THE FORMATION OF VALLEY-WALL ROCK GLACIERS

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THE FORMATION OF VALLEY-WALL ROCK GLACIERS

by

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I ALISON MACLEAN hereby certify that this thesis has been composed by myself, that it is a record of my own work, and that it has not been accepted in partial or complete fulfilment of any other degree or professional qualification.

Signed .. .. Date 25th April '91

I was admitted to the Faculty of Science of the University of St. Andrews under Ordinance General No. 12 on *1st October 1985* and as a candidate for the degree of Ph.D. in *October 1986*.

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UNRESTRICTED

To Mum and Dad

Abstract

In recent years, the study of rock glaciers has increased remarkably. Substantive progress has been made, particularly in understanding the formation of rock glaciers that have developed adjacent to existing or former valley or cirque glaciers. However, our understanding of valley-wall rock glaciers that are located at the base of talus slopes remains scant. Published work exhibits little consensus on the formation of valley-wall rock glaciers and several hypotheses remain under vigorous debate. The major objective of the research reported in this thesis has been to test the generality and feasibility of seven major models of valley-wall rock glacier formation using both empirical and theoretical evidence. The primary conclusion is that only one of these models, the segregation ice model, emerges as a general model of valley-wall rock glacier genesis. The model assumes that a thin layer or several thin layers of segregated ice are overlain by interstitially frozen sediments and an unfrozen mantle of coarse debris. A wide range of empirical and theoretical findings are shown to be consistent with the implications of the segregation ice model. Detailed observations on the morphology, sedimentology and distribution of active, inactive and relict valley-wall rock glaciers studied in Switzerland, northern Norway and Scotland provided a range of findings that support this model. Theoretical evidence was obtained by modelling a number of different density models that reflect different distribution of internal ice by applying a simple laminar flow equation to field measurements. Although only the segregation ice model appears to be valid at a general level, the possibility cannot be excluded of alternative modes of valley-wall rock glacier formation

under particular circumstances. Snow avalanching, deformation of snowbank or matrix ice, and basal sliding under conditions of high hydrostatic pressure all constitute possible contributing mechanisms of formation and movement in particular cases.

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Table of contents

LIST OF FIGURES

Figure: **Page No.:**

LIST OF TABLES

LIST OF PLATES

Chapter 1

Introduction

1.1 Aims and rationale

Over the past few decades, the study of rock glaciers has increased remarkably. Before the publication of a seminal paper on rock glaciers by Wahrhaftig & Cox in 1959, the topic tended to receive little more than passing comment in general textbooks. In recent years, however, a profusion of published work has reported on many aspects of these landforms. Substantive progress has been made, yet the expanding literature base continues to raise as many questions as it answers. Recently, this lack in clarity surrounding rock glacier research was highlighted by the publication of a symposium volume on rock glaciers (Giardino *et al.*, 1987). The volume, intended as a summary of current rock glacier research, disappointingly dwells on the complexity of rock glaciers and the lack of consensus on their origin.

Even the term *rock glacier* continues to generate controversy. A useful definition, however, is provided by Wahrhaftig (1987), who states that the term should be restricted to 'accumulations of blocky detritus, extending outward and downslope from talus cones or from glaciers or the terminal moraines of glaciers, that have extended thus in large part through the deformation of interstitial ice or clear ice bodies within them'. Wahrhaftig's definition draws a valid distinction between two types of rock glacier; those that have developed adjacent to existing or former valley or cirque glaciers, hereafter referred to as *valley-floor rock glaciers*, and those located at the base of talus slopes, hereafter referred to as *valley-wall rock glaciers*.

Over the years, much of the published work has concentrated on the study of valley-floor rock glaciers to the exclusion of valley-wall forms. Indeed, whereas it is now generally accepted that valley-floor rock glaciers are predominantly glacially derived, several hypotheses of valley-wall rock glacier formation remain under vigorous debate. The purpose of this thesis is to test the feasibility of these hypotheses of valley-wall rock glacier formation using both empirical and theoretical evidence.

1.2 Structure of thesis

The rock glacier literature comprises a plethora of site-specific studies and regional morphometric analyses, but very few in depth papers on specific research topics. Papers devoted entirely, for example, to the study of debris input mechanisms, sedimentology, or the mechanics of rock glacier movement are conspicuous by their rarity, whilst many of the site-specific studies contain pages of frustratingly circular argument on less important topics such as terminology and taxonomy. The purpose of Chapter 2, *Rock Glaciers - A Review*, is to identify and evaluate, from the existing literature base, areas of real progress from areas that require additional study, and to suggest approaches to solving some of the major 'unknowns' in rock glacier research. Problems of definition, classification, and identification are considered in the first section of the chapter and working definitions are adopted. Subsequent sections document recent advances in understanding rock glacier distribution, morphology, sedimentology, and processes and rates of rock glacier movement. A concluding section contains an evaluation of the major theories of rock glacier formation for both valley-floor and valley-wall forms.

Review of the literature suggests that seven models of formation constitute the crux of the current debate on the origin of valley-wall rock glaciers. The aim of Chapter 3, *Models of Valley-Wall Rock Glacier Formation*, is to present and discuss each of these models. The generality of each model of formation is considered briefly on the basis of evidence reviewed in Chapter 2. More detailed testing, however, is left to subsequent chapters following the collection of additional empirical and theoretical evidence.

Field evidence pertinent to testing the proposed models was collected from four field areas (Chapter 4) that contain valley-wall rock glaciers of varying activity status, namely the Lyngen Peninsula in northern Norway, the Valais and Graubünden areas of the Swiss Alps, and the Highlands of Scotland. The influence of a range of environmental and geomorphic factors on the distribution of valley-wall rock glaciers within these study areas is also investigated in Chapter 4.

Chapter 5, *Rock Glacier Morphology*, is devoted to a study of the morphological characteristics of the active, inactive and relict rock glaciers that were studied in these four areas. The length, width, thickness, volume and surface gradient of rock glaciers were calculated to define a range of rock glacier dimensions for use in later theoretical calculations. Statistical analyses were also used to determine if morphometric regularities exist within the survey data. In the second half of the chapter, attention is focused on attempts to define common morphological characteristics for several surface microrelief features, including frontal and transverse ridges. In addition, the morphological development of valley-wall rock glaciers is considered using site-specific information from each of the study areas.

Chapter 6 is devoted to a consideration of rock glacier sedimentology. Each of the three principal constituents of active valley-wall rock glaciers, namely coarse debris, fine sediments and ice, is investigated in turn. First, the implications of clast-size distribution patterns for rates and processes of debris input are discussed. Second, observations on the general distribution and characteristics of fine sediments within valley-wall rock glaciers are described using data obtained from granulometric and chemical analyses. In the third section, an attempt is made to ascertain the origin of ice found within valley-wall rock glaciers by comparing the isotopic content of rock glacier ice with ice of known provenance. The chapter concludes by outlining some criteria that may be used to distinguish valley-wall rock glaciers from other morphologically-similar features. Thus Chapters 4, 5 and 6 summarise respectively the locational, morphological and sedimentological characteristics of the rock glaciers studied, and comprise a data base against which subsequent theoretical calculations may be tested.

Theoretical aspects of the rheology of rock-ice mixtures are poorly understood. Through application of a simple laminar flow equation to several conceptual models of rock glacier formation, an attempt is made in Chapter 7 to determine boundary conditions for rock glacier movement. A number of different models are proposed, each based on different assumptions concerning the nature and origin of internal ice, and their feasibility is tested using field measurements of rock glacier thickness and average surface gradient. The seven hypotheses of formation outlined earlier in the thesis are then evaluated through consideration of the implications of the theoretical and empirical findings of the previous three chapters. Chapter 7 concludes with a detailed summary of valley-wall rock glacier genesis.

The final chapter (8) restates the most important conclusions reached in this research and evaluates possible avenues for future work on this topic.

Chapter 2

Rock Glaciers - A Review

2.1 Introduction

First documented at the beginning of the present century, rock glaciers remain phenomena of great interest and debate. Following the publication of a seminal paper on Alaskan rock glaciers (Wahrhaftig & Cox, 1959), rock glacier research increased markedly. Yet, in spite of some progress, fundamental questions regarding the formation, flow mechanisms, and behaviour of these landforms remain unanswered.

The explanation of rock glaciers is complicated by several factors. Most notably, information about the internal structure of rock glaciers has proved very difficult to obtain because of the extremely coarse nature of constituent debris. There also exists much terminological confusion within the literature on this topic. Since the term *rock glacier* first appeared (Capps, 1910), it has been applied somewhat indiscriminately to a variety of glacial and non-glacial features that are in some cases not even morphologically similar. The use of the term *rock glacier* itself is unfortunate, and has probably hindered both explanation of the formation of rock glaciers and interpretation of their geomorphic and environmental significance because of the inherent implication that glacial ice is a necessary component. In addition, several possible origins have been proposed for rock glaciers. Any attempt to define, classify, and interpret a range of features that may contain both genetic and morphological differences is likely to be problematic, and as a result many of the definitions that have been proposed fail to characterise satisfactorily all rock glaciers.

The aim of this review is to summarise and evaluate the wide-ranging views found within the rock glacier literature and to identify those pertinent research questions that remain to be answered satisfactorily. Problems of definition, classification, and identification are discussed in the first section of the chapter before a review of rock glacier morphology and sub-surface information is presented. A detailed discussion of rock glacier distribution, flow mechanics, and origin follows, and the chapter concludes with a general discussion of the major 'unknowns' in rock glacier research.

2.2 *Definition and classification*

2.2.1 *Introduction*

Numerous definitions of the term *rock glacier* have been proposed (e.g. Wahrhaftig & Cox, 1959; White, 1976). Many of these definitions are based solely on morphological description (e.g. Spencer, 1900; Potter, 1972), whilst others include details of internal structure and some imply genesis (e.g Barsch, 1978, 1987; Haeberli, 1985). Most researchers, however, view rock glaciers as masses of angular debris that contain internal ice and which move slowly downslope to produce lobate or tongue-shaped landforms with steep fronts and sides.

In the early literature on the topic, definitional diversity appears to have resulted partly from researchers either using different terminology or misinterpreting other landforms for rock glaciers. Almost twenty years before Capps (1910) coined the term *rock glacier*, Kendal (1891) provided an account of a possible rock glacier which he described as "a moraine-like mound", near Snowdon. In 1900, Spencer described a tongue-shaped rock glacier located downslope from a glacial cirque in the San Juan Mountains, Colorado, as a "peculiar form of talus" (Spencer, 1900, p. 188), and Rohn (1900) described rock glaciers in Alaska as "talus slopes that move". Other early researchers who proposed alternative terms for rock glaciers included Tyrrell (1910) who called them "chrystocrenes", and Cross and Howe (1905), Patton (1910), Kesseli (1941), Griffiths (1958) and Wilhelmy (1958) who referred to them as "rock streams". Descriptions of rock glaciers as "landslides" (Howe, 1909), "blocky moraines" (Chaix, 1923), and "fossil glaciers" (Brown, 1925) added to a terminological confusion that persists to the present-day.

Recent workers have tended to follow one of two approaches towards definition. The first approach is dependent upon morphological description. This approach is exemplified by Potter (1972, p. 3027), who defined a rock glacier as "a tongue-like or lobate body of usually angular boulders that resembles a small glacier, generally occurs in high mountainous terrain and usually has ridges, furrows and sometimes lobes at its surface, and has a steep front at the angle of repose." The second approach towards definition builds upon the first by relating surface morphology to possible modes of formation. For example, Barsch (1978, p. 349) stated that "an active rock glacier is a tongue or lobate shaped body of frozen talus of 'porridge-like' appearance, which creeps downslope at an average speed of 5 to 100cm/a (maximum: 500cm/a) and which is cemented by interstitial ice, the content of which changes from place to place and from individual to individual."

Although more detailed, Barsch's definition implies that all active rock glaciers are "cemented" by "interstitial ice" and move downslope by "creep". Such assertions, however, are not upheld by many workers who regard rock glaciers as polygenetic (e.g. Corte, 1976; White, 1981; Johnson, 1984), and so genetic definitions of this sort must be regarded as premature. Even so, definitions based solely on morphological description, although more acceptable as general definitions, tend to be simplistic and of limited value.

A universally-accepted definition of all rock glaciers remains elusive. Nevertheless, a general definition is both justifiable and necessary to delimit the general set of landforms termed rock glaciers, and to differentiate these from other similar landforms. Even in the recent past rock glaciers have been confused with protalus ramparts, ice-cored moraines, and landslide deposits (e.g. Østrem, 1971; Sissons, 1980a; Lindner and Marks, 1985). For the purposes of simplification it seems appropriate to adopt a working definition for all rock glaciers. Thus, in the author's opinion, a rock glacier may be defined as a multi- or single-ridged accumulation of coarse angular debris that forms either at the base of talus slopes in periglacial environments (hereafter referred to as a *valley-wall rock glacier)* or adjacent to an existing valley or cirque glacier or ice-cored moraine (hereafter referred to as a *valley-floor rock glacier*). When active, a rock glacier contains ice that may be either massive or in the form of a matrix or both, and frozen sediments. Possible sources of such ice are infiltrating precipitation, meltwater runoff from upslope, snow avalanches, and burial of snowbeds or glacier ice. Rock glacier movement is thought to occur mainly through the downslope deformation of such ice. All rock glacier types may be more concisely defined as massive accumulations of debris that move or have moved partly as a result of internal deformation of ice.

The terms *valley-wall* and *valley-floor rock glacier* will be adopted in this thesis in order to reduce confusion over terminology. Their usage is intended primarily to reflect the ease with which rock glaciers may be correctly and usefully classified within the field solely on the basis of location. By translating from other authors' terminology a list of broadly equivalent terms for each type of rock glacier was constructed (Table 2.1). The length of Table 2.1 is ample proof that a simplified terminology of rock glaciers is needed. The division of rock glaciers into valley-wall and valley-floor forms reflects genetic differences as well as the more obvious locational ones, and Figure 2.1 illustrates the development of this simple two-fold classification in the form of a flow diagram. The first type of rock glacier, the valley-wall form, is periglacial in origin and is believed

VALLEY-WALL ROCK GLACIERS	*VALLEY-FLOOR ROCK GLACIERS*	*AUTHOR*	
Lobate rock glaciers	Tongue-shaped rock glaciers	Domaradzki, 1951; Wahrhaftig and Cox, 1959; White, 1976.	MORPHOLOGICAL
	Rock streams	Cross and Howe, 1905; Patton, 1910; Kesseli, 1941; Wilhelmy, 1958; Griffiths, 1958.	
Talus-derived rock glacier		Humlum,1984; Johnson, 1981.	COMPOSITIONAL
Peculiar form of talus		Spencer, 1900.	
	Blocky moraines	Chaix, 1923.	
Talus rock glaciers		Barsch, 1977a.	
Talus terraces		Liestøl, 1962, Chandler, 1973.	
Valley-wall rock glacier	Cirque floor rock glacier	Outcalt & Benedict,1965.	
Talus-foot rock glacier		Dumbell,1984.	
Protalus lobes		Richmond,1962.	LOCATIONAL
Piedmont talus glacier		Smith,1973.	
Protalus rock glacier		Gray, 1972.	
	Moraine rock glacier	Lindner and Marks, 1985.	
	Landslides	Howe, 1909.	GENETIC
	Fossil glacier	Brown, 1925.	
Avalanche rock glacier		Johnson, 1978	
Primary	Secondary	Corte, 1976.	
← Chrystocrenes →		Tyrell, 1910.	

Table 2.1 List of broadly equivalent terms for valley-wall and valley-floor rock glacier.

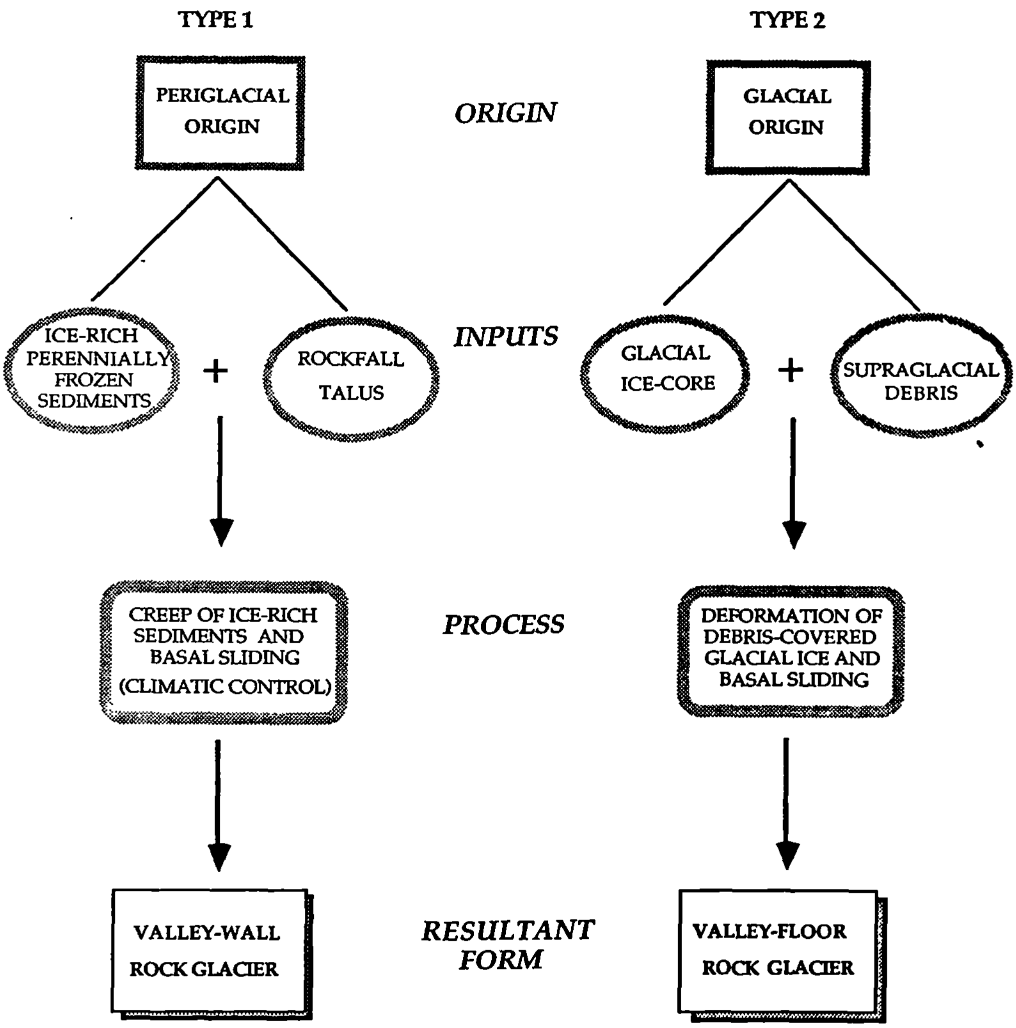


Figure 2.1 Simplified two-fold classification of rock glaciers on the basis of origin, inputs and process of formation.

by the majority of researchers to form by internal deformation of ice-rich perennially frozen sediments within the lower parts of talus slopes. The second type of rock glacier, the valley-floor form, is thought by most authors to be glacially derived and to require massive glacial ice and supraglacial debris for its formation. The flow diagram, which is presented initially as a framework against which some of the many rock glacier classifications that have been proposed may be analysed (e.g. Domaradzki, 1951; Wahrhaftig and Cox, 1959; Barsch, 1969, 1987; J.P. Johnson, 1974; White, 1976; Corte, 1987), will be discussed and developed further in section 2.2.7. Previous approaches towards classification can be subdivided into those that are based on morphology or activity, those that are mixed-criterion classifications, those that are based on genesis or constituent materials, and those that form part of a continuum of landforms.

2.2.2 *Classification by morphology*

The most widely-quoted classification that is based on morphology was initially proposed by Domaradzki (1951) in a study of "block streams" in the Upper Engadine valley, Switzerland, and was subsequently developed by Wahrhaftig and Cox (1959). Domaradzki divided rock glaciers into two groups, namely tongue-shaped and lobate. By identifying differences in plan form, topographic position, and length to width ratio for over 200 rock glaciers in Alaska, Wahrhaftig and Cox confirmed Domaradzki's two-fold classification and added a third somewhat subsidiary category, which they termed spatulate; a tongue-shaped rock glacier but with an enlargement at the front. As Wahrhaftig and Cox believed that their three-fold classification represents three stages of rock glacier development, their terms lobate and tongue-shaped are not directly interchangeable with valley-wall and valley-floor, as they envisaged that tongue-shaped rock glaciers can form when lobate rock glaciers from the sides and back of a cirque meet in the centre and the resulting mass flows down-valley from the cirque. However, as Wahrhaftig and Cox observed that lobate forms are also commonly found to extend out from the base of talus slopes usually at right angles to tongue-shaped rock glaciers, many authors employed their morphological classification because it is readily applicable to the majority of rock glacier forms. For example, Richmond classified rock glaciers into either "tongue-like or lobate masses of rubble, locally with a core of till-like debris" (Richmond, 1962, p.20). Tongue-shaped rock glaciers as described by Domaradzki (1951) and White (1976) correspond to the rock streams of earlier workers, to the cirque floor rock glaciers of Outcalt and Benedict (1965), and to the moraine rock glaciers of Lindner and Marks (1985) (see Table 2.1). Their length to width ratio as proposed by Wahrhaftig and Cox is greater than 1.0. Conversely, lobate forms supposedly have a length to width ratio of less than 1.0, for they are generally broader than they are long. Lobate rock glaciers are the equivalent of the protalus lobes of Richmond (1962), the valley-wall rock glaciers of Outcalt and Benedict (1965), the piedmont talus glaciers of Smith (1973), the talus-derived rock glaciers of Humlum (1984), the talus-foot rock glaciers of Dumbell (1984), and the protalus rock glaciers of Lindner and Marks (1985).

2.2.3 *Classification by activity*

A three-fold classification based on current or recent activity was proposed by Wahrhaftig and Cox (1959). This classification, which divides rock glaciers into active, inactive, or reactivated features, does not include fossil or relict features such as those located in Scotland and described by Sissons (1975, 1979). A more comprehensive three-fold classification based on activity might be one in which rock glaciers were divided into those that are currently active, currently inactive or intermittently active, and fossil or relict. Rock glacier activity is commonly recognised by the gradient and degree of vegetational colonisation of the frontal slope, and by the angle at which the frontal slope joins the upper surface of the rock glacier (Wahrhaftig and Cox, 1959). Thus, active rock glaciers generally have steep frontal slopes that are free of vegetation and are often composed of much finer debris than the upper bouldery surface of the rock glacier. A sharp break of slope usually marks the boundary between the frontal slope and the upper surface of the rock glacier. Conversely, inactive rock glaciers commonly exhibit gentler frontal slopes that are partly or wholly vegetated. The angle at which the frontal slope meets the upper surface is less pronounced and the rock glacier profile appears more rounded. Classification by activity is a useful method for distinguishing rock glaciers although some rock glaciers have been found that do not exhibit the morphological criteria specified for their activity status (Wahrhaftig and Cox, 1959). Such exceptions highlight the difficulties inherent in any descriptive classification.

2.2.4 *Mixed-criterion classifications*

A mixed-criterion classification involving both activity and ice content was suggested by Barsch and King (1975). They identified three groups:

1) active rock glaciers, which show actual downslope movement;

2) inactive rock glaciers, which no longer show any movement but which still contain ice; and

3) fossil rock glaciers, which are ice-free and have collapsed.

There are, however, two major difficulties with this classification. First, long-term geodetic investigations or detailed photogrammetrical analyses are required to establish rock glacier movement, and, second, ice-content is difficult to determine without time-consuming excavations or detailed geophysical studies. Therefore, in terms of application, the morphologically-based, single-criterion scheme devised by Wahrhaftig and Cox is more useful than the mixed-criterion scheme proposed by Barsch and King.

An alternative and more satisfactory mixed-criterion classification was proposed by Barsch (1969). He identified three major groups:

1) rock glaciers at the base of talus slopes;

2) rock glaciers below end moraines:
 a) large valley rock glaciers
 b) small rock glaciers, partly end moraines which are still moving;

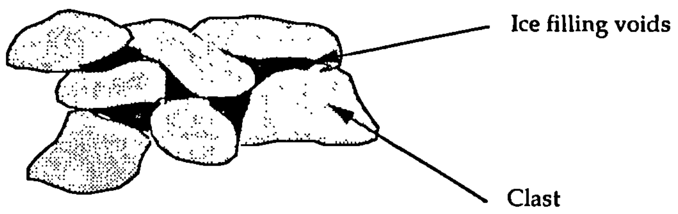
3) special forms: glaciers completely covered by rubble.

The major advantage of Barsch's scheme is that location is used in addition to constituent materials to distinguish a special type of rock glacier (3). As discussed earlier, location is a valuable classificatory parameter because it is one of the easiest to define and apply. It might have been appropriate, however, to include origin as an additional criterion, if only in the widest sense. For example a three-fold grouping that separates rock glaciers into those that are certainly of glacial origin, possibly of glacial origin, and of definite non-glacial origins may be helpful. By 1987, Barsch had decided that he could divide rock glaciers into two major types, namely 'talus rock glaciers' and 'debris rock glaciers', neither of which is related to ice glaciers. His classification, which is perhaps too narrow, reflects his view that all rock glaciers are creep phenomena in frozen sediments.

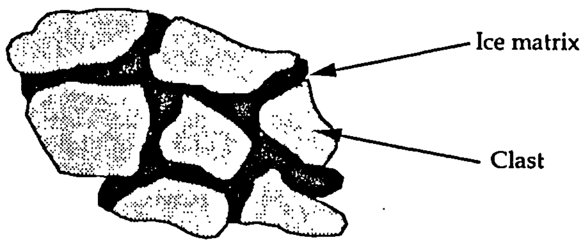
White (1976, 1981) disagreed with Barsch's 1969 classification. He stated that "only three rock glacier types exist: ice-cored tongue shaped, ice-cemented tongue-shaped, and lobate" (White, 1976, p.81). White believed that the distinction that Barsch drew between groups 2.a), 2.b), and 3) was unnecessary as they are all tongue-shaped rock glaciers. However, the subdivisions proposed by Barsch are worthwhile for they indicate that tongue-shaped rock glaciers can originate in a variety of locational settings and at different scales by essentially the same process (i.e. the deformation of glacial ice and its associated supraglacial debris). As mentioned previously, many difficulties are encountered when trying to determine the ice-content of rock glaciers. Attempts have been made to distinguish ice-content by studying rock glacier morphology, and White, in particular, is a strong advocate for such an approach when classifying rock glaciers in the field. For example, morphological features such as closed depressions and conical pits are typically used to indicate *ice-cored rock glaciers* and their absence suggests *ice-cemented forms* (White, 1976).

As in other aspects of rock glacier research, the terminology used to describe ice-content is very confusing and at times quite illogical. For example, the term *ice-cemented* (e.g. Barsch, 1987; Vick, 1987; Vitek & Giardino, 1987) implies that a landform is cemented by ice and, thus, by definition, incapable of movement. Likewise, the term *interstitial* (e.g. Spencer, 1900; Wahrhaftig and Cox, 1959) means that ice fills only the interstices and does not separate the clasts. The shear strength of such a rock-ice mixture in which clast to clast contact persists is greater than one in which no ice exists and only air fills the voids. Thus, the development of only interstitial ice sensu stricto is unlikely to be sufficient for rock glacier movement to occur. For the purposes of simplification, the standard terms described below and illustrated in Figure 2.2 will be used throughout the thesis.

1) Interstitial ice - ice fills voids but clast to clast contact remains

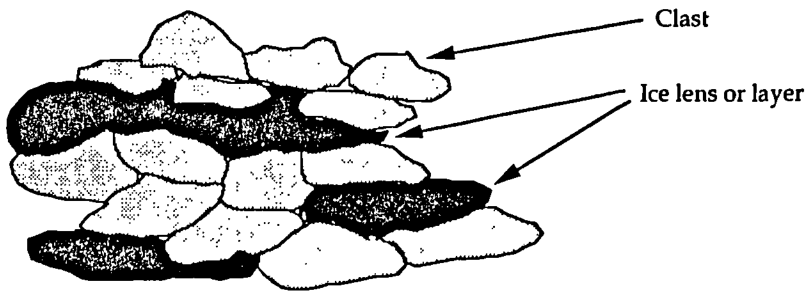


2) Matrix ice - ice supporting clasts



Shear strength of frozen sediments comprising matrix ice is significantly less than the shear strength of frozen sediments comprising interstitial ice as ice separates clast-to-clast contacts.

3) Segregation ice - layers of ice or lenses separating clasts



Shear strength of frozen sediments comprising segregation ice varies depending on the degree of ice segregation and hence the degree of clast separation.

Figure 2.2 Definitions of interstitial, matrix and segregation ice

1) The term *interstitial ice* describes ice that fills interstices between clasts but does not separate clasts.

2) The term *matrix ice* refers to ice that separates and supports clasts.

3) The term *segregation ice* identifies ice that occurs as layers or lenses often separating clasts.

2.2.5 Classification by origin

Corte (1976) recognised two main types of rock glacier which he termed *primary* and *secondary* features. He suggested that primary forms are "real rock glaciers" formed from talus cones and protalus ramparts, whereas secondary rock glaciers are those derived by melting of valley glaciers and are in effect debris-covered glaciers. He stated that it was difficult to distinguish one type of rock glacier from the other, especially where the debris-covered glacier is in its final melting stages. Morphologically this may be true. However, locational differences between each type of rock glacier should allow for easy differentiation. Thus, a combination of genetic and locational parameters may provide a more useful classification particularly if used in conjunction with terms other than primary and secondary, which are very vague. Potter (1972) also recognised that rock glaciers could be divided on the basis of rock glacier origin. His groups, which he termed *ice-cemented* and *ice-cored* correspond closely to Corte's primary and secondary features. Ice-cemented forms, which Potter believes develop as a result of the movement of talus and matrix ice, require that the climate be cold enough for permafrost to exist. On the other hand, ice-cored forms, which are less dependent on purely climatic conditions, develop when glacier ice becomes covered with rock debris. Thus, both authors recognise that some rock glaciers are periglacial in origin while other forms are glacially-derived. In a recent paper, Corte (1987) provided an alternative classification that outlines the complexities of possible origins. He divided rock glaciers into seven types on the basis of genetic factors. However, his new classification, which includes terms such as 'cryogenic gelifluction debris mantle rock glaciers' and 'technogenic (artificial rock glaciers)', introduces ambiguities and is not an improvement on his earlier classification.

2.2.6 Classification by constituent materials

J.P. Johnson (1974) suggested that rock glaciers could be distinguished on the basis of their constituent materials. He devised a five-fold classification in which the dominant component of rock glaciers is either:

1) glacier ice;
2) non-glacier ice (interstitial);
3) lenses of non-glacier ice (snowbank origin) or permafrost;
4) rubble; or
5) some combination of 1-4.

Sub-surface information is rarely available however, and so this classification is clearly unworkable in most instances. Johnson also tentatively proposed a second classification which he represented graphically (Figure 2.3), and in which the relative percentages of glacial

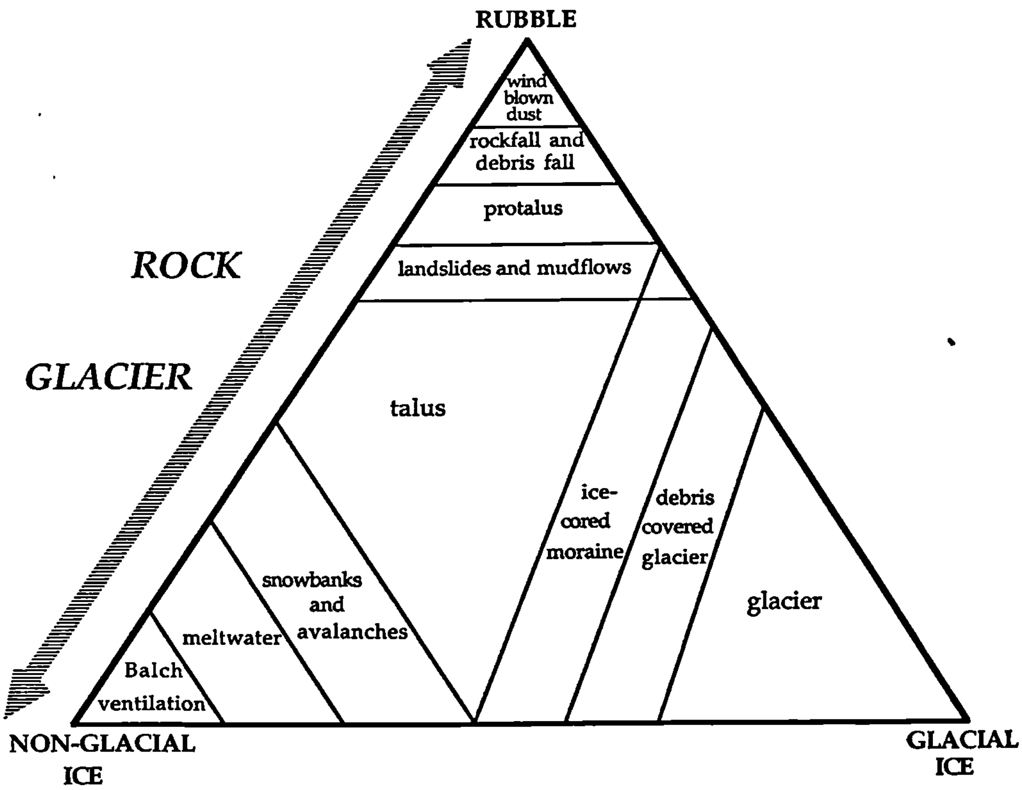


Figure 2.3 Classification of rock glaciers based on constituent materials (after Johnson, 1974).

ice, non-glacial ice, and rock rubble are indicated along each side of a triangle. He proposed the classification as "a crude attempt to show what needs to be considered" (Johnson, 1974, p. 87). However, this classification is illogical both in design and in content, and contributes little to the understanding of rock glaciers. As mentioned previously, there is usually little point in employing internal content in classification as sub-surface information is normally very limited.

2.2.7 Classifications as part of a continuum

More recently, attempts have been made to classify rock glaciers as part of a continuum of landforms. For example, Lindner and Marks (1985) presented a classification that includes all debris accumulations formed at the foot of mountain slopes and glacier snouts in South Spitsbergen. They proposed that talus, protalus ramparts, and protalus rock glaciers (valley-wall rock glaciers) form one continuum, and that debris-covered glaciers, ice-cored moraines, and moraine rock glaciers (valley-floor rock glaciers) comprise another. Similarly, a study of rock debris transport and deposition by valley glaciers in South Georgia led Birnie (1978, p.1) to conclude that "variations in the area of contributing rockwall and the receiving glacier show that rock glaciers, protalus ramparts, and ice-cored moraines form part of a continuum of depositional landforms." Birnie suggested that a high debris supply and a small glacier would result in the formation of a rock glacier; conversely, if debris supply is low and there is a large amount of glacier ice , an ice-cored moraine will develop. The intermediate situation is represented by a protalus rampart where both debris supply and glacier are small. Birnie's classification, however, implies that no difference in process exists between the formation of ice-cored moraines, rock glaciers and protalus ramparts. Such assertions are not upheld by the majority of researchers (e.g. Wahrhaftig & Cox, 1959; Barsch, 1977b; White, 1979; Haeberli, 1985), however, and so his proposition must be considered dubious and simplistic.

Johnson (1983) also proposed a continuum of forms, but did not include other phenomena in his classification. At one end of his continuum, he suggested a simple unmodified talus slope, and at the other end a complex form of rock glacier characterised by a highly complicated ridge morphology. Shakesby *et al.* (1987) observed a similar morphological and developmental continuum of talus-derived landforms in Rondane, Norway. However, they added a "push-deformation moraine" category, which they regarded as the most complex type of talus-derived landform. Shakesby *et al.* and Johnson's continua are certainly easier to accept than those proposed by Birnie and by Lindner and Marks, particularly for valley-wall rock glaciers. The other continua must be regarded as more speculative, because the transitional forms implicit in them have not all been reported, except in the case of Birnie's classification which is dubious in terms of process.

2.2.8 Discussion

Numerous rock glacier classifications have been proposed and yet a universally accepted classification remains as elusive as a universally accepted definition. In order to provide a simplified rational framework against which much of the information contained within this review can be discussed, the initial flow diagram proposed in section 2.2 (Figure 2.1) has been developed further to include many useful classificatory parameters outlined in the literature. Thus, in the revised flow diagram (Figure 2.4), valley-floor rock glaciers are subdivided on the basis of location into three groups. The first group, which in terms of occurrence appears to be the largest, includes those rock glaciers that are, or were at some time, connected to and formed downslope from valley glaciers. The second group comprises those rock glaciers that form downslope from small corrie glaciers, and the third group consists of those that develop downslope from ice-cored moraines. All valley-wall rock glaciers form at the base of talus slopes and further subdivision of this type is unnecessary. Rock glacier types are illustrated schematically in Figure 2.5.

The most straightforward valley-wall model, which is periglacial in origin, comprises rockfall debris and ice-rich perennially frozen sediments (matrix ice) (cf. Haeberli, 1985), whereas the glacially-derived valley-floor model requires massive glacial ice and supraglacial debris (cf. Corte, 1976). At this simplified level, which emphasises that rock glaciers are not merely glacial phenomena (as the term itself implies), valley-wall and valley-floor forms are unconnected in terms of inputs and origin. In practice, the two types are rarely this distinct as inputs can vary and in some cases rock glaciers may contain a combination of genetically-distinct ice types and a variety of debris inputs (Figure 2.4).

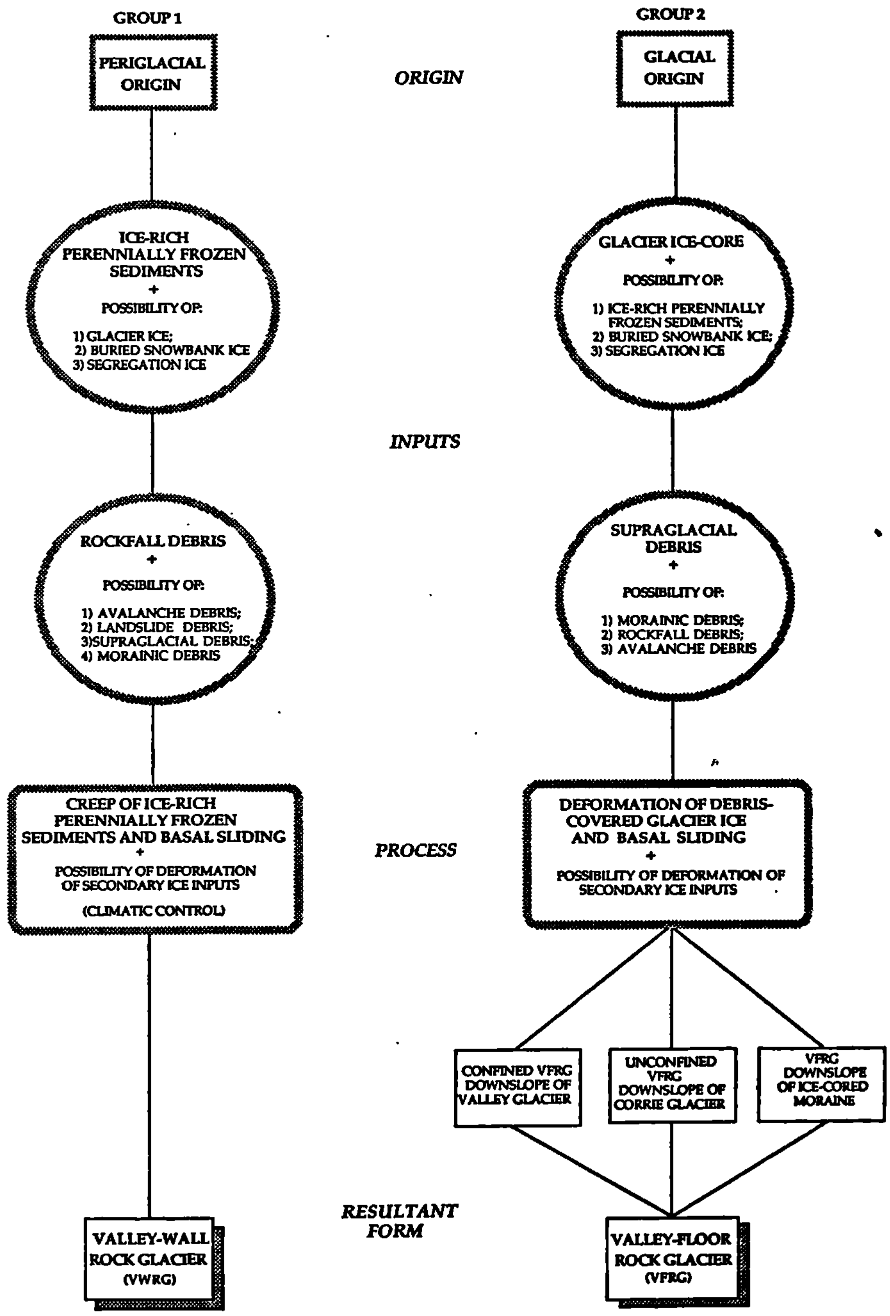


Figure 2.4 Classification of rock glaciers on the basis of origin, inputs and process of formation.

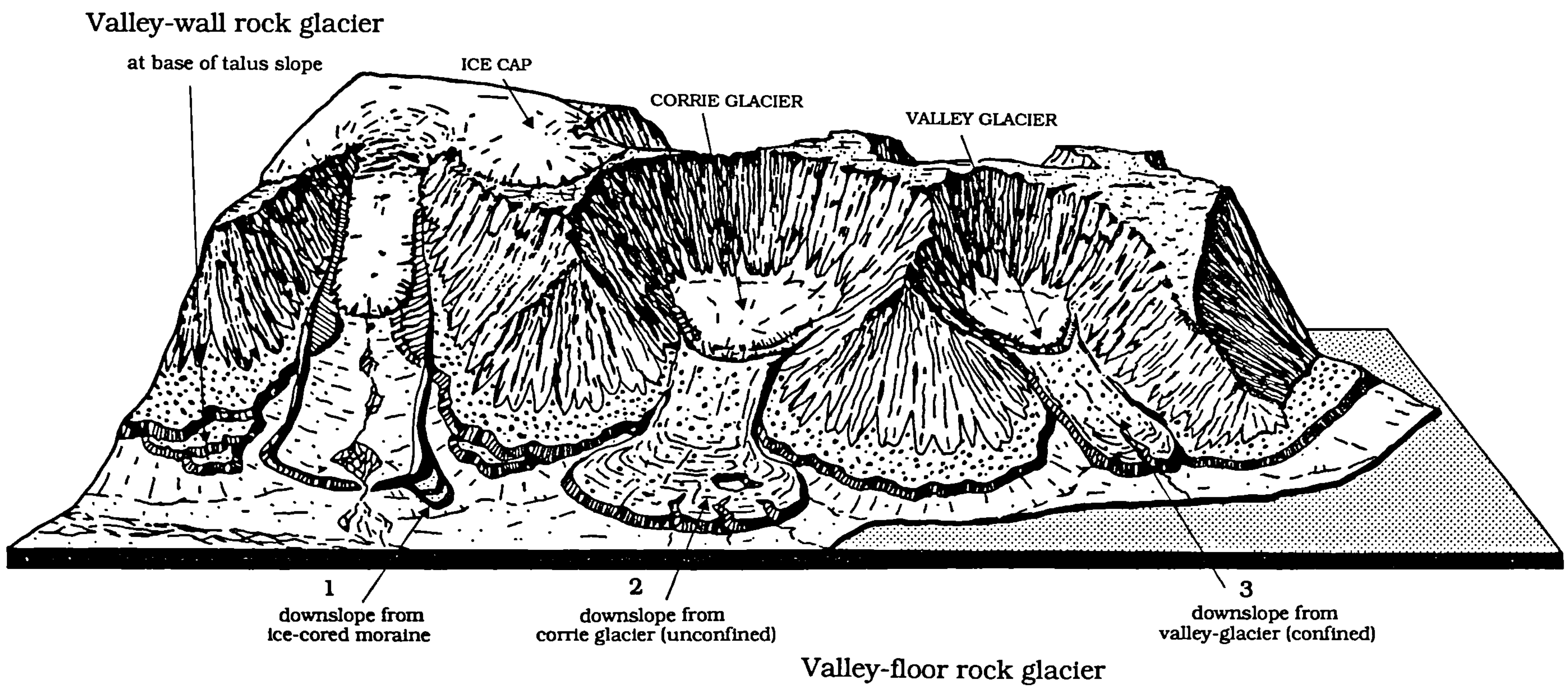


Figure 2.5 Locational classification of rock glaciers (after Humlum, 1982).

In terms of process, the simplest situation for valley-wall rock glaciers describes a steady state, climatically-controlled formation in which creep of ice-rich perennially frozen sediments and rockfall debris produces the resultant form. Thus the climate must be cold enough for permafrost to exist. On the other hand, valley-floor forms are shown in the diagram to result from the retreat of debris-charged glaciers and the subsequent deformation of a massive glacier ice-core. Their formation therefore is not controlled purely by climatic conditions but depends on the initial existence of a glacier, the volume of debris supply, and the rate of glacier ablation. Clearly, however, if matrix ice developed subsequent to retreat, which implies the existence of at least sporadic permafrost, movement may also be due, in part, to creep. In addition, deformation of both snowbank and segregation ice may also contribute to rock glacier movement.

In conclusion, the flow diagram outlined in Figure 2.4 incorporates information on location, genesis, and constituent materials, and represents a partial summary of the ideas presented in this section. Each of these parameters will be reviewed and discussed in more detail in the following sections.

2.3 *Distribution*

2.3.1 *Introduction*

For many years it has been known that active rock glaciers occur primarily in continental sub-polar and mid-latitude alpine environments; they occur only rarely in mountains with abundant precipitation or in lowlands (Wahrhaftig and Cox, 1959; Corte, 1976; Barsch, 1978; Luckman & Crockett, 1978; Washburn, 1979; Morris, 1981, 1987). Less well understood, however, is the relative importance of periglacial versus glacial conditions in explaining the present distribution of both active and inactive rock glaciers (Thompson, 1962; Johnson, 1979; Harris, 1981). Many authors believe that while some rock glaciers require glacial ice for their formation, others require only matrix ice and are therefore not of glacial origin (e.g. Wahrhaftig & Cox, 1959; Outcalt and Benedict, 1965; White, 1971; Potter, 1972; Giardino & Vitek, 1988). Haeberli (1985) and Barsch (1987), however, stated that all rock glaciers are essentially periglacial in origin and form by creep of permafrost, although Haeberli cites examples in which glacial ice occurs within rock glaciers but does not contribute significantly to movement. On the other hand, Whalley (1974) maintained that only features with a glacial origin are true rock glaciers. The extreme viewpoints represented by Haeberli and Whalley are apparently more a reflection of definitional disagreement than differences concerning origin, as Whalley, for example, believes that valley-wall rock glaciers (as defined in section 2.2) should not be termed *rock glaciers*.

2.3.2 *Controls on distribution*

Patterns of rock glacier distribution have been mapped by several researchers (e.g. Corte, 1976; Barsch, 1978; Birnie, 1978; Luckman and Crockett, 1978; White, 1979; Ellis and Calkin, 1979; Harris, 1981; Calkin *et al.*, 1987) who have suggested that a number of environmental and geomorphological factors are critical in determining areas of rock glacier development. Individual selection of these factors, however, appears to have been biased by each author's conclusions regarding rock glacier definition and genesis. For example, Haeberli (1985), who classified rock glaciers as essentially periglacial phenomena, identified three major requirements for rock glacier development: an abundant supply of debris, inclined topography, and permafrost. On the other hand, Whalley, who believed that all rock glaciers are "part of a glacier-debris transport system" (1974, p. 1), stated that glacial ice and not permafrost, was essential for their development, maintaining that "although permafrost can be present in an area of rock glaciers and may prolong the existence of the ice core, it is not necessary for their formation" (Whalley, 1983, p. 1396). Not surprisingly, this uncertainty surrounding the definition of rock glaciers is reflected in the variety of environmental and geomorphological parameters chosen to explain their distribution.

a) Altitude

Altitude has often been studied with respect to its role in influencing rock glacier distribution (Luckman & Crockett, 1977; White, 1979; Morris, 1981, 1987; Humlum, 1984, 1988). Active rock glaciers, which can tolerate warmer temperatures and a more arid climate than active glaciers (Richmond, 1962; White, 1971; Potter, 1972; Kerschner, 1978), generally occupy a narrow altitudinal zone above the tree line and below the zone of glacial ice cover (Luckman & Crockett, 1977; Hassinger & Mayewski, 1983). Two studies that drew distinctions between rock glacier types identified further altitudinal subdivisions within the zone of potential rock glacier distribution. Both Madole (1972) and Ellis and Calkin (1979) noted that throughout their study areas (San Juan Mountains, Colorado, and the east-central Brooks Range, Alaska, respectively), valley-wall rock glaciers occur within an altitudinally-intermediate periglacial zone between valley-floor rock glaciers which are higher and undeformed talus which is lower. These authors all regard valley-wall and valley-floor rock glaciers as morphologically and genetically distinct. Madole found very little altitudinal overlap between each type of deposit, whereas Ellis and Calkin noted that valley-wall rock glaciers occur over a greater altitudinal and geographical range, and have a more variable aspect, than valley floor forms. More specifically, they observed that active valley-wall rock glaciers and talus cones often occur in adjacent sites, and that in such situations topographic factors determine location. Contradictory evidence was proposed by White (1979) who also worked in the San Juan Mountains, Colorado, as he found no link between rock glacier head altitude and form; both valley-wall and valley-floor features occur throughout the zone of rock glacier formation.

Further altitudinal relationships have been proposed concerning rock glacier distribution. White (1979) found that mean altitudes of the uppermost reaches of active rock glaciers in the San Juan Mountains of southwestern Colorado are on average 178m higher than those of inactive rock glaciers, which indicates that temperature has a critical control over activity. Evidence presented by several authors suggests that rock glacier snout elevations could probably be used to establish the lower altitudinal limit for permafrost (Barsch & Updike, 1971; Barsch, 1978; White, 1979). Indeed, Barsch (1978, p.351) stated that "the lower limit of active rock glaciers ... seems to be of greater importance for the description of the permafrost distribution than all other climatically-controlled equilibrium lines".

Recently, Humlum (1988) identified two interesting altitudinally-related concepts for rock glaciers. The first of these, which he termed the "rock glacier appearance level", indicates a critical summit altitude below which rock glaciers do not occur, whilst the "rock glacier initiation line altitude" was defined as the rock glacier head altitude. Humlum proposed that when these altitudes are mapped for a sample of rock glaciers and compared with present-day regional glaciation levels and equilibrium line altitudes, different rock glacier generations can be identified. Moreover, he stated that his methodology will enable the differentiation of glacially and periglacially-derived rock glaciers from the same generation. The validity of this approach, which as yet has been applied only to one small region in Greenland where rock glaciers are numerous, remains to be tested.

b) Aspect

Northerly orientation has been recognised by many as favourable for the development of rock glaciers in the northern hemisphere (Wahrhaftig & Cox, 1959; Blagbrough & Farkas, 1968; Gray, 1972; Carrara & Andrews, 1975; White, 1976; Barsch, 1978), and preferred southerly orientations of both active and inactive rock glaciers were observed by Corte (1976) in the Andes. Local climatic variables such as increased snow drifting and reduced insolation and ablation at these preferred sites are critical in explaining the pronounced asymmetry in rock glacier distribution that occurs throughout many mountain ranges (Carrara & Andrews, 1975; Corte, 1976). Luckman and Crockett (1977) proposed that these same local climatic variables are also partly responsible for the same preferred orientation of cirques and present-day glaciers in Jasper National Park, Alberta. However, they noted that rock glacier distribution exhibits a much stronger preferred aspect than do cirques, which suggested to them that "... rock glaciers have a narrower climatic tolerance than true glaciers" (1977, p. 545).

Although northerly aspects appear to favour the development of rock glaciers (in the northern hemisphere), their location is not limited to north-facing sites and some rock glaciers have been found on slopes facing directly southwards (Barsch, 1978; Ellis & Calkin, 1979; Humlum, 1984). Observations have indicated that, in such situations, rock glacier elevations tend to be higher than on north-facing slopes unless local shading effects compensate for the higher levels of solar insolation received on south-facing slopes. The term *orientation independent* was introduced by Morris (1981) to describe the distribution of rock glaciers whose location is due at least in part to the influence of topographic shading. He established that such topographic shading effects are significant in explaining rock glacier distribution in Alaska.

Ellis and Calkin (1977) noted that in Alaska rock glaciers tend to be 100m higher on south-facing slopes than on north-facing ones. A similar altitudinal differentiation was observed by Barsch (1978), who found that the lower limit of active rock glaciers in the Swiss Alps is consistently 200m less on north-facing slopes as opposed to south-facing ones. Altitudinal relationships between rock glacier activity and aspect were developed further by Ellis & Calkin to include analysis of rock glacier type. They established that on both north- and south-facing slopes, active and inactive valley-wall rock glaciers occur over a wider altitudinal range than active and inactive valley-floor forms. Surprisingly, they discovered that the lower limit of inactive valley-floor rock glaciers was higher on north-facing slopes than on slopes that faced south, and that the lower limit of inactive valley-wall forms on south-facing slopes was higher than the corresponding limit for valley-floor forms. Haeberli (1985) also noted that rock glaciers in the Swiss Alps that are exposed to the south are generally smaller than those rock glaciers that face north at corresponding altitudes.

c) Regional climatic trends

Variations in rock glacier elevation across a region are, however, only partly explicable in terms of the effects of aspect and local shading. For a sample of over one hundred rock glaciers, Luckman and Crockett (1977) identified a significant rise in rock glacier head elevation eastwards across the Jasper National Park area in Alaska. They noted that precipitation decreased and continentality increased from west to east, and from this they deduced that these climatic trends have been directly responsible for the change in rock glacier elevation across the area. Ellis and Calkin (1979) found that in Alaska active rock glaciers are concentrated north of the Continental Divide, and they suggested that this pattern reflects lower temperatures associated with the North Slope's arctic regime as opposed to the more continental climate of the Alaskan interior.

d) Geology

Geological control of rock glacier distribution has been found to manifest itself at two levels. Luckman and Crockett (1977) noted that location is dependent first upon large-scale structural and lithological factors that have determined mountain form and distribution, and second upon bedding and jointing characteristics of specific lithologies, which control the intensity of rockfall processes. They found that in Alberta rock glaciers occur primarily on outcrops of massive quartzite because this rock forms large rockwalls that produce large blocky debris on weathering. The interbedded limestones and shales in the area produce smaller clasts on weathering and do not support rock glacier development. Elsewhere, other evidence has been found which indicates that lithology and rock fractures are important determinants in rock glacier distribution. In Alaska, for example, Wahrhaftig and Cox

(1959) observed that rock glaciers occur mainly on granites, basalts, and quartzites, whereas in the central Andes rock glaciers have largely developed on andesitic and porphyritic rocks (Corte, 1976). More specifically, Gray (1972) found that rock glaciers are absent below fine weathering shales and phyllites in one part of his study area but are in relative abundance in an area of blocky-weathering quartzites and dolomites.

Several authors have also commented on the intensity of rock fractures on rock glacier bedrock source walls. Evin (1985, 1987) recognised that rock glaciers in the southern Alps are often associated with large-scale fissuration in rockwall source areas. She determined that rockwalls that exhibit high joint density are more likely to be associated with rock glacier development than rockwalls with low joint density. Further evidence from fossil rock glaciers in Scotland indicates that bedrock source walls may be controlled by joint planes and faulting in massively bedded rocks (Holmes, 1984). Morris (1981), however, established that while the relative size of rock glaciers is partly controlled by the intensity of rockfall processes, which itself partly reflects joint density in the source rockwalls, jointing only becomes an important determinant of rock glacier development above a certain topoclimatic threshold. His work in the Sangre de Cristo Mountains, Southern Colorado, led him to state that "a highly fractured bedrock wall may not produce a rock glacier in a mild environment, while a highly massive bedrock wall in a severe topoclimate may" (Morris, 1981, p.335). Moreover, Papertezian (1973) working in the Banff area of Canada, noted a strong correlation between the presence of faults and rock glacier development. However, Luckman and Crockett (1977) in a study based on the area just to the north of Banff, proposed that this correspondence is insignificant and merely fortuitous. They concluded that "the major cliffs and cirque walls in which these thrust faults outcrop are also, by virtue of their size, the most suitable topographic sites for rock glacier development" (Luckman and Crockett, 1977, p. 545).

2.3.3 *Factors influencing rock glacier distribution: discussion*

The above review suggests that the development of rock glaciers requires that a number of conditions be met. As rock glaciers require the development or survival of internal ice, temperature and moisture availability appear to be the primary climatic variables controlling rock glacier distribution. Surprisingly, White (1979) found that only very small temperature differences existed between active and inactive rock glaciers in Colorado, which suggested to him that other factors such as bed gradient, debris input, and the depth of the unfrozen surface boulder layer also operate to control activity.

In a study also based in Colorado, Morris (1981) attempted to summarise all the factors that influence the formation and preservation of rock glaciers. He proposed that "the relative size of contemporaneous rock glaciers is controlled by the intensity of the rockfall processes and the preservation of an ice matrix/core which induces flow" (Morris, 1981, p.329). The topoclimatic factors that determine the potential for the preservation of an ice matrix were found to be altitude, insolation reduction by shading, and position of cirque or wall with respect to snow drifting and avalanching. He established that joint density of the bedrock source walls controlled at least in part the rate of production of debris by rockfall.

More recently, Haeberli (1985) suggested a new model (Figure 2.6) of boundary conditions for the distribution of active rock glaciers which, by his definition, are creep phenomena in frozen sediments. On a large-scale he proposed three major constraints. These are:

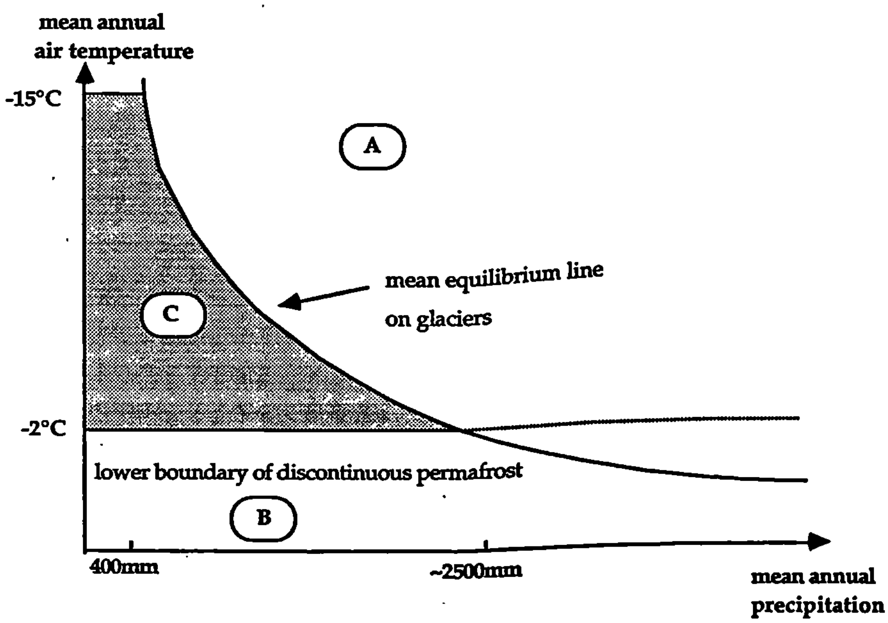
1) mean annual air temperature below about -1°C to -2°C;

2) an annual precipitation less than 2500mm; and

3) a minimum basal slope of 5° to permit creep.

The gradient of 5° is related to the stress distribution in permafrost and to the thickness of perennially frozen sediments. On slopes greater than 30-35° (i.e. those slopes supposedly steeper than the angle of repose for dry accumulation of unconsolidated sediments), Haeberli suggested that the quantity of material necessary for rock glacier formation does not exist in nature.

On a local scale, Haeberli outlined two further factors that he considered important for the origin and development of an individual form. These are:



A : Accumulation zone of glaciers, no debris can accumulate

B : Zone without permafrost, where debris can accumulate but is not perennially frozen

C : Potential zone of rock glacier formation, debris accumulating and can be perennially frozen

Figure 2.6 Temperature and precipitation boundary conditions for rock glacier formation (from Haeberli, 1985).

1) the type of debris source that delivers the material to build up the rock glacier; and

2) the thermal conditions acting at the surface of the creeping permafrost, which are responsible for the formation and existence of ground ice.

On a regional scale, Haeberli noted that the zone of potential rock glacier distribution (below the equilibrium line on glaciers and above the lower permafrost limit) is likely to be absent in wet maritime mountain regions, where the equilibrium line on glaciers is often lower than the lower limit of permafrost. This contention appears to be supported by observations made by Thompson (1962), who noted the apparent absence of rock glaciers in mountains facing the Pacific in North America except on the Alaskan Peninsula at the northern limit of the prevailing west winds in winter. In contrast, Thompson found that rock glaciers were numerous in the drier interior.

In dry mountain regions where the equilibrium line for glacier ice is far above the lower boundary of permafrost distribution, or even lies above the altitude of mountain summits, glaciers may not exist, yet active rock glaciers may be present (Haeberli, 1985). Thompson (1962), who noted that heavy snowfall apparently prevents rock glacier development, believes that rock glaciers can be used as indicators of moderate or light snowfall. Supporting evidence for this contention was provided by Sissons (1979) who inferred that valley-wall rock glaciers in the Cairngorms, Scotland, had developed during the Loch Lomond Stadial under conditions of very limited snowfall. Thus it would appear that a relatively dry periglacial climate is conducive to rock glacier development. Clearly, however, such conditions need not necessarily apply to valley-floor rock glaciers that have developed from the retreat of debris-rich glaciers. Many mountain regions possess transitional climates between maritime and continental, and in such regions Haeberli suggested that rock glaciers and glaciers can exist side by side and sometimes even in direct contact with one another.

Differences between maritime and continental climates and how they affect rock glacier distribution have been studied by Harris (1981). He found that differences between the distribution of valley-wall and valley-floor rock glaciers are critical in determining whether rock glaciers can be used to indicate permafrost limits, as has been suggested by Barsch (1978) and others. He found that active valley-floor rock glaciers have similar limits for freeze/thaw indices to those at the equilibrium line on glaciers, while the limits for active valley-wall rock glaciers show more resemblance to values for climatic stations in zones of permafrost. Harris concluded that it is such differences in the distribution of valley-wall and valley-floor rock glaciers that explain why active valley-floor rock glaciers can be used to indicate the limits of permafrost areas in the Alps but cannot be used in the same way in continental mountain areas, for example in Alberta and the Northwest Territories.

2.3.4 Palaeoclimatic implications of rock glaciers

The work of Wahrhaftig and Cox (1959) in Alaska first established that rock glaciers could be used to reconstruct former climates. Most authors have assumed that the development of rock glaciers is characterised by low temperatures under conditions of relative aridity (e.g. Thompson, 1962; Blagbrough & Farkas, 1968). Three main categories of palaeoclimatic inferences can be drawn from rock glaciers. These are:

1) precipitation (snowfall) inferences based on relationships with contemporaneous glacier limits (e.g. Sissons, 1979; Humlum, 1984, 1988);

2) the presence of discontinuous or sporadic permafrost and a mean annual air temperature of less than *c.* -2°C (note that the value of *c.* -2°C is based on valley-wall rock glaciers and does not necessarily relate to valley-floor forms) (e.g. Barsch, 1977; Haeberli, 1982); and

3) more detailed mean annual air temperature estimates based on altitudinal differences between active and fossil rock glaciers (e.g. Kerschner, 1978).

Recent palaeoclimatic reconstructions based on rock glaciers use the distribution of fossil rock glaciers and the reconstructed equilibrium line altitudes of contemporaneous glaciers to estimate palaeo- temperatures and former precipitation totals (e.g. Kerschner, 1978; Haeberli, 1982; Humlum, 1984). Fundamental to this approach is the conception of rock glaciers as essentially permafrost features, such that their lowermost occurrence indicates the lower limit of discontinuous permafrost at the time of rock glacier development.

Several authors have found that the lower limit of development of active valley-wall rock glaciers appears to relate to mean annual air temperature of about -2°C (Barsch, 1977c; Kerschner, 1980; Haeberli, 1982). Humlum (1984) noted, however, that the altitudinal trend of valley-wall rock glaciers also appears to be related to precipitation, and tentatively suggested that fossil valley-wall rock glaciers might also represent a source of information on palaeoprecipitation.

Kerschner (1978) attempted to reconstruct palaeoclimatic parameters from a study of large fossil rock glaciers of inferred Younger Dryas age (*c.* 11,000-10,000 yr. B.P.) situated at 500 to 600m below a belt of presently-active rock glaciers in the Western Tyrol. From the altitude of the lower limit of fossil rock glaciers Kerschner calculated a decrease in annual temperature from the present of *c.* 3-4°C. By analysing the equilibrium line on present-day glaciers in the same area, he found that summer temperatures had apparently decreased by *c.* 2°C. He therefore deduced that a more continental climate with cooler summers (2°C below present) and much colder winters (5-6°C below present) existed in this part of the Alps during Younger Dryas time. Clearly, however, further investigations are needed to validate this technique, particularly in areas where cross checks can be made.

2.4 Rock glacier morphology

2.4.1 Introduction

Much literature on rock glaciers confines itself to morphological description. Such studies vary from detailed investigations of single features (e.g. Potter, 1972) to regional morphometric analyses of large samples of rock glaciers (e.g. White, 1979). Considerable amounts of information have been collected in such surveys, although speculation about the origin of rock glaciers has, as already mentioned, resulted in some confusion concerning nomenclature (J.P. Johnson, 1974).

In general terms, rock glaciers are masses of debris with a surface topography that gives the impression of slow plastic or viscous flow (Whalley, 1974; White, 1976; Haeberli, 1985). The downslope and lateral margins of active rock glaciers are generally well-defined and are usually characterised by steep slopes, the gradient of which corresponds to the "angle of repose" for rockfall debris (Wahrhaftig and Cox, 1959, p. 387). In this dissertation the angle of repose for rockfall debris is taken to lie between the angle of initial yield and the angle of residual shear (cf. Statham, 1976, 1977). The angle of initial yield, which determines the maximum angle a talus slope will attain before avalanching, is generally a few degrees steeper than the angle of residual shear. If debris supply is continuous, the maximum slope angle will be determined by the angle of residual shear.

At the surface of rock glaciers a relatively thin layer of coarse debris overlies a diamicton containing much finer grained sediments which constitute most of the rock glacier volume (Fisch *et al.*, 1977). The depth of the coarse debris mantle, usually no more than a few metres, is often evident at the frontal margins of most active rock glaciers (Haeberli, 1985). Around the base of the frontal and lateral slopes a talus apron is usually found (White, 1976). This talus is believed to have slipped and rolled down from the coarse surface layer as a result of rock glacier movement (Smith, 1973), which also prevents vegetation from developing on the marginal slopes of active features (White, 1979). The presence of vegetation on the frontal faces of rock glaciers is, therefore, a commonly used indicator of inactivity (Corte, 1976).

2.4.2 Surface microrelief

Rock glaciers often have a very well-developed complex surface microrelief which Haeberli (1985) has attributed to two processes, namely "the production and disappearance of ice at the surface, and the variation of rock glacier flow in space and time " (Haeberli, 1985, p. 15-16). Each process is thought to produce a distinctive group of microrelief forms. Specifically, those surface features which occur randomly across rock glaciers are believed to correspond to the first processes, whereas others exhibiting more pronounced regularities have been directly related to the movement of the rock glacier (Wahrhaftig and Cox, 1959; Haeberli, 1985).

Irregular forms such as meandering and closed depressions and small knolls are common microrelief features. Many such features were described by Blagbrough and Farkas, (1968) who likened them to glacial kame and kettle topography. A similar irregular pattern of lakes and depressions on rock glaciers in the Andes was observed by Corte (1976) who described the microrelief features as "thermokarst facies". Such features have been interpreted as the result of differential melting of ground ice (Corte, 1976; Haeberli, 1985). Additionally, many of the prominent longitudinal ridges and depressions that frequently extend down the length of some rock glaciers are also believed to be formed by wastage of subsurface ice (Wahrhaftig and Cox, 1959; Potter, 1972).

Regular forms, notably transverse ridges and depressions, are known to be directly connected to rock glacier movement (Messerli and Zurbuchen, 1968). Transverse depressions have been interpreted as the result of compressive flow and are most common near the downslope margins of rock glaciers (Haeberli, 1985). Wahrhaftig and Cox (1959) noted similarities between transverse ridge systems on rock glaciers and those on lava flows and glaciers and observed an increase in rock glacier thickness at transverse ridges, which would be expected if velocity had decreased and width was restricted. Some longitudinal ridges and depressions also reflect rock glacier movement and have been shown to develop where rock glaciers exhibit extending flow (Messerli and Zurbuchen, 1968).

The exact mode of formation of transverse ridges and depressions, however, remains unclear. Some researchers have proposed that they are the result of internal thrusting along shear planes (White, 1987), whereas others have suggested they form because of a decrease in viscosity with depth (Wahrhaftig and Cox, 1959). In addition to these uncertainties, Haeberli observed that surface microrelief on active rock glaciers may not relate to present-day stress fields but instead may reflect past movement.

Several studies have reported crude sorting and orientation in rock glacier sediments that may relate to their formation or movement

(Wahrhaftig & Cox, 1959; Barsch, 1971; Potter, 1972; Yarnal, 1982; Giardino & Vitek, 1985). For example, trend surface techniques, which were used to evaluate the nature and distribution of the surface sediments of a rock glacier in British Columbia (Yarnal, 1982), indicated that long-term rock glacier creep causes downward sifting of finer debris and preferred downslope orientation of clasts. Such sediment distribution patterns, however, may take hundreds of years to develop. Clast alignment was found by Giardino and Vitek (1985) to have a tendency to approach parallel to the trend of a surface feature such as a ridge or depression. They suggested that rock glacier fabric could be used to differentiate rock glaciers from other morphologically-similar phenomena such as ice-cored moraines and kame deposits.

Despite a variety in planimetric form, most rock glaciers generally show similarities in terms of surface morphology; indeed such similarities are commonly used in rock glacier identification either in the field or from aerial photographs (e.g. Outcalt and Benedict, 1965; Smith, 1973), although morphological similarities with landforms such as ice-cored moraines has led to some problems in identification (Barsch, 1971; Østrem, 1971). However, some large-scale morphological differences do exist between valley-wall and valley-floor rock glaciers and these require clarification.

2.4.3 Valley-floor rock glaciers

Valley-floor rock glaciers are tongue-shaped masses of poorly sorted angular debris that extend downvalley, either from the lower end of valley or corrie glaciers or from ice-cored moraines. The length, breadth, and height dimensions of these features vary widely. In their study of Alaskan valley-floor rock glaciers, Wahrhaftig and Cox recorded lengths of from 100-1600m, widths of 70-750m, and heights of 10-80m, although they also noted an extreme case where a rock glacier attained a height of 122m. Similar lengths were also recorded in White's (1971) study of valley-floor rock glaciers in the Colorado Front Range.

