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# A review on ice-cores from temperate glaciers: processes, signal preservation, and paleoclimatic significance

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# A review on ice-cores from temperate glaciers: processes, signal preservation, and paleoclimatic significance

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**Abstract.** Temperate glaciers, characterized by ice at the pressure melting point and the coexistence of solid and liquid water, are generally considered unsuitable as natural archives because meltwater undermines the paleoclimatic signals they hold. Historically, ice-core studies have favoured cold glaciers. However, the ongoing atmospheric warming is driving many formerly cold portions of glaciers toward temperate conditions. As such, the relevance of temperate ice as potential paleoclimate archives is increasing. Assessing its ability to record environmental signals has become a priority for ice core science. This review synthesizes more than seventy years of research on temperate ice-cores, tracing the evolution of scientific approaches from pioneering efforts in the 1950s to recent projects across the globe. The behaviour of ice-core proxies -including soluble and insoluble impurities, water stable isotopes, gases, radionuclides, and organic compounds- is discussed in the context of meltwater-related post-depositional processes. By compiling and comparing evidence from diverse settings, this work highlights both the challenges and the emerging opportunities for retrieving meaningful information from temperate glaciers. Understanding how climatic and chemical signals are modified, preserved, or lost in rapidly transforming glaciers is essential for sustaining the role of ice-core science in a warming world.

## 1. INTRODUCTION

*Snowflakes fall to Earth and leave a message*

Henri Bader (1907 Brugg, CH – 1998 Miami, US)

Glacier ice represents an invaluable natural archive, preserving ordered signals of past environmental and climatic conditions. The science of extracting such information began when, during an overwintering expedition in Greenland, Ernst Sorge realized that digging deep into the snow and firn layers was equivalent to accessing increasingly ancient information **(Sorge, 1933)**. Since then, the paleoclimatic exploitation of glaciers has advanced enormously, supported by the development of drilling systems capable to probe the thickness of glaciers through ice-cores. The latter have revolutionized our understanding of Late Pleistocene and Holocene climate history.

Ice-core science divides into two branches, one centred on polar ice sheets, the other on mountain glaciers. Polar ice sheets provide continuous archives over the last hundreds of thousands of years, offering insights into long-term hemispheric and global processes **(EPICA, 2004)**. In contrast, mountain glaciers typically cover shorter intervals, from millennia to decades, but with high temporal resolution. They enable detailed reconstructions of regional climate variability **(Kang et al., 2002; Kozachek et al., 2017)**. Moreover, mountain ice-cores are especially suited to investigate anthropogenic pollution over the past centuries **(Barbante et al., 2004; Eichler et al., 2023)**, linkages between climate, ecosystems, and human societies **(More et al., 2017; Brugger et al., 2021)**, and high-frequency atmospheric processes **(Knüsel et al., 2005; Lindau et al., 2021)**. Finally, because mid-latitude mountain glaciers are located closer to emission sources of many atmospheric species, they are sensitive to the distribution of short-lived tracers **(Schwikowski et al., 1999; Preunkert et al., 2000)**.

Most ice-cores drilled so far, whether from polar or mountain glaciers, have been retrieved from the cold portions of glacier, where the ice is constantly below its pressure melting point. By contrast, temperate ice exists at the pressure melting point and contains liquid water. Because cold ice preserves stratigraphic signals more faithfully, ice-core science has historically focused on it. Today, however, interest in temperate ice-cores is growing, as climate warming is driving many cold glaciers toward temperate conditions.

Global warming not only causes glacier retreat but also raises englacial temperatures, shifting cold ice toward temperate regimes **(Gabrielli et al., 2010; Marshall, 2021)**. Temperate ice is subject to melting, water percolation, and loss of annual snow deposition, all of which can attenuate, relocate, or erase paleoclimatic signals **(Moser et al., 2024)**. Because of climate change, the number of cold mountain glaciers is thus declining **(Gilbert et al., 2010; Hoelzle et al., 2011)**.

In the era of climate change, it is essential to improve our ability to extract and interpret relict information preserved in temperate ice. This requires a better understanding of the physicochemical processes operating under temperate conditions, in order to disentangle post-depositional alterations from climatic signals. Temperate ice-cores have been sporadically studied in the past, but no comprehensive synthesis exists. The aim of this work is to fill this gap by providing an overview of temperate ice-core science, summarizing past research and outlining priorities for the future.

## 2. WHAT IS TEMPERATE ICE? WHAT ARE TEMPERATE GLACIERS?

The temperature of glacier ice on Earth is always close to the melting point, meaning that its homologous temperature, i.e. the ratio of absolute temperature to melting point in Kelvin, is near one. A large fraction of glacier ice actually exists at the melting point where solid and liquid phases coexist **(Lliboutry, 1971)**. The distribution of liquid water in temperate ice is not uniform. Water in temperate ice fills a network of veins developed at the junctions between ice grains (i.e. ice crystals), in particular at junctions where three grains meet (triple junctions; **Nye & Frank, 1972; Mader, 1992**). Despite their micrometer-scale thickness (Fig 1), these veins interconnect, making temperate ice weakly permeable **(Fowler and Iverson, 2022)**. Other types of liquid inclusions are also found inside ice grains in the shape of µm-mm features (see Fig 1).

The equilibrium temperature at which solid and liquid water phases coexist in temperate ice is not always exactly 0°C. It can be lower because of the depression of the melting point due to pressure and impurities present in the ice **(Harrison, 1972; Lliboutry, 1971)**. The water content of temperate ice is controlled by vein and inclusion geometry. Field data indicate seasonal variations but rarely exceed 5% **(Vallon et al., 1976; Murray et al., 2000)**. Beyond this threshold, ice becomes saturated and excess meltwater drains to form supra-, en-, and subglacial channels **(Shreve, 1972)**.

The thermal regime of ice strongly influences the properties and the behavior of glaciers. A solid material at a homologous temperature of 1 is subject to much higher deformation rates compared to the same material at colder temperatures **(Homer and Glen, 1978)**. This has notable effects on the motion of ice sheets and glaciers, making their temperate portions easily deformable **(Ryser et al., 2014)**. Temperature also influences the processes occurring at the interface with the bedrock. The absence of meltwater in cold-based glaciers strongly reduces basal sliding **(Lloyd Davies et al., 2009)**. On the contrary, sliding and erosion are enhanced at temperate-based glaciers **(Alley et al., 2019)**.

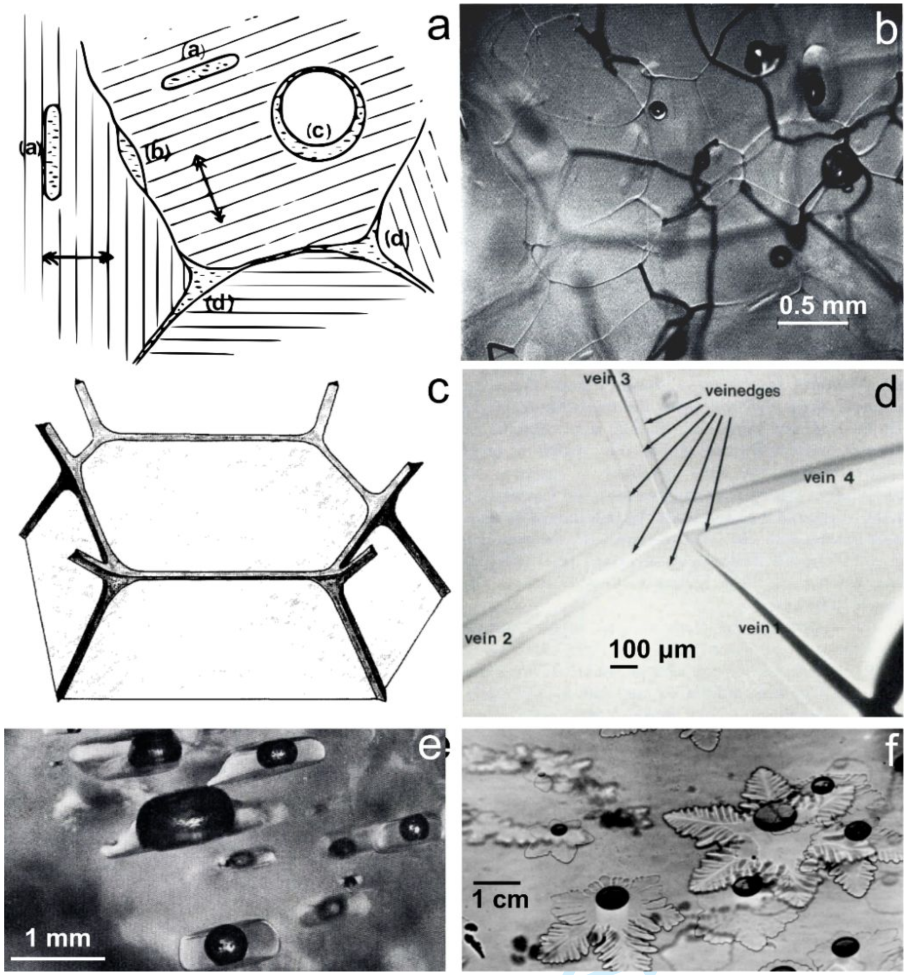


**Fig 1 Microscopic liquid inclusions in temperate ice. Panel a: an overview of the possible types of liquid inclusions, including intra-grain inclusions (a), flat inclusions at the surface separating two ice grains (b), air bubbles surrounded by a film of liquid water (c), inclusions at triple junctions (d); the figure is from Lliboutry (1971). Panel b: the network of liquid water veins developed around ice grains in real temperate glacier ice (from Raymond & Harrison, 1975). Panel c: a sketch representing the geometry of liquid filled veins in temperate ice (from Nye and Frank, 1972). Panel d: liquid water filling the veins found at grain junctions in artificial temperate ice (from Mader, 1992). Panel e: air bubbles surrounded by liquid water pockets in temperate ice (from Raymond, 1976). Panel f: Tyndall's figures in temperate ice: snowflake-shaped cavities filled with liquid water and presenting a central void bubble (from Nakaya, 1956).**

For temperate glaciers, the occurrence of meltwater accelerates the transformation of firn into ice. During partial melting ice-grains round off, enhancing their packing **(Cuffey and Paterson, 2010)**. Packing is also favored by the lubricating effect of water itself. The penetration of winter cold let water-soaked firn to freeze. Ice formed through this mechanism is defined superimposed ice and typically contains little gas as the porosity of firn had been occupied by liquid water before freezing. When superimposed ice forms, the density of firn increases rapidly, reaching the value distinguishing firn from ice (830 kg m$^{-3}$) in a single season **(Koerner, 1970)**. Temperate ice does not entirely consist in superimposed ice, typically layers rich in bubbles are also present, reflecting the alternation between warmer/wetter conditions and colder/drier ones (**Coachman et al., 1958; Vallon et al., 1976**; **Fig 2b**).

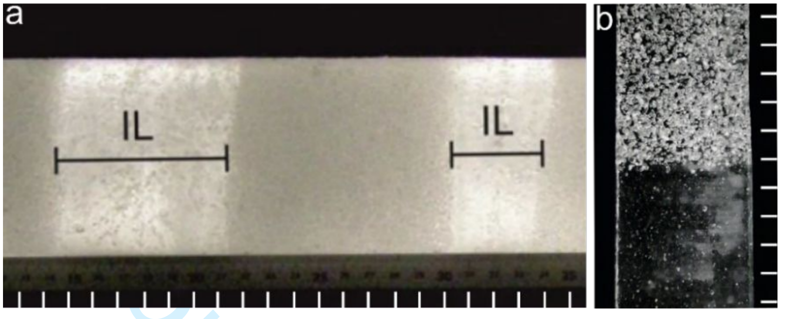


**Fig 2 Effects of meltwater infiltration in firn. (a) Ice lenses (IL) in cold firn with alternating sections of porous firn and more compact ice from refrozen meltwater (ice-core from the Belukha glacier, Siberian Altai, Russia). (b) The resulting, typical structure in temperate ice with bubbly ice alternated with bubble-free compact layers (ice-core from the Adamello glacier, Alps, Italy).**

Glaciologists have been discussing for a long time about the classification of glaciers in relation to their thermal properties. The first of such classifications were independently proposed in the 1930s by Lagally (1932) and Ahlmann (1935) and, given their partial overlap, they are now referred as the *Lagally-Ahlmann classification*, which identifies:

- (1) **cold** (or high-polar) glaciers, which never reach the melting temperature and whose ice is formed through the slow re-crystallization of firn;

- (2) **transitional** (or sub-polar) glaciers, which are basically cold but with some surface melting during summer;

- (3) **temperate** (or warm) glaciers, which consist of ice at the melting point apart from a surface layer which can be cold in winter and temperate in summer. At such glaciers, ice is formed by the re-crystallization of the annual surplus of precipitation through thaw/freezing cycles.

