**SUPPLEMENTARY MATERIAL**

**A. Several examples of radargrams.**

**Fig. S1:** Example of radargram obtained at 250 MHz along a transect (see insert map) with:

a) manual picking shown for end-of-summer 2018 (orange) and 2015 (red) horizons. The blue squares on the top bar and on the insert map correspond to the points precisely measured by differential GNSS. Time-to-depth conversion calculated here for a wave propagation speed of 0.215 m ns-1;

b) same radargram, with simple contrast treatment, without picking to allow the reader to view the raw signal;

c) same radargram with additional energy decay, background removal and deconvolution processes;

d) same radargram with additional energy decay, background removal, deconvolution and dewow processes.

**Fig. S2:** Example of radargram obtained at 250 MHz along a transect (see insert map) with:

a) manual picking shown for end-of-summer 2018 (orange) and 2015 (red) horizons. The blue squares on the top bar and on the insert map correspond to the points precisely measured by differential GNSS. Time-to-depth conversion calculated here for a wave propagation speed of 0.215 m ns-1;

b) same radargram, with simple contrast treatment, without picking to allow the reader to view the raw signal;

c) same radargram with additional energy decay, background removal and deconvolution processes;

d) same radargram with additional energy decay, background removal, deconvolution and dewow processes.

**Fig. S3:** Example of radargram obtained at 250 MHz along a transect (see insert map) with:

a) manual picking shown for end-of-summer 2018 (orange) and 2015 (red) horizons. The dashed circle indicates an area where horizon picking was hardly possible. The blue squares on the top bar and on the insert map correspond to the points precisely measured by differential GNSS. Time-to-depth conversion calculated here for a wave propagation speed of 0.215 m ns-1;

b) same radargram, with simple contrast treatment, without picking to allow the reader to view the raw signal;

c) same radargram with additional energy decay, background removal and deconvolution processes;

d) same radargram with additional energy decay, background removal, deconvolution and dewow processes.

**B. Multi-element chemical analyses of *S2* and *F2* firn cores.**

Firn cores chemistry can assist surface mass balance studies in mountains areas. As the chemical composition of summer and winter snow layers is generally different, the chemistry can allow to identify past buried horizons (see for example Preunkert and others, 2000; Preunkert and others, 2001; Eichler and others, 2001; Ginot and others, 2010; Sold and others 2015).

The *S2* and *F2* firn cores were decontaminated in the IGE cold rooms by removing the external part using a band saw. Then they were continuously sampled and stored frozen in pre-washed (ultrapure water) polyethylene flasks at a resolution of 10 cm and 5 cm for the 21 m and 8 m cores respectively (a total of 211 and 155 samples, respectively). Each sample was analysed for major ionic composition (cations: Na+, NH4+, K+, Mg2+, Ca2+; anions: CH3COO-, HCOO-, Cl-, NO3-, SO42-, C2O42-) using Ion Chromatography (IC) at the IGE laboratory using a Dionex© ICS3000 dual ion chromatography system instrument equipped with AS11-HC and CS12 columns down to sub-ppb level and with a high level of accuracy (6 standard calibrations, relative s.d. < 2%).

Fig. S4 and S5 showcase the 21 m *S2* core (S4) and the 8 m *F2* core (S5) icy and dirty layer results. Also included are the density and analyses of the 11 chemical ionic species (nitrate NO3-, sulphate SO42-, ammonium NH4+, acetate, CH3CO2-; formate, HCO2-; chloride, Cl-; oxalate, C2O42-; sodium, Na+; potassium, K+; magnesium, Mg2+; calcium; Ca2+).

**Fig. S4:** Density and chemical ionic species (nitrate, sulfate, ammonium, formate, acetate, chloride, oxalate, sodium, potassium, magnesium, calcium) profiles from the *S2* core with icy and dirty layers highlighted in grey and red respectively. Density and chemical measurements do not start at the surface because, due to the inconsistency of the fresh snow, the first 40 cm of snow were not sampled.