The surface microrelief of valley-floor rock glaciers is usually very pronounced (Wahrhaftig and Cox, 1959). Transverse ridges and depressions at right angles to the long axis and generally convex downslope in plan are often developed near the break in slope between the main mass of such features and the steep frontal face (White, 1979). Where transverse features exist a concave break in slope is present immediately upslope of the most proximal ridge (Wahrhaftig and Cox, 1959). Longitudinal ridges and depressions commonly extend down the length of such features and on some rock glaciers the longitudinal depressions join downvalley, thereby truncating the intervening ridges (Outcalt and Benedict, 1965). Both transverse and longitudinal ridges commonly have basal widths of 3-10m and have rounded crests, which are usually 1-10m above the depressions. Depressions have basal widths of 1-5m and are generally V-shaped in cross section (Wahrhaftig and Cox, 1959). Upslope of transverse ridges, the rock glacier surface may be characterised by small knolls and depressions (Potter, 1972). Large spoon-shaped depressions occur at the heads of many valley-floor rock glaciers. The origin of these depressions has been explained by differential ablation of an ice tongue that is protected downslope by a thick cover of insulating debris, but relatively free of debris in an upslope direction (Wahrhaftig and Cox, 1959; Outcalt and Benedict, 1965). Surface morphology on relict valley-floor rock glaciers is more subdued, and overall height is generally much less (Corte, 1976).

2.4.4 Valley-wall rock glaciers

Valley-wall rock glaciers are single or multiple lobes of debris that extend out from the base of talus slopes and are generally as broad and sometimes much broader than they are long. Wahrhaftig and Cox (1959) noted that almost 90% of the valley-wall rock glaciers in their study area were less than 500m long, although some reached widths of 3500m. In comparison with valley-floor rock glaciers which trend downvalley, valley-wall rock glaciers usually have a very limited travel distance towards the centre of the valley, with little downvalley deflection.

The microrelief structures of some valley-wall rock glaciers can be very complex with numerous sets of transverse ridges and furrows that may merge, bifurcate, or die out at irregular intervals (Smith, 1973). Other valley-wall forms may have much simpler structures, perhaps consisting of only one transverse ridge (Dumbell, 1984). Longitudinal ridges are not as common on valley wall rock glaciers and where they do occur tend to die out or become discontinuous before they reach the downslope margin of the rock glacier (Smith, 1983). Transverse ridge asymmetry has been noted on rock glaciers in the Andes and in Norway (Corte, 1976; Dumbell, 1984). Dumbell observed that the distal slopes of transverse ridges were consistently steeper by several degrees than the proximal slopes.

2.4.5 Discussion

Surface morphological differences do not always exist between valley-wall and valley-floor forms. In addition, several authors have noted that even large-scale rock glaciers may exhibit few surface features on which possible differentiation could be based (e.g. Wahrhaftig & Cox, 1959; Lliboutry, 1977; Haeberli, 1985). Lliboutry observed that rock glaciers in their earliest stages of development in the Andes of Santiago often do not contain ridges and depressions. He stated that such surface forms, therefore, should not be considered as fundamental or characteristic features of rock glaciers. Thus, surface morphology cannot be used as the sole basis from which valley-wall forms may be distinguished from valley-floor rock glaciers.

2.5 Subsurface information

2.5.1 Introduction

The internal constitution of rock glaciers is poorly understood, largely because the extremely coarse nature of the surface boulders and the presence of ice in active rock glaciers makes excavation very difficult (White, 1976). During the past few years, however, several investigations have provided important new data on the internal structure of these features, data that may help in the determination of rock glacier flow mechanisms. This section summarises the subsurface information that is currently available.

2.5.2 *Rock glacier excavations*

Several methods of examining the internal structures of rock glaciers have been employed. The most direct method of investigation is trenching, and several attempts at excavation have been made. As early as 1910, Capps dug into seven active valley-floor rock glaciers in Alaska and in each case encountered ice. Although Capps used the term *interstitial* and commented that the ice was not massive, he stated that the ice often formed with the clasts as "a breccia with the ice as a matrix" (1910, p.362). Thus, it is reasonable to assume that Capps found matrix ice in which the clasts were, at least in part, supported by ice. At each site he was able to reach clast-free ice at a depth of about 0.5m near the heads of the rock glaciers; he was unable, however, to reach ice in shallow excavations near the rock glacier fronts. From these early observations, Capps attributed rock glacier flow to the creep of ice that he inferred to have formed through the freezing of meltwater or rain. Most other early workers who observed matrix ice within active rock glaciers (e.g. Ives, 1940; Wahrhaftig and Cox, 1959; Thompson, 1962; Johnson, 1967; Blagbrough and Farkas, 1968), agreed with Capp's hypothesis, although others (e.g. Barsch, 1969; Shroder, 1973) felt that matrix ice could explain the movement of some but not all rock glaciers.

More recent excavations have also revealed the presence of both matrix and interstitial ice. In 1971, White dug into the front of the Arapaho rock glacier in the Colorado Front Range and encountered matrix ice at a depth of 0.5m below the surface boulder layer. Fisch *et al.* (1977) excavated a rock glacier within the Wallis Alps in Switzerland to a depth of 10m and found "angular debris cemented by interstitial ice beneath a thin layer of unfrozen coarse blocky debris at the surface" (Fisch *et al.*, 1977, p.245). An analysis of the ice in which the debris fragments were embedded revealed that the size of the ice crystals was "far too small to be of glacial origin" (Fisch *et al.*, 1977, p.245).

Lenses and masses of debris-free ice of non-glacial origin have also been observed within rock glaciers. In the central Andes in Argentina, Wayne (1981) dug into the side of an active valley-wall rock glacier that was unconnected to glacial ice and found ice within the rock debris approximately one metre below the base of the surface boulder layer, which at the excavation site was also one metre thick. At three metres depth (one metre below the permafrost table) Wayne encountered a lens of clear ice 0.1m thick. Below this ice lens, rock debris was found within a frozen sand-silt matrix. Wayne proposed ice segregation as an origin for such lenses of non-glacial ice. He suggested that if a layer of permafrost exists in talus, groundwater percolating through the talus below the frozen near-surface layers will be confined under sufficient hydrostatic pressure to keep it in contact with the base of the permafrost, and segregation of ice lenses will occur. Wayne also reported that further excavation nearer the centre of the rock glacier was impossible because of the thickness of the boulder cover. From excavations made in conjunction with surface mine workings in Colorado, Vick (1981) noted that both interstitial and clear ice occurred within valley-wall rock glaciers. At one rock glacier site, he found a layer of clear ice 0.9 to 1.1m thick that extended over a distance of 45m. The ice layer contained several horizontal silt seams, which according to Vick, indicated that the ice had originated as snow drifts that had accumulated over several seasons. A similar origin for such layers of clear ice within valley-wall rock glaciers had been suggested by Outcalt and Benedict (1965).

A major problem of such limited excavations is that they fail to detect any spatial variations in ice content or type. For example, White (1976) reported that while his excavation of a downvalley section of Arapaho rock glacier revealed only interstitial ice, earlier work (White, 1971) had established that the upvalley part of the rock glacier contained an ice core. Thus data obtained solely from limited excavations cannot always be considered representative, and therefore does not provide a reliable base from which to distinguish between ice-cored and "ice-cemented" rock glaciers, particularly as the former may grade into the latter (White, 1976).

More extensive rock glacier excavations have also been undertaken. During the construction of a new highway through the Sangre de Cristo Mountains in southern Colorado, for example, a rock-crushing machine was used to obtain material for the highway base from a valley-floor rock glacier (Johnson, 1967). The excavation encountered ice at valley floor level at a distance of about 10m from the front of the rock glacier. This ice hampered excavation and operations were subsequently moved to the upper surface of the rock glacier. At this site ice was encountered at depths of three to six metres. Johnson noted that the "depressions and ridges in the highly irregular surface of the ice closely parallel the flow structures on the surface of the rock stream. The ice forms the matrix of the rock stream and apparently permits the rock stream to flow downhill in a manner similar to that of a normal valley glacier composed entirely of ice" (Johnson, 1967, p. D217-8). He also reported that the rock glacier contained a large amount of fine debris, although very little of it was incorporated in the ice, and suggested that water from melting snow drains quickly downwards through the openwork boulder deposits to the surface of the ice where it may form a new ice layer.

Other excavations have revealed buried glaciers as ice cores within rock glaciers rather than non-glacial ice. Miners tunnelling into a rock glacier in the San Juan Mountains of Colorado exposed a core of ice, which Brown (1925) interpreted to be of glacial origin. Elsewhere, Outcalt and Benedict (1965) reported that in the Colorado Front Range they found many examples of true ice glaciers grading downvalley into debris-covered glaciers and finally into rock glaciers with ice cores. Benedict (1973, p. 520) further reported that "during the summer of 1966, erosion by a meltwater stream [in the Arapaho rock glacier, Colorado Front Range] exposed a 220 metre long vertical section of buried ice, extending along the axis of the rock glacier from the shallow depression at its rear to a position about 400 metres behind its front". The exposed ice varied in thickness from 1.0 to 9.8 metres and was essentially debris-free, though it contained a few stones up to 60mm in size. Debris overlying this ice core varied in thickness from 0.2 to 2.4 metres and was composed of two units: an upper layer of large angular boulders, which Benedict attributed to rockfall debris, and a basal layer of poorly sorted sand and gravel containing a few cobbles, which Benedict interpreted as ablation till due to ice melt at the base of the unfrozen surface boulder layer.

Ice samples obtained from the ice core of the Arapaho rock glacier were compared by Benedict with ice samples obtained from recent snow bank ice. The comparison revealed that the ice forming the core of the rock glacier was of glacial origin and was markedly different from the ice of snowbank origin in terms of the size and shape of individual ice crystals, the degree of interlocking boundaries, and the number and distribution of air bubbles within the samples.

One interesting subsurface feature noted by Potter (1972) was the presence of crude sorting in the surficial debris layer of the rock glacier. Samples removed for size analyses showed that coarse fragments are dominant at the surface and that a zone of fines exists just above the debris-ice contact. Three processes, which were initially proposed by Wahrhaftig and Cox (1959), were used by Potter to explain the predominance of coarse blocks on the surface:

1) fine material sifts downwards between the large surface clasts due to rock glacier movement;

2) rainfall and meltwater washes fine material downward; and

3) abrasion of angular clasts beneath the surface produces fine debris.

Potter further suggested that the relatively high percentage of finer material found near the base of the debris mantle "may in part also reflect addition of material from the underlying ice" (p.3042), though he produced no supporting evidence for the operation of this process.

Excavations have shown therefore that valley-floor rock glaciers contain a core of glacial ice that may grade downvalley into either matrix or interstitial ice of non-glacial origin. Glacial ice, however, has yet to be found within valley-wall rock glaciers. Instead, matrix, interstitial and segregation ice have all apparently been found within valley-wall rock glaciers.

2.5.3 *Geophysical investigations*

The use of drilling and geophysical techniques such as resistivity profiling and bore-hole investigations can be effective in investigating the internal structure of rock glaciers (e.g. Potter, 1972; Fisch *et al.*, 1977; Barsch *et al.*, 1979; Hassinger and Mayewski, 1983). Bore hole measurements and shallow core drillings were carried out on an active rock glacier near the Grubengletscher, Wallis, Swiss Alps by Barsch *et al.* (1979), who concluded from the evidence obtained that the rock glacier "is permafrost, by definition, and the material forming the rock glacier is frozen silty sand very rich in ice" (p.224). They determined that the ice content of the rock glacier was approximately 90% by volume just below the base of the active layer, which was established to lie at a depth of about 0.55-0.85m, near the base of the openwork bouldery layer. They suggested that ice content reached an average value of 50 to 70% by volume several metres below the level of the permafrost table. Core analysis revealed that ice was present in the pore spaces and as coatings around the rock particles. In many parts of the core, ice also occurred in the form of lenses and layers up to 0.6m thick.

Seismic refraction soundings taken on the same rock glacier confirmed some of Barsch's conclusions. P-wave velocity values presented by Haeberli *et al.* (1979) indicated a decrease in ice content by about 30% within the uppermost few metres beneath the permafrost table. Mean P-wave velocity values were also used by Haeberli *et al.* to deduce that the mean ice content of the frozen debris was about 50%, which is in reasonable agreement with Barsch's proposed average ice content value of about 50 to 70%.

From drilling undertaken in the same region, Barsch (1978) discovered that average particle size decreased with depth, noting that below the blocky surface boulders are rare and embedded in material of gravel to silt size. In the same paper, Barsch stated that every seismic sounding he had taken on active rock glaciers at the base of talus slopes or below the snouts of glaciers indicated a composition of frozen talus (this, however, excludes those rock glaciers defined by Barsch as glaciers completely covered by rubble; see section 2.2.3). Clear ice, either in lens or layer form, was explained by Barsch as the product of segregation, or as ice derived from snow avalanches. He concluded that there was no evidence to suggest that the ice is glacier ice.

Geophysical data obtained from rock glaciers outside the Swiss Alps are generally similar to those presented by Barsch (1978), Barsch *et.al.* (1979) and Haeberli *et al.*, (1979). For example, in 1983 Hassinger and Mayewski used a combination of shallow seismic refraction and electrical resistivity profiling to study the internal structure of Antarctic rock glaciers. From their results they constructed a structural compositional model based on eleven rock glaciers, which consisted of an ice-free layer from 0.3 to 3.0m thick, underlain by either frozen debris containing "interstitial ice" or ice. They stated that precise determination of the ice content in the various refractor zones inferred from the geophysical evidence was in some cases impossible because of interpretational difficulties of the data.

A geophysical study of a single rock glacier from the Absaroka Mountains, Wyoming, by Potter (1972) also emphasised the limitations of seismic refraction techniques. By using both seismic refraction profiling and limited excavations, Potter suggested that the rock glacier was ice-cored. He concluded that the upvalley two-thirds of the rock glacier are composed of a 1.0-1.5m thick continuous layer of debris over relatively clean glacier ice, and that the downvalley one-third has a thicker debris mantle of the order of 2.0-3.0m thick over ice of unknown debris content. Seismic wave velocities in the debris mantle ranged between 240 and 340ms^{-1} and those of the underlying ice between 2440 and 3350ms^{-1}. Potter suggested that the results were accurate to within 10 to 15% for true debris depth. He did not know, however, whether the higher velocity material beneath the debris mantle in the downvalley third was ice-rich frozen debris or clean ice because he found that the seismic refraction profiles provided insufficient evidence to distinguish between frozen ground and clean ice. The conclusions drawn by Potter regarding the internal structure of the rock glacier are thus based largely on limited shallow excavations made in the debris mantle. A re-evaluation of the Galena Creek rock glacier by Barsch (1987b) found that "all evidence supports the fact that no difference occurs in geomorphology, seismic information, rheology, etc., from a *normal* ice-cemented rock glacier. Therefore, it seems appropriate to abolish the model of the so-called *ice-cored rock glacier* "(p.52). His viewpoint seems rather extreme considering the numerous well documented examples of glacially derived valley-floor rock glaciers (e.g. Wahrhaftig & Cox, 1959; Corte, 1976).

Some doubts exist regarding the usefulness of seismic velocities for determining the sub-surface nature of rock glaciers. Citing Rothlisberger (1972), Hassinger and Mayewski (1983) agreed that P-wave velocities of clean ice and material with interstitial ice overlap within a certain range and may be indistinguishable. Barsch (1978) also recognised some of the problems inherent in geophysical investigations. He found that the seismic wave velocity of frozen talus is between 3000 and 4000ms^{-1}, whereas for glacier ice the velocity is around 3600ms^{-1}. Rothlisberger

(1972) noted that such variability is characteristic for frozen material and is due to changes in ice content from place to place. Seismic refraction techniques therefore require further refinement before they can be applied with confidence to future studies of the internal characteristics of rock glaciers.

Electrical resistivity profiling has been shown to be a more powerful tool than seismic refraction in subsurface investigations of rock glaciers, although as yet few studies have adopted this approach. Fisch *et al.* (1977) carried out extensive electrical resistivity soundings on two active rock glaciers in the Walliser Alpen, Switzerland. From their results they concluded that the two study rock glaciers consisted of "perennially frozen debris and not of glacier ice nor of buried snowbank ice as it is often suggested" (p.239). Perennially frozen debris was defined by them as "interstitial ice or perhaps also ice lenses/segregated ice" (p.255). Resistivity values measured within the rock glaciers ranged between 10,000 Ωm and 300,000 Ωm. Such values contrast markedly both with those of temperate and slightly cold alpine glacier ice or buried snowbank ice (10^6 Ωm to more than 10^7 Ωm), and with those of unfrozen debris, which in high alpine regions are of the order of 1000 to 5000 Ωm. Resistivity determinations are therefore appropriate to discriminate between ice, perennially frozen debris, and unfrozen debris. As disclosed in the previous section (2.5.2), excavations carried out at both study sites confirmed that the rock glaciers are composed of perennially frozen debris. Only in the rock glacier head region was buried glacier ice observed, and even then it was surrounded by thick frozen sediments.

Mean thickness of the permafrost layer within the rock glaciers was reported to be of the order of 20 to 30m with a maximum value of about 70m. Below the high resistivity permafrost layer, a thick lower resistivity layer was encountered. This Fisch *et al.* interpreted as partly or completely unfrozen material, which suggested to them that rock glaciers are not frozen to the underlying bedrock and that rock glacier movement may be due in part to basal sliding.

2.5.4 Hydrological investigations

Johnson (1981) adopted an alternative approach towards investigating the subsurface constituents of a rock glacier in the Yukon. Using tracing experiments he identified two independent drainage systems within a valley-wall rock glacier. The first drainage system was composed of surface and near-surface water draining both talus and rock glacier. The second deeper system was due to the drainage of Grizzly Creek, a stream that flowed from an adjacent moraine into the side of the rock glacier and had its resurgence at the terminus of the rock glacier. The resurgence of Grizzly Creek occurred along a sharply-defined boundary across the whole width of the rock glacier. Johnson hypothesised that the planar nature of the drainage surface was due to "an interstitial ice content rather than a sedimentary structure or a massive ice core" (p.1429), partly on the grounds that planar sedimentary structures are unlikely to occur within a flow landform with a contorted surface structure. He argued that the internal frozen surface could not be due to an ice core because the morphology of ice-cored landforms has consistently been shown to be more irregular than the surface of the study rock glacier.

Although Johnson's (1981) paper is useful, the unusual characteristics of this study site preclude wider application of this methodology in determining the internal structure of rock glaciers. Thus electrical resistivity profiling appears to be the most effective tool currently available for investigating the internal structure of rock glaciers because unlike seismic investigations it is possible to discriminate clearly between different facies that constitute the rock glaciers. Matrix ice has been found in all valley-wall rock glaciers and both glacial and matrix ice occurs in valley-floor rock glaciers. Of particular interest has been the discovery that the basal layers of some active valley-wall rock glaciers may be partly unfrozen, suggesting that rock glacier movement may include some degree of basal sliding.

2.6 Rock glacier movement

2.6.1 Introduction

Rock glacier movement is neither well understood nor easily investigated. Various techniques have been employed to study movement and these can be grouped into three main categories that form the basis of this section. Geodetic investigations have been the standard method for many years and several long-term studies have furnished important information about changes in annual movement (e.g. Jackson & MacDonald, 1980; Benedict *et al.*, 1986). Photogrammetrical observations are becoming more precise and present an overall picture of the distribution of surface velocities (Haeberli *et al.*, 1979). Other methods, such as detailed studies of rock glacier frontal movement (Wahrhaftig & Cox, 1959), and tree-ring analyses (e.g Giardino *et al.*, 1984; Shroder & Giardino, 1987), provide additional information on the development of rock glaciers.

2.6.2 *Geodetic investigations*

For many years, geodetic surveys have been used to study rock glacier movement (e.g. Barsch, 1969; White, 1987). The most common procedure involves measuring the displacement of painted boulders from an initial theodolite-constructed profile. Surface boulder displacements, however, provide information on surface movement but do not necessarily reflect the movement of sub-surface material. The accuracy of geodetic surveys depends largely on being able to relocate and resurvey target boulders, and evidence suggests that problems of relocating undisturbed boulders are considerable (e.g. White, 1971). However, White estimated that the error limits of this technique, which are approximately ±0.002m, are relatively low compared with average annual movement rates, and noted that velocity values obtained from long-term geodetic surveys are in reasonable agreement with data obtained from photogrammetric or other analyses.

Average annual velocities of rock glacier surfaces have been obtained from many studies and the results are summarised in Table 2.2. Kerschner (1978), compiled a list of 29 cases from the literature that gave a

ROCK GLACIER MOVEMENT m/year	*LOCATION*	*AUTHOR*
0.40 to 1.61	Swiss Alps	Chaix, 1923
0.36 to 0.69	Alaska	Wahrhaftig and Cox, 1959
0.05 to 0.23	Colorado	White, 1971
0.3 to 0.60	Alberta	Osborn, 1975
0.03 to 2.38	Swiss Alps	Barsch& Hell, 1975
0.05 to 1.00	Worldwide survey	Barsch, 1977
0.05 to 3.57	Worldwide survey	Kerschner, 1978
0.02 to 0.82	Swiss Alps	Haeberli et al., 1979
0.20 to 1.30	Middle Asia	Gorbunov, 1983
0.06 to 0.19	Colorado	Benedict et al., 1986
5.0	Swiss Alps	Pillewizer, 1957
0.68	Colorado	Bryant, 1971
0.80	Wyoming	Potter, 1972
3.35	Peru	Lliboutry, 1977
0.08	France	Evin, 1985
0.03	Antarctica	Hassinger & Mayewski, 1983

Table 2.2 Average annual surface movement of rock glaciers.

range of values between 0.05-3.57my^{-1}, with a mean of 0.75 and a standard deviation of ±0.75my^{-1}. The greatest observed value of 5.00my^{-1} was obtained on a valley-floor rock glacier in the Alps that had a gradient significantly greater than that of the average surface inclination for rock glaciers of between 15° ± 5° (Kerschner, 1978). Wahrhaftig and Cox (1959) noted that surface velocities for valley-floor forms are generally higher than those of valley-wall rock glaciers but supporting evidence for this is hard to find in the literature as few authors differentiate between rock glacier types when presenting velocity data.

Geodetic surveys have also indicated that velocity profiles across active rock glaciers adopt a parabolic form in the same manner as glaciers, with the highest velocities occurring towards the centre of the feature (Barsch, 1969; Benedict *et al.*, 1986; White, 1987). Annual velocities, however, change both in amount and in direction of displacement (e.g. Osborn 1975; Jackson & MacDonald, 1980). For example, a 25-year resurvey of Arapaho rock glacier, Colorado, demonstrated that rates of movement varied considerably from year to year (Benedict *et al.*, 1986). In addition, White (1971) found that surface velocity values on Arapaho, the same valley-floor rock glacier, decreased downslope from the middle of the rock glacier to the front, indicating that compressive flow is taking place in the downvalley half of the feature. A downvalley decrease in velocity values could also indicate net ablation as with glaciers. Supporting evidence for such a decrease in velocity was provided from a study of a valley-floor rock glacier in Tungsten, North West Territories, by Jackson & MacDonald (1980). They suggested, however, that at their study site the decrease in velocity may be the result of lateral expansion of the snout, which was not possible upslope due to constriction by the valley walls. Jackson and MacDonald argued that surface velocity was not consistently proportional to surface slope and pointed to other factors such as variations in thickness, underlying topography, and ice content to explain differential mobility. Haeberli (1985) also noted that thickness of the rock glacier permafrost strongly influenced movement rates, presumably by increasing shear stresses.

It is interesting to note that rates of creep in discontinuous permafrost are considerably less than average annual velocities of rock glacier surfaces. Savigny (1980), for example, measured permafrost creep in an area of discontinuous permafrost in the Canadian Arctic and reported velocities of 2.5-3.0 mm yr^{-1}. The measured velocities appeared to be closely related to the distribution of ice lenses; no internal deformation was established.

Differential movement across transverse ridges on Arapaho rock glacier was observed by White (1987), who constructed a network of lines across three transverse ridges and measured the displacement of large boulders over a period of eight years. Extensional movement was noted near the rock glacier centre with compression near the margins, confirming earlier observations of flow velocity increases towards the centre of the rock glacier. However, movement parallel to, transverse to, and diagonal to flow direction alternated repeatedly, extending one year and compressing the next. He interpreted these "accordion-like" movements as the result of possible internal overthrusting along internal shear planes because of pressure exerted at the downslope end of relict buried glacier ice inside the rock glacier. Such a theory is plausible but very difficult to test, as field evidence of internal shear planes would be difficult to obtain.

Seasonal variations in rock glacier movement have been studied recently using geodetic techniques on the Gruben rock glacier in the Alps (Haeberli, 1985). Higher rates of movement in the lower parts of the rock glacier occurred in autumn and early summer, and slower movements in winter and early spring. Conversely, the upper part of the rock glacier showed movement peaks in early spring and minima in winter and early summer. Haeberli likened this pattern of movements to those of an earthworm which advances its anterior end first and then pulls up its posterior end. He noted that similar observations have been made on temperate glaciers and are believed to relate to water flow conditions (Hodge, 1974).

2.6.3 Photogrammetric investigations of rock glacier movement

Photogrammetrical determinations of rock glacier movement are normally made by measuring displacements of large distinct boulders on rock glaciers over time periods determined by the dates of the aerial photographs. Difficulties are sometimes encountered in re-identifying targets if the quality or contrast of the photographs varies (Haeberli, *et al.*, 1979). Error limits for photogrammetry were approximated at ±0.05m by Barsch and Hell (1976), which are high compared with some annual velocities. Horizontal surface velocities measured using photogrammetry range from a few centimetres per year to a metre per year for rock glaciers with typical surface inclinations (Kerschner, 1978). Flow lines constructed for the Gruben rock glacier by Haeberli *et al.* (1979), indicate that movement largely follows the general valley-slope and that flow vectors are in most cases at right angles to the contour lines of the rock glacier surface. By relating flow lines to surface structures they noted that an absence of transverse ridges on some rock glaciers can be attributed to the predominance of extensional flow. They suggested that many rock glaciers without transverse ridges may not have been recognised as rock glaciers, especially small initial forms. They further suggested that incipient valley-wall rock glaciers may often have been incorrectly classified as protalus ramparts. Velocity changes in time and space measured by photogrammetrical means have been detailed for several rock glacier sites, and all provide supporting evidence for the patterns found using geodetic surveys.

2.6.4 Other measurements of rock glacier movement

Several other methods have been used to measure rock glacier movement. Tree-ring analysis has been advanced as an alternative method for providing relative data on rock glacier movement (Shroder, 1978; Giardino *et al.*, 1984). For example, movement of a rock glacier on Mount Mestas, Colorado, was investigated for a period from the 15th century onwards using a sample of almost three hundred trees. However, tree-ring analysis has failed to provide any new evidence apart from site-specific development studies, and the application of this method is limited to those few rock glaciers below the treeline that contain sufficient amounts of fine debris on the surface to support extensive tree-growth.

Detailed observations of rock glacier fronts have proved more effective in providing information on both horizontal and vertical velocity profiles. Wahrhaftig and Cox (1959), who were the first to suggest that much information could be inferred from the form of frontal slopes, observed that a front that is entirely free of vegetation indicates that the rock glacier is experiencing differential forward motion, with the upper part moving faster than the lower part. If movement was entirely by basal sliding then the entire front should remain undisturbed, whereas continuous removal of the upper part of the front and partial burial of the lower part occurs if the upper part of the rock glacier is moving forward faster than the lower part. In addition, they suggested that the sharp angle at the top of the frontal slope and the steepness of the frontal slope, both of which are characteristic of all active rock glaciers, could only be maintained in a climate in which mass-wasting predominates by constant renewal.

In order to understand how this upward increase in velocity occurs Wahrhaftig and Cox directed their attention to the height of talus aprons at the front of rock glaciers, which they suggest is highly indicative of the vertical velocity profile of rock glaciers. Figure 2.7 illustrates the concept. No talus should accumulate if movement is due to basal sliding alone, and conversely, if movement is due solely to surface flow, the whole front should be covered by coarse boulders. In nature, talus apron heights generally vary between 10 and 75% of the rock glacier height with a mean value of 45%. This indicates that rock glacier movement is due neither to surface movement nor entirely to basal sliding, but probably to some type of internal deformation possibly accompanied in some cases by basal sliding. Thus, low ratio values of apron height to total rock glacier height are indicative of steep velocity gradients at or near the base of the rock glacier, whereas high ratio values suggest steeper velocity gradients

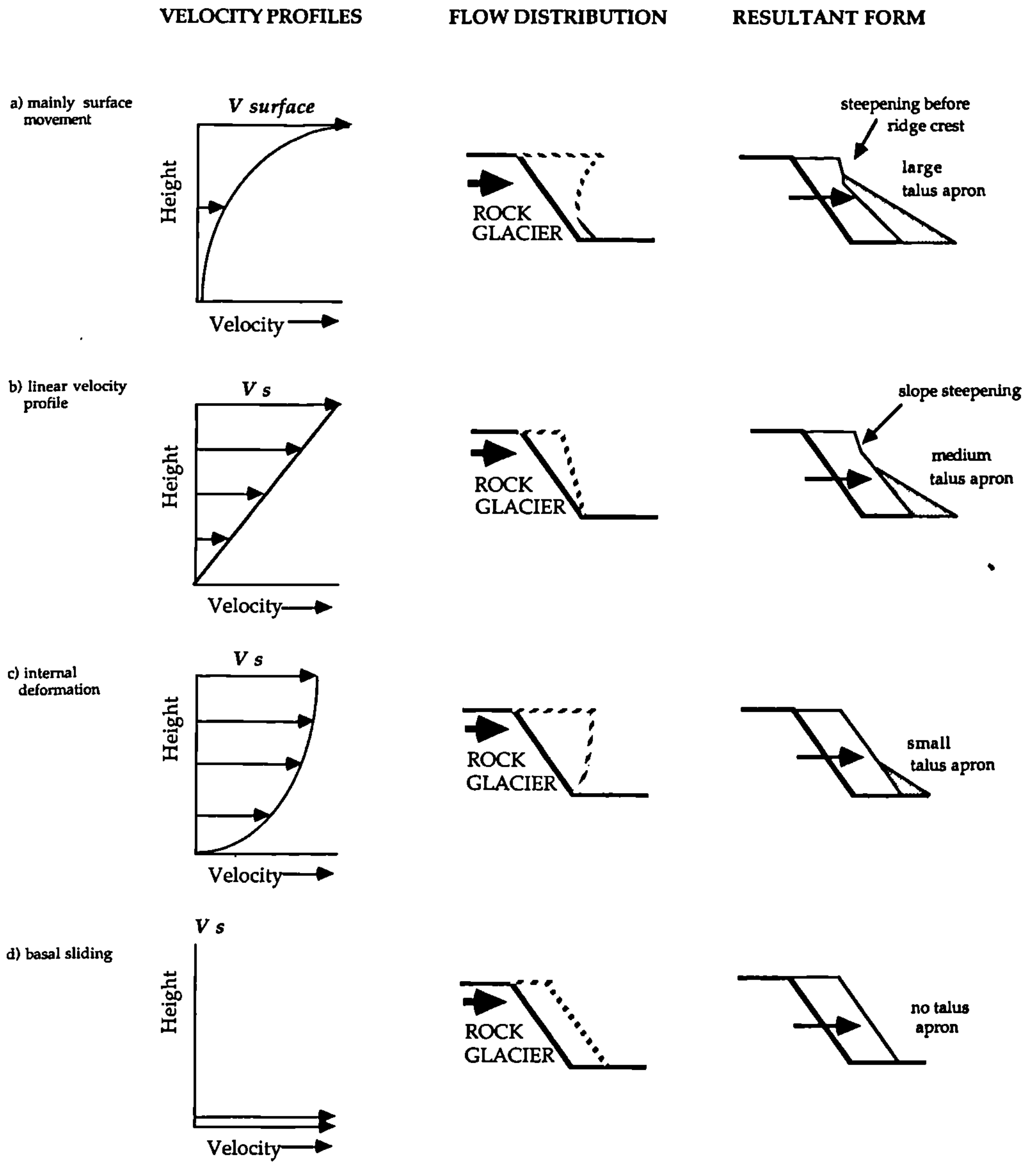


Figure 2.7 Various velocity distributions within rock glaciers (after Wahrhaftig and Cox, 1959).

nearer the surface of the rock glacier. A more even distribution of velocity changes throughout the whole thickness of the rock glacier would be reflected in ratio values approaching the mean value of approximately 45%. Clearly, therefore, the rate of rock glacier advance is on average about 0.5 times the surface velocity. These relationships are summarised in Figure 2.8. In addition, the observation that talus apron height is a relatively constant fraction of the front of actively advancing rock glaciers demonstrates that active rock glaciers re-incorporate their surface boulders by overriding the frontal talus apron (Haeberli, 1985). One further implication of this would appear to be that as the base of rock glaciers consists of coarse unfrozen talus debris, basal sliding may be enhanced if high pore-water pressure develops in this zone of confined rock debris at the base of the rock glacier. Eisbacher (1979) has suggested that in sturzstroms unfrozen sub-surface boulders act as ball-bearings and enhance downslope movement. Such a mechanism for rock glaciers seems unlikely, however, as rock glacier movement rates are significantly smaller than those of sturzstroms.

2.7 Rock glacier origin

2.7.1 Introduction

Many morphologic features of rock glaciers imply viscous or plastic flow, but because of uncertainty regarding the internal structure of these landforms, specific flow mechanisms remain obscure. The aim of this section is to review the major theories of rock glacier formation for both valley-floor and valley-wall forms. The formation of valley-wall rock

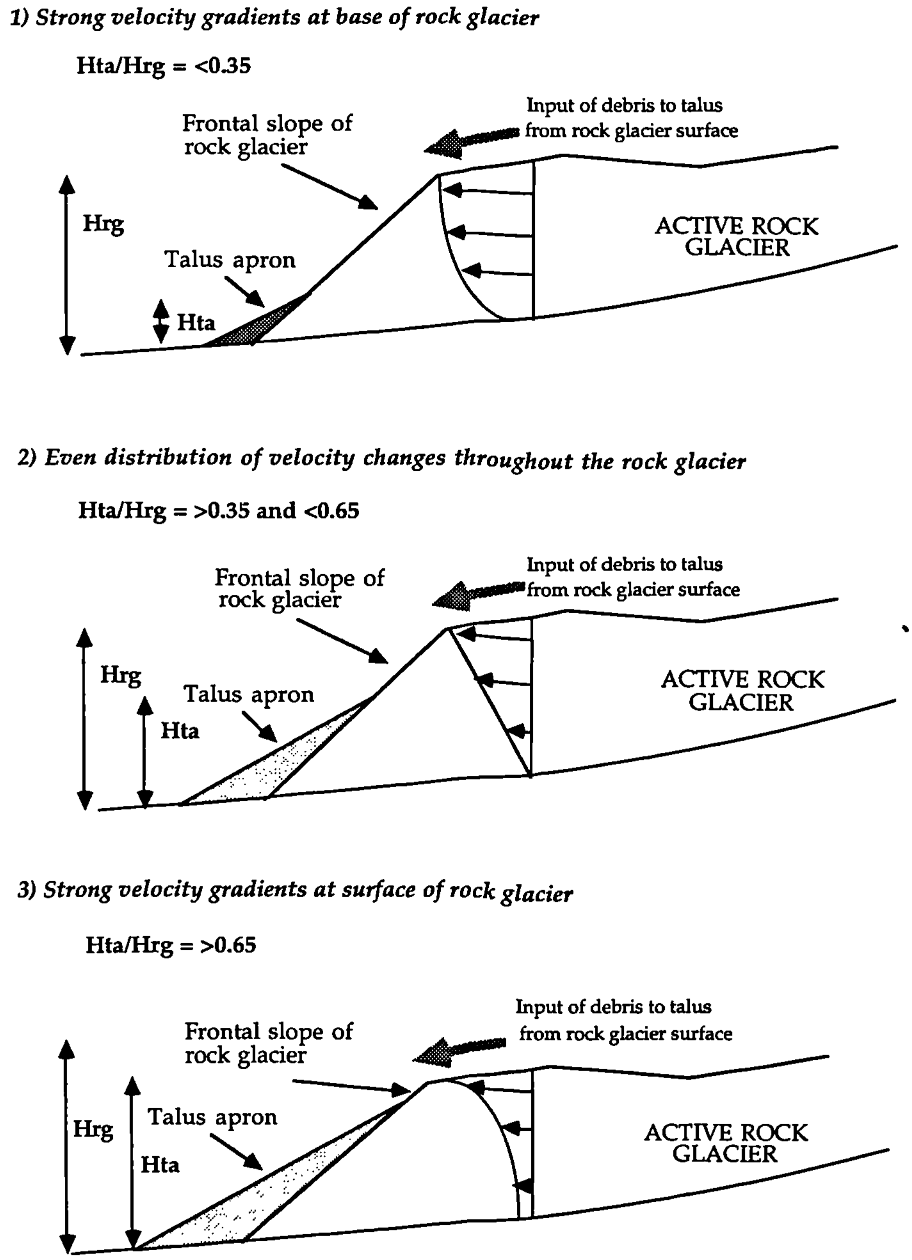


Figure 2.8 Relationship between talus apron height and height of rock glacier front.

glaciers, which is less well understood, will be discussed in greater detail in Chapter 3.

Any explanation of rock glacier flow and formation must account for the existence of at least four major characteristics:

1) surface flow-like structures such as transverse ridges;

2) rock glacier movement, although at several orders of magnitude slower than that of true ice glaciers;

3) steep fronts which tend to be more gentle on inactive features; and

4) well-defined margins particularly in the downslope sections of rock glaciers.

Theories of rock glacier formation may be grouped into three major categories. First, rock glaciers may be periglacial in origin, and form from the creep of ice-rich perennially frozen sediments; second, rock glaciers may be glacial in origin, with movement resulting from the deformation of a glacier-ice core; and third, rock glaciers may not require the presence of ice for movement to occur. Each of these will be discussed separately.

2.7.2 Periglacial origin

The first, and most widely-cited explanation (e.g. Capps, 1910; Wahrhaftig & Cox, 1959; Corte, 1976; Haeberli, 1985), is dependent on the presence of non-glacial ice, occurring either as an ice matrix or as layers or lenses of debris-free ice. Wahrhaftig and Cox (1959) suggested that non-glacial ice develops within the debris mass below the surface boulder layers of rock glaciers by refreezing of snow meltwater and groundwater. When the frozen rock-ice mixture yields to the stress caused by the debris overburden, the whole mass creeps downslope.

The matrix or non-glacial ice hypothesis, which was first suggested by Capps (1910), has been supported by many workers following the elaboration of this early concept by Wahrhaftig and Cox (e.g. Johnson, 1975; White, 1976; Haeberli, 1985). However, Whalley (1974) argued that the shear strengths of rock-ice mixtures are simply too great to allow many of the thinner rock glaciers to flow, and on the basis of limited and perhaps dubious evidence (that will be discussed in Chapter 7) he suggested that all rock glaciers must have a core of glacial ice. Wayne (1981) proposed that downslope movement could occur if active rock glaciers contained some kind of ice lenses of sufficient size to permit the lobes of rock debris to move. He envisaged that as rock glaciers and talus slopes may be partly permeable, water from snow melt and rain could flow through the lower part of the debris above the bedrock floor. If this groundwater was able to maintain water pressure against the base of a permafrost layer, segregation of ice lenses could occur, and rock glacier movement may be possible as a result.

One of the major difficulties in explaining rock glacier movement, and one that has prolonged the argument concerning the theoretical aspects of ice content, is that few studies exist on the rheology of rock-ice mixtures. Flow laws that exist for glacier ice may be used to determine the amounts of debris overburden required to initiate movement for variable thicknesses of ice, but flow laws for rock-ice mixtures have as yet to be precisely determined. Only Whalley (1974) has made a tentative attempt to determine the theoretical boundaries for rock glacier movement in which both ice and rock-ice mixtures are considered.

2.7.3 Glacial origin

The second group of hypotheses requires the presence of glacier ice. Outcalt and Benedict (1965) proposed that rock glaciers may form when debris accumulates and covers true ice glaciers. They suggested a sequence of events in which a glacier advances down-valley until it stagnates. Climatic amelioration then causes clean ice at the head of the glacier to ablate more rapidly than debris-covered ice at the glacier terminus. This leaves a tongue-like lobe of debris-covered ice preserved downvalley and a depression at the rear of the glacier. A subsequent change in climate may allow a second ice tongue to advance into the depression, override the lobe in front of it, and then stagnate on to its surface. They argued that complex rock glaciers develop if several of these ice advances occur. It seems unlikely, however, that glaciers would advance and retreat frequently enough to produce each of the many transverse ridges that are often found on rock glaciers. In addition, transverse ridges do not appear to be stacked one on top of the other as would be expected if successive ice advances had produced each ridge. Whalley (1974) agreed that a glacier ice core is necessary for movement, although he did not accept the idea of successive ice advances overriding earlier debris lobes. Instead, he proposed a generalised model of rock glacier flow for valley-floor forms that is based on a glacier's response to debris supply. He stated that rock glaciers are different from ordinary glaciers only in terms of the amount of debris that they carry on their surfaces. If true, Whalley failed to provide any explanation as to why rock glaciers move much more slowly than glaciers, although intuitively this may be because glacier-cored rock glaciers are glaciologically dead, and so movement is effectively residual with shear stresses marginally above the yield stress of glacier ice.

Whalley's glacier - rock glacier continuum is dependent on the accumulation of sufficient debris to cover the ice surface and so preserve the ice from melting over a long period of time. The origin of this debris may be either ablation or subaerial material. A subaerial origin seems more likely for the debris because of the limited potential for erosion at the heads of these glaciers in small valleys or cirques. In addition, the angular nature of the debris suggests passive (supraglacial or englacial) transport. However, in order to provide a complete terminus debris cover of rockfall debris, several restrictions to formation seem essential. Johnson (1980) suggested that geologically the area must be suitable for the rapid production of debris and that morphologically the valley must be adequately confined so that rockfall debris will cover the whole of the glacier surface. Field evidence, however, does not appear to substantiate these severe limitations as rock glaciers occur in quite diverse geologic and morphologic situations.

2.7.4 Ice-free origin

Some researchers have examined the possibility that ice need not be an integral part of rock glacier formation (Maxwell, 1979; Johnson, 1980, 1984, 1987). Similarities between characteristics of valley-wall rock glaciers and those of landslides and mudflows led Johnson (1980) to suggest that ice-free movement mechanisms might provide an alternative explanation for valley-wall rock glacier formation. He provided field evidence from mine tailing dumps in the Canadian Rockies in which slope failure and subsequent flow occurred in coarse material deposited at high slope angles, and where there was no ice content and no evidence of high water tables. Johnson (1978) stated that "if this type of mechanism could be applied to rock glaciers, the ice content which is found within the rock glacier may be of secondary origin, developing from snowmelt or rainfall after the primary flow mechanism has operated."

Maxwell (1979) recognised the potential of catastrophic mechanisms of rock glacier formation when undertaking extensive fieldwork in the Yukon, and illustrated his work by reference to the Duke valley-wall rock glacier. This feature is derived from the accumulation of talus on a valley-side structural bench, and its morphology is typical of most valley-wall rock glaciers. Talus has extended over the bench and has moved downslope to the valley floor, producing complex transverse ridges. Down the length of the feature there are a series of retrogressive slab failures in the talus material, indicating the possible importance of a failure mechanism as opposed to a flow mechanism. These slab failures suggest the possibility that a talus slope may reach a critical condition, such as oversteepening of the slope during a period of rapid accumulation, may fail and generate mass movement. However, the literature on the mechanics of talus slope accumulation suggests that such a sequence of events is rather rare (e.g. Statham, 1976). In addition, slab failure is liable to be very localised (e.g. Howarth & Bones, 1972). Maxwell's initial work suggests that if failure occurs the individual failed units would reach the base of the slope, thus offering the potential for rock glacier formation. As with glacial forms, secondary movements may occur in these non-glacial forms due to melt of inherited ice, settling, or periglacial processes, and these may be responsible for later movement.

In conclusion, three general categories are used to explain rock glacier formation. The first explanation is that rock glaciers are periglacial features, and require the development of ice-rich perennially frozen sediments for movement to occur. Second, some authors maintain that rock glaciers require the presence of glacial ice, whilst a few argue that ice need not be an integral part of rock glacier formation. Valley-floor rock glaciers appear to be strongly related to ice glaciers, whereas the origin of valley-wall forms is less well understood. More detailed discussion of valley-wall formation is provided in Chapter 3. The working definition of rock glaciers proposed in section 2.2-1 precludes the possibility that rock glaciers have an ice-free origin. In Chapter 3, however, ice-free models of formation are considered together with several glacial and non-glacial models, so that the validity of each hypothesis may be tested in subsequent chapters by a combination of field and theoretical evidence.

2.8 Summary

A review of the literature suggests that rock glaciers may be divided into two major types, namely valley-floor and valley-wall forms. Valley-floor rock glaciers appear to be predominantly glacially derived and require the deformation of glacial ice for movement to occur, though creep of ice-rich perennially frozen sediments may also contribute to movement. They are thought to be less directly dependent on climatic conditions than valley-wall forms and may form in regions where the mean annual air temperature is 0°C or even higher. Valley-wall forms are less well understood. Many researchers believe that they comprise perennially frozen sediments that slowly move downslope by creep, although several other hypotheses have been proposed, including a glacial ice-cored model of formation similar to valley-floor forms. Active valley-wall forms apparently occur above the lower boundary of discontinuous permafrost and below the equilibrium line on glaciers, in regions where the mean annual temperature is below -1° to -2°C.

Excavations and geophysical investigations suggest that below a coarse boulder layer, valley-wall rock glaciers contain frozen fine-grained sediments supersaturated with ice, in which ice content decreases from nearly 100% by volume just below the base of the active layer to around mean values of 50% a few metres below. Non-frozen sediments may exist underneath rock glacier permafrost. Valley-floor rock glaciers have been found to contain buried glacial ice and sometimes ice-rich perennially-frozen sediments.

Rock glacier velocities are much smaller than those of ice glaciers. Geodetic and photogrammetrical evidence suggest that rock glaciers may advance at relatively constant rates for many years, although pronounced variations in speed are detectable over short time-scales such as seasons. The morphology of rock glacier fronts and their associated talus aprons can be used to infer qualitative information on vertical velocity profiles. Morphological features such as transverse and longitudinal ridges and depressions frequently occur on both types of rock glacier. Transverse ridges and depressions are believed to represent areas of compressional movement whereas extensional movement may produce longitudinal ridges and depressions. Therefore, transverse ridges and depressions may be absent from incipient rock glaciers and should not be considered as fundamental morphological features.

Much progress in rock glacier research has been made but many questions remain open. In particular, much work is needed on both the theoretical and practical aspects of the rheology of rock-ice mixtures before a better understanding of rock glacier movement can be achieved. In addition, the boundary conditions for debris inputs are poorly understood and the precise origin of ice, particularly within valley-wall rock glaciers, remains to be satisfactorily established. Several aspects of surface morphology have yet to be fully explained and further potential exists for relationships to be drawn between morphology and both structure and origin, particularly by comparing active, inactive and relict forms.

Chapter 3

Models of Valley-Wall Rock Glacier Formation

3.1 Introduction

Although many hypotheses of rock glacier formation have been proposed during the past century (e.g. Howe, 1909; Wahrhaftig & Cox, 1959; Maxwell, 1979; Haeberli, 1985), most can be rejected in the light of present knowledge. For example, Cross and Howe (1905) envisaged that rock glaciers evolved by the large-scale movement of talus over snowbanks on to the floors of corries. Their hypothesis, which stipulates that after initial formation rock glacier movement is limited to a small amount of settling, is clearly false given known movement rates which in some cases are estimated to have been relatively constant for hundreds of years (Haeberli, 1985). In previous studies, emphasis has been placed on the formation of valley-floor rock glaciers which seem to be strongly related to true ice glaciers. The mechanisms of valley-wall rock glacier formation, which may be quite different from those associated with glacier debris systems, are still very poorly understood, and several hypotheses of formation have been postulated that have yet to be subject to rigorous testing. The aim of Chapter 3 is to present and discuss each of these models of valley-wall rock glacier formation. Empirical and theoretical evidence will be used to test the validity of each model of formation in subsequent chapters.

3.2 *Models of formation*

3.2.1 *Introduction*

Recent studies continue to provide new data and new hypotheses of valley-wall rock glacier formation (Chandler, 1973; White, 1976; Swett *et al.*, 1978; Johnson, 1984). One of the earliest detailed studies of these landforms was made by Liestøl (1962), who noted that valley-wall rock glaciers are a characteristic feature of several arctic regions. In 1973, Smith undertook a photogeologic study of valley-wall rock glaciers in which he described the deposits as "a glacier-like body of blocky detritus extending outward from the foot of a talus slope" (p.70). A recent morphological and sedimentological study of a series of similar talus-foot deposits was made by Dumbell (1984) in the Lyngen Peninsula, Norway. She concluded that valley-wall rock glaciers form by deformation of a glacial ice core, the process being controlled by a series of thresholds dependent on the rate of debris supply. Outcalt and Benedict (1965), however, suggested that valley-wall rock glaciers form as a result of the creep of matrix ice formed from avalanching and subsequent burial of snow by rockfall debris. The opposing views expressed by Dumbell and Outcalt and Benedict echo the earlier *ice-cored* versus *ice-cemented* (matrix ice) debate for valley-floor rock glaciers.

Seven major models of formation form the crux of the current debate on the origin of valley-wall rock glaciers. These are:

1) the glacier ice-cored model, in which burial of an ice core by rockfall debris leads to downslope deformation as a result of shear stresses set up by the accumulating rockfall overburden debris (e.g. Whalley, 1974; Dumbell, 1984);

2) the matrix ice model, in which ice-rich sediments develop within talus by the refreezing of snow and meltwater and slowly creep downslope; (e.g. Haeberli, 1985; Vick, 1987);

3) the segregated ice model, in which the deformation of massive segregated ice at depth within talus deposits causes downslope movement (e.g. Wayne, 1981);

4) the firn field model, in which perennial snowpatch ice behind a protalus rampart becomes buried as a result of subsequent rockfall (e.g. Liestøl, 1962);

5) the hydrostatic pressure model in which a sufficient build up of hydrostatic pressure causes movement of rock debris by basal sliding (e.g. Giardino, 1983; Shroder, 1987);

6) the retrogressive slope failure model, in which individual slab units stack up after limited forward movement, producing complex ridge morphology (e.g. Maxwell, 1979); and

7) the avalanche model, in which successive debris and snow-debris avalanche slides produce small stepped basal extensions at the foot of talus slopes (e.g. Johnson, 1984, 1987).

3.2.2 The glacier ice-cored model

The first model is dependent on the presence of a glacial ice core and is similar to the ice-core hypothesis of formation for valley-floor rock glaciers (e.g. Whalley, 1974; Lindner & Marks, 1985). The glacier ice-cored model is illustrated in Figure 3.1. During glacier stagnation, ablation material accumulates and talus develops supraglacially at the foot of valley walls. As deglaciation progresses, debris from the valley walls accumulates at the margins of stagnating valley glaciers and buries an ice core. Dumbell (1984) envisaged that the buried ice core will subsequently

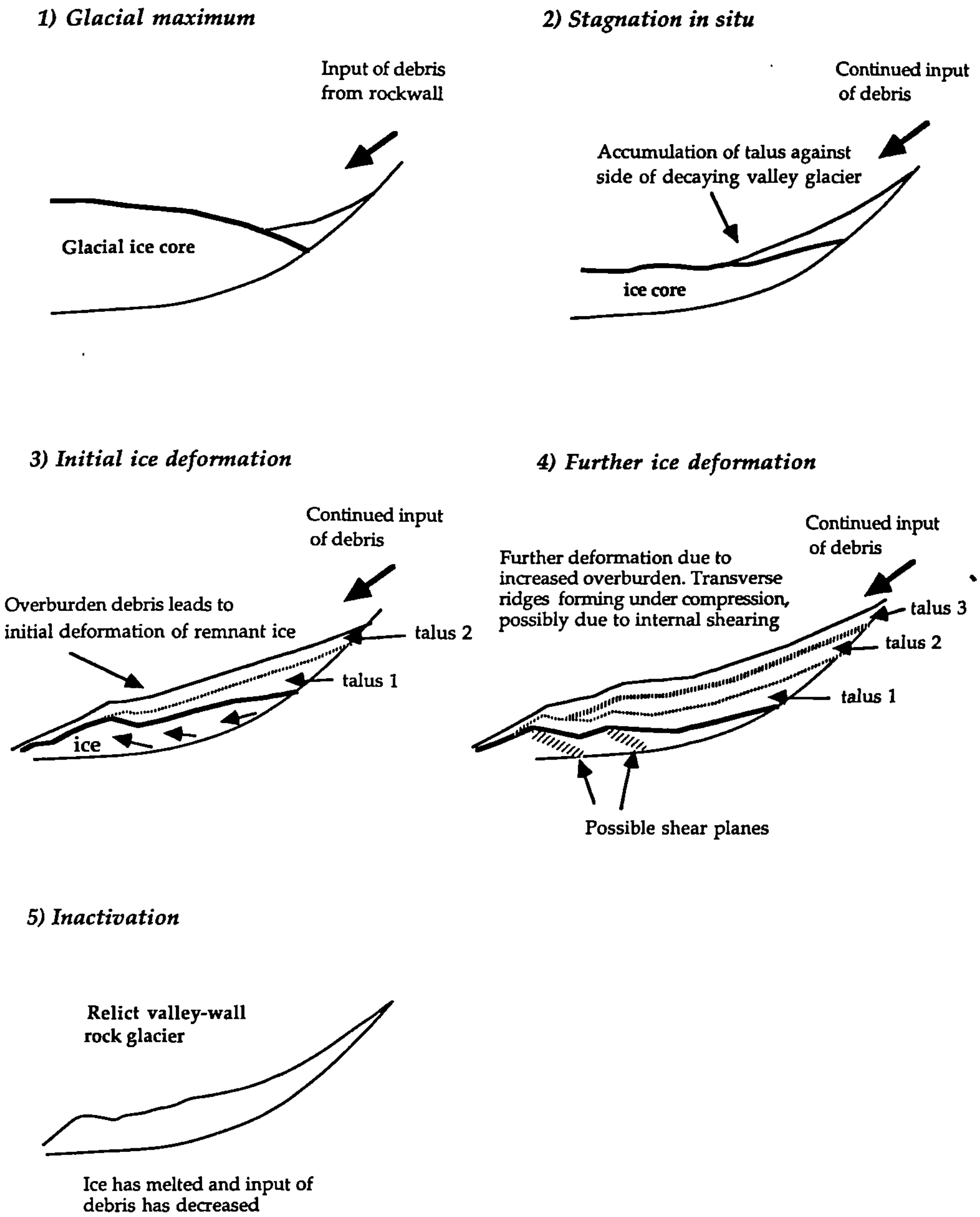


Figure 3.1 The glacier ice-cored model of valley-wall rock glacier formation.

deform as a result of the increase in shear stress caused by an increase in overburden debris, and that transverse ridges will form along internal shear planes. Continued supply of debris may activate a second shear zone, thus forming a new ridge. Further ridges will develop until a critical threshold level is reached at which either the ice core becomes too small or the input of debris insufficient for movement to occur. Reduction in debris input and ice-core size may both result from climatic amelioration.

Valley-wall rock glaciers, however, have been found in areas that have never experienced glaciation (e.g. Blagbrough & Farkas, 1968), and so the glacier ice-cored model must be rejected as a *general* model of formation. As yet, though, conclusive evidence remains to be found which demonstrates that the deglaciation model is invalid for all valley-wall rock glaciers. Pronounced morphological similarities between valley-floor and valley-wall rock glaciers tend to strengthen arguments in favour of a similar origin for at least some valley-wall and valley-floor features.

A glacier ice-core hypothesis was also presented by Johnson (1984), although he suggested that as melt of the buried ice core continued, the whole talus slope would become progressively more unstable until it finally failed. He maintained that field evidence from the Yukon proved that landslides caused by failure and flow of the whole slope, in situations similar to those outlined by Dumbell, produced a long concave basal extension to the talus slope without the formation of any transverse ridges. Given large-scale rockfall it is possible to envisage slope failure of the sort described by Johnson. However, gradual ice-core burial by discrete rockfall events would more likely result in internal deformation and slow downslope movement.

Three further difficulties with the glacial ice-core hypothesis remain. First, Johnson noted that there are often marked discrepancies in the age of valley-wall rock glaciers over short distances along one valley-wall. The deglaciation model requires that following glacier stagnation, sufficient debris buries and preserves an ice-core; if debris supply is initially limited, a rock glacier will not form. However, it is possible that if an ice-core is preserved by an initial input of debris, subsequent increases in debris may result in the formation of much younger rock glaciers. Second, he also noted that there are large morphologic variations in valley-wall rock glaciers where slope conditions such as angle, aspect, and geology are apparently uniform. Third, Liestøl (1962) noted that many valley-wall rock glaciers in Spitsbergen are found near the coast where it is improbable that stagnating glacial ice would be found beneath rockwalls. These difficulties though, only negate the ice-core hypothesis as a *general* explanation.

3.2.3 The matrix ice model

The second hypothesis, the matrix-ice model, is favoured by many workers (e.g. Wahrhaftig & Cox, 1959; Smith, 1973; White, 1976; Corte, 1978; Haeberli, 1985; Vick, 1987). In this model (Figure 3.2), snow

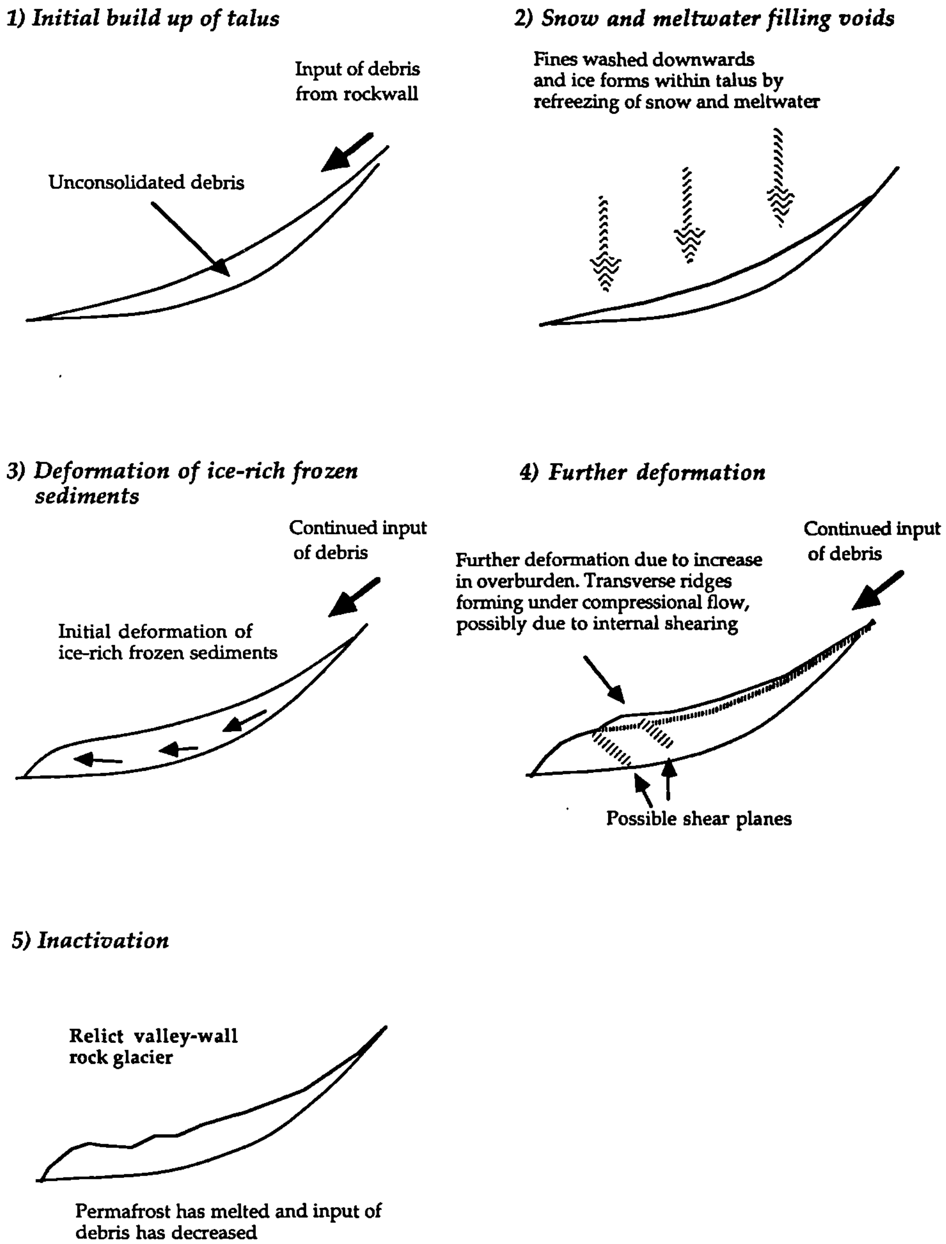


Figure 3.2 The matrix-ice model of valley-wall rock glacier formation.

meltwater accumulates and refreezes within the lower layers of talus material and ice-rich frozen sediments slowly creep downslope under the weight of accumulating overburden debris. If the shear strength of the ice-rich frozen sediments can be exceeded by shear stresses generated by sufficient overburden debris, a series of threshold values similar to those suggested for the ice-cored model could produce a series of inclined shear planes and attendant talus ridges. In the model climatic controls are crucial as the mean annual air temperature must be low enough for permafrost to develop. Moreover, excess moisture must also be present in order for sediments rich in ice to form rather than merely pore ice in which clast-to-clast contact remains (see section 2.2.3). Whalley (1974) maintained that the height of debris required to overcome the shear strength of ice-rich frozen sediments would be much greater than the thickness of rock glaciers. At present, the rheology of ice-rich frozen sediments is largely unknown and detailed calculations must be made of the overburden stresses that would be required to induce strain in various rock-ice mixtures before definitive evidence can be found to validate or disprove this general model for valley-wall rock glacier formation.

3.2.4 The segregation ice model

Support for the segregated ice model of formation has come mainly from Wayne (1981). He has shown that if ice lenses of sufficient size are able to form within talus slopes, then the internal shear strength of these deposits would decrease and movement could occur. The concept is illustrated in Figure 3.3. A continual downwards movement of the perennial freezing front, a necessity for ice segregation, may result from either climatic deterioration or from the gradual downslope movement and consequent thinning of overburden debris after initial formation. In such circumstances, very massive ice lenses can form. It is possible that lenses and layers of segregated ice may, in addition to either glacial ice or perennially-frozen sediments, explain the formation of valley-wall rock glaciers. The segregation and matrix models are both potentially *general* models of formation.

3.2.5 The firn field model

The fourth model of formation requires the presence of a perennial firn field over which rockfall debris slides and accumulates at the base forming a protalus rampart. An increase in debris from the rockwall may result in complete burial of the firn, with subsequent compaction and deformation of buried ice. A major difficulty with this hypothesis as a *general* model of formation is that it is rare for perennial firn fields to be of sufficient size to explain the formation of the largest valley-wall rock glaciers. Liestøl (1962) suggested that a repeated series of these firn fields may merge to form broad valley-wall rock glaciers, although such a situation is likely to be rare. Several authors have suggested that valley-wall rock glaciers may have evolved from protalus ramparts by the continued supply of debris (e.g. Corte, 1976; Haeberli, 1985) (Figure 3.4). It is possible also to envisage a situation in which large-scale rockfall over a firn field may incorporate both buried snowpatch ice and

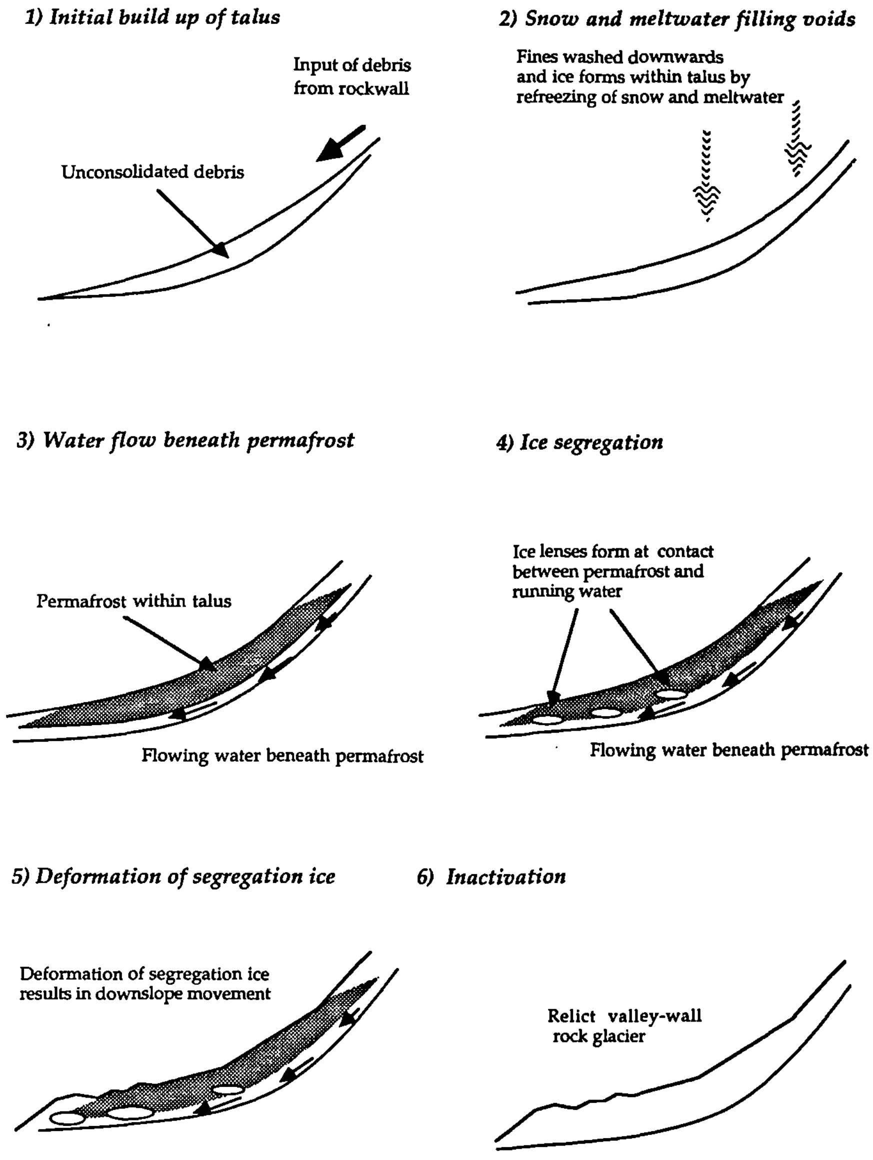


Figure 3.3 The ice segregation model of valley-wall rock glacier formation.

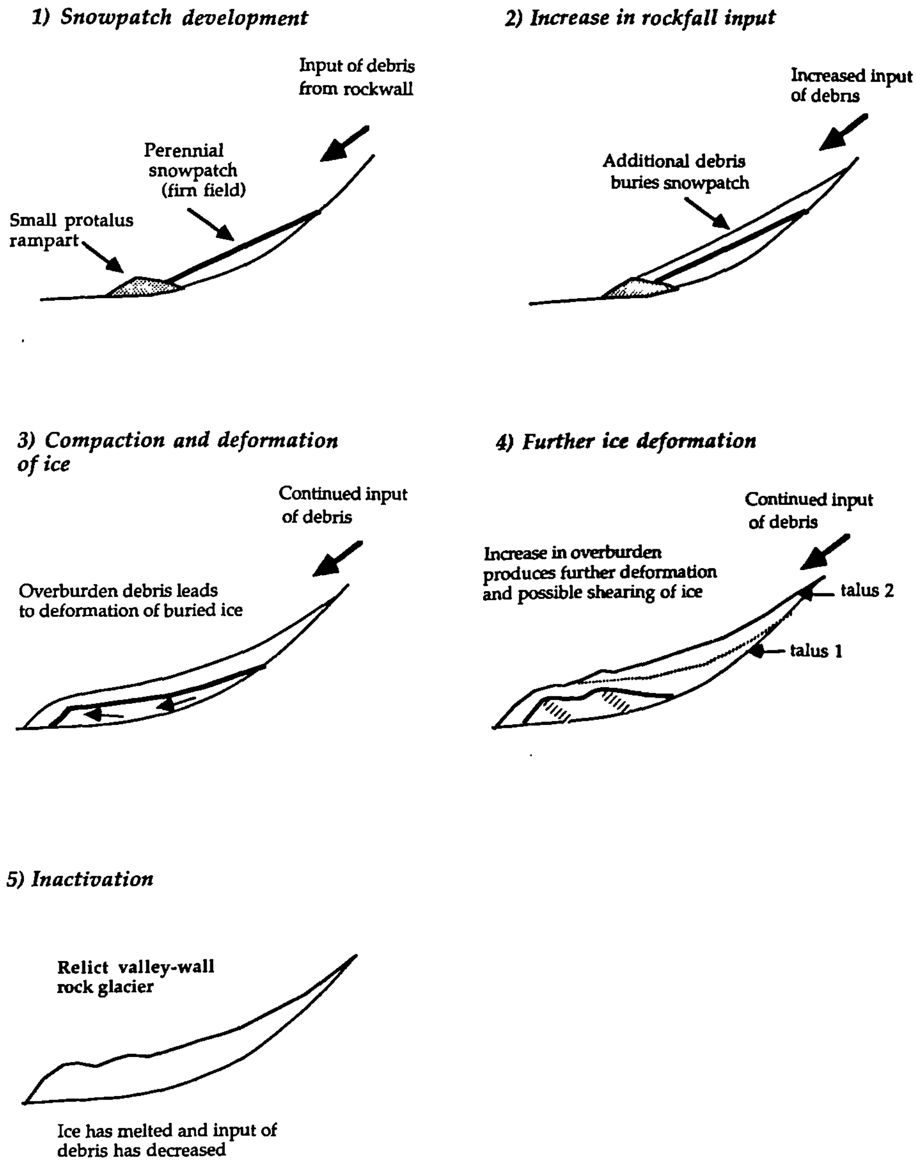


Figure 3.4 The firn field model of valley-wall rock glacier formation.

permafrost which may then promote downslope movement. Morphological similarities between protalus ramparts and incipient valley-wall rock glaciers are very pronounced. In such a situation, however, protalus ramparts will only evolve into rock glaciers if there is an increase in debris input.

3.2.6 The hydrostatic pressure model

The fifth model, which was proposed by Maxwell (1979) and supported by Giardino (1983) and Shroder (1987), is based on the flow of rock glaciers induced by hydrostatic pressures (Figure 3.5). Maxwell assumed that the presence of hydrostatic pressure conditions that he found in the Yukon can be used to infer the former occurrence of pressures sufficient to allow slope failure. Hydrostatic pressure has been known for many years as a cause of rock slide movement (e.g. Whalley *et al.*, 1983). The presence of an impermeable layer within the rock glacier, which could be produced by the refreezing of summer snowmelt, is thought to allow high pressure to build up, eventually causing slope failure. Hydrostatic pressure may, together with the creep of ice-rich frozen sediments or the deformation of glacial ice, contribute towards the formation of valley-wall rock glaciers.