The elegance of this classification lies in its simplicity. However, its oversimplification fails to describe the variable thermal structure of glaciers. More complete classifications were proposed in the 1950s **(Court, 1957)**. Later, as it became clear that a single glacier can hardly be described by a single thermal regime, a new approach emerged. It was based on the fact that most glaciers are characterized by portions with different thermal characteristics. Some portions can lie in a temperate state while others can be cold. Glaciers responding to these features, and they are the majority, are defined polythermal **(Miller, 1976)**.

## 3. ICE TEMPERATURE, ICE-CORES AND CLIMATE CHANGE

Considering ice-cores, the temperature of the glacier at the drilling site is of outmost importance. Until now most ice-cores, including the ones from mountain glaciers, have been retrieved at sites where the ice is cold. The absence of melting guarantees the preservation of paleoclimatic records embedded into glacier ice. As such, a fundamental parameter to assess the feasibility of a potential drilling site is the vertical distribution of temperature within the glacier **(Schwikowski et al., 2013; Kutuzov et al., 2016)**.

Fig 3 provides examples of borehole temperature profiles measured at drilling sites. In most cases, temperature is well below 0°C, implying that the drilling sites are glaciologically cold. Only one location (Blue Glacier) shows a flat temperature profile corresponding to 0°C, revealing temperate conditions. Alto dell'Ortles glacier shows a composite behavior due to the recent atmospheric warming: the upper 30 m are temperate, below the ice is cold **(Gabrielli et al., 2016)**.

Ice temperature in glaciers is strictly controlled by atmospheric temperature. Glaciers present a cold portion where the mean annual air temperature is below –15 °C. At this condition air and ice temperatures are nearly identical a few meters below the surface **(Zagorodnov et al., 2006)**. At warmer sites, even sporadic summer melt allows water to percolate into the firn and refreeze (**Fig 2a**), releasing latent heat. This process significantly warms the ice. Where mean annual air temperature exceeds –15 °C, ice can be up to 15 °C warmer than the mean air temperature, leading to temperate conditions despite a negative annual mean **(Zagorodnov et al., 2006)**.

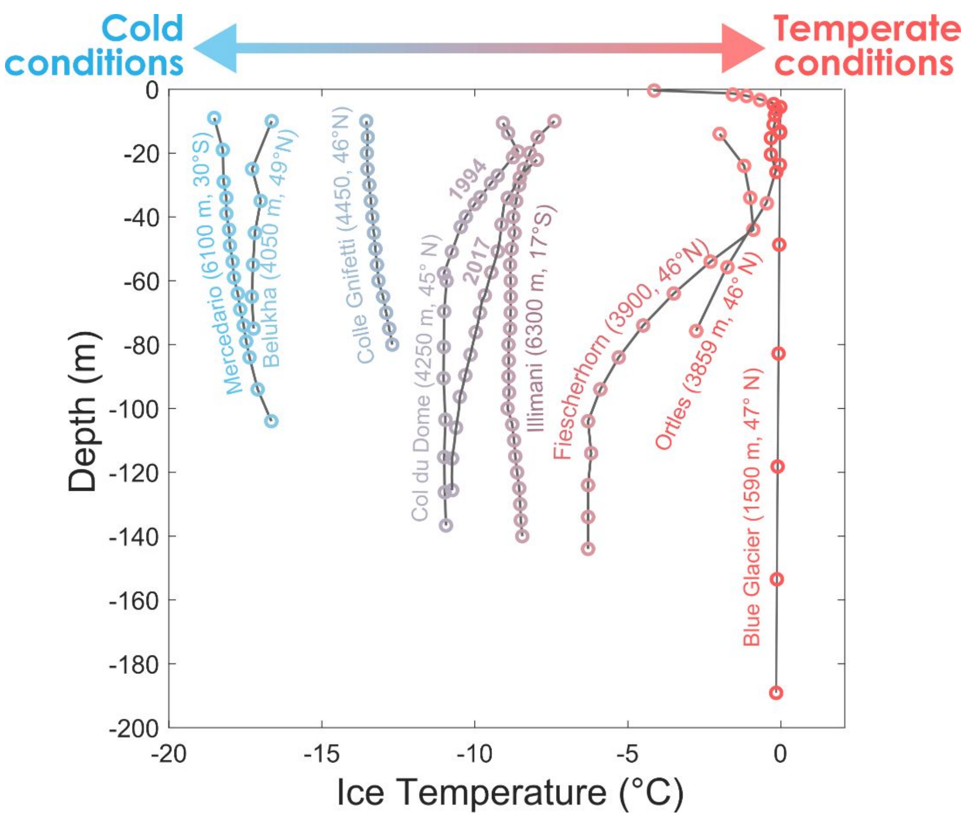


**Fig 3 Borehole temperature profiles measured at ice-core drilling sites. Data from: Cerro Mercedario, Argentinean Andes (Schwerzmann, 2006; Vimeux et al., 2009); Belukha, Siberian Altai (Olivier et al., 2003); Colle Gnifetti, Alps (Schwerzmann, 2006); Col du Dome, Alps (Vincent et al., 2020); Illimani, Bolivia (Vimeux et al., 2009)) Fiescherhorn, Alps (Schwerzmann et al., 2006); Ortles, Alps (Gabrielli et al., 2016); Blue Glacier, Olympic Mountains (Harrison, 1972). For the Col du Dome site two temperature profiles (1994 and 2017), are shown.**

Climate change is altering the thermal state of glaciers. It could be argued that this translates into higher ice temperatures, but this is not always observed. Paradoxically, small glaciers now lying below the equilibrium line are cooling, the loss of the snow/firn cover, an effective winter insulator, exposes them ice to the penetration of winter cold waves **(Huss**

**and Fischer, 2016)**. By contrast, the upper parts of large mountain glaciers, situated above the equilibrium line, are undergoing marked warming. Cold sectors of glaciers are particularly vulnerable: they warm through direct atmospheric heat exchanges and through latent heat from refreezing of infiltrating meltwater **(Cuffey and Paterson, 2010)**. Temperate glaciers respond differently: they are primarily sensitive to mass loss, since any additional energy input is directly available to melt ice.

At Col du Dôme (Mont Blanc, Alps), englacial temperatures increased from the surface down to 100 m depth between 1994 and 2017 (Fig 3; **Vincent et al., 2020**). Comparable trends are reported from Monte Rosa (Alps), where temperatures at 20 m depth rose by 6-7 °C between 1991 and 2008 **(Hoelzle et al., 2011)**. These rates are an order of magnitude higher than concurrent atmospheric warming due to latent heat effects **(Gastaldello et al., 2025)**. At Alto dell'Ortles (Alps), a site formerly characterized by cold ice, the upper 40 m have already transitioned to temperate conditions **(Gabrielli et al., 2016)**.

## 4. CHEMICAL AND PHYSICAL PROCESSES IN TEMPERATE ICE

Glacier ice forms from atmospheric precipitation, and its composition reflects atmospheric conditions and constituents. Yet signals preserved in ice are not completely stable and undergo post-depositional alteration, from polar ice sheets to temperate mountain glaciers. Documented processes include: (1) snow redistribution by wind and partial loss of precipitation **(Fisher et al., 1983)**; (2) snow/firn sublimation **(Ginot et al., 2006)**; (3) re-emission of volatile species **(Wagnon et al., 1999)**; (4) gas fractionation in firn **(Huber et al., 2006)**; (5) impurity mobilization during recrystallization or meltwater percolation **(Moser et al., 2024)**; (6) chemical diffusion **(Cuffey and Steig, 1998)**; (7) layer thinning from ice flow; and (8) englacial geochemical reactions **(Baccolo et al., 2021)**. Some of these processes occur regardless of thermal regime (e.g., wind redistribution, layer thinning), others are typical of cold polar contexts (volatile re-emission, gas fractionation, englacial reactions), while several are enhanced under temperate conditions due to liquid water. Here we summarize the main results on post-depositional processes occurring in temperate ice.



### 4.1. Salts, ionic species and self-purification of temperate ice

In his pioneering work, Renaud (1949) researched the distribution of impurities into glacier ice exploiting fractional melting, with the first fraction corresponding to the outer layer of ice grains (crystals) and the last one to their central portions (Fig 4a). He utilized conductivity as a proxy for the concentration of ionic salts. These are his main results: (1) "*Whatever its origin the glacier grain is always composed of a crystal of pure ice surrounded by a saline skin*"; (2) "*The salinity of the skin is greater in the ice of the névé* [the accumulation area of glaciers] *than in that of the glacier proper* [intended as the glacier tongue]"; (3) "*The purity of the ice constituting the nucleus of the crystal is greater in the ice of the terminal region than in that of the névé*". The glaciologist first observed that the distribution of salts into glacier ice grains is not uniform. They are concentrated in the outer layer. This is visible in Fig 4a, where for the two ice specimens (cold and temperate ice) the highest conductivity is observed in the first melting fraction, corresponding to the outer part of grains. Although the trend is similar for cold and temperate ice, in temperate ice the conductivity is lower, revealing that temperate ice is purer than cold ice **(Renaud, 1949)**.

Temperate ice is not only purer than cold ice, but also purer than the snow from which it forms, as first evidenced by Gorham (1958). Moreover, temperate ice is not always pure in the same way: the deeper (and the older) the temperate ice, the purer it is **(Harrison and Raymond, 1976; Bouard, 1977)**. Harrison & Raymond (1976) reported that "*Blue Glacier ice may contain substantially less salt than ice from the polar regions, even though it is located within 60 km of the Pacific Ocean*". The two authors also found an inverse correlation between salt concentration and ice grain size (Fig 4b).

During recrystallization, ionic impurities incompatible with the ice lattice are expelled to grain boundaries **(Wolff, 1996)**. In cold ice these impurities remain immobile and can be preserved for hundreds to thousands of years, enabling a correct reading of paleoclimatic signals **(Stoll et al., 2023)**. In temperate ice, however, grain boundaries host water-filled veins (Fig 1a-d). Expelled impurities dissolve and are transported with liquid water **(Harrison and Raymond, 1976)**.

Meltwater present in junction veins thus becomes enriched in solutes and eventually flushes into the macroscopic drainage system of channels and cavities of the glacier, leading to large-scale impurity removal and purification **(Davies et al., 1982; Glen et al., 1977)**.

Since liquid water is constantly present in temperate ice, the purification is continuous: the older and deeper the ice, the purer it is **(Harrison and Raymond, 1976)**. The process is also rapid. Most salts are removed before firn turns into ice. This is because at temperate glaciers the deep part of firn is saturated with meltwater accumulating above the transition to ice (water table). Within soaked firn, solid–liquid exchanges are enhanced, accelerating impurity removal and the formation of extremely pure solid ice **(Harrison and Raymond, 1976; Davies et al., 1982)**. The process can be observed in Fig 4d: below the transition to ice, ionic species rapidly reach a stable and extremely low concentration.

