Snow accumulation and ice flow at Dôme du Goûter (4300 m), Mont Blanc, French Alps

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ABSTRACT. Glaciological experiments have been carried out at Dôme du Goûter (4300 m a.s.l.), Mont Blanc, in order to understand the flow of firn/ice in this high-altitude Alpine glacierized area. Accumulation measurements from stakes show a very strong spatial variability and an unusual feature of mass-balance fluctuations for the Alps, i.e. the snow accumulation does not show any seasonal patterns. Measured vertical velocities which should match with long-term mean mass balance are consistent with observed accumulations. Therefore, the measurement of vertical velocities seems a good way of quickly obtaining reliable mean accumulation values for several decades in such a region.

A simple flow model can be used to determine the main flowlines of the glacier and to propose snow/ice age of core samples from the two boreholes drilled down to the bedrock in June 1994. These results coincide with radioactivity measurements made to identify the well-known radioactive snow layers of 1963 and 1986. We can hope to obtain ice samples 55–60 years old about 20 or 30 m above the bedrock (110 m deep). Below, the deformation of the ice layers is too great to be dated accurately.

1. INTRODUCTION

Very few studies have been carried out on high Alpine glacierized areas (above 4000 m a.s.l.), for obvious reasons: access difficulties, cold temperatures and problems with altitude. Nowadays, these glaciers elicit an increased interest because they are seen to represent precious atmospheric archives and they seem to be suitable for climate and geochemical studies. Nevertheless, this kind of research requires knowledge of the firn/ice flow process, and at present this is limited by lack of observation and understanding of the following unusual conditions of the very high Alpine area:

The firn is very thick: it can reach more than 50% of the total thickness and, of course, influences the behaviour of the glacier flow.

The glacier is cold.

Surface conditions are unknown: melting rarely occurs at this altitude, but very little information about accumulation has been collected. For example, nothing is known about the seasonal pattern of mass balance.

Glaciological experiments have been performed at Dôme du Goûter, located at 4300 m a.s.l. on the way to the summit of Mont Blanc (Fig. 1), in order to better understand the flow of firn/ice in this area. The aim of this study is to establish the main flowlines and to compare the chronology obtained from a flow model with the results of radioactive measurements made in two deep boreholes. Sections 3–7 deal with the results of glaciological measurements required for the interpretation of ice cores, namely,

accumulation data from stakes;

velocity data from stake surveys;

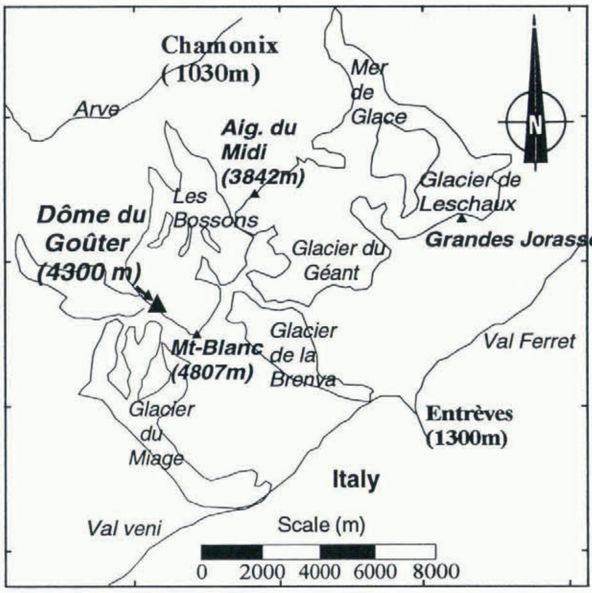

Fig. 1. Location of Dôme du Goûter, Mont Blanc area.

thickness data from radar measurements; and

snow/ice age for ice core samples, from artificial radioactivity methods.

Sections 8 and 9 present a simple firn/ice flow model for determining the snow/ice age of core samples down to the bedrock.

2. PREVIOUS STUDIES

The first studies of snow and glaciers in the vicinity of the Dôme du Goûter summit were carried out by J. Vallot at the end of the 19th century (Vallot, 1913). In 1973 and 1974,

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experiments were carried out by the Laboratoire de Glaciologie of Grenoble (Lliboutry and others, 1976) and showed the temperature pattern in the snow from the top of Mont Blanc down to Aiguille du Midi. In 1980, snow was cored 20 m deep at Col du Dôme, and the chronology of core samples was established from beta radioactivity measurements, deuterium and tritium content (Jouzel and others, 1984). In 1986, a deep ice core was drilled (70 m deep) near Col du Dôme for geochemical purposes (De Angelis and Gaudichet, 1991). In 1990, a surface snow geochemical study was performed at Col du Dôme and compared to others at very high-altitude sites (Maupetit, 1992). These studies provided some accumulation data; we have used only those for which the geographic positions of the observations are well known, because of the very strong spatial accumulation variability.

3. ACCUMULATION MEASUREMENTS

Twenty-six stakes (4 or 5 m long) were set up between 6 June 1993 and 22 June 1995 in the snow on Dôme du Goûter (see Fig. 4a), in order to measure both snow accumulation and surface velocity. Ablation is negligible at this altitude, and the annual mass balance can be assimilated to the accumulation. Because of large accumulation values, the stakes were replaced several times; moreover, this site is difficult to access without a helicopter, especially during winter, and a lot of stakes were lost under the snow cover. At each visit, a drilling core or a pit was dug (at stake 4, between the Dôme and the Col du Dôme) in order to determine the snow density near the surface (Fig. 2). For accumulation determination, we suppose that the bottom tip of the stake is attached to the snow; therefore, the accumulation is calculated as the difference in distance between the lower tip of the stake and the surface of the snow for two dates (in water equivalent). The accumulation measurements are summarised in Table 1, together with the precipitation measured in Chamonix.

