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Col du Dôme (Mt Blanc Massif, French Alps) suitability for ice-core studies in relation with past atmospheric chemistry over Europe

By SUSANNE PREUNKERT[1,2]*, DIETMAR WAGENBACH[2], MICHEL LEGRAND[1] and CHRISTIAN VINCENT[1], [1]*Laboratoire de Glaciologie et Géophysique de l'Environnement, CNRS, BP96, 38402 St. Martin d'Hères, France;* [2]*Institut fur Umweltphysik, Universität Heidelberg, INF 229, 69120 Heidelberg, Germany*

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ABSTRACT

The site of Col du Dôme glacier, located at 4250 m a.s.l. nearby the Mont Blanc summit (French Alps), was investigated for its suitability for reconstructing the anthropogenic perturbation of the atmosphere chemistry over Europe via glacio-chemical ice core studies. For this purpose, a 126 m long ice core drilled close to bedrock has been dedicated for glacio-chemical studies. Major ions (Na^+, NH_4^+, K^+, Mg^{2+}, Ca^{2+}, Cl^-, NO_3^-, and SO_4^{2-}) and D/H isotope ratios have been measured with high seasonal resolution along the upper 60 m of this core (covering 13 years). For dating by annual layer counting, a highly resolved ammonium profile was obtained for the entire core. To assess the spatial representativity of the chemical signals obtained from this core, additional chemical profiles were obtained from two shallow firn cores (13 and 20 m long) drilled at about 100 m from the deep ice core. All ice core parameters show regular seasonal variations with low winter and high summer values. Mean summer to winter ratios (averaged over the period 1981 to 1994) are close to 4 for NO_3^-, SO_4^{2-} and Ca^{2+}, but reach 14 for NH_4^+. While the shape of the mean seasonal cycles of NH_4^+, NO_3^-, and SO_4^{2-} show a flat winter minimum followed by persistently high summer concentrations, more flickered variations are observed in the case of Ca^{2+}. In contrast to the chemical species, δD shows a smoothed seasonal cycle. The chemical impurity levels and the δD content in snow deposits from Col du Dôme have been compared with those from Colle Gnifetti (4450 m a.s.l., located in the Swiss Alps, 80 km east of Col du Dôme) over a time period of 10 years. This comparison suggests that the two sites may experience similar atmospheric pollution conditions throughout the whole year, at least for NH_4^+, NO_3^-, and SO_4^{2-}. Precise dating of the ice core drilled in 1994 was achieved by annual layer counting using the NH_4^+ stratigraphy. The latter reveals that the glacier of Col du Dôme records well preserved snow deposits, arising from summer as well as from winter precipitation, over, at least, the last 75 years. However, the seasonal signal of the δD content appears to be disturbed at increasing depth, in particular below of 115 m. A systematic decrease in the ratio of the winter to summer net snow accumulation found with increasing depth is shown to exert an important influence on mean temporal ice core trends for parameters underlying a strong seasonal variation. The comparison of chemical impurities between Col du Dôme and Colle Gnifetti indicates that glacio-chemical ice core records from Col du Dôme will provide seasonally resolved records over the 20th century which are at least representative on a regional scale.

* Corresponding author.
e-mail: ps@glaciog.ujf-grenoble.fr.

1. Introduction

Polar ice cores represent a unique opportunity to reconstruct some key aspects of the chemical composition of our past atmosphere. However, large ambiguities exist in the source region apportionment of short lived species trapped in ice deposits, making the extraction of such records at various sites over the globe mandatory. Experiencing 10 m firn temperatures below the water freezing point, some mountain glaciers in addition to the polar ice sheets, Greenland and Antarctica, open the possibility to reconstruct several aspects of the past atmospheric chemistry over various regions of the world. This includes glaciers located in tropical regions such as the Huascaran ice cap in the Peruvian Andes (Thompson et al., 1995b), the Guliya (Thompson et al., 1995a) and Dunde (Thompson et al., 1988) ice caps located in Tibet, as well as Devon Island ice cap in the Canadian Arctic (Koerner, 1977), Mount Logan (Holdsworth and Peak, 1985) in western Canada, and Alpine sites (Wagenbach, 1993).

More specifically, long term ice core records from cold Alpine glaciers may help to infer the impact of human activities on climate variability and long term climatic changes at the European scale. Present information obtained on Alpine ice cores are mainly available from the Monte Rosa region (Colle Gnifetti) (Haeberli et al., 1983; Wagenbach, 1989). Due to the relatively low annual net snow accumulation of 0.2–0.4 m of water equivalent (w.e.) at this site, the ice records span more than several centuries. These Colle Gnifetti ice core studies have already revealed an enhancement of sulphate and nitrate concentrations over the 20th century as a result of increasing anthropogenic SO_2 and NO emissions (Wagenbach and Preunkert, 1996). However a straightforward interpretation of these records in term of past environment and climate remains limited by our present poor knowledge of their spatial and seasonal representativity (Wagenbach, 1993). A pilot study (Maupetit et al., 1995) achieved in snow deposited from 1988 to 1991 at the Col du Dôme (Mont Blanc Massif) showed a rather high snow accumulation rate ($\sim$1.5 m w.e. yr^{-1}), and well-preserved winter layers. In contrast to Colle Gnifetti, this site was expected therefore to permit the examination of past changes of atmospheric chemistry at a seasonal (winter/summer) resolution at least over the last few decades or so.

Here, we examine the usefulness of the Col du Dôme site ice core records with respect to their local and regional spatial variability, their accessible time resolution, as well as with respect to their potential extending into the pre-industrial era. For these purposes basic glaciological properties of the Col du Dôme site are discussed, along with highly temporal resoluted records of major ions, obtained from a 126 m long ice core and a 20 m long ice core recently extracted at that site.

2. Present knowledge of glaciological characteristics of the Col du Dôme glacier

Scientific work carried out in the vicinity of the Mont Blanc summit has been initiated by Joseph Vallot some hundred years ago. These pioneer meteorological and glaciological observations have been published at the turn of the last century in the *Annales de l'Observatoire Météorologique du Mont Blanc (1893 to 1905)*. In the 1970s, pioneer ice core studies have been carried out by the Laboratoire de Glaciologie et Géophysique de l'Environnement, Grenoble (LGGE) on several shallow firn cores (Lliboutry et al., 1976; Jouzel et al., 1977; Briat, 1978), followed by numerous investigations of glacio-chemical parameters at various drill sites (Fig. 1, Table 1) in this area. These previous studies gave information on some key glaciological properties of the Col du Dôme glacier site.

2.1. Firn temperature

The study of the firn temperature profile permits to evaluate one of the key characteristic of a high-elevation mid-latitude glacier site: its location with respect to the recrystallization and cold infiltration zone (i.e. absence of regelated layers and melt water percolation over more than one annual snow layer (Shumsky, 1964), respectively). Vallot (1893) reported a 15 m firn temperature at Mont Blanc Summit of -16.7°C (4807 m a.s.l., Fig. 1) at the beginning of this century. Lliboutry et al. (1976) concluded that the "firn cap" regime encountered at the Mont Blanc summit belongs to recrystallization zone. On the other hand, the

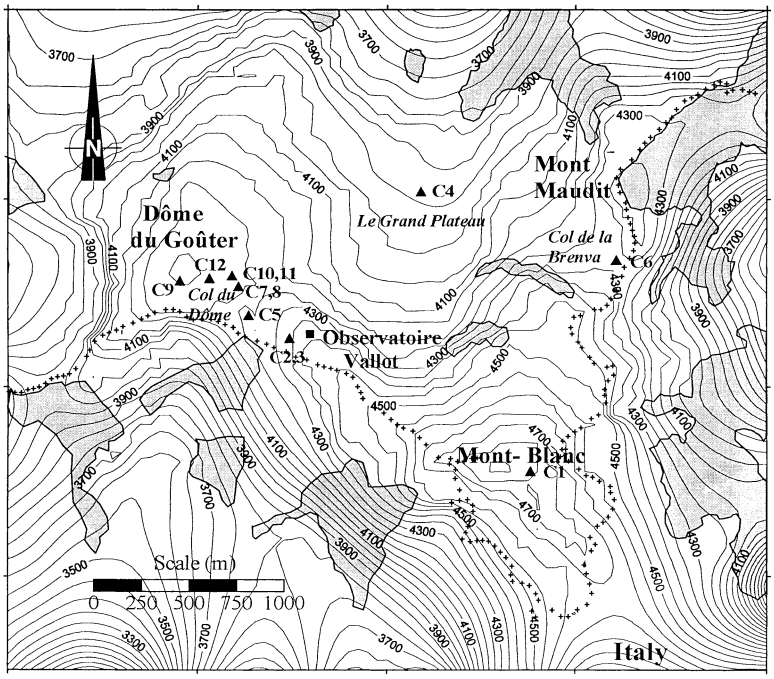

Fig. 1. Map of the Mont Blanc summit area showing the drill sites listed in Table 1. Glaciers are entered with elevation lines, rock regimes are marked in grey.