The firn/ice transition is critical in controlling the distribution of ionic impurities. At this depth, their concentration can locally increase, as illustrated in Fig 4d (area highlighted in red). The concentration of ions increases between 12 and 15 m depth, reaching its maximum values for sulfate, nitrate, $Mg^{2+}$, pH and $Cl^-$. This enrichment results from the accumulation of liquid water and eluted impurities within the water table developed above the transition. Below this depth, liquid water drastically decreases, accompanied by a corresponding decrease in the concentration of impurities **(Davies et al., 1982)**.

The removal of ionic impurities during melting has implications for hydrology and ecosystems. In the 1970s, anthropogenic acidification led to snowfall enriched in sulfuric and nitric acids **(Likens et al., 1979)**. Studies showed that more than 60% of the acidity ($H^+$) stored in snow was released within the first 30% of meltwater, with significant ecological impacts **(Johannessen and Henriksen, 1978)**. Alongside $H^+$, nitrate, sulfate, and some heavy metals were also concentrated in the early melt fractions (Fig 4c). This phenomenon, known as the ionic pulse or acid shock **(Bales et al., 1989)**, is strictly related to the self-purification of temperate ice: both involve the preferential removal of soluble species during melt. The acid shock does not only relate to polluted snow, but also to natural ionic species present in the snowpack **(Li et al., 2006)**.

Over the past 40 years, research on the distribution and mobility of ionic species in melting glacier ice has greatly advanced, with particular focus on the differential behaviour of individual ions. During melting and recrystallization, some ions are preferentially eluted, depending on their solubility in ice. Numerous elution sequences have been proposed to describe the preferential order of ionic release under temperate conditions (**Moser et al., 2024** and references therein). Highly soluble, ice-incompatible ions such as sulfate, nitrate, and $Ca^{2+}$ are eluted first, while more compatible ions like chloride and ammonium are only marginally fractionated **(Eichler et al., 2001; Vega et al., 2016; Avak et al., 2019; Trachsel et al., 2019)**. Most studies on elution sequences have been carried out on seasonal snow **(Avak et al., 2019)** or on sporadic melt/refreeze events which occur in cold glacier firn/ice **(Eichler et al., 2001)**. Yet the few investigations on temperate ice show broadly similar patterns, though with continuous rather than episodic fractionation **(Davies et al., 1982)**. By normalizing sequences compiled by Moser et al. (2024), a mean elution order can be expressed as:

$$SO_4^{2-} > NO_3^- > Ca^{2+} > Mg^{2+} > K^+, Na^+ > Cl^- > NH_4^+$$

**Equation 1**

Only ions previously reported in at least ten sequences are included in Equation 1; others have been less frequently studied, including the anion of methane sulfonic acid (MSA), behaving between $NO_3^-$ and $Ca^{2+}$ **(Moore et al., 2005; Spolaor et al., 2021)**; $I^-$, between $Ca^{2+}$ and $Mg^{2+}$ **(Spolaor et al., 2021)**; $Br^-$ and $H^+$, between $Na^+$ and $Cl^-$ **(Brimblecombe et al., 1985; Herreros et al., 2009; Spolaor et al., 2021)**; and $F^-$, typically the least fractionated ion **(Eichler et al., 2001; Ginot et al., 2010)**. Equation 1 should be regarded as indicative, since local conditions strongly affect mobility: for example, at maritime sites $Na^+$, owing to its abundance, is often the first ion to be eluted **(Spolaor et al., 2021)**.

The differential mobility of ions during melting enables the use of concentration ratios as melt indicators in ice-cores. If one species elutes before another, its residual concentration decreases faster, altering original ratios characteristic of fresh snow. Two ratios are commonly used: $Cl^-/Na^+$ in non-maritime contexts, where melting increases the ratio by preferentially removing $Na^+$ **(Eichler et al., 2001)**, and $Mg^{2+}/Na^+$ in maritime contexts, where melting decreases the ratio by preferentially removing $Mg^{2+}$ **(Iizuka et al., 2002)**. An additional index was proposed by Grinsted et al. (2006): $log(Na^+/Mg^{2+})$.

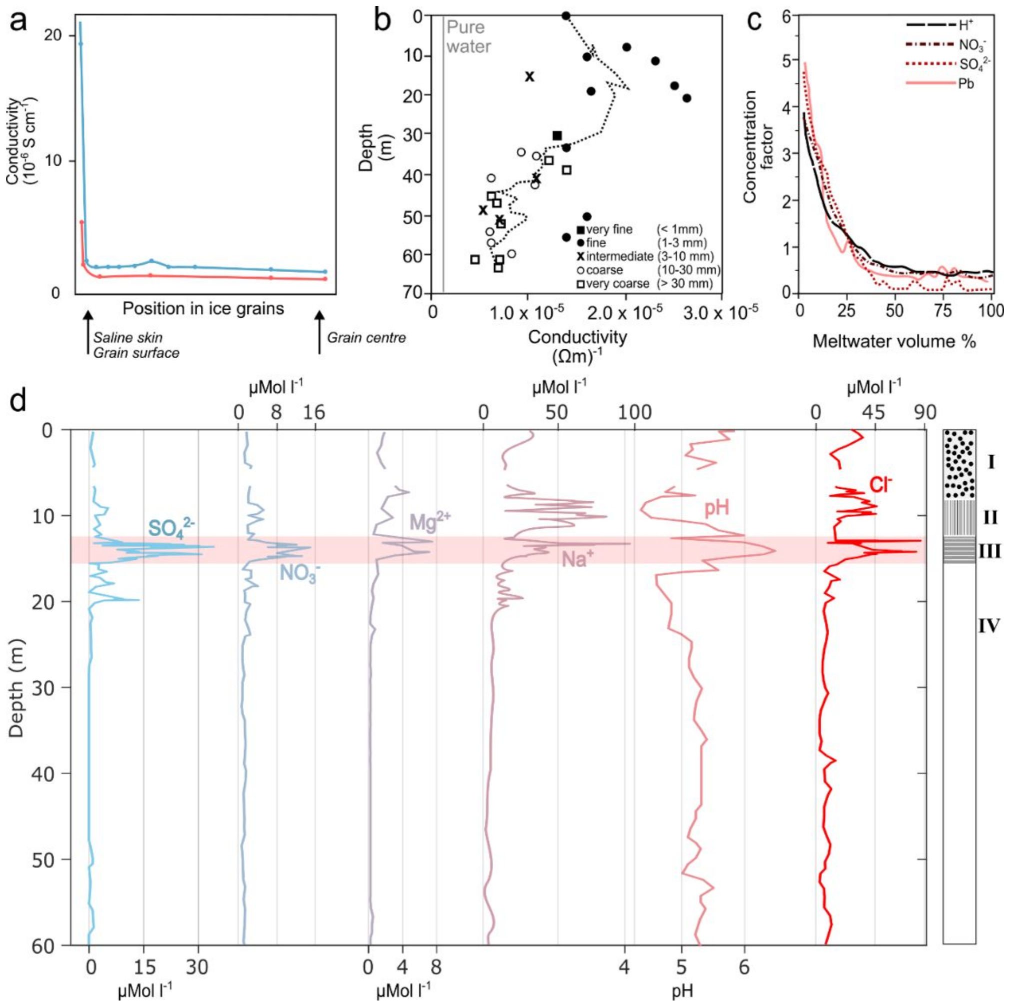


**Fig 4 Temperate ice self-purification. Panel a (redrawn from Renaud, 1949): the conductivity of meltwater fractions produced from different portions of glacier ice grains (in blue a sample of cold ice, in red of temperate ice). Panel b (redrawn from Harrison & Raymond, 1976): conductivity and ice grain size vs. depth in a temperate glacier; the dotted line is the moving average (sampling window: 5). Panel c (redrawn from Johannessen & Henriksen, 1978): impurities in fractions of meltwater from melting snow; data are expressed as the ratio between the concentration observed in meltwater and in the bulk snow sample. Panel d (redrawn from Davies et al. 1982): ionic species in a temperate ice-core; on the right information about the firn/ice structure (I, snow and low density firn; II, consolidated porous firn; III, ice with vertically-connected bubble systems; IV, bubbly compact ice); the red area highlights section III of the core.**

Analyses of meltwater from temperate glaciers show a fraction of meltwater being highly enriched in solutes relative to snow, ice, or bulk runoff **(Tranter et al., 2002)**. The enrichment is attributed to englacial mineral weathering but may also result from impurities expelled into the microscopic vein network during self-purification. Such a mechanism would explain the chemical anomalies of basal temperate ice, where salt concentrations are elevated and dominated by ice-incompatible ions (Hubbard & Sharp, 1989; and references therein). Progressive enrichment of percolating meltwater with these impurities may therefore contribute to the distinct chemistry of basal ice.

### 4.2. Radionuclides

Beginning in the 1960s, newly developed techniques to measure natural and artificial fallout of radionuclides in glacier ice were first applied in polar regions, and soon extended to mountain glaciers. A first study carried out on a temperate glacier revealed that radionuclide concentrations in temperate firn did not match those in fresh precipitation **(Picciotto et al., 1967)**. Specifically, levels of $^{90}Sr$ (an artificial fission product, half-life 28.8 yr) in firn were one tenth of those measured in snowfall, whereas $^{210}Pb$ (a natural fallout product, half-life 22.3 yr) showed comparable concentrations. At the time, the authors could not explain these differences, though they suspected meltwater played a role.

The contrasting behaviors are now understood as a consequence of differing chemical mobility, analogous to ionic species. Strontium, present as the soluble $Sr^{2+}$ cation and poorly soluble in ice, readily segregates into meltwater and shows little affinity for solid substrates **(Kaplan et al., 2014)**. In contrast, Pb is far less mobile. Atmospheric $^{210}Pb$, produced from $^{222}Rn$ decay, attaches to aerosols -especially mineral dust- before being scavenged by precipitation **(Baskaran, 2011)**. Once deposited with snow, it is already bound to particulate matter, making it less susceptible to post-depositional remobilization induced by meltwater **(Gäggeler et al., 2020)**.

Early studies on the vertical distribution of fallout radionuclides in temperate glaciers showed that radioactive peaks linked to specific events, such as the 1986 Chernobyl accident, broadened and decreased in intensity over time **(Ambach et al., 1989)**. On Austrian temperate glaciers, gross beta activity measurements revealed that a few months after the accident the fallout was confined to a snow layer less than 1 m thick. Two years later the peak extended across 5 m of firn and, cnosidering radioactive decay, had lost ~35% of the originally deposited inventory (**Ambach et al., 1988, 1989**; Fig 5a-e).