Previous ice-core studies conducted at the *Col du Dôme* pass (4 250 m a.s.l.), a nearby site 700 m greater in altitude, ~4.5 km southwest of the *Col du Midi* pass, have shown that a clear seasonal cycle is preserved in snow layers for most of the chemical species, with levels higher in summer than in winter (Preunkert and others, 2000). In particular, the authors revealed mean summer to winter ratios close to 4 for nitrate (NO3-) and sulfate (SO42-), and as high as 14 for ammonium (NH4+). These seasonal variations originate mainly from the more limited upward transport of air masses from the boundary layer in winter than in summer (Lugauer and others, 1998; Preunkert and others, 2001). Given that Preunkert and others (2001) observed such ratios at atmospheric aerosol monitoring sites above 1 250 m a.s.l., a large summer to winter ratio would presumably be recorded in the *Col du Midi* snow layers as well, although the glacier is known to be temperate at that site (Vallon and others, 1976).

In the *S2* core, from the surface down to a depth of 4.3 m, the snowpack corresponded to the fresh winter snow accumulated since the end-of-summer 2018 (see Section 5.2). The density increased with the depth, from 260 to 430 kg m-3, and the mean value of the ion concentrations was 76.9±61.0 (s.d.), 27.1±14.2 and 12.0±13.9 ppb for NO3-, SO42- and NH4+, respectively. The sample between 4.32 and 4.42 m was far more concentrated with 661.1, 294.9 and 359.7 ppb, respectively. This corresponds to ratios of 8.6, 10.9 and 30 for NO3-, SO42- and NH4+ respectively, compared to the mean levels in the first 4.3 m, i.e. higher than those recorded at the *Col du Dôme* pass by Preunkert and others (2000) between summer and winter snows. This snow layer, between 4.32 m and 4.42 m, was characterized by coarse grains with a 1 cm ice layer at the bottom, and clearly corresponded to the end-of-summer 2018 surface layer (Table 2). Chemical concentrations of several species were therefore higher in the end-of-summer snow layers than in the winter layers, as expected from the *Col du Dôme* ice core studies (Preunkert and others, 2000). We observed a secondary peak at a depth of 3.8 m for chemical species (particularly for nitrate), but without any clear change in the snow grain size. This might be explained by residual convection of air masses that can occur during warm periods in autumn, when the summer snow layer has already been covered by fresh snow whose chemical characteristics are those of precipitation coming from the free troposphere, undisturbed by advections from the boundary layer.

In contrast, below the 4.4 m end-of-summer 2018 layer, no clear seasonal peak could be easily detected in the chemical profiles. The measured concentrations were quite stable between 4.5 and 21.5 m with mean values (± s.d.) of 27.8±25.5, 10.5±16.4 and 30.4±17.2 ppb for NO3-, SO42- and NH4+ respectively. For nitrate and sulfate, these mean values were lower than the winter concentrations of the upper snow layers (0 to 4.3 m) and quite constant below 7 m. Ammonium levels were greater than the 2018-2019 winter values with greater variability than for nitrate and sulfate.

There were many ice layers below 4.4 m deep (Fig. S4). These ice layers had a thickness ranging from 1 cm to 13 cm. The presence of a large number of ice layers (41 between the depths of 4.4 m and 21.5 m) clearly indicates melting occurs during the summer at that site and that the meltwater percolates into the firn and re-freezes at deeper levels. Moreover, at the *Col du Midi* pass, recent summer precipitation includes rain (M. Marcuzzi, caretaker of the *Refuge des Cosmiques*, see Fig. 1, personal communication, 2020), which also percolates into the firn. This consequently redistributes the high ionic content of summer precipitation to the lower winter snow layers.

Other studies in the Alps (Eichler and others, 2001) or the Andes (Ginot and others, 2010) report better conservation of the annual profile of nitrate and ammonium compared to sulfate when melting occurs in the firn. In a study of the Findelengletscher, which is also a temperate glacier, Sold and others (2015) were able to identify four annual layers based on chemical analyses of ammonium and δD despite summer melting. In our case, the ionic chemical analyses were not sufficient to clearly identify the past end-of-summer horizons.