temperature close to −7.3°C observed at the Grand Plateau (3950 m a.s.l., Fig. 1) suggests that this area settles into the cold infiltration zone. More unclear is the case of the Col du Dôme (4250 m a.s.l.), where a firn temperature of −12.8°C was observed at 15 m depth in 1911 (Vallot, 1913) and no regelated layers were detected in the snow cover at that time. However, more recently, Lliboutry et al. (1976) found a firn temperature of −10.5°C and noticed the presence of a thick regelation layer. From these observations, the authors concluded that the Col du Dôme area has been located in the recristallisation-regelation zone (i.e. restriction of melt water percolation over less than one annual snow layer (Shumsky, 1964)) at the beginning of this century, but revealed characteristics of the cold infiltration zone in 1973. Based on the study of another ice core drilled at the Col du Dôme in 1980, Jouzel et al. (1984) reported a firn temperature of −12.4°C, close to the one observed by Vallot in 1911. These different pictures may be related to an unusually strong percolation having occurred in 1973, as suggested from Lliboutry et al. (1976). They may also reflect a large spatial variability of firn temperatures even at the small spatial scale of the Col du Dôme area, as it has been observed at the Colle Gnifetti (4450 m a.s.l., Mont Rosa massif, Switzerland). On this glacier, near surface firn temperatures show a systematic difference of 2°C between the southerly and northerly exposed slopes of the saddle, located in the cold infiltration zone and the recrystallization-regelation zone, respectively (Alean et al., 1983; Haeberli and Funk, 1991).

2.2. *Snow accumulation rate*

Other key information to assess the suitability of Col du Dôme for glacio-chemical studies are the spatial and seasonal variations of the snow accumulation. Although having been established for time periods limited to a few years, data reported in Table 1 suggest snow accumulation rates at the Col du Dôme ranging from 0.5 to 2.4 m w.e. yr^{-1}. Thus, a large variability of up to a factor of 5 exists within the area of about 0.4 km^2. A more detailed snow accumulation rate

Table 1. *Key parameters recorded in various firn and ice cores of the Mont Blanc Massif, for drill site positions see Fig. 1*

Drillsite	Year of drilling	Core length (m)	Mean accumulation rate of the core (m w.e. yr^{-1}) (covered time period (yr))	Analysed parameters	Ref.	No. in Fig. 1
Mont Blanc Summit	1973	16.7	2.8 (3)	firn temperature, ^3H, δD, gross β activity	Lliboutry et al. (1976) Jouzel et al. (1977)	C1
Col du Dôme	1973	24.5	—	firn temperature	Lliboutry et al. (1976)	C2
Col du Dôme	1974	21	~ 0.5[a] (26)	Na, Al, Cl, K, Ca, V, Mn, Fe, Cu, Zn, Cd and Pb	Briat et al. (1978)	C3
Grand Plateau	1974	—	—	firn temperature	Lliboutry et al. (1976)	C4
Col du Dôme	1976	30.6	~ 0.5[a] (36 ± 5)	acidity	Delmas and Aristarain (1978)	C2
Col du Dôme	1980	20	1.1 (11)	firn temperature, δD, ^3H, gross β activity	Jouzel et al. (1984)	C5
Col de Brenva	1984	10	~ 2[a] (2.5)	major ions, acidity	Ronseaux and Delmas (1988)	C6
Col du Dôme	1986	70	~ 1.5[a] (30)	Al, Ca, Na and K	De Angelis and Gaudichet (1991)	C7
Col du Dôme	1991	15	~ 1.5 (3.5)	major ions, acidity, δD	Maupetit et al. (1995)	C8
Col du Dôme	1991	—	0.61 (8)	^{137}Cs	Vincent et al. (1997)	C9
Col du Dôme	1994	126	2.22[b] (>75)	^{137}Cs, ^{210}Pb partly major ions	Vincent et al. (1997) This study	C10
Col du Dôme	1994	139	2.44[b] (>150)	^{137}Cs, ^{210}Pb CH_4	Vincent et al. (1997) Blanchard (1998)	C11
Col du Dôme	1997	20.5	2.25 (5.5)	major ions	This study	C12

[a] Estimate based on the density profile (C10 core).
[b] Mean accumulation rates from 1986 to 1994.

and surface flow study has been carried out by stake surveys from 1993 to 1995 at Col du Dôme (Vincent et al., 1997). This study reveals that: (1) the net snow accumulation rates are in good agreement with the measured vertical velocities during the investigated time period, leading the authors to conclude that the glacier is near to the steady state; (2) the vertical velocity varies from 0.2 m w.e. yr^{-1} at the Dôme du Goûter to 3.5 m w.e. yr^{-1} in the north east part of the saddle (Fig. 2b); (3) referring to the mean annual precipitation rate at Chamonix (1.25 m w.e. yr^{-1} from 1959 to 1994; data from Meteo France), part of the snow initially deposited in the vicinity of the Dôme du Goûter or in the southern part of the saddle is removed by strong winds, transported northward, and eventually deposited in the relatively wind protected north (or north east) part of the saddle. Note that at Colle Gnifetti, the net snow accumulation rate is only about 0.2–0.4 m w.e. yr^{-1} over the main part of the glacier saddle (Oeschger et al., 1977; Haeberli et al., 1983; Alean et al., 1983). This is at least three times lower than the amount of precipitation observed in surrounding valleys (Kromp-Kolb et al., 1993) suggesting that a dramatic net loss of deposited snow by wind erosion occurs at this site (Wagenbach et al., 1988). Furthermore, Vincent et al. (1997) found a regular year round snow deposition during their three year investigation by stake surveys at the Col du Dôme, while at Colle Gnifetti net the snow accumulation mainly takes place in summer (Haeberli et al., 1983; Wagenbach et al., 1988). Snow accumulation rates of ~ 2 and 2.8 m w.e. yr^{-1} are reported for Col de Brenva (Fig. 1) and Mont Blanc summit, respectively, but

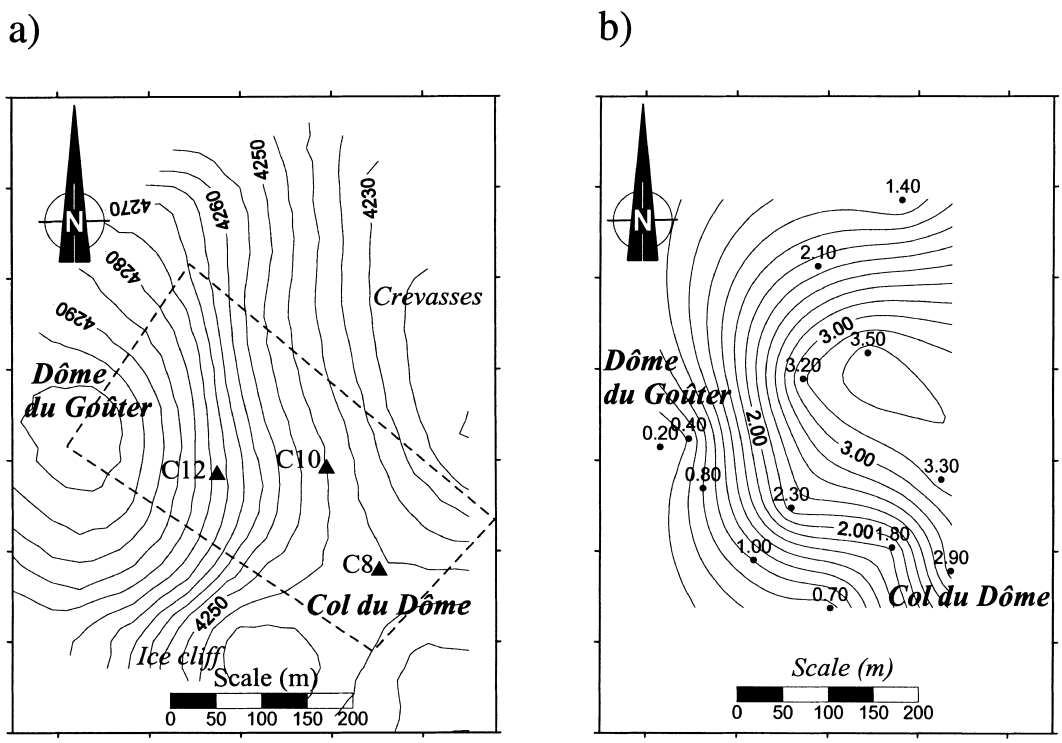

Fig. 2. (a) Map of the Col du Dôme area with positions of investigated ice cores. The dashed line refers to the area where radio echo soundings are available. (b) Vertical velocities observed at Col du Dôme (in m w.e. yr^{-1}) by Vincent et al. (1997) (reprinted from the Journal of Glaciology with permission of the International Glaciological Society).

no information on their spatial variability are presently available.