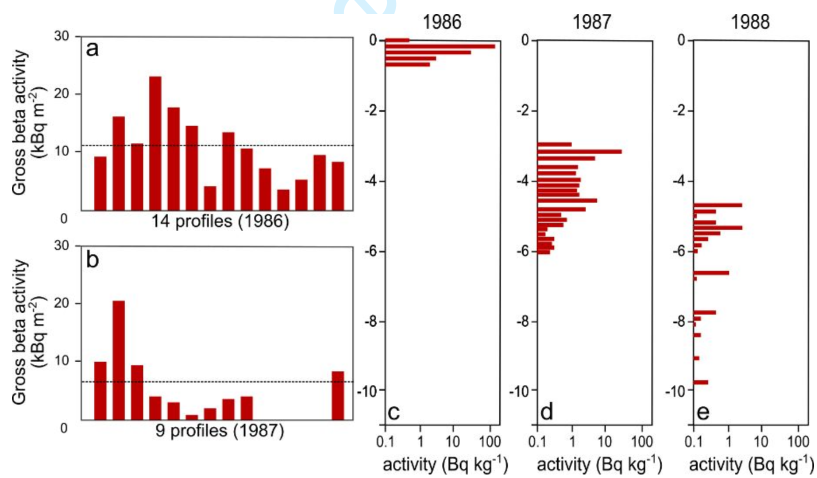


**Fig 5 The distribution of artificial radionuclides in temperate glaciers. Panels a and b show gross beta activity measured in multiple snow/firn vertical profiles at Kesselwandferner (Oetztal Alps, Austria) in 1986, a few months after the Chernobyl accident (a), and in May 1987 (b, redrawn from Ambach et al., 1988). For panels a,b dotted lines refer to mean values observed in the two years, data are expressed in terms of inventories. In panels c,d,e the vertical distribution of gross beta activity concentration into the ice determined at three sites on the same glacier in summer 1986, 1987, 1988 (redrawn from Ambach et al. 1989).**

Such broadening and partial loss reflect the influence of meltwater, which mobilizes soluble and ice-incompatible radionuclides. By contrast, in cold glaciers fallout nuclides stay in their depositional layers and are usable for ice-core chronologies **(Eichler et al., 2000)**. Dating temperate ice with fallout radionuclides remains possible but requires a careful selection of species of interest **(Pavlova et al., 2014; Festi et al., 2021; Di Stefano et al., 2024)**. Among these, tritium ($^{3}H$) and $^{210}Pb$ are the most promising. Tritium, incorporated into water molecules, is unaffected by chemical fractionation. Although snow loss and meltwater percolation reduce its inventories and partly redistribute it into deeper ice **(Oerter and Rauert, 1982)**, its partial mobility has repeatedly allowed its application as a dating tool for temperate ice **(Pinglot et al., 2003; Wel et al., 2011; Di Stefano et al., 2024)**. $^{210}Pb$, strongly associated with insoluble particulate matter, is also well preserved due to its low solubility and association with insoluble impurities **(Von Gunten et al., 1982; Gäggeler et al., 2020)**. In addition, evidence suggests that $^{36}Cl$ may be also relatively stable in temperate ice **(DeWayne Cecil and Vogt, 1997)**.

Other artificial radionuclides -primarily fission products such as $^{137}Cs$, often applied in cold ice-core dating **(Eichler et al., 2000)**- are unsuitable in temperate contexts. Their distribution does not reflect original atmospheric deposition but instead mirrors the abundance of insoluble impurities. Meltwater remobilizes these nuclides and promotes their attachment to dust particles, producing concentration peaks unrelated to the true depositional signal **(Von Gunten et al., 1982; Di Stefano et al., 2019)**.

### 4.3. Trace elements

While dozens of studies have examined the behaviour of ionic species in glaciers when meltwater is present, far fewer have focused on trace elements. Yet several works demonstrate that certain elements -particularly Pb and Hg- can be preserved under temperate conditions and used them as paleoclimatic proxies and for refining ice-core chronologies **(Schuster et al., 2002; Neff et al., 2012; Kang et al., 2015)**. Their relative immobility is attributed to low solubility and strong association with particulate matter, which prevents relocation by meltwater.

Most studies, however, have addressed only individual elements. Broader comparisons are rare, with the most comprehensive insights coming from an ice-core from a cold glacier, but presenting a section affected by severe meltwater infiltration and refreezing through a crevasse **(Avak et al., 2018)**. By comparing 35 trace elements in pristine and melt-affected sections (see Fig 6), the authors found strong mobilization for alkali and alkaline earth metals (Ba, Mg, Sr, Ca, Na) and some transition metals (Cd, Mn, Ni, Co, Zn). These elements are characterized by moderate water solubility and/or incompatibility with the ice lattice. In contrast, elements preserved in melt-affected ice included crustal markers (REEs, Al, U, Fe, Cs) and volatile pollution-related elements (Sb, Pb, Cu, Mo). Their low mobility is explained considering that while in the atmosphere, these elements are adsorbed on mineral dust; once deposited on glaciers they are thus associated with insoluble impurities **(Marx et al., 2008)**.

From these results, three factors were identified as key to trace element preservation under melting: (1) low water solubility, (2) binding to insoluble particles, and (3) low initial concentrations. The latter effect is especially important for ultra-trace elements, which tend to be uniformly distributed within the ice lattice and thus less prone to remobilization. Unlike ionic species, where charge density and solubility in ice largely determine elution, these properties appear less critical for trace elements, except in the case of alkali and alkaline earth metals, whose high charge density and water solubility make them especially mobile (Fig 6).

Follow-up work by the same authors on melting seasonal snow **(Avak et al., 2019)** revealed similar patterns but with systematically stronger elemental depletion compared to melt-affected glacier ice. For example, in the melt-affected ice-core section 72% of Sr was lost compared to the unaffected sections, while in the melting snowpack the loss was limited to 63%. The difference likely reflects the one-off nature of seasonal snowmelt, compared to repeated percolation events through a glacier crevasse **(Avak et al., 2018)**. In truly temperate ice, where interaction between solid and liquid phases is continuous, elemental fractionation could be even more pronounced, though a dedicated study is still lacking.

Other investigations confirm that elements associated with mineral dust and pollution sources tend to be best preserved in melt-affected ice **(Clifford et al., 2023; Potocki et al., 2025; Wong et al., 2013)**.

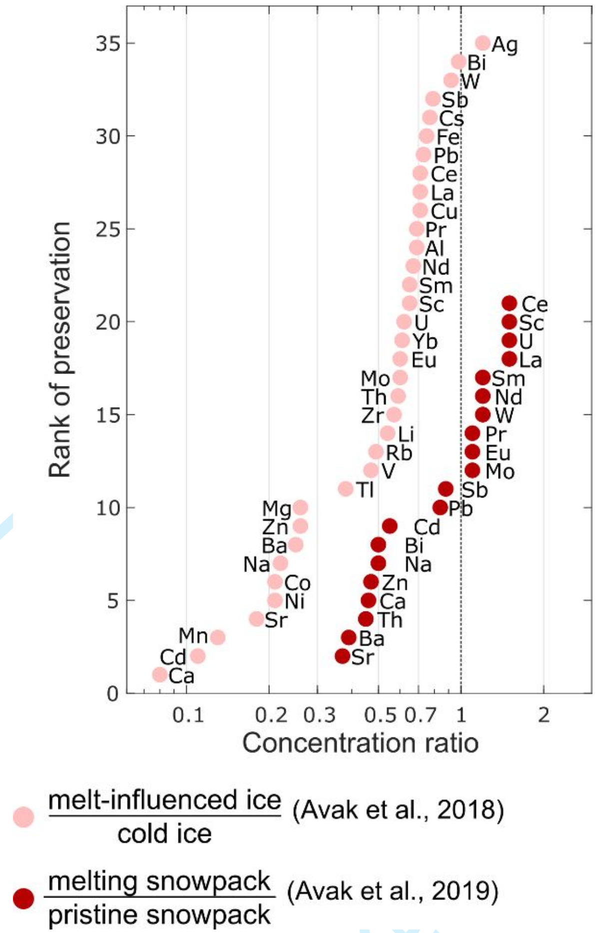


**Fig 6 Fractionation of trace elements in melt-affected snow/ice. For each element the ratio between the median concentration in melt-affected snow/ice and in pristine snow/ice is reported. Pink dots refer represent data from the Grenzgletscher ice-core, which include a section affected by meltwater infiltration from a crevasse (Avak et al., 2018); dark red dots refer to trace element concentrations in melting snow compared to the same snowpack prior to melting (Avak et al., 2019).**

### 4.4. Organics

The occurrence, distribution, and fate of organic compounds in glacier ice remains one of the least explored areas of ice-core research, particularly in temperate glaciers. This is a major gap, since glaciers play a key role in the environmental cycling of organics through the cold-trapping mechanism **(Wania et al., 1998)**.

Pioneering studies, inspired by the discovery of the ionic pulse, showed that organochlorine pesticides (OCPs) and polycyclic aromatic hydrocarbons (PAHs) undergo fractionation during snowmelt, with polar, water-soluble compounds enriched in the first meltwater fractions and insoluble, particle-bound species concentrated later **(Simmleit et al., 1986; Schöndorf and Herrmann, 1987)**. Further studies revealed that contrary to ionic species, whose elution is simply controlled by compatibility with the ice lattice, for organics more factors are involved. They include water solubility (often described by log $K_{ow}$), molecular properties such as functional groups and weight, content of insoluble impurities in snow/ice and ice-grain properties. As summarized by Grannas et al. (2013), five elution profiles are recognized for organics (Fig 7): (a) polar, water-soluble compounds eluted rapidly and enriched in early meltwater fractions; (b) insoluble, particle-bound species retained and concentrated in late fractions; (c–e) intermediate behaviours reflecting differences in functional groups, pH, molecular weight, and affinity for insoluble impurities and snow grains.

In general, lighter molecules within the same compound class tend to be more mobile than heavier congeners **(Grannas et al., 2013; Steinlin et al., 2015)**. Modelling and observations show that less soluble molecules can behave like water-soluble compounds (Fig 7a) in snow with low particle content **(Meyer and Wania, 2011)**. When the particulate matter is low, the bulk of the hydrophobic species dissolves in the liquid phase. In impurity-rich snow, even moderately water-soluble compounds can sorb to particles and elute later than expected, resembling the behaviours displayed in Fig 7b. Snow microstructure also influences elution **(Meyer and Wania, 2011)**. In aged, coarse-grained snow/firn with reduced internal surface area, compounds with strong affinity for the snow grain surface elute early. In contrast, in fresh, fine-grained snow with a larger surface area, these compounds can show a more delayed release (Fig 7c).

For temperate glaciers, where ice is in continuous equilibrium with liquid water rather than undergoing discrete melt events, only part of the elution patterns shown in Fig 7 apply. Mobile, soluble species are progressively depleted from the ice phase, while insoluble, particle-bound species are more effectively retained. The "final-fraction" enrichments typical of seasonal snowmelt (Fig 7b-e) are less relevant for temperate ice.

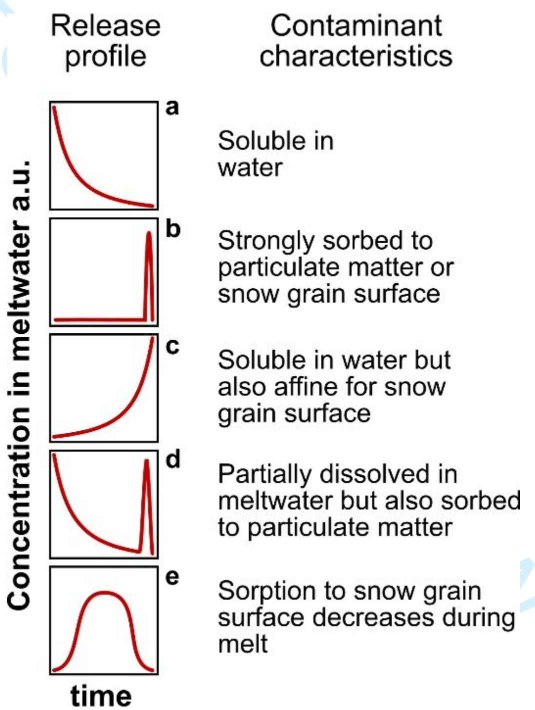


**Fig 7 Different behaviors of organic species observed in laboratory experiments during snowpack melting. Redrawn from Grannas et al., (2013).**

The first study to target organics in temperate ice **(Donald et al., 1999)** analysed OCPs in ice samples from a crevasse in the Canadian Rockies, demonstrating that temperate ice can preserve deposition records, though the impact of meltwater-related postdepositional processes remained unclear **(Gregor and Peters, 2000)**. A study on a Swiss temperate ice-core revealed that ~75% of the original PCB burden had been lost, primarily via revolatilization of lighter congeners and runoff of heavier ones; yet heavy congeners were comparatively well preserved owing to their adsorption on insoluble impurities **(Pavlova et al., 2014; Steinlin et al., 2015)**. Together with observations of contaminated meltwater from retreating glaciers, these findings show that temperate glaciers can act as both a sink and secondary source of persistent pollutants to downstream environments **(Bogdal et al., 2009; Ferrario et al., 2017; Li et al., 2017; Pavlova et al., 2016)**.