**Fig. S5:** Density and chemical ionic species (nitrate, sulfate, ammonium, formate, acetate, chloride, oxalate, sodium, potassium, magnesium, calcium) profiles from the *F2* core with icy and dirty layers highlighted in grey and red respectively. Inconsistencies with fresh snow were also an issue at this site; sampling in the upper 0.2 m of the core was omitted from the analysis.

**C. Equivalence of the submergence velocities measured from GPR measurements and with the stake method.**

Fig. 2 in the main article shows the principle of the method used to calculate the spatial distribution of the surface mass-balance from equation (2). This method, using GPR measurements, is an Eulerian approach that calculates the submergence velocity for fixed points, independent of time. It is similar to the Core method discussed by Réveillet and other (2020). Here we add information about the horizontal flow of the glacier to discuss how it affects the calculation (Fig. S6) and show its equivalence with the stake method (Fig. S7). The elevation of an end-of-summer surface was measured from LiDAR acquisition (“Surface year *N*” in Fig. S6). The elevation of the same layer was then measured after some years using GPR measurements (“Buried horizon of the surface in year *N* at the date *N+i*” in Fig. S6). The elevation difference between the two dates represents the submergence velocity. As discussed in Section 6.2, the submergence velocities determined in our study are spatially integrated values and not purely local values because of the glacier flow (Fig. S6 and Fig. S7).

**Fig. S6:** Schematic representation of the key parameters for estimating the surface mass-balance by the method proposed in this paper. It is similar to Fig. 2 (upper panel) in the main article but include aspects of horizontal displacement. From top to bottom, the three lines correspond to the glacier surface in a reference year *N* (brown line), the surface in year *N+i* (orange line), and the horizon of the year *N* surface layer, moved downward under the effect of submergence, at the date *N+i* (red line).

An enlargement of Fig. S6, with an exaggeration of the surface slope, corresponding to the horizontal distance travelled in *i* years by the glacier is schematised in Fig. S7. The displacement of an accumulation stake between years *N* and *N+i* is also visualized on Fig. S7.

**Fig. S7:** Detailed schematic representation of the principle of the method used to determine submergence velocities from GPR measurements and the comparison with the stake method (Réveillet and others, 2020). From top to bottom, the three lines correspond to the glacier surface in a reference year *N* (brown line), the surface in year *N+i* (orange line), and the horizon of the year *N* surface layer, moved downward under the effect of submergence, at the date *N+i*. The flow line is represented by a black arrow, horizontal and vertical velocities by dashed grey arrows. The thick brown feature represents a year *N* accumulation stake (top left) and its new position after *i* years (bottom right).

An alternative method to determine the submergence velocities uses repetitive measurement at stakes (Réveillet and others, 2020). This method is a Lagrangian approach that follows the location of a point and is time dependent. This method involves measuring the locations of the tops of a stake using differential GNSS measurements. For each measurement, the slope of the surface is calculated from the LiDAR-derived DEM. The submergence velocity (*Vsub*) is obtained from the term (Cuffey and Patterson, 2010, p. 332).

The differences in methods have been explained in detail in Réveillet and others (2020). In both methods, we do not ignore the horizontal displacements. In the GPR method, we infer the submergence velocity directly. In the stake method, we use the vertical velocity *Ws*, the slope of the surface and the horizontal velocities to calculate the submergence velocity. If the surface at year *N* (brown dashed line on Fig. S7) and the same buried surface at year *N+i* (red dashed line in Fig. S7) are parallel, both methods give the same result. The result could differ if the spatial variations of the vertical velocity were large. It is not the case in our study site. In both methods, the submergence velocities determined are spatially integrated values and not purely local values. They take into account the displacement of the glacier over the *i*-year period and correspond to a mean value between the considered point and a few metres upstream.

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