

The Equilibrium Zone on Polar Glaciers

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ABSTRACT

The equilibrium line, connecting points at which glacier mass balance b is equal to zero, is never an isohypse or line at constant altitude. This simple observation casts doubt on the meaning of the equilibrium line *altitude*. The equilibrium line altitude is, however, definable precisely in algebraic or graphical terms, given a plot of mass balance as a function of altitude. It is that altitude at which the fitted or plotted mass balance is zero. The equilibrium line, according to this definition, is an isohypse, but on the glacier surface the spatial distribution of mass balance can be, and generally is, much more complicated than this simple picture suggests. The equilibrium line need not even be simply connected; that is, outliers of accumulation and inliers of ablation can and do occur. We present information from the White Glacier, a valley glacier on Axel Heiberg Island, showing that it is preferable to think in terms of an equilibrium *zone* rather than an equilibrium *line*. The equilibrium zone, defined as that part of the glacier between the lowest altitude at which $b > 0$ and the highest altitude at which $b < 0$, may occupy a vertical range of hundreds of metres. On the White Glacier, the equilibrium zone may occupy ~25 percent of the surface in a typical year, and up to 50 percent of the surface when the equilibrium line altitude is close to the maximum of the glacier's hypsometric curve. Processes within the equilibrium zone require detailed attention if goals such as the accurate remote sensing of mass balance are to be attained.

Key words: Equilibrium line, glaciers, mass balance.

INTRODUCTION

This paper is a commentary on two terms that are widely used in glacier mass balance work. These are *equilibrium line altitude* and *equilibrium line*. The focus here is on polar glaciers, using the White Glacier, Axel Heiberg Island, N.W.T., Canada (Cogley *et al.* 1995) as an example.

It is common in the glacier mass balance literature to use these two terms interchangeably, assuming that they are effectively the same. The equilibrium line altitude, h_0 , is the altitude on the glacier at which, using procedures such as those described below, ablation is estimated to equal accumulation at the end of a balance year. This is a key concept in the calculation and estimation of glacier mass balance (Figure 1). It is used to divide the glacier into the accumulation and ablation zones. Glacier mass balance correlates well with h_0 (e.g. Koerner 1970; Østrem 1975; Braithwaite 1984; Hagen and Liestøl 1990; Cogley *et al.* 1995). Thus where a multi-year mass balance record is available, the mass balance of a year in which measurements are not made can be estimated if a realistic equilibrium line altitude can be identified. The equilibrium line, on the other hand, is the actual boundary between areas of accumulation and ablation, as they occur on the glacier at the end of a summer. For various reasons that are discussed below, this is rarely, if ever, a simple line of equal elevation.

We submit that appreciation of the precise meaning of these two terms is important as glacier mass balance work develops — especially with respect to remote sensing of glacier properties and to understanding of accumulation and ablation processes.