2.3. *Ice flow at Col du Dôme*

In collaboration with the ETH-Zürich, radio-echo soundings have been carried out by the LGGE in 1993 and 1994 at the Col du Dôme (Vincent et al., 1997). Within an area of 400 m by 200 m (Fig. 2a) the glacier thickness increases from about 40 m at the Dôme du Goûter to 150 m at the saddle area. The glacier bed topography appears to be very rugged, showing relatively deep (~15–30 m) and narrow (~30–60 m) canyons. Assuming that the glacier is at steady state, two models have been developed to depict the glacier flow in the Col du Dôme area, including cores C10 and C11 drilled in 1994 (Fig. 1). Using a rheological model for porous ice which is constrained by in situ measured density profiles and firn mechanical properties, a numerical simulation has been performed by Gagliardini and Meyssonnier (1997). A two-dimensional flow model was developed by Vincent et al. (1997), considering the mean mass balance derived from measured vertical surface velocities, using mass continuity and the depth profile of horizontal velocity from the analytical model proposed by Lliboutry (1981). The numerical simulations provided by the two models (Gagliardini and Meyssonnier, 1997; Vincent et al., 1997) are very consistent down to 90 m depth. They are also in agreement with the location of the ice layers corresponding to the radioactive horizons of 1986 (Chernobyl) and 1963 (nuclear bomb tests) detected in the C11 ice core at 39 and 90 m depth, respectively (Vincent et al., 1997). The radioactive horizon of 1954, appearing at 105 m depth in C11 (Vincent et al., 1997) indicates however that at this depth the two models already overestimate the age of the ice by some 20 years (Gagliardini and Meyssonnier, 1997) and 7 years (Vincent

et al., 1997), respectively. Based on the model study by Vincent et al. (1997), the ice located at ~20 to 30 m above bedrock may be 55 to 60 years old. Further down, the extensive deformation of ice layers leads to large uncertainties in the simulation. Flow model calculations applied to the low accumulation Colle Gnifetti site (Haeberli et al., 1988; Wagner, 1994) predict ice ages up to several hundred years for ice 20 m above bedrock.

2.4. Glacio-chemical studies

Investigations of various chemical species deposited at the Col du Dôme site over the last decades have revealed a rather well-marked seasonal signal. Jouzel et al. (1984) reported well-marked seasonal variations of the tritium content and of the total β radioactivity over the years 1970–1980. Although a clear summer maximum of δD (~ -67‰) is generally observed, a few perturbations (i.e. occurrence of a double annual peak, or a lack of well identified summer maximum) have been noticed by Jouzel et al. (1984). Maupetit et al. (1995) showed that all major ions (Na^+, K^+, NH_4^+, Mg^{2+}, Ca^{2+}, Cl^-, NO_3^-, and SO_4^{2-}) exhibit a seasonal signal in the snow deposited between 1988 and 1991, with very low concentrations in winter, enhanced levels in spring and well-marked maxima in summer. Such seasonal patterns are in agreement with atmospheric observations made on the high Alpine sites of Colle Gnifetti, Jungfraujoch (3450 m a.s.l., Bernese Alps, Switzerland) and Hoher Sonnblick (3106 m a.s.l., Hohe Tauern, Austria) (Preunkert and Wagenbach, 1998; Baltensperger et al., 1997; Kasper and Puxbaum, 1998). The low impurity content in winter at the high Alpine sites is due to the very limited contamination of these high altitude regions by polluted boundary layer air masses during that season. Conversely, in spring and summer, upward advection by convective transport significantly affects Alpine sites located well above 4000 m elevation. In addition, disturbances arise within the snow chemistry by the episodic deposition of saharan dust, as reported by Wagenbach et al. (1996), Wagenbach and Geis (1989), Wagenbach (1989) for Colle Gnifetti, and by Ronseaux and Delmas (1988), De Angelis and Gaudichet (1991) for Col du Dôme. The use of alpine ice cores as recorder of such short-term deposition events remains difficult, however. For instance, the Alpine wide strong saharan dust event of 1977 (Haeberli, 1983; Wagenbach and Geis, 1989) is not seen as a major event in the Col du Dôme snow layers (De Angelis and Gaudichet, 1991). Similarly, at the Col du Dôme the Chernobyl radioactivity horizon exhibits inventories varying up to a factor of 120 (Vincent et al., 1997).

3. Sampling and analyses

In 1994 two ice cores were drilled close to bedrock at Col Du Dôme. The core lengths are 126 m for core C10, and 139 m for core C11 (Fig. 1). The C11 core was dedicated to the measurement of trace gases in air bubbles whereas the C10 core was reserved for the snow chemistry. Major ions (Na^+, NH_4^+, K^+, Mg^{2+}, Ca^{2+}, Cl^-, NO_3^-, and SO_4^{2-}), and the stable water isotope Deuterium have been measured with high-resolution (i.e., ~30 samples per year) along the upper 60 m of the C10 core. Extended NH_4^+ measurements have been carried out over the entire C10 core for dating purpose (see Section 6). Additionally, major ions records were obtained along a 20 m long core extracted in 1997 (C12, Fig. 2a) to evaluate the spatial representativity of chemical records in this area.

In addition to ice core decontamination and chemical measurements, which have been carried out at the Institut für Umweltphysik, Heidelberg (IUP) and the Laboratoire de Glaciologie et Géophysique de l'Environnement, Grenoble (LGGE), Deuterium (D) analyses were performed at the Laboratoire des Sciences du Climat et l'Environnement, Saclay (LSCE). As usual, D/H ratios are reported here in δD units denoting the relative deviation from V-SMOW standard (Vienna-Standard Mean Ocean Water) in permille. The core decontamination at Heidelberg was done by using an electric plane device (Fischer et al., 1998), while a lathe was used at Grenoble (Legrand et al., 1993). In both laboratories ion chromatography (IC) procedure currently used for polar and alpine snow samples (Legrand et al., 1993; Maupetit, 1992; Minikin, 1994; Fischer, 1997; Preunkert, 1994) were applied here to determine major ions concentrations. Intercalibrations have shown a quite good agreement between the IC analytical procedures employed in both laboratories (Maupetit et al., 1995). The accuracy of IC determinations are typically of 5 to 10% for winter and summer samples, respectively. The only

one noticeable difference in the methods employed at the two laboratories concerns the use of an autosampler device at the IUP lab which slightly increases the ammonium blank value to 2.1 ± 3.7 ppb (averaged over 152 samples), instead of far less than 1 ppb generally obtained when using standard procedure (including individual sample load) at LGGE.

4. Seasonal variations of major ions and the δD content in the snow cover of Col du Dôme

4.1. Seasonal variations along the upper 60 m of the C10 ice core

The density of the core from close to the surface is about $0.35\ g\ cm^{-3}$. The critical density ($0.55\ g\ cm^{-3}$) is reached at 5 m depth and the close-off depth (i.e. density of 0.80–$0.83\ g\ cm^{-3}$) lies between 60 and 70 m depth. Below 100 m, density values remain quasi constant ranging between 0.90 to $0.92\ g\ cm^{-3}$. As previously argued by Jouzel et al. (1984), De Angelis and Gaudichet (1991), and Maupetit et al. (1995), the visual inspection of grain size variation and the occurrence of recristallisation ice layers, as well as the seasonal change of the δD content can be used to identify annual layers and thus help to establish the dating of the Col du Dôme ice cores. Based on the regular occurrence of δD minima (winter layers) and on the occurrence of ice layers (summer layers) (Fig. 3), we found the annual accumulation rate varying from 1.25 to 4.50 m w.e. yr^{-1} (mean value of 2.45 m w.e. yr^{-1}) along the upper 60 m of the C10 core, which cover the 1981–1994 time period (i.e. 13 years). Compared to the mean precipitation rate at Chamonix (1.25 m w.e. yr^{-1} from 1959 to 1995, Meteo France data), this high net snow accumulation rate of 2.45 m w.e. yr^{-1} within the C10 core suggests an accumulation surplus by drifting snow there. However, the regu-