3.2.7 The retrogressive slope failure model

The remaining two models of formation do not require the presence of either glacial or non-glacial ice. Both models are based on the idea that

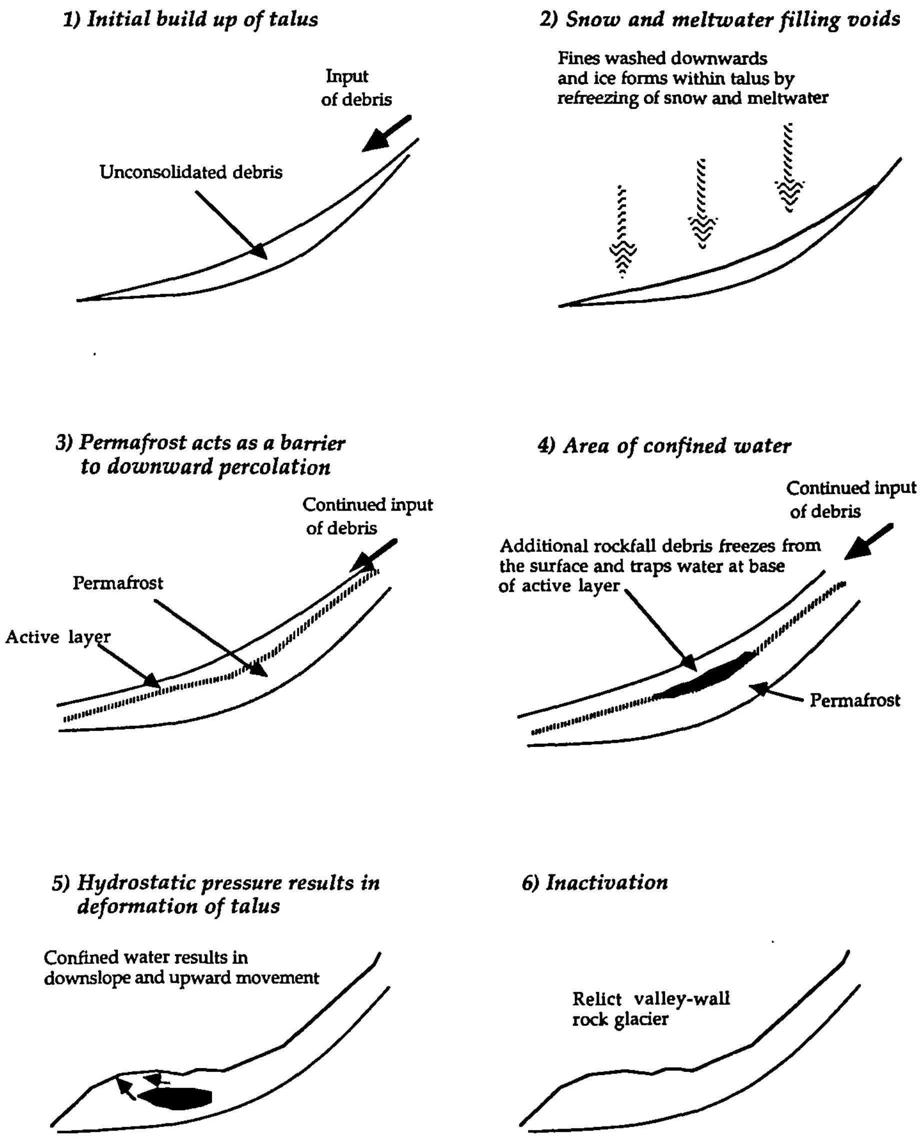


Figure 3.5 The hydrostatic pressure model of valley-wall rock glacier formation.

valley-wall rock glaciers may in fact be the result of one or more catastrophic events, followed by a period of much more subdued change in form (e.g. Maxwell, 1979; Johnson, 1984, 1987). Johnson and Maxwell recognised that it is hard to envisage any type of catastrophic flow that would produce complex morphologies with such limited basal extensions while also allowing for continual movement after initial catastrophic formation, in agreement with known movement rates. Five factors, however, led them to believe that catastrophic processes should be introduced into the debate on valley-wall rock glacier formation. These factors are:

1) there are large variations in rock glacier morphology where slope conditions are uniform;

2) there is an absence of large scale periglacial features in areas of rock glacier occurrence, which implies that less severe periglacial conditions may have been in operation than some mechanisms require;

3) there is an absence of downslope deflections of valley-wall rock glaciers;

4) several periods of flow are apparent on some rock glaciers where there is no evidence of reactivation of the older unit by the younger; and

5) there are discrepancies in the age of some rock glaciers over short distances along one valley wall.

All of these factors, however, can effectively be explained without invoking ice-free catastrophic formation. For example, variations in morphology may be the result of temporal or volumetric variations in the input of debris, and secondly, large-scale periglacial features may be absent as such features are not always associated with areas of discontinuous permafrost. Downslope deflections have been found on some features (e.g. Dumbell, 1984), and their absence elsewhere may simply reflect low valley-floor gradients. Several periods of flow may be explained by invoking the matrix-ice or segregation-ice models as can discrepancies in the age of features along one valley-wall. Despite their poor reasoning for invoking a catastrophic origin, the validity of the two models requires testing.

Maxwell's first catastrophic ice-free model of rock glacier formation was developed using a rock mechanics approach. He suggested that valley-wall complexes can develop by retrogressive slab failures, in which failure on one part of the slope initiates failure of an adjacent unit upslope. He developed three possible movement mechanisms in which the slabs either:

1) stack up after limited forward advance;

2) override the preceding slab; or

3) join up with the preceding slab.

Figure 3.6 illustrates the first of these concepts. Basal extension of these slides would be limited because the first slab, which has low momentum, would act as a barrier to successive slab movement. Evaluation and testing of this model centres on the problem of establishing whether large cracks could develop in what is essentially unconsolidated debris. Even if talus deposits were frozen, it is improbable that tension cracks sufficient for large-scale slab failure could form.

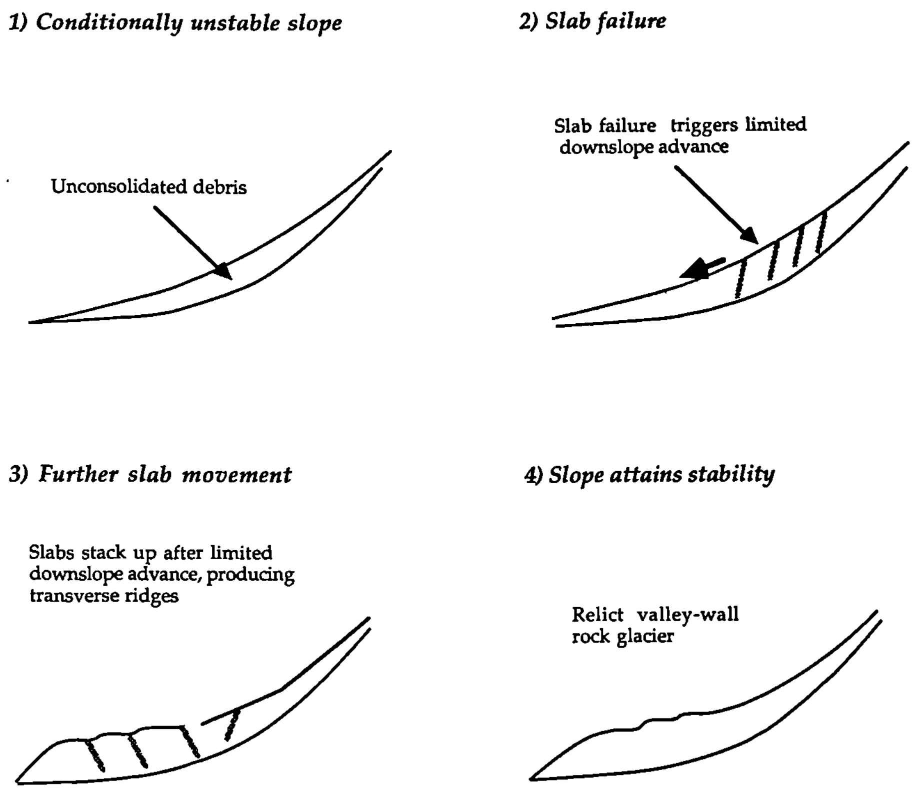


Figure 3.6 The retrogressive slope failure model of valley-wall rock glacier formation.

3.2.8 The avalanche model

The avalanche model, also proposed by Maxwell (1979), was used to explain some of the more simple forms of valley-wall rock glaciers in which complex flow ridges are absent. Small snow avalanches comprising snow and rock debris are believed to accumulate at the base of talus slopes, producing lobate-shaped features with depressions between the lobes and talus slope indicating limited travel distance away from the slope (Figure 3.7). Internal deformation is absent and so the structure has a rather simple form, which suggests that avalanching cannot be proposed as a *general* model of formation for all valley-wall forms. In addition, valley-wall rock glaciers formed by avalanching would have to be associated with avalanche talus slopes, and many are not. A further test for this model would be to attempt to differentiate morphologically between simple valley-wall rock glaciers and small-scale features formed by avalanching. Effectively, this model simply describes the formation of avalanche boulder tongues (c.f. Luckman, 1978). Rockfall avalanches may, however, play an important role in the formation of valley-wall rock glaciers by providing the loading necessary to initiate ice deformation at the base of talus slopes. Vick (1981) stated that the role of landsliding "was not to bring the body of the rock glacier material to its present position, but rather to supply the debris necessary for the initiation of major creep movement".

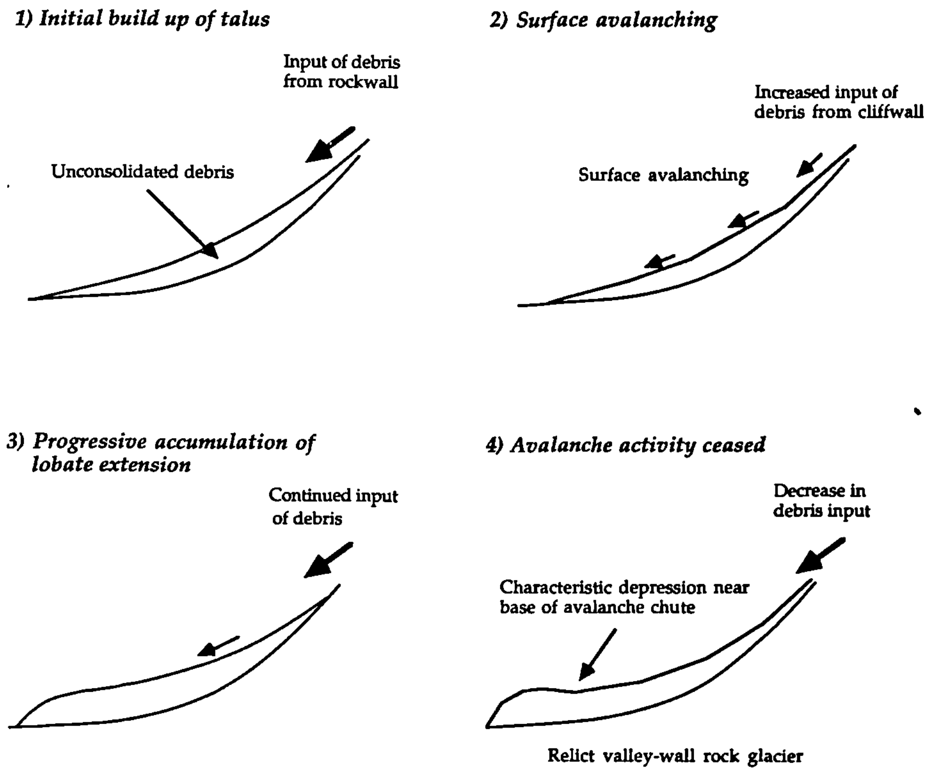


Figure 3.7 The avalanche model of valley-wall rock glacier formation.

3.3 Summary

In conclusion, many aspects of valley-wall rock glacier formation remain open to debate. Review of the literature on this topic indicates that a wide range of processes and mechanisms of movement have been proposed to explain these landforms. Proposed movement characteristics range from steady state to catastrophic, and current theories suggest that the features themselves may or may not be related to periglacial or glacial activity. The generality and feasibility of the proposed models of formation outlined above will be tested both empirically and theoretically in the remainder of this thesis.

Chapter 4

Study Areas and Rock Glacier Distribution

4.1 Introduction

The aims of this chapter are twofold. First, to provide a general introduction to the four major study areas in Switzerland, Norway, and Scotland, and second, to examine rock glacier distribution within these study areas. Following an introductory rationale and a short discussion on the field methodology used to determine rock glacier activity, each study area is considered in terms of climate, geology, topography and, where relevant, glacial history. Distribution of the study rock glaciers is then examined by focusing on five controlling factors, namely aspect, altitude, lithology, bedrock source wall characteristics and topographic location. The regional data obtained from these analyses is used to test the validity of previously-proposed distributional boundary conditions, and finally, a model is suggested that outlines the most important local factors controlling the distribution of valley-wall rock glaciers.

4.2 Study areas

4.2.1 Rationale

Variations in valley-wall rock glacier morphology, distribution, and dimensions have given rise to seven major hypotheses of formation, each of which was outlined in the previous chapter. Some authors believe that a single mechanism explains the formation of all valley-wall rock glaciers, whilst others suggest that the landforms are polygenetic. Although individual hypotheses may be valid for specific rock glaciers, investigations of single features may not reveal more general explanations applicable to all valley-wall rock glaciers. As discussed in Chapter 1, the fundamental aim of this thesis is to test the validity of each model of valley-wall rock glacier formation using a range of theoretical and empirical evidence. In order to do this, a large sample of geographically and morphologically diverse active, inactive and relict valley-wall rock glaciers was selected for field study. Several known relict rock glaciers in Scotland, together with some previously unstudied Scottish valley-wall rock glaciers, were chosen to represent the relict stage of development. Two areas in Switzerland, Valais and Graubünden, were selected for a field study of active rock glaciers on the basis of an initial aerial reconnaissance. In addition, the field sample includes inactive and relict valley-wall rock glaciers from the Lyngen Peninsula in northern Norway.

Thus, the rationale in choosing spatially diverse field areas in Switzerland, Norway, and Scotland lies in the need to obtain sufficient field evidence from rock glaciers at different stages of development to test each model of formation. The selection of these diverse field areas also presents an opportunity to study the influence of a range of environmental and geomorphic factors on the distribution of valley-wall rock glaciers. In addition, by choosing a large sample of study sites, variations in the characteristics of valley-wall rock glaciers within each activity stage may be interpreted.

Using a combination of morphological, sedimentological, locational and theoretical evidence each model of valley-wall rock glacier formation may be evaluated. For example, field measurements of rock glacier depth may be compared with theoretical depth values derived from the shear strength characteristics of various ice and rock-ice mixtures. Such comparisons of field and theoretical data may allow some hypotheses of valley-wall rock glacier formation to be rejected. In addition, by including examples of each activity stage in the analysis, morphological differences between active, inactive, and relict valley-wall rock glaciers may be studied. A detailed evaluation of this developmental continuum may provide information that would otherwise be unavailable in a limited analysis of only active valley-wall rock glaciers. For example, volumetric comparisons between active and relict rock glaciers of similar size may provide information on the amount of ice present in active features. Clearly, direct evidence of rock glacier processes must be obtained from active rock glaciers. Previous research, however, has tended to ignore the potential of relict valley-wall rock glaciers as a source of complementary information.

4.2.2 *Rock glacier activity*

Distinctions between active, inactive, and relict rock glaciers may be difficult to detect. Activity is best determined by long-term geodetic or photogrammetric investigations of rock glacier movement. However, the large number of study sites and the relatively short-term nature of this research precluded such an approach. Instead, rock glacier activity was determined principally from morphological evidence. Thus an active rock glacier, which by definition is currently subject to downslope movement attributable to the process (or processes) of formation, commonly exhibits a very steep frontal slope that is free of vegetation and is composed of much finer debris than the remainder of the rock glacier surface. A sharp angle usually marks the boundary between the upper surface of the rock glacier and the top of the frontal slope. The presence of internal ice in active rock glaciers may result in ponding of surface water. By contrast, on relict rock glaciers frontal slopes are generally less steep and covered by boulders that may be lichen-covered. Vegetation may also be present on all or part of the rock glacier surface. Although the margins of relict rock glaciers generally remain well-defined, their slopes are more gentle and sharp breaks of slope are often absent. In addition, as relict forms do not contain internal ice, surface water does not accumulate.

Morphological evidence, therefore, is usually sufficient to distinguish active rock glaciers from relict features. Inactive rock glaciers, on the other hand, are less easy to identify solely on the basis of morphological evidence. They no longer exhibit downslope movement, though melt of internal ice may cause localised secondary movement. Frontal slopes of inactive rock glaciers often retain a relatively steep gradient until internal ice melts, and if periglacial conditions persist, prolonged snowcover may inhibit the growth of lichens and vegetation. In addition, surface ponds may occur on inactive rock glaciers as internal ice impedes surface drainage. Morphologically, therefore, inactive and active rock glaciers may be very similar. However, when a rock glacier becomes inactive, new material is no longer added to the frontal slope as differential forward movement has ceased. Consequently, the frontal slope loses its 'fresh' appearance. Moreover, if rock glacier deactivation results from a reduction in debris supply rather than from climatic amelioration, periglacial mass-wasting may significantly reduce the 'sharpness' of inactive features. Rock glacier inactivity may also be assumed if geomorphic events such as rockfall avalanches and debris flows, which modify rock glacier morphology, themselves remain unaltered by subsequent rock glacier movement. Thus, for each of the study rock glaciers, activity was determined from morphological and, where possible, sedimentological evidence. Activity status is shown in Tables 4.1 to 4.3 below, which also summarise the characteristics and classification of each rock glacier site in each of the four main study areas discussed in the following sections.

4.2.3 Study areas in Switzerland

Two major study areas are located in Switzerland (Figure 4.1). The first, in Valais, lies between 46°2' and 46°8'N latitude and 7°25' and 7°46'E longitude and contains several rock glaciers near Arolla, Fèrpecle, Zinal, and Gornergrat. The second study area in Graubünden includes rock glaciers at Flüela-Wisshorn, Schwarzhorn, Radüner, Chilbiritzenspitz,

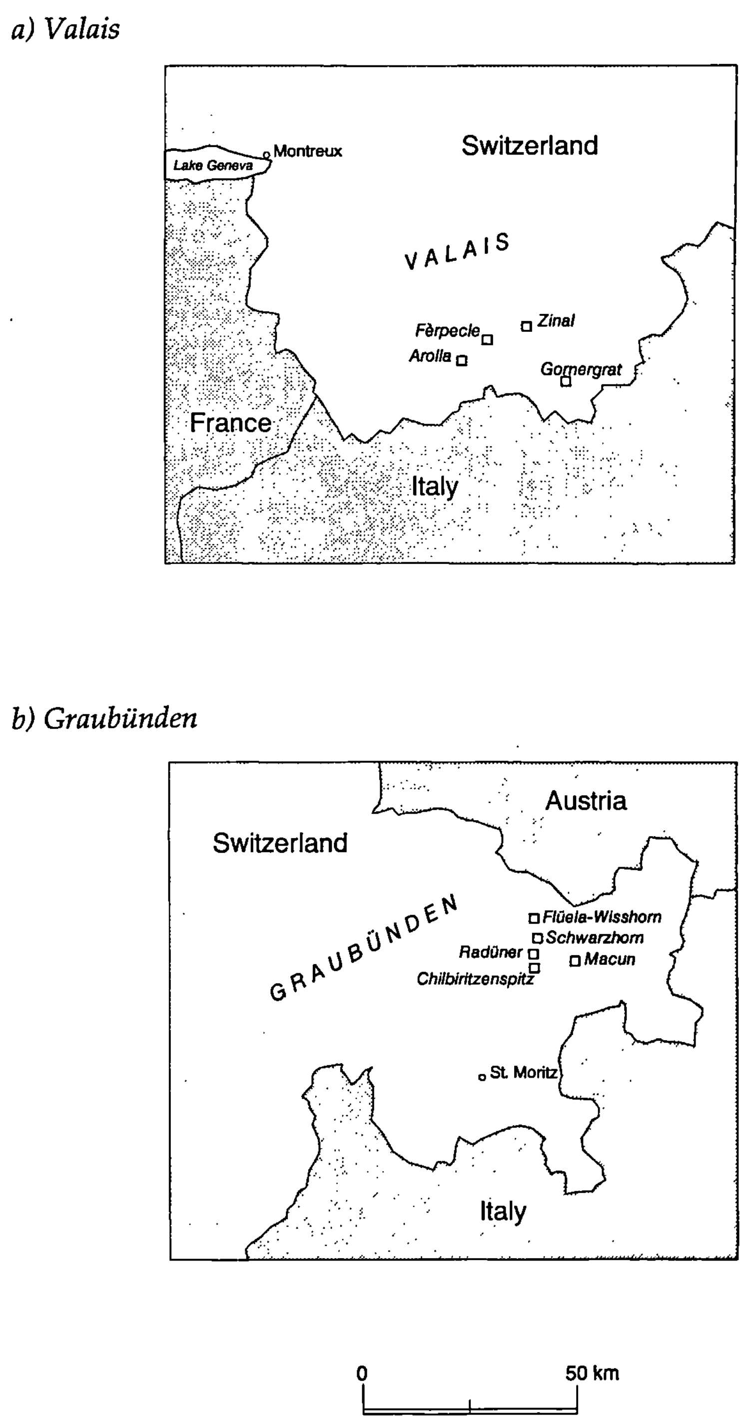


Figure 4.1 Location map of rock glaciers studied in Valais and Graubünden, Switzerland.

STUDY SITES	SITE LABEL	ALTITUDE (metres)	ASPECT (°N)	LITHOLOGY	CLASSIFICATION	ROCK GLACIER ACTIVITY	GRID REFERENCE Swiss survey
Arolla 1	A1	2600	45	Chloritic schist	Valley-wall rock glacier	Inactive	6005 0959
Arolla 2	A2	2650	95	Chloritic schist	Valley-wall rock glacier	Inactive	6004 0962
Arolla 3	A3	3180	140	Chloritic schist	Valley-wall rock glacier	Active	6005 0977
Zinal 1	Z1	2560	10	Chloritic schist	Valley-wall rock glacier (complex)	Inactive/active	6165 1074
Zinal 2	Z2	2600	355	Chloritic schist	Valley-wall rock glacier (complex)	Inactive	6169 1075
Gornergrat	GO	3000	350	Schist	Valley-wall rock glacier	Inactive/active	6268 0928
Flüela-Wisshorn 1	FW1	2620	275	Gneiss	Valley-floor rock glacier (unconfined)	Inactive	7925 1814
Flüela-Wisshorn 2	FW2	2580	255	Gneiss	Valley-floor rock glacier (unconfined)	Inactive	7921 1817
Flüela-Wisshorn 3	FW3	2660	180	Gneiss	Valley-wall rock glacier	Inactive	7925 1818
Schwarzhorn	SC	2720	90	Hornblende schist	Valley-wall rock glacier (complex)	Inactive	7918 1790
Macun 1	M1	2760	100	Schist	Valley-wall rock glacier	Inactive/active	8048 1787
Macun 3	M3	2720	350	Schist	Valley-wall rock glacier	Inactive	8062 1779
Chilbiritzenspitz	CZ	2500	345	Mica schist	Valley-wall rock glacier (complex)	Inactive	7921 1762
Radüner	RA	2720	200	Mica schist	Valley-wall rock glacier	Inactive/active	7919 1774
Macun 4	M4	2600	350	Schist	Valley-floor rock glacier (confined)	Active	8062 1787
Macun 5	M5	2760	285	Schist	Valley-floor rock glacier (confined)	Active	8060 1777
Fèrpecle	FP	2400	230	Schist	Valley-floor rock glacier (confined)	Inactive	6082 1042
Zinal 3	Z3	2600	310	Chloritic schist	Valley-floor rock glacier (confined)	Active	6176 1077

Table 4.1 Characteristics and classification of study rock glacier sites in Switzerland.

and Macun, and lies between 46°40' and 46°48'N latitude and between 9°56' and 10°12'E longitude.

Both study areas lie within the Alpine chain, which trends approximately east-west across southern Switzerland never exceeding a total width of 100km. In Valais, several mountains attain altitudes in excess of 4000m in a region of typical alpine glaciated topography. The study area here comprises north-south trending valleys that increase in altitude towards the south where they terminate in glaciated plateaux. Topographically, Graubünden is less rugged and the highest peak, Schwarzhorn, attains an altitude of 3147m, which corresponds to the elevation of the highest valley-wall rock glacier in Valais. A similar pattern of north-south trending valleys is broken by the valley of the River Inn which flows eastwards through part of the study area. Small corrie glaciers remain in a few of the high valleys, although all have been retreating since the last glacial maximum in the middle of the 19th century.

Geologically, the Alps represent one of the most complex regions in the world. The Valais region, which is composed mainly of crystalline rocks, is characterised by numerous thrusts and nappes. A detailed geological map of the area has yet to be published, but the lithology of each rock glacier site is given in Table 4.1. Generally, the geology is dominated by gneisses, schists and amphibolites, all of which usually produce massive blocks on weathering. Gabbro and serpentinite representing a Mesozoic ophiolite are also found in a narrow band trending from Arolla in the southwest to Zinal in the northeast. Graubünden, which has experienced strong tectonism, is characterised by strongly metamorphosed rocks such as gneisses, mica schists and amphibolites (Rutten, 1969).

The study areas are located within the inner Alpine climatic zone, which exhibits marked climatic differences between the north and south slope climatic provinces (Wallén, 1977). The dual sheltering effects of mountain ranges to the north and south produce a more continental and drier climate in the interior valleys of Valais and Graubünden. Annual precipitation values as low as 530mm have been recorded in Valais. Mean annual precipitation in both study areas is less than *ca.* 700mm, although with increasing elevation, precipitation totals may increase up to *ca.* 2500mm (Wallén, 1977). The north and south slope climatic provinces are characterised by mean annual precipitation values of *ca.* 1200mm and *ca.* 1800mm respectively, with extreme precipitation totals in excess of 4000mm in the Bernese Alps. In Valais, the scanty precipitation is fairly uniformly distributed over the year, whereas in Graubünden there is a summer precipitation maximum and a winter or spring minimum (Barry, 1981). In both study areas, precipitation falls mainly as snow between November and April.

In Valais and Graubünden, the equilibrium line on glaciers lies above 3000m, although equilibrium line altitudes in Valais are approximately 50m higher on average than in Graubünden (Barsch, 1978). In both study areas, the timberline is about 2250m, and the lower limits of permafrost are approximately 3600m and 2550m, respectively (Barsch, 1978). In Valais, winter brings very strong temperature inversions to the snow-covered valleys so that temperature minima of less than -30°C have been observed (Wallén, 1977).

4.2.4 Study areas in Norway

The Lyngen Peninsula in northern Norway lies between 69°15' and 70°00'N and between 19°30' and 20°00'E (Figure 4.2). The central part of the peninsula is dominated by north-south trending gabbroic mountains that have formed from a layered ophiolite. The central core of gabbro is bounded on either side by several metasedimentary groups of lower relief. Figure 4.3 illustrates the main geological divisions in Lyngen, and the lithology of each rock glacier site is outlined in Table 4.2. All but one of the rock glaciers studied are located on Lyngen gabbro, which weathers to produce massive blocks and extensive talus slopes.

The highest peak in the study area is Jiekkevarre (1833m), and several mountain summits above 1400m support small cold-based ice-caps (Gordon *et al.*, 1986). Rock glacier elevations range from 100m at Trollvatnet to 900m at Gjerdalvdalen. At 900m, mean annual air temperature in Lyngen (based on climatic records for Tromsø and Skibotn and assuming a saturated adiabatic lapse rate of 0.6°C for 100m) is *ca.* -3.0°C, with January and July averages of *ca.* -12°C and *ca.* 7.8°C, respectively (Ballantyne, 1987). Present-day mean annual temperature at 100m, which is the elevation of the lowest rock glaciers at Trollvatnet

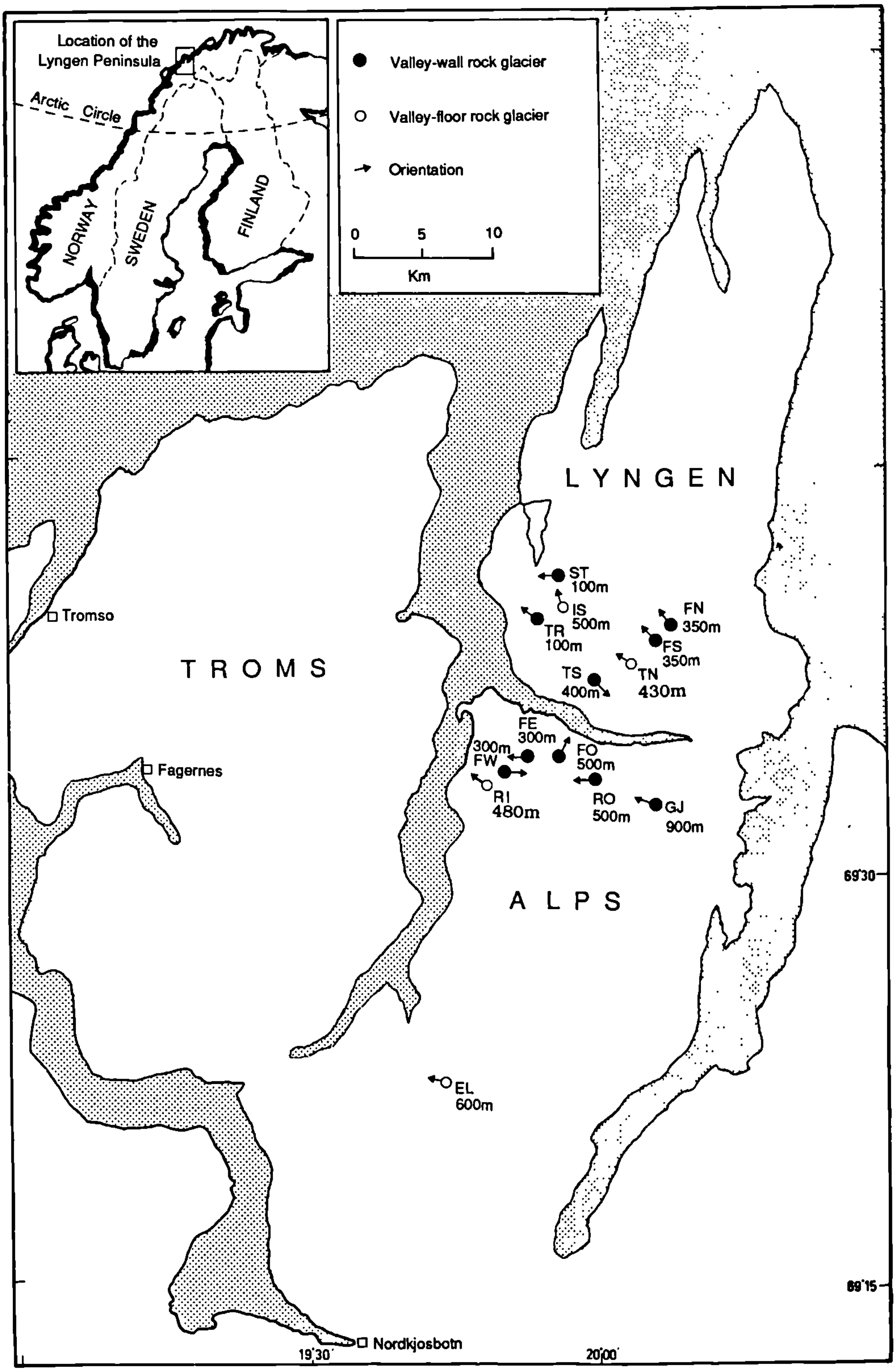


Figure 4.2 Location map of rock glaciers studied on the Lyngen Peninsula, northern Norway, showing rock glacier orientation and altitude. Site labels are identified in Table 4.2.

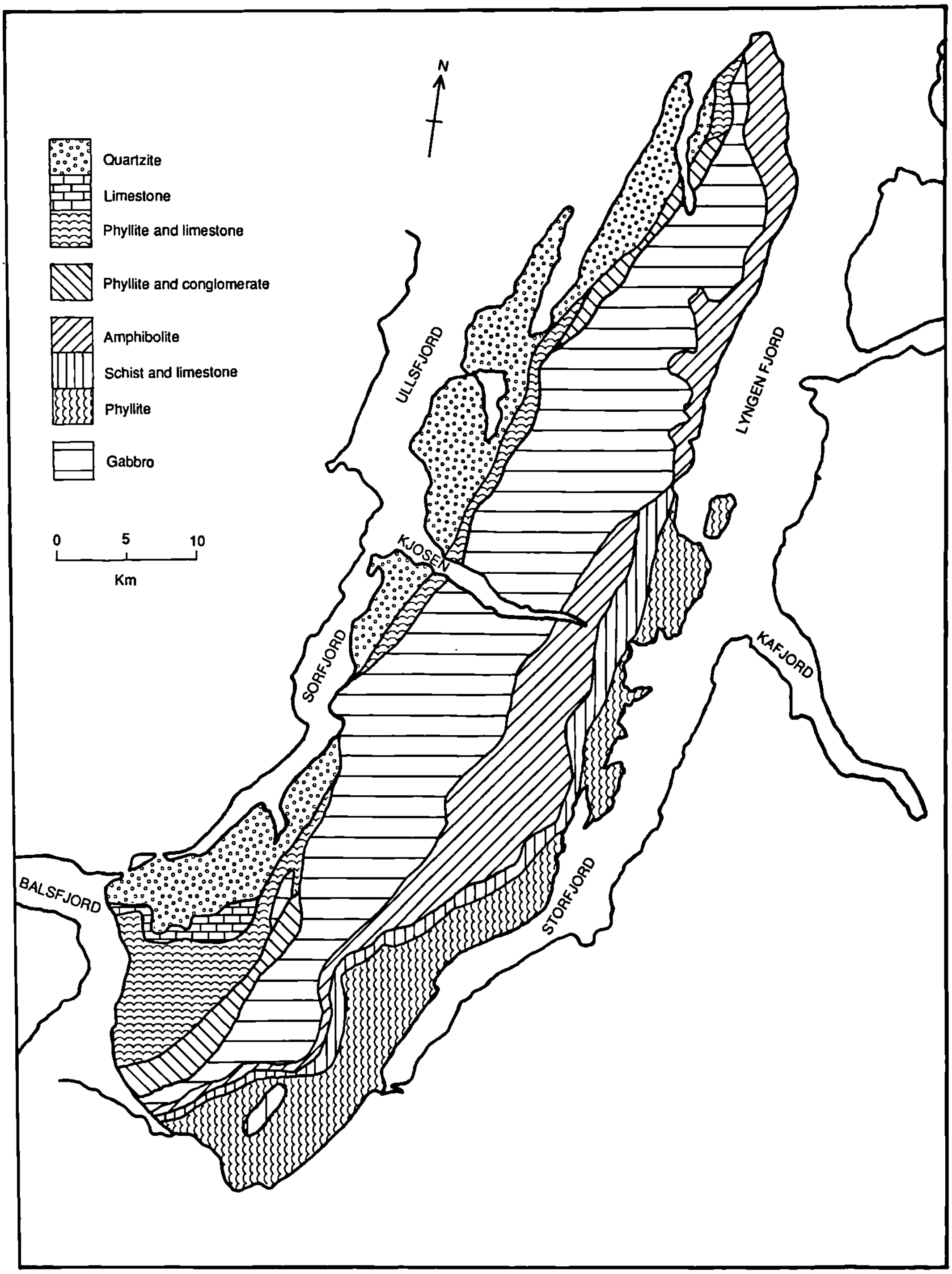


Figure 4.3 Geological map of the Lyngen Peninsula(after Randall, 1971).

STUDY SITES	SITE LABEL	ALTITUDE (metres)	ASPECT (°N)	LITHOLOGY	CLASSIFICATION	ROCK GLACIER ACTIVITY	GRID REFERENCE Norwegian survey
Fornesdalen West	FW	300	80	Gabbro	Valley-wall rock glacier	Inactive	34WDC 562167
Fornesdalen East	FE	300	270	Gabbro	Valley-wall rock glacier (complex)	Inactive	34WDC 567175
Rodbergdalen	RO	500	270	Gabbro	Valley-wall rock glacier	Inactive/relict	34WDC 625175
Trollvatnet	TR	100	305	Gabbro	Valley-wall rock glacier	Inactive/relict	34WDC 579283
Tverrelvdalen	TS	400	135	Gabbro	Valley-wall rock glacier	Inactive	34WDC 617237
Tytebaerdalen	TN	430	300	Gabbro	Valley-floor rock glacier (Unconfined)	Relict	34WDC 643246
Gjerdelvdalen	GJ	900	290	Amphibolite	Valley-wall rock glacier	Inactive	34WDC 655155
Fastdalen North	FN	350	315	Gabbro	Valley-wall rock glacier	Inactive/relict	34WDC 661260
Fastdalen South	FS	350	315	Gabbro	Valley-wall rock glacier	Inactive/relict	34WDC 658258
Stortinddalen	ST	100	275	Gabbro	Valley-wall rock glacier (complex)	Inactive	34WDC 594301
Isskardtindane	IS	500	345	Gabbro	Valley-floor rock glacier (Valley-confined)	Inactive	34WDC 592283
Forholtskardet	FO	500	65	Gabbro	Valley-wall rock glacier	Inactive	34WDC 595183
Ellendalen	EL	600	270	Gabbro	Valley-floor rock glacier (Valley-confined)	Active	34WDB 516967
Rissmålsskardet	RI	480	300	Gabbro	Valley-floor rock glacier (Valley-confined)	Inactive	34WDC 530147

Table 4.2 Characteristics and classification of study rock glacier sites in Norway.

and Stortinddalen, is *ca.* 1.8°C. Mean annual sea-level precipitation declines eastward across the Lyngen Peninsula from *ca.* 850mm in the west to *ca.* 600mm in the east. Winter precipitation falls mainly as snow.

In northern and western Norway the Late Weichselian ice sheet *ca.* 20,000-18,000 yr BP. extended westwards of the present coastline and terminated on the continental shelf (Corner, 1980). Glacial readvances during the Lateglacial period produced two extensive moraine systems in Lyngen; namely the Skarpnes moraines (12,500-12,000 B.P.) and the Tromsø-Lyngen moraines (11,680-10,150 B.P.) (Corner, 1980). Ballantyne (pers. comm.) suggests that on the Lyngen Peninsula three Preboreal readvances and up to five Neoglacial advances or readvances have occurred, forming a complex and compact sequence of Holocene moraines. The moraine systems provide an indication of rock glacier age. For example, in Fornesdalen, a north-south trending valley south of Kjosen Fjord, deglaciation began after the Tromsø-Lyngen readvance *ca.* 10,500 yr BP. A complex valley-wall rock glacier on the eastern side of Fornesdalen overlaps the Tromsø-Lyngen moraine deposited at the maximum of this readvance, suggesting that rock glacier development began after 10,500 yr B.P., probably during the Preboreal. On the opposite side of the same valley, Fornesdalen West valley-wall rock glacier, overlaps a Preboreal moraine (cf. Plates 5.1 and 5.2 below) that is thought to date from between 9400 yr. B.P. and 9900 yr. B.P. (Ballantyne, pers. comm). This indicates that Fornesdalen West rock glacier is younger than 9900 yr. B.P.

4.2.5 Study areas in Scotland

Rock glaciers in Scotland have previously been reported in the Cairngorms (Sissons, 1979; Chattopadhyay, 1984), in Wester Ross (Sissons, 1975), and on the Isle of Jura (Dawson, 1977). These sites together with previously unreported valley-wall rock glaciers in Scotland were chosen as representative examples of relict rock glaciers. The location of the study rock glaciers in Scotland is presented in Figure 4.4. Their diverse geographic distribution provides a wide range of aspects, elevations, and lithologies (Table 4.3).

Four relict valley-wall rock glaciers are located in the northwestern part of the Cairngorm Mountains at elevations in excess of 550m. The Cairngorms are underlain by a large pluton that was intruded into the surrounding Moinian metasediments, and are composed mainly of medium to coarse grained granite. The plateau forms the most extensive area of land above 1000m anywhere in Britain and is heavily dissected by glacial troughs and corries; many of these corries were at least partially occupied by glaciers during the Loch Lomond Stadial of *ca.* 11,000-10,000 yr BP. (Sissons, 1979). However, each of the rock glaciers lies outside the Loch Lomond Stadial glacier limit; and as the Loch Lomond Stadial represents the only period of permafrost conditions known to have affected upland Scotland following ice-sheet deglaciation (Ballantyne, 1984), it is virtually certain that these rock glaciers developed at this time.

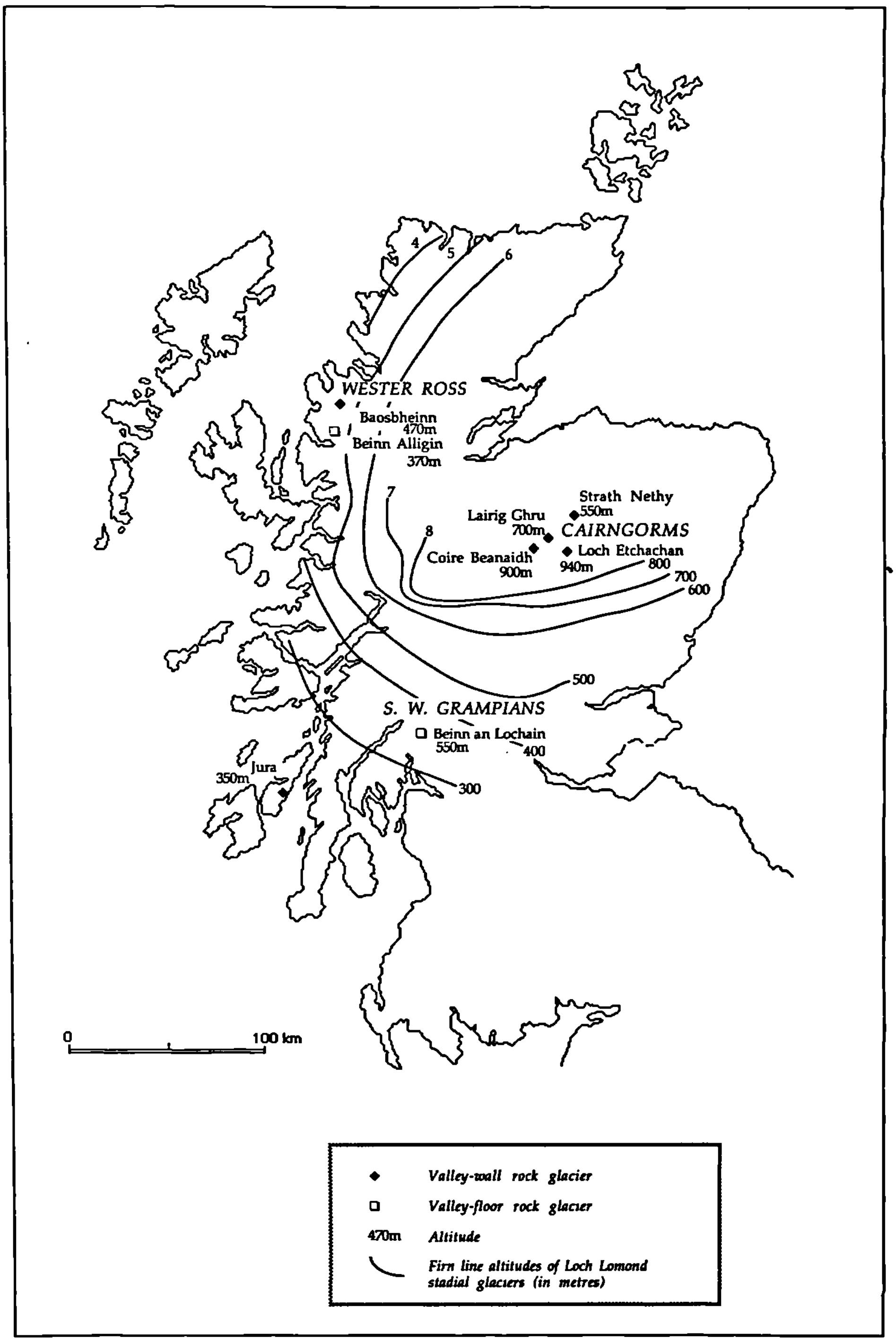


Figure 4.4 Location map of relict rock glaciers in Scotland showing altitude in metres of the frontal slopes of the rock glaciers and reconstructed regional firn line altitudes of Loch Lomond Stadial glaciers at their maximal extent (firn line altitude from Sissons, 1980).

STUDY SITES	SITE LABEL	ALTITUDE (metres)	ASPECT (°N)	LITHOLOGY	CLASSIFICATION	ROCK GLACIER ACTIVITY	GRID REFERENCE
Strath Nethy	SN	550	90	Granite	Valley-wall rock glacier (complex)	Relict	NJ 024065
Loch Etchachan	LE	940	90	Granite	Valley-wall rock glacier	Relict	NJ 007008
Lairig Ghru	LG	700	265	Granite	Valley-wall rock glacier	Relict	NH 965033
Coire Beanaidh	CB	900	40	Granite	Valley-wall rock glacier	Relict	NH 954012
Beinn an Lochain	BL	550	320	Schist	Valley-floor rock glacier	Relict	NN 226084
Baosbheinn	BB	470	320	Sandstone	Valley-wall rock glacier	Relict	NH 854677
Beinn Alligin	BA	270	130	Sandstone	Valley-floor rock glacier	Relict	NG 865605
Jura	JU	350	80	Quartzite	Valley-wall rock glacier	Relict	NG 521749

Table 4.3 Characteristics and classification of study rock glacier sites in Scotland.

The very small size of the glaciers that formed in the northwest Cairngorms during the Loch Lomond Stadial indicates very limited snowfall at this time in this area, as temperatures were significantly lower than at present. According to Sissons (1980), the existence of permafrost down to sea-level during the stadial, which he believes is indicated by the presence of fossil frost wedges, indicates a mean annual air temperature at sea-level of less than -1°C. Based on this figure of -1°C, he estimated a mean January temperature of no higher than -14°C at 1000m for the Cairngorms during the stadial. Assuming a saturated adiabatic lapse rate of 0.6°C/100m, mean January air temperatures at the rock glacier sites, which range in elevation from 550m to 940m, would have been between *ca.* -11.3°C and *ca.* -13.6°C respectively. However, some doubt exists as to the presence of these stadial frost wedges (cf. Ballantyne, 1980), and so the mean January air temperature figures quoted above may be unrealistic. In addition, available data on present-day ice wedge formation suggest that a mean annual air temperature of -5°C, and not -1°C as believed by Sissons, provides an upper limit for wedge formation (cf. Ballantyne, 1984). Thus, if frost wedges formed during the stadial at sea-level, mean January air temperatures at the study rock glaciers would have been much lower than suggested above. Annual precipitation values of *ca.* 500-600mm have been suggested for the stadial at an altitude of 1000m in the northwest Cairngorms (Sissons, 1980).

Of the remaining rock glaciers studied in Scotland, two are located in Wester Ross, one at Beinn an Lochain in the southwestern Grampians, and one in the Paps of Jura. Each site, apart from the Beinn Alligin rock glacier, is located in an area that was glacier-free during the Loch Lomond Stadial. Recent investigations of the Beinn Alligin 'rock glacier' suggest that the feature may not be a true rock glacier. The entire deposit, which is over 1.2km long and 400m wide, lies within the readvance limits proposed by Sissons (1975, 1977) and was interpreted by him as a valley-floor rock glacier that formed by the reactivation of a thin remnant of decaying glacier ice under the weight of rockslide debris near the end of the Loch Lomond Stadial. Such an explanation seems plausible, although the very poorly defined frontal margin suggests that the feature did not move downslope as a true rock glacier. Instead, the lack of debris near the downslope margins of the feature, and the possible existence of a more extensive remnant of glacier ice than previously thought (cf. Ballantyne, 1987b), suggests that the feature may more accurately be described as a supraglacially-transported rockslide or avalanche deposit.

Lithologically, the Wester Ross rock glaciers are composed of Torridon Sandstone, Beinn an Lochain is principally schistose, and the Jura rock glacier is formed of quartzite debris. Mean annual stadial temperatures at each of these sites would have been *ca.* 1° or 2°C higher than in the Cairngorms, and precipitation, particularly in the south-west Grampians, would have been much greater: Sissons (1980) suggested a maximum annual precipitation value of *ca.* 3000-4000mm for the south-west Grampians.

The lack of relict valley-floor rock glaciers in Scotland (apart from those at Beinn Alligin and Beinn an Lochain) suggests that glacier retreat in response to rapidly warming conditions may have been too rapid for valley-floor rock glacier formation. Evidence from insect assemblages indicates very rapid climatic amelioration at the end of the stadial, over a timescale of 500 years or less (cf. Bishop & Coope, 1977).

4.2.6 Summary

Four main field areas in Switzerland, Norway, and Scotland have been selected so that a variety of active, inactive, and relict valley-wall rock glaciers could be studied. Examples of active and inactive rock glaciers were chosen from two presently glacierised regions in the Swiss Alps, whilst several inactive and relict valley-wall rock glaciers were selected from the Lyngen Peninsula in Arctic Norway. A number of Scottish relict rock glaciers comprise the remainder of the study sample. Their diverse geographic distribution provides a wide range of lithologies, altitudes, and aspects. Factors controlling rock glacier distribution within these study areas are discussed in the following section.

4.3 Factors influencing rock glacier distribution

4.3.1 Introduction

Previous research on the distribution of rock glaciers indicates that a number of environmental and geomorphic factors partly or wholly determines areas favourable for rock glacier development (cf. section 2.3 above). As with ice glaciers, temperature and precipitation are the most important climatic constraints controlling rock glacier distribution However, several authors have observed that only very small temperature differences may exist between areas supporting active rock glaciers and neighbouring areas with only inactive rock glaciers, which suggests that other variables also operate to control activity and distribution (e.g. White, 1979; Morris, 1981). Here, the mapped distribution of rock glaciers in the study areas is examined in terms of five factors which previously have been found to exert a controlling influence on distribution. The five factors are aspect, altitude, lithology, bedrock source wall characteristics, and topographic location.

4.3.2 Aspect

Rock glacier aspects within the study areas in Switzerland, Norway, and Scotland are summarised in Figure 4.5. Pronounced asymmetry is clearly evident as rock glaciers are generally favoured by northerly aspects. Similar preferred orientations in the northern hemisphere have been recognised by many other researchers (cf. section 2.3 above). The asymmetry apparent in Figure 4.5 can be explained in terms of reduced

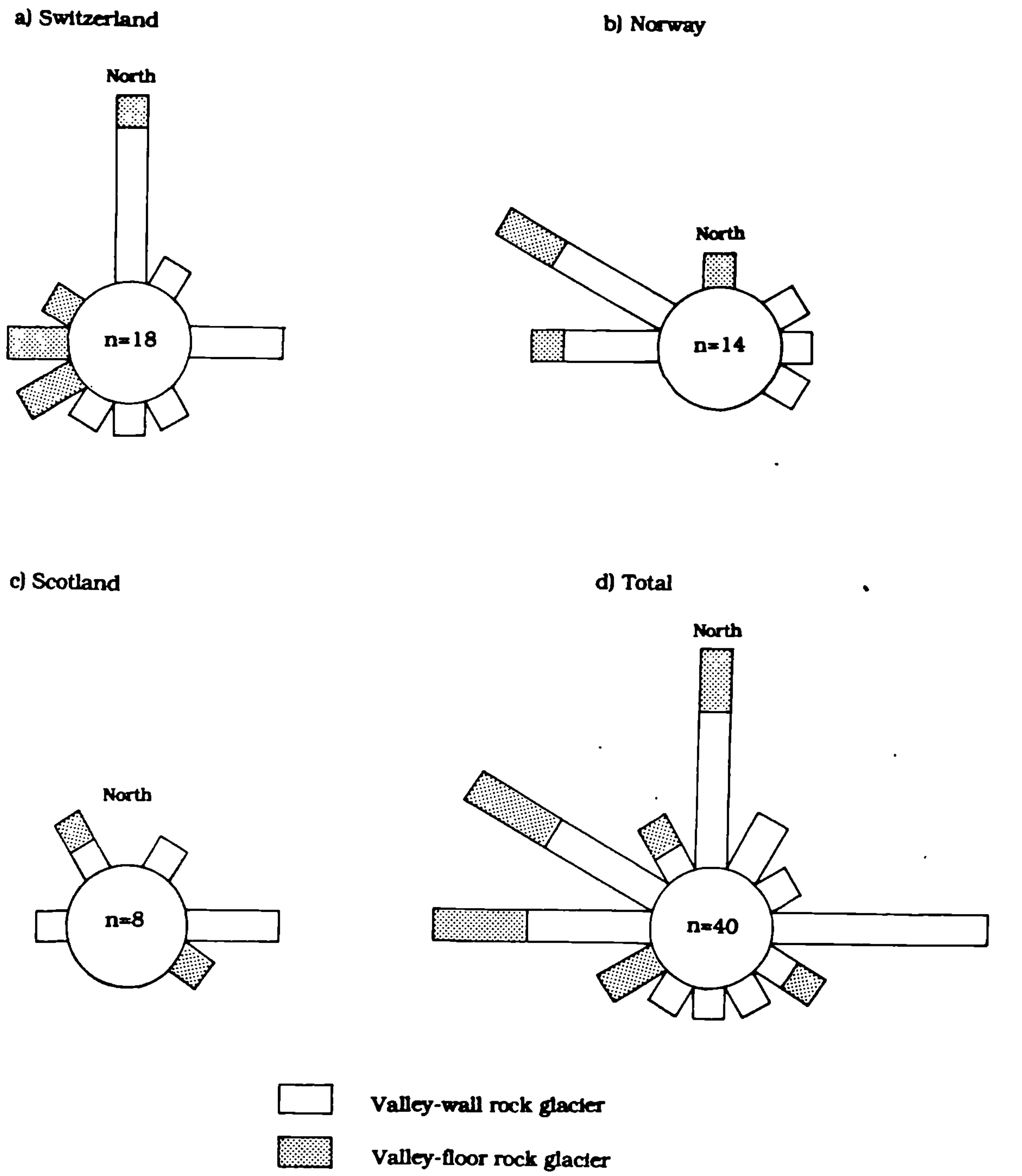


Figure 4.5 Orientation of valley-wall and valley-floor rock glaciers studied in a) Switzerland, b) Norway, c) Scotland and d) all study areas.

insolation, particularly as those rock glaciers in the sample that have southerly aspects experience reduced insolation due to topographic shading. Clearly, if insolation is reduced, mean annual temperatures will be lower and permafrost can develop. Moreover, it has been noted in several studies of regional rock glacier distribution (e.g. Barsch, 1978; Ellis & Calkin, 1979) that unless topographic shading reduces received levels of insolation, south-facing rock glaciers generally occur at higher elevations than those facing north. Barsch (1978), for example, found that in the Swiss Alps the lower limit of active rock glaciers is 200m higher on south-facing slopes than on north-facing ones. Temperature gradients between north-facing and south-facing slopes and the limited availability of sites at higher altitudes provide two additional reasons for few rock glaciers experiencing southerly aspects.

Increased snow drifting may also partly explain the lack of southerly orientations indicated in Figure 4.5. Bouët (1972) noted that in the Swiss Alps snow-cover duration at 1500m is on average 160 days on northern slopes, but 30 days less on southern slopes. In addition, in each of the study areas, northeasterly sites receive the greatest amounts of windblown snow. The small number of rock glaciers with northeasterly aspects (cf. Figure 4.5) and the higher proportion of ice glaciers with such aspects (as indicated on regional survey maps), reflects the importance of increased snowcover for glacier development. At northwesterly sites, where snowcover is greater than at southerly sites but not as prolonged as for northeast-facing sites, rock glaciers are more numerous, which suggests that such sites favour the development of ice glaciers rather than rock glaciers.

On the Lyngen Peninsula a north-northwest orientation predominates (Figure 4.5b). Only one rock glacier exhibits a southerly aspect; at this site the 965m peak of Tytebærtind lies to the south of the rock glacier, (Tverrelvdalen), thus shading the rock glacier site and reducing received insolation. The term 'orientation-independent' is generally applied to sites such as Tverrelvdalen to explain the distribution of rock glaciers with more southerly aspects.

Rock glacier aspects within the study areas in Switzerland are shown in Figure 4.5a. The strongest component is to the north, although a significant number of rock glaciers face either east or west, with surprisingly few rock glaciers facing north-west or north-east. A strong easterly component is also found in the mapped pattern of rock glacier distribution in Scotland (Figure 4.5c). These east and west facing rock glaciers are due to the predominance of north-south trending valleys particularly in Valais, and in the Cairngorms, but also on the Lyngen Peninsula and in Graubünden. In Fornesdalen in Lyngen, for example, valley-wall rock glaciers have developed on opposite sides of the same north-south trending valley.

The orientation of several valley-floor rock glaciers is also presented in Figure 4.5. Once again, a north-northwest orientation appears to be most favourable for rock glacier development. However, the south-facing sites do not appear to experience topographic shading, suggesting that valley-floor rock glaciers, which are believed to be predominantly glacially derived, have a wider climatic tolerance than valley-wall rock glaciers.

In sum (Figure 4.5d), north and northwesterly aspects are most common and southerly aspects least common in rock glacier distribution due to differences in received insolation. Preferred east and west aspects in some study areas are largely the result of large-scale structural control, and increased amounts of windblown snow appear to favour the development of ice glaciers rather than rock glaciers at locations with northeasterly aspects.

4.3.3 Altitude

Rock glacier site altitudes vary from 100m in Norway to over 3000m in Switzerland (Tables 4.1 to 4.3). Due to differences in latitude, it is not valid to consider altitudinal differentiation between active, inactive, and relict rock glaciers for the study sample as a whole. In Lyngen, for example, an active rock glacier in Ellendalen occurs at an elevation of only 600m, whereas in the Swiss study areas the lower limit of active rock glaciers is *ca.* 2500m, which corresponds to the lower limit of discontinuous permafrost (Barsch, 1978). Elsewhere, in the Arctic, active rock glaciers occur at or near sea-level (e.g. Humlum, 1984). Surprisingly, even within individual field areas, active and inactive rock glaciers have been found to occur in overlapping altitudinal zones. Moreover, it was noted that in Switzerland many valley-wall rock glaciers located above the minimal altitude for activity are inactive, which suggests that rock glacier activity must be determined by non-climatic as well as climatic factors. Altitudinal differentiation between valley-wall and valley-floor rock glaciers was not possible because of the small size of the sample of valley-floor rock glaciers.

In Scotland a distinct regional altitudinal pattern emerged. Rock glacier altitudes were found to increase in elevation from 270m in the southwest (Jura) to 940m in the northeast (Cairngorms). This general altitudinal trend corresponds fairly closely with that of the regional trend of reconstructed equilibrium firn line altitudes of Loch Lomond Stadial glaciers mapped by Sissons (1980; Figure 4.4). Sissons interpreted this trend to be largely the result of decreasing precipitation from a maximum in the southwest Grampians to a minimum in the northern Cairngorms. Ballantyne and Kirkbride (1986) noted that the altitudinal distribution of Lateglacial protalus ramparts in Scotland was remarkably consistent with the reconstructed regional firn line altitudes and suggested that this altitudinal trend could provide a realistic indication of former patterns of regional snowfall. As mentioned above, where snowfall is high, the development of ice glaciers is favoured more than the development of valley-wall rock glaciers or protalus ramparts. Thus it would appear that both valley-wall rock glaciers and protalus ramparts were not able to develop in areas or at altitudes where snow accumulation was relatively high. Clearly, the above argument is constrained by the assumption that both snow accumulation and ablation must have decreased from the southwest to the northeast (cf. Ballantyne & Kirkbride, 1986).

In the study areas in Norway and in Switzerland distinct regional altitudinal patterns are absent, although the lack of a trend in the Norwegian study area may be due, at least partly, to its much smaller geographical extent compared with the Scottish study area. On the Lyngen Peninsula, the range in rock glacier elevations of 800m is very slightly greater than in Switzerland and Scotland, and, despite a present-day decrease in precipitation from west to east across the peninsula, no related increase in rock glacier elevations was found. Instead, considerable local variation in rock glacier altitudes occur that apparently reflect age, activity and local topographic control. The active valley-floor rock glacier in Ellendalen, which lies at an elevation of 600m lies both above and below the inactive and relict valley-wall rock glaciers on the peninsula. As mean annual air temperature at the Ellendalen site is *ca.* -1.2°C, with January and July averages of *ca.* -10.2°C and *ca.* +9.6°C, respectively (Ballantyne, 1987), temperature and precipitation values prevalent at the time of greatest rock glacier development on the peninsula may not have been markedly different from present-day climatic conditions. For example, in order for mean annual air temperature to be less than -1°C to -2°C (i.e. the critical temperature above which valley-wall rock glaciers do not form) at the lowest rock glacier sites at 100m, a drop in temperature of only 3°C would be required from present-day conditions.

In Switzerland the altitudinal range in rock glacier elevations is very slightly less than in Norway. The lowest inactive valley-floor rock glacier at Férpecle (2400m) is 780m below the highest active valley-wall rock glacier at Arolla (3180m). Both valley-wall and valley-floor rock glaciers were found to occur both above and below present-day glacier equilibrium firn line altitudes. Active valley-wall rock glaciers only occur above the lower boundary of discontinuous permafrost, although, as mentioned above, several rock glaciers above this lower boundary were found to be inactive.

4.3.4 Lithology

The lithological characteristics of each study site, which are summarised in Tables 4.1 to 4.3, indicate that many rock types contribute to the formation of rock glaciers, with massive rocks such as gneiss, granite, and gabbro being the most favourable lithologies. Sandstones, amphibolites and sericite schists are also important rock types, the last-mentioned particularly in the Swiss Alps.

In Lyngen, an interesting and marked lithological pattern emerged. The main geological feature of the peninsula is a central core of gabbro which is bounded on both sides by several metasedimentary groups including quartzites, limestones, and amphibolites (Randall, 1971; Figure 4.3). Of the fourteen rock glaciers in the study area, thirteen have source areas on the gabbro. The exception is the small Gjerddelvdalen rock glacier which is located on the amphibolites of the Kjosen Formation. In part, though, this formation resembles sheared gabbro and chemical evidence indicates an igneous parentage (Randall, 1971). Such a marked pattern suggests that rock glacier distribution is strongly influenced by lithology. However, of the other rock types in Lyngen, several, such as limestone and schist, also produce massive rocks on weathering and Evin (1987) has noted the suitability of such lithologies for rock glacier development in the Swiss Alps. It is suggested, therefore, that topography is of equal, if not greater, importance in determining rock glacier distribution. Valley-wall rock glaciers form at the base of talus slopes generally beneath high cliff-walls. The greatest topographic relief on the peninsula is provided by the gabbro and by the amphibolites of the Kjosen Formation. The other lithologies, which do not support rock glaciers, are not as resistant as gabbro and their relief is much lower. Thus it would appear that the most favourable lithologies for rock glacier development are those that produce massive blocks on weathering, although rock glaciers have only developed where relief is sufficient.

A marked lithological pattern is absent from the study areas in Scotland and Switzerland. In Scotland, granites predominate in the Cairngorms whilst rock glaciers comprised of massive sandstones and quartzites are found elsewhere in the study area. In Switzerland, rock types such as gneiss and schist are most common at the rock glacier sites.

4.3.5 Bedrock source-wall characteristics

Several authors have noted that the distribution of rock glaciers may be at least partially controlled by joint planes and faulting in massively bedded rocks. (e.g. Morris, 1981; Evin, 1985, 1987). In the Cairngorms, three types of jointing have developed in the granitic exposures: a set of vertical joints that are emphasised by glacial erosion and occur on the precipices of the corrie headwalls, a type of laminar pseudo-bedding that occurs as horizontal sheeting in the surface layers of the rock, and a series of dilation joints that have developed parallel to rock faces (Sugden, 1968; Whittow, 1977). Rock slope failure in the Cairngorms appears to be in the form of translational slab slides that are related to the failure of sheeting along these dilation joints parallel to the rock-face. The scars resulting from translational slab slides are evident in the bedrock source wall above Strath Nethy rock glacier in the Cairngorms (Plate 4.1). Five of the eight Scottish valley-wall rock glaciers, namely Strath Nethy, Beinn Alligin, Baosbheinn, Beinn an Lochain, and Coire Beanaidh, are clearly associated with rock-slope failures, which appear to be an important debris supply mechanism for rock glacier development. In Norway and Switzerland, the rock glaciers studied do not appear to be associated with rock-slope failures, although structural trends of joints could be detected in many of the blocky-weathering source walls. It is possible that the alignment of these joints, which is often near-horizontal, may have inhibited rock-slope failures.

Plate 4.1 Relict valley-wall rock glacier at Strath Nethy, Cairngorm Mountains, showing the very pronounced frontal margin and travel distance away from the talus slope. A rock slope failure scar can be seen at the top of the photograph and jointing on the rockwall suggests extensive translational sliding.

Another related factor in rock glacier formation seems to be the height of the rockwall. In Table 4.4, the height of the talus slope and the height of the rockwall above each of the study rock glaciers is summarised. As can be seen, rockwall heights are generally greater than those of the talus slopes below. In Norway, where gabbro is the dominant lithology, large talus slopes have developed beneath high rockwalls (Plate 4.2). Similarly, in the Cairngorms in Scotland, weathering of massive granitic rockwalls has produced slightly less extensive talus slopes, whilst in the Swiss study areas, where less massive rock types predominate, the majority of rock glaciers are associated with much smaller talus slopes. An interesting aspect of the data presented in Table 4.4 is that rockwalls above active rock glaciers may be quite small whilst those above nearby inactive rock glaciers can be much larger. For example, the rockwall above the inactive rock glacier Arolla 1 (400m) is almost four times the size of the rockwall associated with the active rock glacier Arolla 3 (120m). This suggests that rockwall height is not as important a factor in controlling the local distribution of valley-wall rock glaciers as temperature and precipitation.

4.3.6 Topographic location

All valley-wall rock glaciers investigated are located at the base of talus slopes, generally beneath fractured cliffwalls of homogeneous lithology, and most have formed on slopes inclined gently away from the foot of the talus. However, at Gornergrat in the Swiss Alps, a valley-wall rock glacier has developed partly on a reverse slope (i.e. inclined towards the

Plate 4.2 Rockwall and active talus slopes above the complex valley-wall rock glacier on the eastern side of Fornesdalen, Lyngen. The height of the rockwall, up to 700m, is greater than the 300m high talus slopes.

Study site	Rockwall height (m)	Talus height (m)	Ratio of rockwall to talus height	Lithology
In Switzerland:				
Arolla 1	400	300	1.3	Schist
Arolla 2	180	80	2.3	Schist
Arolla 3	120	70	1.7	Schist
Zinal 1	90	50	1.8	Schist
Zinal 2	150	50	3.0	Schist
Flüela-Wisshorn 3	180	100	1.8	Gneiss
Schwarzhorn	250	120	2.1	Schist
Macun 1	100	80	1.3	Schist
Macun 3	60	70	0.9	Schist
Chilbiritzenspitz	170	80	2.1	Schist
Radüner	100	60	1.7	Schist
Gornergrat	30	60	0.5	Schist
In Norway:				
Fornesdalen West	500	250	2.0	Gabbro
Fornesdalen East	700	300	2.3	Gabbro
Rodbergdalen	450	200	2.2	Gabbro
Trollvatnet	200	150	1.3	Gabbro
Tverrelvdalen	200	100	2.0	Gabbro
Gjerdelvdalen	150	150	1.0	Amphibolite
Fastdalen North	250	150	1.7	Gabbro
Fastdalen South	300	150	2.0	Gabbro
Stordinddalen	800	300	2.7	Gabbro
Forholtskardet	500	200	2.5	Gabbro
In Scotland:				
Strath Nethy	250	150	1.7	Granite
Loch Etchachan	100	60	1.7	Granite
Lairig Ghru	200	100	2.0	Granite
Coire Beanaidh	80	100	0.8	Granite
Baosbheinn	120	100	1.2	Sandstone
Jura	150	200	0.7	Quartzite

Table 4.4 Rockwall and talus heights for each of the study rock glaciers, together with lithology. Rockwalls are generally greater in height than the talus slopes below.

talus foot), which suggests that in unusual topographic situations, rock glacier movement may be 'uphill' against the gradient of the underlying surface if debris supply is sufficient. On the other hand, where valley-wall rock glaciers have moved outwards into a steeply-inclined valley, some downvalley deflection of the rock glacier may have occurred as is the case at Forholtskardet in Lyngen (Plate 4.3). In sum, all valley-wall rock glaciers are located at the base of talus slopes and most beneath high fractured cliffwalls.

4.4 Discussion

The primary factors controlling the regional distribution of valley-wall rock glaciers, and indeed valley-floor rock glaciers and ice glaciers, are those of temperature and precipitation. Boundary conditions have been proposed by several authors, most notably Haeberli (1985) who suggested that active rock glaciers require a mean annual temperature below *ca.* -1°C to -2°C and an annual precipitation less than 2500mm (cf Fig. 2.3-1). His third major constraint was that active valley-wall rock glaciers require a minimum basal slope of 5° to permit creep.

The validity of these boundary conditions has been tested using the study sample. Active valley-wall rock glaciers studied in Switzerland all fall within the limits of Haeberli's conditions in terms of temperature and precipitation. Mean annual temperature values at the active rock glaciers are consistently less than -3°C (Barsch, 1978) and annual precipitation values are below *ca.* 700mm. Precipitation values in the

Plate 4.3 Two valley-wall rock glaciers in Forholtskardet, Lyngen, that have been deflected down a steeply inclined valley. Figure for scale is indicated by black arrow.

high mountains may increase up to a maximum value of *ca.* 2500mm (Wallén, 1977) which corresponds with the upper limit suggested by Haeberli. Interestingly, the field areas in Valais and Graubünden, both of which Barsch (1978) found to contain very high densities of rock glaciers, correspond to two of the driest areas in Switzerland, consistent with earlier ideas concerning low precipitation requirements for rock glacier development (e.g. Thompson, 1962; Harris, 1981). However, the basal slope gradient limitations imposed by Haeberli are invalid for two Swiss valley-wall rock glaciers, Gornergrat and Macun 1, which lie on gentler slopes. At Gornergrat, some rock glacier movement has apparently been 'uphill' against the basal gradient.

The relict rock glaciers in Scotland, which almost certainly developed during the Loch Lomond Stadial, are thought to have experienced temperature and precipitation regimes within the limits suggested by Haeberli. Mean annual temperatures at the rock glacier sites are unlikely to have exceeded *ca.* -5°C (based on Sissons, 1980; Ballantyne, 1984). Maximal precipitation values of *ca.* 500-600mm have been suggested by Sissons for the Cairngorms during the stadial. In the southwest Grampians, where precipitation levels reached a probable maximum of *ca.* 3000- 4000mm/yr), valley-wall rock glaciers are absent. The Norwegian study rock glaciers, (for which palaeoclimatic estimates are unavailable), and the Scottish rock glaciers all occur on gently-sloping sites with gradients greater than 5°. Thus, although the two climatic constraints suggested by Haeberli appear valid for the rock glaciers investigated, the slope gradient limitation appears to be too restrictive.