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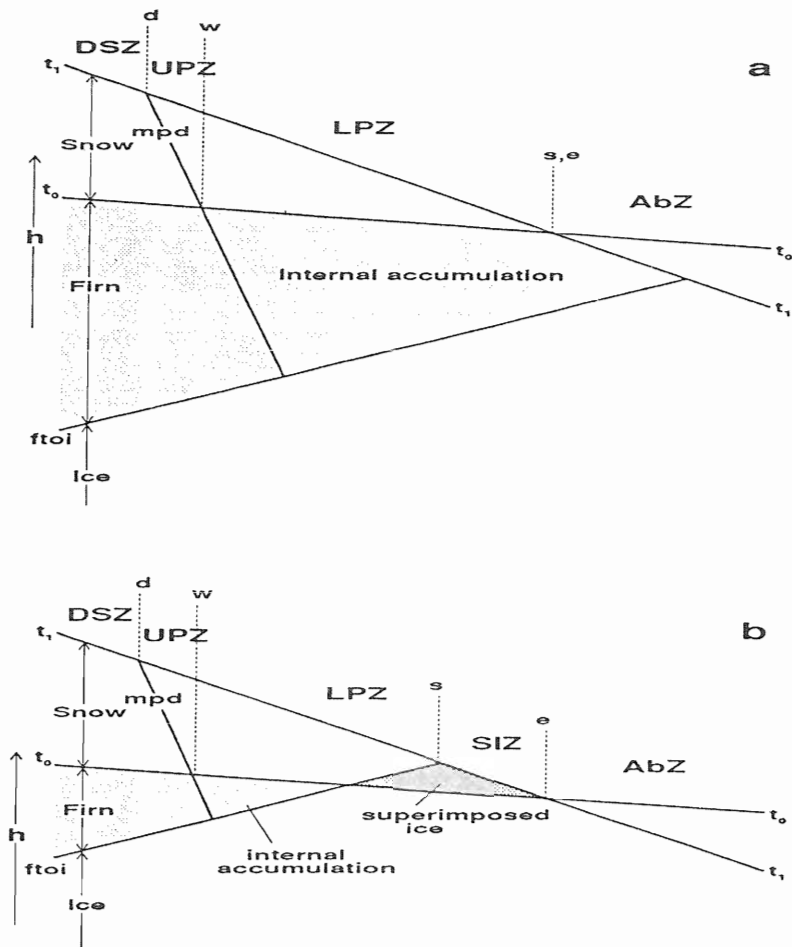


Figure 1. Concepts of glacier zonation, adapted from Benson (1959) and Müller (1962). a. The common situation on temperate glaciers. t_0 represents the glacier surface in profile at the start of a balance year; t_1 represents the surface at the end of the year. t_0 and t_1 intersect at the equilibrium line e, which is also the snowline, s. The line labelled mpd, which may outcrop at the surface as the dry snow line d, is the maximum depth to which snow meltwater (melted at the surface) percolates before refreezing or before encountering the effectively impermeable barrier ftoi. ftoi is the boundary between firn and glacier ice. The conventional zones are: DSZ, the dry snow zone (generally absent on all glaciers but the large ice sheets); UPZ and LPZ, the upper and lower percolation zones (separated by the wet snow line, w, which has no surface expression); and AbZ, the ablation zone. Internal accumulation, by the refreezing of meltwater below t_0 , may occur in the wedge $t_0 - t_1 - ftoi - mpd$. b. The ideas of case a adapted to a situation which is common on polar glaciers. Case b differs from case a only in that the surface ftoi is shifted up-glacier. In consequence ice (not snow) added to the glacier during the balance year is exposed on the surface t_1 between s and e, which no longer coincide. (This ice may be derived either from refrozen percolating meltwater or from surface transfer by slush avalanches; cf. Müller 1962.) s and e define the superimposed ice zone SIZ.

MASS BALANCE PROCEDURES

Mass balance measurements using stakes are a standard way of monitoring changes in the mass of glaciers (Østrem and Brugman 1991). They involve tracking mass receipts (accumulation, b_+)

and losses (ablation, b_-), usually on an annual basis. Key concepts and terms of the most common, stratigraphic, approach to this are shown in Figure 2 (cf. Figure 4 of Østrem and Brugman 1991).

Measurements of b_+ and b_- are made at points distributed over the glacier surface.