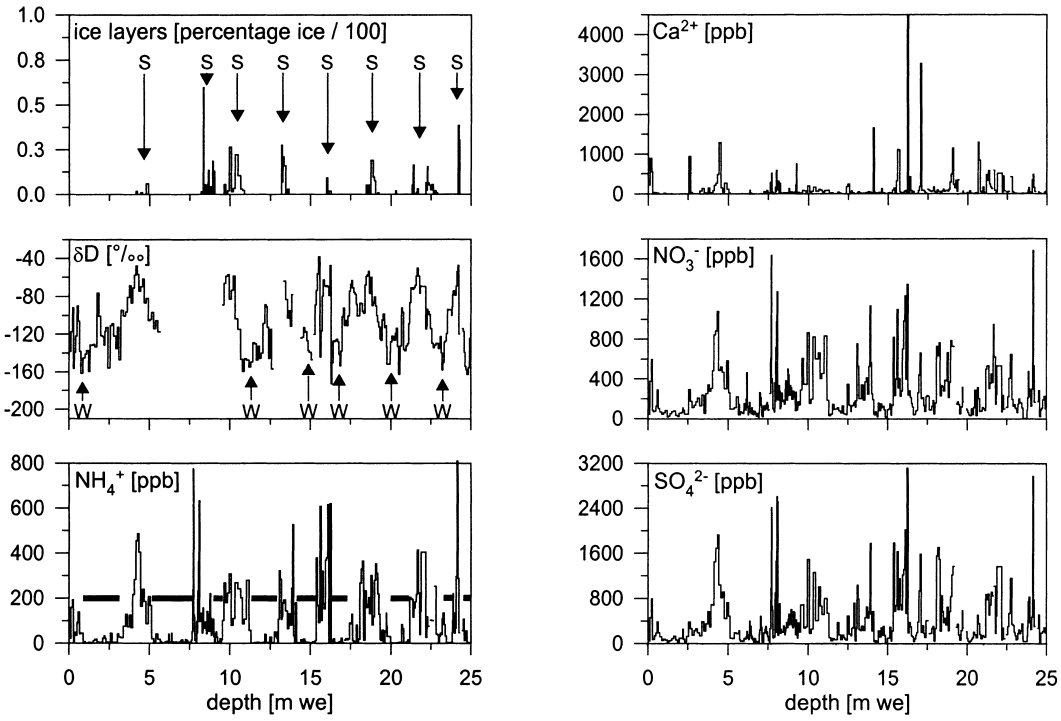

Fig. 3. Stratigraphy and chemical depth profiles of the upper part of the C10 ice core. The regular occurrence of δD minima (vertical arrows denoted *W*) and ice layers (vertical arrows denoted *S*) indicate winter and summer snow layers. Horizontal lines reported in the ammonium profile denote the annual winter snow layers (see text).

lar formation of ice layers in summer very likely prevents the perturbation of the stratigraphy by wind redistribution of snow over more than one year. In addition to summer ice layers which are up to 5 cm thick, vertical ice veins resulting from percolation had been detected four times over the 13 years of the record. Their maximum vertical extend reaches 0.20 m w.e., which remains small compared to the annual firn layer thickness of 1.25 to 4.50 m w.e..

As seen in Fig. 3, depth profiles show seasonal variations for all ions and δD, with very low winter and high summer values. Summer to winter mass ratios for individual samples may reach up to 800 for NH_4^+, 120 for SO_4^{2-}, 60 for NO_3^-, and 500 for Ca^{2+}. Given the quite large interannual variability of the annual snow accumulation, we divided each annual layer into two parts corresponding to winter and summer snow accumulation. Note that summer and winter means summer halfyear and winter halfyear, defined by the distinct atmospheric conditions prevailing at high Alpine sites, the presence and absence of upward advection of air masses from the boundary layer, respectively. The dissection cutting has been based on the NH_4^+ profile which exhibits the strongest seasonal signal as seen in Fig. 3. The frequency distribution of NH_4^+ concentrations (not shown) indicates a bimodal distribution with a low concentration mode below 10 ppb and a second mode centred on $\sim$200–300 ppb. The boundaries of the winter half year snow pack have been identified by requiring at least 3 consecutive samples to significantly exceed the 10 ppb NH_4^+ (background) level. Indeed, in more than 90% of samples assumed to be winter snow, NH_4^+ concentrations are lower than 10 ppb. Using this criterion, we found that 46 % of the net snow deposition may be attributed to winter snow precipitation in the upper 60 m of C10.

Each summer and winter layer was divided into 6 equally spaced (in w.e.) parts. A weighted interpolation was applied to years containing more or less than 12 samples. These calculations were performed for NH_4^+, Ca^{2+}, NO_3^-, SO_4^{2-}, and δD. Finally, a 13-year mean was calculated for each 1/12 annual accumulation (Fig. 4) in order to derive mean seasonal cycles. In Fig. 4, we also report the mean seasonal temperature cycle observed at 2472 m a.s.l. in the Alps (Grosser St. Bernhard, Switzerland, (Aschwanden et al., 1996)), indicating an approximate monthly time scale of the mean seasonal cycles in C10 by matching the temperature and the δD seasonality.

The seasonal cycles of major ions show a flat winter minimum from September to February followed by persistent high concentrations from March to August, in particular for NH_4^+, NO_3^-, and SO_4^{2-}. For Ca^{2+}, a more flickered seasonal cycle is observed (Fig. 4), likely due to the sporadic inputs of Saharan dust, which mainly occur from spring to fall. The mean summer to winter ratio remains close to 4 for SO_4^{2-}, NO_3^- and Ca^{2+}, and reaches 14 for NH_4^+ (Table 3). The relatively huge summer to winter mass ratio for ammonium suggests that, in addition to the atmospheric dynamics, a strong seasonal cycle of ammonia emissions may also contribute to the observed variations. Indeed, the seasonality of NH_3 emissions exhibits a strong maximum in summer in the pre-alpine areas (Puxbaum, 1993), in relation to agricultural activities and NH_3 soil emissions by bacterial activity. In contrast to ammonium, Cl^- and Na^+ reveal a low summer to winter ratio (close to a factor of two, Table 3) likely mirroring a quite complex picture. In addition to the transport, various changes in sources may modulate the Na^+ deposition in precipitation at high elevation European sites throughout the year. For instance, we can speculate that the sea salt content of the free troposphere at mid latitudes is higher in winter than in summer because of the more intense cyclogenesis over the Atlantic Ocean at that time. Conversely, the higher dust input from spring to fall over alpine sites depicted by the calcium profile may be bring Na^+ inputs because of the presence of Na^+ in airborne continental aerosols.

As a consequence of matching with the seasonal air temperature cycle, the mean seasonal δD cycle of C10 shows minimum in late winter (February) and maximum in summer (July). In contrast to chemical species, the δD profile exhibits a smoothed seasonal variation, similarly to the seasonal changes of air temperatures (Fig. 4). Since the shape and the summer/winter amplitude of the δD depth profile do not change significantly with increasing depth (Fig. 3), we can exclude that water vapour diffusion in firn contributes to the smoothed pattern of δD ice signal. This is also consistent with laboratory experiments on diffusion of deutered water vapour which suggest