Other classes of persistent organic pollutants (POPs) show similarly variable behaviours. PAHs, with high log $K_{ow}$ values (>5), are strongly particle-bound and poorly mobile, while compounds such as γ-HCH (log Kow ≈ 3.7) display mixed elution between early and late meltwater fractions **(Simmleit et al., 1986)**. For per- and polyfluoroalkyl substances (PFAS), mobility depends on chain length: short-to-medium chain PFAS (≤9 C atoms) are more water-soluble and readily relocated by meltwater, while longer-chain PFAS are comparatively less mobile **(Plassmann et al., 2011; Kirchgeorg et al., 2016; Hartz et al., 2023; Zhou et al., 2024)**. However, differences in dust content can alter expected patterns, allowing relatively hydrophobic PFAS to behave like soluble compounds **(Hartz et al., 2023)**.

Among PFAS, trifluoroacetic acid (TFA) is receiving particular attention. Owing to its short-chain structure, high polarity, and solubility, TFA is highly mobile in the environment with liquid water and has been identified as an emerging contaminant of concern. Its presence in glacier ice and meltwater is being increasingly investigated, with recent studies focusing on its release from melting glaciers and subsequent downstream transport **(Wu et al., 2024; Zhou et al., 2025)**. Studying TFA in temperate ice-cores would surely contribute to a better understanding of its behaviour and partitioning in glacial environments.

Far fewer studies have examined naturally occurring organic compounds in temperate ice. One investigation reported that 12 targeted secondary organic aerosols, being soluble and mobile, were rapidly lost during meltwater percolation **(Müller-Tautges et al., 2016)**. Yet, these compounds represent a tiny fraction of the tens of thousands of organic molecules expected in atmospheric aerosols and glacier ice. Recent developments allow the detection of hundreds of organic aerosol tracers in snow and ice samples without any a priori knowledge of their molecular composition **(Burgay et al., 2023)**. This non-targeted approach has been applied to an ice-core strongly influenced by melting, yielding ~250 identified molecules **(Huber et al., 2025)**. Of these, ~170 were completely eluted from the ice layers most affected by meltwater percolation, predominantly small, polar and water-soluble species. In contrast, a suite of larger, less soluble molecules remained preserved. Despite strong meltwater alteration, temperate ice can retain some stratigraphic information for specific organic tracers.

Both natural and anthropogenic organics in ice are strongly affected by post-depositional processes including revolatilization, photochemical and microbial degradation, particle deposition, and meltwater-driven fractionation **(Grannas et al., 2013)**. Even in cold ice the concentration of organic species decreases, mostly because of UV-mediated reactions **(Jaffrezo et al., 1994; Hermanson et al., 2020)**.

### 4.5. Insoluble impurities

Glacier ice contains solid particles such as mineral dust, pollen, algae, black carbon, and organic fragments. Unlike dissolved species, these particles do not directly interact with the ice molecular lattice and are largely unaffected by meltwater fractionation or grain-boundary processes, making them comparatively stable under temperate conditions.

This stability has enabled their use as environmental and climatic proxies in temperate ice-cores **(Uetake et al., 2006; Neff et al., 2012; Kaspari et al., 2020; Mangili et al., 2025)**. Many insoluble impurities are deposited seasonally on glaciers and their stratigraphy remains well preserved regardless of ice thermal regime. Summer surface melt further enhances these signals by concentrating impurities in the residual layers, especially in regions with little or no summer accumulation, such as the Andes **(Thompson et al., 1979; Reis et al., 2022)**. According to this, seasonal variations in the concentrations of algae **(Kohshima et al., 2007)**, pollen **(Nakazawa et al., 2004; Festi et al., 2017; Takeuchi et al., 2019)**, dust **(Neff et al., 2012; Reis et al., 2022)**, black carbon **(Pavlova et al., 2014; Festi et al., 2021)**, and organic fragments **(Uetake et al., 2006)** have been used to identify annual layers and refine temperate ice-core chronologies. Multi-proxy approaches have also been successfully applied to identify a seasonal signal in temperate ice-cores **(Mangili et al., 2025)**.

### 4.6. Water stable isotopes

Ratios of water stable isotopes δ-Oxygen and δ-hydrogen isotopes ($\delta^{18}O$, δD) are key ice-core proxies linked to temperature and climate changes **(Jouzel et al., 1997)**. While their post-depositional alteration is minor in cold glaciers, temperate contexts strongly modify the original signal, hampering their paleoclimatic application. Despite this, the earliest investigations into the hydrogen and oxygen isotopic composition of glacier ice were actually conducted on temperate glaciers.

Early studies on temperate ice carried out in the late 1950s and 1960s **(Epstein & Sharp, 1959; Sharp et al., 1960; Deutsch et al., 1966)** showed enrichment in heavy isotopes in ice and firn relative to fresh snow. This was attributed to spring melt and summer rain, which introduced isotopically heavier water into the glacier. Smoothing of seasonal variability was also observed and interpreted as a consequence of repeated melt-refreeze cycles **(Epstein and Sharp, 1959; Deutsch et al., 1966)**. Isotopic seasonality persisted only in the upper firn, disappearing a few meters below the surface **(Sharp et al., 1960; Deutsch et al., 1966)**. Fig 8b shows the marked difference between the records from cold and temperate glaciers in the same region. While the cold site presents a well-expressed seasonality, the two temperate sites display a much more smoothed record **(Schotterer et al., 2004)**.

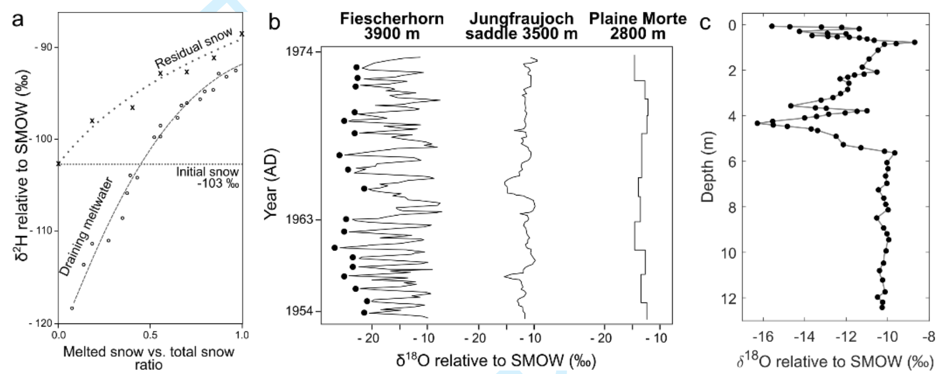


**Fig 8 Distribution and alteration of water stable isotopes in temperate ice. Panel a: changes in δD of meltwater and residual snow during a melting experiment (redrawn from Herrmann et al., 1981). Panel b: vertical distribution of $\delta^{18}O$ in firn and ice from different Alpine glaciological contexts: Fiescherhorn (cold conditions), Jungfraujoch saddle and Plaine Morte (temperate conditions with varying degrees of meltwater percolation). For the Fiescherhorn record, black dots mark isotopic minima corresponding to winter snowfall (redrawn from Schotterer et al., (2004). Panel c: typical $\delta^{18}O$ distribution in shallow temperate firn (data from a core drilled at the Northern Patagonia Icefield), showing a seasonal signal near the surface and homogeneous composition below a few meters (data from Yamada, 1987).**

A major step forward came with Árnason (1969), who showed that percolation alone could not explain the enrichment in heavy isotopes observed in temperate ice. The author observed that temperate ice can present an isotopic composition heavier (less depleted using the δ notation) than the isotopically heavy summer precipitation. He concluded that in temperate firn and ice, isotopic fractionation must also take place in addition to smoothing and homogenization. Experiments confirmed that early meltwater released by melting snow/ice is depleted in heavy isotopes and late fractions enriched (**Árnason, 1969; Herrmann et al., 1981**; see Fig 8a). Árnason (1969) proposed a balance equation (Equation 2) to estimate annual melt losses from the difference in terms of isotopic composition between fresh snow precipitation and firn/ice.

$$R_i q + R_w(1-q) = R_p$$

**Equation 2**

In Equation 2 the mean concentration of a heavy isotopes ($^{18}O$ or $^{2}H$) in total annual precipitation is represented by $R_p$; $q$ is the fraction of annual precipitation remaining on the glacier as firn and ice; $1\text{-}q$ is the fraction of precipitation lost through melting; $R_i$ and $R_w$ are the concentrations of heavy isotopes in the remaining and lost fractions respectively. The equation is highly simplified and has limited utility when applied to real-world data. This is primarily due to the assumption that the isotopic composition of melting firn and ice is homogenous, a condition that may hold only in highly maritime environments, where the seasonal variability in snowfall isotopic composition is minimal. A more advanced approach was proposed by Búason (1972), who developed a set of equations to describe more comprehensively the isotopic evolution of temperate firn and ice affected by exchanges with liquid water.

Árnason (1981) added a process to the discussion: persistent isotopic exchange within the saturated basal firn layer above the firn-ice transition (see Fig 8c). In this zone, firn is permanently saturated with meltwater, promoting continuous exchanges between the solid and liquid phases. Unlike in the shallow firn, where melting and infiltration of rain are limited to summer, the basal firn layers experience persistent isotopic exchanges and homogenization (Fig 8c).

Studies at Vernagtferner **(Stichler et al., 1982; Oerter et al., 1985)** first considered *deuterium excess* ($d = \delta^2H - 8\delta^{18}O$), a second-order parameter influenced by non-equilibrium fractionation during phase changes. Oscillations in $d$ were found to be more clearly expressed than those in the individual isotope ratios. This made the deuterium excess a useful parameter for identifying seasonal cycles and for providing dating constraints **(Stichler et al., 1982)**. A difference between oscillations of $\delta$-variables and $d$ in temperate ice concerns their origin. Smoothed and degraded variations in $\delta^{18}O$ and $\delta^2H$ reflect the isotopic seasonality of precipitation. Variations in $d$ arise as a result of post-depositional processes **(Stichler et al., 1982)**. This is because during melting, the $d$ value of the residual solid fraction decreases due to ongoing isotopic fractionation **(Martinec et al., 1977)**, producing a seasonal post-depositional signal characterized by summer minima and winter maxima **(Stichler et al., 1982)**. Research on variations in $d$ in temperate ice has remained stagnant. Only recently were the results obtained at Vernagtferner replicated and confirmed in other contexts **(Zhou et al., 2014)**.

Subsequent work (e.g., **Yamada, 1987; Koerner, 1997; Yuanqing et al., 2001**) largely confirmed previous findings. An aspect that emerged is that the degree of degradation of isotopic signals in temperate glaciers is not always the same. Limited melt can allow annual layers to remain identifiable below the firn–ice transition **(Neff et al., 2012; Schwikowski et al., 2013)**, whereas strong melt obliterates signals within a few years. Another emerging trend is that glaciers once considered cold show isotopic alteration due to warming and melt, both in polar **(Spolaor et al., 2024)** and alpine contexts **(Clifford et al., 2023)**.

Recent advances have shifted toward hydrology. Water stable isotope analyses of river discharge are used to distinguish contributions from snow, rainfall, and groundwater. However, isotopic fractionation during melt prevents a direct comparison between the signature of the melting snowpack/ice and that of the resulting meltwater. To account for this, experimental and theoretical approaches have been developed **(Taylor et al., 2001; Zhou et al., 2014; Ham et al., 2019; Noor et al., 2023; Nyamgerel et al., 2024)**. A novelty of these works is the consideration of factors such as melting rate - where lower rates enhance isotopic fractionation **(Taylor et al., 2001; Noor et al., 2023)**- and firn saturation, with the maximum efficiency of fractionation observed in non-saturated firn **(Ham et al., 2019)**. But while such studies have advanced our understanding of isotopic processes during melting, they have not yet been systematically applied to temperate ice-cores for enhancing the interpretation of isotopic records.

