly trapped air bubbles. Furthermore, cooling induces

thermal contraction, which generates internal tension cracks (thermal contraction cracking) that further compromise structural integrity (Ngo and Pellet 2018). Mechanical properties of ice are summarized in Tables 3, for strength and elastic properties (Cuffey and Paterson, 2010) and in Table 4, for shear strength (Davies et al., 2000).

Table 3. Mechanical properties of ice at -10°C.

Properties	Mean value (at −10°C)	Observation
Compressive Strength	5 to 25 MPa	At -10°C; increases as temperature decreases
Tensile Strength	0.7 to 3.1 MPa	Highly sensitive to existing micro-cracks independent of temperature.
Shear Strength	0.5 to 2 MPa	Depends on crystal orientation
Young Modulus	9.0 to 9.5 GPa	Elasticity of pure ice.
Poisson's ratio	0.33	Elastic behavior.

Table 4. Shear strength at different temperatures.

Temperature °C	Strength (MPa)	Interface State
-15	2.0 to 4.0	Very stable; brittle failure
-2	0.5 à 1.0	Onset of plasticity (creep)
0	approx. 0.1	Sliding promoted by a liquid water film

2.3 Ice-Rock interface behavior

The adfreeze strength of the ice-rock interface serves as the primary stabilizing "glue" for high-mountain rock walls. According to Davies and al. (2000), this bond is governed by three mechanisms: chemical adhesion (hydrogen bonding), mechanical interlocking (penetration into micro-pores and asperities), and friction (post-bond resistance).This shear strength is highly temperature-dependent and can be modeled using Jellinek's power law (Jellinek, 1959):

$$\tau = a\,(-T)^n \tag{4}$$

Where τ Adfreeze bond stress (MPa)
T Temperature in degrees Celsius (the negative sign ensures a positive value)
a, n Constants depending on the substrate material

For rough rock surfaces, $n \approx 1$, yielding a near-linear relationship between temperature and strength. Interface behavior is categorized into three thermal zones:

0°C to −1°C (Critical Zone): Adhesion is negligible due to the formation of a liquid water film. The friction coefficient drops abruptly, making this the highest-risk period for rockfall.

−1°C to −10°C (Sensitivity Zone): Strength is extremely sensitive; a 2°C drop can double the adfreeze bond as the ice lattice mechanically anchors into the rock.

Below −10°C (Stability Zone): The interface bond often exceeds the intrinsic shear strength of the ice itself.

Beyond temperature, shear strength is dictated by surface roughness (e.g., granitic surfaces exhibit $\tau > 2$ MPa vs. 0.5 MPa for polished metal) and strain rate, where rapid loading yields higher apparent strength compared to the stress-relaxation of slow creep.

Table 5. Orders of magnitude of adfreeze strength at different temperatures.

Temperature (°C)	Strength (MPa)	Interface State
0	0,05 à 0,1	Presence of a liquid film (near zero)
-2	0,2 à 0,5	Plastic transition: ice still "slides"
-5	0,8 à 1,2	Strong adhesion: onset of "cryo-anchoring"
-10	1.5 à 2.5	Rigid ice: failure occurs within the ice itself

3 Rock behavior and cyclic fatigue

In rock masses, many discontinuities—such as fissures or fractures—are non-persistent, meaning they are interrupted by intact segments known as rock bridges. The crystallization pressure exerted by ice on fissure walls induces a redistribution of stresses at the crack tip, which can be quantified using

Linear Elastic Fracture Mechanics (Torsello 2025).The stress intensity factor (K) defines the magnitude of tensile stress at the crack tip. If this value exceeds the material's fracture toughness (K_{IC}), the fissure propagates(Atkinson 1987). According to the Griffith criterion, failure occurs when the energy release rate reaches a critical threshold.

3.1 Mechanical fatigue and subcritical crack growth

Cyclic freeze-thaw action induces mechanical fatigue within rock bridges through subcritical crack growth, allowing fractures to propagate even when stresses remain below the rock's critical threshold ($K_I < K_{Ic}$). This process begins with microscopic damage accumulation in the Fracture Process Zone (Gatelier et al. 2002), where hysteresis prevents micro-cracks from fully closing upon thawing, leading to irreversible structural weakening. According to Paris' Law (Paris and Erdogan 1963), this cyclic propagation is highly sensitive to stress amplitudes, creating a self-reinforcing feedback loop. As fractures extend, they increase the void volume available for meltwater infiltration, which generates higher cryostatic pressure in subsequent cycles. Ultimately, this cumulative fatigue reduces the effective cross-section of the rock bridges until the remaining structural capacity is exceeded, triggering a sudden, catastrophic collapse.