The Dôme du Goûter summit is exposed to very strong winds, and observed accumulation (stake 1) is very low (about 45 cm w.e. a^{-1}). Because of strong winds in the south of this area (in the vicinity of the pass), accumulation values there (stakes 3 and 12) are low and erratic. In the northeast part, by contrast, one can observe very high accumulations

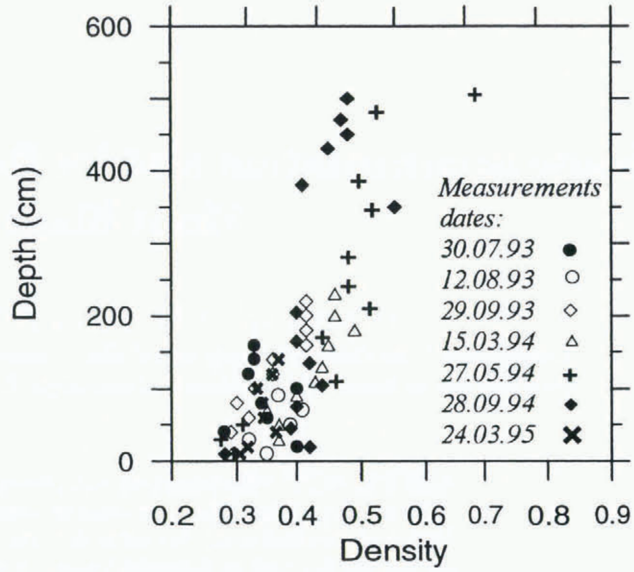

Fig. 2. Surface density measurements from pits and drilling core.

(2–4 m w.e. a^{-1}). Therefore, spatial variability is very important; accumulation can easily be multiplied by two or three, 100 m away. Another important result of these 2 years of observation is that mass balance does not show any seasonal pattern: summer and winter accumulations are very similar (like the precipitation data for Chamonix).

The observed accumulations have been compared with precipitation at Chamonix for the same period (Fig. 3). Stake measurements from areas exposed to strong winds are not taken into account, because they are not representative of local precipitation. Proportional functions between each stake measurement and the Chamonix precipitation are determined (1.3, 1.7 or 2.3; cf. Fig. 3) in order to estimate missing values in Table 1 and to calculate mass balance from 1 June to 31 May for 1993–94 and 1994–95 (Fig. 4b and c).

4. VERTICAL VELOCITIES AND LONG-TERM MASS BALANCES

Vertical velocities have been determined from topographic surveys (Table 2) and slope corrections. Vertical velocities were calculated from $ws = w - u \tan \alpha$, where u and w

Table 1. Observed accumulations at Dôme du Goûter (cm w.e.) and estimated annual mass balance (1 June–31 May; cm water)

	Observed accumulations							Estimated mass balance	
Stakes	*6 June–29 July 1993*	*29 Sept. 1993*	*15 March 1994*	*31 May 1994*	*28 Sept. 1994*	*24 March 1995*	*22 June 1995*	*1993–94*	*1994–95*
1 (Dôme)	31	26	7	0	0	0	27		27
2, 14	31	17	6	30	5	0	> 40	64	45
3	26	25	5	20	4	27	32	76	63
12	11	20	3	17	3			51	
4	46	60	76	25	67	>160	49	213	266
5, 22, 23, 24	52	42			71	>140		211	245
6, 7, 8, 20, 21	40	48			65	>160	242	269	
11	30	37	60	50	134	147		183	325
10, 15, 16, 17	39	>68			120	>150	>150	315	392
26 (Col)					74	92		181	200
9	29	24	84	34				175	224
19					112			314	371
Precip. Chamonix	21.7	27.7	55.4	31.6	53.6	91.9	32.2	136.4	177.7

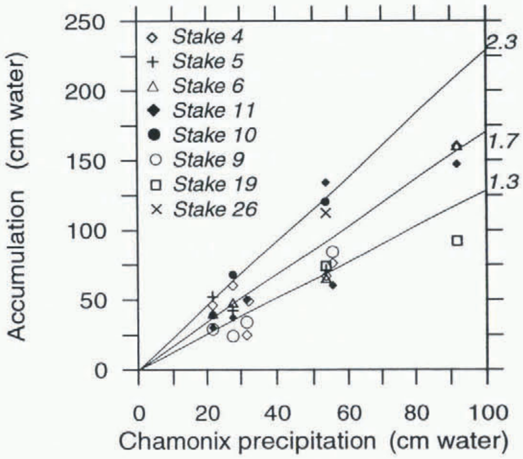

Fig. 3. Observed accumulation (cm water) at Dôme du Goûter with Chamonix precipitation (for different time lapse).

are the measured horizontal and vertical components of the surface velocity and $\tan\alpha$ is the surface slope (Paterson, 1981). These values as an independent result from direct accumulation measurements, expressed in m w.e., should match the long-term average mass balances if we suppose the Dôme glacier is in steady state (Fig. 4d). The spatial distribution of the vertical velocities data seems to be in good agreement with observed accumulations during 1993–94 and 1994–95 (Fig. 4b and c). Annual mean Chamonix precipitation is 1.25 m w.e. between 1959 and 1995. In 1993–94, we can see that Chamonix precipitation (1.36 m w.e. a^{-1}) is very close to this mean value, and the Dôme's accumulations are also close to the vertical velocities (except for stake 26 at Col du Dôme and stake 5). By the way, although vertical velocities require positioning measurements and slope determination, they seem to provide the best way of easily obtaining reliable mean accumulation values for several decades.