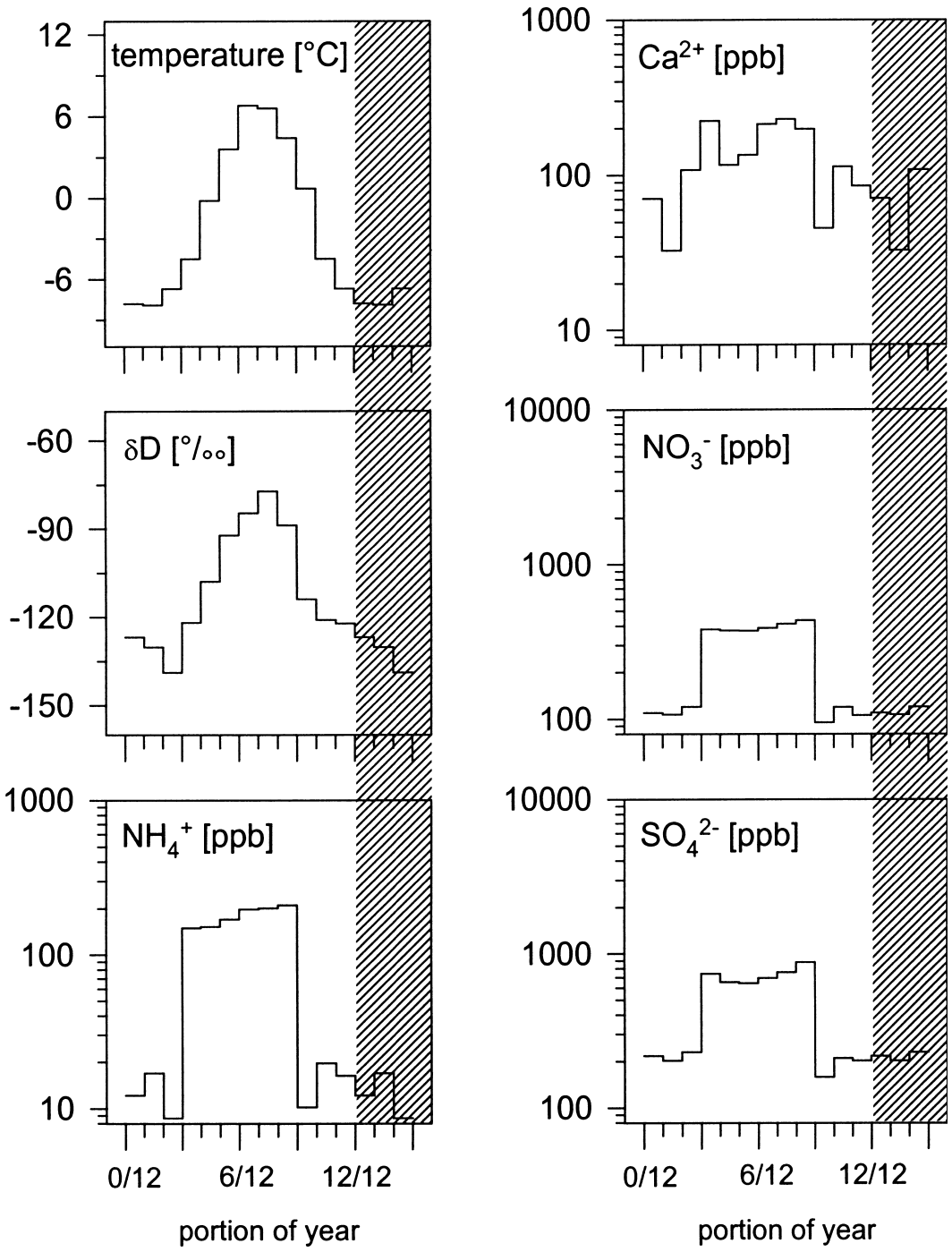

Fig. 4. Seasonal cycles of δD, NH_4^+, Ca^{2+}, NO_3^- and SO_4^{2-} from the C10 core averaged over 13 years. Monthly time scale is indicated by matching the δD annual cycle with the 30-year mean annual cycle of air temperatures (top left panel) observed at Grosser St. Bernhard (2427 m a.s.l., Switzerland) (data from Aschwanden et al., 1996).

that this phenomenon remains limited to less than 15 cm long in firn over one to two years (Jean-Baptiste et al., 1998), while within C10 the transition section summer to winter is about 1 m.

4.2. *Spatial variability of seasonal variations within the glacier site of Col du Dôme*

The seasonal pattern seen in the snow chemistry at the C10 Col du Dôme site is compared with that obtained at two neighbouring sites. Two shallow cores (C12) and (C8) were drilled 100 m away from the C10 site in 1997 and 1991 (Maupetit, 1992), respectively (Fig. 2a). With ~11.5 and 6.5 m w.e. length, the C12 and C8 cores span around 6 and 3 years, respectively. They were dated by using the same procedure as applied for the C10 core. Winter snow layers represent 40% and 44% of the annual net snow accumulation in C12 and C8 cores, respectively, what is in excellent agreement with the C10 core (46%). Considering the strong wind scouring at such exposed glacier sites, the chemical stratigraphies of the 3 cores are in good agreement, as illustrated by the ammonium profiles (Fig. 5).

In order to achieve an accurate comparison of winter and summer concentrations of chemical species present in ice from different sites, we have to account for the high spatial variability of sporadic events like episodic saharan dust inputs in our ice records (see Subsection 2.4). For instance, among five layers marked by huge calcium contents in core C10, four are clearly seen in cores C8 and C12, but their Ca^{2+} inventory differs up to a factor of 4. These sporadic calcium events are caused by deposition of large amount of alkaline dust over the Alps. As acidic species are readily scavenged by mineral dust particles, those events may add considerably to the snow inventory of sulphate and, to a lesser extend, of nitrate (Maupetit and Delmas, 1994). Following Wagenbach et al. (1996), such events can be identified in alpine ice as calcium-rich layers exhibiting a negative acidity (i.e. alkaline samples).

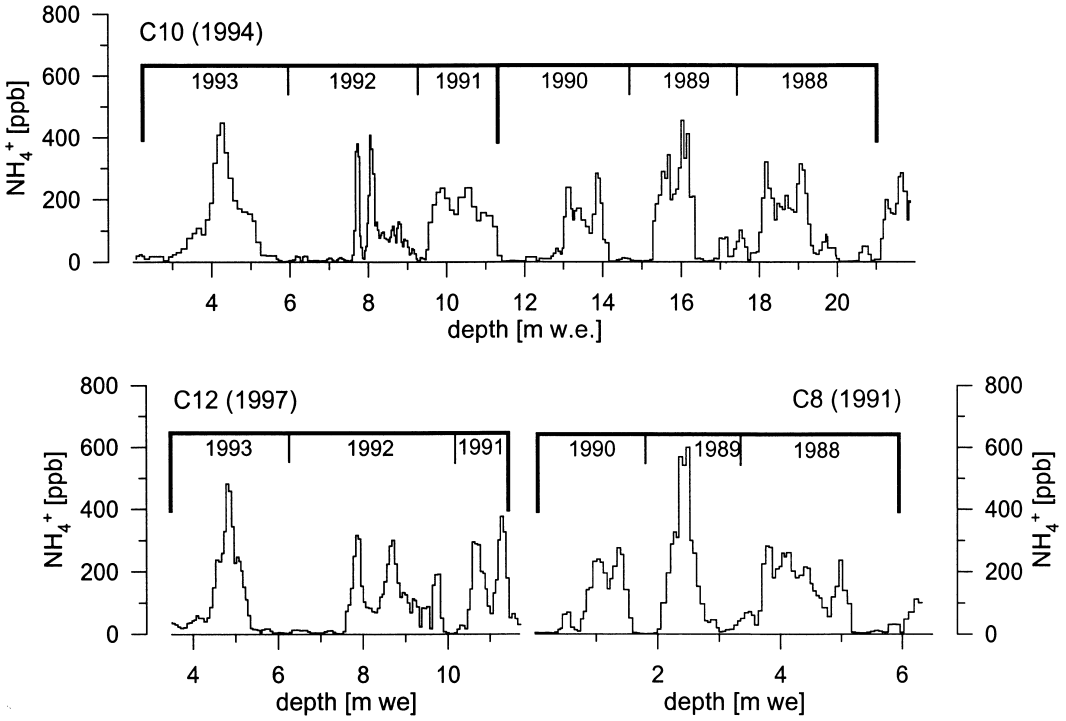

Fig. 5. Smoothed (3 sample running average) NH_4^+ profiles of the C10 core (top) and two shallow firn cores: C12 (left bottom), C8 (right bottom, adapted from Maupetit (1992)). The two latter cores have been extracted at a distance of about 100 m to C10 (Fig. 2a).

Table 2. *Seasonal snow chemistry variation over the 1988–1993 time period of the C10 core as compared to shallow firn cores C8 and C12*

		Na^+	NH_4^+	Ca^{2+}	Cl^-	NO_3^-	SO_4^{2-}
Summer	C10 mean concentration (ppb)	25 ± 14	180 ± 58	62 ± 28	43 ± 17	440 ± 101	620 ± 131
	mean square deviation (ppb)	15 ± 16	45 ± 53	48 ± 51	16 ± 17	99 ± 95	155 ± 173
	relative mean square deviation (%)	61 ± 63	25 ± 29	77 ± 83	38 ± 41	22 ± 22	25 ± 28
Winter	C10 mean concentration (ppb)	11 ± 8	13 ± 9	29 ± 17	17 ± 10	120 ± 35	180 ± 54
	mean square deviation (ppb)	3.4 ± 2.5	2.3 ± 2.0	21 ± 26	4.8 ± 2.7	29 ± 26	75 ± 62
	relative mean square deviation (%)	31 ± 22	17 ± 16	72 ± 86	28 ± 28	23 ± 21	42 ± 35

Alkaline samples were identified by calculating the ionic balance between anions and cations, acidic samples corresponding to an excess of anions, alkaline samples showing a deficit of anions. Using this method, we conclude that 13% of C10 samples, 8% of C12 and 19% of C8 have been affected by such sporadic, excess alkaline dust events. These data have been discarded from our calcium records. For sulphate, the observed concentrations were corrected for the terrigenous contribution according to $[SO_{4\,corr}^{2-}] = [SO_4^{2-}] - 0.59[Ca^{2+}]$, where the factor 0.59 represents the mean SO_4^{2-} to Ca^{2+} mass ratio observed by Wagenbach et al. (1996) in pre-industrial mineral dust horizons at Colle Gnifetti.

To compare the mean summer and winter concentrations among the three cores, we calculated water weighted mean concentrations for each summer and winter half year. To give a proportion for the spatial variability within Col du Dôme, square deviations for summer and winter layers were calculated between C10 and the shallow cores (C12, C8) for each year. Overall means have been then calculated by averaging the annual deviation values over the concurrent period from 1988 to 1993. Relative deviations given in Table 2 are normalised to the corresponding C10 mean concentrations.