### 4.7. Gas content

In cold glaciers, snow and firn contain air that is trapped in bubbles without significant gas fractionation, so their composition reflects the atmosphere at the time of entrapment, making them reliable archives. In temperate glaciers, however, ice forms mainly through the refreezing of meltwater-saturated firn, producing superimposed ice. Because prior to ice formation firn pores are filled with water rather than air, in temperate ice the air content is much lower than in cold ice. Ice-cores from temperate glaciers typically show compact, bubble-free layers alternating with strata containing variable bubble concentrations (**Coachman et al., 1958; Vallon et al., 1976**; Fig 2). The total air content of temperate glaciers is therefore linked to the frequency of melt events: the more frequent the events, the smaller the fraction of air entrapped within the ice. The relationship has been applied to reconstruct past summer temperatures in a an ice-core from the Himalaya, with encouraging results **(Hou et al., 2007)**. Scarce information about the vertical distribution of gases in temperate glaciers confirm that the deeper the ice, the lowest the air content due to the increasing degree of metamorphism and interaction with meltwater **(Hubbard et al., 2000)**.

After the initial descriptive studies on gas content in temperate ice **(Bader, 1950)**, it became clear that, differently from cold glaciers, gas composition in temperate ice does not reflect atmospheric air. Because atmospheric gases differ in solubility, refreezing in the presence of liquid water fractionates them, altering bubble composition **(Coachman et al., 1958)**. Such deviations can be used to distinguish ice not impacted by melting from pristine ice **(Hou et al., 2025)**. Additionally, the presence of liquid water enables englacial chemical reactions that consume and/or produce specific gases **(Coachman et al., 1956)**. Gas composition in temperate ice shows depleted $O_2$, $N_2$, and Ar but elevated $CO_2$ **(Coachman et al., 1956; Weiss et al., 1972)** and $CH_4$ **(Burns et al., 2018)**. $CO_2$ is particularly affected due to its high solubility in water and to the dissolution of carbonate mineral particles present in the ice **(Stauffer and Berner, 1978; Tranter et al., 2002)**. The decrease in gas content observed at temperate glaciers over depth is linked to the expulsion of gases at the bottom of the glacier, mostly via meltwater, with implications on subglacial weathering processes **(Hubbard et al., 2000)**.

Research on gases in temperate ice is not very active now, likely because they cannot be used for paleoclimatic purposes. Nonetheless, renewed interest in englacial biogeochemistry highlights the potential of gas analyses in temperate glaciers to shed light on englacial microbial activity and weathering processes **(Hubbard et al., 2000; Sharp & Tranter, 2017)**.

## 5. TEMPERATE ICE-CORE DRILLING TECHNIQUES

One of the earliest ice-cores ever drilled was extracted from a temperate glacier in 1950 **(Miller, 1954)**. That first attempt immediately revealed the main challenge of coring temperate ice: the presence of liquid water. The drilling relied on a rotary mechanical system adapted from geological prospecting, equipped with diamond blades that produced 5.4 cm diameter cores. The small barrel size made it difficult to recover intact firn sections, as the fragile material was often crushed, preventing the retrieval of stratigraphically continuous samples.

A key issue was the interaction of drill cuttings with liquid water. In cold ice, residues remain unconsolidated and can be removed between runs, but in temperate conditions they form a sticky slurry that can clamp the drill. At Taku Glacier, researchers attempted to flush cuttings by supplying water to the drill head. Although partially effective, this required casing the firn section and pumping large volumes of water from crevasses to maintain a stable borehole water level. Most drilling still had to proceed under dry conditions, resulting in slow progress and fractured cores. Despite these limitations, several cores were retrieved, the deepest reaching 90 m **(Miller, 1954)**.

To overcome these issues, thermal drills were developed. Instead of cutting, they melt a ring of ice around the core using a heated metallic ring powered by electric resistance. These systems have since yielded high-quality, unfractured cores from temperate glaciers (**Taylor, 1976; Schwikowski et al., 2014; Zagorodnov & Thompson, 2014**; and references therein). Their main drawback is the high power demand of the melting heads, which almost always requires combustion-based generators. Unlike electro-mechanical drills, which can be relatively easily powered by portable solar arrays, thermal drills cannot realistically rely on solar energy. This has major logistical implications, since transporting solar panels is considerably easier than moving large amounts of fuel and heavy generators. Only a few experimental tests with solar-powered thermal drills have been attempted **(Naftz et al., 1996)**.

Although thermal drills are now recognized as the most suitable technology for temperate glaciers, there remain contexts where electro-mechanical systems perform better. This is the case for glaciers with high concentrations of solid impurities, such as mineral dust or volcanic ash. In such layers, particles act as thermal insulators, hindering heat transfer and blocking thermal drill operations, a problem documented in Iceland **(Árnason et al., 1974)** and on Kilimanjaro **(Zagorodnov et al., 2002)**. Another situation in which electro-mechanical drills remain advantageous on temperate glaciers is during spring or early summer. At this time of the year, a layer of cold, low density snow and firn overlies temperate ice. Thermal systems can struggle to penetrate this material because their weight is insufficient to ensure a steady downward progress of the instrument. A practical solution is the use of modular drilling tools, which allow to begin with an electro-mechanical head and then switch to a thermal drill once temperate ice is reached **(Zagorodnov et al., 2002; Schwikowski et al., 2014)**.

## 6. WHAT HAVE TEMPERATE GLACIERS BEEN TELLING US IN 75 YEARS OF RESEARCH?

This section reviews the main projects that, from the 1950s onward, focused on ice coring in temperate glaciers (see also Fig 9 and Table 1). While interest in the properties of temperate ice predates this period, earlier efforts were largely limited to surface studies using snow pits or shallow trenches and are not considered.

### 6.1. The pioneering epoch

The first three successful deep ice-cores were drilled between 1950 and 1951 **(Langway, 2008)**, one of them at Taku Glacier in Alaska, the thickest temperate glacier on Earth **(Nolan et al., 1995)**. In summer 1950, as part of the Juneau Icefield Research Project, a core was retrieved from the glacier's accumulation basin. At the time, little was known about the preservation of paleoclimatic signals in glacier ice, and the main objective of these efforts was to test drilling methods. Even so, the Taku core provided valuable results. Technically, it offered the first lessons on drilling temperate ice, while scientifically, it yielded early data on the vertical evolution of ice texture and englacial temperature **(Miller, 1954)**. Despite these promising beginnings, research soon shifted toward cold glaciers, recognized as more suitable for paleoclimate reconstructions.

Interest in temperate sites resurfaced nearly 20 years later, when shallow cores from Icelandic glaciers were analyzed to explore the behavior of water stable isotopes in temperate glaciers **(Árnason, 1969)**. A few years later, two landmark drilling campaigns advanced the study of temperate glaciers. In 1971, a 180 m core was extracted from the Vallée Blanche plateau (Alps, France), followed in 1972 by a 65 m core from Blue Glacier (Rocky Mountains, U.S.A.).

At Vallée Blanche, research focused on the transformation of snow to temperate ice. The stratigraphy described by Vallon et al. (1976) offered the first comprehensive view of this transition, from cold winter snow to fully temperate firn and ice. A novel outcome was the identification of an internal water table: by measuring borehole water levels, the authors tracked its seasonal rise and fall in relation to surface melt and meteorological variability, providing pioneering understanding of temperate glacier hydrology.

The Blue Glacier project provided key insights into the microscale structure of temperate ice, describing grain size evolution, impurity and bubble distributions, and the geometry of liquid veins **(Harrison and Raymond, 1976; Raymond, 1976; Raymond and Harrison, 1975)**. These studies showed that recrystallization and continuous meltwater interactions modify physical and chemical properties with depth (see Fig 4b), laying the groundwork for conceptual and theoretical models about the physics of temperate glaciers.

### 6.2. Tropics

From the late 1970s, the lack of low-latitude ice-core records spurred drilling efforts on tropical glaciers. Cold glaciers in these regions are confined to the highest peaks, where logistics are challenging, while most tropical glaciers are fully temperate, even in their accumulation zones. As a result, tropical temperate glaciers have received more and earlier attention than in other regions.

Since 1974, a series of campaigns were launched atop the Quelccaya Ice Cap in the Peruvian Andes. Several ice-cores have been extracted since then to bedrock, drilling about 160 m of ice. Initially the glacier was polythermal, being temperate in the upper tens of meters and cold below, allowing the preservation of seasonal markers such as dust, radioactive tracers, and stable isotopes. This enabled the reconstruction of ~1,500 years of Andean climate variability **(Thompson et al., 1979, 1985)** and trace element pollution **(Uglietti et al., 2015)**. Today, however, enhanced percolation and ice warming are threating Quelccaya ice as paleoclimatic archive **(Thompson et al., 1993; Clifford et al., 2023)**.

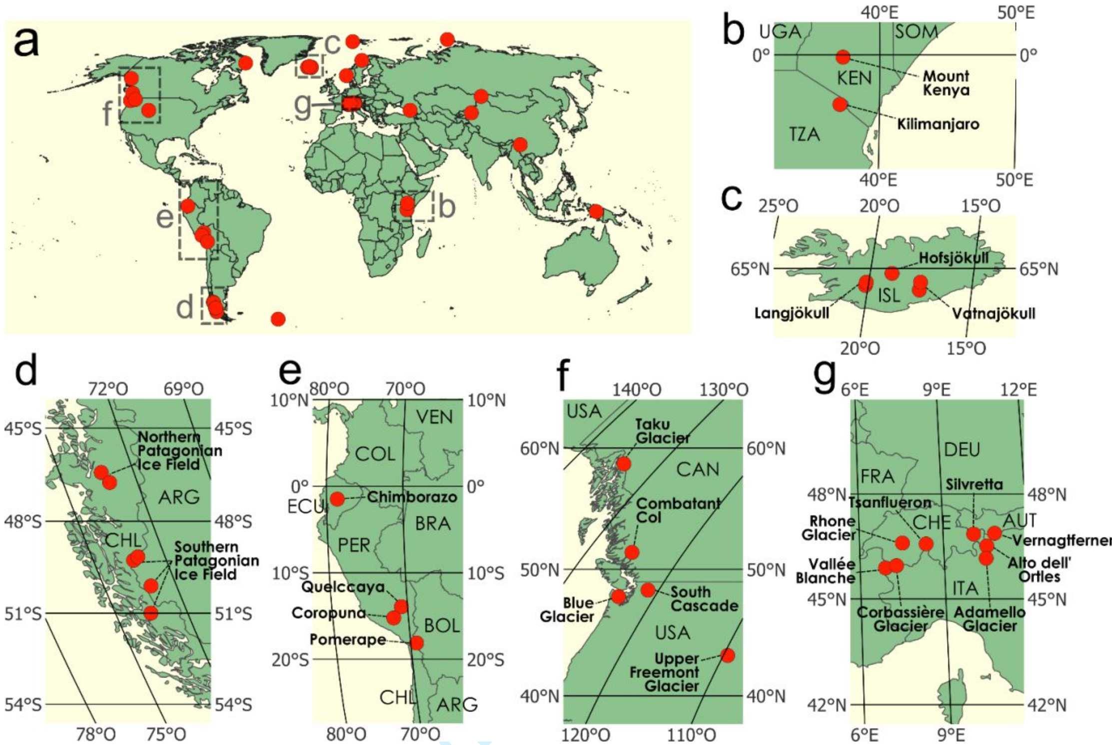


**Fig 9 The geographic distribution of the temperate ice-core drilling projects discussed in this work.**

In 1999-2000, two cores from Chimborazo (Ecuador) reached 54 m. The 1999 ice was likely cold but close to temperate, whereas by 2000 volcanic ash from the Tungurahua eruption starting in October 1999 had darkened the glacier surface, warming the firn and increasing meltwater percolation **(Schotterer et al., 2003)**. Comparing the 1999 and 2000 ice-core records, it was possible to investigate the migration and redistribution of ionic species (mostly volcanic-derived mineral acids) within the ice column, providing an unrepeatable opportunity to observe live meltwater-related post-depositional processes **(Ginot et al., 2010)**. At Nevado Coropuna (Peru), cores up to 163 m were recovered in 2003. Heavy melting severely degraded most proxies, though impurity-rich layers still recorded the impact of positive ENSO phases, and some signals from pollen and diatoms were preserved **(Herreros et al., 2009; Weide et al., 2017)**. A shallow core from Pomerape (Bolivia) produced no usable data due to excessive melting **(Vimeux et al., 2009)**.