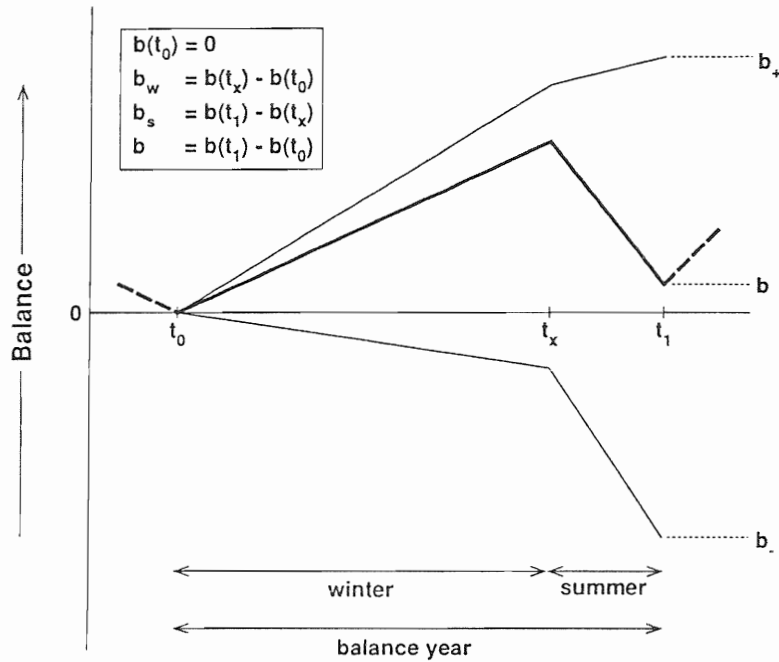


Figure 2. Mass balance at a point on a glacier as a function of time during a balance year. The (stratigraphic) year begins at t_0 , generally in the autumn, when the total mass is at its (– annual) minimum and $b(t)$ is defined to be zero. Accumulation b_+ and ablation b_- change throughout the year, but their rates usually differ sharply between winter and summer. Winter ends and summer begins at time t_x , when $b = b_+ + b_-$ is at its (– annual) maximum. The balance year ends at t_1 , when the net balance for the year is recorded as $b \equiv b(t_1)$ and both b and t may be reset to zero. The winter balance b_w and summer balance b_s are defined in the inset.

Where $b_+ > b_-$, the glacier is said to have a positive mass balance (it has increased in mass); where $b_+ < b_-$, the glacier has a negative mass balance. However, it is not necessary to measure b_+ and b_- separately. The annual net balance b can be estimated by tracking the change of elevation of the glacier surface, usually with respect to the top of a stake, successive readings being made one year apart. (Of course the density of the substance gained or lost must also be estimated.)

The net balance B in units of mass per unit area per unit time for the whole glacier is, by definition,

$$B = \int_A b \, dx \, dy / \int_A dx \, dy, \quad (1)$$

where A is the area of the glacier and $b \equiv b(x,y,\Delta t)$ represents the mass balance at each point on the glacier over the time span $\Delta t = t_1 - t_0$.

As a basis for extrapolating from the measurement sites to the rest of the glacier surface, it is usual to assume that elevation controls accumulation and ablation. Thus extrapolation from point measurements is done only in the vertical direction. To operationalize this, measurements are

grouped within elevation bands to obtain a band average, $b(h)$. This reduces the bias from the possibly uneven distribution of measuring sites with elevation and permits the weighting of different elevation bands by their areas $a(h)$ (Cogley *et al.* 1995; Østrem and Brugman 1991). It is generally desirable, as an intermediate step, to estimate the point mass balance as a continuous function of elevation, $\beta(h)$ say. The relationship of $\beta(h)$ to $b(h)$ and to stake measurements of b is illustrated in Figure 3.

Thus in operational terms the whole-glacier mass balance may be expressed as

$$B = \int_H a(h) \beta(h) \, dh / \int_H a(h) \, dh, \quad (2)$$

where the integrals range from the minimum to the maximum elevation of the glacier. By means of this procedure, the glacier is divided into an upper, accumulation, zone (where there has been a net gain in mass) and a lower, ablation zone. The boundary between these zones is the altitude h_0 at which $\beta(h)$ is estimated to be zero.