In summer, annual mean square deviations range from 22% to 25% for NH_4^+, NO_3^-, and SO_4^{2-}, while Ca^{2+} shows a larger square deviation of 77%. Such a difference might be due to an enhanced variability of the non Saharan dust influenced Ca^{2+} background, or due to low level Saharan dust events being still included in the background population. The mean winter square deviations for NH_4^+, Ca^{2+} and NO_3^- are similar to their respective summer values. For SO_4^{2-}, the higher square deviation in winter compared to summer (Table 2) cannot be attributed to an incomplete correction of concentrations from the saharan dust contribution, since quite similar values are obtained when only acidic samples are considered. The variability of Na^+ and Cl^- are relatively high in summer (61% and 38%, respectively) when compared to other non crustal species. Although the source apportionment of these species is not yet established, the higher square deviation observed for sodium in summer may support the assumption of a significant dust contribution for this species.

5. Spatial variability of seasonal variations on a regional scale

An attempt was made to compare summer and winter means of chemical species and δD between Col du Dôme and Colle Gnifetti. At the low accumulation site Colle Gnifetti which is located $\sim$80 km east to Col du Dôme only a small fraction of winter snow is preserved (see Subsection 2.2). Here seasonal means have been examined along a 5.6 m w.e. long firn core (further-on denoted HK) showing relatively regular seasonal cycles and covering the time period 1982–1991. Given the annual accumulation rate of $\sim$0.37 m w.e. yr^{-1}, the identification of summer and winter

layers was still possible, and the winter snow contribution was found to account for 18% of the annual net snow accumulation (Preunkert, 1994). Seasonal subdivision, as well as the correction from saharan dust impact and the calculation of mean summer and winter values have been done by using same procedures as employed for the C10 core. Taking the overall summer and winter means (from 1982 to 1991) for each species, we calculated averaged deviations between C10 and HK normalised to the C10 means (Table 3). In spite of the more difficult subdivision between summer and winter layers in the HK core, a good overall agreement can be observed for both mean summer and winter concentrations between the two glacier sites. In summer, the δD content is lower at Colle Gnifetti with respect to Col du Dôme by 5.5‰. This is expected, given the 200 m higher elevation and the more easterly position of Colle Gnifetti. In winter, however, the δD values are higher by 10‰ in HK than the C10 core, likely reflecting the incomplete deposition of winter precipitation at Colle Gnifetti due to its lost through wind scouring.

The significance of slight differences detected for the mean summer and winter ion concentrations between the two glacier sites has been assessed by a Students test (t-test). For each species, the t-test was applied to the difference record (R), which was built up by subtracting individual seasonal means of HK from the corresponding ones of C10 over the 1982–1991 time period. The R-values (10 samples) were expected to be trend free and should be t-distributed around zero (i.e., the expected value (ξ_0) is equal to zero) under the assumption that the concentration record of C10 and HK is identical. The level of significance (α) has been chosen equal to 0.05 which leads to a critical t-value of 2.26. Calculated values of t, as well as the values of σ (standard deviation) and ξ (mean value) are reported in Fig. 6a (summer) and 6b (winter) for each species.

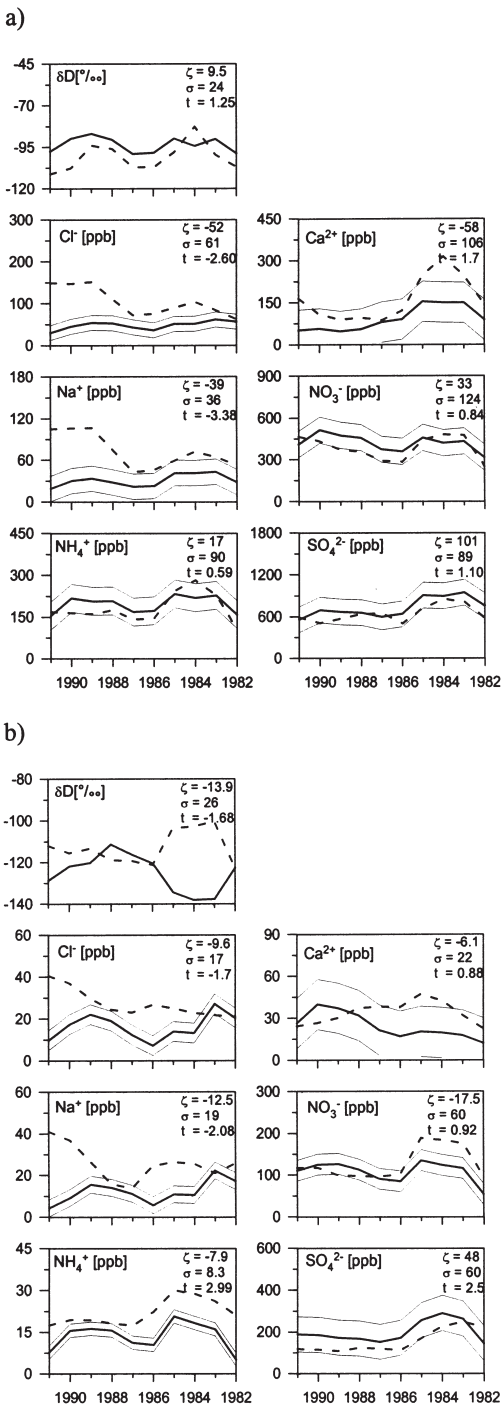

Fig. 6. Three point running averages of annual summer (6a) and winter (6b) contents of chemical species and δD at Col du Dôme (C10, solid lines) and Colle Gnifetti (HK, dashed lines). Together with the C10 curves, the respective relative mean square deviations reported Table 2 have been here applied (thin solid lines enveloping the thick solid lines). For each species, the t-test values of ξ, σ and t (see text) are included.

Table 3. *Mean 10-year (1982–1991) summer and winter concentrations of chemical species and δD at Col du Dôme (C10) and Colle Gnifetti (HK); in addition to the summer to winter concentration ratios, the differences between HK and C10 relative to C10 are given*

			Na^+	NH_4^+	Ca^{2+}	Cl^-	NO_3^-	SO_4^{2-}	δD[a)]
Col du Dôme (C10)	mean concentration (ppb)	summer	30 ± 19	200 ± 67	94 ± 60	47 ± 18	430 ± 120	740 ± 230	−93.5 ± 13
		winter	13 ± 12	14 ± 10	25 ± 16	17 ± 14	110 ± 57	200 ± 80	−126 ± 13
	ratio	summer/ winter	2.3 ± 2.6	14 ± 11.3	3.8 ± 3.4	2.6 ± 2.5	3.9 ± 2.3	3.7 ± 1.9	32.7 ± 18.4[b)]
Colle Gnifetti (HK)	mean concentation (ppb)	summer	70 ± 42	180 ± 110	186 ± 145	100 ± 60	380 ± 190	660 ± 290	−99 ± 18
		winter	25 ± 13	20 ± 9	37 ± 13	27 ± 8	134 ± 91	145 ± 85	−116 ± 23
	ratio	summer/ winter	2.8 ± 2.3	9.3 ± 7.0	4.7 ± 4.3	3.7 ± 2.5	2.8 ± 2.4	4.5 ± 3.3	17 ± 29.2[b)]
$\frac{HK-C10}{C10}$ (%)		summer	133	−7	87	112	−12	−11	5.8
		winter	92	42	48	59	23	−28	8

[a)] (‰).
[b)] Summer–winter difference.

For relative small values of $|t|$ (<1.10), for which the level of significance could be chosen higher than 0.3, the test is not sensitive enough. These small t-values could be due to a dating error of the HK core, leading to enhanced σ values. Therefore, in addition to the t-test values, we also report in Fig. 6 smoothed time records from both cores.

In summer, both the t-test and smoothed records show that Cl^- and Na^+ signals are significantly different in the two cores. Conversely, for NH_4^+, NO_3^-, SO_4^{2-} and Ca^{2+}, no significant differences are assessed by the t-test, and a good agreement of the temporal trends can be observed. Again, note the very large square deviation for Ca^{2+} at Col du Dôme. The δD signal at HK shows generally lower levels with respect to the C10 ones in summer. However, the t-test suggests that these differences are not significant.

In winter, for NH_4^+, NO_3^- and SO_4^{2-} similar temporal trends are observed in both cores. For SO_4^{2-} and NH_4^+, the t-test suggests a significant difference between C10 and HK. However, due to the high spatial variability at Col du Dôme (Fig. 6b), the lower SO_4^{2-} concentrations, with respect to the C10 ones, at HK are not necessarily caused by a systematic spatial difference between Col du Dôme and Colle Gnifetti. The higher winter NH_4^+ values for HK, compared to C10, may be explained with a regular lack of mid-winter snow preservation at Colle Gnifetti. In spite of this lack of winter snow deposition at the HK site, the good agreement on temporal trends seen for NH_4^+, NO_3^- and SO_4^{2-} (Fig. 6b) suggests that the chemical signals recorded in snow at the two sites correspond to similar air chemical situations. For Na^+, Cl^-, Ca^{2+} and δD the t-test results suggest that the records of both cores are not significantly different in winter.