The first attempt on African glaciers dates to 1977, when a shallow core (<10 m) was drilled on Kilimanjaro (Tanzania). Aside from noting successful drilling **(Davies et al., 1977)**, no results were published. The glaciers were revisited in 2000, yielding Holocene-spanning records (~12,000 years; **Thompson et al., 2002**). Radiocarbon ages for organic fragments allowed to establish the ice-core chronology. While early reports suggested temperate conditions **(Davies et al., 1977)**, no meltwater was found during the 2000 drilling **(Thompson et al., 2002)**. Later work confirmed temperate conditions **(Kaser et al., 2010; Yoshikawa et al., 2021)**, with discrepancies likely reflecting cooling and thermal instability in small, thin ice bodies once firn cover is lost **(Huss and Fischer, 2016)**. Attempts on Mount Kenya (Kenya) produced no usable records due to massive percolation and loss of annual accumulation **(Thompson & Hastenrath, 1981)**. Elsewhere, a 2010 core from Puncak Jaya (Indonesia) provided evidence that in recent years climatic signals present in tropical temperate ice-cores are irreparably degrading **(Permana et al., 2019)**.

Overall, tropical glaciers contributed significantly to the early development of temperate ice-core science. Yet rising equilibrium-line altitudes and intensified meltwater percolation are rapidly erasing seasonal stratigraphy. Within only a few decades, several once-promising sites have become largely unsuitable for paleoclimate reconstruction **(Thompson et al., 2021b)**.

### 6.3. Mid-latitudes

Unlike the tropics, many mid-latitude mountain ranges still harbor cold accumulation zones, allowing extraction of cold ice-cores. Temperate ice coring has therefore focused mainly on regions lacking cold glaciers (e.g., Patagonia, non-polar North America), though other areas such as the Alps and maritime sub-polar regions have also been targeted for reasons of accessibility or local scarcity of cold ice.

*Patagonia*

The Northern and Southern Patagonian Ice Fields form the largest mid-latitude ice masses. Both are fully temperate **(Lliboutry, 1956; Warren and and Sugden, 1993)**. Early drilling on San Rafael Glacier in 1985 (Northern Ice Field, 1300 m a.s.l.) confirmed the temperate regime, with most seasonal signals erased and only shallow residual features usable for chronology **(Yamada, 1987)**. Density discontinuities, possibly marking former summer surfaces, were the only stratigraphic features usable to derive a tentative chronology. Later work showed that although firn at ~1500 m a.s.l. was temperate, the highest plateaus above 2000 m still preserve paleoclimatic signals **(Matsuoka and and Naruse, 1999; Vimeux et al., 2008)**. In the Southern Ice Field, shallow cores collected in 1986 and 1999 revealed partial preservation of water stable isotope records and seasonal distribution of snow algae, enabling the identification of annual layers and revealing extremely high accumulation rates (~15 m w.e. $yr^{-1}$; **Aristarain & Delmas, 1993; Shiraiwa et al., 2002; Kohshima et al., 2007**). Additional cores from Pio XI Glacier showed that isotopic stratigraphy could survive under near-temperate conditions, provided no water table developed **(Schwikowski et al., 2013)**. Overall, results indicate that the highest sectors of both fields may still preserve (or might have preserved) readable paleoclimatic information, though no deep drilling has yet been carried out.

*Non-polar North America*

In non-polar North America, the Upper Fremont Glacier (Wyoming, US) yielded a 160 m temperate ice-core in 1991 that provided a chronology from ~1770–1991 CE, capturing the water stable isotope signal of the Little Ice Age and recent warming trends **(Naftz et al., 2002, 1996)**. A second core drilled in 1998 preserved mercury records, enabling the reconstruction of detailed atmospheric deposition histories for this anthropogenic-related species **(Schuster et al., 2002)**. Later work refined the age model using local tree-ring chronologies **(Chellman et al., 2017)**. A 141 m core extracted from Combatant Col in the Canadian Rockies in 2011 revealed high accumulation (>4 m w.e. $yr^{-1}$), with limited meltwater percolation to the depth of the annual firn layer and preserved seasonal signals of metals, dust, and black carbon back to the early 1970s **(Neff et al., 2012)**. In 1994, a 158-meter core was extracted from the temperate South Cascade Glacier (Washington, US) to focus on the role of light-absorbing particles in melt processes and mass balance **(Kaspari et al., 2020)**. The core was dated using multiple proxies, including $^{210}Pb$ and $^{3}H$ peaks, annual layer counting, volcanic tephra and mass balance records. The resulting chronology spans from 1840 to 1991 CE **(Kaspari et al., 2020)**.

*Alps*

The Alps have hosted several important temperate ice-core projects. After Vallon et al. (1976) drilled on Mont Blanc, attention focused on Vernagtferner in Austria, where cores extracted between 1976 and 1984 (up to 81 m) advanced the understanding of water stable isotopes and radionuclide behaviour in temperate ice **(Drost and Hofreiter, 1982; Oerter and Rauert, 1982; Von Gunten et al., 1982; Stichler et al., 1982; Baker et al., 1985; Oerter et al., 1985)**. At Silvretta Glacier (Switzerland), a 101 m temperate core retrieved in 2011 provided first direct evidence of secondary release of organic pollutants by temperate glaciers **(Pavlova et al., 2016, 2014)**. The ice-core was dated through nuclear dating ($^{210}Pb$ and $^{3}H$), annual layer counting (black carbon) and glaciological data (annual mass balance). At Alto dell'Ortles glacier (Italy), coring in 2011 to 75 m revealed polythermal conditions: cold deeper layers (below 30 m) yielded a 7000-year record (dated by radiocarbon), while the upper temperate portion was dated by means of Fukushima radioactive fallout, pollen, and glaciological modeling **(Festi et al., 2015; Gabrielli et al., 2016; Carturan et al., 2025)**.

More recently, Corbassière Glacier (Switzerland) and Adamello Glacier (Italy) showed how fast signal degradation can occur at high altitudes. At Corbassière, shallow cores collected in 2018 and 2020 revealed a surprising transition to temperate conditions above 4000 m in a few years **(Huber et al., 2024, 2025)**. At Adamello glacier several cores were obtained in the former accumulation basin of the glacier, where the mass balance has been negative since the 2010s. The deepest core reached 224 m. Results confirmed the deterioration of paleoclimatic signals due to the complete removal of firn and pervasive meltwater percolation, though proxies such as pollen, black carbon, $^3$H, and environmental DNA survived at least partially and allowed a chronology to be developed for the upper ~50 m of the core **(Festi et al., 2021; Varotto et al., 2021; Maggi et al., 2023; Di Stefano et al., 2024)**. The Adamello glacier is an interesting site for testing the preservation of paleoclimatic records in conditions particularly adverse to the preservation of stratigraphic signals, where the most recent record has already been lost. The Alps also remain unique in utilizing temperate ice-cores for structural and rheological studies. Drilling at Tsanfleuron Glacier in the 1990s and Rhône Glacier more recently has provided insights into ice fabrics, debris content, and flow properties under temperate conditions **(Hubbard et al., 2000; Hubbard et al., 2003; Tison & Hubbard, 2000; Hellmann et al., 2021)**.

*Central Asia*

In Central Asia, efforts began later, largely because cold glaciers dominate the region. The Djantugan Ice Plateau in the Caucasus was cored in the 1980s, with the deepest ice-core (93 m) allowing mass balance reconstructions since the 1930s despite the temperate regime **(Popovnin, 1999)**. In China's Yulong glaciers (the southernmost glacierized region of Eurasia), shallow cores showed smoothing of the water stable isotope signal owing to the occurrence of a water table, preventing further drilling **(Yuanqing et al., 2001)**. Despite not being fully temperate, some drilling activities were carried out on Sofiyskiy glacier (Russian Altai), heavily impacted by summer melting. Pollen and algae provided seasonal signals despite melt, highlighting the potential of coarse biological markers to date melt-affected ice **(Nakazawa et al., 2004; Uetake et al., 2006)**. Finally, the Grigoriev Ice Cap in the Tien Shan, long considered cold, now shows increasing melt effects, with water stable isotopes and ion-related signals degraded but dust and biological proxies still informative **(Machguth et al., 2024)**.

### 6.4. Subpolar and polar regions

*Scandinavia*

To investigate the early-melt "ionic pulse," a 61 m temperate ice-core was drilled in 1980 at Folgefonna Ice Cap (southern Norway). The core clarified how ions partition between solid and liquid phases in temperate ice, underpinning elution sequences **(Davies et al., 1982)**. In 1996, an exploratory core (34 m) from the small polythermal Riukojietna ice cap (Sweden; bedrock ~105 m) assessed the potential of subpolar glaciers for paleoclimatic constructions. Stratigraphy showed sharp discontinuities between facies, attributed to intervals of negative mass balance: bubble-poor firn/superimposed ice overlying bubble-richer ice likely formed under colder, drier conditions (**Pohjola et al., 2005**). Records of major ions and water stable isotopes helped to develop an interpretation according to which ice formed during the 20th century overlies ice dated to the Little Ice Age with a stratigraphic gap, but chronological control was limited.

*Iceland*

Despite hosting Europe's largest ice masses, Iceland has seen relatively few ice-core studies due to the absence of cold glaciers and proximity to Svalbard/Greenland. After early work by Arnason (1969), a 415 m core on western Vatnajökull primarily served drill development **(Árnason et al., 1974)**. On Hofsjökull, at least two cores were recovered (longest 101 m), where dust-rich (summer melt) layers enabled dating of the shallow section and offered preliminary insight into signal preservation (**Thorsteinsson et al., 2002** and references therein).

*Svalbard*

The thermal structure of Svalbard glaciers is complex, with most glaciers being polythermal. They are typically characterized by a shallow surface layer of cold ice, which is thinnest -or even absent- in the upper accumulation zones and thickens toward the terminus. Additionally, the largest and thickest glaciers are believed to host a basal layer of temperate ice **(Sevestre et al., 2015)**. Moreover, seasonal surface melt promotes percolation/refreezing that significantly modifies the ice stratigraphy. From the 1960s to late 1990s, Soviet and Japanese teams drilled multiple sites (e.g., Lomonosovfonna, Austfonna, Vestfonna, Høghetta Ice Dome, Grønfjord), retrieving cores up to 565 m deep (Koerner, 1997; Zagorodnov, 1998; and references therein). Visual stratigraphy showed substantial superimposed ice (≈34% at Lomonosovfonna to near-100% at lower-elevation Austfonna/Vestfonna; **Fujii et al., 1990; Zagorodnov, 1998**), consistent with temperate/polythermal regimes, though direct temperature profiles are lacking. Early Svalbard cores yielded multi-millennial records (Fujii et al., 1990; Koerner, 1997; Zagorodnov, 1998), but their reliability was questioned due to strong melt and unrecognized negative mass-balance episodes **(Vaykmyae et al., 1985; Koerner, 1997; Iizuka et al., 2002)**. Later work at the highest, coldest sites showed better preservation of water stable isotopes, major ions, and radionuclides **(Pohjola et al., 2002; Pinglot et al., 2003; Moore et al., 2005; Wel et al., 2011; Wendl et al., 2015; Osmont et al., 2018)**, shifting attention toward cold-regime targets. Recent observations, however, indicate growing vulnerability even at cold sites, in particular because of firn warming related to latent-heat input from percolation/refreezing. At Holtedahlfonna, temperatures in 2005 were below melting throughout the column but reached −0.4 °C at 15 m **(Beaudon et al., 2013)**. Four shallow cores were drilled between 2012 and 2019 at the same glacier. While in 2012 seasonal isotopic cycles were clearly visible, they became undetectable by 2019 due to enhanced summer melt **(Spolaor et al., 2024)**, consistent with expanding firn water tables on major Svalbard glaciers **(van den Akker et al., 2025)**.