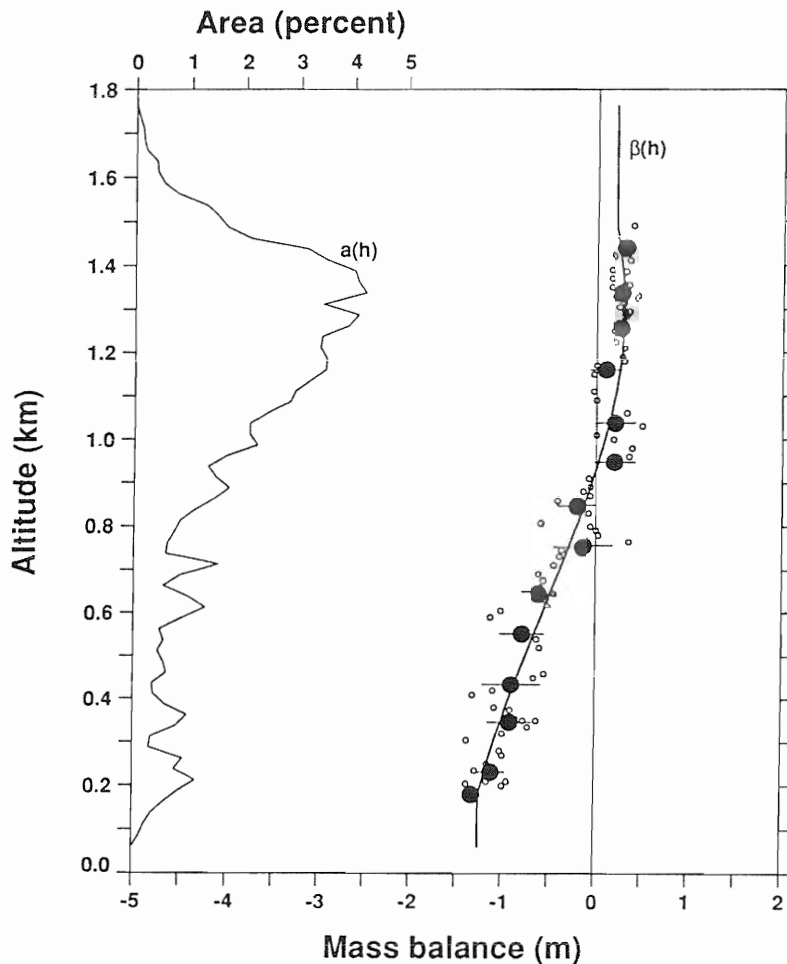


Figure 3. The function $\beta(h)$ fitted to the grouped elevations $b(h)$ on the White Glacier for the balance year 1960-61. $b(h)$ (solid circles, with error bars representing ± 1 standard deviation) is the average of all stake mass balances (open circles) measured within each 100m elevation band, while $\beta(h)$ is a polynomial (of order 3) fitted to the observed $b(h)$ by least squares. Automated and manual procedures for the fitting of mass-balance curves are discussed in Cogley et al. 1995. The equilibrium line altitude h_0 is the altitude at which $\beta(h_0)=0$. The curve labelled $a(h)$ is the hypsometric curve of the glacier.

THE EQUILIBRIUM LINE ON TEMPERATE GLACIERS

The progression of a temperate glacier through a mass balance year is shown schematically in Figure 2. Snow accumulates over the entire surface during winter, and the glacier is snow-covered at the beginning of summer. As the summer proceeds, snow and ice at the surface of the glacier melt, beginning at lower altitudes. The melt proceeds up-glacier until, at the end of the summer, it reaches an altitude at which all the previous winter's snow, but no glacier ice has melted. The quantity h_0 , explained in the last

section, is an estimator of this altitude. Above it, some snow will have melted, up to an altitude at which the entire winter snowpack survives. In this scenario, all the snow which is unmelted at the end of the summer is net accumulation for the glacier. As snow usually has a higher surface reflectance than glacier ice, the boundary between ablation and accumulation — the equilibrium line — can be detected in autumn, before the first snowfall of the winter, by human eye and other means.

However, it is clear that, even in the simplest of situations, this equilibrium line will rarely, if ever, be a line of constant altitude. The amount of winter snow present and the energy

available for melt — key variables in determining the amount of ablation — vary from place to place on a glacier at any given elevation. For example, irregularities in the glacier surface affect snow accumulation, and rates of melt are often higher (or lower) near glacier margins than towards the centre of the glacier (e.g. Adams 1966).