Haeberli's model, however, is also restrictive in the sense that it does not encompass the complete range of factors that operate to control rock glacier distribution. Although temperature and precipitation are the major climatic constraints, each of the additional factors outlined in the previous section has been shown to be of some importance in determining local distribution. In order to provide a more general summary of valley-wall rock glacier distribution, a diagram has been developed that identifies the range of factors discussed in this chapter (Figure 4.6). The controlling factors, which are highly correlated, cannot be summarised quantitatively in the same way that boundary conditions were defined for temperature and precipitation. For example, although rock glaciers favour a northerly orientation, some may face southwards because of unusual shading conditions or topographic control, and so precise limits cannot be placed on aspect. Similarly, in terms of altitude, valley-wall rock glaciers have formed above the lower boundary of discontinuous permafrost only where debris supply has been sufficient. Therefore, instead of extending Haeberli's model by including a list of poorly-defined constraints the summary diagram is in the form of a flow chart. Five main interdependent variables controlling rock glacier distribution, namely altitude, aspect, topographic location, source rockwall height and lithology, are represented at the first level of the flow chart. The first three of these factors determine, to varying degrees, local temperature and precipitation conditions. Lithology, headwall height and joint density together with freeze-thaw patterns determine debris supply characteristics. Where a balance is reached between debris

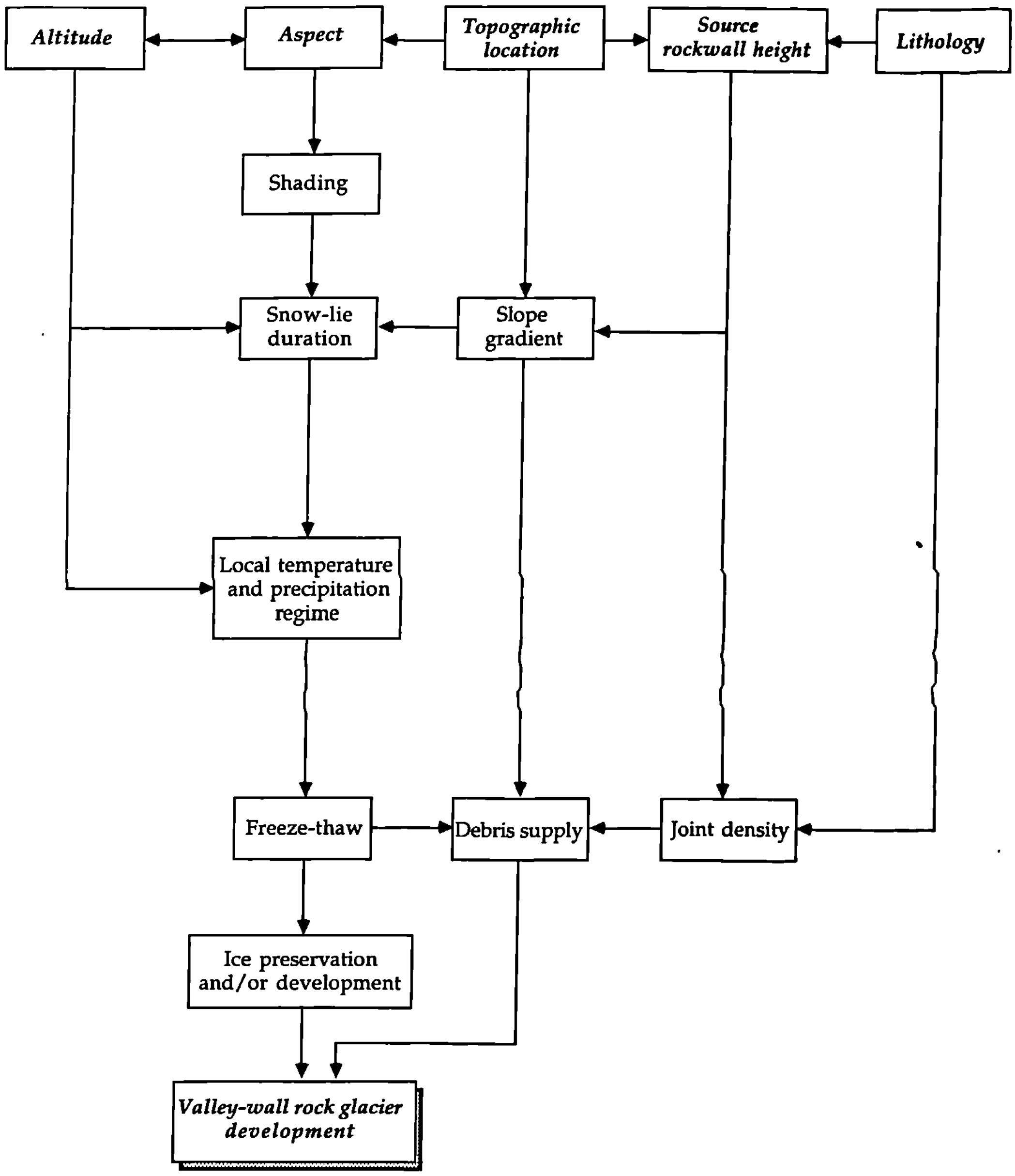


Figure 4.6 Summary diagram of factors controlling the local distribution of valley-wall rock glaciers at any latitude.

supply and ice preservation and/or development, valley-wall rock glaciers may develop.

4.5 Conclusions

The main aims of this chapter were to describe four major study areas and to examine rock glacier distribution within these areas. The study areas in Switzerland, Norway and Scotland contain examples of active, inactive, and relict valley-wall rock glaciers that have formed on a wide range of lithologies, altitudes, and aspects. The relict rock glaciers in Scotland almost certainly formed during the Loch Lomond Stadial (*ca.* 11,000 - 10,000 yr. B.P.), as this is the only period of permafrost conditions known to have affected upland Scotland following ice-sheet deglaciation. On the basis of dated moraine systems, several inactive and relict rock glaciers studied on the Lyngen peninsula in Norway seem to have formed during the Preboreal (*ca.* 9900 - 9400 yr. B.P.). A valley-floor rock glacier in Ellendalen, Lyngen, which is presently active, almost certainly dates from the Little Ice Age. Several of the rock glaciers studied in Switzerland are presently active.

The primary factors that control the regional distribution of valley-wall rock glaciers are temperature and precipitation. Previously proposed boundary conditions (Haeberli, 1985), which were validated using field data, suggest that active valley-wall rock glaciers require a mean annual temperature below *ca.* -1°C to -2°C and an annual precipitation less than 2500mm. In addition, several factors, namely aspect, altitude, lithology, bedrock source-wall characteristics, and topographic location, were found to exert a controlling influence on the local distribution of rock glaciers. Consideration of these factors resulted in eight main conclusions:

1) Rock glaciers are generally favoured by north and northwesterly aspects; the few that experience more southerly aspects tend to receive reduced levels of insolation because of topographic shading. In Scotland and in the Graübunden region of Switzerland, easterly aspects are more common due to large-scale structural control.

2) Active rock glaciers form above the lower boundary of discontinuous permafrost. Some rock glaciers, however, that occur above the discontinuous permafrost boundary are inactive, which indicates that both climatic and non-climatic factors must operate to determine local distribution.

3) In Scotland, the altitudinal distribution of relict rock glaciers is largely consistent with reconstructed regional firn line altitudes of Loch Lomond Stadial glaciers, which suggests that snow accumulation and ablation must have decreased from a maximum in southwest Scotland to a minimum in the northeast near the Cairngorm Mountains.

4) Rocks such as gneiss, granite and gabbro, which all produce massive blocks on weathering, appear to be the most favourable lithologies for rock glacier formation.

5) Rock glacier distribution is partly influenced by the intensity of fractures in the bedrock source-wall. Five rock glaciers in Scotland are associated with rock-slope failures that have occurred along joints, which

in the Cairngorms, are generally aligned parallel to the rockwall surfaces.

6) All valley-wall rock glaciers form at the base of talus slopes, generally beneath fractured rockwalls.

7) In general, the size of the rockwall above a rock glacier, which is partly controlled by lithology, is greater than the height of the talus slope.

8) A wide range of factors that are highly correlated determine the local distribution of valley-wall rock glaciers.

Chapter 5

Rock Glacier Morphology

5.1 Introduction

This chapter presents data on the morphological characteristics of a sample of active, inactive and relict valley-wall rock glaciers in Switzerland, Norway and Scotland. Following a brief introductory rationale, data collection methods are summarised. Rock glacier dimensions and gradients are then examined and long profiles are illustrated. Despite variability in the form of these profiles, morphometric relationships were found to exist between several parameters including rock glacier length, width and thickness, surface and frontal slope gradients, frontal height and slope form. Statistical analyses of these relationships are presented and discussed. Attention is then focused on surface microrelief, and a detailed analysis is made of frontal slope morphology, transverse ridge morphology, longitudinal and closed depressions, and surface ponds. The chapter concludes with a summary discussion.

5.2 Rationale

As outlined in the rationale section of the previous chapter, a large sample of geographically and morphologically diverse active, inactive and relict valley-wall rock glaciers was selected for field study so that the validity of each model of valley-wall rock glacier formation proposed in Chapter 3 could be rigorously tested using a sufficiently large data base. In this chapter, field evidence relating to the morphological characteristics of valley-wall rock glaciers is presented. Three major aspects of research reported in this chapter will be carried forward into subsequent chapters that, first, examine the theoretical boundary conditions for valley-wall rock glacier formation and, second, combine this theoretical evidence with field evidence to test each model of formation. These three lines of enquiry are: 1) an examination of the range of rock glacier dimensions, including length, width, thickness, volume and surface gradient, in order to provide an envelope of values within which theoretical calculations may be set; 2) the identification of statistically significant morphometric relationships both in terms of large-scale rock glacier dimensions and small-scale surface relief features; and 3) a description and analysis of the morphological development of valley-wall rock glaciers and the implications of this developmental continuum in terms of process. In addition to these three major aims, this chapter presents site-specific morphological information that will also be used to test models of valley-wall rock glacier formation in subsequent chapters.

5.3 Data collection

Following preliminary aerial photographic selection of suitable study sites, all morphometric measurements on rock glaciers were made in the field. From an initial sample of forty rock glaciers whose distribution was described in Chapter 4, twenty-four valley-wall and three valley-floor rock glaciers were selected for detailed morphological examination. At each site at least one long section was surveyed using either a theodolite or an abney level, tape and ranging rods. The theodolite used for these measurements was a Sokkisha TM6 model, which has a resolution to the nearest six seconds of arc. Repeatability tests using abney levels suggest that individual angular measurements are reproducible to within ±0.5°. The subjectivity inherent in both types of survey was reduced to a minimum by precise positioning of survey points at distinct breaks of slope and at regular intervals between breaks of slope. Measurements obtained by theodolite are marginally more accurate than those obtained using abney levels. However, as it was not possible to survey some large surface depressions unless the theodolite was moved from the frontal ridge crest, an advantage of surveying by abney level was that continuous profiles could be obtained without losing time relocating the theodolite.

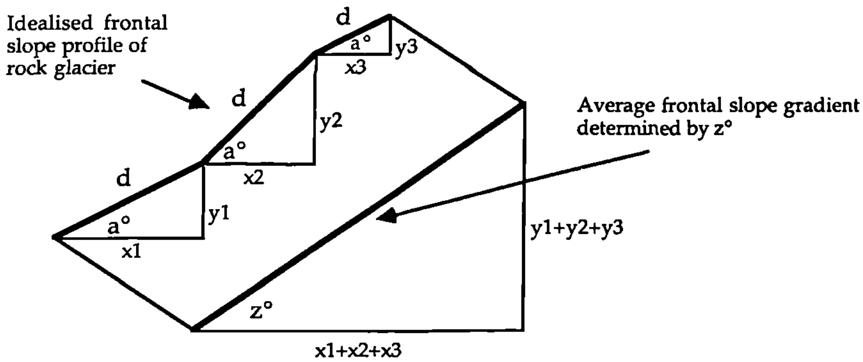
The maximum distance between sitings on long profiles was established at 30m, although extra sitings were carried out at each break of slope. Sitings for more detailed surveys of frontal and transverse ridges were made every 5m or more frequently whenever a break in slope occurred. Each rock glacier was surveyed along its medial long profile, and at several of the larger sites additional long profiles were surveyed on either side of the medial line. In addition, the distal and proximal slopes of lateral and frontal ridges were surveyed at 30m intervals along the ridge crests.

The survey data for each rock glacier was used to construct representative profiles generally along the medial long profile of the rock glacier. In order to determine if statistically significant morphometric relationships could be derived from the survey data, various dimensions and gradients were calculated, namely rock glacier length and width, frontal slope height, average surface gradient, frontal slope gradient, rock glacier thickness and slope form.

In previous morphological studies, different definitions of rock glacier length and width have produced non-comparable measurements. Whilst the toe of the frontal slope near the midpoint of the rock glacier provides a clearly identifiable downslope limit to rock glacier length, the upslope limit of valley-wall rock glaciers is less distinct. In the past, the upslope limit has rather vaguely been defined as where the rock glacier merges with the talus upslope. On inactive and relict rock glaciers, in particular, post-depositional mass-movement has often obscured the talus/rock glacier margin. For consistency, the upslope limit of rock glacier development in this study was determined to be where the surveyed profile exceeds an average gradient of 30° by a maximum slope distance of 30m. Rock glacier width was measured wherever the rock glacier was at its greatest lateral extent, which was rarely at the frontal margin.

The height of the frontal slope of each rock glacier was calculated for the midpoint of each feature and additionally in many cases at 30m intervals along the frontal and lateral margins of the feature. In addition, average frontal slope gradients were determined along medial long profiles and elsewhere on the frontal margins of the rock glaciers. Each of these parameters was calculated from the survey data using simple trigonometry as illustrated in Figure 5.1. The average surface slope gradient of the rock glaciers above the frontal ridge was similarly determined, allowing for negative slope angles on the proximal slopes of transverse ridges.

Rock glacier slope form was examined using three parameters. The most straightforward slope form parameter was calculated by dividing the mean gradient of the uppermost third of the long profile by that of the lowermost third. Values greater than one suggest predominantly concave slopes, values less than one predominantly convex slopes and values approaching one indicate an approximately rectilinear slope. It was noted, however, that this method assigns values that approach rectilinearity to rock glaciers that exhibit pronounced convexo-concave surfaces, as concave and convex sections of similar size tend to cancel each other out. Thus, two additional indices proposed by Church et. al. (1979) were used to provide further information on rock glacier slope form; these are the index of departure from linearity and the index of convexity/concavity, and are defined in Figure 5.2.



For each slope facet within the profile, slope distance (d) and slope gradient (a°) were measured using a theodolite or abney level and tape. Change in elevation (y) for each slope facet was then calculated using the equation:

$$y_n = (d \sin a°)_n$$

and true horizontal distance was calculated using:

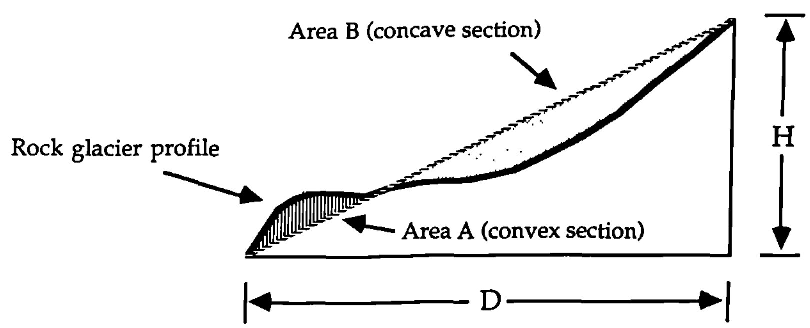
$$x_n = (d \cos a°)_n$$

These values were then summed to give overall change in elevation and horizontal distance.

To obtain the average gradient of any section of the rock glacier (z°), the following equation was used:

$$z° = \tan^{-1} (y1+y2+y3) / (x1+x2+x3)$$

Figure 5.1 Method used to calculate average surface gradients and frontal slope gradients from survey data.



Standardising factor F = HD/2 (which equals the area of the right-angled triangle defined by the limits of the slope)

Index of departure from linearity = (A + B)/F

Index of convexity/concavity of the slope = A/B

Figure 5.2 Definition of the index of departure from linearity and the index of convexity/concavity. Based on Church et al., 1979.

By surveying a profile of the hillslope adjacent to a rock glacier an estimate of rock glacier thickness could be determined. Precise reference points had to be established between the rock glacier profile and the adjacent hillslope profile so that transects could be compared. For example, at Fornesdalen West rock glacier (Plates 5.1 and 5.2) a profile for estimating thickness was surveyed on the hillslope to the north of the feature. The position of a lateral moraine was used as a reference point between the medial long profile of the rock glacier and the long profile of the adjacent hillslope. At many study sites, however, profiling to determine thickness was not possible as the adjacent hillslope was too dissimilar either in form or in orientation. In consequence, rock glacier thickness at such sites had to be calculated by assuming a regular decline in slope gradient under the rock glacier. Clearly, this assumption introduces a degree of imprecision but where comparison of the two methods was made only small discrepancies resulted from the assumption of a regular decline in the gradient of the underlying terrain (see Table 5.1, below).

Finally, by using a combination of long profile data, thickness estimates and width measurements, estimates of rock glacier volume were determined. On four rock glaciers, a series of between two and four long profiles indicated that for any position on the medial long profile, thickness is relatively constant across the entire width of the rock glacier unless rock glacier length varies markedly within one feature (Figure 5.3). For example, thickness variations within the first three rock glaciers shown in Figure 5.3, namely Trollvatnet, Rödbergdalen and Fornesdalen

Plate 5.1 Fornesdalen West valley-wall rock glacier on the Lyngen Peninsula, Norway. Frontal ridge rests on a lateral moraine. Post-formational avalanching and debris flow activity has partially buried surface microrelief on southern (left) part of rock glacier. Inner transverse ridges can be seen on northern half.

Plate 5.2 Lateral view of Fornesdalen West rock glacier. In the foreground at the left, the proximal slope of the lateral moraine can be seen. A profile for estimating rock glacier thickness was surveyed on the hillslope in the foreground, which is to the right of the rock glacier in Plate 5.1.

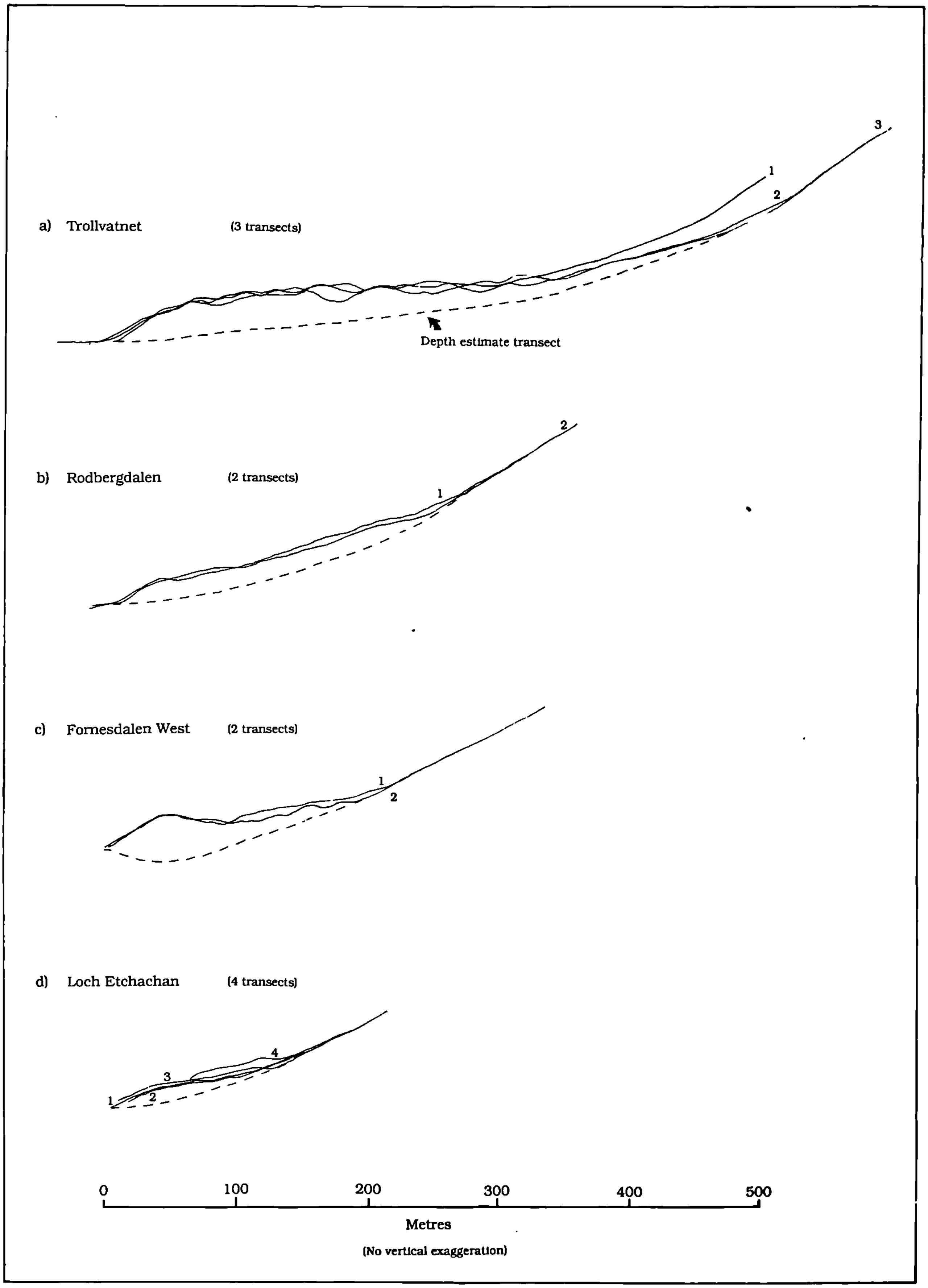


Figure 5.3 Long and depth profiles for four study rock glaciers showing that depth is relatively constant across the entire width of the feature (a, b, c), unless rock glacier length varies markedly within one rock glacier (d).

West, are relatively limited; however, in the fourth example, at Loch Etchachan, the thickness of the rock glacier is more variable largely because rock glacier length ranges from 180m at the position of the medial long profile (transect number 1 in Figure 5.3d) to less than 75m at the lateral margin (transect number 4). Therefore, when calculating the volume of rock glaciers that display pronounced variations in length, additional length and thickness transects had to be surveyed towards the lateral margins of the features, as medial data could not be extrapolated across the entire width of the rock glacier with a sufficient level of precision. Then, by subdividing the length of the rock glacier into a number of slices for which average width could be measured and thickness estimated, an estimate of rock glacier volume was established by summing the volumes of the individual slices (Figure 5.4). Adjustments made in terms of void space, to determine rock volume as opposed to rock glacier volume which includes void space, are discussed in section 5.4.4 below, where volumetric calculations are presented.

5.4 Valley-wall rock glacier dimensions

5.4.1 Introduction

Tables 5.1 to 5.3 summarise the dimensions and gradients of the study rock glaciers and show the inferred activity status of each rock glacier. Rock glacier activity, as discussed in section 4.2.1 above, was determined mainly using morphological criteria and, where possible, sedimentological evidence. The valley-wall rock glaciers studied extend 70 to 600m beyond talus slopes and are 10 to 35.5m in depth. Widths

Rock glaciers are of relatively constant depth across their width at any given position along the medial long profile unless rock glacier length varies markedly within one feature. Rock glacier volume is best estimated using a combination of long profiles, depth estimates and width data as shown below.

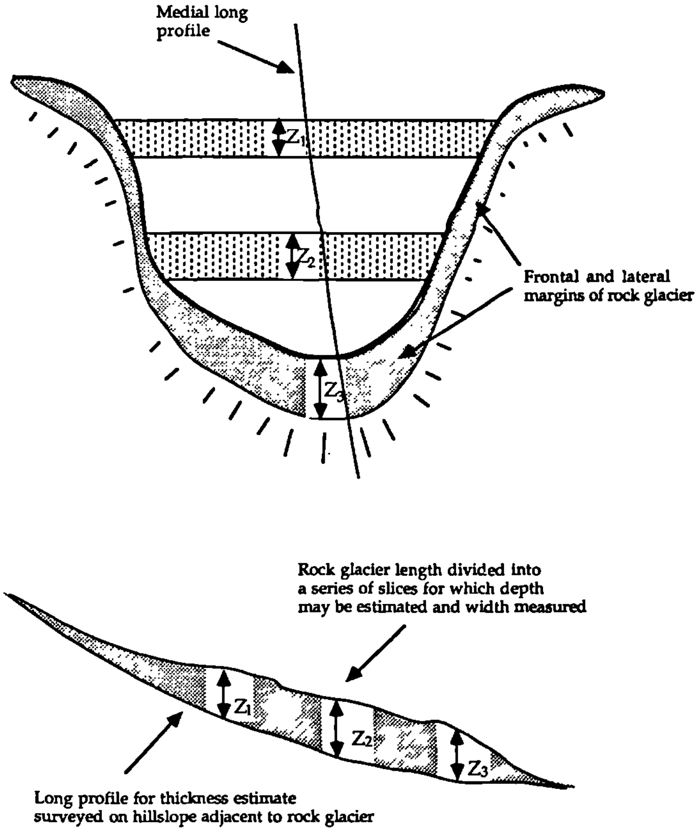


Figure 5.4 Method used to determine rock glacier volume assuming a relatively constant depth at any point on the medial long profile.

	Study site	Activity status	Width (m)	Length (m)	Length/Width	Mean surface slope angle	Mean frontal slope angle	Height of rock glacier front (m)	Profiled maximum depth estimate (m)	Slope decline maximum depth estimate (m)
Valley-wall rock glaciers	Arolla 1	Inactive	250	600	2.40	17.6°	31.4°	29.4	35.5	35
	Arolla 2	Inactive	60	70	1.16	22.1°	35.1°	11.1	11.0	11
	Arolla 3	Active	150	375	2.51	16.7°	37.8°	84.6	-	15
	Zinal 1	Inactive/Active	80	180	2.25	21.0°	35.7°	26.6	-	20
	Zinal 2	Inactive	300	400	1.33	14.2°	28.2°	14.7	-	25
	Flüela-Wisshorn 3	Inactive	120	100	0.83	26.9°	33.3°	16.8	-	16
	Schwarzhorn 1	Inactive	130	225	1.73	14.2°	29.9°	12.2	-	15
	Schwarzhorn 4	Inactive	100	75	0.75	23.9°	34.1°	16.5	-	10
	Macun 1	Inactive/Active	200	175	0.87	13.1°	33.0°	27.3	-	25
	Macun 3	Inactive	75	100	1.33	20.9°	39.3°	19.5	17.5	17
	Chilbiritzenspitz	Inactive	550	475	0.86	16.9°	32.5°	31.4	-	33
	Radüner	Inactive/Active	250	230	0.92	18.4°	40.7°	24.9	-	28
	Gornergrat	Inactive/Active	350	170	0.48	3.3°	26.6°	5.6	-	10
Valley-floor rock glacier	Fluela-Wisshorn 1	Inactive	100	480	4.80	16.9°	33.1°	9.0	-	15

Table 5.1 Selected dimensions and gradients of rock glaciers studied in Switzerland

	Study site	Activity status	Width (m)	Length (m)	Length/Width	Mean surface slope angle	Mean frontal slope angle	Height of rock glacier front (m)	Profiled maximum depth estimate (m)	Slope decline maximum depth estimate (m)
Valley-wall rock glaciers										
	Fornesdalen West	Inactive	500	320	0.64	14.6°	33.6°	24.9	30	35
	Fornesdalen East (total)	Inactive	4500	350	0.14	-	-	-	-	-
	Fornesdalen East 1	Inactive	300	330	1.10	11.1°	31.0°	14.4	-	16
	Fornesdalen East 2	Inactive	250	350	1.40	10.4°	28.0°	11.4	-	15
	Rodbergdalen	Inactive/Relict	700	250	0.36	16.2°	28.5°	17.7	-	17.5
	Trollvatnet	Inactive/Relict	550	510	0.93	9.7°	32.0°	30.9	32	34
	Tverrelvdalen	Inactive	400	250	0.63	14.5°	30.7°	19.4	-	25
Valley-floor rock glacier										
	Tytebaerdalen	Relict	420	775	1.84	18.6°	31.2°	20.7	-	18

Table 5.2 Selected dimensions and gradients of rock glaciers studied on the Lyngen Peninsula, northern Norway

	Study site	Activity status	Width (m)	Length (m)	Length/Width	Mean surface slope angle	Mean frontal slope angle	Height of rock glacier front (m)	Profiled maximum depth estimate (m)	Slope decline maximum depth estimate (m)
Valley-wall rock glaciers	Strath Nethy	Relict	2400	350	0.15	17.3°	33.3°	26.0	-	28
	Loch Etchachan	Relict	350	180	0.51	17.4°	26.3°	10.2	12.0	10.5
	Lairig Ghru	Relict	425	200	0.47	19.9°	29.9°	34.8	-	16
	Coire Beanaidh	Relict	600	200	0.34	12.4°	23.0°	9.7	-	10
	Baosbheinn	Relict	485	180	0.37	13.0°	30.0°	31.0	-	19
Valley-floor rock glacier	Beinn an Lochain	Relict	250	500	2.00	25.7°	34.1°	26.0	-	17

Table 5.3 Selected dimensions and gradients of rock glaciers studied in Scotland

range from 60m in the case of simple lobate forms to over 4.5km for the complex rock glacier at Fornesdalen East. Such pronounced dimensional variations hint at the complexities involved in attempting to provide a general explanation of valley-wall rock glacier formation. In addition, rock glaciers of markedly contrasting dimensions often occur within small lithologically homogeneous regions such as in the Montagne d'Arolla in Valais, Switzerland, where a rock glacier only 70m long and 60m wide is located less than 200m from one that is 600m long and 250m wide (Plates 5.3 and 5.4).

Preliminary visual inspection of the long profiles (Figures 5.5 to 5.8) indicates that at the majority of sites rock glacier surfaces slope gently upslope of the frontal ridge then gradually steepen to merge with talus. The number of inner transverse ridges varies from nil to thirteen. Some rock glaciers have reverse slopes which form depressions that are filled by meltwater. Downslope of the frontal crest, rock glaciers terminate in generally steep slopes of 23° to 41°. The aims of this section are to examine each large-scale dimensional characteristic and to present the results of statistical morphometric relationships derived using linear regression.

5.4.2 Rock glacier length and width

Length, width, and length to width ratios for each rock glacier studied are summarised in Tables 5.1 to 5.3. A widely-used classification of rock glaciers is based on the assumption that valley-wall rock glaciers exhibit a

Plate 5.3 Arolla 1 valley-wall rock glacier in Valais, Switzerland, showing two distinct units of movement. This inactive rock glacier is 500m long, 250m wide, and approximately 35m thick at its maximum.

Plate 5.4 Arolla 2 is a much smaller inactive valley-wall rock glacier located 180m from Arolla 1. This feature is only 70m long, 60m wide and approximately 11m thick at its maximum. Note the concentration of large clasts along the crest ridge. Figure for scale is standing on crest ridge, indicated by arrow.

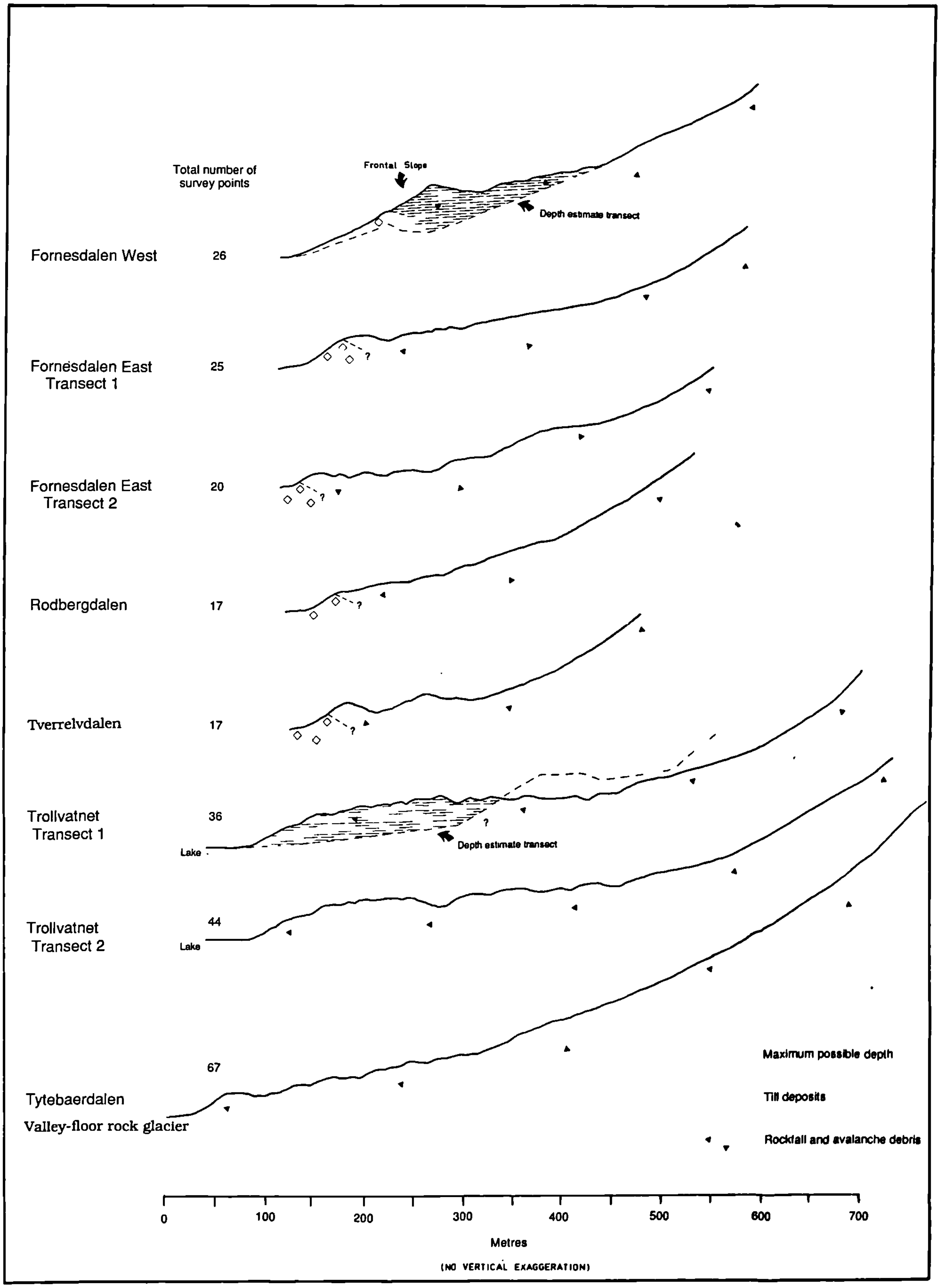


Figure 5.5 Long profiles of seven rock glaciers studied on the Lyngen Peninsula, northern Norway. All are valley-wall rock glaciers except Tytebaerdalen, which is a valley-floor rock glacier.

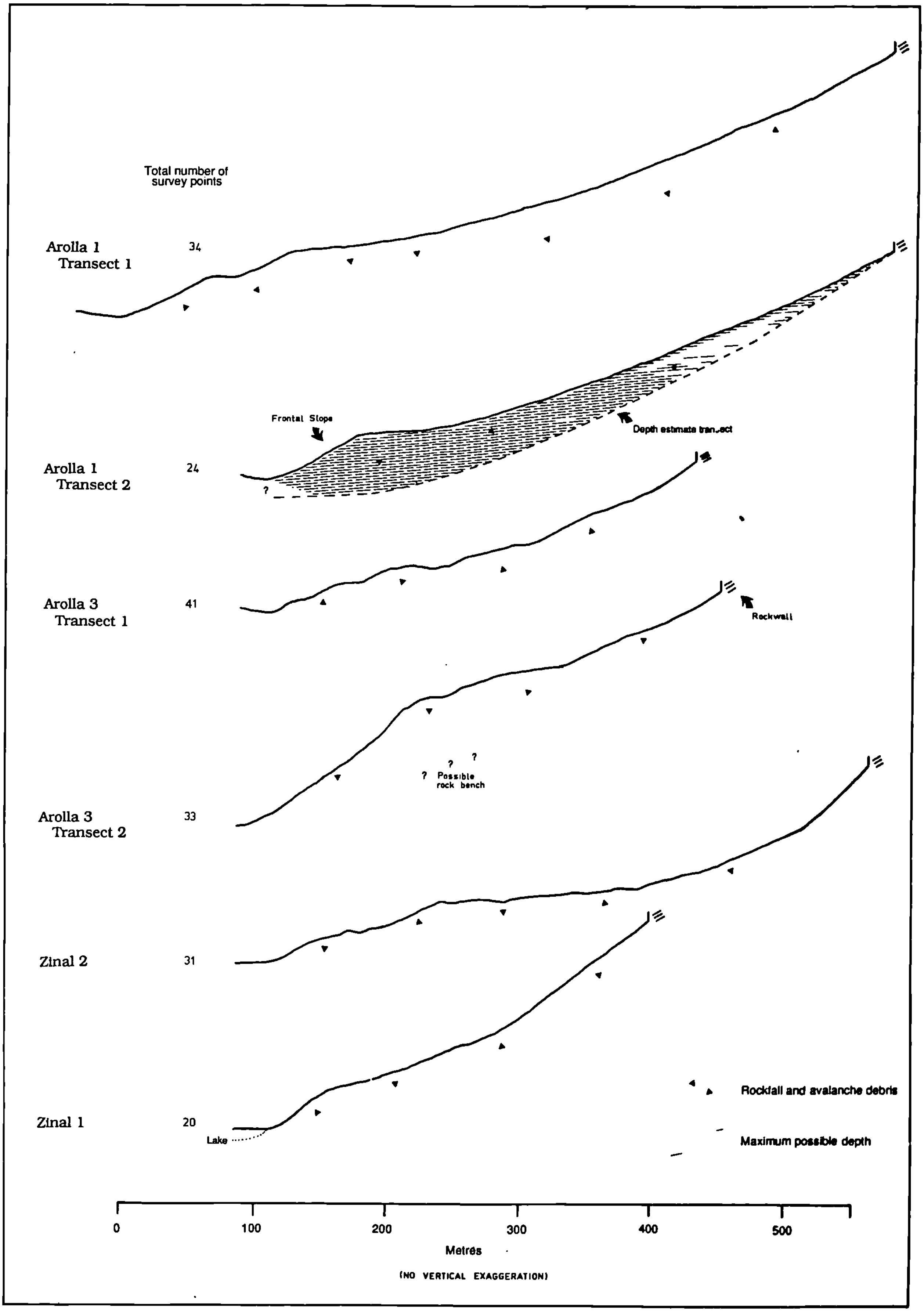


Figure 5.6 Long profiles of four rock glaciers studied in Switzerland.

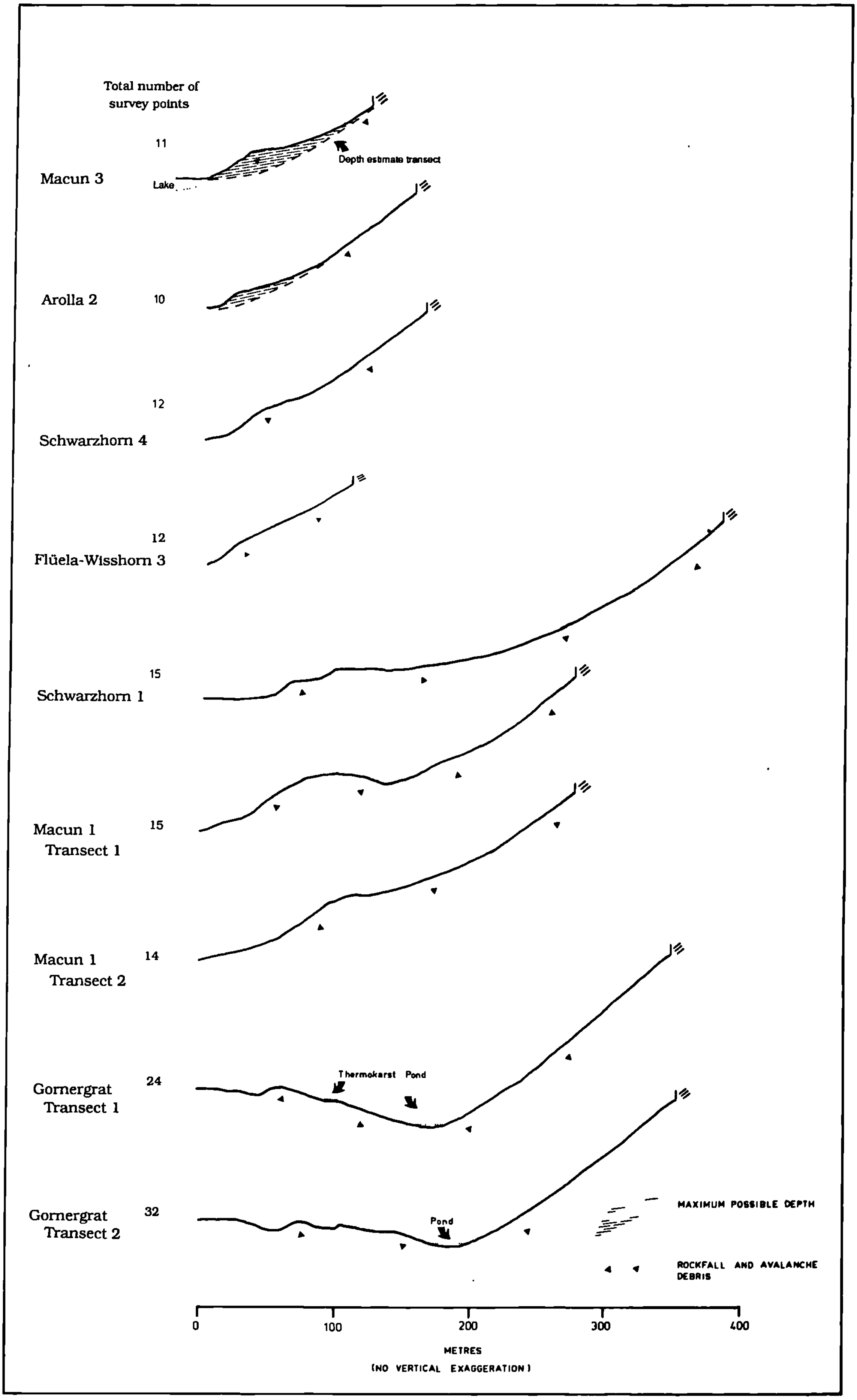


Figure 5.7 Long profiles of seven rock glaciers studied in Switzerland.

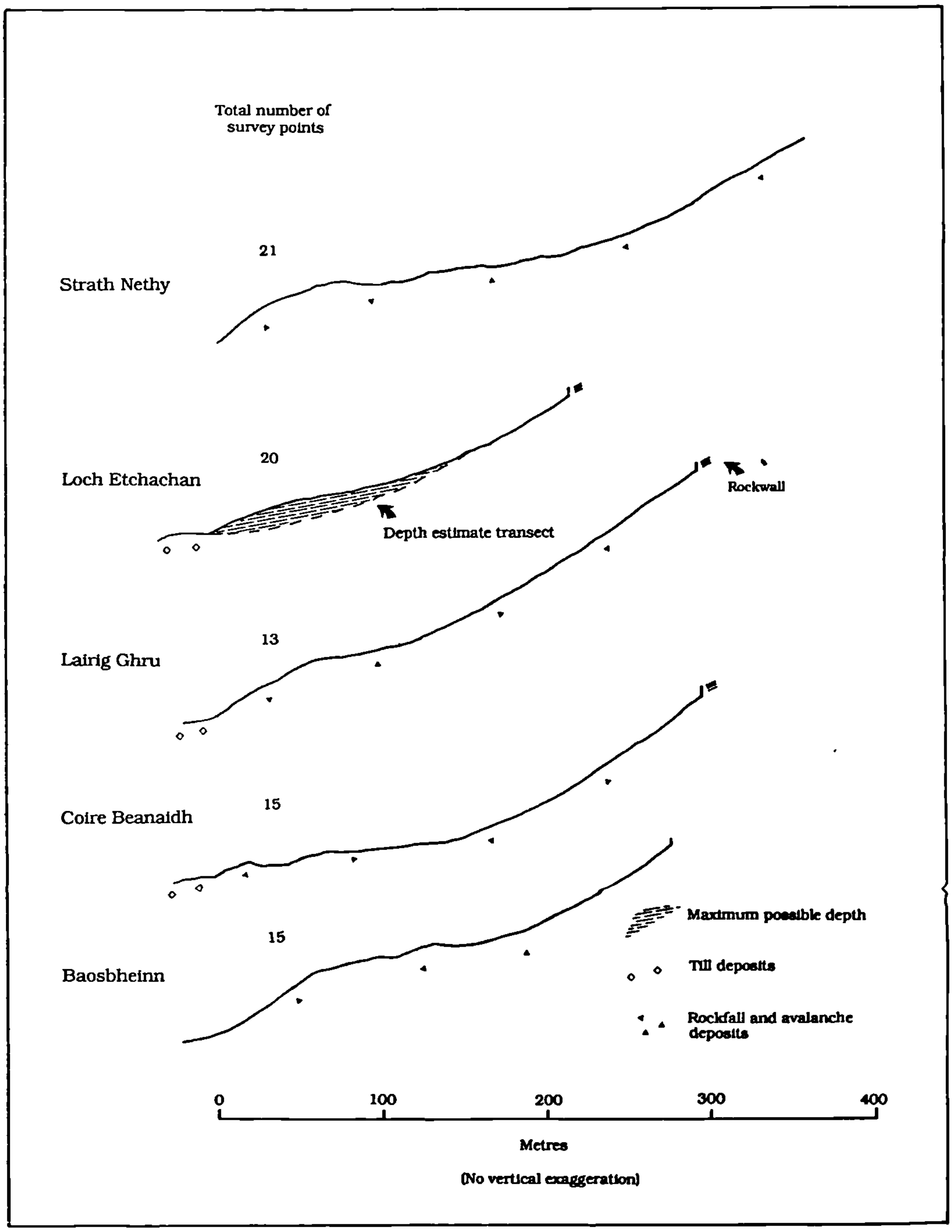


Figure 5.8 Long profiles of five rock glaciers surveyed in Scotland.

length to width ratio of less than one, whereas the same ratio for valley-floor rock glaciers is greater than one (e.g. Domaradzki, 1951; Wahrhaftig & Cox, 1959; Richmond, 1962; Outcalt & Benedict, 1965; see section 2.2.2). Using the data set to test the validity of this scheme, it is notable that for 9 out of 24 valley-wall rock glaciers length exceeds width, which indicates that the scheme is inappropriate. Perhaps part of the confusion concerning length to width ratios has arisen because of a tendency for coalescing valley-wall rock glaciers along a valley-side to be described as a single feature. Such complex forms, which many consider typical of valley-wall rock glacier development, are much wider than long; however, they are often a combination of several individual units of movement that may or may not be wider than they are long. For example, Plate 5.5 shows complex valley-wall rock glacier development near Zinal, in Valais, Switzerland. Within this feature several individual rock glaciers may be identified, and Plate 5.6 shows one such rock glacier lobe that is located towards the western part of the feature (right side of Plate 5.5). Its medial long profile (Zinal 1) is shown in Figure 5.6 together with a long profile of Zinal 2, which is located farther along the valley-wall. The rock glaciers at Zinal 1 and 2, which are both longer than wide, are markedly different in terms of length, surface gradient and inner transverse ridge development.

A similar complex valley-wall rock glacier occurs on the eastern side of Fornesdalen on the Lyngen peninsula in northern Norway. Here, valley-wall rock glaciers have formed over a distance of 4.5km, although within this lateral extent development has not been spatially

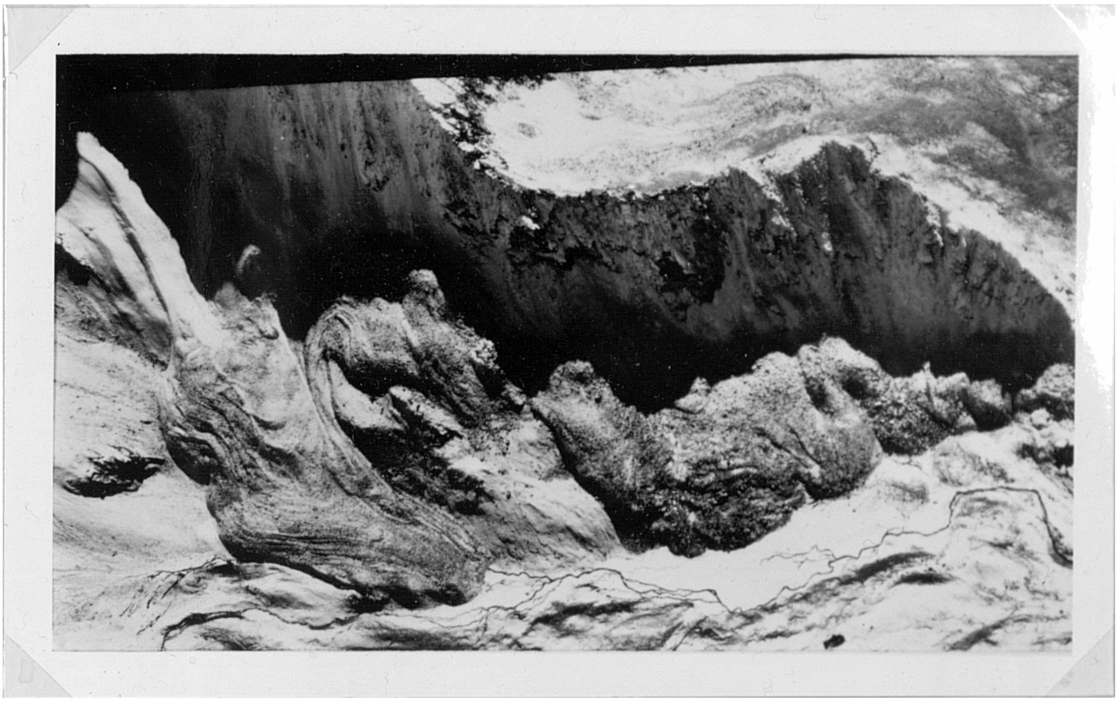


Plate 5.5 Aerial photograph of rock glacier development at Zinal. The heavily ridged rock glacier towards the left of the photograph is an active valley-floor rock glacier that merges with a corrie glacier at its upslope margin. To the right, extensive valley-wall rock glacier development in which travel distance away from the talus slopes exceeds 400m.

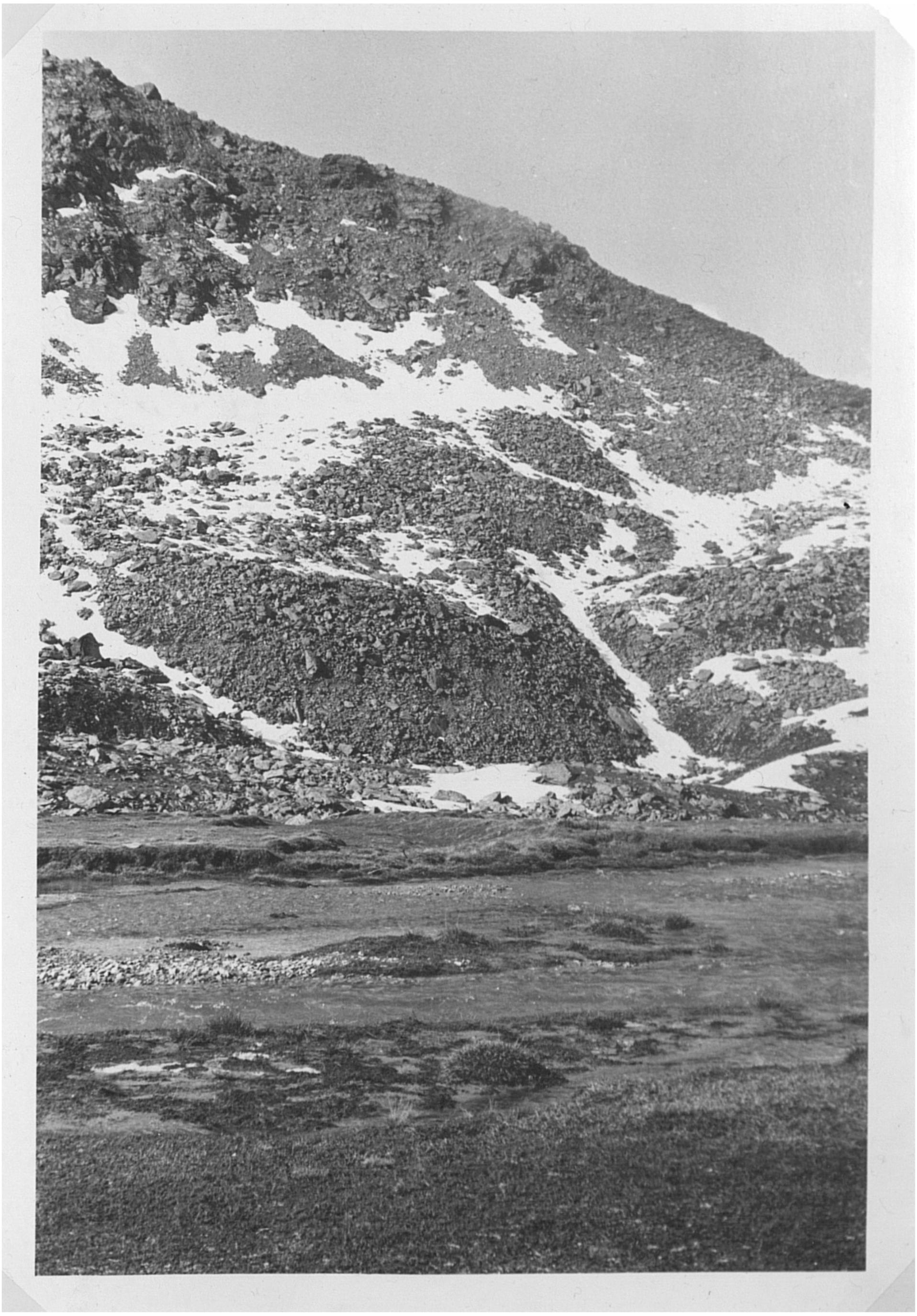


Plate 5.6 An inactive or active valley-wall rock glacier that is located at the western end of the complex rock glacier at Zinal (right side of Plate 5.5). Individual slope facets on the frontal slope exceed 45°.

continuous. Plate 5.7, which is an aerial photograph of Fornesdalen and its accompanying overlay, shows variations and discontinuities in rock glacier length and width along the length of this feature. The positions of the individual valley-wall rock glaciers Fornesdalen East 1 and 2, both of which are longer than wide, are indicated.

As discussed in Chapter 3, it is thought by several authors (e.g. Whalley, 1974; Liestøl, 1962; Dumbell, 1984; Vick, 1987) that rock glaciers develop when debris input exceeds a threshold point at which the shear strength of the underlying ice (or rock and ice mixture) is exceeded and movement occurs. If this is the case, it is most unlikely that this threshold will be reached simultaneously along several kilometres of a valley-wall if debris input is by a series of small-scale rockfall events. It is arguable, therefore, that neither Zinal nor Fornesdalen East should be considered as one rock glacier; instead they should be viewed as a series of contiguous individual valley-wall rock glaciers.

A third example of a complex valley-wall rock glacier is found at Strath Nethy in the northwestern part of the Cairngorm Mountains where rock glacier development extends almost continuously for 2.4km at the base of an east facing valley slope (Plate 5.8). The rock glacier at Strath Nethy, however, differs from those in Fornesdalen East and at Zinal because the frontal margin is more continuous and individual units of movement cannot be identified. It seems unlikely that modification of this relict feature since its development during the Loch Lomond Stadial could have subdued more pronounced units of movement. Instead it seems

FORNESDALEN, LYNGEN

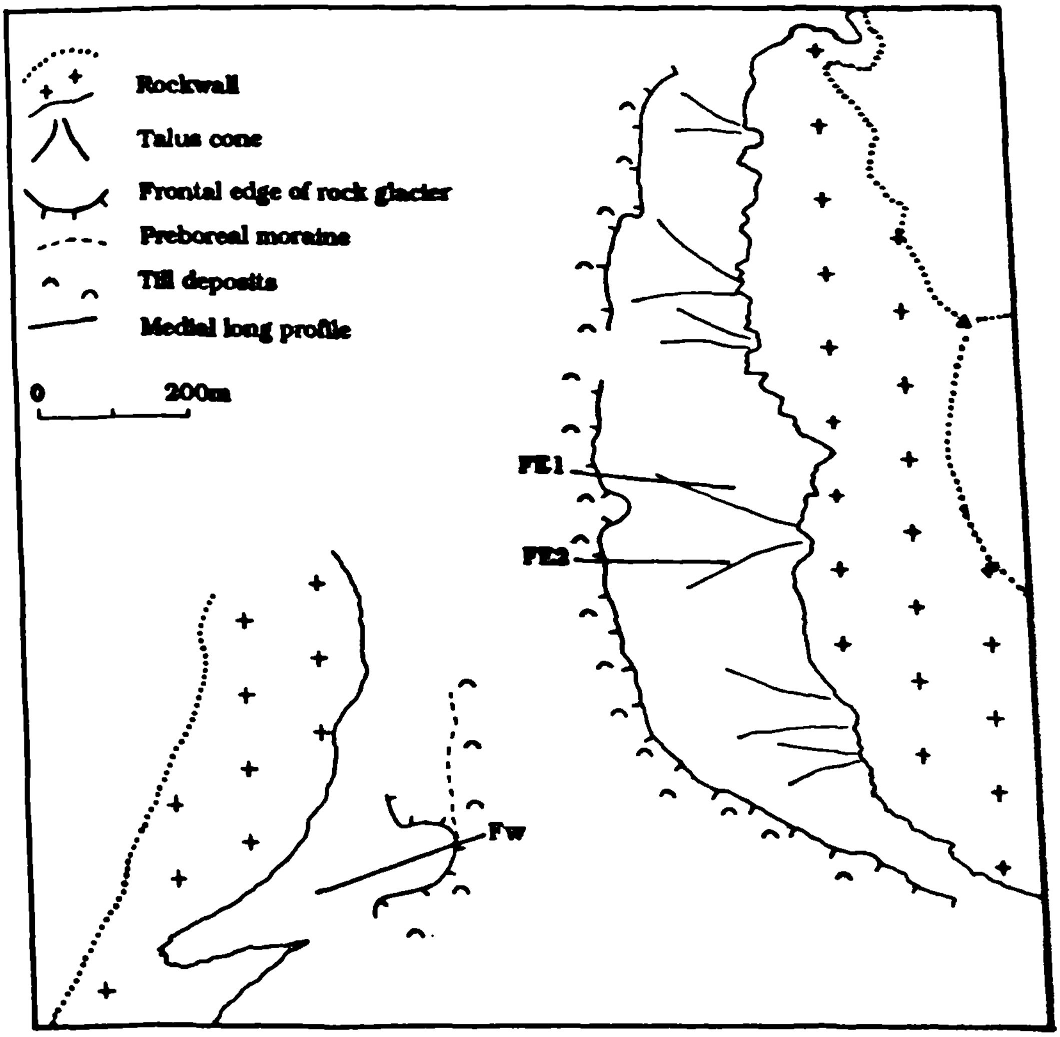


Plate 5.7 Fornesdalen on the Lyngen Peninsula, northern Norway. Fornesdalen West (FW) and Fornesdalen East (FE) valley-wall rock glaciers are marked on overlay. FW has terminated against the proximal slope of a lateral moraine (see Plate 5.1). FE lies on extensive till deposits and extends almost continuously for over 4.5km along the west-facing valley-slope at the base of extensive talus slopes.

FORNESDALEN, LYNGEN

Plate 5.7 Fornesdalen on the Lyngen Peninsula, northern Norway. Fornesdalen West (FW) and Fornesdalen East (FE) valley-wall rock glaciers are marked on overlay. FW has terminated against the proximal slope of a lateral moraine (see Plate 5.1). FE lies on extensive till deposits and extends almost continuously for over 4.5km along the west-facing valley-slope at the base of extensive talus slopes.

Plate 5.8 A complex relict valley-wall rock glacier at Strath Nethy in the Cairngorms, Scotland. Here the frontal margin is more continuous in length and height than at Zinal and individual rock glacier units cannot be detected. Rock glacier formation may have been more catastrophic.

that the formation of this feature was more catastrophic, and that debris input was more continuous both spatially and temporally, than at Zinal or Fornesdalen. Evidence that large-scale rockfall events occurred at Strath Nethy is provided by the presence of rock-slope failure scars resulting from translational sliding in the cliffwalls above the rock glacier (see section 4.3.5). The possibility that the origin of some rock glaciers is associated with the occurrence of rock slope failures is discussed further in Chapter 7.

5.4.3 Slope gradients and form

As explained in section 5.3, average gradients of frontal slopes and rock glacier surfaces upslope of the frontal ridge were calculated from the survey data (Tables 5.1 to 5.3). Average frontal slope gradients for the Swiss study rock glaciers range between 26.6° and 40.7° with individual slope facets on the active valley-wall rock glacier at Arolla 3 exceeding 50°. Very steep slope facets of 45° were recorded on the rock glaciers at Zinal 1 and Radüner, both of which are regarded as possibly active features (see Table 5.1). Average frontal slope gradients of inactive and relict rock glaciers on the Lyngen Peninsula and relict rock glaciers in Scotland ranged from 28.0° to 33.6° and from 23.0° to 33.3°, respectively. Maximum individual slope facet angles of less than 40° on these inactive and relict rock glaciers probably reflect modification by mass-movement processes. Frontal slope morphology is considered in detail in section 5.5.2.

Average gradients of rock glacier surfaces upslope of the frontal ridge were found to lie in a surprisingly wide range of values from 3.3° to 26.9°. However, the end members of this range account for much of the variance. Gornergrat rock glacier exhibits an average surface gradient of only 3.3°, whilst Flüela-Wisshorn 3 has a mean gradient of 26.9°. The rock glacier at Gornergrat is morphologically distinct from the rest of the sample because of a reverse basal slope gradient and is discussed in detail later in this section. At Flüela-Wisshorn, which is only 100m long, debris from recent rockfall avalanches has accumulated behind the frontal ridge, thus increasing an already high average slope gradient that reflects limited rock glacier length. When these end values are removed, the reduced range in average gradients is from 9.7° to 23.9°. Mean gradients of inactive and relict valley-wall rock glaciers in Norway and Scotland do not exceed 19.9°. Of the four Swiss rock glaciers whose gradients are greater than 20°, it is thought that only one may be active; they are all, however, very small features with at most one inner transverse ridge, and as travel distance away from the talus slope is limited they remain very steep. Clearly, the rock glacier surface gradients presented in Tables 5.1-5.3 are partly a function of the fact that the upslope extent of rock glaciers is defined by a 30° slope gradient; thus, although valid comparisons and contrasts may be drawn between rock glaciers within the data set, comparisons with other rock glacier data sets must be approached cautiously.

Viewed at a large scale, the form of each valley-wall rock glacier generally falls into three sections. The first section comprises a convex frontal ridge, 10 to 35m in height and 23° to 41° in average distal gradient. Upslope of the frontal ridge, average slope gradients decrease to between 6° and 20° in an approximately rectilinear section of variable length that includes all transverse ridges and depressions. In the uppermost concave section, average gradients increase until the rock glaciers merge with either active or inactive talus slopes. Variations in rock glacier slope form were examined quantitatively using three parameters. The first is a dimensionless parameter, termed slope form, that was derived by dividing the mean gradient of the uppermost third of the long profile by that of the lowermost third. Slope form values for the study sample range from 0.55 to 14.80 and are presented in the first column of Table 5.4. As discussed in section 5.3 above, this parameter was found to be inadequate in describing rock glacier slope geometry, as some rock glaciers that recorded near rectilinear slope form values, comprised pronounced convex and concave sections of similar size and gradient. Thus, two additional indices were included in the analysis, the results of which are shown in Table 5.4. The first is the index of departure from linearity, which summarises the departure from a rectilinear slope (Figure 5.2). Values range from 0.01 for the rock glacier Flüela-Wisshorn 3, which exhibits the most rectilinear slope, to 0.25 for Fornesdalen East 2 rock glacier, which shows the greatest departure from linearity. The second index indicates the relative amounts of convexity and concavity. Thus, Arolla 1, a long concave rock glacier records a ratio of 96.35 which contrasts with a value of 0.02 for the rock glaciers Arolla 3 and Schwarzhorn 4 , both of which exhibit marked convexity.

	Study site	*Index of slope form	*Index of departure from linearity	*Index of convexity/ concavity
Valley-wall rock glaciers				
	Arolla 1	1.49	0.17	96.35
	Arolla 2	0.91	0.05	0.22
	Arolla 3	0.55	0.10	0.02
	Zinal 1	0.84	0.06	0.22
	Zinal 2	1.13	0.14	1.22
	Flüela-Wisshorn 3	1.05	0.01	4.00
	Schwarzhorn 1	0.97	0.17	2.26
	Schwarzhorn 4	0.85	0.09	0.02
	Macun 1	0.94	0.23	0.89
	Macun 3	0.93	0.09	0.04
	Chilbiritzenspitz	0.93	0.17	0.11
	Radüner	0.98	0.12	0.32
	Gornergrat	14.80	-	-
	Fornesdalen West	1.81	0.18	8.08
	Fornesdalen East 1	1.49	0.20	8.08
	Fornesdalen East 2	2.03	0.25	53.36
	Rödbergdalen	1.21	0.11	2.81
	Trollvatnet	1.84	0.24	2.73
	Tverrelvdalen	1.86	0.16	3.30
	Strath Nethy	1.68	0.16	2.32
	Loch Etchachan	1.19	0.12	45.33
	Lairig Ghru	1.00	0.09	0.78
	Coire Beanaidh	1.34	0.16	9.09
	Baosbheinn	0.92	0.10	0.20
Valley-floor rock glaciers				
	Flüela-Wisshorn 1	2.05	0.23	99.20
	Tytebaerdalen	1.64	0.20	112.73
	Beinn an Lochain	1.15	0.08	15.40

Table 5.4 Slope form, index of departure from linearity and index of convexity/concavity for all study rock glaciers. (*Indices are defined in section 5.3)

Slope form indices and visual analysis of medial long profiles identified three morphologically 'anomalous' valley-wall rock glaciers that require some discussion. The most unusual long profile was of the rock glacier at Gornergrat (Figure 5.7) which also recorded the highest slope form value of 14.80. Its profile was so unusual that the indices of departure from linearity and convexity/concavity could not be calculated. Preliminary aerial photographic examination of Gornergrat revealed a lobate-shaped plan form typical of many valley-wall rock glaciers, and topographically the site was suitable for rock glacier development at the base of a north facing talus slope at 3000m in Valais, Switzerland (Plate 5.9). However, field examination immediately highlighted two pronounced morphological peculiarities. First, the feature has actually moved up a reverse basal slope (Plate 5.10), and second, there is a large depression at the base of the talus slope that is partly filled by meltwater. The drop in elevation from the crest of the frontal ridge to the surface of the lake that partially fills this depression is approximately 28m. The presence of internal ice is inferred from the occurrence of large surface ponds (Plate 5.11), but the average frontal slope gradient of 26.6° is less than might be expected for a rock glacier that still contains ice even if movement has ceased. However, it was noted that one section of the frontal slope appeared much steeper than 26°, perhaps reflecting localised activity. Interestingly, if the feature is still active, movement would be predominantly *backwards* towards the talus slope as basal and surface gradients are both steeper in that direction. In Plate 5.12, extremely large boulders can be seen along the frontal ridge crest whereas smaller clasts predominate over the rest of the rock glacier. Rock glacier

Plate 5.9 The valley-wall rock glacier at Gornergrat, which lies at an altitude of 3000m, south-east of Zermatt, Switzerland. Plan view shows typical lobate-shaped feature at the base of a talus slope. Extent of landform corresponds with lighter grey clasts.

Plate 5.10 Lateral view of Gornergrat illustrating the reverse basal slope over which the landform has moved.

Plate 5.11 View down on to Gornergrat rock glacier. From the lake at the base of the talus slope to the frontal ridge crest, elevation increases by ~28m.. The occurrence of surface ponds strongly suggests the presence of internal ice.

Plate 5.12 Frontal margin of Gornergrat rock glacier. Extremely large clasts in excess of 35m^3 occur on the frontal ridge. Much smaller clasts predominate over the rest of the rock glacier surface between the frontal ridge and the talus.

thickness at the base of the talus is apparently almost negligible but increases towards the frontal slope where a maximum thickness of 10m is reached. Clearly, the morphological characteristics of Gornergrat differ from those of other valley-wall rock glaciers studied and so the morphological data relating to Gornergrat (Table 5.1) were removed from the statistical analyses presented in section 5.4.5.

One possible explanation for the formation of the Gornergrat feature is that a small steeply sloping glacier, or a perennial snowbed developed at this site over which a single massive rockfall or a series of large-scale rockfall events cascaded, to be dumped at the base of the steep ice surface. Such an explanation may explain how anomalously large clasts in excess of $35m^3$ travelled over 180m up a reverse slope. Further support for this explanation is provided by the observed outward increase in thickness as the majority of clasts would have been transported over the ice surface with only a few deposited close to the talus slope following ice melt. Finally, additional support for this conclusion comes from the presence of well developed lateral ridges at Gornergrat, which may in fact be lateral moraines. Although the origin of Gornergrat is debatable, it is highly improbable that it was transported to its present position by the slow downward deformation of internal ice, and so it should not be classified as a true valley-wall rock glacier.

At Macun 1, an inactive valley-wall rock glacier in Graubünden, Switzerland, there is a marked depression in the surface of the rock glacier at the foot of the talus slope, as at Gornergrat, although the rock glacier has not moved up a reverse basal slope (Figure 5.7; Plate 5.13).

Plate 5.13 Lateral view of Macun 1 valley-wall rock glacier. Upslope of the frontal ridge, the rock glacier trends downwards towards the base of the talus where a lake has formed, as at Gornergrat (figures for scale can be seen on this reverse slope). The unusually large frontal slope partly reflects steeply sloping basal topography.

This depression was partly filled by meltwater when the feature was visited in July 1987. Similar negative gradients on the upper surfaces of valley-wall rock glaciers have been reported elsewhere (eg. Calkin *et. al.*, 1987), and are believed to be the result of subsidence due to the melting of unusually large amounts of buried ice, of either glacial or periglacial origin. It should be noted that internal ice may persist within inactive rock glaciers despite the presence of such depressions.

The active valley-wall rock glacier Arolla 3, which lies at an altitude of 3180m in the Valais region of Switzerland, recorded the lowest slope form value (0.55) and the most convex slope (0.02) of the study sample. The reason for the unusually convex form of this rock glacier lies in an extremely (84m) high frontal slope (Figure 5.6; Arolla 3, transect 2). Field examination, however, revealed a pronounced rock bench located adjacent to the rock glacier that is thought to continue underneath the frontal slope of the rock glacier. If debris had not cascaded over this rock bench, it seems reasonable to assume that the true height of the rock glacier would be much less than 84m. The frontal slope at Arolla 3 is discussed in detail in section 5.5.2 below and is illustrated in Plate 5.16.