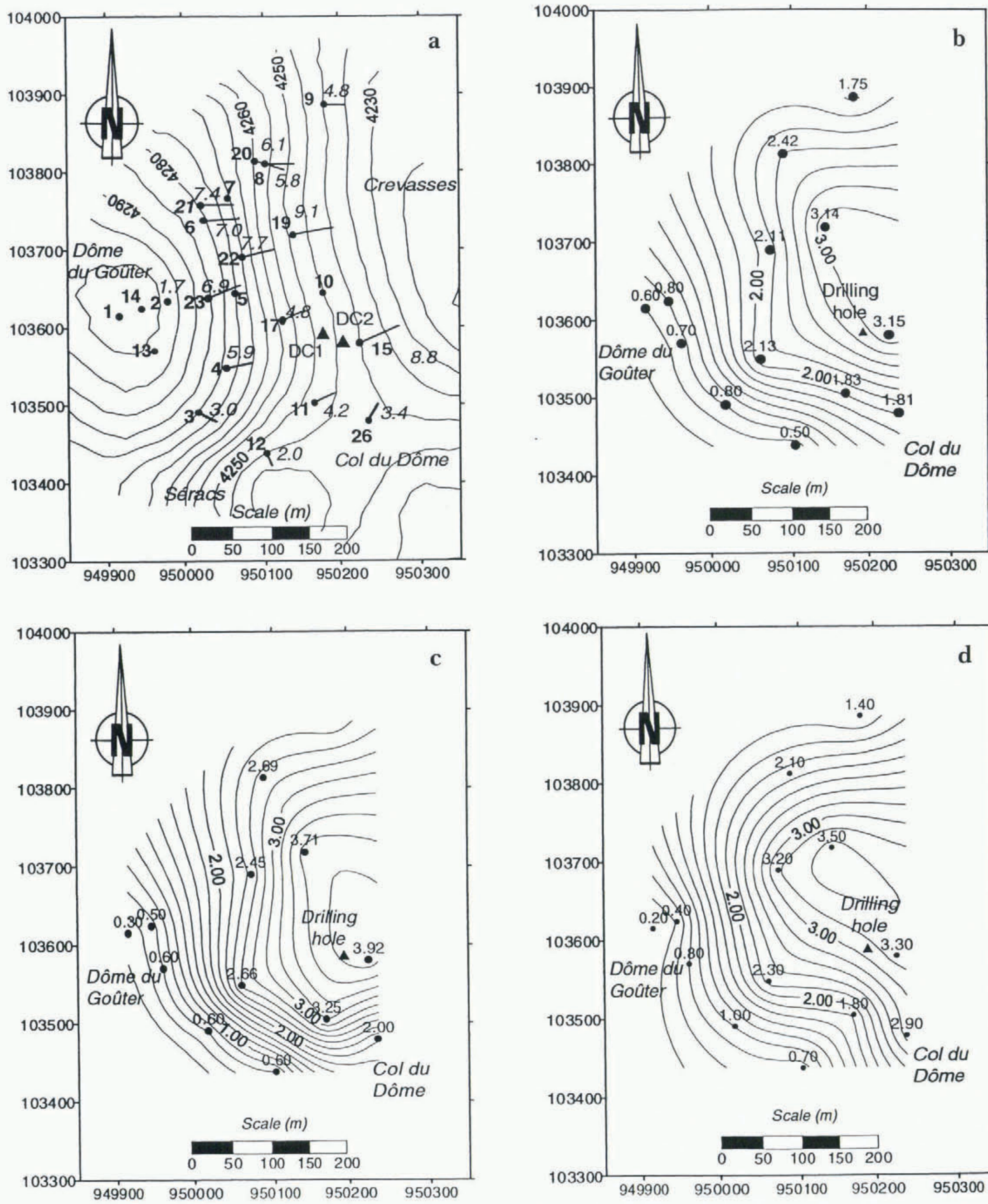

Fig. 4. (a) Surface topography, stakes, and annual horizontal velocities (m a^{-1}) in italics. (b) Observed accumulation, 1993–94 (m w.e. a^{-1}). (c) Observed accumulation, 1994–95 (m w.e. a^{-1}). (d) Measured vertical velocities (m w.e. a^{-1}).

Table 2. Horizontal and vertical velocities; stakes were observed with topographic means on 6 June 1993, 29 September 1993 and 31 May 1994. (Two or three values are sometimes given for different periods observed, in order to show the scattering of the values)

Stake	Annual horizontal velocities m year^{-1}	Annual vertical velocities m water year^{-1}
1		0.06, 0.2, 0.4
2	1.7, 2.4, 1.7	0.9, 1.0, 0.9
3	3.0, 2.0, 3.4	1.1, 0.65, 1.0
4	6.2 , 5.7	2.3, 2.2, 3.2
6	7.0	2.7
8	6.1	1.9
9	4.8	1.4
11	4.3, 4.0	1.8, 1.8
12	2.00, 1.50, 2.4	0.7, 0.6, 0.7
13		0.8
14		0.4
15, 16	8.7, 8.9	3.3, 3.3
17, 18	4.8	3.9, 4.4
19	9.1	3.5
20	5.8	2.2
21	7.4	2.8
22	7.7	3.5
23, 24	6.9	3.2
26	3.4	2.9

5. HORIZONTAL VELOCITIES (Fig. 4a)

Horizontal velocities have been estimated from the position of the bottom tip of these stakes, considering tilt and orientation of the stakes (two points have been measured along each stake at each observation). Because of the slope and the strong creep of snow, stakes tilt with time, and observed tilt can reach a deviation of 20° from the vertical in steep slopes. Figure 4a shows that horizontal velocities are more or less perpendicular to contour lines. Furthermore, the observed tilts of the stakes after a few months suggest that a large part of horizontal velocities is attributable to firn deformation (about 80 m thick at the drillhole site shown in Figure 4a).

6. RADIOACTIVE DATING OF ICE CORES

The radioactive fallout from atmospheric thermonuclear tests (Picciotto and Wilgain, 1963), conducted mainly in 1954 and 1961–62, and from the Chernobyl accident in 1986 (Pourchet and others, 1988) provide well-known radioactive levels in glaciers, and enable an absolute dating. Ice cores have been drilled in the Dôme du Goûter area since 1973 (see Table 4). After density measurements, the snow samples were melted and filtered (Delmas and Pourchet, 1977), then analyzed in the laboratory for global-beta (Pinglot and Pourchet, 1979) and gamma radioactivity (Pinglot and Pourchet, 1994).

6.1. Spatial accumulation variation

Analysis of samples from ice cores drilled in 1976, about 200 m southeast of stake 26 at Col du Dôme (P2), showed a mean annual accumulation (MAA) of about 0.34 m w.e. a^{-1} (1970–76), with a large annual variability by a factor of 2 (personal communication from M. Pourchet, 1995). A 20 m ice core (P11) was taken out in 1980 precisely at the saddle of Col du Dôme (Jouzel and others, 1984), close to stake 26. The MAA was 1.09 ± 0.23 m w.e. a^{-1} (1970–79). In the neighbourhood of Col du Dôme, a MAA of 1.75 m w.e. a^{-1} for 1988–90 was determined (Maupetit and others, 1995). In 1993, shallow ice cores were also taken for Chernobyl determination: at the summit of Dôme du Goûter (P20), the MAA is 0.61 m w.e. a^{-1} (1986–93), as compared with 0.45 m w.e. a^{-1} (1993–95) measured with stakes.