The equilibrium line will also be much less easy to detect in years when it separates snow from firn (roughly, snow persisting from the previous mass balance year), as in Figure 1a, rather than snow from glacier ice. Case *a* occurs whenever a year of relatively high ablation follows a year of relatively low ablation. In case *a*, the equilibrium line is a snow/firn boundary rather than a snow/ice boundary and, on a temperate glacier, up to three boundaries occur on the surface: firn/ice, snow/firn and, at the highest altitude (if at all), unmelted snow/partially melted snow. Thus detection and interpretation of the equilibrium line on temperate glaciers may be more difficult than calculations and representations of h_0 , as in Figure 1, would suggest. The situation is even more complex on polar glaciers, which contain most of the globe's ice.

THE EQUILIBRIUM LINE ON POLAR GLACIERS

On polar glaciers the progression through an ablation season, sketched in the previous section, is complicated by the refreezing of snowmelt at, and possibly beneath, the base of the cold winter snowpack. This refrozen melt accumulates and, where exposed at the surface, is referred to as superimposed ice. Here, again, summer melt proceeds from lower to higher levels on the glacier (Figure 1b). After the entire snowpack at a point has melted, the underlying superimposed ice has to be remelted before glacier ice can be melted. At the end of the summer, in this case, the equilibrium line is located where all the winter snowpack *and* all that season's superimposed ice, but no glacier ice, has melted (i.e. where $b(h)=0$). However, immediately above this line, accumulation is entirely in the form of superimposed ice. Higher up, accumulation is superimposed ice *and* snow, and higher still it is entirely in the form of snow (Figure 1b). The classic zonation for this scenario, developed from Benson (1959), is shown in Müller (1962). In this case, the boundaries on the glacier surface are: superimposed ice/glacier ice; partially melted snow/superimposed ice; and unmelted snow/partially melted snow. The second of these is

again the snowline, but the *first* is the equilibrium line, at the lower end of the superimposed ice zone.

Again, it is very unlikely that the equilibrium line will be a line of equal altitude, for the reasons given above for the temperate case, but also because of factors controlling superimposed ice formation.

Furthermore, in this case, remote detection of the equilibrium line is difficult, if not impossible, because it involves distinguishing between the surface reflectance of superimposed ice and glacier ice (Jung-Rothenhäusler 1993; Cogley *et al.* 1995), or — where a year of intense ablation follows a year of less intense ablation — distinguishing between newly-formed and one-year-old superimposed ice.

Figures 3 to 6 illustrate the points made above. Figure 3 is a plot of mass balance versus altitude for the White Glacier for 1960-61. Stake mass balance data for the same year, plotted along the length of the glacier, are shown in Figure 4. In both of these diagrams h_0 is identifiable as the altitude at which the thick curve crosses the zero balance axis. Figure 5 (Adams 1966) is a map of the equilibrium line as interpreted by field observation for the main stream of the White Glacier for that year.

The marked variations in altitude of the equilibrium line on the White Glacier in 1960-61 may be ascribed in part (Adams 1966) to large slush avalanches (transferring mass down-glacier) and accumulation of snow and superimposed ice in a large hollow at about 750m (Figure 5). Indeed, in 1960-61 (as in other years) there was no single, continuous, boundary between accumulation and ablation zones on the White Glacier. Rather, there were substantial outliers of accumulation within the "ablation zone" and inliers of ablation in the "accumulation zone" (see also Figure 6). This situation is likely to be common on all but the most simple of glaciers.

CONCLUDING REMARKS

An "equilibrium line" at constant altitude h_0 is a simplification, albeit a useful simplification, of reality. Given enough information, it is possible to detect the boundary between accumulation and ablation at the end of a summer, so that an isarithm of zero mass balance can be drawn on a map (e.g. Adams 1966; Rogerson 1986; Wolfe 1994). But this isarithm will not normally be simple and it will not be a line of constant altitude.