*Russian Arctic*

Glaciers in Franz Josef Land and Severnaya Zemlya share Svalbard-like thermal structures. Drilling campaigns recovered several long (> 100 m) cores and multi-millennial records from these regions (Kotlyakov et al., 2004; Opel et al., 2013 and references therein). Although in these ice-cores summer melt has been identified to deteriorate paleoclimatic signals, analyses of water stable isotopes, major ions, heavy metals and radionuclides suggest that the impact is moderate, allowing the partial preservation of records **(Koerner, 1997; Pinglot et al., 2003; Fritzsche et al., 2005; McConnell et al., 2019)**. In Severnaya Zemlya, redistribution of mobile ions was leveraged as a dating tool: seasonal melt–refreeze cycles produced ion oscillations that, combined with density and visual stratigraphy, resolved annual layers **(Weiler et al., 2005)**. Unlike Svalbard, recent assessments of how ongoing warming is transforming Russian Arctic firn/ice archives remain scarce.

*Canada*

On Baffin Island, Penny Ice Cap is a key site for melt-affected archives. Although still glaciologically cold, its firn is warming toward the melting point, and summer surface melt has been pervasive throughout the Holocene. Consequently, numerous cores from Penny Ice Cap have served as benchmarks for reconstructing multimillennial records from melt-impacted ice (**Zdanowicz et al., 2012** and references therein). Similar trends toward temperate conditions are emerging across Canadian Arctic ice caps, at least in shallow layers **(Fisher et al., 2012)**, threatening archives that in many cases extend to the last glacial period.

*Southern Ocean*

South Georgia is largely glacierized by temperate ice. In 2015, the first ice-core drilling attempt was conducted on a glaciated plateau of the main island, at approximately 850 m a.s.l.. A 15-meter long firn core was retrieved to assess the preservation of glaciochemical signals **(Potocki et al., 2025)**. Stable water isotopes showed a few seasonal oscillations in the upper ~6 m but became homogeneous below. Trace elements likewise declined to near-constant values at the same depth. The site was thus deemed unsuitable for deeper drilling. Higher, colder plateaus might preserve signals better but remain unexplored.

| Glacier | Location | Year | Depth (m) | Investigated Proxies | Ref. |
|---|---|---|---|---|---|
| Kilimanjaro | Tanzania | 1977; 2000 | Up to 51 | Water stable isotopes, radiocarbon, radionuclides, visual stratigraphy, | Davies et al. (1977); Thompson et al. (2002) |
| Mount Kenya | Kenya | 1978 | Up to 13 | Radionuclides, insoluble particles, water stable isotopes, density | Thompson & Hastenrath (1981) |
| Quelccaya | Peru | 1974-2018 | Up to 164 | Water stable isotopes, visual stratigraphy, conductivity, radionuclides, insoluble particles, trace elements | Thompson et al. (1979, 1985, 1986); Clifford et al. (2023) |
| Chimborazo | Equador | 1999; 2000 | Up to 54 | Water stable isotopes, ionic species | Schotterer et al. (2004); Ginot et al. (2010) |
| Nevado Coropuna | Peru | 2003 | Up to 163 | Water stable isotopes, radionuclides, ionic species | Herreros et al. (2009) |
| San Rafael Glacier | Patagonia | 1985 | 39 | Visual stratigraphy, water stable isotopes, pH, conductivity | Yamada (1987) |
| Glaciar Nef | Patagonia | 1996 | 14.5 | Visual stratigraphy, water stable isotopes, ice grain size | Matsuoka & and Naruse (1999) |
| Southern Patagonia ice cap | Patagonia | 1986 | 13 | Visual stratigraphy, water stable isotopes, ionic species | Aristarain & Delmas (1993) |
| Glaciar Tyndall | Patagonia | 1999 | 46 | Visual stratigraphy, water stable isotopes, ionic species, snow algae | (Shiraiwa et al., 2002; Kohshima et al., 2007) |
| Gorra Blanca Norte | Patagonia | 2001 | Up to 5 m | Water stable isotopes, ionic species | Schwikowski et al. (2006) |
| Pío XI glacier | Patagonia | 2006 | 50 | Visual stratigraphy, water stable isotopes, ice temperature, ionic species | Schwikowski et al. (2013) |
| Puncak Jaya | Indonesia | 2010 | 32 | Water stable isotopes, insoluble particles, ionic species | Permana et al. (2019) |
| Blue Glacier | Olympic Mts. | 1972 | 65 | Visual stratigraphy, microscopic observations, ice temperature, crystalline texture and structure | Raymond & Harrison (1975); Raymond (1976); Harrison & Raymond (1976) |

| | | | | | |
|---|---|---|---|---|---|
| Upper Freemont Glacier | Rocky Mts. | 1991; 1998 | 160 | Water stable isotopes, ionic species, mercury | Naftz et al. (1996, 2002); Schuster et al. (2002) |
| Combatant Col | Coast Mts. | 2011 | 141 | Water stable isotopes, black carbon, lead, insoluble particles | (Neff et al., 2012) |
| South Cascade Glacier | North Cascades | 1994 | 158 | Black carbon, insoluble particles, organic content, water stable isotopes, radionuclides | Kaspari et al. (2020) |
| Vallée Blanche | European Alps | 1971 | 180 | Visual stratigraphy, physical properties, crystalline texture and structure, liquid water content | Vallon et al. (1976) |
| Vernagtferner | European Alps | 1976-1984 | Up to 81 | Visual stratigraphy, water stable isotopes, radionuclides | Oerter & Rauert (1982); Von Gunten et al. (1982); Oerter et al. (1985); Baker et al. (1985) |
| Silvretta Glacier | European Alps | 2011 | 101 | Black carbon, radionuclides, polychlorinated biphenyls | Pavlova et al. (2014) |
| Alto dell'Ortles | European Alps | 2011 | Up to 75 | Water stable isotopes, radionuclides, trace elements, pollen | Gabrielli et al. (2016) |
| Corbassière Glacier | European Alps | 2018; 2020; 2025 | Up to 100 | Water stable isotopes, ionic species | Huber et al, (2024) |
| Adamello Glacier | European Alps | 2015-2021 | Up to 224 | Water stable isotopes, black carbon, ionic species, trace elements, pollen, radionuclides | Festi et al. (2021); Maggi et al. (2023); Di Stefano et al. (2024) |
| Tsanfleuron Glacier | European Alps | 1996; 1997 | Up to 45 | Visual stratigraphy, crystalline texture and structure | Hubbard et al. (2000); Tison & Hubbard (2000) |
| Rhone Glacier | European Alps | 2017 | 80 | Crystalline texture and structure | Hellmann et al. (2021) |
| Lexyr Glacier | Caucasus | 1980s | Up to 93 | Visual stratigraphy, water stable isotopes | (Popovnin, 1999) |
| Baishui No. 1 Glacier | Jade Dragon Snow Mts. | 1999 | 10 | Water stable isotopes, pH, ionic species | Yuanqing et al. (2001) |

| | | | | | |
|---|---|---|---|---|---|
| Sofiyskiy glacier | Altai | 2001 | Up to 25 | Water stable isotopes, ionic species, pollen, fungi, bacteria | Nakazawa et al, (2004; Uetake et al. (2006) |
| Grigoriev Ice Cap | Tien Shan | 1960s-2018 | Up to 87 | Water stable isotopes, radionuclides, ionic species, black carbon, ice temperature | Machguth et al. (2024) |
| Folgefonna | Scandinavia | 1980 | 61 | Ionic species, pH | Davies et al, (1982) |
| Riukojietna | Scandinavia | 1996 | 34 | Visual stratigraphy, water stable isotopes, conductivity, ionic species | Pohjola et al. (2005) |
| Taku Glacier | Alaska | 1950 | 90 | Visual stratigraphy, ice temperature, ionic impurities (attempt) | Miller (1954) |
| Icelandic glaciers | Iceland | 1967-68 | Up to 30 | Water stable isotopes | Arnason (1969) |
| Svalbard glaciers | Svalbard | 1960s-201 | Up to 565 | Water stable isotopes, radionuclides, ionic species, visual stratigraphy, black carbon | Koerner (1997); Pohjola et al. (2002); Pinglot et al. (2003); Moore et al. (2005); Spolaor et al. (2024) |
| Akademii Nauk ice cap | Severnaya Zemlya | 1999-2001 | 724 | Water stable isotopes, ionic species, radionuclides | Weiler et al. (2005); Opel et al. (2013) |
| Penny ice cap | Baffin Island | 1953-2011 | Up to 334 | Visual stratigraphy, water stable isotopes, ice temperature | Zdanowicz et al., (2012) |
| Briggs Glacier | South Georgia | 2015 | 15 | Water stable isotopes, trace elements, ice temperature | Potocki et al. (2025) |

**Table 1 Information about the major ice-core drilling projects involving temperate glaciers.**

### 6.5. CONCLUSIONS AND PERSPECTIVES

At the dawn of ice-core science, temperate and cold glaciers received similar attention, with several projects launched in the 1950s. It soon became clear, however, that cold glaciers offered far more reliable paleoclimatic archives. Two melt-related processes make temperate glaciers especially challenging: (1) part of the annual snow accumulation is lost to melt, introducing stratigraphic discontinuities; and (2) the production, persistence, and percolation of meltwater deteriorate most proxies. Virtually all proxies in temperate ice are subject to smoothing, fractionation, and remobilization, with insoluble impurities being the least affected.

Once the impairment of glaciochemical records in temperate ice was recognized, interest in ice-cores from temperate glaciers declined. The few studies conducted on temperate ice-cores between the 1960s and 1980s focused primarily on improving our understanding of the chemical and physical processes typical of temperate glaciers rather than on palaeoclimatic reconstruction. In recent decades, however, climate change has renewed the interest for them. Many cold glaciers are increasingly affected by melting events and their temperature is approaching the transition threshold to a temperate regime, threatening the integrity of the records they have preserved until now. To keep exploiting glaciers as paleoclimatic archives, we need to improve our ability to read deteriorated proxies in ice-cores drilled at temperate glaciers.

This review is intended to offer a synthesis of processes affecting temperate ice-cores. This will hopefully serve as a starting point for future research. While qualitative knowledge of meltwater impacts is already substantial, quantitative approaches to reconstruct residual paleoclimatic signals from temperate glaciers and disentangle them from postdepositional effects remain limited. Melting and meltwater percolation cause loss of information. To enable the best possible utilization of the residual signals of temperate ice, it is necessary to complete the missing information by drawing from other sources. Available examples have used meteorological data, data from other paleoclimate archives (such as tree rings), and glaciological mass balance data. The combination of information from multiple sources allow to calibrate the impact of post-depositional processes on proxies and to consider periods with negative mass balance, otherwise hardly detectable.

Future efforts should thus focus on refining quantitative methods for integrating the impact of melt-related post-depositional processes into the analysis and interpretation of signals from temperate ice-cores. As glaciers progressively warm, temperate ice will become in many regions the primary, if imperfect, repositories of glaciological climate information, making their understanding essential for the future of ice-core science.

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## 8. AUTHOR CONTRIBUTION

GB conceived the idea of this work and wrote the manuscript with contributions from all authors (FB, AE, TJ, MS).