6.2. Effect of wind on ^{137}Cs redistribution

Strong winds scour the Dôme summit and remove snow continually. The ^{137}Cs from Chernobyl deposits clearly confirms this fact. An intermediate ice core (P15, at the saddle of Col du Dôme; 4250 m; December 1986) showed a ^{137}Cs deposit of 538 Bq m^{-2} (Pourchet and others, 1988) from Chernobyl. At the Dôme summit the remaining deposit is as low as 10 Bq m^{-2}, while at drilling site 2 it is 3000 Bq m^{-2}. However, for drilling site 1 (30 m away), the Chernobyl deposit is only 25 Bq m^{-2}, demonstrating the effect of wind scouring on short-duration events like Chernobyl. The ^{137}Cs from thermonuclear-test fallout (at time of deposition, in 1963) is 3000 Bq m^{-2} for both ice cores. As can be seen, the ^{137}Cs fallout from Chernobyl can be as large as from the nuclear tests (ice core 2).

6.3. Deep ice cores

In 1986, a deep ice core was drilled (70 m; P14) but without an accurate geographic position: its location is at least 100 m away from Col du Dôme, not very far from the 1994 deep drilling cores. The 1963 thermonuclear-tests deposit was detected at 55.7 m depth. As previously determined in this 1986 ice core (Pinglot and Pourchet, 1995), the thermonuclear-tests layer presents two distinct maxima, in 1962 and 1963 (Table 3). In 1994, the ^{137}Cs radioactive layers (1962–63 and 1986 (Chernobyl)) and ^{210}Pb content were determined in both deep ice cores 1 and 2 (Table 3). As will be seen later, it is impossible to neglect the thinning effect (due to vertical strain rate) and the deposit surface origin of deep layers. Interpretation of these ice cores' dating results requires flow-model computations.

6.4. ^{210}Pb profiles in deep ice cores

^{210}Pb (a natural isotope with 22.3 years half-period from ^{238}U to ^{226}Ra and ^{222}Rn) was jointly determined with ^{137}Cs by gamma spectrometry (Fig. 5). Both profiles (ice cores 1

Table 3. Depths of layers resulting from atmospheric nuclear tests (1954–55 and 1962–63) and Chernobyl (1986) in deep ice cores

Drilling site	Drilling date	Deposition date	Depth Minimum m	Depth Maximum m	^{137}Cs deposition Nuclear tests Bq m^{-2}	^{137}Cs deposition Chernobyl Bq m^{-2}
P14	1986	1955	69.07	69.39	2014	
		1962	57.39	57.54		
		1963	55.67	55.83		
Core 1	1994	1963	85.95	87.67	3000	
		1986	40.09	41.02		25
Core 2	1994	1954	103.63	104.51	3005	
		1963	90.28	91.16		
		1986	38.61	39.60		3000

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Table 4. Mean annual accumulation and ^{137}Cs fallout from Chernobyl. (Since the drilling holes are shallow, the thinning effect due to vertical strain rate is neglected)

Stake drilling site	*Budget years*	*Annual accumulation* m w.e.	*^{137}Cs deposition* Bq m^{-2}	*Source*
Col du Dôme				
P2	1970–76	0.34		M. Pourchet (personal communication, 1995)
P11	1970–79	1.09		Jouzel and others (1984)
P15	June–Dec. 1986	0.96	538	Pourchet and others (1988)
Near to P11–P15	1988–90	1.75		Maupetit and others (1995)
Dôme summit				
P7	1970–76	0.33		M. Pourchet (personal communication)
P20	1986–93	0.61	10	
Dôme				
P21	1986–93	1.62		
P10 Mont Blanc summit	1970–73	2.80		Jouzel and others (1977)

and 2 in 1994) exhibit three different states. From the surface down to about 85 m, ^{210}Pb decreases, as expected. But there is a strong ^{210}Pb deposition increase roughly by a factor of 6 for ice layers under 85 m, in both ice cores.

These increases are definitely not due to thermonuclear tests. The disturbed ^{210}Pb signal may be due to the presence of crevasses near the surface-deposit origin point. As much as 28 Bq kg^{-1} of ^{210}Pb concentration was measured on an ice core (P7, near the summit) exposed to in-depth crevasses.

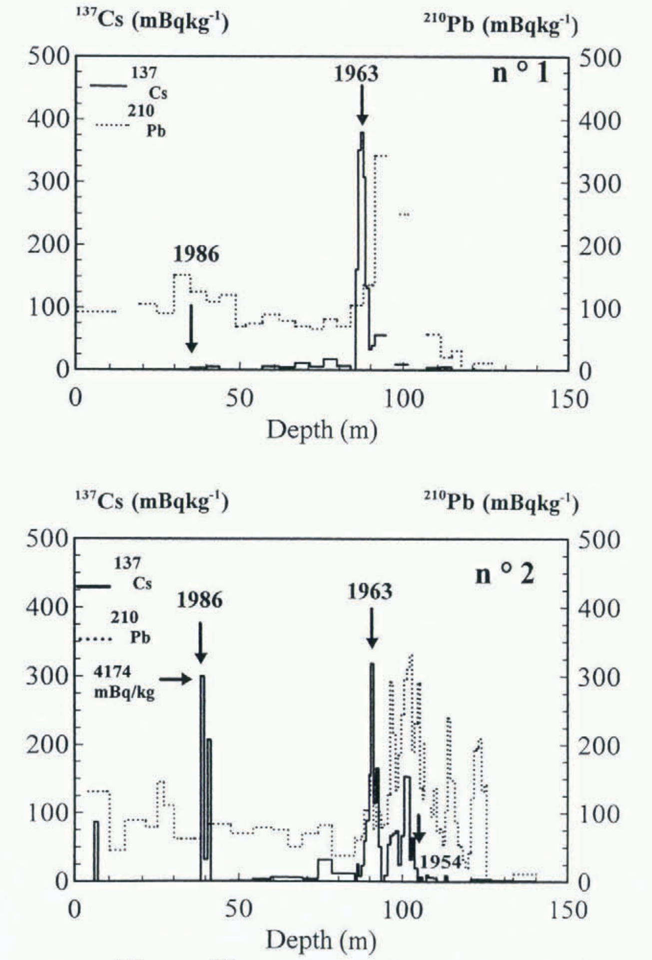


Fig. 5. ^{137}Cs and ^{210}Pb activities at Col du Dôme du Goûter (ice cores 1 and 2; 1994), vs depth.