This suggests that Colle Gnifetti and Col du Dôme likely experience similar atmospheric conditions throughout the whole year, at least for species such as NH_4^+, NO_3^- and SO_4^{2-}. Based on the comparison of two shallow firn cores, Maupetit et al. (1995) suggested that during summer, the Col du Dôme is more influenced by polluted air masses than the Colle Gnifetti. Such a conclusion is not supported by this study, which shows no significant difference between Colle Gnifetti and Col du Dôme for NH_4^+, NO_3^- and SO_4^{2-} in the snow deposited during the summer season. This apparent discrepancy likely comes from an underestimation of the true summer levels at Colle Gnifetti made in the study by Maupetit et al. (1995), where summer means from Col du Dôme have been compared with overall mean concentrations from Colle Gnifetti (assumed to be dominated by summer precipitation).

6. Stratigraphy and dating of the C10 core

With the aim to establish a chronology of the C10 core, the upper 119 m were subdivided in summer and winter sections based on the ammonium profile, as detailed in Subsection 4.1. In the lower part of the core, the winter snow layers are very thin (Fig. 7) and are sometimes made up by one single sample (i.e., 4 cm w.e.) below the 10 ppb NH_4^+ criterion. We identified winter layers between 103 and 119 m (77–91 m w.e.) depth via the relative NH_4^+ minima, though sometimes exceeding 10 ppb. Possibly, winter sections are smaller here than our depth resolution of 5 cm. For the very last 7 m of the core, annual layer counting was not possible by this method, since the NH_4^+ record shows no regular variations. Note

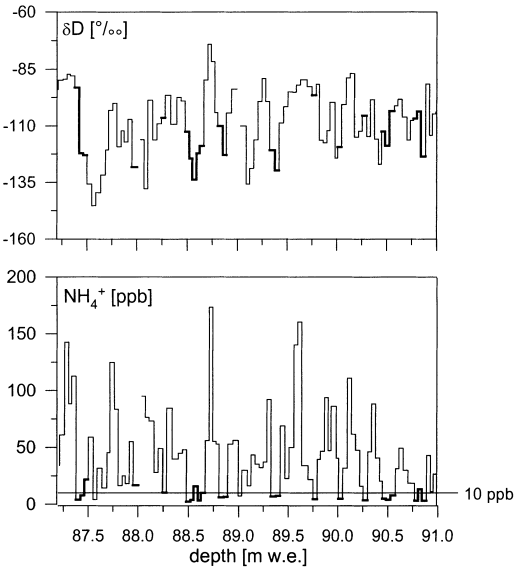

Fig. 7. δD and NH_4^+ raw data of the C10 core between 87 and 91 m w.e. depth. Winter snow layers, as identified from the ammonium stratigraphy (see text), are indicated as thick lines.

that this absence of the observance of regular NH_4^+ variations is not due to our depth resolution of 5 cm, since continuous, high resoluted NH_4^+ measurements performed over the last 20 m of C10 (depth resolution $\sim$7 mm w.e., (Fuhrer et al., 1993)) yield similar results (S. Sommer, 1995, personal communication).

6.1. Dating of the C10 core

The dating established by counting annual layers along the NH_4^+ profile (Fig. 8) has been compared with various time horizons and a chronology derived from an ice flow model. Time horizons are gained from ^{137}Cs measurements (Vincent et al., 1997), as well as from the calcium record of dust horizons. The dust horizons of 1977, 1947 and 1936/37 had also been detected in snow deposits at Colle Gnifetti (Haeberli et al., 1983; Jung, 1993). The theoretical time scale reported as a solid line in Fig. 8 refers to the 2D flow model from Vincent et al. (1997) applied to the C10 core. The annual layer counting is in good agreement with the three ^{137}Cs horizons (1986, 1963, and 1954). Assuming that the dust horizons of 1947 and 1936/37 are correct, a systematic overestimation of $\sim$5 years arises from the NH_4^+ dating at 114 m depth (i.e., 86 m w.e.), corresponding to an age of $\sim$60 years. The core chronology derived from the flow model still agrees within $\pm$5 years with the NH_4^+ stratigraphy down to 90 m depth (i.e., 65 m w.e. corresponding to the last 30 years). However, at 100 m depth (i.e., 74 m w.e., Fig. 8), the model overestimates the age by some 18 years, while the NH_4^+ stratigraphy is still in agreement with the ^{137}Cs horizon of 1954. This systematic error of the model may be related to an underestimation of the glacier thickness at the drill site. Radio echo soundings suggested an ice thickness of about 122 m (core length of 126 m) but this thickness determination has a quite large uncertainty of $\pm$20 m (S. Suter, 1999, personal communication). The model (dashed line in Fig. 8) and the stratigraphical time scale are in better agreement when a glacier thickness of 140 m is assumed, which remains within the range of the ice thickness determination (from 102 to 142 m).

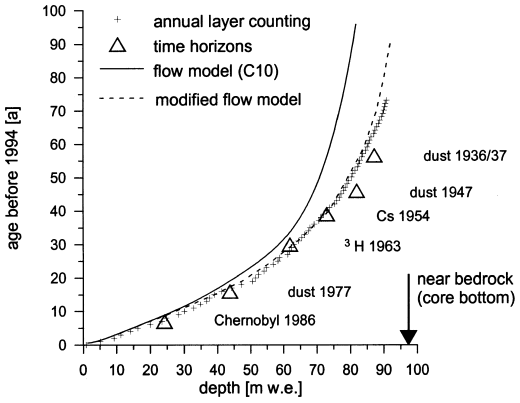

Fig. 8. Comparison of different depth-age relation in the upper 119 m of the C10 core (see text).

Given the good agreement between mean NH_4^+ summer concentrations at Col du Dôme and Colle Gnifetti (Section 5), we compared the mean NH_4^+ summer concentrations observed between 116 and 119 m depth within the C10 core (i.e., 88–91 m w.e.) with concentrations of the monotonic increasing NH_4^+ time trend during this century, gained from a Colle Gnifetti ice core (dominated by summer snow) (Jung, 1993; Wagenbach and Preunkert, 1996). The well established chronology of this core suggests that the C10 ice located at around 116–119 m depth (i.e., 88–91 m w.e.) is 75 ± 10 years old. This is again in agreement with our dating obtained by annual layer counting along the NH_4^+ profile.

6.2. The seasonal δD signal in deep C10 ice layers

While ammonium minima coincide rather well with δD minima from the surface down to 60 m depth (Fig. 3), a more complex picture is observed further down from 115 to 119 m depth (Fig. 7). Here, the ammonium minima do not always correspond to layers low in δD (Fig. 7). In addition, while the seasonal δD amplitude within the upper 60 m ranges from -60 to -160‰ (Fig. 3), -90 to -130‰ (i.e., a decrease by more than a factor of 2) are seen between 115 and 119 m depth (i.e., 87–91 m w.e.). This may be due to water vapour diffusion occurring during firnification. The model of Johnsen (1977) gives the ratio R of the final (post firnification) to the initial (near the surface) amplitude of an annual water isotope cycle as a function of its wavelength (A).

$$R = \exp\left(\frac{-2\pi^2 L_f^2}{A^2}\right) \quad (1)$$

Hereby L_f (the average displacement due to vapour water diffusion) was roughly estimated for the firn temperature (~ -12.5°C) and the accumulation rates prevailing at the Col du Dôme, by extrapolating the theoretical L_f values given by Cuffey and Steig (1998).

For the upper most 60 m of the C10 site, with a mean accumulation rate of 2.45 m w.e. yr^{-1}, an average displacement (L_f) of less than 13 cm (in w.e.) is obtained, leaving the R value higher than 0.95. This confirms our conclusion made in Subsection 4.1 that smoothing by diffusive water vapour transport is relatively unimportant in this part of the core. On the other hand, further down in the core, the ice has been originally deposited upstream the borehole of the C10 site. Based on the ice flow model by Vincent et al. (1997), the ice located at $\sim$115–120 m depth (i.e., 87 to 90 m w.e.) comes from sites where surface accumulation rates range from 0.5 to 0.3 m w.e. only. From such surface snow accumulation conditions, an average displacement (L_f) of $\sim$11 to $\sim$14 cm w.e. is expected, which, according to eq. (1), would correspond to R values of 0.5 to 0.01, respectively. Thus, the decreasing amplitude of the δD variations between upper layers and 115–119 m depth may be attributed to smoothing by diffusive mass transport in connection with the decrease of the net snow accumulation upstream the C10 site. Furthermore, Johnsen (1977) pointed out, that for R lower than 0.05, the seasonal δD cycle is obliterated after the firnification. Hence, the seasonal δD cycle of C10 starts to become obliterated at a core depth of 115 to 119 m, and δD fluctuations seen around 119 m depth and in deeper ice layers may be likely not connected anymore to seasonal variations.