5.4.4 Rock glacier thickness and volume

Long profiles were surveyed, where possible adjacent and parallel to each valley-wall rock glacier, to provide an estimate of rock glacier thickness. If such surveying was impossible because of a change in orientation or form of the adjacent talus slope, thickness estimates were derived by assuming a regular decline in slope gradient underneath the rock glacier. Estimates of maximum rock glacier thickness are summarised in Tables

5.1 to 5.3 and range from 10 to 35.5m. Usually the estimated thickness is greatest beneath frontal ridge crests and decreases gradually upslope until rock glacier and talus merge. This general pattern may be seen on long profile figures where surveyed estimates of thickness are depicted (Figures 5.5 to 5.8). Rock glacier thickness data are used in calculating the theoretical boundary conditions for rock glacier formation in Chapter 7.

Using a combination of long profile data, together with thickness and width measurements, it was possible to calculate an approximate estimate of rock glacier volume. In order to identify the range of rock glacier volumes that occur within the study sample, the volume of the largest and smallest individual valley-wall rock glaciers was calculated. The method used to calculate rock glacier volume is explained in detail in Figure 5.9 with reference to Trollvatnet, an inactive, possibly relict, rock glacier on the Lyngen Peninsula, Norway, which is the largest individual rock glacier in the study sample. Ground and aerial photographic views of the Trollvatnet rock glacier are shown in Plates 5.14 and 5.15. The first stage of the calculation involves constructing a surveyed scale plan using a combination of theodolite and abney level transects (Plot A). The scale plan is required in order to determine precise width measurements for any point on the medial long profile. Lateral and frontal slope gradients are depicted on Plot A together with the position of transverse ridges and closed depressions. Long profile and thickness profiles of Trollvatnet are combined and illustrated in Plot B so that thickness measurements may be obtained for any position on the rock glacier. (In Figure 5.3 it was shown that rock glacier thickness is relatively constant across the width of Trollvatnet for any position on its

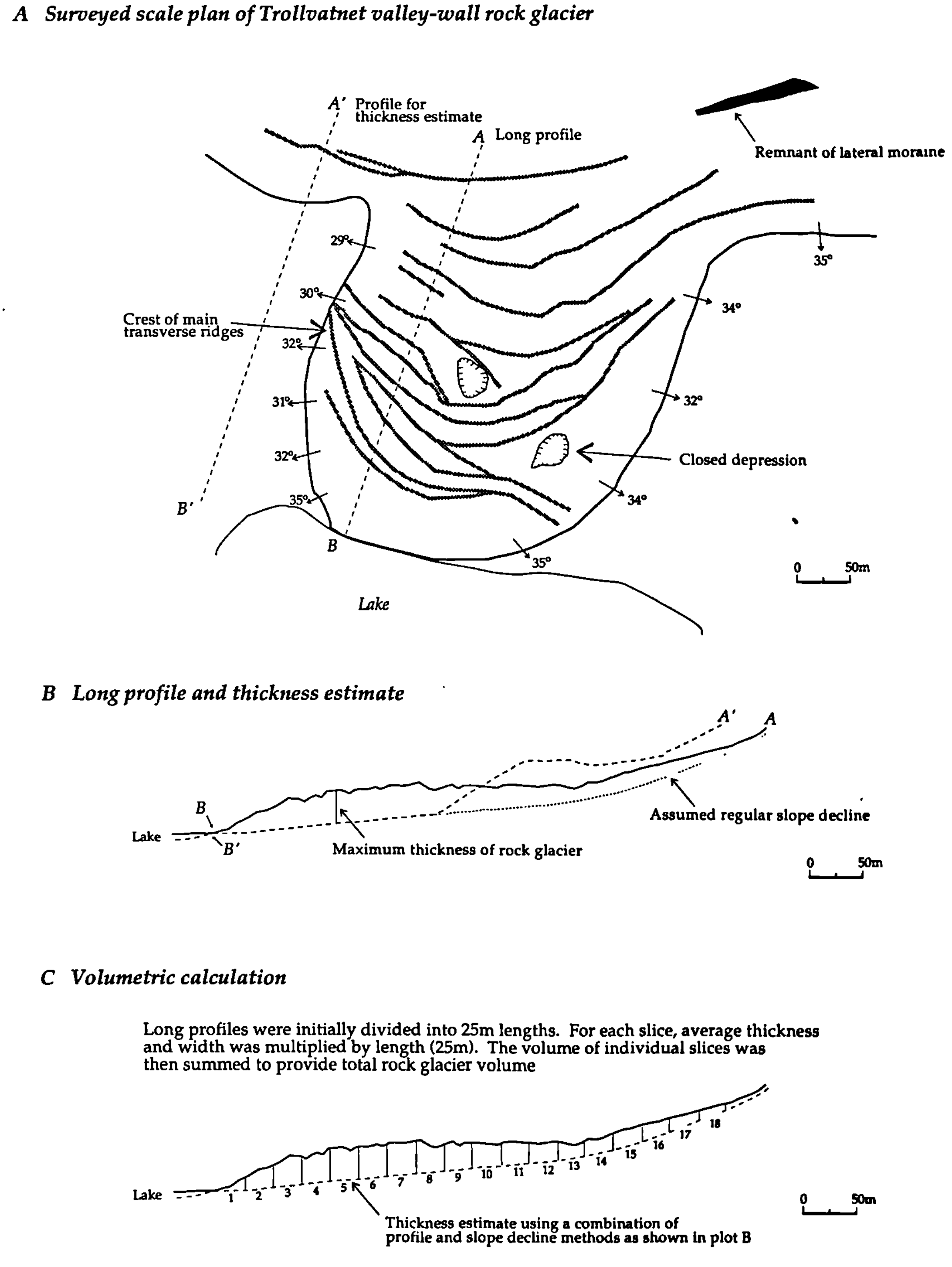


Figure 5.9 Trollvatnet valley-wall rock glacier showing surveyed scale plan, long and depth profiles, and volumetric calculation.

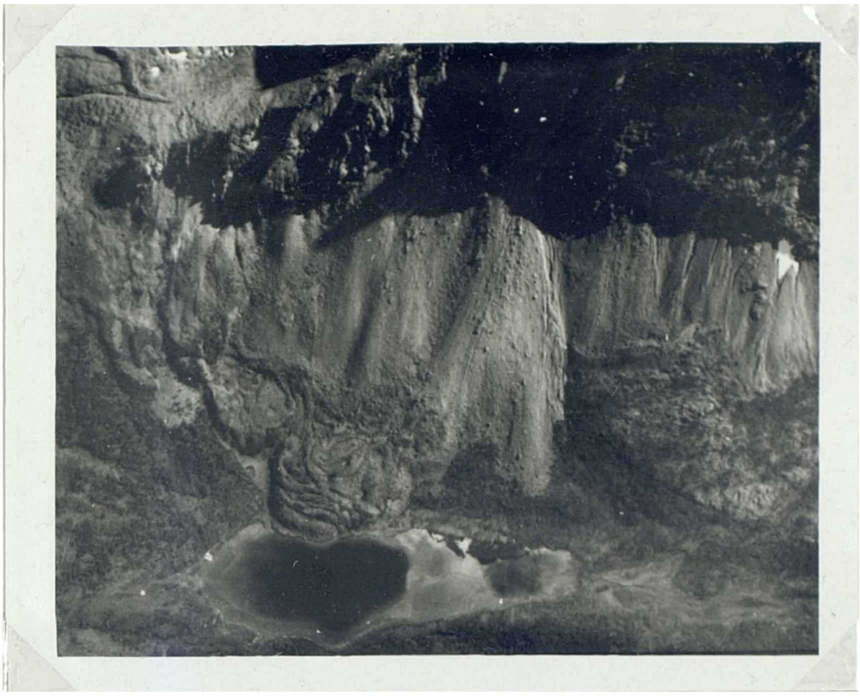


Plate 5.14 Aerial photographic view of Trollvatnet valley-wall rock glacier. Vertical view displays well-developed, but discontinuous, transverse ridges.

Plate 5.15 View of Trollvatnet from talus upslope showing vegetation growth on the dark weathered gabbroic mass of the rock glacier. This rock glacier, which is an inactive or relict feature, terminates in a lake.

medial long profile.) In Plot C, rock glacier length is divided into eighteen 25m slices so that variations in thickness down the length of the rock glacier may be accounted for in volumetric calculations (Plot C).

Using the method outlined above, the total volume of Trollvatnet was calculated to be *ca.* 2.11 x 10^6m^3. In contrast, the volume of the smallest valley-wall rock glacier in the study sample, Arolla 2 (Plate 5.4), an inactive rock glacier in Valais, Switzerland was estimated at *ca.* 2.26 x 10^4m^3, almost 100 times smaller than Trollvatnet. Given that the upper few metres of most rock glaciers comprise an openwork boulder layer beneath which finer material predominates and forms the majority of the rock glacier volume, it is assumed that void space occupies between 20% and 25% of the total volume. When suitable adjustments for void space are made the equivalent volumes of rock are between *ca.* 1.58 x 10^6m^3 and *ca.* 1.69 x 10^6m^3 for the Trollvatnet rock glacier and between *ca.* 1.69 x 10^4m^3 and *ca.* 1.81 x 10^4m^3 for the Arolla 2 rock glacier. Barsch (1977b) estimated that on average every individual active rock glacier transports a volume of between 1.2 and 1.6 x 10^6m^3 of talus and ice. Trollvatnet lies just outside the upper end of this average range. The rock glacier at Trollvatnet, however, is inactive and ice content if any is minimal, which suggests that volume when active was significantly greater than at present. The large volume of material comprising Trollvatnet rock glacier emphasises the importance of valley-wall rock glaciers as agents of mass-wasting.

The area of contributing free faces above the rock glacier at Trollvatnet is *ca.* 180,000m^2, which implies an extremely large average rockwall retreat of *ca.* 11.7m during the period of rock glacier movement. Similarly, the area of contributing free faces at Arolla 2 is *ca.* 4000m^2, which implies a lower, yet still impressive rockwall retreat of *ca.* 4.7m. It is possible that an unknown volume of talus that existed prior to rock glacier formation may have become incorporated within the rock glaciers during formation; if so, rockwall retreat figures presented above may overestimate true rockwall retreat amounts. As volumetric calculations are both time-consuming and somewhat fruitless unless the duration of rock glacier activity is known, so that rockwall retreat rates can be estimated, it was decided not to calculate the volume of each rock glacier in the data set. In Chapter 7, however, further consideration is given to the volume of the Scottish rock glaciers, whose profiles are shown in Figure 5.8. An attempt is made to compare possible rates of rockwall retreat with previously calculated rates for the Loch Lomond Stadial in order to infer information on the processes and rates of rock glacier formation.

5.4.5 Morphometric relationships

Despite considerable variability in the form of the rock glacier profiles, bivariate statistical analyses reveal quite strong morphometric regularities. Of the twenty-four valley-wall rock glaciers chosen for detailed field study, twenty-two are included in the following statistical analyses. As mentioned above, Gornergrat rock glacier was excluded because of its highly unusual morphometric characteristics. In addition, the rock glacier Arolla 3 was removed because of its peculiar frontal slope, which apparently is due to underlying topographic control and not directly to the formation process. Figure 5.10 presents the results of

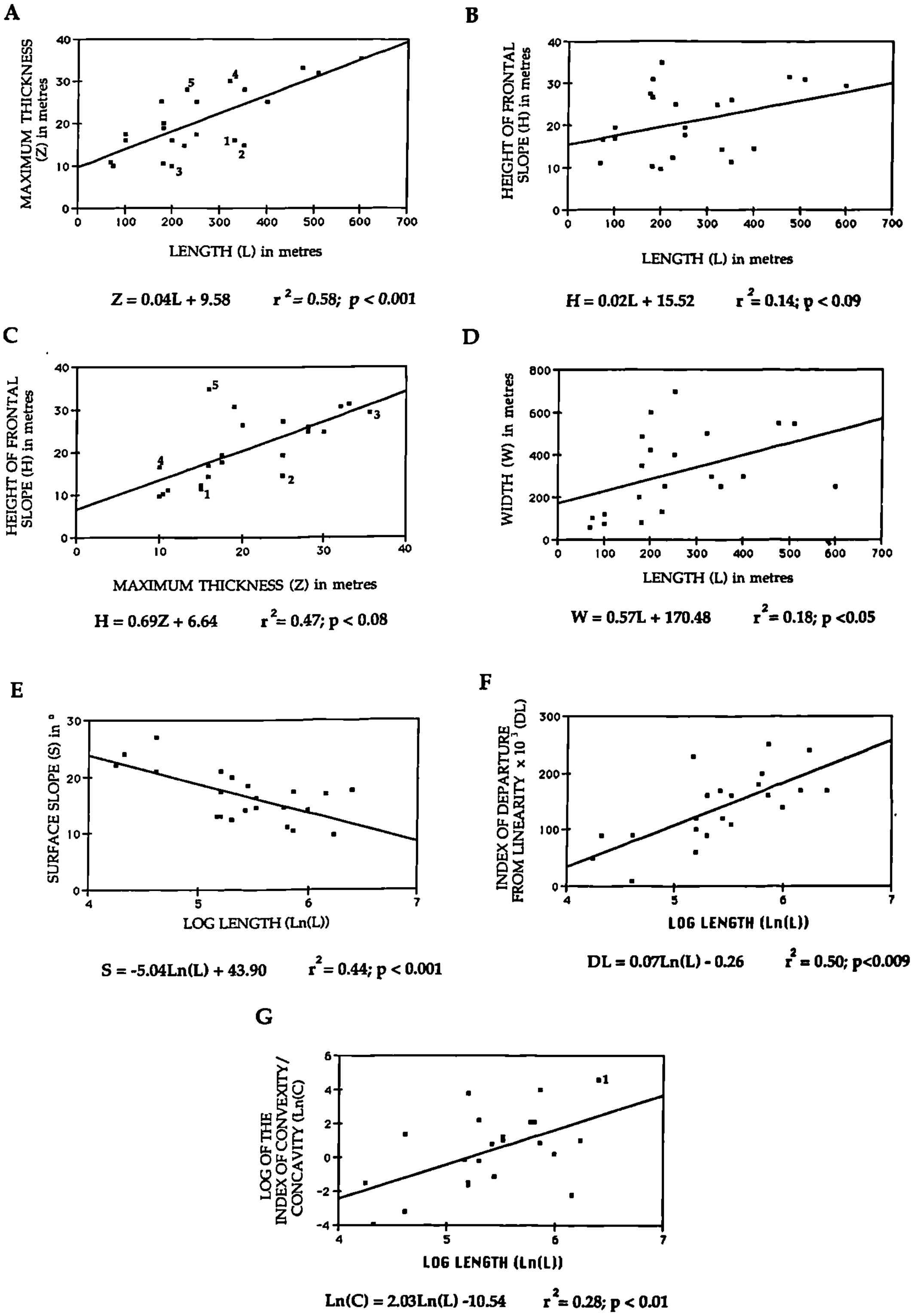


Figure 5.10 Morphometric relationships for valley-wall rock glaciers

simple linear regression analyses. Plot A shows that a good positive correlation exists between rock glacier length and maximum depth. A similar correlation was observed by Calkin et al. (1987) for valley-wall rock glaciers in the central Brooks Range, Alaska. As length/depth ratios appear to be similar for active, inactive and relict valley-wall rock glaciers, depth must remain a relatively constant fraction of rock glacier length from initial formation to deactivation. This finding has important implications for the threshold characteristics of rock glacier formation and development, which will be analysed in detail in Chapter 7. The thickest rock glacier in the study sample, Arolla 1, reaches a maximum depth of 35.5m, although depths up to 60m have been reported elsewhere (cf. Wahrhaftig & Cox, 1959; Calkin et al., 1987). In Plot A, it is interesting to note that the regression line passes well above the origin, which suggests that a depth threshold of debris is required before rock glacier movement can occur.

The majority of rock glaciers in which thickness is statistically less than expected have developed on top of till deposits, as at Fornesdalen East (labelled '1' and '2' on Plot A, Figure 5.10) and Coire Beanaidh, Cairngorms ('3'). A possible explanation for this is that both till and rock glacier may have deformed simultaneously; clearly, however, such secondary deformation is impossible for rock glaciers forming on bedrock. Boulton and Jones (1979) noted that glaciers that rest on beds of deformable sediment often exhibit a lower equilibrium profile than expected. They believe that in such conditions a large proportion of forward movement may be due to deformation of the bed rather than of the glacier, and suggest that if bed transmissibility is low, water pressures may build up to a threshold point at which the bed itself begins to deform. It seems reasonable to assume that deformation of unconsolidated sediments beneath rock glaciers, such as the till deposits in Fornesdalen, may contribute to rock glacier movement. In contrast, for those rock glaciers that are slightly thicker than expected, topographic control has apparently restricted further rock glacier movement downslope. For example, at Fornesdalen West ('4' on Plot A) and at Radüner ('5'), the rock glaciers have terminated abruptly against moraine ridges.

A much poorer correlation exists between rock glacier length and frontal slope height (Plot B), suggesting that the height of the frontal slope should not be used as a surrogate for maximum thickness as it has been in previous studies (eg. Wahrhaftig & Cox, 1959). Plot C, which tests directly the relationship between maximum depth and frontal slope height, shows that over 50% of the variance in the face height values must be explained by variables other than maximum thickness. If rock glaciers override till deposits such as at Fornesdalen East (labelled '1' on Plot C), maximum thickness apparently occurs upslope of the frontal ridge. In addition, if large inner transverse ridges occur where the basal gradient beneath the rock glacier is still relatively small, the height of the frontal slope will also underestimate the maximum depth of the rock glacier. For example at Zinal 2 ('2') and at Arolla 1 ('3'), the maximum thickness of the rock glacier occurs beneath inner transverse ridges upslope of the frontal margin. Alternatively, if the basal gradient beneath rock glaciers is relatively steep, for example at the rock glaciers at Schwarzhorn 4 ('4') and Lairig Ghru ('5'), rock glacier thickness is often less than the height of the frontal slope.

In Plot D, the relationship between length and width is explored for the sample of valley-wall rock glaciers. As discussed earlier, classification of rock glaciers into valley-floor and valley-wall forms, on the basis of their length to width ratio, is misleading. Interestingly, the regression equation provides supporting evidence for this claim. For valley-wall rock glaciers smaller than approximately 400m in length, the regression equation predicts that width will tend to exceed length. However, for valley-wall rock glaciers greater than 400m in length, the equation indicates that width will tend to be less than length. This evidence supports the conclusion that the length/width classification scheme should not be used. In this plot, there is no logical independent variable as there appears to be no *a priori* reason for supposing that length or width should explain the variance in the other parameter. Clearly, the lateral extent of the contributing rockwall and the topographic characteristics of the site are likely to play an important role in determining rock glacier width, whereas rock glacier length is likely to have been determined largely by the availability of debris at the site, which is partly dependent on the size of the contributing rockwall, and possibly, by basal topography.

In the fifth of the regression plots shown in Figure 5.10, mean surface slope gradient above the frontal ridge crest is plotted against the natural logarithm of rock glacier length (Plot E). A distinct negative relationship with a correlation coefficient of 0.44 may be observed, such that surface slope gradients tend to be much greater for small valley-wall rock glaciers than for larger ones. This pattern is linked to the form of the rock glacier and, in particular, to the length of the mid section of the rock glacier directly upslope of the frontal ridge where inner transverse ridges occur. Many of the smaller steep rock glaciers possess at most one inner transverse ridge, whereas larger ones often exhibit several transverse ridges most of which exhibit negative proximal slope gradients that reduce overall surface slope gradient.

The sixth regression analysis investigates the statistical relationship between rock glacier length and the geometry of the rock glacier slope (section 5.4.3), by plotting the index of departure from linearity against the natural log of rock glacier length (cf. section 5.3 and Figure 5.2 above; index values increase as the slope departs from linearity). The plot clearly shows that as length increases, rock glaciers tend to become less rectilinear in form. This finding, when interpreted in conjunction with Plot G, in which the natural log of the index of convexity/concavity is plotted against log length, indicates that where rock glaciers are of limited length, the steep generally convex frontal slope forms a greater proportion of the total length of the rock glacier, thereby resulting in a slightly overall convex form (convexity/concavity index values increase as concavity increases). As length increases, rock glaciers generally become progressively convexo-concave as the convexity of frontal slope is matched by the concavity of the upper section of the rock glacier. With further increases in rock glacier length, the frontal slope gradually decreases in relative importance and overall slope form tends to becomes more concave; for example, the concave rock glacier Arolla 1 labelled '1' in Plot G, which is 600m long.

The relationship between slope form and rock glacier length may be viewed from a developmental angle in which the typically convex slope form of an incipient rock glacier becomes progressively convexo-concave, then predominantly concave if rock glacier length increases sufficiently before deactivation. It must be emphasised, however, that of the rock glaciers studied many appear to have become inactive while rock glacier length was still very limited, either as a result of a reduction in debris supply or due to climatic amelioration. Thus, small rock glaciers may either be active, inactive or relict; they are not necessarily all active features in the initial stages of formation.

5.4.6 Summary

The research reported in this section can be summarised in five major points:

1) Preliminary analysis of the dimensional data (Tables 5.1 to 5.3) and visual inspection of the long profiles (Figures 5.5 to 5.8) suggested considerable variability in rock glacier form. However, some fairly strong morphometric regularities were detected by bivariate statistical analyses. In particular, a strong positive relationship was observed between rock glacier length and maximum thickness. In addition, the geometry (convexity/concavity) of the rock glacier slope is related to rock glacier length.

2) Despite these morphometric regularities, the occurrence of 'anomalous' characteristics in several of the valley-wall rock glaciers studied emphasises that definition and classification of rock glaciers by morphology alone is problematical.

3) The form of each valley-wall rock glacier comprises three parts: a convex frontal ridge with average gradients of between 23° and 41°, an approximately rectilinear section of variable length with gentler gradients of between 6° and 20°, and an uppermost concave section, in which average gradients increase until the rock glacier merges with the talus above.

4) Valley-wall rock glaciers are not always wider than they are long (cf. Wahrhaftig and Cox, 1959), and those within the study sample that extend for over 600m along a valley-wall should not be considered as one rock glacier; instead they should be viewed as a series of contiguous individual valley-wall rock glaciers.

5) Assuming that void space occupies 20% of the total volume of a rock glacier, the volume of rock within the study rock glaciers ranges from $1.81 \times 10^4 m^3$ for the smallest rock glacier at Arolla 2 to $1.69 \times 10^6 m^3$ for the largest rock glacier at Trollvatnet. Measurements of rock glacier thickness and volume have important implications in terms of defining boundary conditions for the formation of valley-wall rock glaciers and as such will be considered further in Chapter 7.

5.5 *Surface microrelief*

5.5.1 *Introduction*

Differential melting of ground ice (including unequal distribution of ice prior to melt) and rock glacier movement are thought to be the two main processes responsible for the formation of a variety of surface microrelief features common to many valley-wall rock glaciers (cf. Wahrhaftig & Cox, 1959; Potter, 1972; Corte, 1976; Haeberli, 1985; section 2.4.1 above). Some small valley-wall rock glaciers, however, are almost devoid of such features. The aim of this section is to describe in detail the characteristics of each type of surface microrelief feature and to establish if any common morphological characteristics occur. Every valley-wall rock glacier in the study sample possesses a frontal ridge and most possess at least one inner transverse ridge. In addition to an examination of the characteristics of frontal and transverse ridges, the longitudinal and closed depressions and surface ponds that occur on several rock glaciers are described. Surface microrelief is most easily studied on active rock glaciers as post-formational modification by rockfall avalanches, snow avalanches or debris flow activity, may obscure part or all of the surface features on inactive or relict rock glaciers. However, it is useful to examine active, inactive and relict rock glaciers to establish if differences in surface microrelief characteristics are affected by deactivation.

5.5.2 *Frontal ridge morphology*

Without exception, the frontal ridges of all valley-wall rock glaciers studied are well-defined. Detailed examination of a large number of these margins led to the identification of four characteristic units based on morphological and sedimentological evidence. Rock glacier sedimentology, which is examined in detail in Chapter 6, is briefly considered here in order to define more precisely the characteristics of the frontal slope units. The four unit subdivision (Figure 5.11) also forms a basis from which distinctions may be drawn between active and inactive/relict rock glaciers. As outlined in section 4.2 above, the frontal slopes of active rock glaciers tend to be characterised by steeper gradients, greater instability, a higher percentage of fines, and an absence of vegetation and lichens. However, it must be emphasised that in some cases such distinctions are very finely drawn. In arid or polar climates, vegetation growth may be severely restricted, so rock glaciers that have been inactive for several years may remain free of vegetation or lichens. Moreover, in areas where snowcover extends beyond nine months of the year vegetation growth may be similarly inhibited. In addition, relatively steep slopes can be maintained for longer periods in polar climates than in more temperate climates (Shaw and Healy, 1977). Whilst recognising that differentiation between active and inactive rock glaciers remains problematical, the morpho-sedimentary units are as follows.

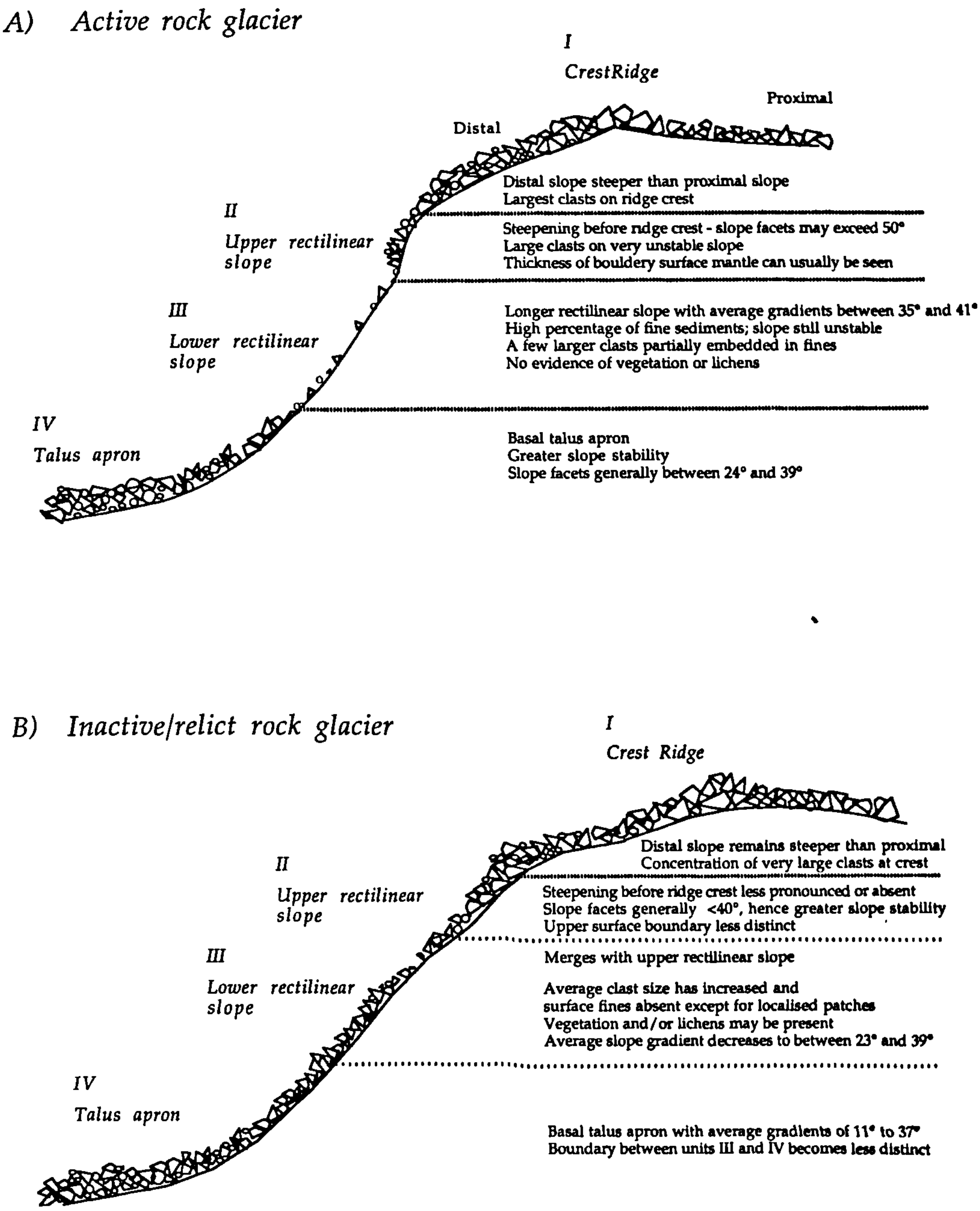


Figure 5.11 Schematic diagram showing frontal slope morphological units for both active and inactive/relict valley-wall rock glaciers.

I. Frontal Ridge Crest

Crest ridges generally comprise a steep distal and gentle proximal slope that usually join at a sharp crest on active features and a slightly more subdued crest on inactive and relict rock glaciers. On some rock glaciers, however, proximal slopes do not occur; instead, the crest levels at the top of the distal slope and then rises again forming a benched crest. This is particularly common on small rock glaciers where inner transverse ridge development is limited or absent. On active rock glaciers, average gradients for distal slopes range from 35° to 47° and for proximal slopes from 5° to 10°. Similar measurements made on inactive and relict rock glaciers produced a range of values between 22° and 33.5° for distal slopes and 6° and 22° for proximal slopes. Crest ridges are also characterised by a concentration of very large clasts that form an openwork mixture of boulders (see Plate 5.4). Sorting of clasts on proximal slopes is generally very poor.

II. Upper rectilinear slope

This unit comprises an approximately rectilinear slope downslope of the crest ridge and contains the steepest gradients on any part of a rock glacier. On active features, the slope is extremely unstable and ranges from 43° to 50° in average gradient with individual slope facets of up to 53°. Unit length, which varies between 3 and 7.5m, provides an approximate estimate of the thickness of the bouldery surface mantle. The downslope boundary of the upper rectilinear unit on active rock glaciers is often a visible break between large clasts of unit II and much finer deposits at the top of unit III. On unit II, surface fines are almost totally absent and clast size varies from small to large in an openwork boulder framework. On inactive and relict rock glaciers, the upper rectilinear slope is characterised by average gradients of between 29° and 40°, with individual slope facets as high as 45°.

III. Lower rectilinear slope

On active, inactive and relict rock glaciers, gradients on the lower rectilinear slope are generally less steep than on unit II. On average, gradients range from 35° to 41° for active features and from 23° to 39° for inactive and relict rock glaciers. On active rock glaciers, this unit is underlain almost exclusively by fine deposits that are thought to constitute most of the rock glacier volume. The majority of large clasts that become dislodged from the upper rectilinear slope roll or slide down the entire length of the relatively smooth face to unit IV, although a few are embedded in the fine deposits of unit III. On inactive rock glaciers, extensive surface fines do not occur but may be present at shallow depth and clasts of all sizes occur on the surface. The boundary between units II and III was not detected on relict valley-wall rock glaciers in Scotland nor on some inactive rock glaciers in Norway.

IV. Talus apron

Unit IV comprises fall-sorted clasts that have accumulated mainly by rolling and sliding down the frontal slope. Slope angles on active rock glaciers range from 24.5° to 39° and on inactive and relict rock glaciers from 11° to 37°. The height of unit IV ranges between 15 and 40% of the total height of the frontal slope. The boundary between units III and IV on inactive and relict rock glaciers is less distinct as surface fines are largely absent from the lower rectilinear slope. As mentioned in section 2.6 above, talus aprons maintain a relatively constant height of the frontal slope throughout their formation, which suggests that clastic debris probably becomes reincorporated within the moving rock glacier.

Morphological differences between the frontal slopes of active and inactive rock glaciers may be accounted for mainly in terms of the presence or absence of differential forward movement. It is widely believed (e.g. Wahrhaftig & Cox, 1959; Haeberli, 1985; White, 1987; cf. section 2.6 above), that the upper layers of active rock glaciers move downslope at a greater velocity than the rest of the feature producing very high slope gradients in unit II and basal talus aprons. Over a six hour period spent working on the active frontal slope at Arolla 1, seven large clasts rolled down from the upper rectilinear slope towards the base of the slope. The tracks of several other clasts could be detected on late-lying snow. Several smaller slope failures on the lower rectilinear slope comprising mostly fine material slipped downslope during this period. Such movements do not include several that were directly caused by traverses of the frontal slope during surveying and sediment collection. A few larger clasts that were embedded in the fine deposits of the lower rectilinear slope seemed to act as temporary storage traps for other clasts. However, the occurrence of failure scars on the fine deposits and the observation that such *storage traps* are limited in size suggests that a threshold size is reached at which slope failure occurs. On deactivation, steep slope gradients are not maintained and lower average slope angles are a response to slumping. Melt of internal ice may also decrease average slope gradient. During and following deactivation, clasts that become dislodged from the top of the slope tend to become wedged in the interstices of the other clasts and fewer of them reach the basal talus. Thus the surface of the lower rectilinear slope supports more large clasts on inactive and relict rock glaciers than is the case for active rock glaciers. As mentioned above, on some inactive and relict rock glaciers it was impossible to detect any increase in gradient from unit III to unit II.

The principles outlined in the above discussion can be illustrated by reference to a selection of study rock glaciers. Plate 5.16 is a photograph of the active frontal slope at Arolla 3. Although it is unusually high because of an underlying rock bench, the frontal slope exhibits many of the characteristics outlined in Figure 5.11. The predominance of fines underlying the lower rectilinear slope contrasts with the cover of large clasts on the steep short upper rectilinear slope. There is no evidence of vegetation and clasts do not support lichens. Late-lying snow obscures the extent of the basal talus apron. In Plate 5.17, which shows the active frontal margin of a valley-floor rock glacier at Férpecle, the fall-sorted talus apron is clearly visible. Here, the boundary between the upper and lower rectilinear slopes is very distinct and the thickness of the thin bouldery surface mantle can be seen.

In contrast to the two previous photographs, inactive (Macun 3) and relict (Lairig Ghru) frontal margins are illustrated in Plates 5.18 and 5.19,

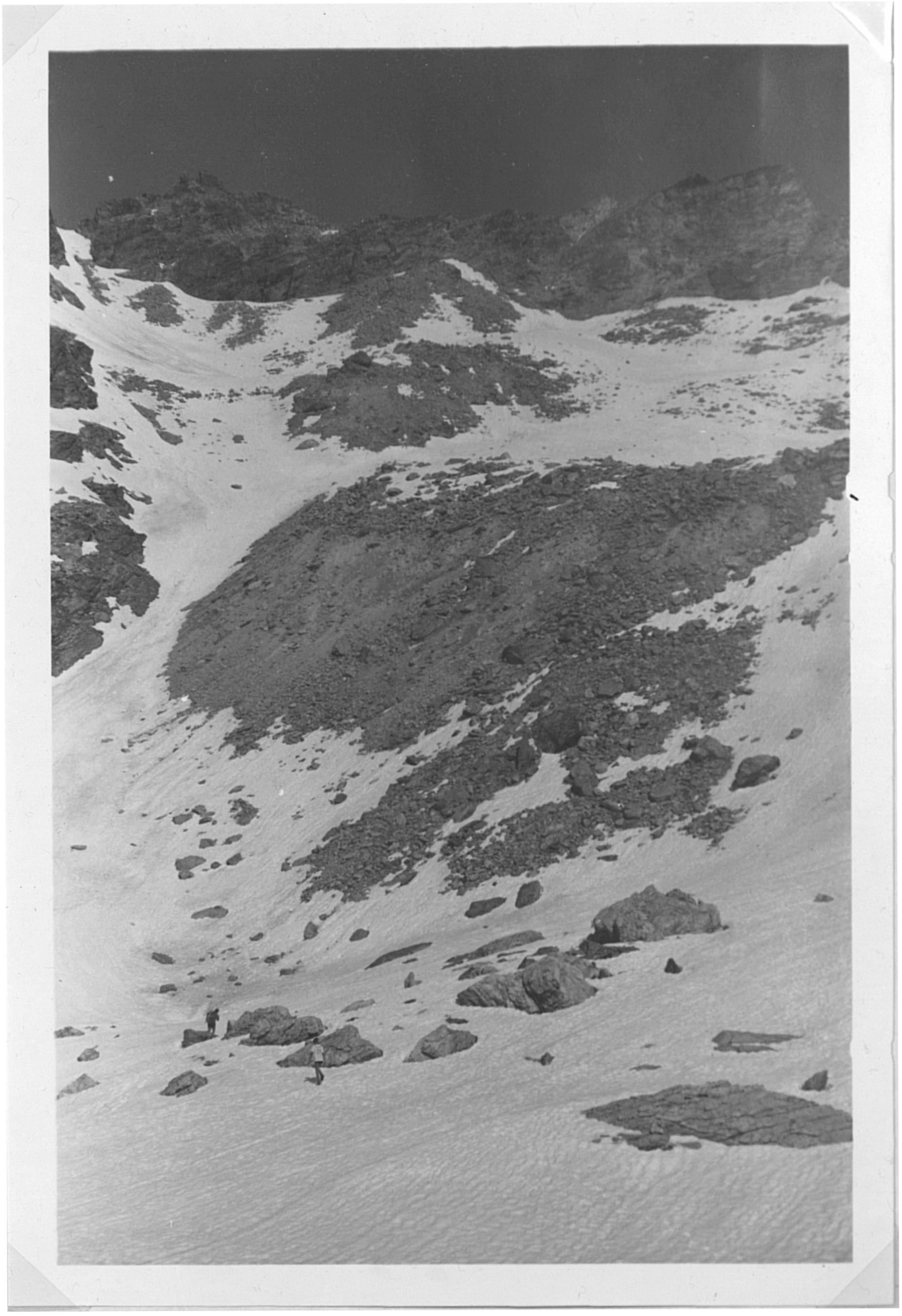


Plate 5.16 Frontal slope of the active valley-wall rock glacier at Arolla 3 showing a predominance of fine material at the surface. Individual slope facets exceed 50° and the slope is very unstable. Vegetation and lichens are absent. The unusually high 85m frontal slope is due to a rock bench over which the rock glacier has moved. Figures in foreground for scale.

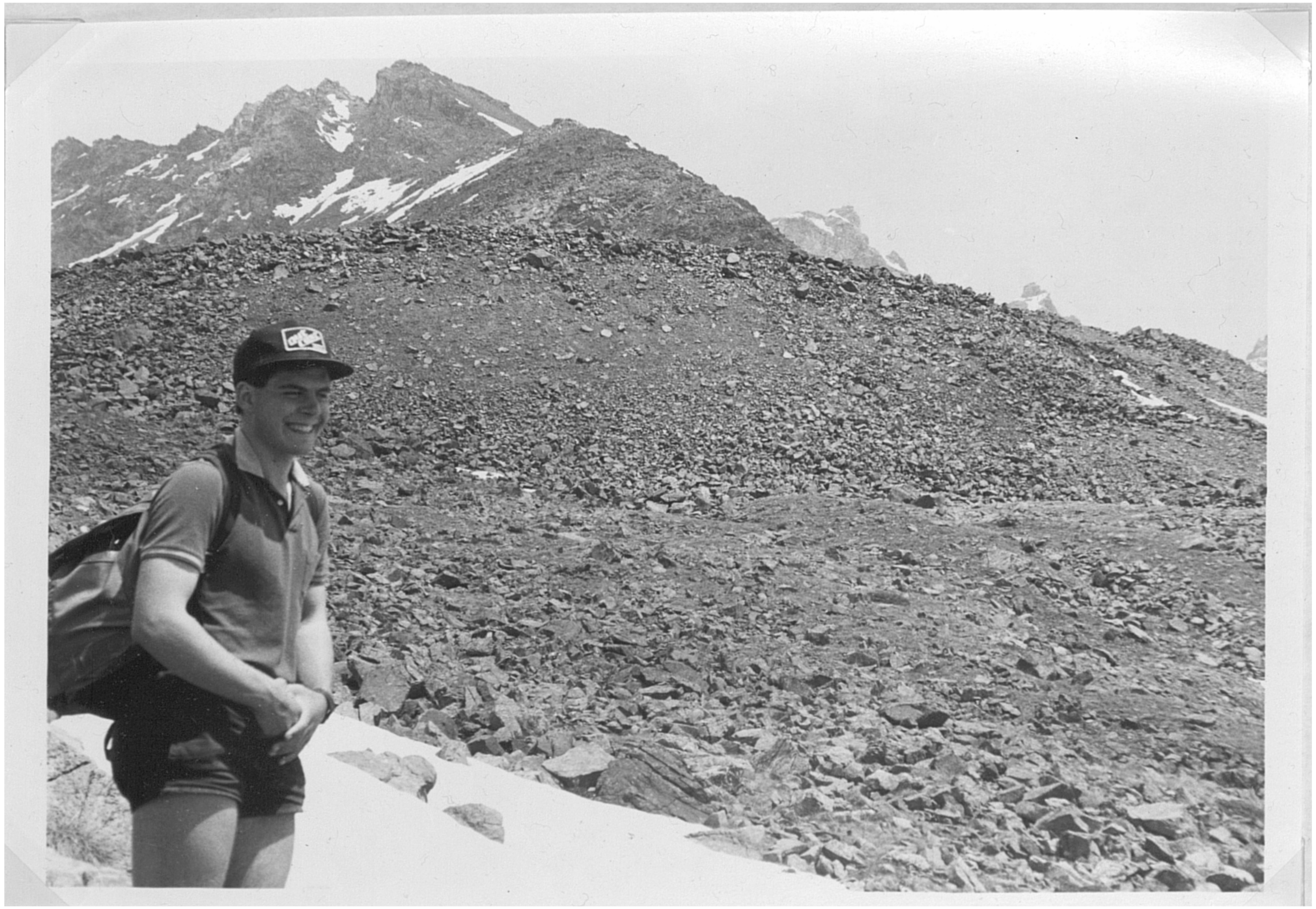


Plate 5.17 Frontal slope morphology of an active valley-floor rock glacier at Fèrpecle, Switzerland, showing three of the four morphological units outlined in Figure 5.11. Below the ridge crest (which is out of view), the bouldery upper rectilinear slope merges downslope with the less steeply sloping lower rectilinear slope which is underlain by predominantly fine material. At the base of the frontal slope, the fall-sorted basal talus apron can be seen.

Plate 5.18 Lateral view of Macun 3, a small inactive valley-wall rock glacier in Graubünden, Switzerland. Large clasts occur down the entire length of the frontal slope. In the background, much finer deposits can be seen on the frontal slope of an active valley-floor rock glacier.

Plate 5.19 A relict valley-wall rock glacier in the Lairig Ghru, Cairngorms. Extensive vegetation occurs on most of the rock glacier and on the relict talus slope. The boundaries between units II, III, and IV are obscured, although parts of a basal talus apron remain.

respectively. At Macun 3, surface fines are absent and larger clasts occur down the entire length of the face. In the background of Plate 5.18, the frontal margin of an active valley-floor rock glacier provides a visual contrast to Macun 3; on the active feature, which has a 'fresher' appearance, surface fines are clearly evident and the frontal slope is steeper. The relict rock glacier in the Lairig Ghru supports extensive vegetation and individual clasts are extensively covered by lichens.

The effect of late-lying snow on valley-wall rock glacier development has rarely been considered. At the beginning of the 1987 Swiss field season, towards the end of June, the occurrence of extremely extensive patches of late-lying snow, which obscured many of the study rock glaciers, emphasised the probable widespread nature of prolonged snowcover. Some of the most pronounced effects of extended snowcover may be seen on inactive rock glaciers. For example, late-lying snow inhibits and in some cases prevents the development of vegetation and lichen growth. At a valley-floor rock glacier in the Macun valley in Switzerland, lichen development was quite pronounced towards the top of the frontal slope but decreased progressively in extent towards the base of the slope.

5.5.3 Transverse ridges and longitudinal depressions

Of the twenty-four valley-wall rock glaciers in the study sample all but five contain at least one inner transverse ridge, with a maximum of thirteen ridges occurring at Trollvatnet rock glacier (Plate 5.14).

Transverse ridges, which are generally absent on rock glaciers less than 100m long, tend to increase in number for progressively larger rock glaciers, although some large rock glaciers, such as Rödbergdalen and Arolla 1, contain only one and two inner transverse ridges respectively. Generally, however, although more numerous on larger rock glaciers, transverse ridges tend to become less continuous as they frequently merge, bifurcate or die out in closed depressions. In contrast, on smaller rock glaciers transverse ridges often extend across the width of the rock glacier as continuous features. All ridges, whether continuous or not, are convex downslope in planform, which suggests that their formation may be related to rock glacier movement. Typically, they are located immediately upslope of the frontal ridge, although where they are numerous they extend up to 350m upslope of the frontal ridge. No apparent relationship exists between ridge size and position on the rock glacier, as large and small ridges are intermixed and size does not increase or decrease in any preferred direction.

Marked cross-sectional asymmetry was observed on all transverse ridges of active, inactive and relict rock glaciers. Average slope gradients on distal slopes, which ranged from 9° to 37°, proved consistently steeper by between 3° and 20° than average gradients on the corresponding proximal slopes, which all fall within a range of values from 1° to 23°. Crests of transverse ridges, even on active features, are generally rounded and not as sharp as frontal ridge crests, which perhaps reflects a different mode of origin. The height difference between ridge crest and adjacent depression ranges from 1 to 20m. Figure 5.12 shows two

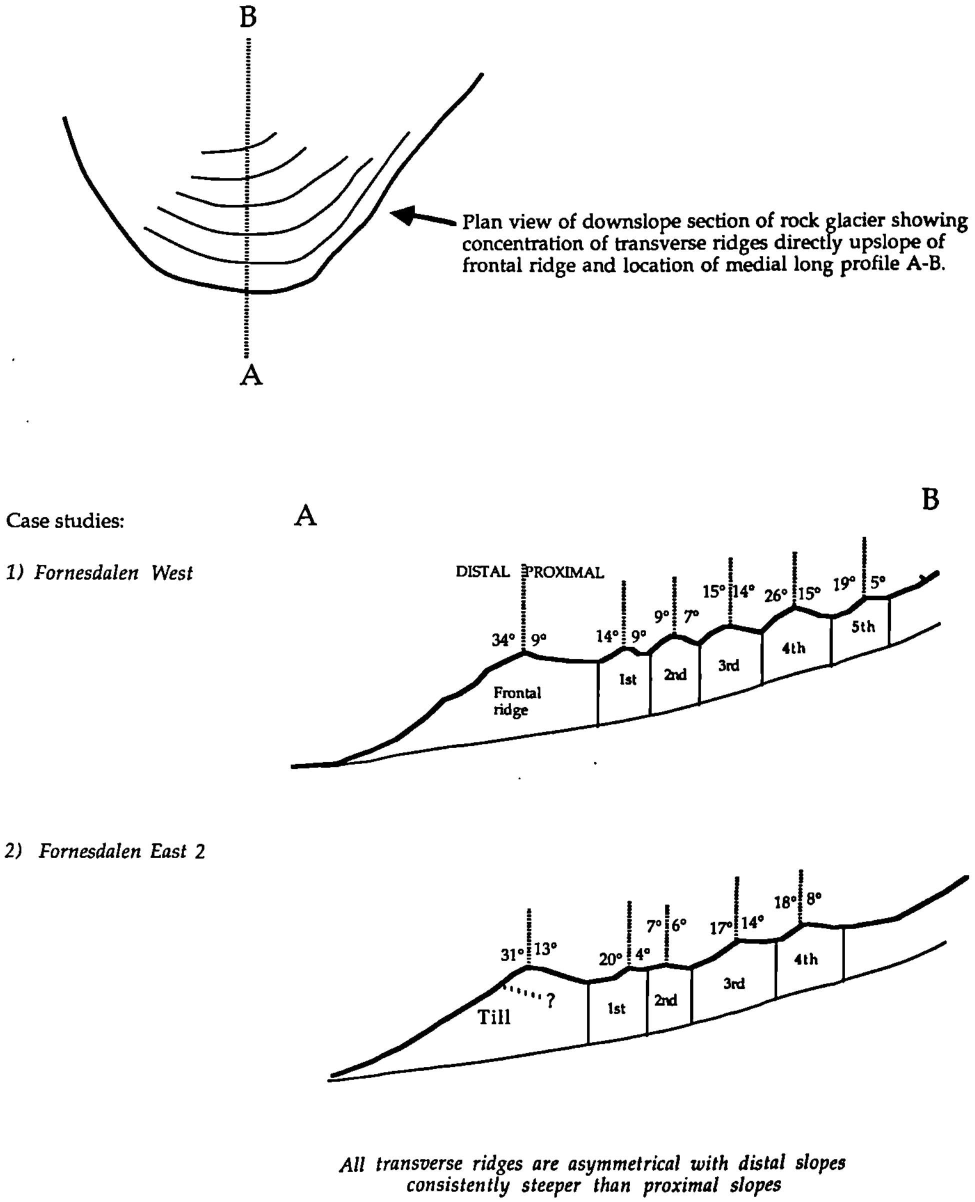


Figure 5.12 Asymmetry of transverse ridges on Fornesdalen West and Fornesdalen East 2 valley-wall rock glaciers. On all study rock glaciers, distal slope angles on transverse ridges are consistently steeper than proximal slope angles.

case-studies illustrating transverse ridge asymmetry on Fornesdalen West and Fornesdalen East 2 valley-wall rock glaciers, and Plate 5.20 shows the northern half of Fornesdalen West in which the transverse ridges examined in Figure 5.12 can be seen. On the photograph, depressions between ridges can be identified by the linear accumulations of large clasts. The origin of inner transverse ridges remains unclear. However, their shape and location suggest that they are related to rock glacier movement, whilst their asymmetry indicates the possibility that internal shearing as a result of compressive flow may occur. The shearing hypothesis is considered in more detail in section 6.2.3 below and in Chapter 7. It is also possible that melt-out of internal ice could produce ridge and depression structures that resemble inner transverse ridges.

Whereas transverse ridges are well-defined ridge features with intervening depressions, microrelief structures on valley-wall rock glaciers that are aligned parallel to the direction of movement should more accurately be described as longitudinal depressions. On valley-wall rock glaciers they are less common than transverse ridges, and appear to be morphologically distinct. In cross-section they are approximately symmetrical with average slope gradients of between 18° and 30°. Crude fall-sorting is usually evident on the flanks of these depressions. Longitudinal depressions on the study rock glaciers vary in depth from 2m to 5m and in length from 10m to 65m. They have developed at right angles to transverse ridges and trend downslope often in a sinuous manner where they are generally truncated by the frontal ridge. The

Plate 5.20 Northern half of Fornesdalen West rock glacier showing transverse ridges (middle of photograph) and a small longitudinal depression towards the left of the photograph in which late-lying snowpatches remain. The location of a large closed depression is marked by a large snowpatch at the right of the photograph (see Plate 5.21).

morphological characteristics of longitudinal depressions resemble those of melt-out structures, in which the rectilinear sides are approximately symmetrical and form a 'V-shape'. An analogous situation might be the typically straight-sided pits that form in sand or sugar during partial removal of the material through a basal funnel. It seems possible that the linear form of these depressions may represent the surface expression of subsurface drainage channels for the movement of meltwater within active features. For example, at two active valley-floor rock glaciers in Ellendalen, Norway and in Zinal, Valais, meltwater streams were seen emerging at the base of frontal slopes. Increased melt-out along the paths of these streams if they are reused for several years may produce longitudinal depressions as a result of melting of internal ice along the path of water flow. Their location appears to be random and they occur on only three of the valley-wall rock glaciers studied (Plate 5.20). It is possible that their development may be restricted to inactive or relict rock glaciers that contained unusually high amounts of internal ice.

5.5.4 *Closed depressions and surface ponds*

On several of the valley-wall rock glaciers studied, closed depressions occur at random over the surface of the feature, sometimes truncating transverse ridges. On the study rock glaciers, they range in depth from shallow features to depressions greater than 15m deep and 35m in diameter. Generally, they are composed of coarse debris, with an apparent absence of fall-sorting (Plate 5.21). Average slope gradients

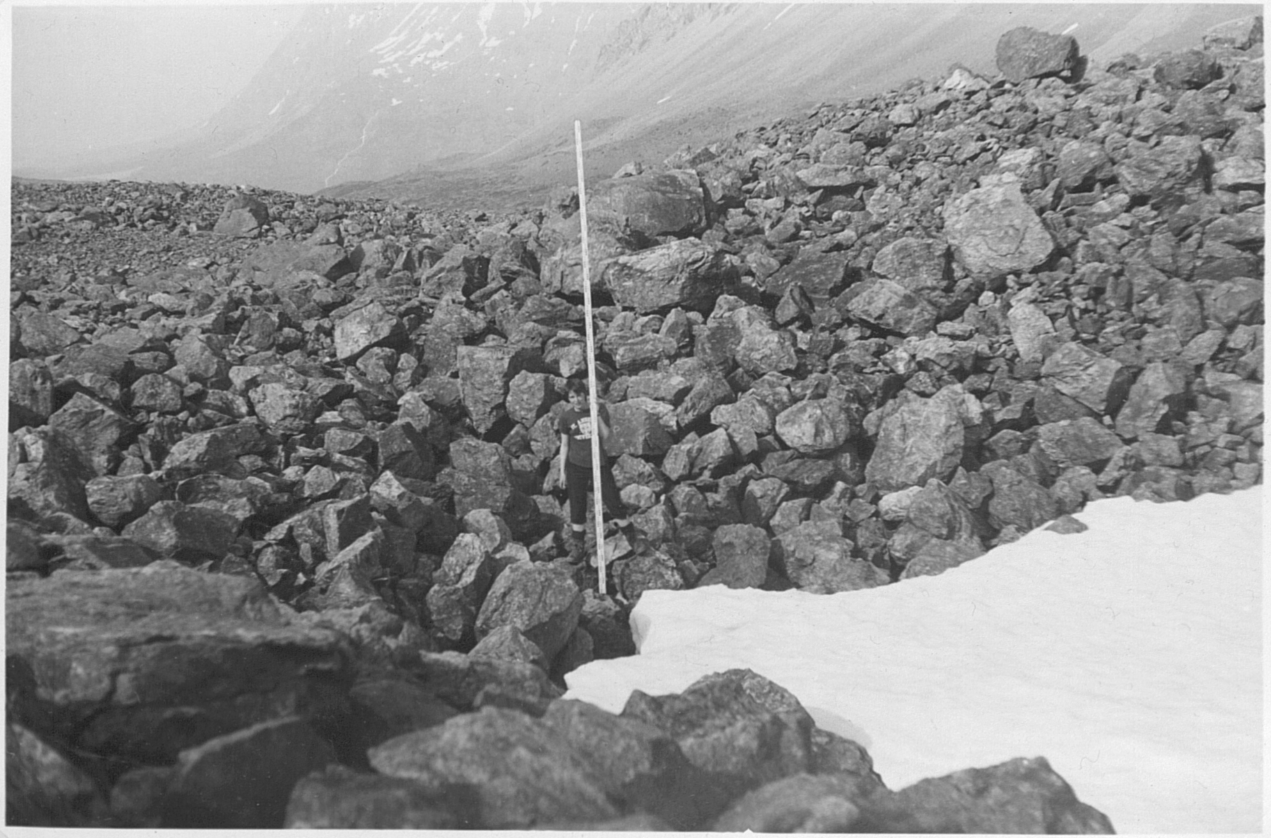


Plate 5.21 A large closed depression on Førnesdalen West rock glacier. The depression lies just upslope of the frontal ridge and can be seen on the medial long profile of the rock glacier in Figure 5.5. The height of the staff is 4m.

range from 21° to 31°. Such depressions may be the result of localised melting of large masses of either glacial or non-glacial ice.

Surface ponds often form in association with closed depressions on valley-wall rock glaciers. These ponds, which occur on several study rock glaciers, are generally shallow and less than 20m in diameter. Their occurrence strongly suggests the presence of an internal ice core or ice-rich frozen ground within at least part of the rock glacier because if the ground was unfrozen drainage would presumably occur. It seems unlikely that high water tables could prevent the drainage of these ponds given their favoured location upslope of the frontal ridges of rock glaciers. Additionally, such ponds appear to occur only on active and recently inactive rock glaciers; they have most probably accumulated through seasonal melt of internal ice, rainfall, and surface snow in closed depressions. One further interesting characteristic of the study sample was that lakes were found at the foot of the frontal ridges of inactive and relict rock glaciers at Zinal 1, Macun 3, and Trollvatnet. As such lakes were probably present before the rock glaciers formed, lake-water may have precipitated ice melt within the features and caused premature deactivation.

5.5.5 Summary

Five major points have emerged from the research reported in this section concerning the surface microrelief of valley-wall rock glaciers:

1) A four-unit subdivision based on a combination of morphological and sedimentological evidence and comprising a frontal ridge crest, an upper and lower rectilinear slope, and a basal talus apron, was found to be helpful in defining the common characteristics of the frontal slopes of valley-wall rock glaciers. Differences between active and inactive/relict forms were described within this framework.

2) Transverse ridges that occur on almost all of the study rock glaciers were found to be markedly asymmetrical, possibly reflecting internal shearing.

3) The morphology of longitudinal depressions, however, suggests that they are collapse features not necessarily related to rock glacier movement.

4) Closed depressions also appear to be collapse structures due to the melt of large localised masses of internal ice. On rock glaciers that still contain internal ice, some of these closed depressions have become filled with water.

5) The extent of surface microrelief features appears to depend largely upon the distance of rock glacier movement away from the base of the talus slope, and possibly, ice content.

5.6 Conclusions

The research reported in this chapter on rock glacier morphology may be summarised in three general conclusions.

1) A range of large-scale dimensional values for valley-wall rock glaciers was presented, including data on length, width, thickness, volume, average gradient and slope geometry. Bivariate statistical analysis of this data set has shown that several of these dimensional parameters are related; in particular, a strong positive correlation exists between rock glacier length and thickness. These findings have important ramifications regarding the possible mechanisms of valley-wall rock glacier formation (Chapter 7).

2) The frontal ridges of all valley-wall rock glaciers studied are well-defined. The morphology and sedimentology of these frontal ridges provide a useful means for classifying rock glaciers in the field as either active or inactive/relict. It seems likely that morphological differences between the frontal slopes of active, inactive and relict valley-wall rock glaciers may be explained mainly in terms of the cessation of differential forward movement following deactivation and consequent degradation of frontal slopes by mass-movement and melting of internal ice.

3) The form and location of inner transverse ridges suggest that they are related to rock glacier movement and possibly to internal shearing. Conversely, longitudinal and closed depressions that apparently occur randomly across a rock glacier, appear to be related to the melt-out of internal ice.

Chapter 6

Rock Glacier Sedimentology

6.1 Introduction and rationale

This chapter presents the results of research on the sedimentological characteristics of valley-wall rock glaciers in order to provide additional empirical evidence against which the models of valley-wall rock glacier formation outlined in Chapter 3 may be evaluated. Thus, Chapters 4, 5 and 6, which summarise respectively the locational, morphological and sedimentological characteristics of the study sample, comprise an extensive data base against which subsequent theoretical calculations may be tested.

Active rock glaciers comprise three principal sedimentological constituents, namely coarse debris, fine sediments and ice. The main aim of this chapter is to investigate each of these constituents in turn, and to identify any general patterns or regularities that may help explain the formation of valley-wall rock glaciers. Thus, in section 6.2, the characteristics of the coarse clastic component of valley-wall rock glaciers are examined and a series of clast-size analyses is undertaken for a sample of the study rock glaciers. In section 6.3, the general characteristics and distribution of fine sediments within valley-wall rock glaciers is described, and the results of granulometric and chemical comparisons of rock glacier, till and moraine sediments are presented. In the next section, the origin of rock glacier ice, the third major constituent of active glaciers, is investigated by comparing the isotopic characteristics of samples of ice obtained from active and inactive rock glaciers in Switzerland with ice obtained from neighbouring glaciers, snowbanks and ground ice. In the final section of the chapter, field evidence presented in Chapters 4, 5 and 6 is summarised by outlining criteria that may be used to distinguish *valley-wall rock glaciers from* other morphologically-similar features. The chapter concludes with a summary of the main conclusions.

6.2 Rock glacier clast analysis

6.2.1 Introduction

The upper surface of each active and inactive rock glacier and most relict rock glaciers in the study sample comprises an openwork boulder layer of debris that generally appears to fine downwards from very coarse surface boulders to much finer sediments. Field observations indicate that rock glacier clasts are mainly angular; active rock glaciers may contain some very angular clasts whilst sub-angular clasts are often found on relict valley-wall rock glaciers. As discussed in section 2.5.2 above, three factors are believed to explain the predominance of large clasts on the surface of rock glaciers and the presence of crude sorting within the surface layer (e.g. Wahrhaftig & Cox, 1959; Potter, 1972). These are: 1) rock glacier movement, which causes the downward sifting of fine material between larger surface clasts; 2) the abrasion of coarse material beneath the surface, which produces fine debris; and 3) rain and meltwater, which wash fine material downwards.

Following deactivation, the openwork boulder layer is often preserved, which suggests the importance of precipitation in washing wind-borne or weathered fines through the surface clast layer. For example, some of the relict rock glaciers studied in Scotland such as those at Coire Beanaidh and Loch Etchachan, which are believed to have formed approximately 11,000-10,000 yr BP. (cf. section 4.2.4 above), still contain an openwork boulder mantle although at others, including those at Strath Nethy and the Lairig Ghru, the voids between the surface clasts have become infilled with fine wind-borne sediments and vegetation.

Fabric studies such as those described by Giardino and Vitek (1985, 1988) were not attempted at any of the study sites as component clasts, particularly those derived from massive gabbros and granites, are often insufficiently elongate for the measurement of preferred orientation. Instead, clast-size analyses were undertaken to determine if any sedimentological regularities exist on valley-wall rock glaciers. For example, clast-size distributions on inner transverse ridges and frontal slopes were examined in order to infer as much information on their formation as possible. In addition, clast size was measured at randomly selected sites along medial long profiles to determine if any correlation exists between distance downslope and clast size so that process links between rockfall talus, which is known to exhibit a downslope increase in clast size, and rock glaciers may be investigated.

6.2.2 Data collection

At each sampling site chosen for clast-size analysis, either 50 or 100 clasts were measured from the surface boulder layer. A minimum size of 15mm was established for the intermediate axis of a clast. At the majority of sites, samples were chosen at random by sampling radially outwards from a central point. Clast-size measurements on the slopes of inner transverse ridges, however, were sampled along transects that ran from the base to the top of each slope along the line of the medial long profile of the rock glacier. Ten valley-wall rock glaciers were included in the sample. The most detailed measurements were made on six valley-wall rock glaciers studied in northern Norway. Clast-size measurements on valley-wall rock glaciers studied in Switzerland were more difficult to make because of late-lying snow. Thus at the Swiss study sites, measurements could be obtained only for frontal ridge clasts.

The mean diameter of a clast is most accurately established by measuring its three principal axes, namely length, breadth and depth. However, it is generally accepted (e.g. Briggs, 1977) that for most purposes the 'b' or intermediate axis provides a sufficiently accurate estimate of mean clast diameter. Thus, given the aim of establishing the generality of clast-size relationships at numerous sample points over as many rock glaciers as possible, the 'b' axis measurement was adopted in this study. All measurements were made with steel rules and were rounded up or down as appropriate to the nearest 5mm. As clast-size data are generally skew, the Mann-Whitney U Test, which evaluates the significance of differences between paired samples of non-normally distributed interval scale data (e.g. Matthews, 1981) was used to test for statistically-significant differences between samples.

6.2.3 Downslope distribution and proximal/distal clast-size patterns

Figures 6.1 to 6.3 summarise the results of clast-size measurements on six valley-wall rock glaciers studied on the Lyngen Peninsula in northern Norway. Two aspects of clast-size distribution were investigated using these measurements. First, the data were used to determine if a relationship exists between downslope distance and median clast size, and second comparisons were drawn between median clast sizes on the distal and proximal slopes of inner transverse ridges.

When the six plots of clast-size distribution are viewed together, it is clear that there is no consistent relationship between median clast size and distance downslope; instead, there exist pronounced variations in clast size both within and between the rock glaciers, all of which are composed of massive weathered gabbro. For example, median clast sizes on Fornesdalen East 1 (FE1; Figure 6.1) and on Rodbergdalen (RO; Figure 6.3) fall largely between 100 and 150mm with no apparent increase or decrease in a downslope direction. On Tverrelvdalen rock glacier (Figure 6.2) however, the range of median clast sizes is greater, from 100mm to 250mm, and clasts appear to decrease in size with increasing distance from the source-wall. Conversely, Trollvatnet (TR) rock glacier, which contains the largest clasts of the Lyngen rock glaciers, appears to

CLAST SIZE DISTRIBUTION

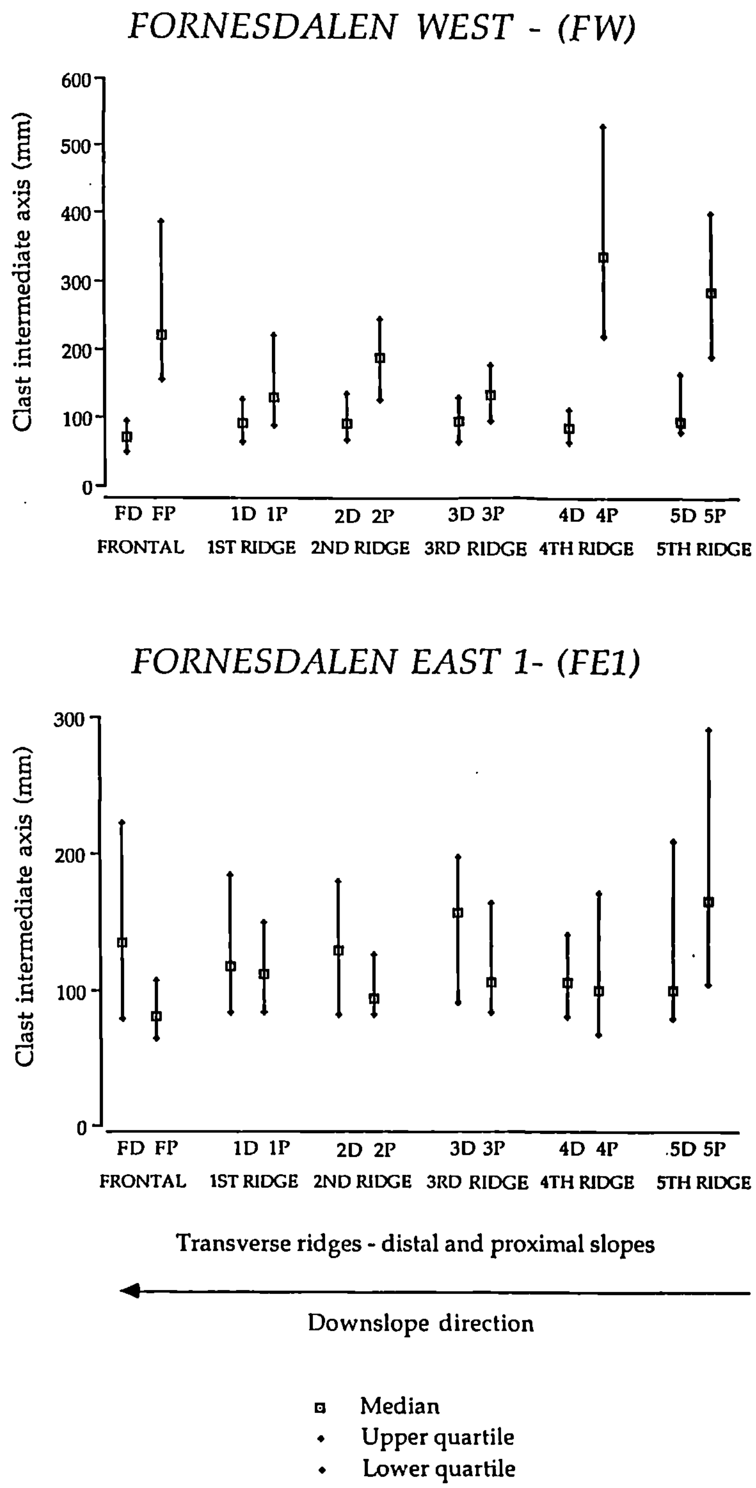


Figure 6.1 Downslope distribution and proximal/distal clast size patterns on Fornesdalen West and Fornesdalen East 1 valley-wall rock glaciers.

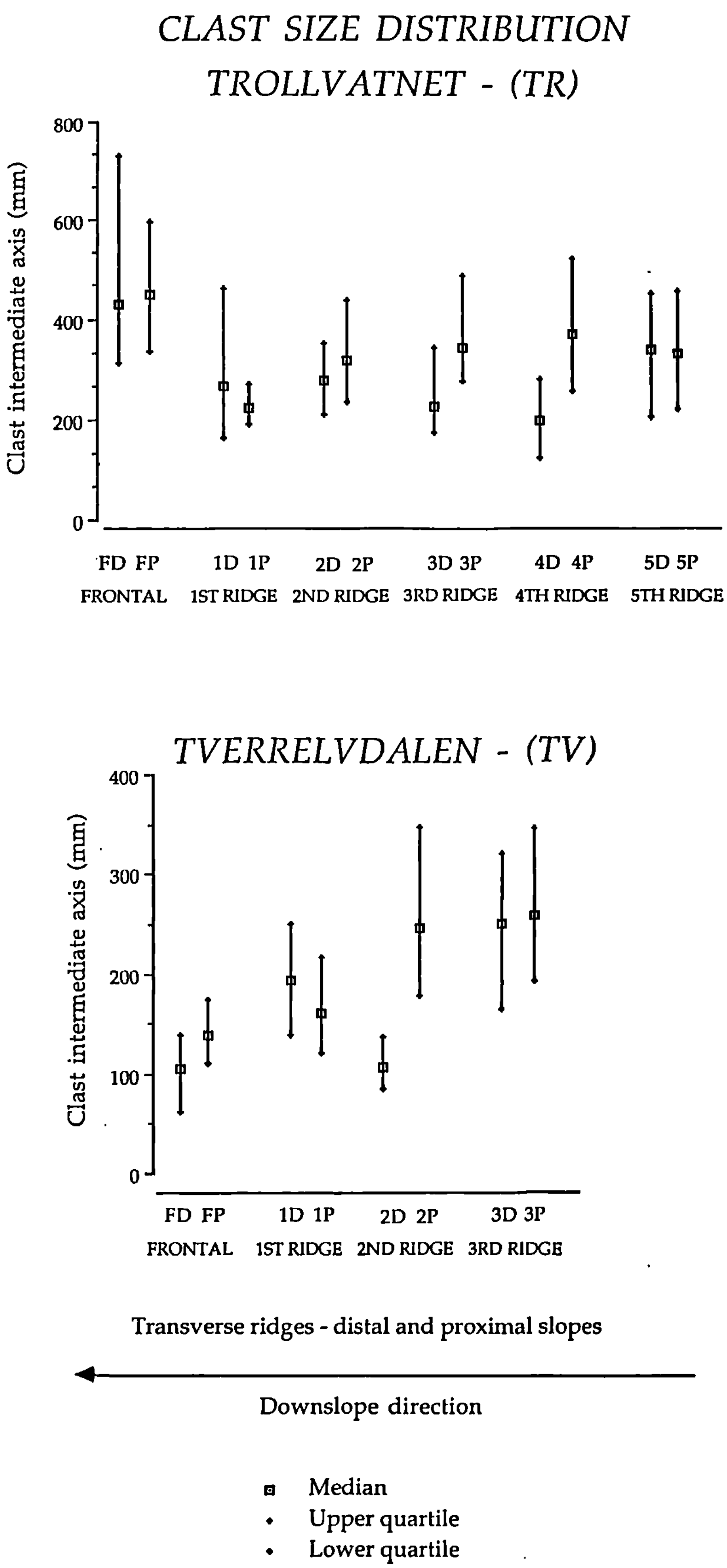
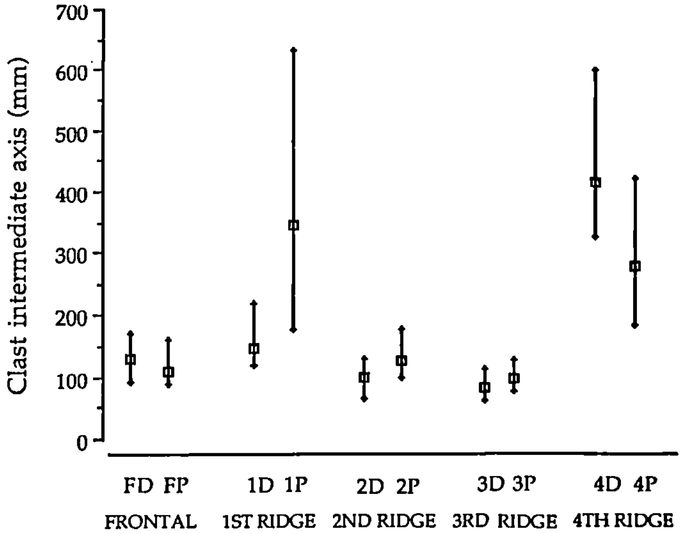


Figure 6.2 Downslope distribution and proximal/distal clast size patterns on Trollvatnet and Tverrelvdalen valley-wall rock glaciers.