7. RADAR MEASUREMENTS

Radio-echo soundings were made from 1 to 5 June 1993, along eight profiles. Four additional profiles were measured in June 1994 where determination of the glacier bed was not possible with the measurements of the first campaign. We used radar devices, which were built following the designs of Jones and others (1989) and Wright and others (1990), i.e. with two output stages which generate inverse pulses. Changes were made in the layout of the free-running pulse generator and in choosing the power devices (Funk and others, 1995). The speed of electromagnetic wave propagation in the ice (in our case, cold ice) has been assumed to be identical to the value found at Colle Gnifetti (Monte Rosa, Valais, Switzerland), 175 m μs^{-1} (Wagner, 1994).

In the case of Dôme du Goûter, the determination of the bed topography from the radio-echo travel time is a problem which is much more delicate than for a plane ice shield, because the glacier bed is very rough in the investigated area, with relatively deep, short and narrow valleys. The field measurements were performed in such a way as to obtain reflections from the glacier bed situated more or less in a vertical plane with the measurement points at the glacier surface, allowing the determination of the glacier bed in two dimensions. The surface of the glacier bed was constructed as an envelope of all ellipse functions, which give all the possible reflection positions to a certain travel time between sending and receiving antennae.

Interpretation of the data obtained was difficult because of multiple reflections from the bottom, characteristic of a rough glacier bed (Fabri, unpublished). The radio-echo sounding profiles intersect at several points, allowing results to be checked. The resulting bed geometry is shown in Figure 6.

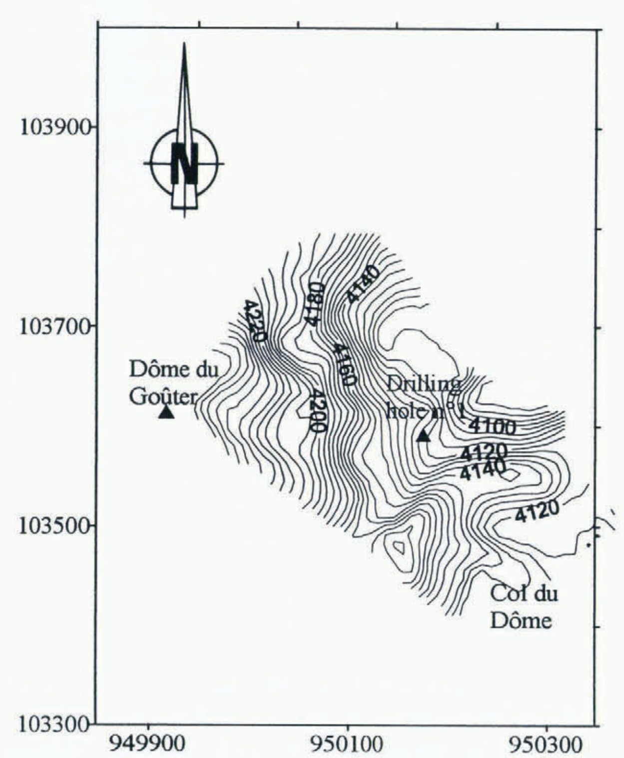
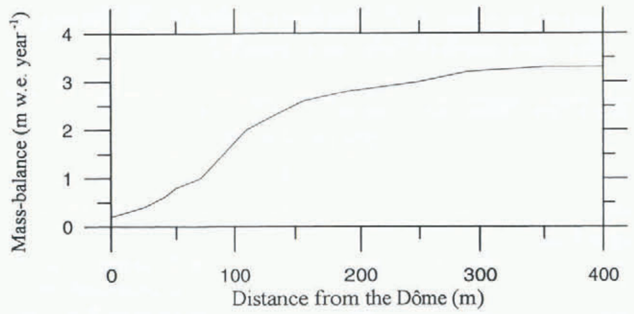


Fig. 6. Bed topography from radar measurements.

The accuracy of the calculated ice thickness is determined, in part, by the accuracy of the measurement of the time delays and the antenna spacing. Additional errors may arise due to neglect of the passage of the bottom return through a firn layer. If the firn layer is 20 m thick this may give rise to 3–4 m error in the depth estimate. Other errors may arise because the smooth envelope of the reflection ellipses is only a minimal profile for a narrow, deep valley-shape bed topography, with the result that the ellipse equation will be governed by an arrival from a reflector situated toward the side, and thus not directly beneath the point of observation. One can easily imagine a glacier bed geometry for which no density of observation will yield the correct depth near the centre of a valley-shape bed topography. Further errors may be introduced by assuming that all reflection points lie in the plane of the profile rather than an ellipsoid. These errors may be important in our case, because the bed topography changes rapidly in all directions. The two boreholes reached the glacier bed at 126 and 140 m. The ice thickness as derived from the radio-echo soundings at the same sites is 122 and 130 m, respectively. The agreement is seen to be excellent, 4 and 10 m difference between the radio-echo and borehole depths.



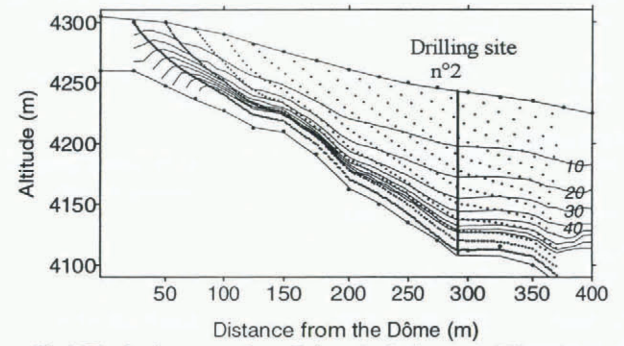


Fig. 7. Vertical transect from Dôme du Goûter to drilling site 2 with flowlines (dash) and isochrones.