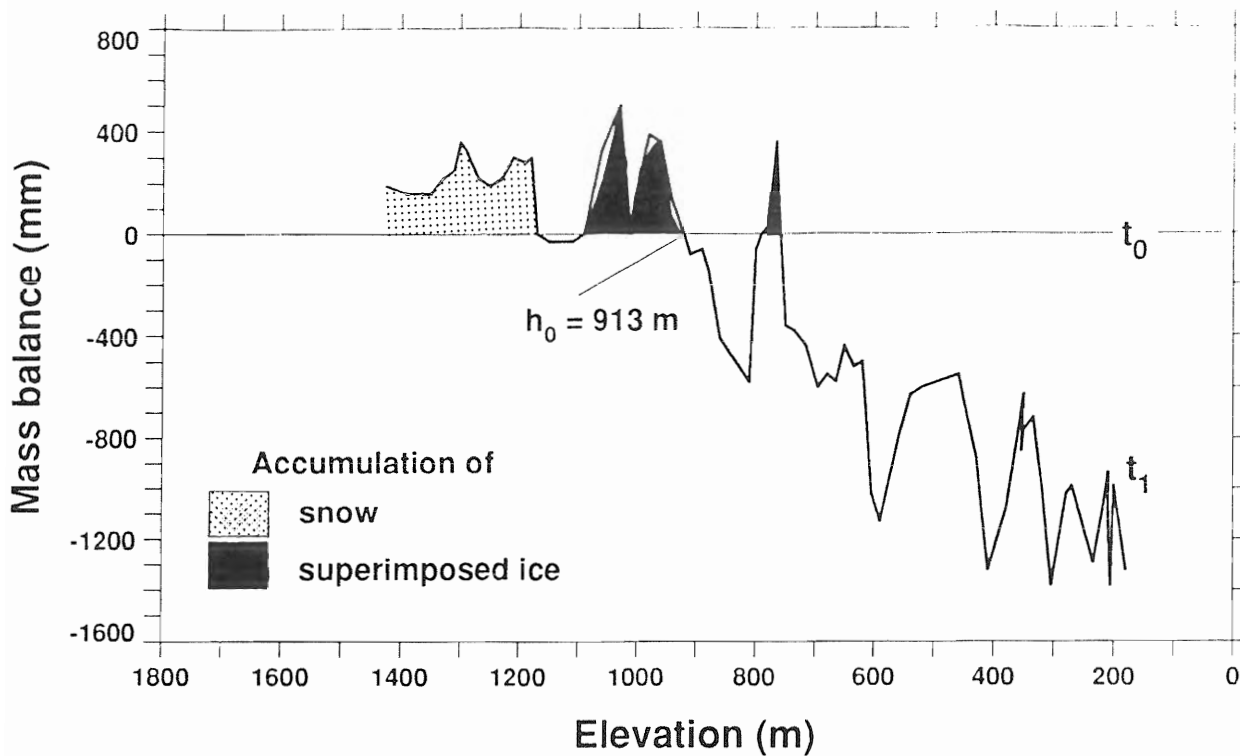


Figure 4. Point mass balance b as observed at stakes on the White Glacier in 1960-61. The stakes are those of the so-called Longitudinal Profile, which followed closely the centreline of the glacier. t_0 represents the glacier surface at the start, and t_1 the surface at the end, of the balance year.

The practical consequences of this complexity are far-reaching. For example the goal of substituting satellite-based remote sensing of glacier mass balance for expensive programmes of field measurement remains elusive after more than twenty years of effort (Østrem 1975; Jung-Rothenhäusler *et al.* 1992; Jung-Rothenhäusler 1993). In large part this must be due to unappreciated complexity in spatial patterns on the glacier surface. Most proposals for estimating mass balance by remote sensing rely on being able to distinguish, in some part of the electromagnetic spectrum, newly-added mass (snow and superimposed ice) from newly-exposed old mass (firn, glacier ice, and old superimposed ice). There seems to be no realistic alternative as yet to continued detailed study in the field of the spatial relations between these surface types. A good physical understanding of melting, refreezing and other processes, and of associated forms, in the vicinity of the equilibrium line altitude is one

prerequisite for progress in remote sensing of glacier mass balance.