CLAST SIZE DISTRIBUTION

FORNESDALEN EAST 2- (FE2)



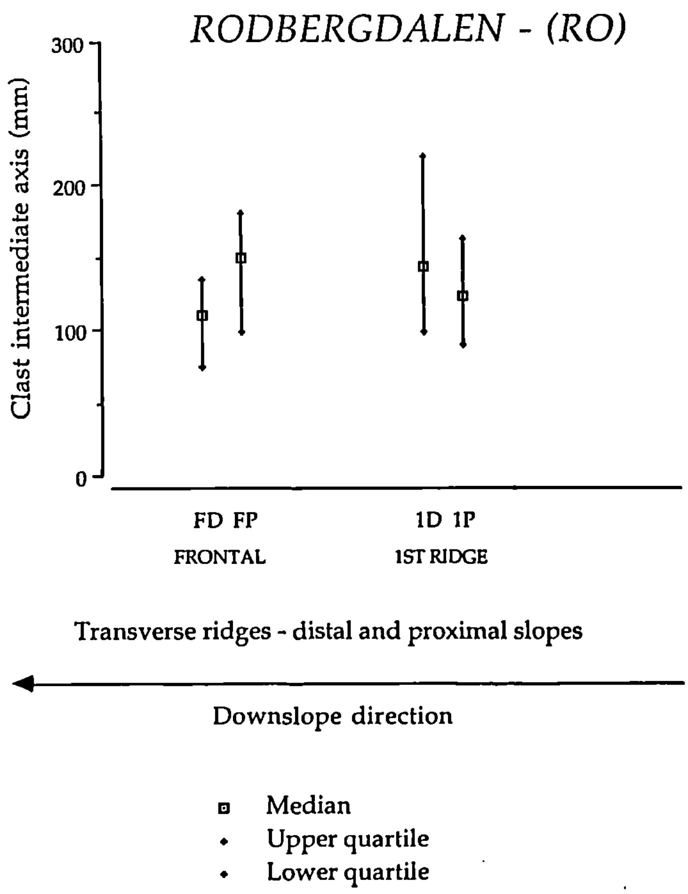


Figure 6.3 Downslope distribution and proximal/distal clast size patterns on Fornesdalen East 2 and Rodbergdalen valley-wall rock glaciers.

exhibit a slight increase in median clast size with increasing distance from the source-wall. The remaining two rock glaciers, namely Fornesdalen West (FW) and Fornesdalen East 2 (FE2), exhibit large variations in median clast size which appear to occur randomly down the length of these rock glaciers.

The absence of a general downslope increase in rock glacier clast size, except perhaps at Trollvatnet, contrasts markedly with the pattern of clast-size distributions on rockfall talus cones, which generally exhibit marked increases in clast size with increasing distance downslope (e.g. Church *et al.*, 1979). Given the close spatial association of talus slopes and valley-wall rock glaciers, the pronounced differences in clast-size distributions appear to reflect not only an abrupt process boundary but also provide some clues concerning the rate of debris input. For example, as large surface clasts generally occur down the entire length of rock glaciers and are not concentrated at downslope margins, deposition of large clasts at the foot of talus slopes during rock glacier formation must have kept pace with rock glacier movement.

Variations in median clast size between each valley-wall rock glacier shown in Figures 6.1 to 6.3 are pronounced, despite lithological consistency. For example, the upper quartile values of clast-size diameter on Rodbergdalen (RO) and on Fornesdalen East 1 (FE1) do not exceed 300mm whereas the comparable values on Trollvatnet and on Fornesdalen East 2 (FE2) exceed 600mm. This variability is not explained by differences in rock glacier size, as FE1 and FE2, which are adjacent to one another, have broadly similar lengths of 200 and 250m respectively. Instead, variations in the size of surface clasts probably reflect local differences in the source-wall characteristics of each rock glacier; for example, rock glacier clast size may be influenced by joint-spacing in the source-wall.

As can be seen from Figures 6.1 to 6.3, there are pronounced differences in median clast-sizes between *the proximal and distal slopes of the inner* transverse ridges on each rock glacier. In general, distal slopes tend to be characterised by smaller clasts. For example, on FW (Figure 6.1) median clast size is markedly greater on proximal slopes (Plate 6.1). This pattern is generally true for rock glaciers FE2, TR and TV, although on one or two ridges of each of these rock glaciers proximal clasts are marginally smaller than clasts on corresponding distal slopes. Field evidence indicates that on these inactive rock glaciers only distal slopes tend to be fall-sorted (Plate 6.2). Moreover, fall-sorting exists on slopes with gradients as shallow as 15°, suggesting that such slope angles have decreased since deactivation. There appears to be no relationship between clast size and transverse ridge size.

Although the exact mode of transverse ridge formation is unknown, several workers have suggested that they are shear ridges (e.g. White, 1987). Interestingly, the observed differences in clast size and slope gradient between the proximal and distal slopes of these asymmetric transverse ridges (cf. section 5.5.3 above) may be explained in terms of the shearing process. For example, the angle of shear for a block of

Plate 6.1 Inner transverse ridges on Fornesdalen West rock glacier, Lyngen. Distal slopes are characterised by smaller clasts than proximal slopes. Depressions between ridges can be identified by the curvilinear accumulations of large clasts. Figures for scale are circled.

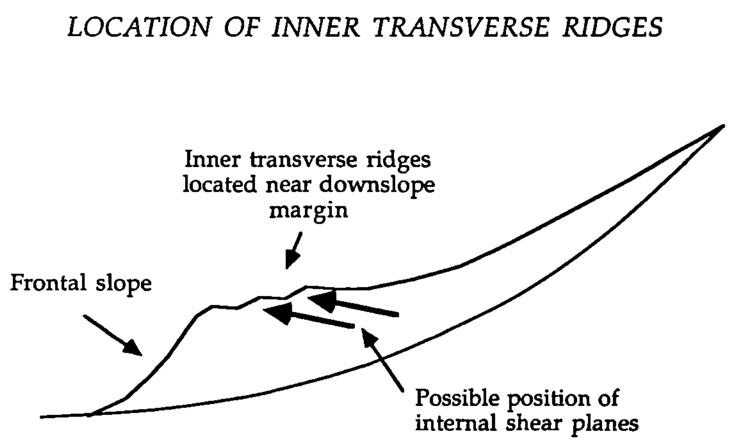
Plate 6.2 Inner transverse ridges and depressions on Fornesdalen East rock glacier, Lyngen. To the left of the photograph, the gentler proximal slope supports non fall-sorted large clasts. To the right of the figure standing in the depression, the steeper distal slope supports generally finer clasts.

material on a horizontal bed rarely exceeds the angle of initial yield of coarse bouldery debris (Means, 1976), which normally lies between about 35° and 40° (Chandler, 1973; Statham, 1976). Thus, the upper sheared surface or, in this case, the upper surface of the rock glacier which would become the proximal slope of the ridge, would remain largely undisturbed and would not be subject to fall-sorting as its gradient would remain below the angle of initial yield (Figure 6.4). Moreover, as rock glacier surfaces are generally characterised by coarse bouldery debris, the proximal slope of the ridge would likewise be characterised by large clasts. In contrast, the gradient of the distal ridge slope, which is the emergent or 'scarp' face, would be greater than the angle of residual shear; thus fall-sorting would occur and finer material would be exposed at the surface by shearing.

Surprisingly, clast sizes on the distal slopes of the rock glacier at FE1 are all larger than on the corresponding proximal slopes, except for the fifth inner ridge. However, it is possible that differential ice-melt during rock glacier deactivation may substantially alter the form of these inner transverse ridges and that the very strong relationships observed at Fornesdalen West rock glacier, for example, would not always be preserved elsewhere.

6.2.4 Clast size-analyses of frontal slopes

Figures 6.5 to 6.7 summarise clast-size distributions on the frontal slopes of three rock glaciers studied in Switzerland. At Radüner valley-wall



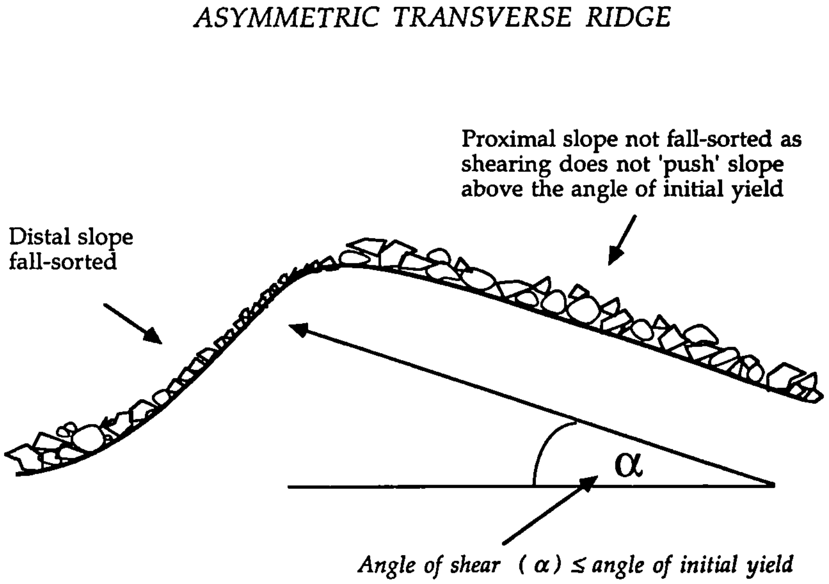


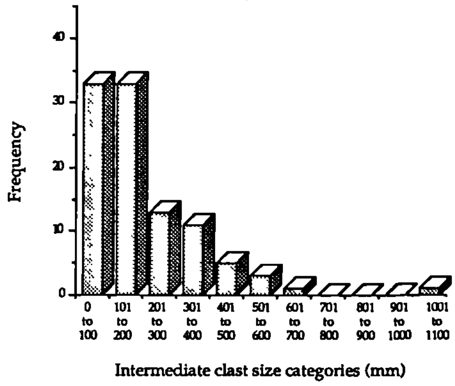
Figure 6.4 Clast size differences on the proximal and distal slopes of inner transverse ridges as explained by internal shearing along an angle less than the angle of initial yield.

CLAST SIZE DISTRIBUTION
FRONTAL SLOPE

RADÜNER VALLEY-WALL ROCK GLACIER

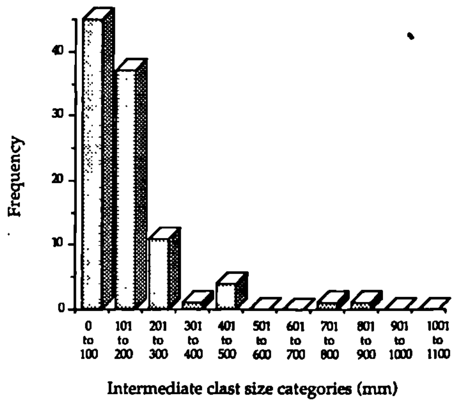
A) FRONTAL RIDGE CREST

Median: 165.0
Mean: 194.1
Skewness: 1.8
n = 100



B) MID-POINT OF FRONTAL SLOPE

Median: 112.5
Mean: 145.2
Skewness: 2.9
n = 100



C) BASE OF FRONTAL SLOPE

Median: 217.5
Mean: 241.5
Skewness: 1.1
n = 100

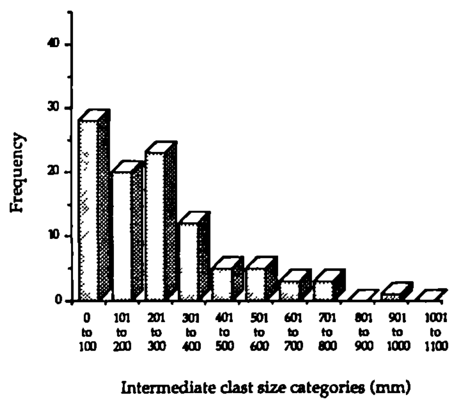


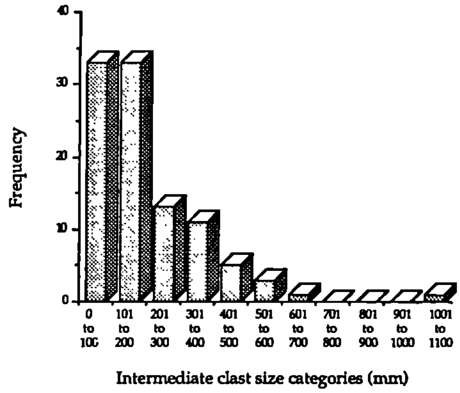
Figure 6.5 Clast size distribution on the frontal slope of Radüner valley-wall rock glacier, Switzerland.

CLAST SIZE DISTRIBUTION
FRONTAL SLOPE

RADÜNER VALLEY-WALL ROCK GLACIER

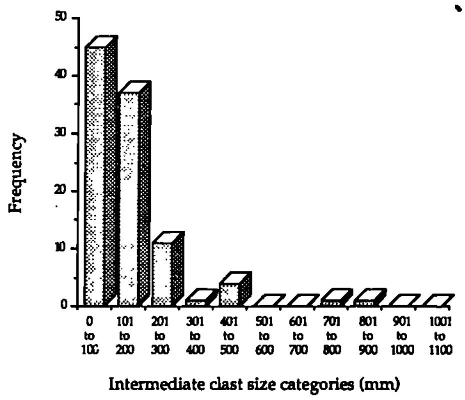
A) FRONTAL RIDGE CREST

Median: 165.0
Mean: 194.1
Skewness: 1.8
n = 100



B) MID-POINT OF FRONTAL SLOPE

Median: 112.5
Mean: 145.2
Skewness: 2.9
n = 100



C) BASE OF FRONTAL SLOPE

Median: 217.5
Mean: 241.5
Skewness: 1.1
n = 100

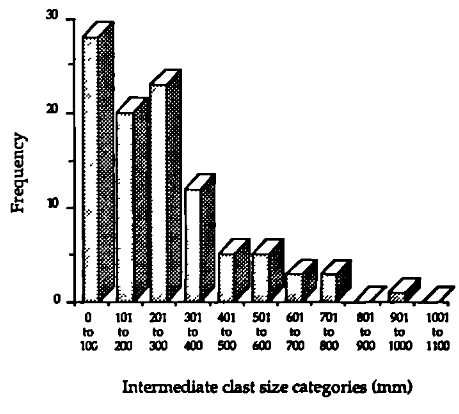


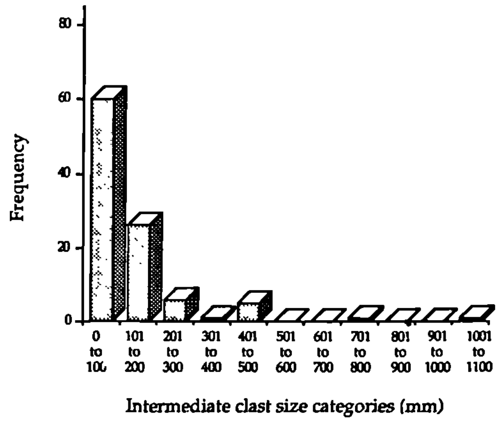
Figure 6.6 Clast size distribution on the frontal slope of Zinal 1 valley-wall rock glacier, Switzerland.

CLAST SIZE DISTRIBUTION
FRONTAL SLOPE

AROLLA 2 VALLEY-WALL ROCK GLACIER

A) FRONTAL RIDGE CREST

Median: 80.0
Mean: 130.4
Skewness: 3.4
n = 100



B) MID-POINT OF FRONTAL SLOPE

Median: 60.0
Mean: 69.0
Skewness: 1.3
n = 100

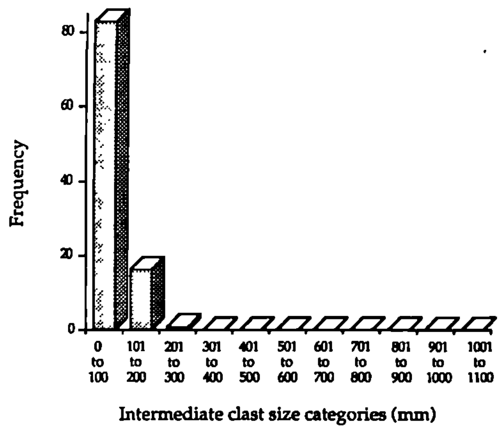


Figure 6.7 Clast size distribution on the frontal slope of Arolla 2 valley-wall rock glacier, Switzerland.

rock glacier, which is thought to be either active or recently inactive, 100 clasts were sampled from the base of the frontal slope, 100 from midway up the frontal slope and 100 from the frontal ridge crest (Figure 6.5). The average gradient of the frontal slope is 47°. The highest proportion of large clasts (median=217.5mm) was found in the basal talus apron at the foot of the frontal slope whereas the mid-point of the frontal slope is characterised by generally small clasts. The median value of clasts sampled from the frontal ridge crest (165) falls between the other two median values (112.5 and 217.5, respectively). The shape of the histograms clearly shows that all three distributions exhibit a high degree of skew. As discussed in section 5.5.2 above, a few large clasts are embedded in the predominantly finer sediments that characterise the middle sections of frontal slopes. This is reflected in the pronounced skewness of the sample obtained from the mid-point of the frontal slope.

A similar frontal slope clast-size distribution may be seen in Figure 6.6 which summarises clast-size measurements at Zinal 1, an apparently active or recently inactive valley-wall rock glacier with a frontal slope gradient of 48°. At Zinal (Plate 5.6, above), 50 samples were collected from the ridge crest, 50 from the mid-point of the frontal slope, and 50 from the base of the slope. The highly-skew clast-size distribution of the mid-point sample emphasises the predominance of fine material and small clasts that generally occur on lower rectilinear slopes of active or recently inactive rock glaciers (cf. section 5.5.2; Figure 5.12). The frontal ridge crest at Zinal, not the basal talus apron as at Radüner, contains the greatest proportion of large clasts. It is possible, however, that the sample obtained from the base of the slope at Zinal may not have been truly representative as some very large clasts could be seen in a partially frozen lake at the downslope margin of the rock glacier; these clasts could not be measured and were not included in the sample.

At a third rock glacier site, Arolla 2, which appears to be inactive, 100 samples were obtained from the ridge crest and 100 from the mid-point of the slope. Late-lying snow obscured the base of the slope preventing further sampling (Plate 5.4). From the distributions shown on Figure 6.7, it can be seen that over 80% of the clasts sampled from the mid-point of the slope fall into the smallest size category, in which intermediate clast diameters are less than 100mm. On Arolla 2, larger clasts do not appear to have become embedded in the finer sediments of the middle rectilinear slope and the sample obtained from the ridge crest exhibits a higher degree of skew.

6.2.5 Comparison of rock glacier and talus clast sizes

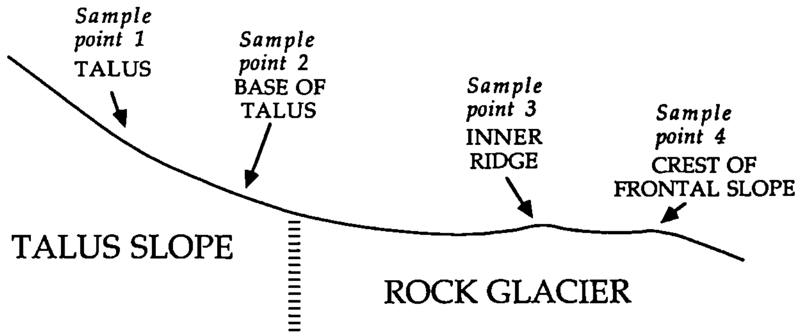
At the majority of rock glacier study sites, it was noted that the largest rock glacier clasts are consistently greater than the largest clasts on corresponding talus slopes. In order to examine this observation in more detail, clast samples were obtained from talus and rock glacier deposits at Flüela-Wisshorn 3 rock glacier. A sample bias was introduced in which the minimum size of the intermediate clast axis was set at 400mm, so that large clasts only would be measured in each sample. Fifty clasts were measured at four locations, namely the frontal ridge crest of the rock glacier, an inner transverse ridge crest, the base of the active talus, and mid-way up the talus (Figure 6.8). Descriptive statistics of the intermediate axis measurements (Figure 6.8) show that median clast size is greatest for the frontal ridge crest of the rock glacier (800mm) and least for the higher of the talus sample points (473mm). Indeed, median clast size increases with distance downslope and the upper quartile value for the base of the talus reaches only 681mm which is much smaller than the corresponding rock glacier values of 935mm and 1122mm. A series of Mann-Whitney U tests indicates that statistically significant differences exist between the sizes of the largest clasts on talus and those on the rock glaciers (Figure 6.8).

Of the nineteen rock glaciers chosen for detailed field study in Switzerland and northern Norway (cf. Figures 5.1 and 5.2), fourteen rock glaciers contain clasts that are visibly larger than adjacent talus clasts. One explanation for this pattern is that during rock glacier formation, debris input was more catastrophic than the discrete rockfall events that produce talus deposits. At the Scottish rock glacier sites, talus slopes are largely relict and often vegetated, thus preventing a similar comparative study. However, as discussed in section 5.4.2 above, there is evidence of large-scale rock slope failures at several of the Scottish rock glaciers.

6.2.6 Summary

Five main conclusions may be drawn from the research reported in this section.

Flüela-Wisshorn 3 valley-wall rock glacier



Intermediate clast size measurements (mm)

	Talus	Base of talus	Inner ridge	Frontal ridge crest
Median	473	600	795	800
Lower quartile	440	490	639	654
Upper quartile	559	681	935	1122

Mann-Whitney U tests of intermediate clast size measurements

	TALUS 1	BASE OF TALUS 2	INNER RIDGE 3
BASE OF TALUS	Reject H_o at $p \leq 0.001$		
INNER RIDGE	Reject H_o at $p \leq 0.001$	Reject H_o at $p \leq 0.001$	
FRONTAL CREST	Reject H_o at $p \leq 0.001$	Reject H_o at $p \leq 0.001$	No significant difference at 0.05

Mann-Whitney U test for non-parametric data:

H_o : median X = median Y

H_a : median X ≠ median Y

Figure 6.8 Location of talus and rock glacier clast sampling sites at Flüela-Wisshorn 3 valley-wall rock glacier together with intermediate clast size summary statistics and Mann-Whitney U Test results.

1) There is no apparent general relationship between median clast size and distance downslope, which suggests that deposition of large clasts at the foot of talus slopes must have kept pace with rock glacier movement.

2) Steep distal slopes of inner transverse ridges tend to be characterised by smaller clasts than gentler proximal slopes, which supports the hypothesis that inner transverse ridges may have formed through internal shearing (cf. section 5.5.3).

3) Mean clast size is not related to the size of the inner transverse ridges nor to the overall size of the rock glacier.

4) The frontal slopes of rock glaciers exhibit similar clast-size distributions in which finer sediments predominate midway down frontal slopes, particularly on active or recently-inactive rock glaciers.

5) At Flüela-Wisshorn rock glacier, the sizes of the largest clasts on the rock glacier are statistically significantly greater than those on the adjacent talus, which suggests that large-scale rockfall events may be associated with the formation of this rock glacier and several others on which rock glacier deposits appear coarser than corresponding talus.

6.3 *Rock glacier fines*

6.3.1 *Introduction*

It has now been widely established both by direct observation and by remote-sensing techniques that the interiors of rock glaciers are comprised mainly of fine sediments and that the layer of coarse blocks and boulders at the surface is relatively thin (e.g. Haeberli, 1975; White, 1976; Fisch *et al.*, 1977; Haeberli, 1985; cf. section 2.5). In this section, observations on the general distribution and characteristics of fine sediments within valley-wall rock glaciers are presented. A combination of surface exposures and excavations was used to study and sample the characteristics of these fine sediments. In addition, samples of fine sediments (<2mm) obtained from two valley-wall rock glaciers in Fornesdalen, northern Norway were compared using chemical and granulometric techniques with fines sampled from adjacent lateral moraines and underlying till deposits to establish if moraine and till deposits became incorporated within the rock glaciers during formation.

6.3.2 *General distribution and description*

Surface fines occur most extensively on the lower rectilinear units of frontal slopes of active rock glaciers, where they are exposed by differential forward movement (cf. section 5.5.2 above). Less extensive exposures of fine sediments may be found on the frontal slopes of many inactive rock glaciers and on the distal slopes of some inner transverse ridges. Unfortunately, due to prolonged snowcover at the field sites in

Switzerland, much of the upper surface of these rock glaciers was obscured throughout the field season, thus preventing detailed investigation of their sedimentary characteristics. On inactive rock glaciers studied in northern Norway, fine surface sediments were frequently observed on the distal slopes of inner transverse ridges. As the base of these deposits was not reached during excavations up to 1.5 metres in depth, they are not thought to be wind-borne accumulations; instead it is possible that they have been brought to the surface by upward shearing during rock glacier movement.

Limited excavations were undertaken at randomly-selected points on several rock glaciers to establish if variations exist in the depth of the surface boulder layer and to examine underlying deposits. Generally, fine material was encountered at relatively shallow depths (<1.5m), particularly on the distal slopes of both frontal and inner transverse ridges. In many of the pits, clasts up to approximately 60cm in diameter were found embedded within the fine matrix (Plate 6.3). At other locations, most notably on the proximal slopes or crests of transverse and frontal ridges, the base of the boulder mantle was not reached despite removing up to 1.5 metres of clasts. The presence of very large surface boulders prevented any sub-surface investigations at several sites.

Attempts were made to use electrical resistivity techniques to investigate the sub-surface distribution of fines and in particular to establish the depth of the unfrozen surface layer on active valley-wall rock glaciers. However, a general absence of surface fines except in localised patches on

Plate 6.3 Small pit excavated in the distal slope of an inner transverse ridge at Trollvatnet, an inactive or relict rock glacier in Lyngen, showing a predominantly sandy matrix in which clasts up to 60cm in diameter are embedded.

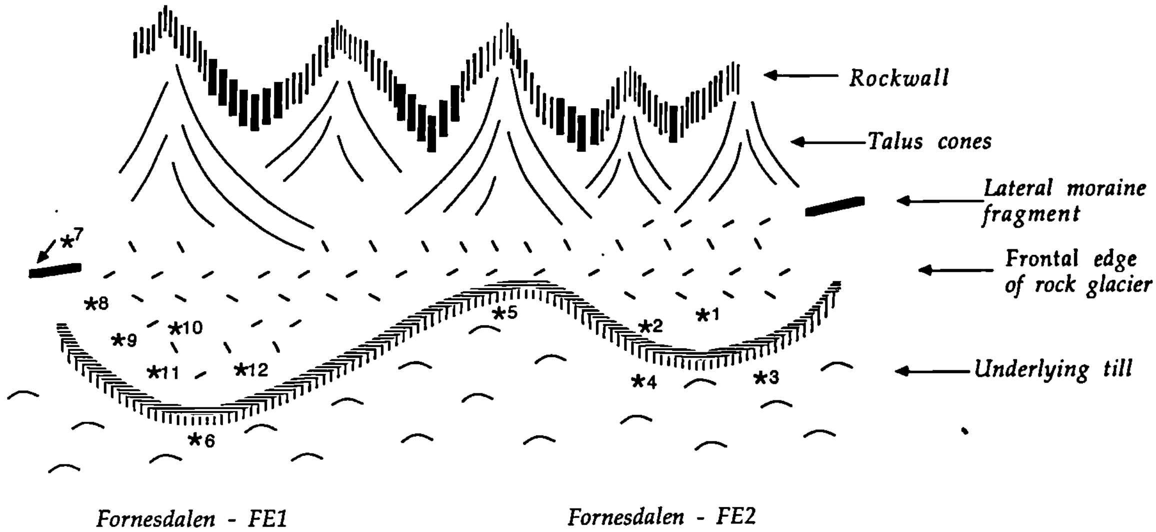
the upper surfaces of rock glaciers prevented any such measurements as it proved impossible to embed the electrodes securely in the deposits. Although the frontal slopes of rock glaciers have more extensive fine deposits, suitable contact points for the electrodes could not be found and no measurements were obtained.

On the Lyngen Peninsula in northern Norway, an interesting correlation was observed between the location of valley-wall rock glaciers and the presence of adjacent lateral moraine fragments. The absence of lateral moraines on slopes directly above many rock glaciers, despite the presence of such moraines on adjacent slopes, suggests that moraine debris may have become incorporated within rock glaciers as they moved downslope. If so, the possibility exists that the fines from these moraines may be incorporated within the rock glaciers. Furthermore, many of the Lyngen valley-wall rock glaciers have formed on top of till deposits. In section 5.4.5 above, the possibility was raised that these till deposits may have been ice-cored at the time of rock glacier formation, and thus may have deformed downslope under the weight of the overlying rockfall debris, possibly deforming upwards to produce inliers of till within the rock glaciers (i.e. forming a diapir). In order to test these hypotheses, samples of fines from rock glaciers, lateral moraines and till deposits were subjected to comparative granulometric and chemical analyses.

6.3.3 Data collection

Samples of fine material were obtained from Fornesdalen East 1 and Fornesdalen East 2 rock glaciers, from the till deposits that underlie the snouts of both of these rock glaciers, and from a lateral moraine fragment near the base of the talus (Figure 6.9). The underlying till deposits, which clearly must pre-date the rock glacier, are thought to be Preboreal in age, dating from *ca.* 9900 to *ca.* 9400 yr BP. (pers. comm., C.K. Ballantyne, 1987). The lateral moraine fragment is believed to be of Tromsø-Lyngen Readvance age, dating from between *ca.* 11,680 and *ca.* 10,150 yr BP. Lithologically, the rock glaciers appear to be comprised exclusively of massive weathered gabbro. Small quantities of dunite erratics were found together with the gabbro in both the underlying till deposits and the lateral moraine fragment.

All fine sediment samples were obtained from a depth of at least 30cm to ensure that they are not wind-blown deposits. Samples were kept in a cold store at 4°C before laboratory analysis. Initially, the samples were dried in an oven for 24 hours at a temperature of 105°C. They were then dry sieved at half phi intervals from 2000 microns (μm) to 63μm. Particle size components less than 63μm were measured at one-third phi intervals in a Coulter Counter (TA2) using a 140 micron tube and a 10% glycerol isoton solution (electrolyte solution). Granulometric results are presented as percentages and cumulative percentages by weight from -1 to 9 phi.



Sample points:

1 Rock glacier fines - FE2
2 Rock glacier fines - FE2

3 Underlying till deposits
4 Underlying till deposits
5 Underlying till deposits
6 Underlying till deposits

7 Lateral moraine fragment

8 Rock glacier fines - FE1
9 Rock glacier fines - FE1
10 Rock glacier fines - FE1
11 Rock glacier fines - FE1
12 Rock glacier fines - FE1

Figure 6.9 Location of samples obtained for granulometric and chemical analyses at Fornesdalen East rock glacier, northern Norway.

In addition to granulometry, the clay mineral content of each sample (<2mm) was analysed using X-ray diffraction, which is a useful technique for differentiating between soils of different chemical composition and origin. Its basic principle is that each clay mineral has an atomic structure which diffracts X-rays in a characteristic pattern (Wells & Smidt, 1978). Samples were oven-dried at 105°C and then passed through a 63μm sieve. They were then dispersed in deionised water and left to settle overnight so that the silt particles could be removed. The resulting clay suspension was placed on glass slides and allowed to dry at room temperature. The samples were analysed on a Philips PW 1010/PW 1349 X-ray diffractometer using Co Ka radiation and scanned from 5° to 65° at a rotation of 1° per minute .

Following X-ray diffraction analysis, X-ray fluorescence spectrography was used to obtain the concentrations of major elements in each sample. The technique operates by irradiating a sample with the broad spectrum of primary X-rays from an X-ray tube and then identifying and measuring the intensity of the secondary, fluorescent X-rays that are emitted by the elements. Initially, samples were ground very finely using an agate mortar and pestle and a fused glass bead was prepared from 0.5g of the sample (<125μm) and 2.5g of Spectroflux 105 with ammonium nitrate as oxidant. The analyses were performed on a Philips PW1212 using a Rh tube for primary excitation.

6.3.4 Results of granulometric and chemical analyses

Granulometry results are presented in Figures 6.10 to 6.13. In Figure 6.10, particle size graphs showing cumulative percentages (0 to 9 phi) are presented for two rock glacier samples and three samples taken from the underlying till at Fornesdalen FE2. The graphs show a separation of the two deposits with rock glacier sediments comprising slightly higher proportions of sand ($<4\phi$) than till samples. Both types of deposit, however, contain less than 1% by weight of clay. Silt comprises between 14 and 22% by weight for rock glacier samples and between 28 and 32% for till deposits.

Particle size analyses of rock glacier, till and lateral moraine samples collected at the adjacent rock glacier, Fornesdalen FE1, are shown in Figure 6.11. The rock glacier samples show similar particle size percentages to those found at FE1 (Figure 6.10). The sample obtained from the lateral moraine is generally finer than those from the rock glacier. Till sample 6 contains marginally lower percentages of clay and silt particles than the rock glacier sediments, but otherwise possesses rather similar granulometric characteristics.

The granulometric data for both sites are summarised in Figures 6.12 and 6.13. In general, till and lateral moraine samples contain smaller amounts of fine, medium and coarse sand (0 to 4 phi) and greater amounts of silt and clay (5 to 9 phi) than the rock glacier samples.

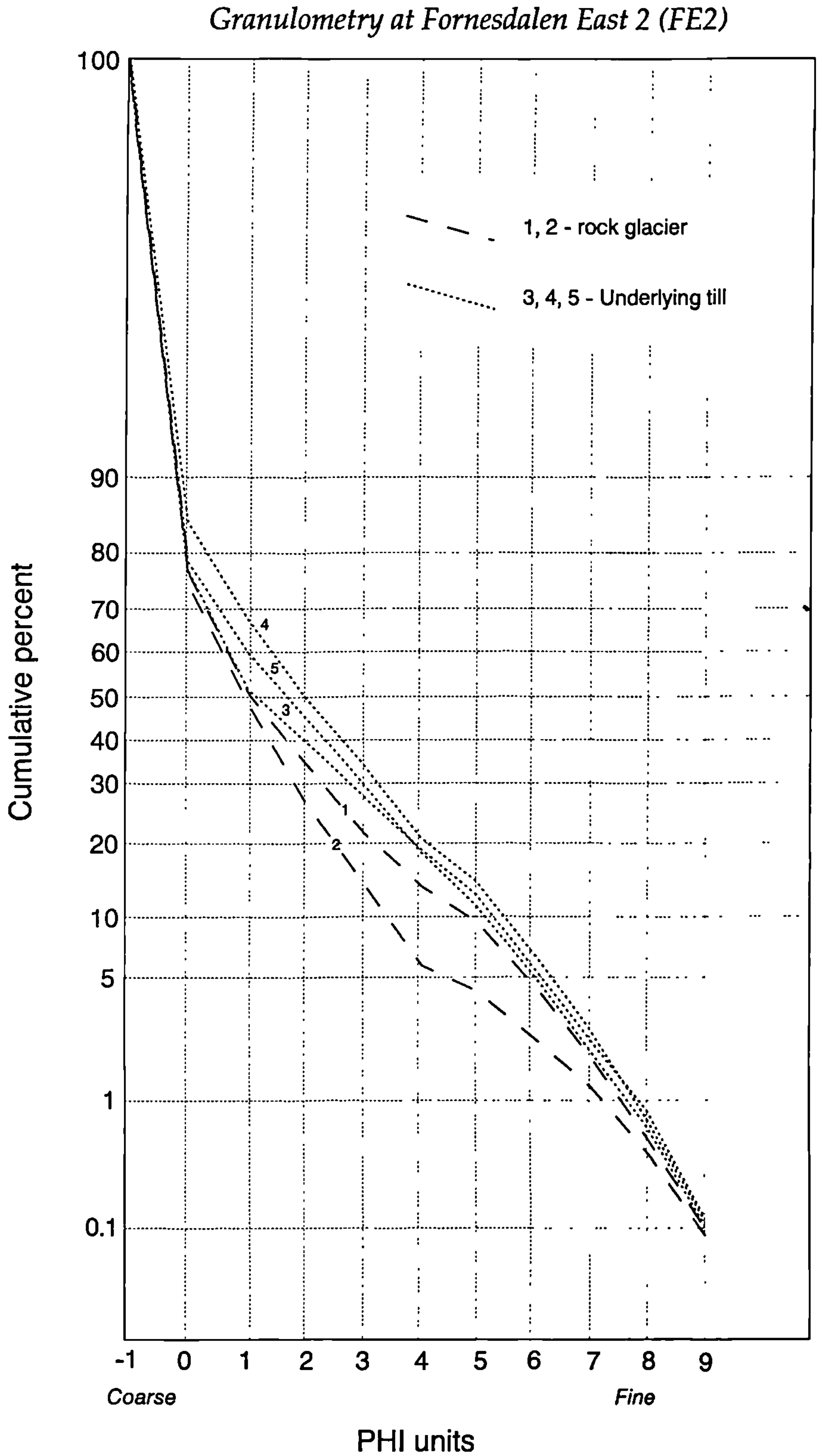


Figure 6.10 Cumulative percentage particle size graphs for two rock glacier and three till deposits sampled at Fornesdalen East 2 valley-wall rock glacier. Sample locations are shown in Figure 6.9.

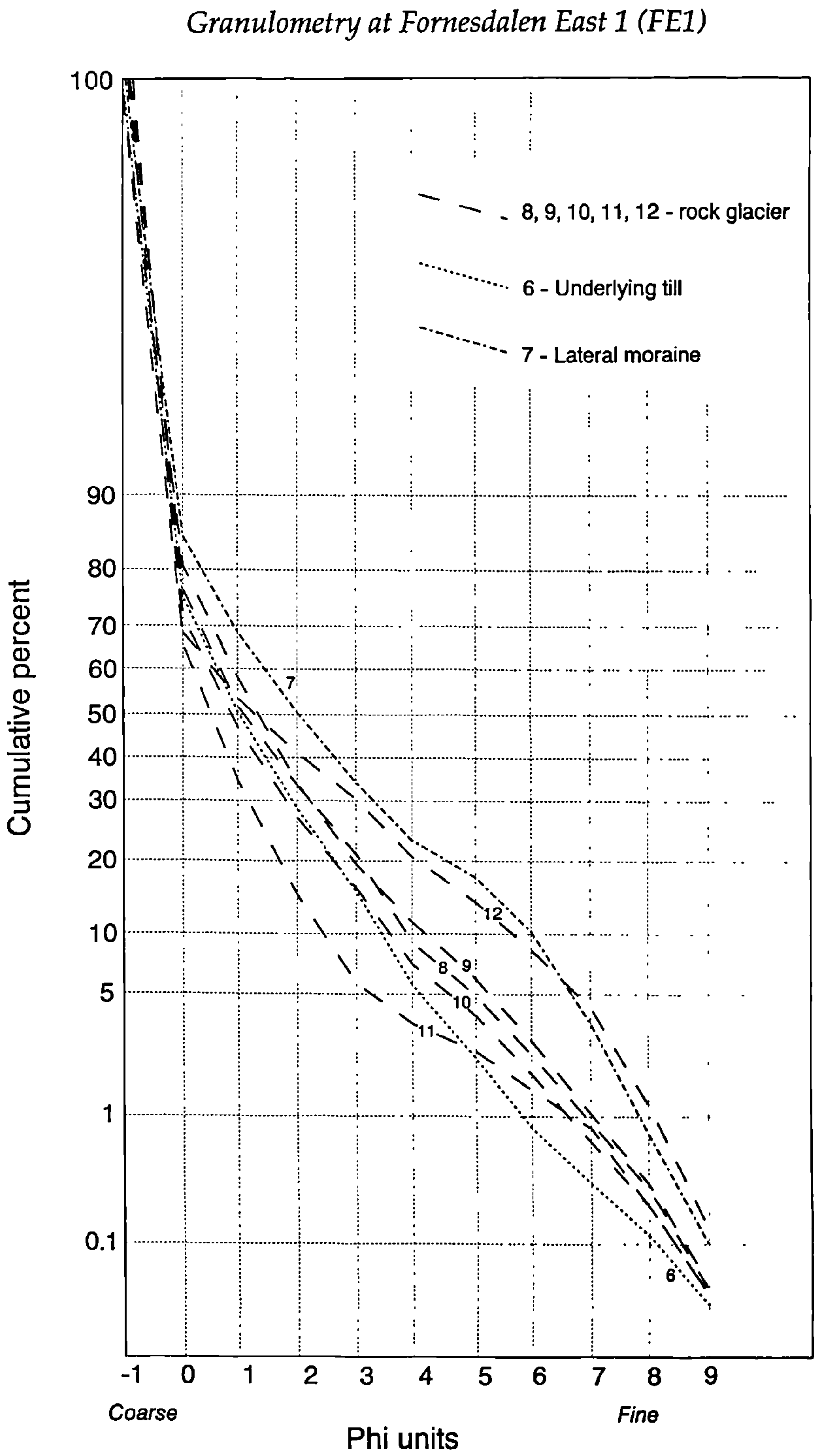


Figure 6.11 Granulometry of rock glacier, till and lateral moraine sediments sampled at Fornesdalen East 1 valley-wall rock glacier.

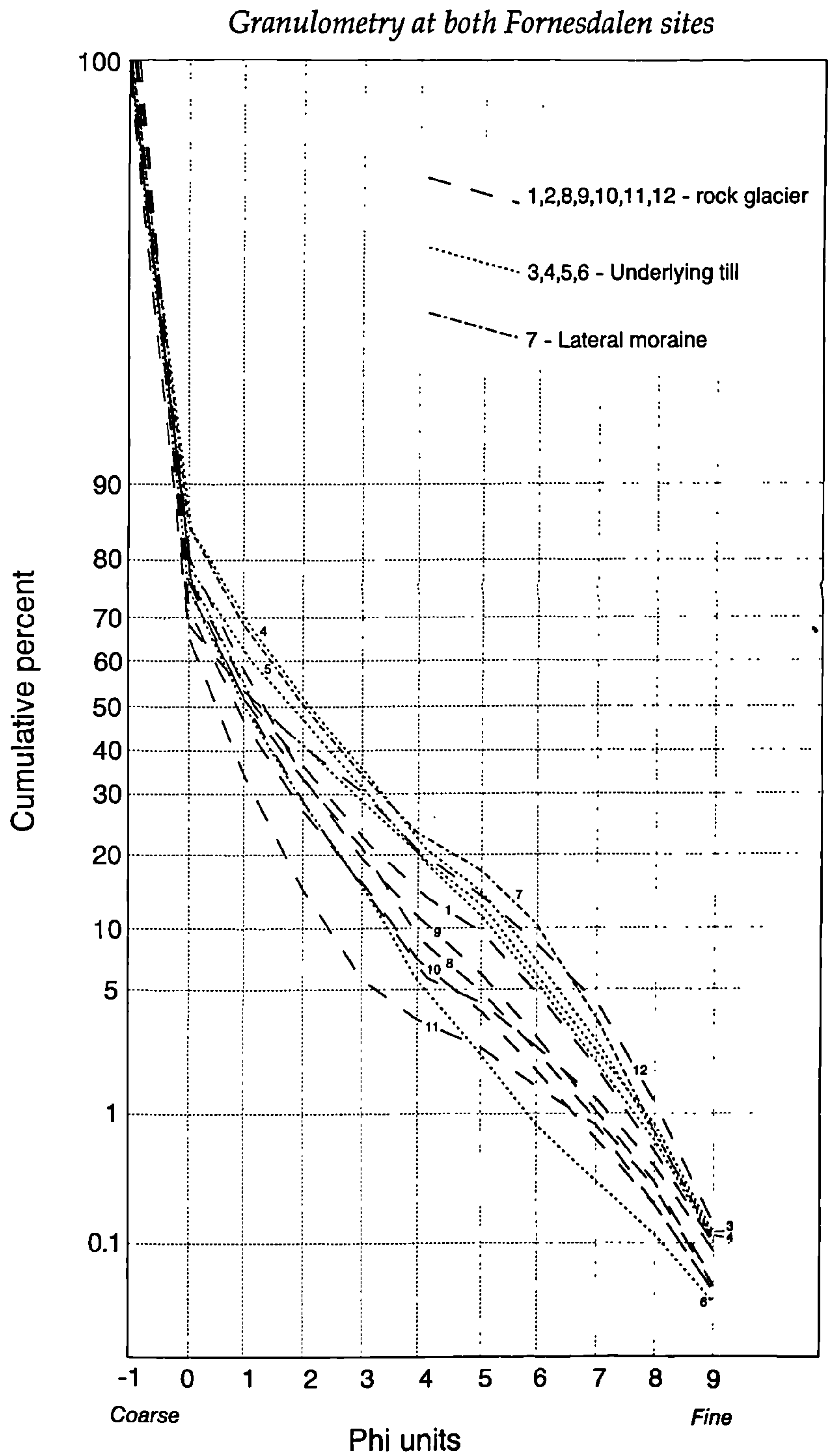


Figure 6.12 Granulometry of all rock glacier, till and lateral moraine sediments sampled in Fornesdalen, northern Norway.

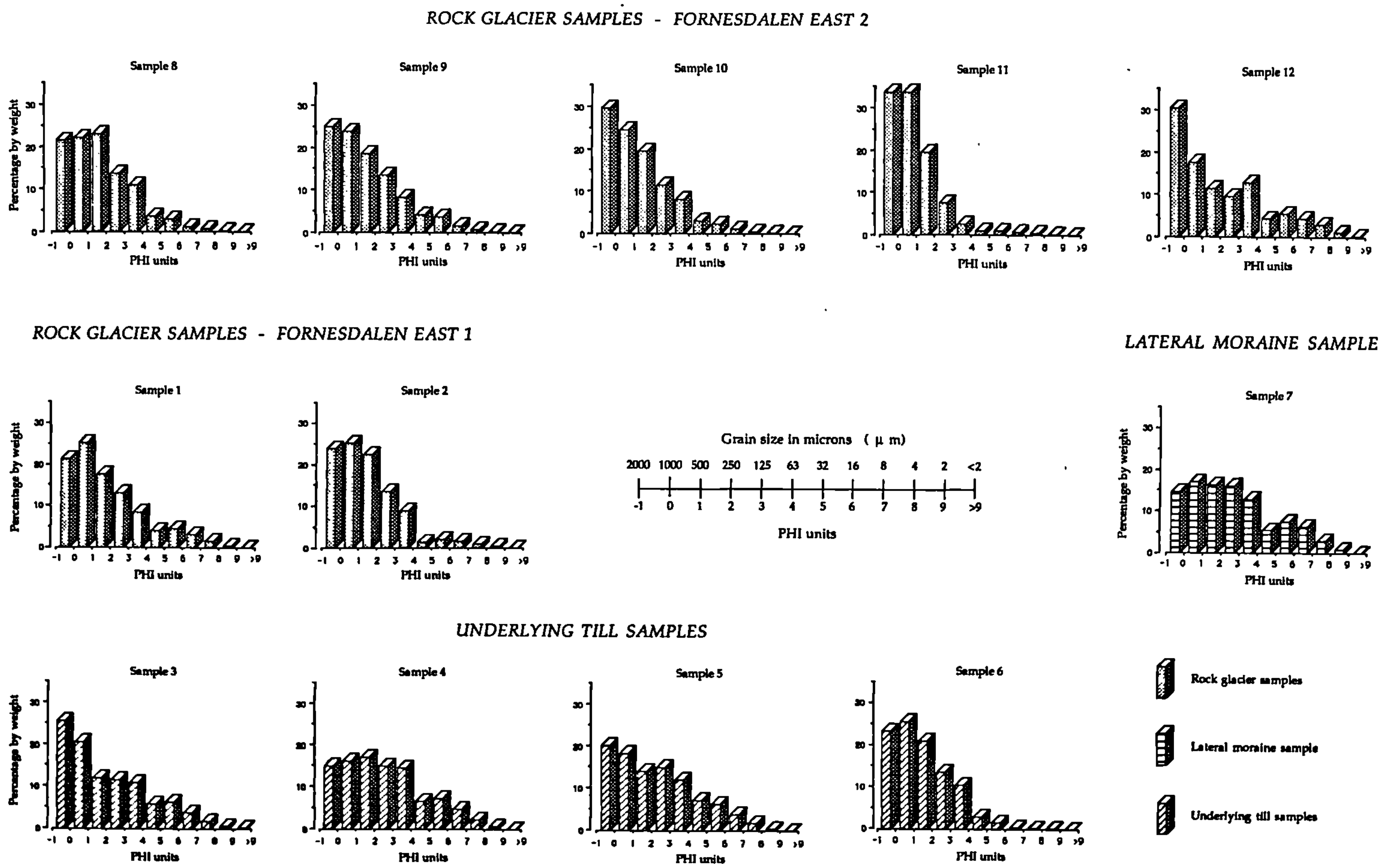


Figure 6.13 Comparative granulometry of samples (<2mm) taken fromrock glaciers, underlying till, and lateral moraine at Fornesdaeln East valley-wall rock glacier northern Norway.

Supporting statistical evidence for these conclusions is presented in Figure 6.14, where the percentage of particles > 0phi, > 1phi, and < 5phi are shown for each sample. As can be seen, rock glacier samples generally contain higher percentages of coarser particles (0phi and 1phi) and lower percentages of finer particles (5phi) than those samples from either the lateral moraine or underlying till. However, Mann-Whitney U test results, which are also shown in Figure 6.14, indicate that the differences discussed above between the rock glacier and underlying till samples are significant only at the 84% significance level.

One further implication of these granulometric results concerns the frost susceptibility of the rock glacier sediments. It is well known that pore size and, therefore grain size, strongly affect the growth and form of ice in a soil (e.g. Washburn, 1979). For example, grain size influences the movement of water to the freezing front because the potential for drawing water to the freezing front increases with decreasing grain size except for very heavy clays (Beskow, 1935). If ice segregation contributes to the formation of ice within rock glaciers (cf. section 3.2.4 above), rock glacier sediments must be frost susceptible (i.e. subject to growth of ice lenses). In Figure 6.15, the textural curves of the Fornesdalen rock glaciers are shown together with Beskow's limits of frost susceptibility (1935) and frost susceptibility criteria proposed by Casagrande (1932) and Terzaghi (1952). As Figure 6.15 shows, the rock glacier samples are at the limits of frost susceptibility as proposed by Beskow, although they may be termed frost susceptible if the less stringent criteria suggested by Casagrande and Terzaghi are adopted. Additionally, the rock glacier

	Rock glaciers %	Underlying till %	Lateral moraine %
> 0 phi	33.6 30.4 29.6 25.0 23.7 21.4 21.3	25.6 23.4 20.1 15.0	14.7
> 1 phi	67.2 54.0 48.8 48.8 48.0 46.5 43.5	49.1 46.2 38.5 31.3	31.6
< 5 phi	13.9 10.0 6.2 5.1 5.0 4.1 2.4	15.0 13.0 12.6 2.5	18.0

Mann-Whitney U tests

	Underlying till > 0 phi	> 1 phi	< 5 phi
Rock glaciers	Cannot reject H_0 at 0.05 (significant at 0.16)	Cannot reject H_0 at 0.05 (significant at 0.16)	Cannot reject H_0 at 0.05 (significant at 0.30)

Mann-Whitney U test for non-parametric data:

H_0: median X = median Y

H_a: median X ≠ median Y

Figure 6.14 Mann-Whitney U Test results of granulometry data.

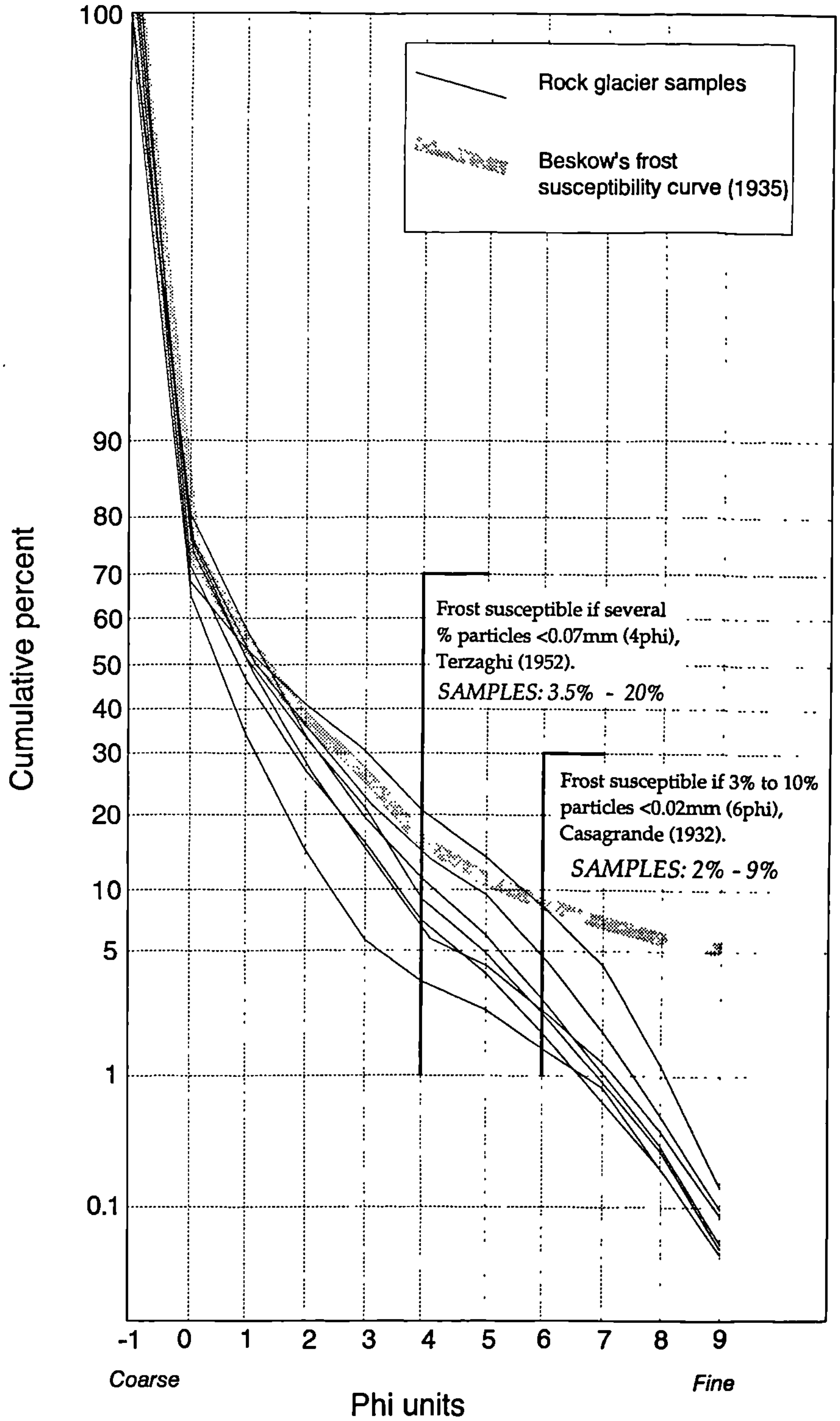


Figure 6.15 Particle size distributions for finer than 2mm fractions from rock glacier deposits in Fornesdalen, plotted against Beskow's frost susceptibility curve (1935) and frost susceptibility criteria proposed by Casagrande (1932) and Terzaghi (1952).

sediments described in this section were obtained from relatively shallow depths and it is probable that rock glacier sediments at greater depths would be finer and hence not as marginal in terms of frost susceptibility.

X-Ray diffraction analyses were undertaken to determine the proportions of different clay mineral groups within each sample. In the resulting traces each clay mineral group may be recognised by diagnostic basal reflections or 'd spacings' that are related to the particular type of layer and interlayer structure of the mineral. In comparative studies, the presence or absence of specific basal reflections is of greater importance than the size of the reflection. Traces were obtained for each of the seven rock glacier samples, four till samples, and one lateral moraine sample. The seven rock glacier traces proved virtually identical in their clay mineral groupings. Similarly, the four till samples proved virtually identical to each other. It is interesting to note the absence of secondary clay minerals in any of the traces which suggests that the fines have been derived largely from granular disintegration of the bedrock and not from chemical weathering. In order to allow visual comparisons of the three types of deposit, one representative trace of each is shown in Figure 6.16.

Clay minerals present in rock glacier fines (middle trace) are very similar to those from the lateral moraine fragment (upper trace). Feldspars and chlorites predominate with significant amounts of quartz and mica, as would be expected from weathered gabbro. However, the trace representing the sample of underlying till shows two significant differences from the trace of the rock glacier sample. First, the till trace contains a feldspar peak at 2.98Å which is completely absent both from the rock glacier samples and the moraine sample. Second, a pronounced

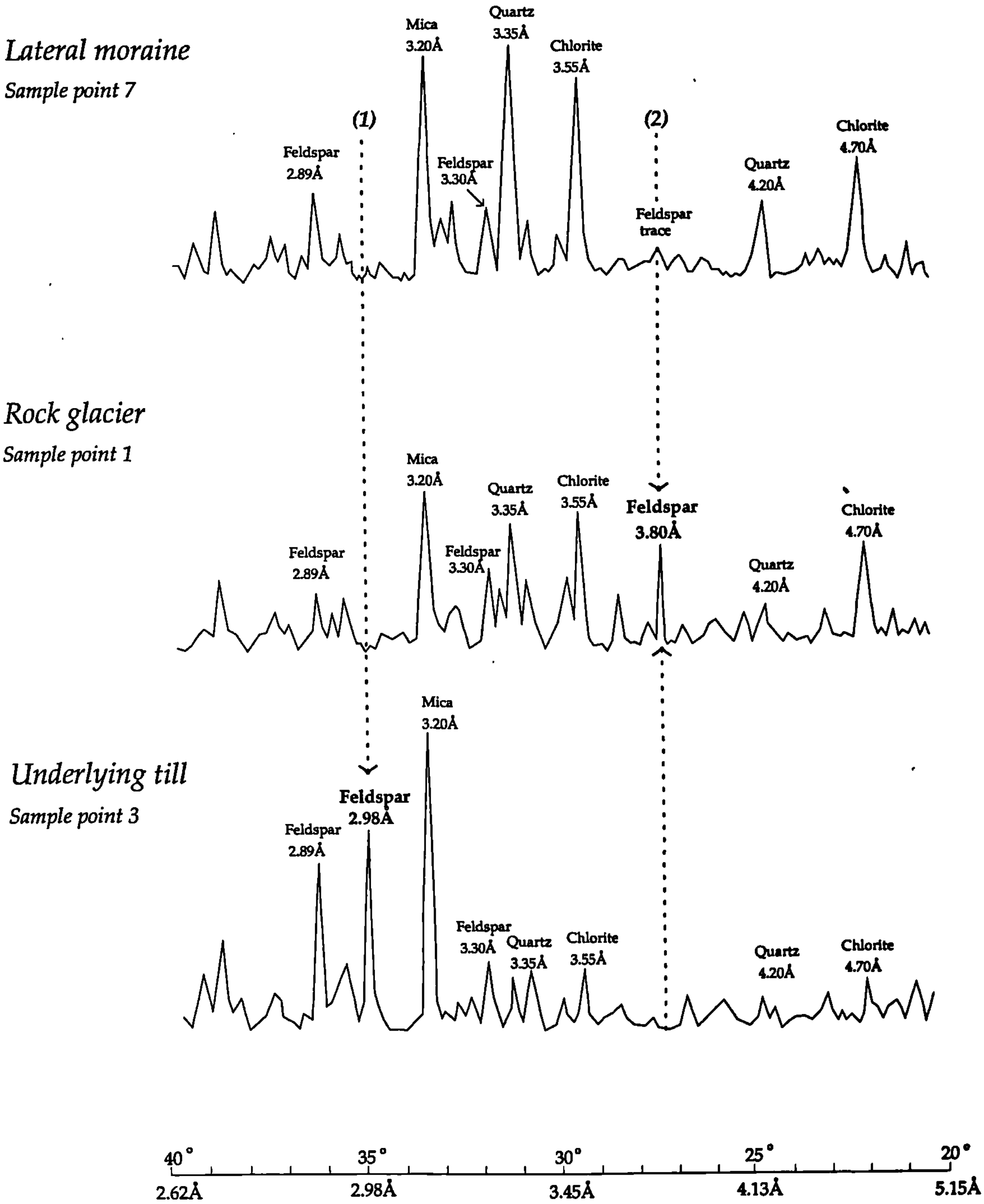


Figure 6.16 X-Ray diffraction traces of fine sediments (<2µm) from a lateral moraine fragment, rock glacier and underlying till deposit at Fornesdalen, northern Norway. Location of sample points is given in Figure 6.9.

feldspar peak at 3.80Å on the rock glacier trace is absent from the till trace. The only notable difference between the rock glacier and lateral moraine traces is that whereas the 3.80Å feldspar peak on the former is pronounced, the same peak on the latter is very subdued. Thus, it appears that there are greater differences between the underlying till and the rock glacier than between the lateral moraine and rock glacier. One possible explanation for this is that lateral moraines and valley-wall rock glaciers consist generally of angular fragments entrained supraglacially from rockfall, whereas till consists generally of subglacially-abraded material with rather different source areas upvalley.

Table 6.1 summarises the concentrations of ten major elements (expressed as oxides) in the various samples, as determined by X-ray fluorescence spectrography. The results, which are shown as percentages, are presented for rock glacier samples obtained from FE1 and FE2 in columns 1 and 2 respectively, for samples from both rock glaciers in column 3, and for the lateral moraine and till samples in columns 4 and 5 respectively. Concentrations of the lowermost four elements shown in Table 6.1 show no apparent differences between the three types of deposits. However, concentrations of the upper six elements indicate that the till deposits are chemically distinct from the rock glacier fines. In particular, till samples contain lower concentrations of silica, aluminium, and potassium, and higher concentrations of magnesium and calcium than either the rock glacier samples or the lateral moraine sample. In addition, sodium concentrations are lower in till samples than in rock glacier samples

X-RAY FLUORESCENCE

Oxide	*Rock glacier Fornesdalen FE1 n=5* %	*Rock glacier Fornesdalen FE2 n=2* %	*Rock glaciers FE1 +FE2 n=7* %	*Lateral moraine n=1* %	*Underlying till n=4* %
SiO_2	48.55 S.D. = 0.47	48.53 S.D. = 0.61	48.54 S.D. = 0.46	48.17	45.93 S.D. = 0.91
Al_2O_3	18.71 S.D. = 0.53	17.58 S.D. = 1.23	18.39 S.D. = 0.86	19.72	16.13 S.D. = 1.78
MgO	9.36 S.D. = 1.01	9.90 S.D. = 1.36	9.51 S.D. = 1.03	9.13	11.99 S.D. = 1.06
CaO	11.33 S.D. = 0.27	12.24 S.D. = 0.85	11.59 S.D. = 0.60	11.06	13.88 S.D. = 0.79
K_2O	0.24 S.D. = 0.08	0.18 S.D. = 0.03	0.22 S.D. = 0.07	0.16	0.13 S.D. = 0.03
Na_2O	1.01 S.D. = 0.09	1.10 S.D. = 0.14	1.03 S.D. = 0.10	0.84	0.90 S.D. = 0.12
Fe_2O_3	10.97 S.D. = 0.76	9.85 S.D. = 0.08	10.65 S.D. = 0.83	11.01	10.34 S.D. = 0.58
TiO_2	0.49 S.D. = 0.08	0.38 S.D. = 0.13	0.46 S.D. = 0.10	0.50	0.42 S.D. = 0.07
MnO	0.16 S.D. = 0.02	0.15 S.D. = 0.01	0.15 S.D. = 0.01	0.16	0.15 S.D. = 0.01
P_2O_5	0.12 S.D. = 0.04	0.09 S.D. = 0.01	0.11 S.D. = 0.04	0.07	0.11 S.D. = 0.03

Table 6.1 X-Ray fluorescence spectography results for rock glacier, lateral moraine and till samples.

although they are similar to the sodium concentrations in the lateral moraine sample. These results suggest, therefore, that till and rock glacier samples have distinctly different concentrations of major elements in their composition, whereas the chemical composition of rock glacier and lateral moraine samples is more similar.

In sum, evidence obtained from X-ray diffraction and X-ray fluorescence analyses indicates that rock glacier sediments at FE1 and FE2 are chemically distinct from underlying till deposits. It therefore appears most unlikely that outcrops of fine sediments observed and sampled on rock glaciers FE1 and FE2 represent inliers of underlying till. Chemical evidence, however, does not negate the possibility that fines from lateral moraines may have become incorporated within the rock glaciers during formation as the major element concentrations and clay mineralogy of the lateral moraine and rock glacier sediments are largely similar. It is likely that most fine sediments that occur within valley-wall rock glaciers have been inherited from pre-existing talus deposits, although clast abrasion, particularly within the surface boulder layer, may produce additional fine material during rock glacier movement.

6.3.5 Summary

Three major conclusions are suggested by the research reported in this section.

1) Beneath a surface boulder mantle of variable depth (0 to >2 metres), fine material forms a matrix in which clasts up to ~60cm in size are embedded.

2) Rock glacier fine sediments sampled in Fornesdalen, northern Norway, are relatively coarse containing less than 1% by weight of clay-sized particles and between 4 and 20% by weight of silt-sized particles; they are therefore at the lower limits of frost susceptibility.

3) The majority of fine material within rock glaciers has probably been inherited from talus slopes although field and laboratory evidence suggests that in some cases lateral moraines may become incorporated within rock glaciers as they move downslope.

6.4 Rock glacier ice

6.4.1 Introduction

The third principal constituent of active and inactive rock valley-wall rock glaciers is ice. As discussed in section 2.5 above, the nature and origin of ice found within rock glaciers is largely unknown; some researchers believe that the ice is glacially-derived (e.g. Whalley, 1974; Lindner & Marks, 1985), others suggest that the ice represents buried snowbank ice (e.g. Liestøl, 1962), while many maintain that refreezing of snow and percolating meltwater or ice segregation explain its origin (e.g. Wayne, 1981; Haeberli, 1985). Indeed, the variety of hypotheses that have been proposed to explain the formation of valley-wall rock glaciers (cf. Chapter 3) has arisen largely because the precise origin of internal ice is unknown.

The main aim of the research reported in this section is to attempt to ascertain the origin of ice found within valley-wall rock glaciers by comparing the isotopic content of rock glacier ice with ice of known origin. Two commonly measured environmental isotopes are deuterium (^{2}H) and oxygen (^{18}O), which are simply atoms of hydrogen and oxygen, respectively, with nuclei of slightly different atomic weights. In this analysis, deuterium concentrations are calculated for ice samples collected from active and inactive valley-wall rock glaciers in Switzerland together with ice from neighbouring glaciers, snowbanks and ice-rich frozen sediments. Deuterium concentrations of approximately one part per million can be determined by using sensitive mass-spectrometric measuring methods combined with detailed laboratory preparation of the water samples. A number of papers have been published that are concerned with stable isotopic analyses of ice, snow and water (e.g. O'Neil, 1968; Ambach, 1976; Grabczak *et al.*, 1983; Souchez & Jouzel, 1984; Boulton, 1986; Stichler, 1987), although ice isotopic procedures have not hitherto been used to ascertain the origin of rock glacier ice. Lorrain and Demeur (1985) however, used isotopic ratios to discriminate among possible ice origins for ground ice masses in the Canadian Arctic and were able to identify buried glacier ice that was locally penetrated by a more recent ice wedge. It is known (e.g. Stichler,

1987) that in natural waters $\delta\,^2H$ and $\delta^{18}O$ values obey the following general relationship:

$$\delta\,^2H = m\,\delta^{18}O + t$$

For waters that have not been subjected to evaporation the value of m is approximately 8 and the average value of t is 10 for precipitation in the northern hemisphere. Water samples that have been affected by evaporation will represent a line with a slope (m) which is usually about 5. Thus, better discrimination of water samples is often obtained by calculating both deuterium (^2H) and oxygen (^{18}O) values. However, due to operational difficulties at the laboratory where the analyses were carried out, only deuterium concentrations could be calculated for the ice samples.

In the remainder of this section, sampling procedures and laboratory methods are discussed before the results of the ice isotopic analyses are presented. Variations in the deuterium isotopic contents of the different ice groups are interpreted in the light of our current understanding of the processes that control isotopic enrichment and depletion. Differences between the ice groups are examined statistically and implications for the origin of rock glacier ice are discussed. The section concludes with a brief evaluation of the usefulness of an isotopic approach when comparing and contrasting ice samples of different origin.

6.4.2 Ice collection and laboratory methods

Ice sampling sites, which are located within two small study areas in Arolla, Valais, and Flüela-Pass, Graübunden, Switzerland (cf. Chapter 4), were determined principally by the initial selection of an active rock glacier site suitable for ice collection. Once rock glacier ice was obtained, samples were collected from neighbouring glaciers, snowbanks and ground ice. In addition to these four main ice types, ice from an inactive rock glacier was sampled and water samples were collected from a nearby lake to provide an estimate of the isotopic composition of present-day surface waters in the region. Lake-water samples are thought to be more representative of local conditions than precipitation, as the general mixing of waters in lake catchment areas reduces the spatial and temporal variations in deuterium concentrations that are present in precipitation samples (Ambach *et al.*, 1976).