8. FLOW MODEL AND AGE OF ICE AS FUNCTION OF DEPTH

Artificial radioactivity measurements allow us to identify with confidence the 1986 and 1963 snow layers. Nevertheless, it is difficult to determine the age of the lower part of the core and to estimate the thickness of annual layers. For this purpose and for the determination of the mean thickness of each annual layer down to the bedrock, a flow model can provide useful information about the age of the firn/ice of the lower part of the ice core. Gagliardini (1995) developed a two-dimensional flow model based on mechanical concepts. Here, a simple two-dimensional flow model was developed assuming that the glacier thickness has not changed significantly since the beginning of the 20th century. This assumption is supported by the dynamic behaviour of the glacier (vertical velocity measurements) which has been shown to be close to steady state. In the model, the mass conservation is considered, in order to obtain the balance horizontal velocity. The mass continuity equation is solved:

$$\partial\left[\frac{u_{\mathrm{m}}H(x)}{\partial x}\right] = b(x) - \frac{\partial[H(x)]}{\partial t}$$

where $H(x)$ is thickness, $b(x)$ is mass balance (m water a^{-1}), x is distance from the Dôme (m), u_{m} is mean velocity, and $\partial[H(x)]/\partial t$ is the thickness variation with time assumed to be 0.

From this equation, the mean balance velocity can be computed for each section. The depth profile of horizontal velocity is derived from the analytical model given by Lliboutry (1981):

$$u(z,H) = \frac{n+2}{n+1}u_{\mathrm{m}}\left[1-\left(\frac{z}{H}\right)^{n+1}\right]$$

where $u(z,H)$ is the horizontal velocity at depth z, H is the glacier thickness and n is the Glen flow-law exponent ($n=3$). The ice temperature is −11°C near the bedrock (personal communication from C. Rado, 1995) and the sliding velocity is assumed to be 0. As a result, the mean horizontal velocity is equal to 80% of the surface horizontal velocity. The vertical velocity $w(x,z)$ is calculated using mass continuity for the local element:

$$\frac{\partial u}{\partial x} + \frac{\partial w}{\partial z} = 0\,.$$

This gives (Ritz, 1987):

$$w[x,H(x)] = b(x)\left\{1 - \frac{1}{H(x)}\int^{H}\frac{5}{4}\left[\frac{1-z^4}{H(x)^4}\right]\mathrm{d}z\right\} + u(x,z)\frac{z}{H(x)}\frac{\partial H(x)}{\partial x} - \frac{\partial E(x)}{\partial x}$$

where the last term, from $u(x,z)$ to the end of the equation, is called kinematic vertical velocity, $b(x)$ is mass balance (m water a^{-1}), z is depth (m), $H(x)$ is thickness (m), $u(x,z)$ is horizontal velocity, $\partial H(x)/\partial x$ is thickness variation, $\partial E(x)/\partial x$ is altitude surface variation, and $w[x,H(x)] = b(x)$ at surface and 0 at bedrock.

Horizontal and vertical velocities expressed in m water a^{-1} are then converted by the following relations:

$$u_{\mathrm{s}} = \frac{u_{\mathrm{w}}}{\rho}\,, \qquad w_{\mathrm{s}} = \frac{w_{\mathrm{w}}}{\rho}$$

where u_{w} is horizontal velocity in m water a^{-1}, u_{s} is horizontal velocity in m snow or ice a^{-1}, w_{w} is vertical velocity in m water a^{-1}, w_{s} is vertical velocity in m snow or ice a^{-1}, and

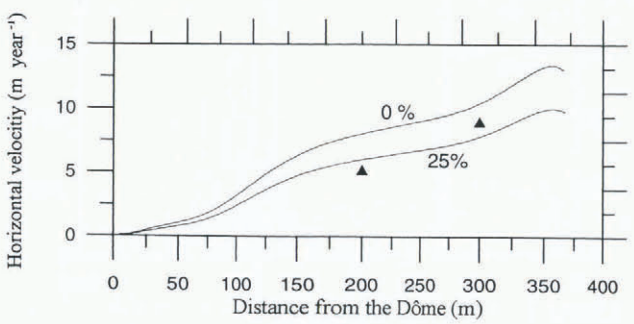


Fig. 8. Surface horizontal velocities from the model results (divergence of 0% and 25%) and measurements.

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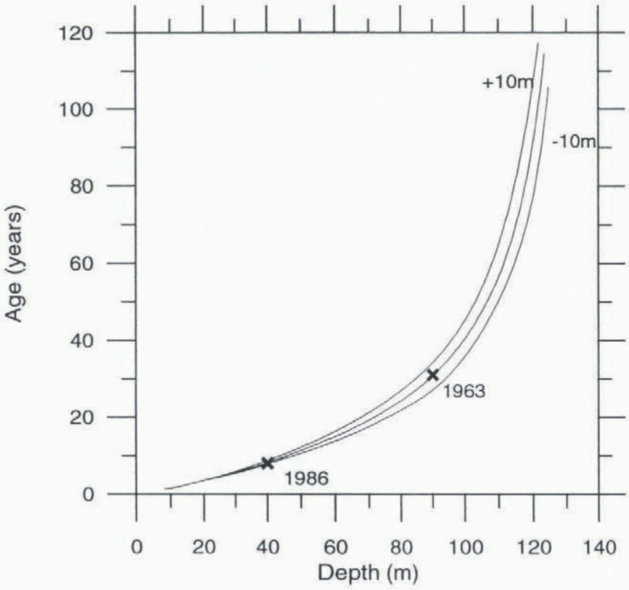

Fig. 9. Snow age vs depth from model results and radioactivity measurements (1986 and 1963). Sensitivity to thickness variation (±10 m).