6.3. *The seasonal variations of the net snow accumulation versus depth*

Based on the ammonium profile, a decrease of the annual layer thickness with depth is observed along the C10 core (Fig. 9). The 10-year means of the snow accumulation rate show an almost linear decrease from 3.3 m w.e. yr^{-1} at the surface to 0.2 m w.e. yr^{-1} at 118 m depth (i.e., 90 m w.e.). The change of the annual thickness with depth, as calculated by the ice flow model (Vincent et al., 1997), is also shown in Fig. 9. Based on the NH_4^+ stratigraphy, the ratio of winter to summer net snow accumulation (denoted as $A_{W/S}$) was calculated for each individual year. As is seen in Fig. 9, $A_{W/S}$ is decreasing from $\sim$1 near the surface to 0.5 at 100 m depth (i.e., 74 m w.e.). This change suggests that spatial variations in the relative amount of summer and winter snow deposition taking place upstream to the C10 borehole.

With regard to the strong seasonal cycle, exhibited by the ice impurity content (maxima in summer), the decrease in the fraction of winter snow deposition with depth may lead to a down-

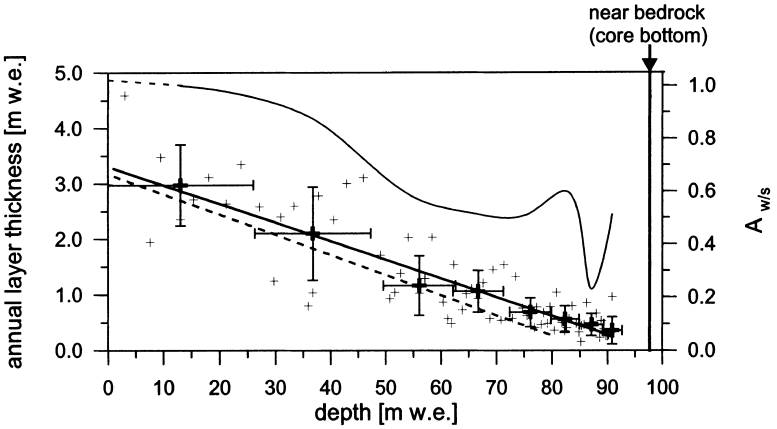

Fig. 9. Change of annual layer thickness along the C10 core: estimates via NH_4^+ stratigraphy are marked with thin crosses, corresponding 10 year means with thick crosses, and their regression with thick line. Results from the flow model of Vincent et al. (1997) are given as dashed line. Smoothed change of winter to summer snow deposition ratio $A_{w/s}$ is indicated by solid curve.

core increase of the mean annual impurity levels which is independent of temporal atmospheric changes. Using the summer (c_s) and winter (c_w) mean concentrations in recent snow of Table 3, the concentration enhancement $\Delta C(z)$ at a depth z with respect to the surface concentration $\overline{C(0)}$ can be estimated as:

$$\Delta C(z) = \overline{C(z)} - \overline{C(0)}, \quad (2)$$

$$\overline{C(z)} = f_w(z) \times c_w + f_s(z) \times c_s, \quad (3)$$

where $f_s(z)$ and $f_w(z)$ represent the summer and winter fractions of the annual layer at depth z, determined via the NH_4^+ stratigraphy, respectively.

Using eqs. (2) and (3), we estimated that (due to the increase of $f_s(z)$ and the decrease of $f_w(z)$ with depth) mean concentrations at a depth of 74 m w.e. would be enhanced by about 20% for NO_3^-, SO_4^{2-} and Ca^{2+} and by 30% for NH_4^+. Thus, such changes of the summer to winter amount of snow accumulation will systematically affect the general trends at the C10 site. Detailed investigations of the spatial variation of seasonal signals in the uppermost snow layers at Colle Gnifetti suggest, that this phenomena may introduce an artificial concentration increase in deep ice cores by up to 100% at this glacier site (Preunkert, 1994).

As seen in Fig. 9, the $A_{W/S}$ values decrease monotonically between the surface and 115 m depth (i.e., 87 m w.e.) apart from a partly recover between 101 and 110 m (i.e., 75 an 83 m w.e.). However, at a core depth of 115 m an abrupt increase of the $A_{w/s}$ values is observed. Provided that this pattern is not caused by an artefact related to a misinterpretation of the NH_4^+ stratigraphy, two possible explanations can be invoked. Such changes may be due to an abrupt increase of the winter snow contribution at approximately 200 m upstream the C10 borehole, or to a discontinuity of the ice flow trajectory starting points in this area, located $\sim$200 m upstream of the C10 borehole. The latter explanation is supported by unexpectedly high ^{210}Pb levels, found between 90 and 110 m depth in C10. The sudden increase of ^{210}Pb concentrations at these depths has been attributed to crevasses in the catchment area (Vincent et al., 1997), which may lead to this discontinuities in the ice flow. If correct, this phenomenon may have disturbed the C10 core chronology. Nevertheless, CH_4 measurements performed in the C11 core (drilled 30 m apart from the C10 site) give no evidence for a discontinuity in the core chronology at this glacier depth (Blanchard, 1998), although the ^{210}Pb profile is disturbed in this core as well. Finally, the comparison of the mean NH_4^+ summer level from 116–119 m deep ice layers from the Col du Dôme core C10 with the NH_4^+ signal incorporated in the glacier of Colle Gnifetti (see Subsection 6.2.) does not support the assumption of a discontinuity in the core chronology of C10.

7. Conclusions

Seasonally resolved measurements of major ions and δD have been performed along three ice cores extracted at the Col du Dôme glacier site. The hereby gained chemical depth records have been combined with the glaciological and glacio-meteorological properties of the glacier site, in view to exploit this glacier site for ice core investigations related to paleo-climatic and environmental issues.

Given the relative high net snow accumulation rate and the nearly complete winter snow contribution in the snow cover of Col du Dôme (as compared to corresponding values from the Colle Gnifetti glacier site), a detailed reconstruction of the mean seasonal cycle of chemical impurities and the δD content could be carried out. Mean seasonal cycles with flat winter minima and persistent high concentrations in summer are revealed for all chemical species and are particularly well-marked for NH_4^+, NO_3^-, and SO_4^{2-} (mean summer to winter concentration ratios ranging from about 4 for NO_3^- and SO_4^{2-} to 14 for NH_4^+). Although clearly visible, the Ca^{2+} seasonal cycle is more flickering due to sporadic Sahara dust inputs. In contrast to chemical species, the δD record shows more smoothed seasonal variations, exhibiting a similar pattern as the seasonal changes of air temperatures adopted from Grosser St. Bernhard (2472 m a.s.l., Switzerland).

Recent mean summer concentrations of the Col du Dôme site are in good agreement with the ones found at Colle Gnifetti 80 km apart, particularly for NH_4^+, NO_3^-, SO_4^{2-} and Ca^{2+}. In winter, a good agreement is again observed for NO_3^- and SO_4^{2-}, while NH_4^+ levels are higher at Colle Gnifetti than at Col du Dôme, probably due to a lack of winter snow deposition at Colle Gnifetti.

This comparison suggests that these two Alpine sites underlie similar atmospheric conditions year-round, at least for NH_4^+, NO_3^-, and SO_4^{2-}. This implies that for these species the two high alpine drilling sites are representative at least on a regional scale in the range of 100 km. Based on well-marked seasonal NH_4^+ cycles, in a 126 m deep Col du Dôme ice core is shown that a relative precise chronology (compared to the existing theoretical chronologies) can be established over the last 75 years (accuracy ± 5 years at an ice age of 75 years), and that snow deposits, made up by summer as well as by winter precipitation, are well preserved in the glacier over this time period.

In reconstructing long-term concentration records special attention has to be paid however to systematic changes in the snow deposition characteristics upstream to the drill site. This concerns the systematic spatial changes of the annual net snow accumulation (as linked to changing contributions by seasons) which may be mirrored in various ice core profiles. Such a glaciological forcing exerts a non-atmospheric influence on the shape of the seasonal cycle, and the mean annual levels of all species underlying a distinct seasonal variation. As a consequence, following findings have to be carefully considered in using the deep ice cores drilled at Col du Dôme in 1994: (1) the seasonal cycle of δD starts to become obliterated at a core depth of 115 m; (2) relative to the near surface regime, annual mean concentration levels of NH_4^+ are about 30% increased at a core depth of 100 m (the respective values for NO_3^-, SO_4^{2-} and Ca^{2+} are close to 20%).

In summary, the ice core drilled at Col du Dôme in 1994 will provide seasonally resolved glacio-chemical and isotopic records covering about the last 75 years. They are expected to contain information of past atmospheric conditions which are representative at least on a regional Alpine scale. This will especially help to fill the information lack, concerning long term Alpine ice core records, representative for atmospheric conditions prevailing during the winter season.

It is not clear yet, whether the Col du Dôme area provides the possibility to extend such seasonally resolved ice core records into the pre-industrial era. Current glaciological investigations include the search for an appropriate drill site to retrieve well preserved pre-industrial ice layers from Col du Dôme.

8. Acknowledgements

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