Sampling sites were grouped into six main classes, namely snowbank ice, glacier ice, ground ice, ice from an active rock glacier, ice from two inactive rock glaciers and lake-water. At each sampling site, several individual samples were obtained for ice isotopic analysis. At glacier sites, an area of ice was cleaned prior to sampling and samples were obtained at varying depths from the glacier surface and also from moulins and crevasses up to 4.2m below the glacier surface. Similarly, on snowbanks, snow and ice samples were obtained from depths of up to 2 metres, and pits were excavated on several talus slopes in order to obtain ground ice samples. More extensive excavations, principally in the frontal slopes of both active and inactive rock glaciers, yielded samples of rock glacier ice (Plate 6.4). Ice obtained from the active valley-wall rock glacier Arolla 3, forms a matrix with and supports fine sediments and small clasts. In Plate 6.5, it can be seen that clear ice up to 2cm thick appears to surround each clast so that clast-to-clast contact is often removed; thus the ice is not simply interstitial or pore ice. No evidence of larger ice lenses was found, although the limited nature of the excavations may not have produced a truly representative sample; the bouldery surface layer of the rock glacier prevented all attempts at excavation elsewhere on the surface of the rock glacier.

Immediately following collection, the sampled ice was placed in covered containers and was allowed to melt entirely. The water was filtered before being placed in self-sealing screw-top 30ml glass bottles that contained a minimum amount of air so that fractionation of the water samples was kept to a minimum. Glass bottles were used to avoid possible slight diffusion through the walls. At each sampling site several characteristics were noted including underlying lithology, sampling depth, site altitude, the slope of the ice surface and a description of the ice. Wherever possible ice samples were obtained from sites that are underlain by the same lithology. All samples were kept as cool as possible in the field and on return were stored at 4°C until analysed in the laboratory.

The isotopic analyses were carried out at the Scottish Universities Research and Reactor Centre at East Kilbride. The basic procedure for obtaining deuterium from water involves passing a gaseous sample of water over a uranium furnace in a high pressure vacuum line, so that the oxygen in the water sample reacts with the uranium to produce uranium oxide leaving hydrogen that can then be analysed in a mass-spectrometer to obtain the relative proportions of the deuterium isotope and the other hydrogen isotopes. Initially, therefore, the water samples were placed in pre-calibrated 5 µl pipettes. To run a sample, the pipette containing the sample was placed in a high-pressure vacuum line. Following evacuation of air from the vacuum line, the pipette was

Plate 6.4 Excavations in the frontal slope of Arolla 3 active valley-wall rock glacier. Ice was reached at depths of approximately 1.5 metres. Note the very steep gradient of the frontal slope and the predominance of fine material.

heated until bubbles could be seen in the sample at which point the sample was released into the vacuum. The gaseous sample was drawn along the vacuum line towards a liquid nitrogen trap. After five minutes, liquid nitrogen cooling had frozen the majority of the sample, and it was possible to extract any non-condensable gases present in the sample. The liquid nitrogen trap was then removed and the sample was drawn through a uranium furnace allowing the oxygen in the gas to react with the uranium. The gas was passed through the uranium furnace for a second time to remove all oxygen. The remaining hydrogen was flamed round the vacuum line and was collected in a charcoal bottle. The hydrogen isotope ratios of the samples were then calculated using a Micromass 602 mass spectrometer and the amount of deuterium was measured. Precision of the measurements using this technique was estimated to be approximately ± 1.3°/oo by a series of repetition tests.

6.4.3 Results and implications

All isotope results are expressed as δ–values *(per mil (°/oo) deviation* from Standard Mean Ocean Water (SMOW) value). As indicated in section 6.4.2 above, samples were collected from six ice groups, namely snowbank ice, glacier ice, ice from an active rock glacier, ice from inactive rock glaciers, ground ice and lake-water. At each sampling site several individual samples were collected from a variety of depths. The mean δ–values of each sampling site are summarised in Table 6.2, together with the mean δ-values for each of the six ice groups. Standard deviations of

ICE ORIGIN	*n*	*MEAN VALUES OF $\delta\ ^2H$*
Snowbank ice - 3 snowbanks	16	-138.5
Site 1	6	-136.2
Site 2	7	-140.8
Site 3	3	-137.7
Glacier ice - 3 sampling sites	15	-123.3
Site A	5	-126.9
Site B	5	-119.6
Site C	5	-123.5
Ground ice - 2 talus sites	6	-121.8
Site A	3	-118.7
Site B	3	-124.9
Active rock glacier ice		
1 rock glacier	9	-120.1
Lake water control		
1 lake	3	-111.6
Inactive rock glacier ice - 2 sites	8	-104.6
Site A	4	-113.6
Site B	4	-95.6

Table 6.2 Mean deuterium δ-values for each sampling site and each ice group.

mean δ-values and additional descriptive statistics for each ice group are summarised in Table 6.3. From these statistics it is clear that snowbank ice exhibits the greatest negative shifts from SMOW, which signifies that it is most depleted in deuterium. Conversely, those ice samples that are most enriched in deuterium (in relative terms) fall within the inactive rock glacier group.

In order to interpret the δ-value variations of the sampled ice, some understanding of the processes involved must be gained. It has long been established that the fractionation of deuterium (^{2}H) and oxygen (^{18}O) isotopes occurs during the freezing of water and the melting of snow and ice (e.g. O'Neil, 1968). The concentrations of stable isotopes are governed by fractionation phenomena which in snow and ice may occur not only at the phase boundary liquid-vapour but also at the phase boundaries solid-liquid and solid-vapour. Four main conditions and processes appear to control the isotopic content of an ice sample:

1) the isotopic content of the initial precipitation;

2) melting and evaporation which results in heavy isotopic enrichment of ice as less energy is required to evaporate lighter isotopes so the relative amount of heavy isotopes such as deuterium will increase;

3) freezing and condensation which results in the depletion of heavy isotopes in ice as air moisture generally has a higher percentage of lighter isotopes;

Ice groups: summary statistics	n	Mean values of δ D	Median values of δ D	Standard deviation	Minimum	Maximum	Range
Snowbank ice	16	-138.5	-137.2	5.54	-149.9	-131.3	18.6
Glacier ice	15	-123.3	-124.1	4.89	-130.6	-115.2	15.4
Active rock glacier ice	9	-120.1	-121.1	3.26	-123.8	-115.4	8.4
Inactive rock glacier ice	8	-104.6	-106.4	10.73	-116.0	-88.7	27.3
Ground ice	6	-121.8	-122.8	4.90	-129.0	-115.3	14.7
Lake water	3	-111.6	-112.4	2.09	-113.1	-109.2	3.9

Table 6.3 Descriptive summary statistics of deuterium δ-values for each ice group.

4) the movement of water vapour within snow and ice due to temperature gradients which results in a vertical and horizontal redistribution of the heavy isotopes;

Initially, within-group variations and absolute group mean δ-values for each ice group are interpreted before the statistical significance of δ-value differences between each ice group is examined. The lake-water samples, which were obtained from a lake at an altitude of 2383m in Flüela-Pass, give an indication of the isotopic content of local surface waters. The mean value of the sampled lake-water is -111.6°/oo with a standard deviation (s.d.) of 2.09. Similar lake-water values have been reported for high-altitude lakes elsewhere in the European Alps. For example, δ-values of between -123°/oo and -115°/oo were obtained for lake-water samples near the Grubengletscher in the Wallis Alps, Switzerland (Souchez & De Groote, 1985). The marginally more negative δ-values recorded at Grubengletscher may possibly be the result of more prolonged lake freezing as freezing will make residual waters more negative (Souchez & De Groote, 1985). In addition, Stichler (1987) has suggested that lower condensation temperatures produce a decrease in the amount of deuterium. As the altitude of the ice-dammed lake at Grubengletscher is higher (2950m) than at Flüela-Pass (2374m), it is not surprising that lower deuterium concentrations are found at Grubengletscher. The small range of δ–value variations of the lake samples (3.9°/oo) is likely to be the result of homogenisation effects of lake-water sample size.

The median and inter-quartile distributions of each ice group, which are plotted in Figure 6.17, clearly show that the δ-values of snowbank ice are the most negative (mean = -138.5°/oo, s.d. = 5.54). As a general rule, isotopic variations in fresh snow depend primarily upon the initial content and the condensation temperature of the precipitation. Stichler (1987) has calculated that because of the temperature dependence of isotope fractionation during condensation, a 100m increase in elevation will cause a decrease in deuterium of about -4 ± 2°/oo. However, this altitudinal effect tends only to be found in fresh snow; the isotope content of pack snow samples may be substantially different to the original isotope content of the snow because of melt, refreezing and sublimation processes. Of the three snowbanks that were sampled, two are located at an altitude of 2250m in the Arolla valley and one at an altitude of 2600m at Flüela-Wisshorn. Samples were obtained from the Arolla snowbanks in early July 1987 when the maximum ablation rate of a neighbouring snowpatch was measured at 31cm per day (Finlay, 1987). Thus, although each of the three snowbank sites is thought to contain perennial ice, it is likely that samples obtained from the two snowbanks at Arolla at depths of less than one metre accumulated solely during the preceding winter. Ice collected from the third snowbank at Flüela-Wisshorn from a depth of about 2 metres may have survived the previous summer melt season.

The pronounced negative shifts from SMOW (and from the local lake-water) of the snowbank ice samples suggest that the snowbanks

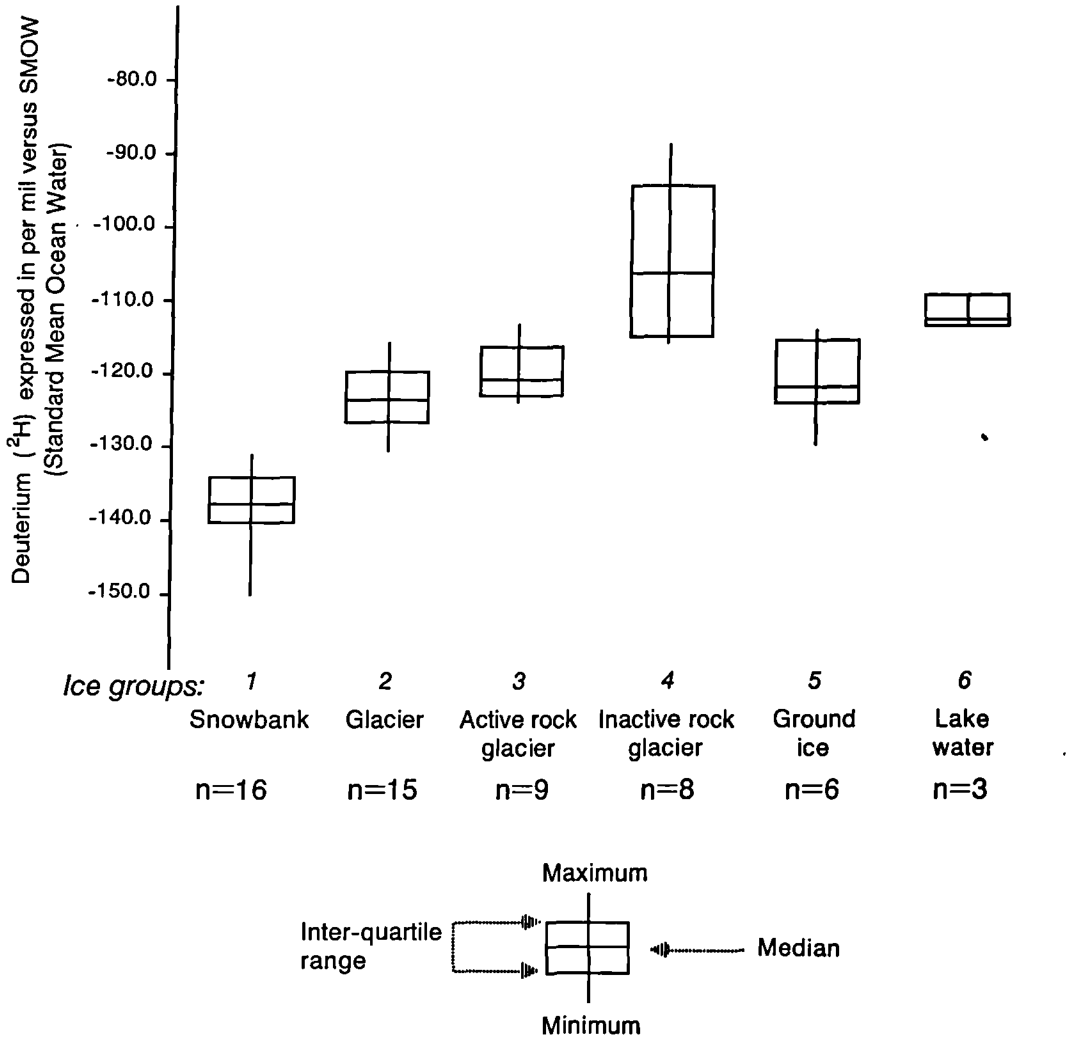


Figure 6.17 Median and inter-quartile distributions of deuterium values for each ice group.

(1983), which were obtained from greater depths in the glacier correspond more closely with the glacier samples obtained from crevasses and moulins in the Bas Glacier d'Arolla.

Isotopic variations in temperate Alpine glaciers appear to be much more complicated than those on cold glaciers of the Greenland and Antarctic ice sheets where measurements of stable isotopes have been used as indicators for dating and computing accumulation rates (Reeh *et al.*, 1978). The precipitation on polar ice sheets falls only as snow and melting of the deposited snow is practically negligible. For temperate glaciers, on the other hand, the effect of melting and evaporating processes as well as the influence of meltwater seepage through the snow and the firn is very important as there is an isotope-exchange process between the snow-firn matrix and percolating water. After the snow is deposited, the deuterium excess of the snowcover will be changed by isotope fractionation processes at the surface which involve primarily evaporation and condensation of atmospheric moisture. As shown above, the uppermost annual layer contains the most isotopically depleted samples. The underlying strata are characterised by a stepwise enrichment of heavy-isotope content (Grabczak, *et al.*, 1983) until the layers become more homogeneous particularly in the transition zone from snow to ice. Processes of fractionation between percolating meltwater and firn cause homogenisation accompanied by an enrichment of heavy isotopes. The depths of the sampled glacial ice and its associated d- values are shown in Figure 6.14. As is to be expected,

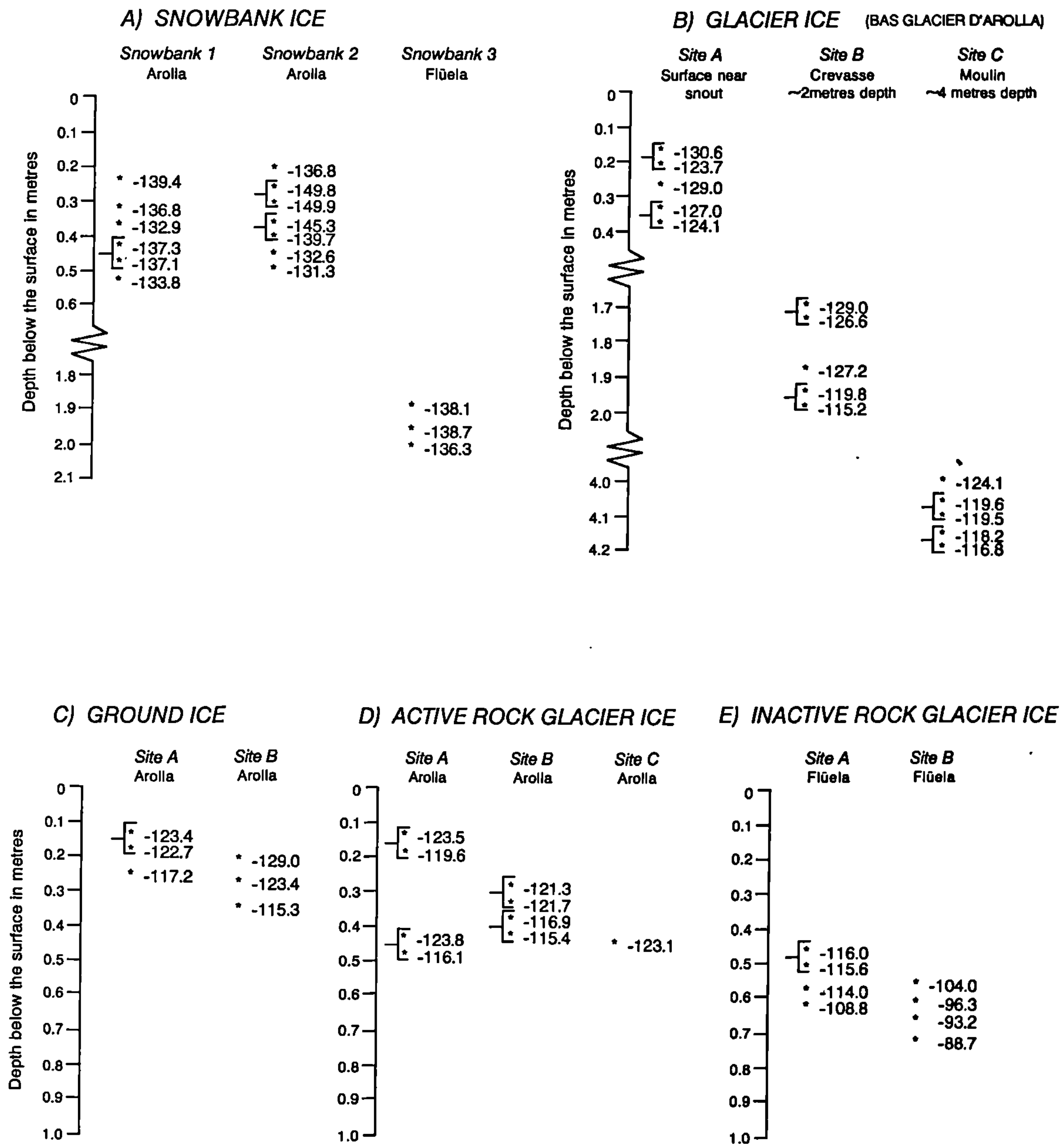


Figure 6.18 Variations in deuterium δ-values with depth for each ice group.

have undergone melt-freeze cycles at the surface and repeated refreezing of meltwater at depth. At such high altitudes it is likely that diurnal freeze-thaw cycles would be pronounced, particularly for several weeks following initial snowfall before daily temperatures fell constantly below 0°C. In snowbanks there is often an enrichment in heavy isotopes with increasing depth, which Stichler (1987) concluded to be the result of the evaporation of snow. In Figure 6.18 the depth of the snowbank samples and their associated d-values are presented. The samples taken from the snowbank site at Arolla 2 show a marked decline with depth except for the uppermost sample; the uppermost enriched deuterium value may be due to diurnal freeze-thaw cycles. The pattern at Arolla 1 is less clear although the most negative value is uppermost. At Flüela, only three samples were measured from clear ice at a depth of 2 metres where a stream had cut through a perennial firn field.

The mean d-value of the glacial ice samples, which was calculated from 15 samples from the Bas Glacier d'Arolla, is -123.3°/oo with a standard deviation of 4.89 and a range of 15.4°/oo from -130.6°/oo to -115.2°/oo. Grabczak *et al.*, (1983) obtained d ^{2}H values of -140 °/oo to -90°/oo relative to SMOW for glacier ice samples obtained from crevasses in the Grubengletscher in the Swiss Alps. However, their d-values that approach -140°/oo relate to samples that were obtained from the surface layers of the glacier which comprise recent falls of snow rather than firn or ice. Thus, these values not unsurprisingly correlate with the snowbank samples, whereas the less negative values of Grabczak *et al.*

samples obtained from greater depths in crevasses and moulins (sites B and C) are generally less depleted in deuterium than samples obtained from nearer the surface (site A).

The third ice group, ground ice, which exhibits the least depleted deuterium concentrations compared with snowbank and glacier ice, has a mean δ-value of -121.8°/oo, a standard deviation of 4.90 and a range from -129.0°/oo to -115.3°/oo relative to SMOW. Ground ice samples were obtained from depths of approximately 18 to 35cm beneath the surface of talus (Figure 6.18). The ice was in the form of pore ice and is likely to have formed by the refreezing of nival meltwater percolating within the talus, a process that is generally accompanied by an enrichment of heavy isotopes. In addition, freezing and condensation processes within the talus will not cause deuterium depletion of the pore ice to the same degree as they would for snow near the surface of snowbanks, so δ-values are less negative than in snowbanks.

Ice samples obtained from two valley-wall rock glaciers at Flüela-Wisshorn, both of which are believed to be inactive, are quite similar in their physical characteristics to those of ground ice found in talus as the ice generally appears to be pore ice that fills the interstices. However, the δ-values of the inactive rock glacier group are quite dissimilar to those of the ground ice group as deuterium concentrations are much lower. The mean δ-value of ice in these inactive rock glaciers is -104.6°/oo with a standard deviation of 10.73 and a very wide range from -116.0°/oo to

-88.7°/oo relative to SMOW. It is possible that the least negative δ-values of the inactive rock glacier group represent older ice as homogenisation effects of percolating meltwater over several years may result in progressively enriched deuterium. The talus ground ice, on the other hand, may have formed from the refreezing of only one years snowmelt. Supporting evidence for this tentative hypothesis comes from the depth of the sampled ice; the ice from the inactive rock glaciers was obtained at depths up to 75cm whereas the ground ice in talus was obtained from much shallower depths (Figure 6.18). The most enriched rock glacier ice samples are at greatest depth.

In contrast to the relatively enriched ice from inactive rock glaciers, the mean δ-value of the 9 active rock glacier samples is -120.1°/oo with a standard deviation of 3.26 and a much smaller range of only 8.4°/oo. Moreover, the ice characteristics of the two groups are noticeably different. For example, ice obtained from the active valley-wall rock glacier Arolla 3 (Plate 5.16,), forms a matrix with and supports fine sediments and small clasts. In Plate 6.5, it can be seen that clear ice up to 2cm thick appears to surround each clast so that clast-to-clast contact is often removed; thus the ice is <u>not</u> simply interstitial or pore ice. Instead an excess of moisture must have been present for a greater volume of ice to form in the rock glaciers. No evidence of larger ice lenses was found, although it should be emphasised that the limited nature of the excavations may not have produced a truly representative sample; the bouldery surface layer of the rock glacier prevented all attempts at

Plate 6.5 a and b. Samples of ice-rich frozen sediments obtained from the active rock glacier, Arolla 3. Note the development of clear ice that supports small clasts.

excavation elsewhere on the surface of the rock glacier.

Before any attempt is made to infer the origin of active valley-wall rock glacier ice, it seems appropriate to determine if the ice groups are statistically independent of one another in terms of their deuterium concentrations. In Table 6.4, the results of Mann-Whitney U tests that examine the significance of differences between the sample medians of the δ-values of each ice group are presented. The null hypothesis of the tests is that paired samples are not significantly different. Tests are regarded as significant at the 95% level or above and the significance level at which the null hypothesis can be rejected for each test is indicated in Table 6.4.

A general glance at Table 6.4 shows that for the majority of tests the null hypothesis is rejected; thus statistically significant differences exist between most of the sample groups. There are three exceptions. The first of these involves ice samples from inactive rock glaciers and lake-water samples. Figure 6.17 shows that the three lake-water δ-values fall entirely within the larger δ-value range of inactive rock glaciers, although deuterium concentrations in inactive rock glaciers tend to be slightly higher on average. Clearly, in a comparative study of genetically-distinct ice types the overlap of these two groups is rather insignificant in terms of identifying ice origin.

Statistical differences between two sample sets using the Mann-Whitney U Test	Snowbank	Glacier	Active rock glacier	Inactive rock glacier	Ground ice	Lake water
Snowbank ice	-	-	-	-	-	-
Glacier ice	Reject H_0 at 0.0001	-	-	-	-	-
Active rock glacier ice	Reject H_0 at 0.0001	Reject H_0 at 0.001	-	-	-	-
Inactive rock glacier ice	Reject H_0 at 0.0001	Reject H_0 at 0.001	Reject H_0 at 0.01	-	-	-
Ground ice	Reject H_0 at 0.0005	No significant difference at 0.05	No significant difference at 0.05	Reject H_0 at 0.01	-	-
Lake water	Reject H_0 at 0.01	Reject H_0 at 0.01	Reject H_0 at 0.05	No significant difference at 0.05	Reject H_0 at 0.05	-

Mann-Whitney U test for non-parametric data:

H_0: median X = median Y

H_a: median X $\neq$ median Y

Table 6.4 Mann-Whitney U Test results of the differences between each ice group.

The two other examples where the null hypothesis was rejected both involve ground ice. In the first, ground ice and glacier ice could not be distinguished on the basis of deuterium levels within the ice. Given the markedly different origins of these ice types this lack of separation is unhelpful. In addition, there is no statistical difference between ground ice and the ice samples from the active rock glacier.

The chemical and physical evidence of the different ice groups, when taken together, allows some tentative conclusions to be drawn regarding the origin of rock glacier ice. No evidence was found in any rock glacier for a buried ice core or large clear lenses of ice. Instead, the ice in the active rock glacier is in the form of a matrix that supports fine sediments and separated clast-to-clast contacts. The volume of ice within the inactive rock glaciers, however, appears to be less as the ice generally fills only the interstices. Deuterium concentrations in ice taken from the active rock glacier are most similar to those of ground ice and least similar to those of the snowbank ice group. The δ-values of the inactive and active rock glacier groups overlap each other and both overlap with the δ-values of the ground ice group. These results are therefore consistent with the hypothesis that rock glaciers contain perennially frozen sediments rich in ice that form largely through the refreezing of percolating snow and meltwater. However, in order to enable the formation of matrix ice rather than interstitial ice, excess moisture must have been present. Given this requirement, it is possible that segregation processes, particularly near the top of the permafrost table, may also have operated. Finally, as the glacier ice samples are statistically significantly different from both the active and inactive rock glacier ice samples (Table 6.4), the hypothesis that valley-wall rock glaciers contain glacier ice cores is inconsistent with these results.

6.4.4 Conclusions

The isotopic measurements presented here constitute, up to the present time, the only comparative measurements to have been made on rock glacier ice, glacier ice, snow and ice from perennial firn fields, ground ice and lake water. They allow us to draw the following conclusions:

1) The majority of the ice groups are isotopically distinct in terms of their deuterium concentrations.

2) The ice that was obtained from the active rock glacier was found to form a matrix with and support small clasts and sediments. If this ice is representative, ice volumes in active rock glacier samples are greater than in pore ice samples. Deuterium concentrations in the ice obtained from the active rock glacier are most similar to those of ground ice and least similar to those of the snowbank group. In addition, deuterium concentrations of the ice samples from the active rock glacier are statistically significantly different from those of the glacier ice samples. These results therefore, are inconsistent with the hypothesis that valley-wall rock glaciers contain glacier ice-cores.

3) The volume of ice found within the inactive rock glaciers appears to be less than that within the active rock glacier, and generally fills only the interstices. Deuterium concentrations within the ice samples of the active and inactive rock glacier groups overlap.

4) Taken together, the physical and chemical evidence strongly suggests that rock glaciers contain perennially frozen sediments rich in ice, the ice having formed mainly through the refreezing of percolating snow and meltwater. However, given that clasts are separated and supported by ice, ice segregation processes probably contribute to the development of rock glacier ice.

5) Deuterium concentrations in snowbank ice, particularly in annual surface layers, are markedly depleted relative to local surface waters because of the fractionation processes involved in freezing and condensation. In addition, deuterium concentrations become gradually enriched with increasing depth.

6) Deuterium concentrations in glacial ice also exhibit a general enrichment with increasing depth, although mean δ-values of glacial ice tend to be less depleted than snowbank ice. This probably reflects homogenisation effects through time, possibly

caused by fractionation between percolating meltwater and firn being accompanied by an enrichment of deuterium.

7) Shallow pore ice samples obtained from talus slopes are more similar than either snowbank or glacial ice to present-day local waters, although they remain depleted in heavy isotopes because of the refreezing of percolating snow or meltwater and the interaction between this meltwater and existing snow and ice. Deuterium concentrations, once again, increase with increasing depth from the surface.

8) Pore ice samples obtained from inactive rock glaciers are more enriched in deuterium than any other ice group. As outlined above, refreezing of percolating meltwater within the ground and processes of fractionation between percolating meltwater and existing pore ice cause deuterium enrichment. It seems possible that progressively greater enrichment would occur through time so the inactive rock glacier samples may represent older ice than that obtained from shallower depths in talus.

Clearly, additional experimental data is needed to clarify the above conclusions. In particular, measurement of both deuterium (^{2}H) and oxygen (^{18}O) would undoubtedly provide better discrimination between the ice groups. Partial separation of different ice groups on the basis of deuterium concentrations alone, however, has shown the potential for an isotopic approach when comparing genetically distinct ice groups.

6.5 Morphologically-similar features

6.5.1 Introduction

Valley-wall rock glaciers have in the past proved difficult to distinguish from morphologically similar but genetically-distinct landforms such as protalus ramparts, avalanche boulder tongues, landslide deposits, and ice-cored moraines (cf. Østrem, 1971; White, 1981; Lindner & Marks, 1985). This section considers those landforms that are most commonly confused with rock glaciers in the field and discusses the possibility of identifying morphological and other criteria that may be used to distinguish each type of landform.

6.5.2 Protalus ramparts

Valley-wall rock glaciers have often been mistaken for protalus ramparts (e.g. Blagbrough & Breed, 1967; Sissons, 1979; Lindner & Marks, 1985). Protalus ramparts are ridges or ramps of predominantly coarse detritus, usually located at or near the foot of talus, that have formed through the accumulation of debris along the downslope margins of perennial snowbanks or firn fields (cf. Ballantyne, 1987a). Protalus ramparts, therefore, are depositional landforms that accumulate by a series of rockfalls and snow avalanches. Locational and sedimentological criteria are virtually identical for both valley-wall rock glaciers and protalus ramparts; both form at the feet of talus slopes and both contain rockfall debris. The main distinction is that protalus ramparts do not move, rather they accumulate as depositional landforms. In a study of nine Lateglacial protalus ramparts in upland Britain, Ballantyne & Kirkbride (1986) found that the distance from the frontal ridge crest to the foot of the adjacent talus slope was limited, and lay between 25 and 67m. In addition, they noted that rampart length rarely exceeded 300m and that the ramparts they studied possess only a frontal ridge with no internal transverse ridges. In contrast, valley-wall rock glaciers in this study extend up to 600m from the feet of talus slopes, possess up to thirteen inner transverse ridges, and their lateral extent is often much greater than 300m. Morphologically, however, the frontal ridges of protalus ramparts and valley-wall rock glaciers show similar characteristics. Ballantyne (1987) noted that on two active protalus ramparts on the Lyngen Peninsula, northern Norway, frontal ridges comprise four typical units, namely a basal talus apron, a middle rectilinear slope, a ridge crest and a gentler proximal slope. The gradients of these slopes are often less steep than those on active valley-wall rock glaciers, although the range of mean gradients of the frontal slopes of rock glaciers and protalus ramparts are not markedly different; mean frontal slope gradients for valley-wall rock glaciers lie between 34° and 41°, whilst those for protalus ramparts lie between 34° and 39.5°. The main distinguishing characteristic appears to be the maximum facet angle of the frontal slope, which on active valley-wall rock glaciers reaches 53°, but does not exceed 43° on active protalus ramparts. This important difference arises because the frontal slopes of protalus ramparts represent "repose slopes formed by the accumulation of cohesionless cascading debris" (Ballantyne & Kirkbride, 1986, p. 664), whereas the frontal slopes of active valley-wall rock glaciers are experiencing differential forward motion, producing very steep gradients near the top of the frontal slope with maximum facet angles at the angle of initial yield.

In view of the differences in dimensions identified above, the greatest difficulties are most likely to be encountered when attempting to discriminate between small inactive or relict valley-wall rock glaciers that lack any inner transverse ridges and protalus ramparts. Clearly, active protalus ramparts require the presence of a perennial snowbed upslope and, as such, are easily identifiable. Once inactive, however, a protalus rampart may be morphologically indistinguishable from a small inactive valley-wall rock glacier that does not contain any inner transverse ridges.

Plate 6.6 shows a small inactive valley-wall landform, Schwarzhorn 4, in Graubünden, Switzerland. Despite an absence of any inner transverse ridges the feature is interpreted as a valley-wall rock glacier because the distance between the crest ridge and the talus exceeds 80m. Haeberli (1985, p. 13), however, has suggested that this landform is a feature of "initial rock glacier development (protalus rampart)". His interpretation highlights an added complication, in that several researchers (e.g. Liestøl, 1962; Corte, 1976; Haeberli, 1985;

Plate 6.6 Small valley-wall rock glacier at Schwarzhorn, Switzerland, which Haeberli (1985) believes has evolved from a protalus rampart.

cf. section 3.2.5 above) believe that valley-wall rock glaciers may evolve from protalus ramparts. Haeberli, and others in favour of a continuum, hypothesise that if snowbed thickening continues beyond the point where plastic deformation of ice occurs at the base of the snowbed, a valley-wall rock glacier may develop. The lateral extent of such rock glaciers, however, is likely to be constrained by the size of the initial protalus rampart. The validity of this interpretation is considered in the following chapter.

6.5.3 Ice-cored moraines

Perhaps the best-documented controversy concerning rock glacier identification is the debate between Østrem (1971) and Barsch (1971) regarding the nature of ice-cored moraines. To a large extent, this debate focused on the difficulties involved in differentiating between ice-cored moraines and valley-floor rock glaciers that form downslope from glaciers. However, the debate has relevance to valley-wall rock glacier identification, particularly in relation to ice-cored lateral moraines.

Barsch believed that all ice-cored moraines described by Østrem are analogous to rock glaciers in the Alps and in the southern Rocky Mountains. He also considered that the origin of ice-cored moraines proposed by Østrem, in which debris is loaded on to a permanent snowbed by glacier advance, is incorrect, as each ridge in the resulting 'ice-cored moraine' could not be related to a specific readvance event at the glacier's snout. Instead, Barsch suggested that an end moraine downslope of a glacier may contain residual glacier ice and that if the amount of ice and morainic debris is sufficient, the lower layers of the mass may deform plastically downslope. Thus, the ridges on Barsch's 'rock glaciers' are directly related to internal movement. Østrem, on the other hand, believed that the main confusion centred on whether the landform was moving or had moved in the past. He conceded that if his 'ice-cored moraines' moved downslope independently, they could then be regarded as 'rock glaciers'.

There are many end or lateral moraines that contain ice but have not moved downslope either because debris overburden is insufficient or because surface slope gradient is too low (e.g. Østrem & Arnold, 1970; Boulton & Eyles, 1979; Matthews & Petch, 1982). Such landforms are without doubt ice-cored moraines. However, as Barsch suggested, many other ice-cored moraines deform and move downslope and therefore should be termed rock glaciers (e.g. Griffey & Whalley, 1979; Vere & Matthews, 1985). Difficulties may arise, however, in attempting to distinguish morphological boundaries between two genetically-distinct landforms that both contain morainic debris and internal ice; clearly, locational and sedimentological criteria are similar for both types of landform. Fabric studies, however, may be used to differentiate rock glaciers from ice-cored moraines. For example, Giardino and Vitek (1985) demonstrated that clast alignment in rock glaciers tends to be approximately parallel to the trend of a surface feature such as an inner transverse ridge. They stated that such alignment is absent on ice-cored moraines and kame deposits. Perhaps the key differentiating factor, however, is the question of movement. Rock glaciers move downslope as a result of internal deformation whereas ice-cored moraines do not; they may however be subject to localised movement as a result of ice-melt. Thus, conclusive evidence to discriminate each landform may come only from detailed long-term movement studies. As a general rule, however, it is likely that multiple-ridged features greater than 100m in length are almost certainly rock glaciers. Additionally, landform configuration may indicate movement. For example, lobate-shaped landforms with crescentic transverse ridges suggest that movement has occurred (cf. Vere & Matthews, 1985). The greatest confusion probably arises over small features that contain only one or two transverse ridges. Morphologically, perhaps only three conclusive discriminating factors have come to light in the present study. The first is that transverse ridges on rock glaciers appear to be always asymmetrical. Detailed measurements of ridge morphology on ice-cored moraines (if transverse ridges exist on ice-cored moraines) are clearly required in order to establish if transverse ridge asymmetry is a valid discriminating criterion. Second, the characteristic four-unit frontal slope described in section 5.5.2 may help to discriminate between ice-cored moraines and rock glaciers, and third, maximum facet angles on active valley-wall rock glaciers may exceed those on ice-cored moraines for the reasons outlined in the previous section.

The present study also suggests that a combination of sedimentological and lithological criteria may provide the easiest way of discriminating between valley-wall rock glaciers and lateral moraines. The lithological characteristics of valley-wall rock glaciers are largely identical to those of the source-wall directly upslope of the rock glacier; they may, in part, contain a few clasts from deforming lateral moraines that may themselves contain erratics. In addition, lateral moraines often possess a well-developed matrix of fines which is usually visible near the surface, whereas the surface of rock glaciers tends to be characterised by an openwork layer of boulders that is largely free of fine material.

6.5.4 Avalanche boulder tongues and rockslide deposits

Morphological expressions of avalanche and landslide activity include small bench-like 'bulges' on talus, avalanche-modified talus, avalanche boulder tongues and large lobate extensions at the base of talus slopes. Such extensions, although limited in terms of microrelief structures, have often mistakenly been identified as rock glaciers. Avalanche boulder tongues are accumulations of rock debris that form through the erosion and deposition by snow avalanches (Rapp, 1959). The slope profile of avalanche-modified talus and avalanche boulder tongues is markedly concave, and most of the latter have an asymmetrical cross-profile that is not typical of valley-wall rock glaciers (Luckman, 1978). In addition, the surface of avalanche boulder tongues is generally smooth and flat, and is often characterised by the presence of avalanche debris trails (Rapp, 1959). Another diagnostic feature of these landforms is that they are usually associated with obvious avalanche tracks of chutes higher up the same slope. Finally, although lengths may be comparable to those of valley-wall rock glaciers, avalanche boulder tongues rarely exceed 200m in width (Rapp, 1959; Gardner, 1970; Luckman, 1978).

Rockslide deposits are accumulations of rock debris that may sometimes resemble avalanche boulder tongues and valley-wall rock glaciers as they are often characterised by concave profiles, and tend to develop in similar topographic settings (Rapp, 1959). Generally, they may be distinguished from avalanche boulder tongues by their rough and uneven surface which often consists of very large boulders (Luckman, 1978). Unlike valley-wall rock glaciers, neither avalanche boulder tongues nor rockslide deposits contain asymmetrical inner transverse ridges; in addition, their frontal slope gradients are generally less than those of valley-wall rock glaciers (Luckman, 1978).

6.5.5 Summary

Table 6.5 summarises a list of morphological and sedimentological criteria that may help to differentiate between morphologically-similar features in the field. In general, the vast majority of valley-wall rock glaciers are readily distinguishable on the basis of morphology alone. A few, however, are more difficult to

DIAGNOSTIC CRITERIA	VALLEY-WALL ROCK GLACIERS			PROTALUS RAMPARTS		AVALANCHE BOULDER TONGUES	ROCKSLIDE DEPOSITS
	Active	*Inactive*	*Relict*	*Active*	*Relict*		
	Field data from Chapters 4, 5 and 6.			*Ballantyne, 1987*	*Ballantyne & Kirkbride, 1986*	*Rapp, 1959; Gardner, 1970; Luckman, 1978.*	*Rapp, 1959; Luckman, 1978.*
Frontal slope gradient	34° - 41°	28° - 39°	23° - 32°	34° - 39.5°	25° - 34°	5° - 15°	
Maximum slope facet	53°			43°			
Surface slope gradient	13° - 21°	10° - 26°	9° - 20°			10° - 30°	10° - 40°
Frontal slope height	25m - 28m	12m - 32m	10m - 34m	15m - 21m	7m - 20m	1m - 5m	
Length	170m - 230m	70m - 600m	180m - 510m	40m - 50m	25m - 67m	330m - 700m	
Width	80m - 250m	60m - 700m	350m - 700m	60m -115m	120m - 620m	70m - 200m	
Depth	10m - 28m	10m - 35m	10m - 34m	5m - 10m	4m - 14m	5m - 25m	
Frontal slope characteristics	Proximal slope gradients 5°-10°; unstable, upper rectilinear slope up to 53°; many fine deposits; basal talus 24°-39°.	Generally only 3 distinct units; more rounded crest; less fines; upper rectilinear gradients to 45°.	3 units; surface fines absent; vegetation and lichens common; basal talus 11°- 36°	Proximal slope gradients 0°- 8°; unstable, upper rectilinear unit up to 43°; basal talus 23°-36°	Maximum distal gradients of 39°; vegetation and lichens may be present.	Fall sorting may be present, often asymmetrical cross-profile.	No basal talus apron and absence of fall-sorting.
Volume	$1.69 \times 10^{4} m^{3}$ - $1.76 \times 10^{6} m^{3}$			$1.20 \times 10^{4} m^{3}$ - $3.90 \times 10^{4} m^{3}$			
Inner transverse ridge characteristics	Between 1 and 13 inner ridges on which distal slopes are consistently steeper than proximal slopes			No inner transverse ridge formation		No inner ridges, generally smooth surface, concave profile	No inner ridges, rough and uneven surface.
Clast angularity	Generally sub-angular and angular clasts. Surface mantle of openwork coarse debris particularly on active features			Dominantly angular/slabby clasts. Non-oriented		Mostly angular to very angular clasts, oriented and perched boulders downslope.	Angular to very angular clasts.
Microrelief features	Closed depressions and surface ponding may occur			Perched boulders		Avalanche debris trails, perched boulders	Perched boulders

Table 6.5 Morphological and sedimentological criteria that may be used to differentiate between valley-wall rock glaciers, protalus ramparts, avalanche boulder tongues and rockslide deposits.

identify, particularly small poorly developed rock glaciers that contain little or no inner transverse ridge development.

6.6 Conclusions

The central theme of this chapter has been to investigate the nature and origin of the three principal constituents of valley-wall rock glaciers, namely coarse debris, fine sediments and ice. In addition, a list of morphological and sedimentological criteria that may help to discriminate between valley-wall rock glaciers and morphologically-similar features was outlined. Ten major conclusions may be drawn from the research reported in this chapter.

1) The surface of most active, inactive and relict valley-wall rock glaciers is characterised by a coarse openwork debris layer that varies in depth from 0 to more than 2 metres.

2) Beneath this boulder mantle, fine sediments form a matrix in which sizeable clasts may be embedded.

3) Rock glacier fine sediments sampled in Fornesdalen, northern Norway are marginally frost-susceptible.

4) A combination of field and laboratory evidence suggests that lateral moraine sediments may have become incorporated within some of the study rock glaciers during formation. It is thought, however, that

the majority of rock glacier fines are inherited from pre-existing talus deposits.

5) Isotopically, ice samples from an active valley-wall rock glacier proved significantly different from both snowbank and glacier ice. Physical and isotopic analyses of a range of ice samples suggest that rock glacier ice forms largely through the refreezing of percolating rain and meltwater, although ice formed by segregation processes probably contributes to total ice volume.

6) Active and inactive valley-wall rock glaciers often exhibit distinctive clast size distributions on their frontal slopes.

7) Mean clast size does not appear to be related to rock glacier size nor is there any apparent relationship between clast size and downslope direction.

8) Pronounced differences in median clast sizes between the proximal and distal slopes of inner transverse ridges are consistent with the hypothesis that these asymmetric transverse ridges formed through internal shearing.

9) Mean clast size on the surface of valley-wall rock glaciers is frequently greater than mean clast size on corresponding talus slopes, suggesting that large-scale rockfall events may be associated with the formation of some valley-wall rock glaciers.

10) The majority of valley-wall rock glaciers are readily distinguishable on the basis of morphological and sedimentological evidence, although small or incipient valley-wall rock glaciers may be confused with other morphologically-similar landforms.

Chapter 7

Valley-Wall Rock Glacier Formation

7.1 *Introduction and rationale*

The aims of this chapter are twofold. In the first main section, field data are applied to several conceptual models of rock glacier movement in order to determine which of these models are theoretically feasible. A simple laminar flow equation is used as the basis for these calculations as field measurements of rock glacier thickness and average surface gradient are available. A number of different density models are proposed, based upon the proposed origin of internal ice (cf. section 3.2), and critical basal shear stresses are calculated for each.

The second aim of the chapter is to consider the implications of both the theoretical findings and the empirical findings discussed in the previous three chapters, so that the hypotheses of formation outlined in Chapter 3 can be evaluated.

7.2 Theoretical boundary conditions for valley-wall rock glacier formation

7.2.1 Introduction

The aim of this section is to attempt to define theoretical boundary conditions for valley-wall rock glacier formation. Attention is focused on the factors that control rock glacier movement. As discussed in section 2.6.4 above, several morphological and sedimentological characteristics of the frontal slopes of active rock glaciers suggest that rock glacier movement is largely a deformational process rather than the sliding downslope of a rigid body of material. For example, a lack of vegetation on the frontal slopes of active valley-wall rock glaciers and the steepness and manifest instability of these active frontal slopes indicate that valley-wall rock glaciers experience differential forward motion, in which the upper layers move downslope faster than the lower layers. In addition, exposures of fine-grained sediments on the lower rectilinear units of the frontal slopes of each active valley-wall rock glacier studied (cf. section 5.5.2 above) suggest that most of the thickness of each rock glacier is involved in the motion, because if only the top two or three metres were moving downslope, fine sediments would not be exposed below this depth on the frontal slope.

It is very difficult to compute stress fields within a large, generally heterogeneous, mass of rock glacier; it is possible, however, to do so within simplified conceptual models. For example, whether a rock glacier moves *en masse* or not is dependent largely upon four main factors, namely average rock glacier density, average thickness, average surface gradient and the yield strength of the constituent material. Field data are available for rock glacier thickness and surface slope gradient, but are not available for rock glacier density. However, by modelling a variety of rock glacier densities based upon the proposed origin of internal ice (cf. section 3.2), basal shear stresses can be calculated for each of these models to establish which are theoretically possible.

7.2.2 *Conceptual models of laminar flow*

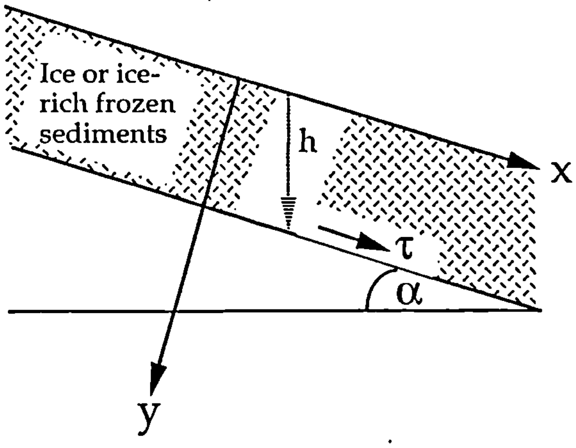
A block of ice of constant thickness h, which rests on a plane of slope a, is used as a first model of rock glacier movement in which the rock glacier is viewed as having a core of ice (Figure 7.1a). In this model, it is assumed that the length and width of the slab are large compared with h, in other words that boundary effects are negligible. In addition, it is assumed that the block does not slide on the plane, but that movement consists solely of internal deformation due to shear stresses generated by self-weight. The shear stress τ_b at the base of a column of ice within such a slab is equal to the component of weight resolved down the slope, namely ρ g h sinα (Paterson, 1981), where ρ equals ice density and g equals gravitational acceleration. If it is assumed that the block of ice is perfectly plastic, the depth of ice required before deformation will occur can be obtained for various measurements of surface slope. Thus:

$$h_{crit} = \tau_b / (\rho g \sin\alpha)$$

$$= (\tau_b \operatorname{cosec}\alpha) / \rho g \qquad (7.1)$$

Clearly, surface and basal slopes of valley-wall rock glaciers are not parallel along their entire length (cf. long profiles depicted in Figures 5.5 to 5.8). However, provided the difference in gradient between the slopes of the upper and basal surfaces is small, τ_b is the same as if the block of

7.1a



$$\tau = \rho\, g h \sin(\alpha)$$

Basal shear stress = density x gravity x height x sin (α)

7.1b

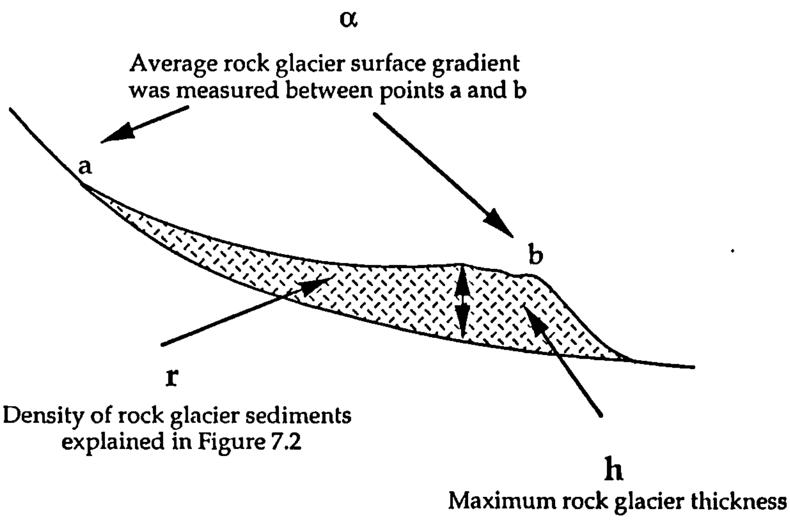


Figure 7.1a Parallel-sided slab of ice or ice-rich frozen sediments for which the basal shear stress equals the downslope component of weight, $\rho g h \sin\alpha$;
7.1b Definitions of rock glacier dimensions used in modelling.

ice had parallel surface and basal slopes (e.g. Paterson, 1981). In addition, as the shear stress at the bed is determined largely by the surface slope, ice flows in the direction of maximum surface slope even though the bed may slope in the opposite direction.

Definitions of rock glacier dimensions measured from the study sample are illustrated in Figure 7.1b. Average rock glacier surface gradient (α) was measured from a point at the top of the frontal ridge to a point upslope where the rock glacier and talus merge (cf. section 5.3). Maximum rock glacier thickness (h) was estimated for each study rock glacier by assuming a regular downslope decline in the gradient of the underlying slope or by profiling the hillslope adjacent to the rock glacier (cf. section 5.3). It should be emphasised that by using a maximum thickness value for 'h' in the equation $\tau_b = \rho\ g\ h\ \sin\alpha$ rather than an average thickness value, which is not available, the resulting calculated basal shear stress values will represent maximum average values, although locally basal stresses may exceed this calculated value.

Four conceptual models of valley-wall rock glacier formation that vary in terms of internal ice content are investigated and are shown in Figure 7.2. Model 1, the glacier ice-cored model, comprises a core of glacier ice that is overlain by a layer of coarse unfrozen debris. Model 2, the matrix ice model, contains perennially-frozen sediments rich in ice, rather than an ice-core. An openwork, unfrozen mantle of coarse boulders overlies the frozen interior of the rock glacier. The ice in this model is viewed as a matrix that supports and separates clasts (cf. Figure 2.2). Ice volumes therefore must exceed those associated with interstitial ice, because if ice

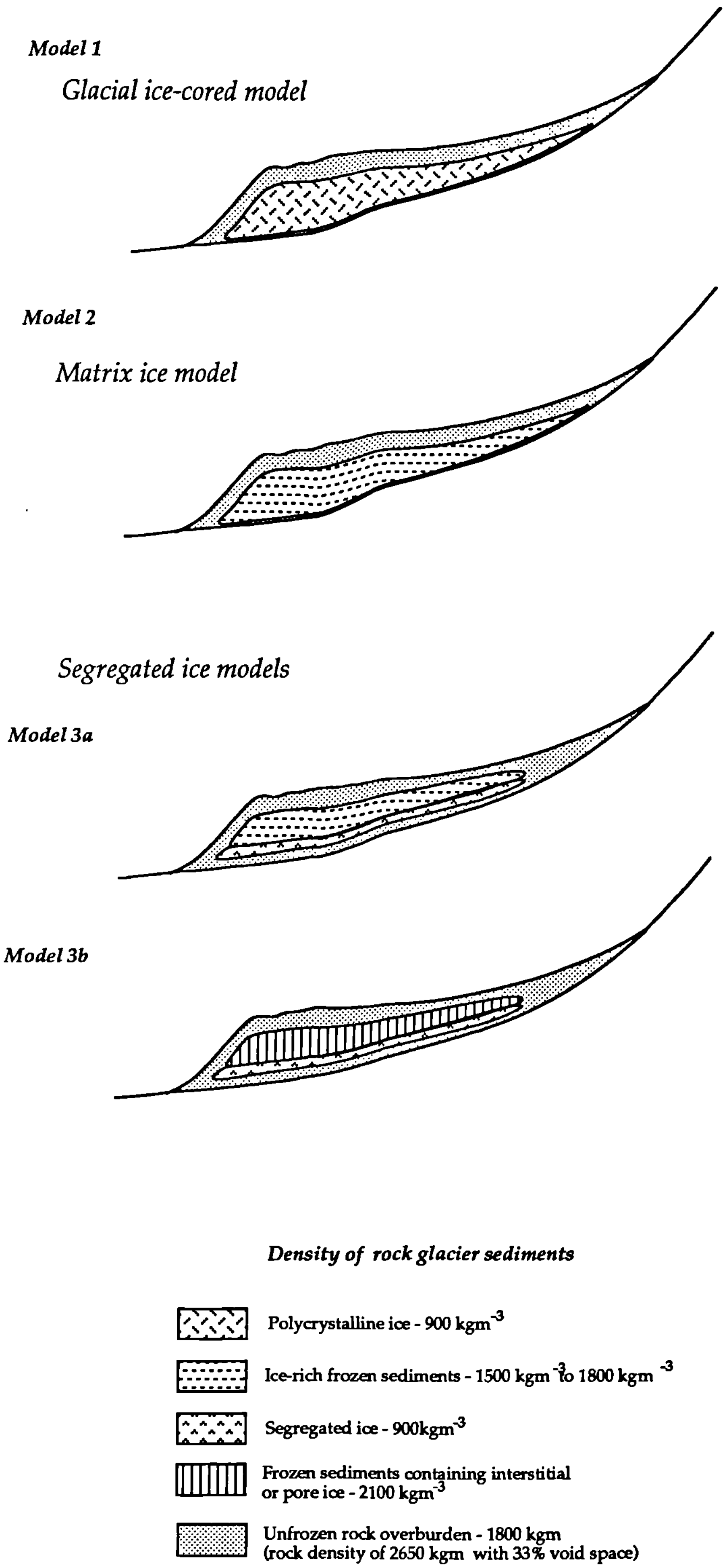


Figure 7.2 Four possible valley-wall rock glacier models based on different internal ice compositions.

filled only the interstices the shear strength of the rock-ice mix would be too great for movement to occur (cf. section 2.2.4; Whalley, 1974). In the third and fourth models, layers or lenses of segregated ice are assumed to have formed below either perennially-frozen sediments rich in ice (Model 3a) or sediments that contain only interstitial ice (Model 3b). Clearly, in models 3a and 3b, a wide range of segregated ice volumes is possible, from small lenses to massive ice bodies. In Figure 7.2, a relatively shallow layer of segregated ice is shown near the base of the rock glacier.

The densities of the constituents in each model are also indicated in Figure 7.2. A density of 900kgm^{-3} is generally accepted for both polycrystalline ice and segregated ice (e.g. Andrews, 1975; Paterson, 1981; Drewry, 1987). The density of perennially-frozen sediments has been estimated by several workers. For example, Wahrhaftig and Cox (1959) suggested an average value of 1800kgm^{-3}, Haeberli seems to have employed a value of about 1700kgm^{-3}, whilst Barsch et al. (1979) thought that the density of such sediments would lie between 1500 and 1800kgm^{-3}. The second model of rock glacier formation therefore assumes a range of densities from 1500kgm^{-3} to 1800kgm^{-3} for perennially-frozen sediments. A density of 2100kgm^{-3} is used for interstitial ice sediments where ice fills all the voids between the rubble. This value is based on a rock density of 2650kgm^{-3}, an ice density of 900kgm^{-3} and a void ratio of one third.

As noted in section 6.2.1 above, the upper surface of each active and most inactive and relict valley-wall rock glaciers is characterised by a relatively shallow layer of unfrozen coarse debris. The depth, and thus weight, of this overburden debris alters the thickness of ice needed before rock glacier movement will occur. New critical thicknesses of ice or ice-rich sediments can be calculated using a slightly modified version of the equation shown in Figure 7.1:

$$\tau_o = (\rho_{rock} h_{rock} + \rho_{ice} h_{ice})\, g \sin\alpha \qquad (7.2)$$

where the density of the overburden debris (ρ_{rock}) is taken to be 1800kgm^{-3}, which allows for 33% void space and a rock density of 2650kgm^{-3}.

In terms of basal shear stress, measurements that have been undertaken for glaciers have shown that critical τ_b values for deformation of polycrystalline ice lie within a range from 0.5 to 1.5 bar (Raymond, 1980; Paterson, 1981). Paterson suggested that it is therefore reasonable to regard glacial ice as a perfectly plastic material with a yield stress (τ_o) of 1 bar. However, in some cases where unconsolidated deformable sediments rather than bedrock form the substrate (cf. section 5.4.5), maximum average basal shear stresses of less than 1 bar have been recorded (e.g. Alley et al., 1986).

Two previous attempts have been made to determine the maximum shear stresses acting at the base of deforming rock glaciers (Wahrhaftig & Cox, 1959; Haeberli, 1985). Using the equation shown in Figure 7.1a, Wahrhaftig and Cox calculated the maximum shear stresses acting within several Alaskan rock glaciers and found that base-stresses of active rock glaciers lie in a range between 1 and 2 bar, whereas the maximum shear stresses in most inactive rock glaciers were less than 1 bar. The authors proposed two possible reasons for this difference. The first, and, they believe, most likely explanation, is that a threshold value (in this case is ~1 bar) must be reached before rock glaciers will move. In other words, when basal shear stress values fall below this threshold, possibly as a result of a decrease in the rate of debris supply or a decrease in self-weight resulting from rock glacier movement, rock glaciers cease to move. The second reason relates to a reduction in rock glacier thickness due to the melt-out of internal ice, which they suggest may also reduce the basal shear stress value. The authors note that these two explanations are not mutually exclusive.

Two assumptions made by Wahrhaftig and Cox have an important bearing on their results and thus require discussion. First, the values of rock glacier thickness used by Wahrhaftig and Cox were estimated solely from the height of the frontal slope using the equation:

$$H = H_V \cos\alpha \tag{7.3}$$

where a is the "slope of the upper surface at the front" of the rock glacier and H_V is the "vertical height of the front" (p.402). Although their methodology is unclear, use of this equation was shown to produce rock glacier thickness values that are consistently *less* than the true vertical height of the frontal slope. In Tables 5.1 to 5.3 it was shown that rock glacier thickness may be *either* greater or less than the height of the frontal slope. Moreover, in section 5.4.5 it was noted that the height of the frontal slope is not always a good surrogate for maximum rock glacier thickness. Thus, the accuracy of the thickness data reported by Wahrhaftig and Cox must be questioned. The second assumption made by Wahrhaftig and Cox concerns the density of rock glacier sediments. They employed a value of 1800kgm^{-3} on the basis that the motion of rock glaciers corresponds to the movement of frozen rubble, and did not consider other possible density values.

More recently, Haeberli (1985), who regards the movement of rock glaciers as the slow downslope creep of "perennially frozen sediments rich in ice", has suggested that 1 bar is a poor approximation of the yield strength of rock glaciers. He believes that estimates of base-stress values can be made only for those rock glaciers whose thickness has been determined by means of tunnelling or geophysical soundings. Accordingly, he does not believe that accurate determinations of rock glacier thickness can be obtained solely from the dimensions of the frontal ridge. He has not used nor commented upon the validity of the method used in this work (cf. section 5.3), whereby long profiles on hillslopes adjacent to rock glaciers are surveyed to obtain an estimate of maximum rock glacier thickness.

To determine values of basal shear stress Haeberli also used the equation $\tau_b = \rho\ g\ h\ \sin\alpha$, although he rightly emphasised that its application is valid only at large scales. At smaller scales, where variations in basal shear stresses within individual rock glaciers are investigated, corrections must be made for the effects of longitudinal stress gradients. Such corrections, however, would require detailed information about surface strain rates, about the form of the flow law, and about local variations in rock glacier thickness, data that are normally not available. Maximum basal shear stress values calculated for the six active rock glaciers studied by Haeberli are 1.10 bar, 1.25 bar, 2.25 bar, 2.50 bar, 2.50 bar and 3.5 bar, and are therefore generally greater than those calculated by

Wahrhaftig and Cox for the Alaskan rock glaciers. Apparently, the generally higher basal shear stress values reported by Haeberli are due mainly to the much larger dimensions of Haeberli's study rock glaciers, at least four of which have been identified by the present author to be valley-floor rock glaciers. (Haeberli does not recognise the existence of valley-floor rock glaciers that contain some amount of glacially-derived ice; instead he views all rock glaciers as masses of perennially frozen sediments that slowly creep downslope.) The thickness of frozen sediments within Haeberli's three rock glaciers that exhibit maximum basal shear stresses of 2.5 and above, namely Kintole, Murtèl and Gruben is >50m, ≤80m and ~100m respectively. These values contrast with those reported in this study (cf. Tables 5.1 to 5.3) and with those reported by Wahrhaftig and Cox who found that the total thickness of both valley-wall and valley-floor rock glaciers in Alaska rarely exceeds 60m.

On the basis of his calculated average six basal shear stress values, Haeberli proposed that a critical value of approximately 2.5 bar seems reasonable for rock glacier movement. His suggestion of such an 'average' value, however, seems illogical as its adoption implies that those rock glaciers with maximum base-stresses of less than 2.5 bar would be incapable of movement. Indeed, three of the six rock glaciers studied by Haeberli fall below this average value, and yet all are active. Thus, instead of using an average value, it would be more appropriate to determine a critical value below which rock glaciers do not move. The lowest maximum basal shear stress calculated by Haeberli is 1.10 bar, which corresponds closely with the 1 bar limit suggested by Wahrhaftig and Cox (1959).

One other worker has attempted to use a laminar flow equation to determine boundary conditions for rock glacier movement. Whalley (1974) calculated that the critical thickness of glacier ice required to induce deformation on a slope of 10° is 23m (1 bar) whereas the value for an ice/rock mixture is 94m (10 bar). Whalley maintained that his calculated critical height value of 94m for an ice/rock mixture indicates that rock glaciers cannot form unless glacier ice is present because rock glacier thicknesses are almost always less than 94m. His conclusion, however, can be disputed by a careful examination of his methodology. First, Whalley employed an inappropriate basal shear stress equation, which he stated is applicable for non-zero slope angles:

$$H_{crit} = 2\tau \;/\; \rho\, g \cos\alpha \tag{7.4}$$

This equation contradicts the reasoning outlined by Paterson (1981) unless Whalley is envisaging a wedge-shaped block of ice in which the basal and upper surfaces exhibit markedly different gradients. As discussed in section 7.2.2, such a situation is unrealistic for valley-wall rock glaciers. Second, Whalley correctly adopted a basal shear stress value of 1 bar for his glacial ice calculations but rather surprisingly employed a basal shear stress of 10 bar for the ice-rock mixture. Although his reasons for doing so are not explained, it is possible that he envisaged the ice in his rock-ice mixture to occupy only the interstices; if so, a 10 bar basal shear stress value may be appropriate as the yield strength of such a mixture would be very high. However, the matrix ice model presented above assumes that the ice fills not only the interstices but actually supports and separates the majority of clasts, and so the yield strength of the matrix ice would be markedly lower than 10 bar. As the methodology employed by Whalley seems flawed, his results must be questioned.

In the following sections, each of the three conceptual models illustrated in Figure 7.2 will be investigated by applying field data collected in this study, and the results will be compared with those presented by Haeberli (1985), Wahrhaftig and Cox (1959) and Whalley (1974).

7.2.3 Model 1 - Glacier ice-cored rock glacier model

As outlined above, the first model under consideration involves a rock glacier that contains a core of glacier ice. Following Paterson's (1981) suggestion that a yield stress of 1 bar adequately defines the yield stress of glacial ice, a basal shear stress of 1 bar is plotted on a graph that shows the thickness of an ice-cored rock glacier as a function of the cosecant of the slope of the upper surface of the rock glacier (Figure 7.3). In other words, the critical thickness of ice required before deformation will occur is shown for various surface slope angles assuming a basal shear stress (τ_b) of 1 bar, an ice density (ρ) of 900kgm^{-3}, and gravitational acceleration (g) of 9.81msec^{-2}. For example, for a surface slope of 15°, the critical height of ice required to exceed a basal shear stress of 1 bar is 43.8m, whereas for a more gentle slope of 10°, 65.2m of ice would be required. Thus, if τ_b is constant, an active ice-cored rock glacier may be relatively thin where the upper surface is steep but must be thick where the surface slope is gentle.

Reported measurements of total rock glacier thickness suggest an upper limit of approximately 60m (e.g. Wahrhaftig & Cox, 1959; study data, section 5.4.5 above). If this limit is used as an initial boundary condition

Polycrystalline ice
1 bar basal shear stress
(Ice-cored rock glacier model)

$$H_{crit} = \frac{\tau \operatorname{cosec} A}{\rho g}, \text{ where } \begin{aligned} g &= 9.81\ \text{msec}^{-2} \\ \rho &= 900\ \text{kg m}^{-3} \\ \tau &= 10^5\ \text{Nm}^{-2} \end{aligned}$$

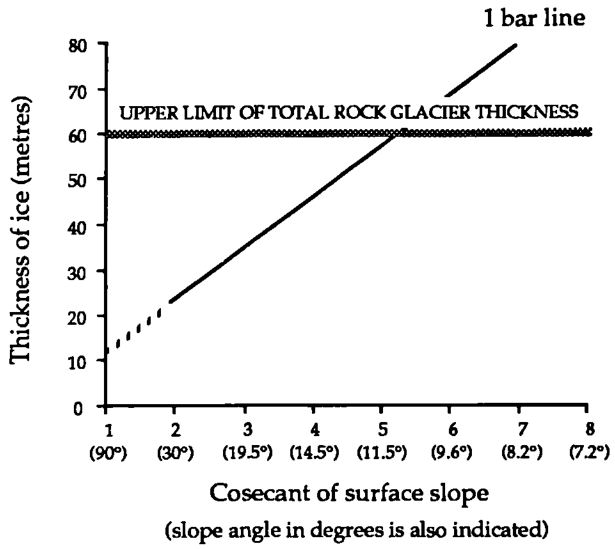


Figure 7.3 Line along which basal shear stress of an ice-cored rock glacier is equal to 1 bar.

in Figure 7.3 (i.e. before including overburden effects), it can be seen that surface slope gradients would need to be greater than approximately 10° for internal deformation to occur if the limiting basal yield stress was 1 bar. Interestingly, average surface gradients of valley-wall rock glaciers reported in the literature range from 10° to 30° (e.g. Wahrhaftig & Cox, 1959), whilst those in the study set lie between 9.7° to 26.9° (cf. Tables 5.1-5.3). However, if we consider the average gradients measured on each study rock glacier and assume a basal stress of 1 bar, ice thicknesses of between 25.0m and 67.2m are needed before movement would occur through internal deformation; yet the thickness of the study rock glaciers ranges from 10m to only 35.5m. This suggests that many of the study rock glaciers are too thin to have contained ice-cores of sufficient size for internal deformation to have occurred, assuming a base-stress of 1 bar. Also, the upper limit of average surface slope gradients (30°) suggests that a critical depth of ice of ~20m is needed before deformation will occur. However, several of the study rock glaciers, including active and inactive features in which ice should still be present, attain a total thickness of less than 20m. Once again, therefore, the total thickness of the study rock glaciers does not seem substantial enough given the assumptions shown in Figure 7.3. However, the effects of the unfrozen rock overburden, which previously have not been considered by other workers, may reduce the thickness of the ice-core to within the limits of the field data.

As illustrated in Figure 7.2a, Model 1 consists of an ice-core that is overlain by an openwork layer of coarse debris. Figure 7.4 shows the thicknesses of this unfrozen overburden layer that are required to induce strain for varying ice-core thicknesses and surface slope gradients. In the

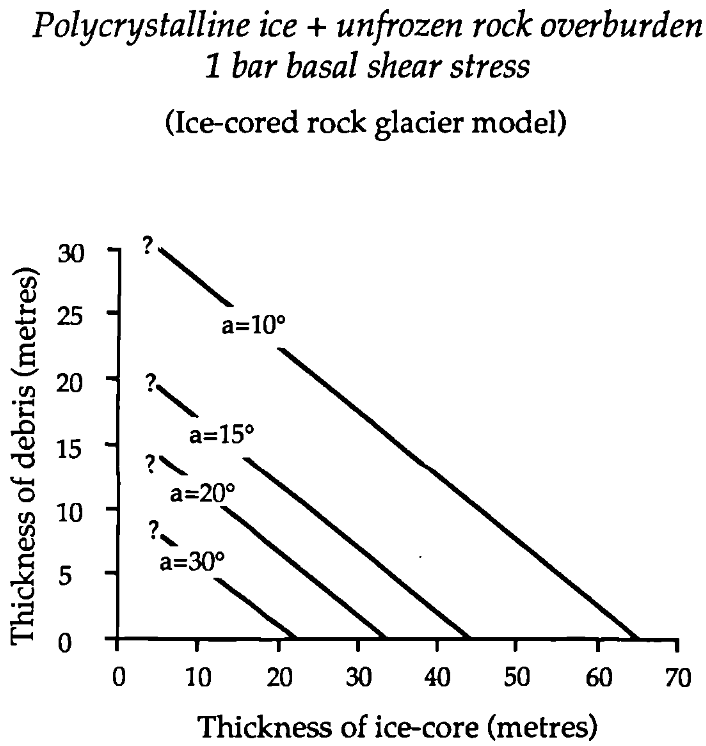


Figure 7.4 Thickness of unfrozen rock overburden required to induce strain for varying ice-core thickness and surface slope gradients of 10°, 15°, 20° and 30° assuming an ice density of 900kgm^{-3}, an overburden density of 1800kgm^{-3} and a critical yield strength of 1 bar.

graph, a yield strength of 1 bar is assumed. Clearly, the lower the slope angle, the greater the amount of debris required if ice depth remains constant. For example, given a surface gradient of 15°, an ice thickness of 20m and a yield strength of 1 bar, 11.9m of debris would be required to induce strain, whereas for a slope of 20° only 6.5m of debris would be needed.