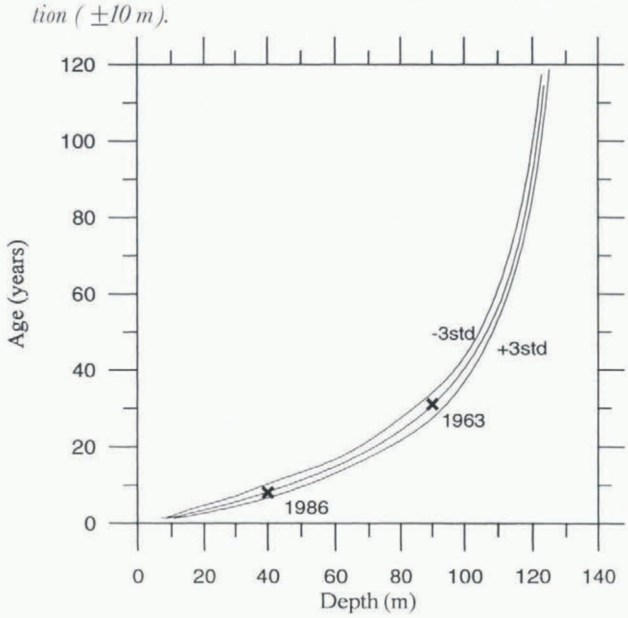

Fig. 10. Snow age vs depth from model results and radioactivity measurements (1986 and 1963). Sensitivity to mass-balance variation (±3 standard deviation).

ρ is local density. Density measured in the borehole is assumed to be representative for the whole firn.

Such a model does not take into account the mass-balance change with time; it considers the mean mass balance obtained from vertical velocities (Fig. 4d).

This model takes into account horizontal flowline divergence by corrections on flux along the main flowline. Nevertheless, measured horizontal velocity directions are almost parallel (Fig. 4a) and horizontal flowline divergence is neglected at first.

Finally, calculated velocity vectors allow us to determine flowlines (Fig. 7) and the ice age from Dôme du Goûter to the drill site. Horizontal velocity values are shown in Figure 8 and have been compared with surface measurements.

9. RESULTS OF FLOW MODEL

The sensitivity of the ice age proposed for ice core 2 has

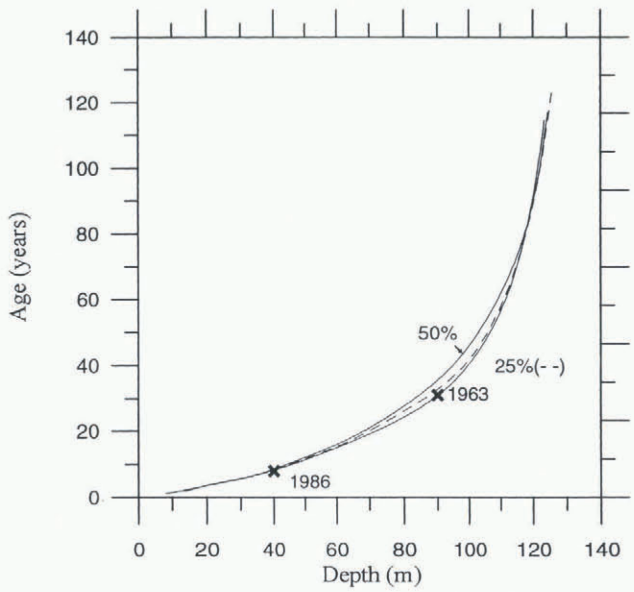

Fig. 11. Snow age vs depth from model results and radioactivity measurements (1986 and 1963). Sensitivity to divergence (50% and 25%).

been studied for some parameters. Figure 9 shows the sensitivity to thickness variation; a variation of bedrock elevation (±10 m), applied on the profile from the Dôme to the drilling core, involves an uncertainty of ±6 years at 110 m depth. Figure 10 shows the sensitivity to the mass-balance variation; for this purpose, the standard deviation (20 cm w.e.) of the Chamonix precipitation is multiplied by 3 to obtain an estimation of the standard deviation ($\sigma =$ 60 cm w.e.) of the mass balance at Dôme du Goûter (according to the results of accumulation measurements). Therefore, uncertainties equal to $3\sigma/\sqrt{n}$, where n is the number of years from now, were introduced in the model, and the age of the ice core has been obtained from these calculations (Fig. 10).

Sensitivity to the horizontal divergence of the flowlines is shown in Figure 11. A divergence value of 50% (discharge/2) gives results similar to the results without divergence. That means that the model is not very sensitive to horizontal velocity modifications. On the other hand, Figures 12–14 confirm the sensitivity of the model to the vertical

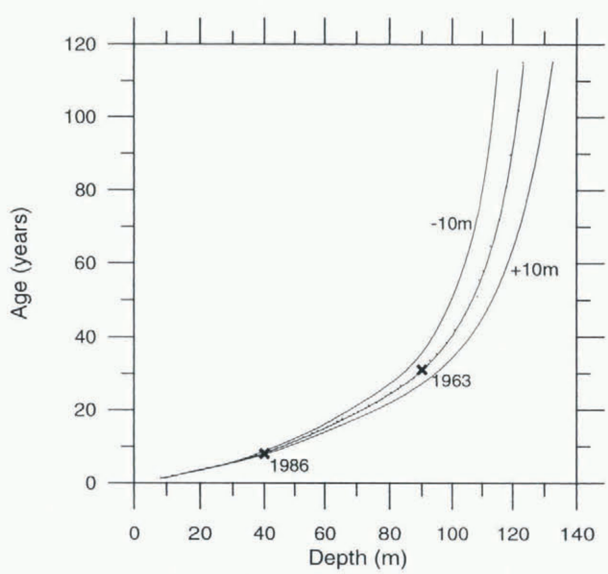

Fig. 12. Snow age vs depth from model results. Sensitivity to the depth of drilling hole (±10 m).