We believe that, for the present, it is more appropriate and productive to think in terms of an equilibrium zone (Figure 6), which may, as on the White Glacier, have a vertical span of hundreds of metres, and may occupy a substantial proportion of the glacier surface. The equilibrium zone may be defined as being bounded below by the lowest altitude at which $b > 0$ and above by the highest altitude at which $b < 0$. If the $\sim 300 - 350\text{m}$ vertical span of the equilibrium zone in 1960-61 is typical, then on the White Glacier the equilibrium zone typically occupies about a quarter of the glacier surface. In a year when h_0 is close to 1350m, which is the maximum of the White Glacier's hypsometric curve (Figure 2; *cf.* Cogley *et al.* 1995), the equilibrium zone will occupy almost one half of the surface. This encourages thought about processes and phenomena, especially remotely-detectable phenomena such as slushflow deposits, which may

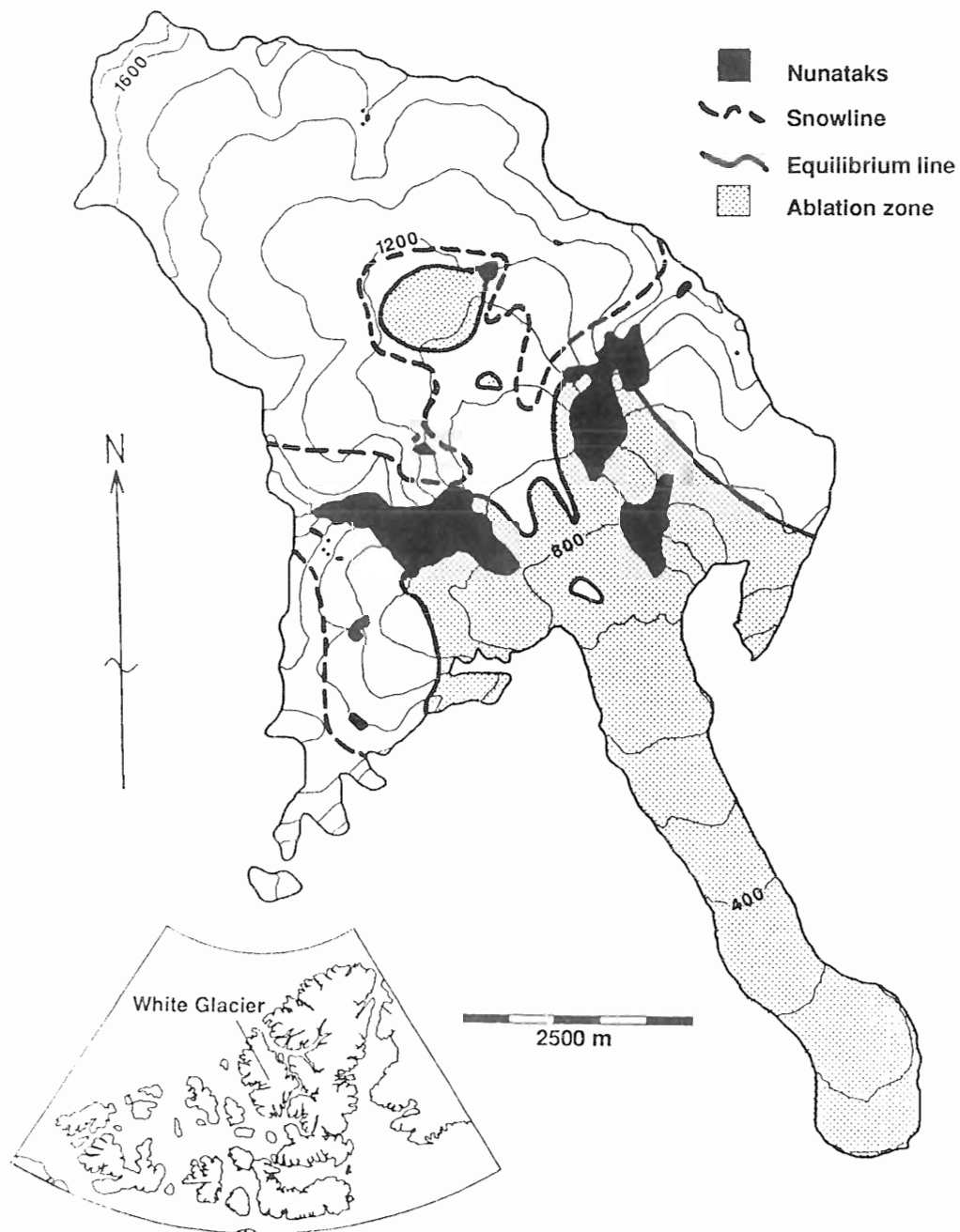
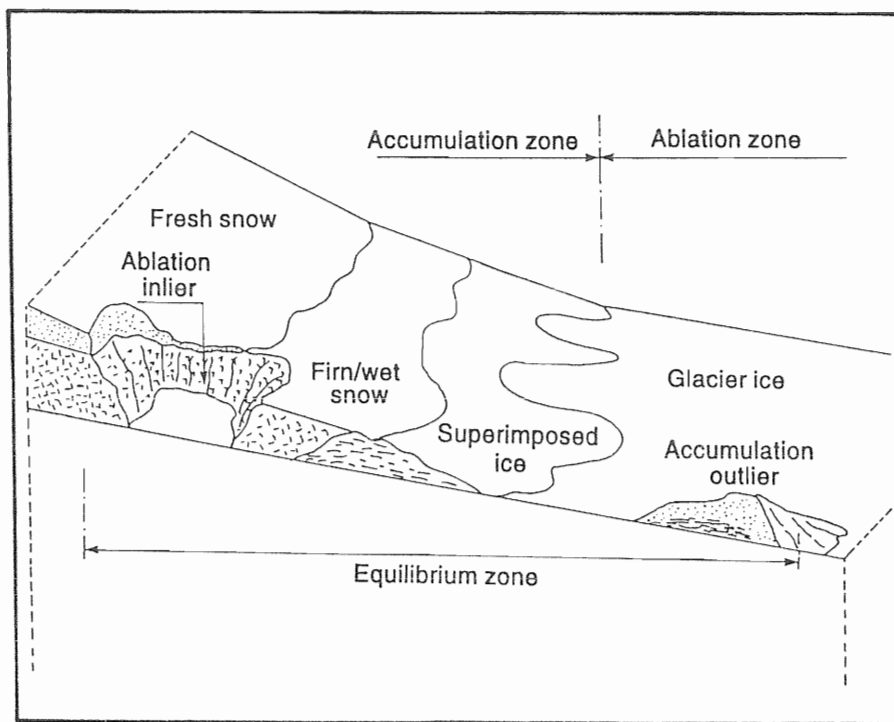


Figure 5. Sketch map of the White Glacier at the end of the 1960-61 balance year (Adams 1966). Slush avalanches can be seen as tongues of net accumulation of superimposed ice extending into the ablation zone, and isolated inliers of ablation and an outlier of accumulation are also visible. The equilibrium zone ranges from -750m to -1200m . The positions of the snowline and equilibrium line, which enclose the superimposed ice zone, become speculative (although based on field inspection) beyond -1km away from the centreline of the glacier.



(after Jung-Rothenhäusler, 1993)

Figure 6. Snow and ice facies in the equilibrium zone, i.e. in the vicinity of h_0 . The equilibrium zone typically occupies an altitude range of several hundred metres on the White Glacier.

eventually allow more precise identification of the equilibrium "line" within the equilibrium zone. Even the extent of the equilibrium zone, however, is difficult to estimate accurately from stake networks of typical density. This approach is particularly appropriate for polar glaciers and others on which accumulation of superimposed ice can force the equilibrium line to significantly lower elevations than the snowline.

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