3.2 Stress corrosion cracking (SCC)

Stress Corrosion Cracking (SCC) reveals that rock failure is a chemo-mechanical process (Atkinson, 1987), explaining why fissures in granitic lithologies propagate solely through moisture exposure. In the presence of water, the strong silicon-oxygen (Si-O) bonds within silicate minerals undergo a hydrolysis reaction ($Si{-}O{-}Si + H2O \rightarrow Si{-}OH + HO{-}Si$), replacing primary atomic bonds with weaker hydrogen bonds. Mechanical stress facilitates this by dilating the crystal lattice at the crack tip, allowing water molecules to penetrate and sever bonds more efficiently. This subcritical propagation is governed by relative humidity and temperature, which accelerates the reaction rate. SCC acts in concert with frost shattering; while freezing provides mechanical dilation, chemical corrosion continues the degradation during thaw periods. Furthermore, the acidity of rainwater (pH $\approx$ 5.6) accelerates the transformation of feldspars into weak clay minerals like kaolinite. Consequently, crack propagation velocity increases exponentially as pH decreases, explaining why collapses often occur in mid-summer without immediate mechanical triggers.

4 Case studies in the Mont Blanc massif

4.1 West face of Les Drus (3,700 m)

The collapse of the West Face of Les Drus is undoubtedly the most iconic example of rock massif failure due to permafrost degradation. In late June 2005, the famous Bonatti Pillar detached from the mountain, involving a volume of approximately 265,000 m³ of granite. This event left a 700-meter-high scar and resulted in the disappearance of several historic climbing routes (Gruber and Haeberli 2007).

In this high-altitude rock wall, freeze-thaw cycles generated cryostatic pressure that increased the stress intensity factor, leading to the fatigue of rock bridges through subcritical crack propagation. Furthermore, rising temperatures melted the permanent ice that filled the fissures and acted as a binder between rock blocks. Meltwater infiltrated the network, generating significant destabilizing hydrostatic pressures at the base of the wall. The stress intensity factor (K_I) suddenly exceeded the critical fracture toughness of the weakened granite ($K_{Ic} \approx$ 1.5 to 2.5 MPa.m). At that precise moment, the final rock bridges failed by catastrophic rupture, releasing elastic energy and plunging 265,000 m³ of rock into the void.

4.2 Aiguille du Midi (3,842 m)

The Aiguille du Midi is a primary site for monitoring rock wall degradation. Unlike the massive collapse at Les Drus, this peak undergoes progressive decohesion driven by freeze-thaw fatigue. Deep borehole sensors show rock temperatures approaching 0°C at 10-meter depths, increasing the stress intensity factor (K_I) at crack tips due to hydraulic pressure and ice melt (Magnin and al. 2017).

To compensate for the loss of ice-mediated cohesion, pre-stressed rock bolts were installed. These anchors compress fissures to maintain K_I below the rock's fracture toughness (K_{Ic}). These must be anchored beyond the active layer—the surface zone that thaws annually, typically 2 to 5 meters deep.

A critical technical challenge is the differing thermal expansion coefficients of steel and rock; winter cooling causes steel contraction, which can spike anchor tension. To manage this, real-time extensometers monitor fissure apertures, providing data for the alert thresholds defined in Table 6.

Table 6. Monitoring Parameters and Thresholds.

Parameter	Stable State	Alert (Fatigue)	Danger (Imminent failure)
Opening rate	< 0.5 mm/year	1 to 5 mm/year	Continuous acceleration
Deep Temperature	< -2°C	-1°C à 0°C	Thaw (> 0°C)
Anchor Tension	Nominal	Fluctuations due to Freezing	Sudden increase or drop

5 Conclusion

The stability of periglacial rock walls depends on a complex coupling between thermodynamics, mechanical, and chemical processes. Permafrost degradation is not merely a freeze-thaw phenomenon but a profound rheological transition. As water cools toward 0°C, its volumetric expansion generates cryostatic pressure, significantly increasing the stress intensity factor (K_I) in the rock at crack tips. Conversely, during thaw periods, hydrostatic pressures exert a similar destabilizing influence.

This study demonstrates that large-scale collapses, such as the Bonatti Pillar at Les Drus, result from the cumulative fatigue of rock bridges. This weakening is driven by subcritical crack growth, where cyclic freeze-thaw stresses act in concert with stress corrosion cracking. While engineering solutions like pre-stressed anchors at the Aiguille du Midi can mitigate stability loss by artificially reducingK_I, the inexorable warming of the rock matrix poses a long-term threat. Integrating these thermomechanical models with real-time monitoring remains essential for predicting future instabilities in a rapidly changing alpine environment.

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