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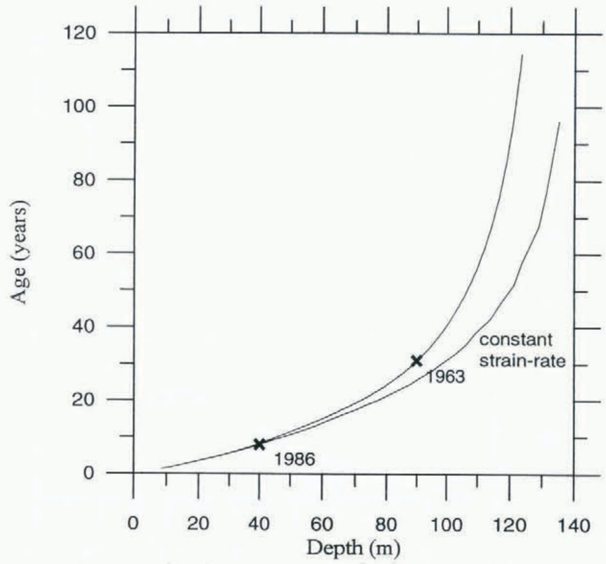

Fig. 13. Snow age vs depth from model results and radioactivity measurements (1986 and 1963). Sensitivity to vertical strain rate (constant and non-constant strain rate).

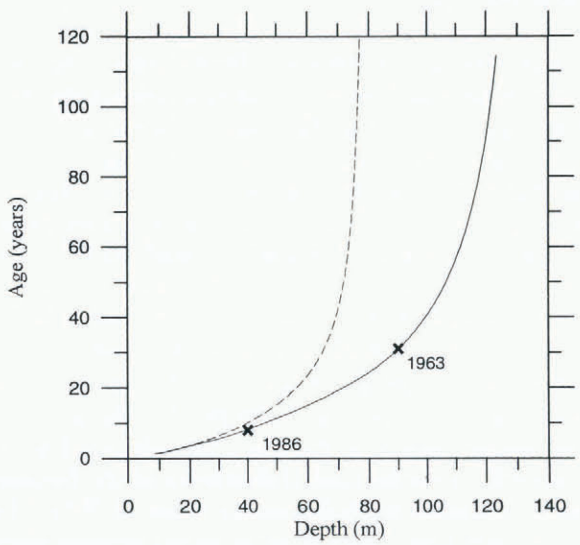

Fig. 14. Snow age vs depth from model results and radioactivity measurements (1986 and 1963). The dashed line does not take into account the kinematic vertical velocity.

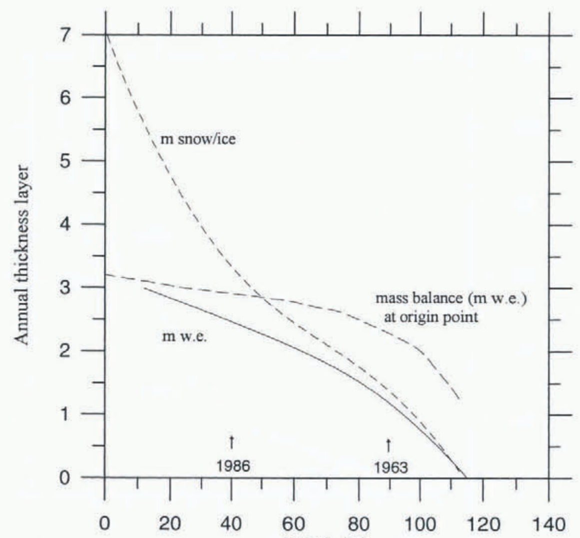

Fig. 15. Annual thickness layer vs depth from the model results in the drillhole, and mass balance at the origin point of the layer.

velocity pattern. First, the sensitivity of the model was studied according to glacier thickness at the site of the drilling core (Fig. 12). Since the model forces the bottom flowlines to follow the bedrock whatever the profile of the bedrock, the thickness at the drilling site directly influences the vertical velocity at this site. This effect shows the limits of such models.

Furthermore, it can be seen that the model is very sensitive to the vertical strain rate. A constant vertical strain rate, for which vertical velocity is $b(1-z)/H$, where b is the mass balance, z is the depth, and H is the thickness, gives very different results (Fig. 13).

Finally, the effect of the kinematic vertical velocity (see section 8) was analyzed (this term allows the bottom flowlines to follow the bedrock). When this parameter is deleted, the model indicates an age of the ice core which is older than inferred from radioactivity measurements (Fig. 14). Thus, it seems that the calculated age is not very sensitive to uncertainties on surface conditions (mass-balance variation) and bottom conditions (bedrock topography uncertainty) between Dôme du Goûter and the drilling site (that does not mean a good knowledge of absolute values of these parameters is not required). However, the parameters linked to the vertical velocity at the drilling site are predominant: vertical strain rate, thickness and mass balance at the drilling site.

Figure 8 allows us to compare the horizontal velocities obtained from the model and the measured velocities. Although the uncertainties about measured velocities are large, it can be seen that the calculated velocities are too high. These differences can be explained by horizontal divergence of the flowlines (about 20–30%) without changing the ice age (Fig. 11). In Figure 15, the mean thickness of annual layers in the drillhole is calculated according to depth. Below 110 m, layers are very thin. In the same figure, the mass balance at the beginning of the flowline is shown. This offers the possibility of distinguishing the parameters which lead to thinning of layers: the deformation of ice layers due to the strain rate, and the spatial distribution of mass balance between the Dôme and the drilling site.

10. CONCLUSIONS

The glaciological surveys carried out on Dôme du Goûter show an important spatial distribution of the accumulation: the mean accumulation varies from 0.3 m w.e. a^{-1} at the top of the Dôme to 3.0 m w.e. a^{-1} in the vicinity of the drilling cores. In addition, an unusual feature of mass-balance fluctuations in the Alps has been pointed out: the snow accumulation in this high-altitude Alpine glacierized area does not show any seasonal pattern; summer and winter mass balances are very similar, which has not often been observed in the Alps. Furthermore, the observed spatial distribution of the accumulation seems to be very similar to the vertical velocities obtained from topographic measurements of the stakes. These values, as an independent result from direct accumulation measurements, match the long-term average mass balances for steady-state conditions. Therefore, the measurement of vertical velocities is a good way of quickly obtaining the reliable mean accumulation values for several decades in such a region.

The flow model described in this paper provides useful information about the firn/ice cores made down to the bedrock. It can be shown that, with a simple flow model and

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some precautions (for example, taking into account the kinematic vertical velocity connected with the bedrock shape), a reliable age/depth relation can be obtained, a result which is supported by radioactivity measurements (1963 and 1986 events).

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