For each of the study rock glaciers, the depth of this debris mantle was estimated from the depth of the upper rectilinear slope on the frontal ridge (cf. section 5.5.2; Figure 5.11, unit II). Mantle depth was then subtracted from total rock glacier thickness to provide an estimate of core thickness. Using these dimensions, it is possible to plot the study rock glacier data under the assumptions of Model 1 using equation (7.2), where ρ_{mantle}=1800kgm^{-3} and ρ_{core}=900kgm^{-3}. To present these results graphically, basal shear stress lines of 0.5 bar and 1 bar are plotted on Figure 7.5 for varying rock glacier thicknesses and surface gradients assuming a standard density of 900kgm^{-3} for total rock glacier thickness. Adjustments, therefore, were made to the true height of the rock glacier mantle as its density is twice that of the ice-core. Thus, the adjusted height of the mantle (H_a) was calculated using the equation

$$H_a = (1800/900)\, H_{mantle}. \qquad (7.5)$$

For each study rock glacier, the adjusted height of the mantle (H_a) was added to the height of the core and the total adjusted height (H_{adj}) was plotted against surface gradient. Thus,

$$H_{adj} = (1800/900)\, H_{mantle} + H_{core}.$$

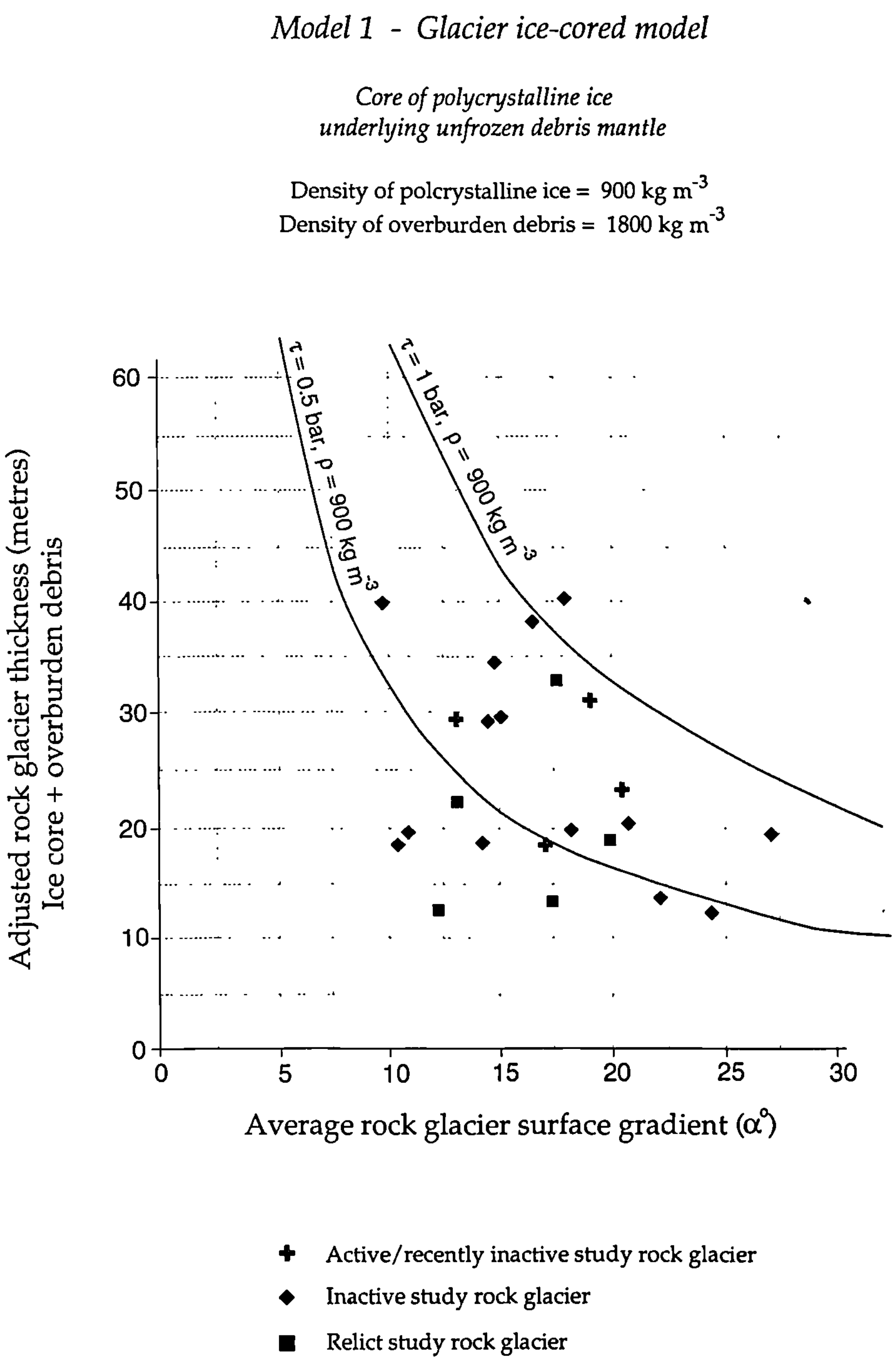


Figure 7.5 Plot of study rock glacier data under the assumptions of Model 1 - the glacier ice-cored model.

In Figure 7.5, the study rock glaciers are divided into those that are active or recently inactive, those that are inactive, and those that are relict. In terms of this model, all but one of the study rock glaciers are predicted to generate maximum basal shear stresses of less than 1 bar. Predicted basal shear stresses of the four active rock glaciers lie between 0.46 bar and 0.88 bar, whilst those of the inactive group range from 0.29 bar to 1.08 bar. Although the relict rock glaciers are plotted on Figure 7.5, the assumptions of Model 1 suggest that they have little validity, as relict features by definition contain no ice. Thus, the thickness of relict Model 1 rock glaciers should be due entirely to overburden debris as the ice-core must have melted. However, thicknesses of relict rock glaciers, which range from 10m to 28m are not markedly less than those for the inactive or active rock glaciers whose thicknesses range from 10m to 35.5m, which suggests that it is unlikely that sizeable ice-cores occurred within the study rock glaciers when active. In addition, the maximum predicted basal shear stress values of the four active or recently inactive valley-wall rock glaciers within the study sample all lie below Paterson's (1981) 1 bar average for polycrystalline ice. As mentioned in section 7.2.2, maximum basal shear stresses less than 1 bar have been reported, particularly if the substrate consists of deformable unconsolidated sediments. However, of the four rock glaciers, only one (Radüner) may lie on top of deformable till; the others all appear to have developed on rigid bedrock.

The evidence presented in this section suggests, therefore, that it is improbable that the majority of the study rock glaciers contained large glacially-derived ice-cores when active. The evidence, however, does not remove the possibility that thinner, perhaps less continuous, masses or layers of glacier ice (cf. Figure 7.4) that become buried by rockfall debris may contribute to the movement of valley-wall rock glaciers. This possibility will be discussed further after Model 2 is investigated.

7.2.4 Model 2 - Matrix-ice rock glacier model

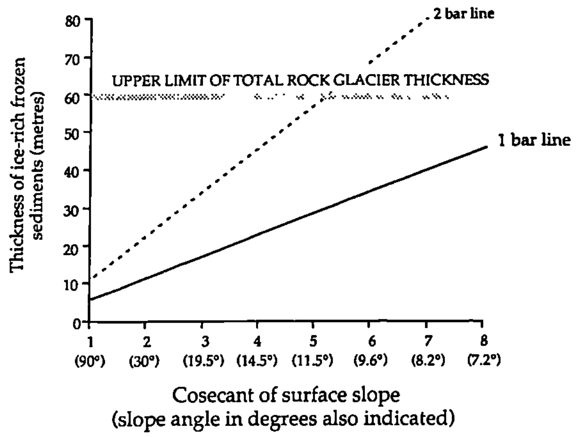
The next stage is to conceptualise a model of a rock glacier that contains perennially frozen sediments rich in ice, rather than an ice-core. The density of these ice-rich sediments is thought to lie between 1500kgm^{-3} and 1800kgm^{-3} (cf. section 7.2.2). Thus, in the calculations that follow, densities of both 1500kgm^{-3} and 1800kgm^{-3} will be used for the ice-rich sediments. In Figure 7.6, the critical thickness of ice-rich frozen sediments required before deformation will occur is presented for various surface gradients. Basal shear stresses of 1 bar and 2 bar are considered because of the uncertainty regarding probable base-stress values, although it is the author's viewpoint that 1 bar is a more accurate estimate of the yield strength of valley-wall rock glaciers (cf. section 7.7.2). Thus, assuming a density of 1800kgm^{-3} (Figure 7.6a) and a basal shear stress of 1 bar, the critical thickness of ice-rich frozen sediments required to induce strain on a slope of 15° is 21.8m. This value is exactly half the depth of polycrystalline ice required under the same yield strength conditions. In other words, downslope movement of a 30m thick valley-wall rock glacier with a surface slope of 15° would occur if it was composed largely of ice-rich frozen sediments (ρ=1800kgm^{-3}) but would not occur if the rock glacier was ice-cored. If the rock glacier was composed of ice-rich frozen sediments with a density of 1500kgm^{-3}, movement would still occur on a slope of 15° as the critical depth of frozen sediments is 26.2m, assuming a basal shear stress of 1 bar.

MODEL 2 - MATRIX ICE MODEL

Ice-rich frozen sediments
1 bar or 2 bar basal shear stress

7.6a

$$H_{crit} = \frac{t \text{ cosec } a}{r\, g}, \text{ where } \quad g = 9.81 \text{ m sec}^{-2}, \quad r = 1800 \text{ kg m}^{-3}, \quad t = 10^{5} \text{ Nm}^{-2}$$



7.6b

$$H_{crit} = \frac{t \text{ cosec } a}{r\, g}, \text{ where } \quad g = 9.81 \text{ m sec}^{-2}, \quad r = 1500 \text{ kg m}^{-3}, \quad t = 10^{5} \text{ Nm}^{-2}$$

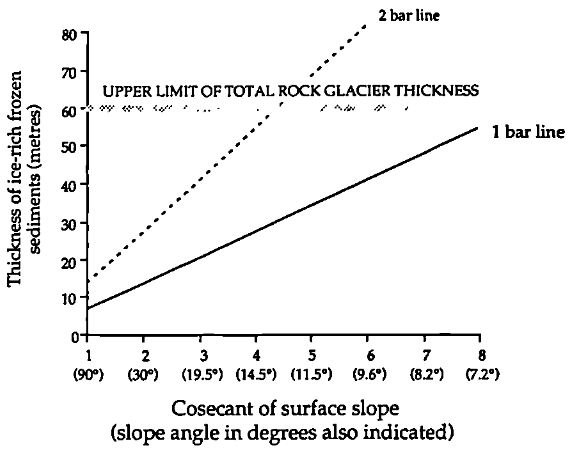


Figure 7.6 Basal shear stress values of 1 bar and 2 bar plotted for varying thicknesses of ice-rich frozen sediments and surface gradient; 7.6a assumes $r = 1800 \text{ kgm}^{-3}$, 7.6b assumes $r = 1500 \text{ kgm}^{-3}$.

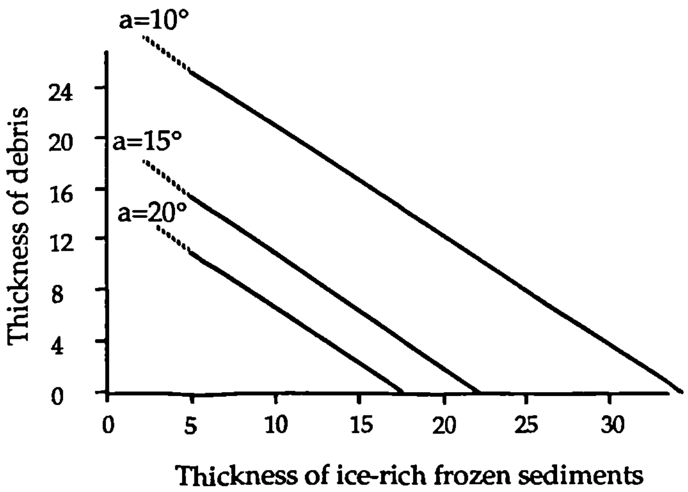
The model shown in Figure 7.6 can be modified to include an overburden layer of unfrozen debris. Thus, Figures 7.7 and 7.8 show the thicknesses of the unfrozen overburden layer that are required to induce strain for varying thicknesses of ice-rich frozen sediments and surface slope gradients. Figure 7.7 assumes a density of $1800 kgm^{-3}$ for the frozen sediments, whereas in Figure 7.8 a density of $1500 kgm^{-3}$ is used. For example, in Figure 7.7, 20m of ice-rich frozen sediments with a surface gradient of 15° and a yield strength of 1 bar, would require approximately 2m of debris to induce strain, whereas for a surface slope of 20° no overburden debris would be required as 20m of frozen sediments will deform under their own weight. If the basal shear stress is 2 bar, almost 23.8m of overburden debris would be required to induce strain in a 20m thick block of ice-rich frozen sediments with a surface slope of 15°. If the density of the perennially frozen sediments is $1500 kgm^{-3}$ (Figure 7.8), slightly greater thicknesses of ice and overburden debris would be required to induce strain, although other things being equal, these thicknesses would still be much less than those required by Model 1.

By using the same methodology that was outlined for Model 1 in the previous section, the study rock glaciers can be plotted under the assumptions of Model 2 (Figures 7.9 and 7.10). In Figure 7.9, a density of $1800 kgm^{-3}$ is used for the frozen sediments, whilst a density of $1500 kgm^{-3}$ is assumed in Figure 7.10. Once again, the study rock glaciers are divided into those that are active, inactive and relict. In Figure 7.9, the mantle thickness of the rock glaciers was not adjusted as the assumed densities of the debris overburden and the ice-rich sediments are the same. As can be seen from Figure 7.9, all of the active and inactive study

MODEL 2 - MATRIX ICE MODEL

Density of ice-rich sediments = 1800kgm^{-3}
Density of overburden debris = 1800kgm^{-3}

1 bar basal shear stress



2 bar basal shear stress

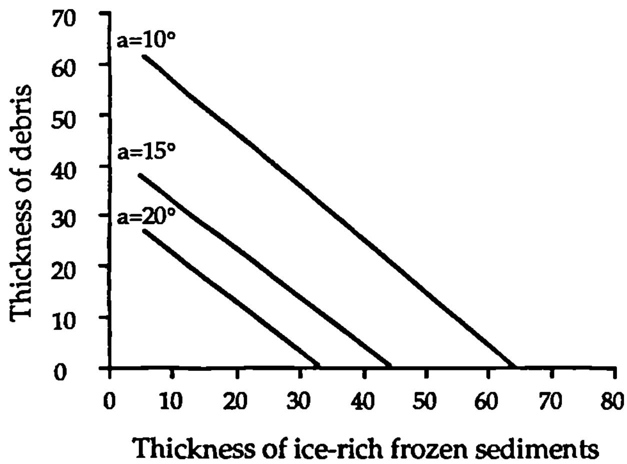


Figure 7.7 Thicknesses of unfrozen rock overburden required to induce strain for varying thicknesses of ice-rich frozen sediments and surface slope gradients of 10°, 15° and 20°, assuming a density of 1800kgm^{-3} for the ice-rich frozen sediments and 1800kgm^{-3} for the overburden debris. The upper plot assumes a critical basal shear stress of 1 bar, whilst a critical base-stress of 2 bar is assumed in the lower plot.

MODEL 2 - MATRIX ICE MODEL

Density of ice-rich sediments = $1500 kgm^{-3}$
Density of overburden debris = $1800 kgm^{-3}$

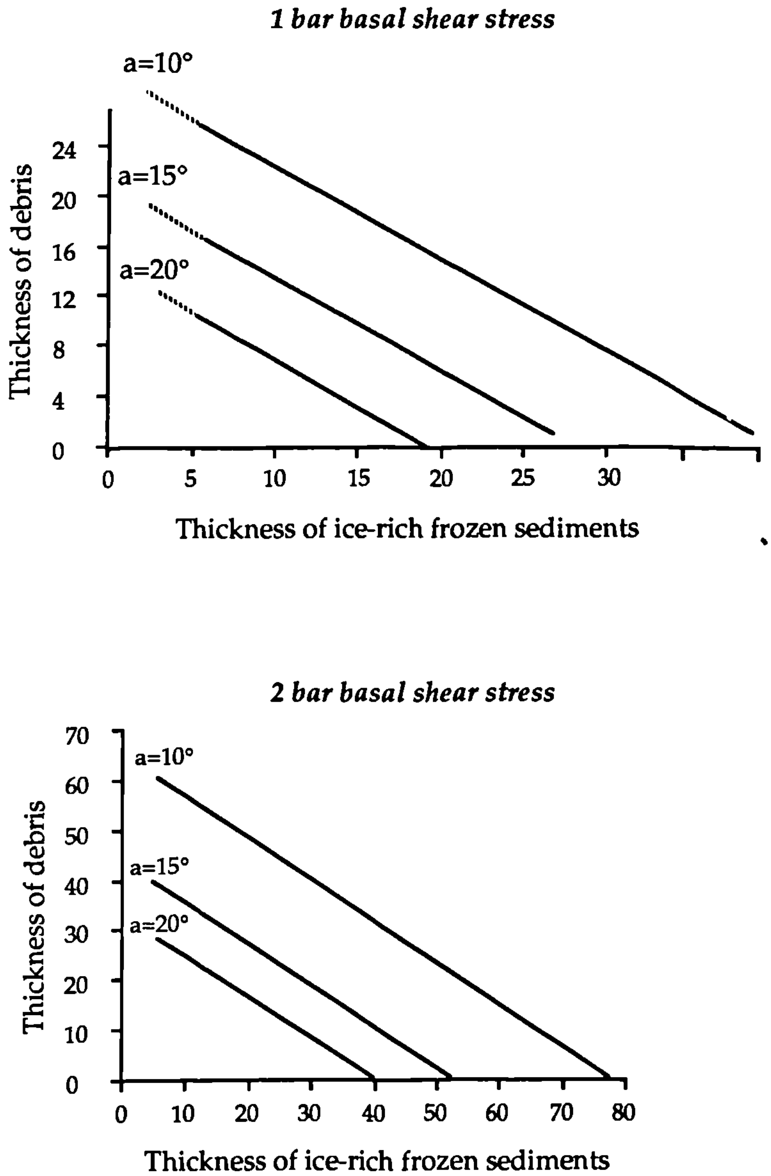


Figure 7.8 Thicknesses of unfrozen rock overburden required to induce strain for varying thicknesses of ice-rich frozen sediments and surface slope gradients of 10°, 15° and 20°, assuming a density of $1500 kgm^{-3}$ for the ice-rich frozen sediments and $1800 kgm^{-3}$ for the overburden debris. The upper plot assumes a critical basal shear stress of 1 bar, whilst a critical base-stress of 2 bar is assumed in the lower plot.

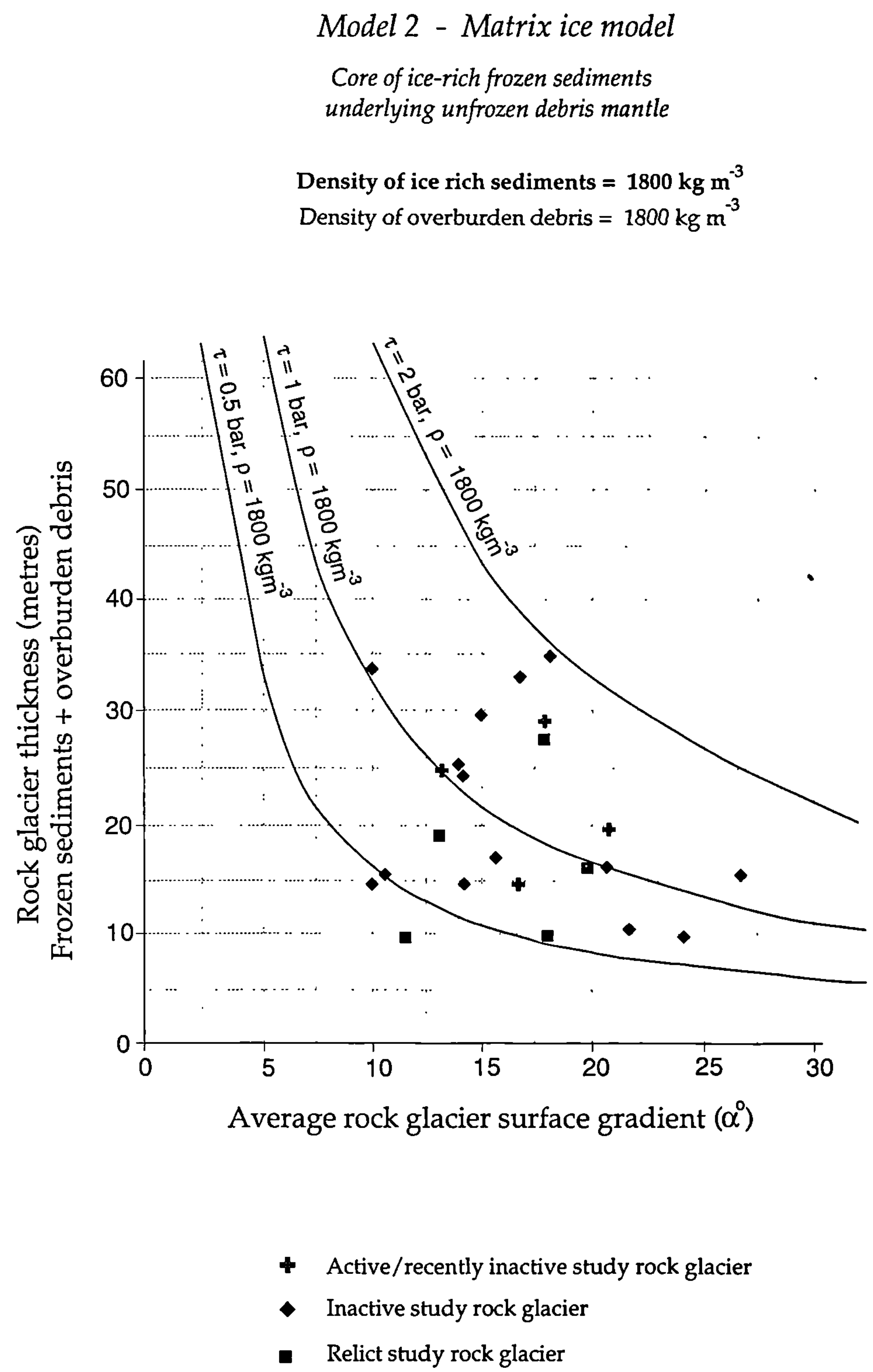


Figure 7.9 Plot of study rock glacier data under the assumptions of Model 2 - the matrix ice model.

rock glaciers exhibit predicted maximum basal shear stresses below 2 bar, and all but one lie above the 0.5 bar line. Predicted basal shear stresses of the four active rock glaciers are 0.76 bar, 1.00 bar, 1.27 bar and 1.56 bar, whilst those in the inactive group range from 0.48 bar to 1.89 bar. Three of the four active rock glaciers, therefore, have maximum basal shear stresses of 1 bar and above; the 1 bar line was thought by Wahrhaftig and Cox (1959) to be a critical threshold below which rock glaciers do not move. The active rock glacier that exhibits a calculated basal shear stress of less than 1 bar is Arolla 3, which is characterised by an anomalously high (85m) frontal slope that is thought to be the result of an underlying rock bench. Given the unusual geological characteristics of the site, it is possible that the estimated value of average thickness for Arolla 3 (15m), which was determined by assuming a regular decline in surface gradient beneath the rock glacier, underestimates the true thickness of the rock glacier. If, for example, average rock glacier thickness was 20m and not 15m, then basal shear stresses would be greater than the 1 bar threshold mentioned above.

In Figure 7.10 where ρ = 1500kgm^{-3}, maximum predicted basal shear stresses of the study rock glaciers are slightly lower, and two of the active rock glaciers fall below the 1 bar basal shear stress line. The validity of the relict rock glacier data shown in Figures 7.9 and 7.10 must again be questioned, although the decrease in thickness following ice-melt would be less than that for Model 1, as ice content of the ice-rich frozen sediments is thought to be approximately 50% by volume (eg. Barsch *et al.*, 1979; Haeberli, 1985).

By assuming a density of 1700kgm^{-3} for the ice-rich frozen sediments in

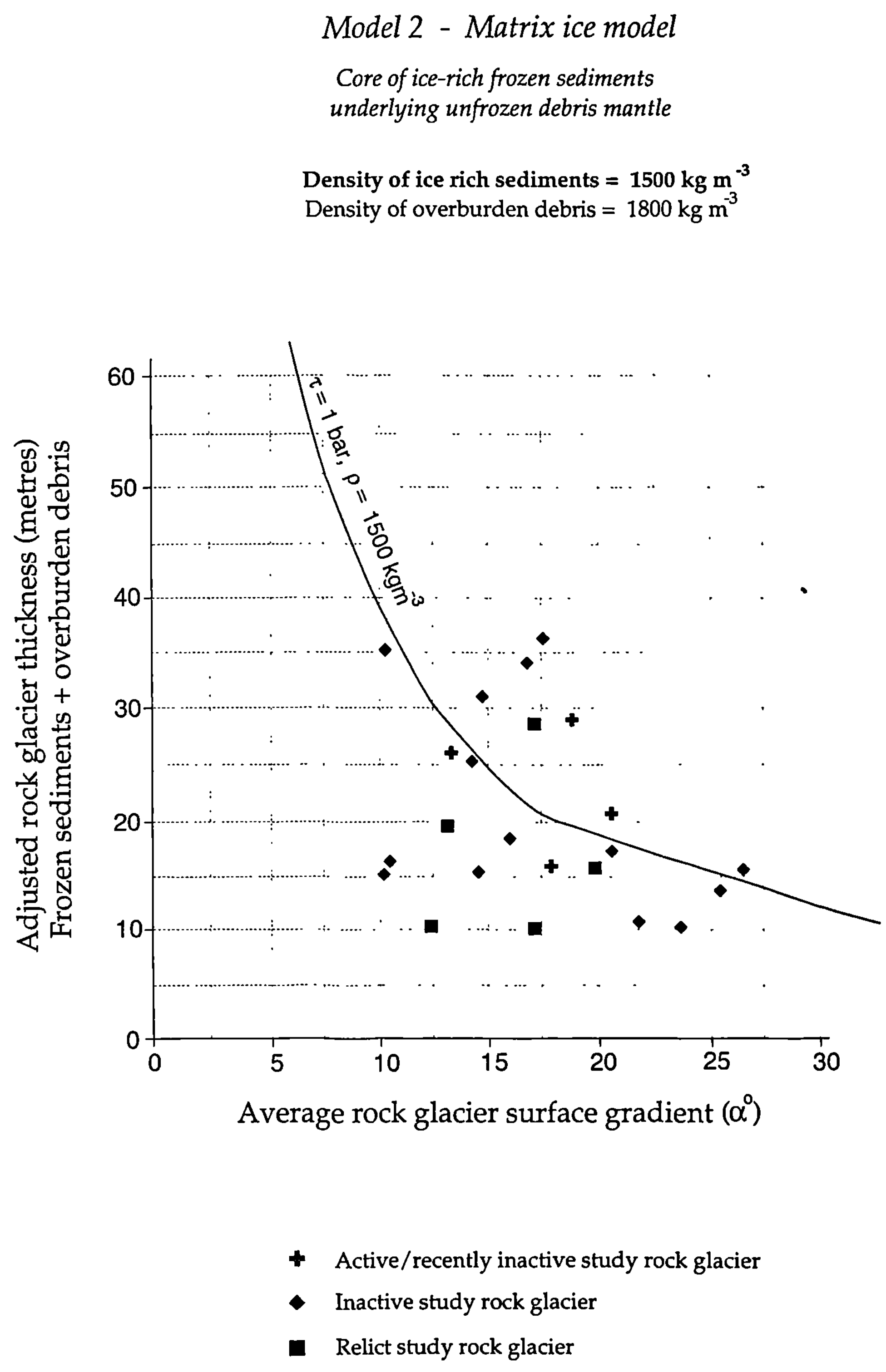


Figure 7.10 Plot of study rock glacier data under the assumptions f Model 2 - the matrix ice model ρ = 1500kg m^{-3}).

Model 2, valid comparisons may be drawn between the maximum basal shear stress values of the four active rock glaciers in this study with those for the six rock glaciers investigated by Haeberli (1985). The values for the four active study rock glaciers are 0.73 bar, 0.95 bar, 1.21 bar and 1.48 bar, whilst those from Haeberli's study are 1.10 bar, 1.25 bar, 2.25 bar, 2.50 bar, 2.50 bar and 3.5 bar. Clearly, Haeberli's rock glaciers exhibit generally higher basal shear stress values, although some overlap occurs between the two data sets. As argued in section 7.2.2 above, Haeberli's lowermost value (1.10) rather than his average value (2.50), is the more important value in terms of determining a critical threshold for rock glacier movement.

An additional finding of these calculations is that critical height values obtained for the ice-cored and matrix-ice models differ markedly from values proposed by Whalley (1974). As outlined in section 7.2.2, Whalley determined that the critical thickness of glacier ice required to induce strain on a slope of 10° is 23m (1 bar) whereas the value for an ice/rock mixture is 94m (10 bar). These values contrast with those obtained from Models 1 and 2, which indicate that 65.2m (1 bar) of glacier ice are required and 32.6m (1 bar) [or 65.2m (2 bar)] of perennially frozen sediments. Whalley's methodology was shown to be inconsistent (cf. section 7.2.2), and so the 94m value he obtained for "an ice/rock mixture" must be regarded as invalid.

In sum, if the study valley-wall rock glaciers when active comprised ice-rich frozen sediments (ρ = ~1500-1800kgm^{-3}) and an unfrozen debris mantle (ρ = 1800kgm^{-3}), predicted maximum basal shear stresses would have been sufficient for movement to occur at only half of the sites.

7.2.5 Model 3 - Segregated ice rock glacier model

The next model under consideration is the segregated ice model of formation. So far, the evidence presented in these theoretical considerations indicates that the critical thickness of ice or frozen sediments is dependent on the assumed density of the rock glacier sediments and surface gradient. For example, the critical thickness of ice-rich frozen sediments (ρ = 1800kgm^{-3}) required to induce strain for any given gradient is exactly half the depth of polycrystalline ice required under the same yield strength conditions, as the density of polycrystalline ice is half that of ice-rich frozen sediments. Using this shortcut, some general observations can be made regarding possible boundary conditions for segregated ice. In Model 3, a wide range of ice volumes is possible from small lenses of segregated ice to massive ice bodies. As such, the ice content could range from that assumed for the glacier ice-cored model to that envisaged in the matrix-ice model. The possibility also exists, however, for segregated ice layers to form in conjunction with interstitial ice *sensu stricto,* in which case ice content would be less than that for the matrix-ice model, while rock glacier density would be greater. With such a combination of segregated ice and interstitial ice, movement would still be feasible as deformation could take place within the segregated ice layer.

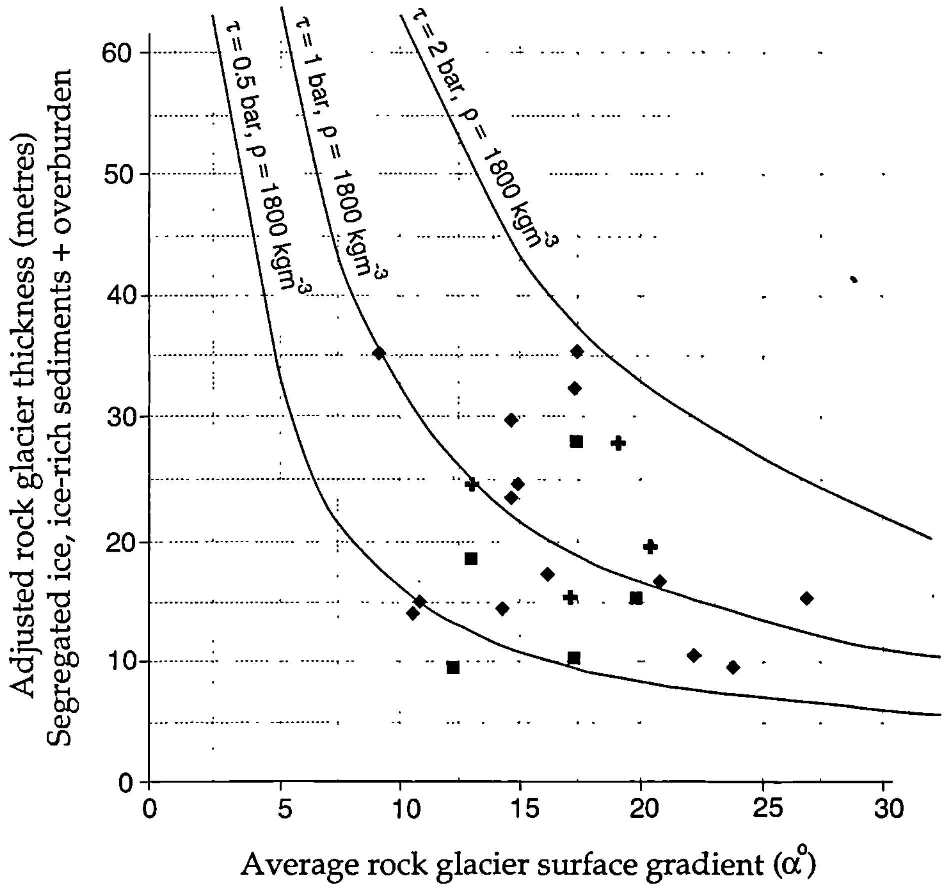
An example of Model 3 (cf. Model 3a, Figure 7.2), assumes a 1 metre thick layer of segregated ice near the base of the rock glacier that is overlain by ice-rich frozen sediments and an unfrozen debris mantle. The study data are plotted against the prediction of this model in Figure 7.11. Predicted

Model 3a - Segregated ice model

Layer of segregated ice with ice-rich frozen sediments underlying unfrozen debris mantle

Assumed thickness of segregated ice = 1m

Density of segregated ice = 900 kg m^{-3}
Density of ice rich sediments = 1800 kg m^{-3}
Density of overburden debris = 1800 kg m^{-3}



- ✚ Active/recently inactive study rock glacier
- ◆ Inactive study rock glacier
- ■ Relict study rock glacier

Figure 7.11 Plot of study rock glacier data under the assumptions of Model 3a - the segregated ice model.

basal shear stress values for different gradients are marginally lower than those for Model 2 where the density of the frozen sediments is assumed to be 1800kgm^{-3}. In addition, the predicted maximum basal shear stress values shown in Figure 7.11 are much higher than those for Model 1, the ice-cored model.

Interestingly, predicted basal shear stress values calculated for the study rock glaciers under the assumptions of Model 3b, which assumes that a 1 metre thick layer of segregated ice is overlain first by sediments in which the void spaces are filled with interstitial ice, and then by an unfrozen debris mantle (cf Figure 7.2), are higher than those obtained in any of the previous models. For example, calculated basal shear stress values for the four active rock glaciers are 0.83 bar, 1.10 bar, 1.40 bar and 1.75 bar, whilst those in the inactive group range from 0.52 bar to 2.13 bar. Figure 7.12 plots the study data under the assumptions of model 3b.

Thus, in terms of the laminar flow of valley-wall rock glaciers, it would appear that field data tend to fit the assumptions of Models 2 and 3 much more readily than they fit Model 1. The implications of these results will be considered in some detail in the latter half of the chapter. It should be emphasised at the moment however, that each of the models outlined above need not be regarded as being mutually exclusive, as different combinations of ice types are physically and theoretically possible. In addition, many authors (e.g. Wahrhaftig & Cox, 1959; White, 1979; Haeberli, 1985; Barsch, 1987) believe that valley-wall rock glacier movement involves some combination of internal deformation and basal sliding. The following section, therefore, provides a consideration of basal sliding.

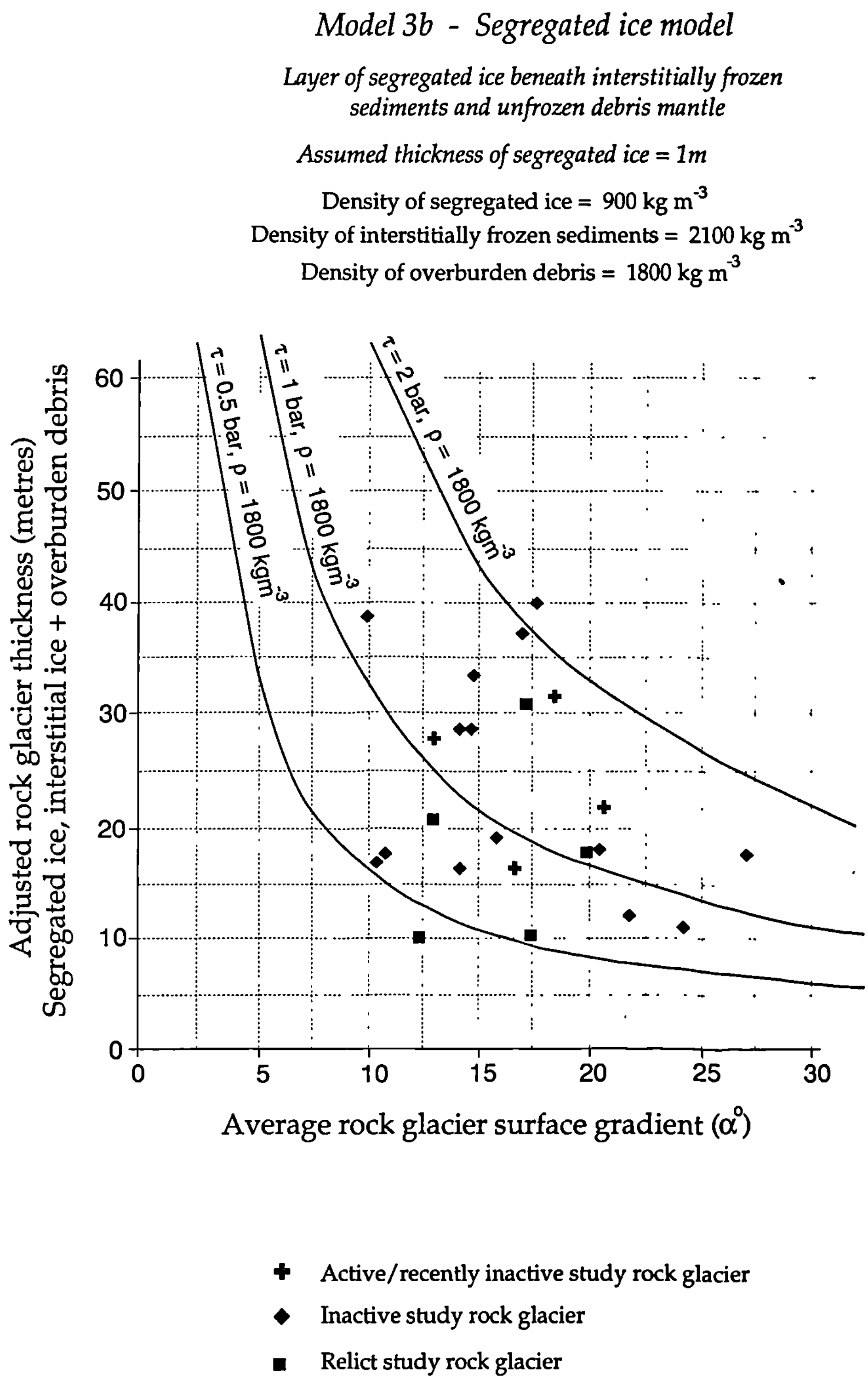


Figure 7.12 Plot of study rock glacier data under the assumptions of Model 3b - the segregated ice model.

7.2.6 Basal sliding

Three lines of field evidence introduced in Chapter 2 indicate that basal sliding may contribute to the movement of some valley-wall rock glaciers. First, electrical resistivity profiling of several valley-wall rock glaciers in Switzerland detected a partly or completely unfrozen layer beneath the main mass of frozen rock glacier sediments, which suggested to Fisch *et al.* (1977) that rock glaciers are not frozen to the underlying bedrock and that movement may be due in part to basal sliding (cf. section 2.5.3). Second, seasonal variations in the movement of the Gruben rock glacier were observed by Haeberli (1985), who likened these to similar movements on temperate glaciers. He suggested that high rates of movement in the downslope section of the rock glacier during the summer and autumn may relate to water flow conditions (cf. section 2.6.2). Although the detailed mechanism of basal movement is not clear, significant amounts of water at the base of rock glaciers may have a 'lubricating' effect as they do in glaciers, probably by reducing effective normal stresses. Large melt-water streams were found emerging from the foot of the frontal slopes of two valley-floor rock glaciers at Ellendalen and Zinal (Plate 7.1), and sub-surface water could be heard beneath the surface of many valley-wall rock glaciers. If plentiful supplies of water are available, hydrostatic pressures may build up within rock glaciers.

The third indirect indication of basal sliding, which was proposed by Wahrhaftig and Cox (1959), is based on the presence of marked talus aprons at the snouts of rock glaciers, which appear to indicate that sliding

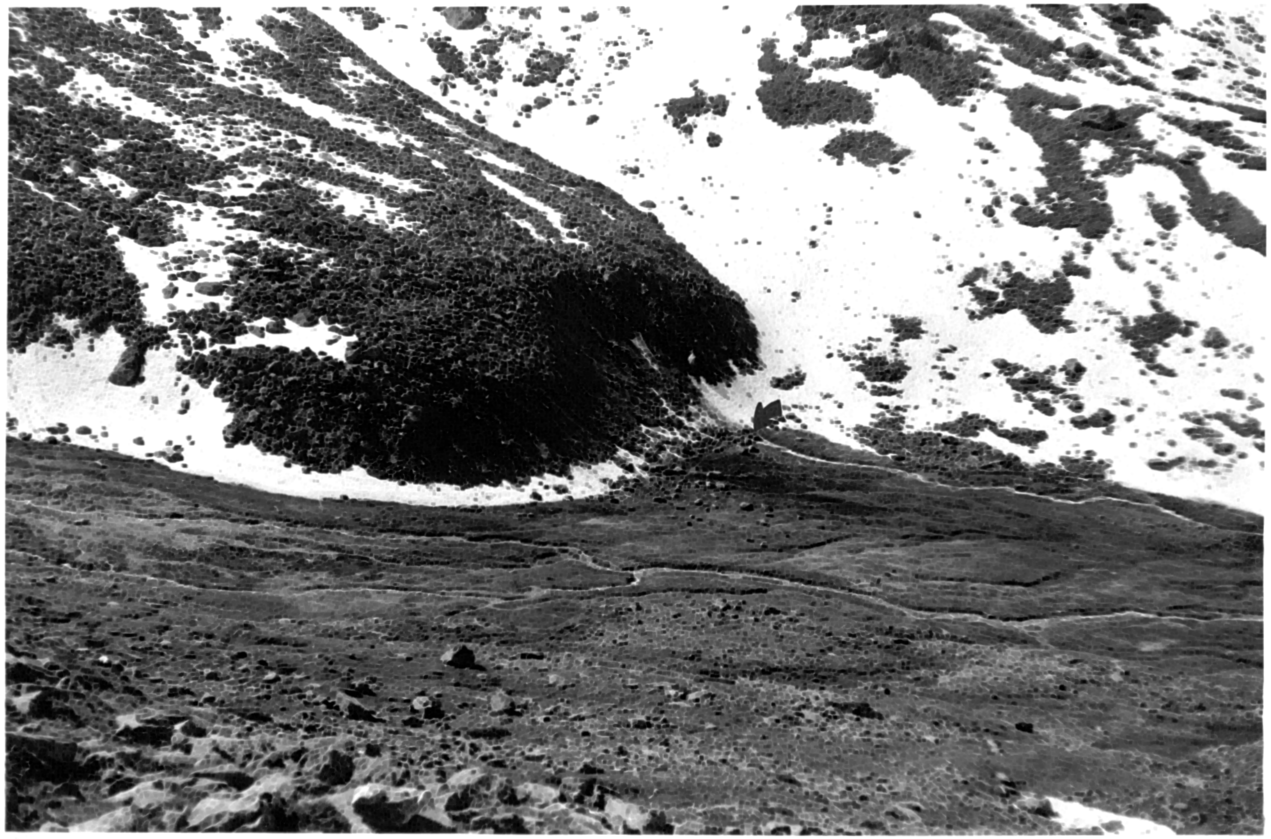


Plate 7.1 Frontal section of valley-floor rock glacier at Zinal, Switzerland, showing stream emerging from base of frontal slope (stream marked with an arrow).

may account for part of their motion (cf. section 2.6.4; Figures 2.7 and 2.8). Additionally, the fact that these talus aprons maintain a relatively constant height suggests that over-run debris may become reincorporated within the rock glacier mass. One further implication of this would appear to be that the base of some rock glaciers may consist of coarse unfrozen talus debris, and basal sliding may be initiated if high pore-water pressure develops in this zone of confined rock debris at the base of the rock glacier. This concept is illustrated in Figure 7.13.

Sliding mechanisms of valley-wall and valley-floor rock glaciers are, however, very poorly understood. Moreover, concepts of glacier sliding are not generally applicable to rock glaciers because of the much smaller velocities involved and the unknown characteristics of the interface at the base of deforming rock glaciers. Thus, the proportion of rock glacier movement that can be attributed to basal sliding is unknown, and probably varies from one rock glacier to the next. However, if basal sliding occurs, valley-wall rock glaciers may move at shear stresses less than those of the critical basal yield stress. This would reduce the critical thicknesses of ice or ice-rich frozen sediments required to induce strain in Models 1 to 3 above.

One important implication of the evidence for basal sliding is that it implies that ice within rock glaciers is near the pressure melting-point. If so, this justifies the plastic models of ice movement outlined above because cold ice, several degrees below the pressure melting point, has a higher yield strength than temperate ice and therefore requires higher shear stresses to initiate strain.

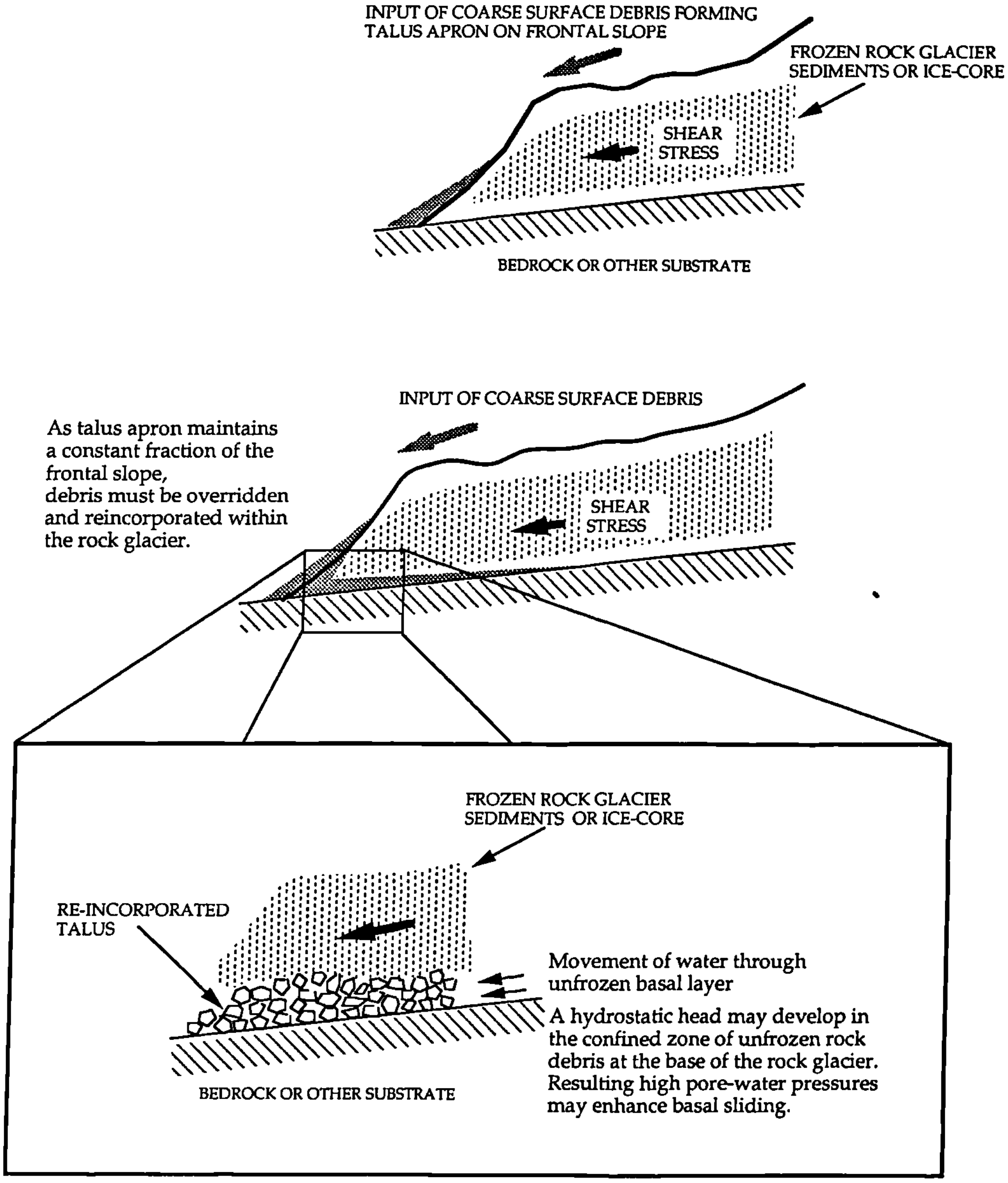


Figure 7.13 Development of high pore-water pressures in unfrozen basal layer of rock glacier.

7.2.7 Summary

The aim of the above analyses was to define theoretical boundary conditions for valley-wall rock glacier formation. From the calculations presented, several significant findings have emerged that have important implications for the formation of valley-wall rock glaciers.

1) Basal shear stress values calculated using field data under the assumptions of model 3b (layer of segregated ice beneath interstitially frozen sediments and an unfrozen debris mantle) are theoretically more feasible than those obtained using any of the other models, as the calculated stress values for model 3b generally exceed threshold limits to movement.

2) In model 1, the glacier ice-cored model, maximum predicted basal shear stresses for the four active study rock glaciers all lie below 1 bar. As this value is regarded as a critical threshold for movement through internal deformation (e.g. Paterson, 1981), model 1 does not appear to explain the movement of these rock glaciers.

3) If the study rock glaciers when active were composed of ice-rich frozen sediments and an unfrozen debris mantle (Model 2), predicted maximum basal shear stresses would have been sufficient for movement to occur at only half of the sites.

4) Predicted maximum basal shear stresses exceed 1 bar for all but one of the active study rock glaciers under the assumptions of model 3b.

5) It is possible that basal sliding contributes to the movement of valley-wall rock glaciers.

In the light of the findings of the above analyses, the feasibility of the seven models of valley-wall rock glacier formation outlined in Chapter 3 may now be considered using both theoretical and empirical evidence.

7.3 Testing the hypotheses of formation

7.3.1 Introduction

The aim of this section is to test the feasibility of each model of valley-wall rock glacier formation outlined in Chapter 3 by considering the field and theoretical findings reported above. Each model of formation is considered individually before a general discussion of valley-wall rock glacier genesis is presented.

7.3.2 The glacier ice-cored model

Early in this thesis (section 3.2.2), the glacier ice-cored model of formation, which is dependent on the presence of glacier ice, was rejected as a *general* model of formation because it is known that valley-wall rock glaciers occur in areas that have never experienced glaciation (e.g. Blagbrough & Farkas, 1968). However, as stated earlier, its rejection as a general model does not negate the possibility that *some* valley-wall rock glaciers may be glacially derived. Indeed, the pronounced morphological similarities that exist between glacially-derived valley-floor rock glaciers and some valley-wall rock glaciers have frequently been cited as arguments in favour of a similar origin for the two categories.

In this section, attention is focused initially on evidence obtained from this study which provides additional support for the rejection of the glacier ice-cored model as a *general* model of formation. Consideration is then given as to whether the model is theoretically valid for any of the study valley-wall rock glaciers. In doing so, particular attention is given to the implications that the glacier ice-cored model has on volume loss during deactivation.

Several lines of evidence have emerged from the research reported in this thesis which support the notion that many, if not all, valley-wall rock glaciers are not glacially-derived. In terms of rock glacier distribution, it was noted that the primary factors that control the regional distribution of valley-wall rock glaciers are temperature and precipitation (cf. Chapter 4). Previously-proposed climatic boundary conditions of valley-wall rock glacier formation (Haeberli, 1985), which were validated using field data from this study, indicate that active valley-wall rock glaciers require a mean annual temperature below *ca.* -1°C to -2°C and an annual precipitation less than 2500mm. Corresponding boundary conditions for ice glaciers and active valley-floor rock glaciers however, are less severe; these features occur in areas where the mean annual temperature is 0°C or even higher. For example, of the fourteen study rock glaciers on the Lyngen peninsula (10 valley-wall and four valley-floor), only the valley-floor rock glacier in Ellendalen is active, despite the fact that mean annual temperature at some of the other more elevated Lyngen sites is lower. Mean annual temperature at the Ellendalen rock glacier is approximately -1°C (section 4.2.4). Such marked differences in the climatic boundary conditions and distributions of valley-wall and valley-floor rock glaciers strongly suggest that valley-wall rock glaciers are not all glacially derived. Furthermore, unlike active valley-floor rock glaciers, all active valley-wall rock glaciers from the study sample occur above the lower boundary of discontinuous permafrost.

In addition, it is significant that in the Scottish Highlands, examples of relict valley-wall rock glaciers, which are thought to date from the Loch Lomond Stadial, all occur outside the limits of the Loch Lomond Readvance, whereas both known examples of valley-floor rock glaciers lie within the limits of this readvance.

Morphological evidence gathered at some of the study sites also points to separate origins for valley-wall and valley-floor rock glaciers. Data presented in Tables 5.1 to 5.3 clearly illustrate that the overall dimensions of valley-wall rock glaciers are smaller than those of glacially-derived valley-floor rock glaciers. In particular, downslope length and thicknesses of valley-wall rock glaciers are usually less than those of valley-floor rock glaciers. Although a small amount of overlap exists between the dimensions of the two categories, differences are sufficient to warrant the adoption of a two-fold size classification, which in itself hints at a genetic distinction.

When attention is focused on the range and morphology of microrelief features on valley-wall and valley-floor rock glaciers, a two-fold classification is not so apparent. Yet the similarity of microrelief features is not surprising, given that both categories of rock glacier probably experience downslope movement of debris by internal deformation of ice, whether the ice is of glacial or non-glacial derivation. Clearly, if ice deformation plays an important role in movement, microrelief structures will exhibit pronounced similarities.

Morphological characteristics of specific study valley-wall rock glaciers provide more explicit evidence for a non-glacial origin. For example, at Fornesdalen West rock glacier on the Lyngen Peninsula in northern Norway (cf Plate 5.1 & 5.2), rock glacier movement appears to have occurred in two distinct phases. As can be seen from the photographs, the central part of the rock glacier has travelled approximately 300m farther downslope than either of the adjoining lateral sections, and rests against a lateral moraine. The morphology of this rock glacier suggests that rock glacier movement began at the approximate height of the present position of the lateral margins and that the central section experienced greater downslope movement possibly as a result of a greater supply of debris. Given that the rock glacier is younger than the underlying lateral moraine, it seems unlikely that a sizeable core of glacial ice could have been preserved on this steep scree slope 300m above an older lateral moraine. Fornesdalen West valley-wall rock glacier cannot therefore be glacial in origin. Other rock glaciers in the study sample, such as those at Trollvatnet and Tverrelvdalen, also overlie lateral moraines and exhibit similar two phase patterns.

Further evidence against a glacial origin for the Lyngen valley-wall rock glaciers is provided by the pattern of glacier decay, as the development of nested sequences of readvance moraines indicates that former Lyngen glaciers retreated actively rather than downwasting in situ (pers. comm. C K Ballantyne). This contradicts one of the assumptions of the glacier-ice cored model, which is that the glaciers must have stagnated in situ in order for sizeable glacier ice-cores to have been preserved.

Sedimentological evidence presented in Chapter 6 provides additional clues to rock glacier origin. Perhaps of greatest relevance to this discussion is the conclusion that the sedimentological structure of relict valley-wall rock glaciers largely resembles that of talus slopes in which a coarse boulder surface layer overlies much finer deposits. This suggests that valley-wall rock glaciers develop through the deformation and downward movement of relatively well established talus slopes. Such evidence is at variance with the notion that a glacial ice-core becomes buried by rockfall debris before talus slopes develop.

Additional sedimentological evidence against the glacier-ice cored model is provided by the results of a series of ice isotopic analyses (section 6.4). Ice obtained from valley-wall rock glaciers in Switzerland was shown to be isotopically distinct from neighbouring glacier ice. Deuterium concentrations of the glacier ice samples were found to be significantly different from both active and inactive rock glacier ice samples. In addition, no physical evidence was found in any rock glacier for a buried glacier-ice core. Indeed the presence of glacial ice within valley-wall rock glaciers has yet to be substantiated by any researcher (cf. Barsch, 1987).

The above empirical findings provide strong support for the initial conclusion that the glacier-ice cored model must be rejected as a *general* model of formation for valley-wall rock glaciers. No amount of empirical evidence, however, can negate the possibility that valley-wall rock glaciers may occasionally result from the deformation of a buried ice-core. In order to take the test a stage further, an attempt was made to establish the theoretical feasibility of the glacier ice-cored model by modelling a number of the factors that control rock glacier movement (section 7.2).

Using field data for average thickness and average surface gradient, predicted maximum basal shear stresses of the four active study valley-wall rock glaciers were found to lie below 1 bar for the glacier ice-core model. As 1 bar is regarded by many as a critical threshold for ice deformation (eg. Paterson, 1981), the movement of the active valley-rock glaciers studied cannot be attributed to the deformation of glacier ice cores.

One final independent argument against the glacier ice-cored model is that only very small differences were recorded in the thickness of active and relict valley-wall rock glaciers; active rock glaciers in the study sample range from 10m to 35.5m in depth compared with a range of 10m to 28m for relict rock glaciers. On deactivation, if rock glaciers contained sizeable ice cores, only a very thin layer of debris should remain. This is clearly not the case. Thus, unless the study grossly underestimates the thickness of the unfrozen debris mantle, which seems unlikely given supporting geophysical evidence from elsewhere (e.g. Fisch *et al.*, 1977; Barsch *et al.*, 1979), this evidence indicates that if a glacial core was present it must have been very thin. However, the thinner the ice-core the less empirically feasible becomes the model.

In sum, the above conclusions strongly indicate that the glacier ice-cored model is not valid for the formation of valley-wall rock glaciers.

7.3.3 *The matrix-ice model*

The matrix ice model assumes that valley-wall rock glaciers are permafrost phenomena that have developed independent of glacier ice. In the past, geophysical investigations and minor excavations have shown that interstitial ice (i.e. ice which fills the voids between particles), and matrix ice (i.e. ice that surrounds and supports individual particles) occur within valley-wall rock glaciers (e.g. Capps , 1910; Ives, 1940; White, 1971; Fisch *et al.*, 1977). Both types of ice were found within active valley-wall rock glaciers in the Swiss field area (see section 6.4). The presence of non-glacial ice within active valley-wall rock glaciers is therefore not in doubt. The question that must be answered is whether the shear strengths of rock-ice mixtures that contain matrix and/or interstitial ice can be exceeded for rock glacier movement to occur.

Early in this thesis (see section 2.2.4), it was stated that the shear strength of a rock-ice mixture in which clast-to-clast contact persists (i.e. interstitial ice content) may actually be greater than one in which no ice exists and only air fills the interstices. For this reason, the model under discussion assumes that matrix ice separates clast-to-clast contacts and that the resulting ice-rich permafrost slowly creeps downslope. The origin of matrix ice, which will be discussed more fully below, has been attributed to the accumulation and refreezing of snow meltwater within the lower layers of talus material where an excess of moisture is present (e.g. Wahrhaftig & Cox, 1959; White, 1976; Corte, 1978).

In many ways, the matrix ice model seems more satisfactory than the glacier-ice cored model as a *general* model of formation. First, the model neither restricts the distribution of valley-wall rock glaciers to those areas that have experienced glaciation nor does it preclude their occurrence in such glaciated areas. Thus the model is consistent with observed distributions. Second, the model accounts for those rock glaciers on the Lyngen Peninsula that record two phases of movement, as the distribution of matrix-ice within talus slopes need not necessarily be limited to one level on the slope. Movement of one part of the talus slope may be triggered by a greater input of debris causing the shear stress within the ice-rich debris to increase. If climatic conditions remain constant and matrix-ice persists, a second section of the slope may deform following a secondary input of debris. Third, volumetric loss on deactivation for this model would be much less than that for the glacial ice-cored model, as the volume of matrix ice is assumed to be significantly less than the volume of an ice-core. Fourth, by proposing a non-glacial mode of formation for valley-wall rock glaciers, the possibility remains that the observed two-fold size classification of valley-wall and valley-floor rock glaciers reflects separate origins.

Sedimentological evidence obtained from the study areas provides further support for the matrix-ice model. It is of great significance that the model conforms to the concept that rock glacier movement involves the deformation of a relatively well established talus slope. This concept, which does not fit the glacier ice-core model, is based on the observed similarity of the sedimentological structure of relict valley-wall rock glaciers and that of talus slopes. Ice isotopic results, presented in section 6.4.3, are also consistent with the hypothesis that valley-wall rock glaciers contain perennially frozen sediments rich in ice that form largely through refreezing of percolating snow and meltwater. Deuterium concentrations in ice from the active rock glaciers were found to be most similar to those of ground ice within talus.

Theoretical analyses described in section 7.2.4 indicate that in terms of the laminar flow of valley-wall rock glaciers, the field data fit the assumptions of the matrix-ice model more readily than they fit the glacial ice-core model. However, neither model is satisfactory. If the study valley-wall rock glaciers when active are assumed to have comprised ice-rich frozen sediments with a density of between 1500 and 1800kgm^{-3} and an unfrozen debris mantle of density 1800kgm^{-3}, maximum basal shear stresses would have been sufficient for movement to occur at only half of the sites on the assumptions of a 1 bar threshold.

In addition, a major drawback to this model concerns the origin of the matrix-ice. Detailed discussion regarding both its formation and definition is absent from the literature. Although the formation of matrix-ice has usually been attributed to the accumulation and refreezing of meltwater within the pore spaces of talus material where an excess of moisture is present, its relationship to other types of ice is poorly understood. It differs from pore and interstitial ice in terms of ice content; matrix-ice separates particle to particle contacts, whereas pore and interstitial ice do not. The higher ice content of matrix-ice is commonly linked to the presence of excess quantities of moisture, which at some sites, exist in the form of snowmelt, rainfall, thaw of the active layer frozen in the previous winter and unusual groundwater conditions. However, the actual process of formation is not understood.

Certainly, however, ice can form a matrix around fine sediments and small clasts within valley-wall rock glaciers. Samples of matrix-ice were obtained from active rock glaciers in Switzerland (section 6.4.3); Plate 5.16 shows clear ice up to 2cm thick surrounding each clast. Numerous investigators have reported the presence of what seems to be matrix-ice. For example, examination of the core of an active rock glacier by Barsch *et al.* (1979) revealed that ice was present in the pore spaces and also "as coatings around rock particles" (p. 215). Although detailed descriptions are lacking, the occurrence of matrix-ice seems to be quite widespread (eg. Lliboutry, 1961; Johnson, 1973; Barsch, 1977). It is possible that the matrix-ice found at Arolla is a type of segregation ice, as Mackay (1971) has noted that "every conceivable gradation between icy muds and pure ice" forms as a result of ice segregation (p. 399).

Notwithstanding the uncertainty regarding matrix-ice formation, important questions pertaining to the theoretical feasibility of its laminar flow remain unanswered. If the ice within the frozen sediments separates clasts only by an order of a centimetre or two, as suggested by the ice samples, it is very probable that downslope creep of such sediments (if indeed it occurs) could not be maintained over any distance, as movement may result in the progressive interlocking of particles leading to an increase in shear stiffness.

In conclusion, although much of the field evidence is consistent with the matrix-ice model, significant difficulties remain concerning the development and behaviour of matrix-ice on a large scale. In addition, the theoretical analyses indicate that maximum predicted basal shear stresses associated with a matrix-ice model are insufficient for movement to occur at half of the rock glacier sites. The modelling implies therefore that the matrix-ice model is inadequate to explain movement unless critical yield stresses are lower than the assumed 1 bar.

7.3.4 The segregation ice model

The segregation ice model assumes that a thin layer or several thin layers of segregated ice near the base of the rock glacier are overlain by interstitially frozen sediments and an unfrozen debris mantle. The model is based on the work of Wayne (1981), who first suggested that the segregation of ice lenses could take place within rock glaciers if groundwater flow is able to maintain water pressure against the base of aggrading permafrost. The origin of segregated ice in periglacial environments is well understood and documented (eg Mackay, 1971, 1983; Washburn, 1979; Williams & Smith, 1989), and so its presence within rock glaciers can be accepted more readily than the presence of matrix-ice.

As the segregation ice model is non-glacial, many of the factors that were relevant to the matrix-ice model also lend support to the adoption of the segregation model as a *general* model of formation. Briefly, these include: 1) the segregation ice model, like the matrix-ice model, does not suffer the distributional limitations of the glacial model; 2) the dual-movement rock glaciers on Lyngen can be explained within its framework as ice segregation may vary both spatially and temporally; 3) volumetric loss on deactivation is likely to be much less than that for the glacial ice-cored model; 4) the two-fold size classification of valley-wall and valley-floor rock glaciers can still be viewed as reflecting genetic differences.

The ice isotopic analyses, which were devised to investigate the origin of rock glacier ice, unfortunately provide inconclusive results in terms of evaluating this model. As indicated in the previous section, the isotopic content of the ice samples obtained from the upper layers of active valley-wall rock glaciers in Switzerland strongly supports the theory of a non-glacial origin. However, evidence of ice segregation was absent from the ice samples, possibly because samples were obtained only from near-surface layers. It was noted, however, that conditions for ice segregation were suitable in terms of temperature and moisture. Theory and observation indicate that the amount of ice segregation that occurs during freezing of a frost-susceptible soil clearly depends in part on water supply. Given the importance of water supply, it is interesting to note that a favoured location for valley-wall rock glaciers is below gullies. An additional impetus for ice segregation is downwards movement of the perennial freezing front, which may result from either climatic deterioration or from the thinning of overburden debris that occurs during rock glacier movement.

One of the most important factors affecting the amount of ice segregation during soil freezing is grain size. The potential for drawing water to the freezing front increases with decreasing pore size and hence decreasing grain size. Several workers have proposed frost susceptibility criteria against which soil samples may be evaluated (eg. Beskow, 1935; Casagrande, 1932; Terzaghi, 1952).

Particle size distributions were constructed for several samples of fine sediments obtained from rock glaciers on the Lyngen Peninsula. The results show that these samples are at the limits of frost susceptibility as proposed by Beskow, although they may be considered frost susceptible if the less stringent criteria suggested by Casagrande and Terzaghi are adopted (cf. Figure 6.15). It should be noted, however, that the sediment samples were obtained from relatively shallow depths (<1.5 metres); geophysical and core evidence suggests that rock glacier sediments at greater depths are finer and hence less likely to be marginal in terms of frost susceptibility (eg. Fisch *et al.*, 1977; Barsch *et al.*, 1979). Furthermore, Mackay (1983) noted that if drainage in coarse-grained sediments is impeded during freezing and pore water pressures increase, "ice segregation might occur even in relatively coarse grained soils" (p.412). Likewise, Williams & Smith (1989) conclude that ice segregation can "occur in permafrost in relatively coarse-grained materials, over long periods" (p.31). Hydrostatic pressures within rock glaciers have been noted by a number of investigators (eg. Giardino, 1983; Haeberli *et al.*, 1979), and are discussed in more detail in section 7.3.6 below.

Strong evidence in support of the segregation ice model comes from the results of the theoretical analyses. Basal shear stress values calculated using field data under the assumptions of ice segregation are more feasible than those obtained using any of the other models. The model that fits field data most closely assumes that a layer of segregated ice of density 900kgm^{-3} develops beneath interstitially frozen sediments (density=2100kgm^{-3}) and an unfrozen debris mantle (density =1800kgm^{-3}). Under this assumption, predicted basal shear stresses exceed 1 bar for all but one of the active study rock glaciers. Some doubt surrounds the estimated value of average thickness for the active rock glacier that exhibits a calculated basal shear stress of less than 1 bar (cf. sections 5.4.3 and 7.2.4).

It is significant that not one line of evidence within the range of empirical and theoretical findings presented in this thesis contradicts the proposition that the segregation ice model represents a *general* model of valley-wall rock glacier formation. Indeed, in spite of the inconclusive results of the ice isotopic analyses, which are thought to reflect the limits of the sample base, numerous findings lend strong support to the notion that ice segregation plays an important role in the formation of valley-wall rock glaciers.

7.3.5 The firn-field model

The fourth model of formation is based on the presence of a firn-field over which rockfall debris slides and accumulates at the base as a protalus rampart. It is envisaged that if debris supply is maintained, the firn may become buried and subsequently deform to form a valley-wall rock glacier.

As indicated earlier in this thesis (cf. section 3.2.5), it is improbable that the firn-field hypothesis represents a *general* model of valley-wall rock glacier formation as it is rare for perennial firn fields to be of sufficient size to explain the formation of the largest valley-wall rock glaciers.

Very wide valley-wall rock glaciers such as at Strath Nethy, where the frontal margin is well defined and continuous over a distance of 2.4km, clearly do not fit into the framework of the model. In addition, Ballantyne & Kirkbride (1986) noted that Lateglacial protalus ramparts in upland Britain rarely exceed 0.3km in length, an order of magnitude smaller than some valley-wall rock glaciers. Liestøl (1962) proposed that a linked series of firn fields may merge to allow the development of broad valley-wall rock glaciers. However, given the continuous nature of the frontal margin at Strath Nethy and the fact that such a linked series of firn fields along one valley-wall was not observed at any of the field sites, this seems an unlikely hypothesis. Furthermore, no evidence was found either in Norway or in the Alps for any transition through the burial of firn fields.

It is possible however, that ice from firn fields may have become incorporated within some valley-wall rock glaciers. The occurrence of massive and apparently non-segregated ice lenses in rock glaciers within non-glaciated areas has been noted and attributed to the burial of snowbank ice. Lliboutry (1961), for example, attributed intermittent ice masses found in rock glaciers in the Chilean Andes to the burial of snowbanks by rockfall. Similarly, Johnson (1973) found intermittent ice masses overlying discontinuous permafrost in rock glaciers in Alaska. Numerous examples of firn-fields occurring in close proximity to many of the study rock glaciers lends further support to the likelihood that snowbank ice may have been incorporated within some valley-wall rock glaciers. In a similar geomorphic context, Østrem (1963, 1971) has shown by crystallographic analyses that snowbank ice commonly becomes incorporated within ice-cored moraines. However, analyses of ice samples removed from active rock glaciers studied in Switzerland provided no hint of the presence of former snowbank ice. Indeed, ice sampled from the study rock glaciers was found to be most similar to ground ice and least similar to the snowbank group. These results do not of course preclude the possibility that snowbank ice is present in other valley-wall rock glaciers.

Some authors, in particular Haeberli (1985), are strong advocates for a landform continuum between protalus ramparts and valley-wall rock glaciers. In section 6.5.2, it was noted that locational and sedimentological criteria are virtually identical for protalus ramparts and incipient valley-wall rock glaciers; both form at the feet of talus slopes and both contain rockfall debris. It is not unreasonable, therefore, to propose that a continuum exists, although direct evidence to support the idea has yet to be found. Strong arguments in favour of a continuum can be obtained however from studying the formation of rock glaciers that develop downslope of lateral and end moraines. The example of one such rock glacier at Bukkeholsbreen in Jotunheimen is used to illustrate the argument. The Bukkeholsbreen rock glacier, which was studied by Vere & Matthews (1985) and visited by the present author in 1986, is a typical example of a simple rock glacier that has formed from a discrete lateral moraine. Sedimentological evidence gathered by Vere & Matthews suggests that its development was triggered by renewed increments of debris following fluctuations in the glacier snout soon after the culmination of the 'Little Ice Age' advance. Two main sources of ice are envisaged; remnants of glacier ice and the subsequent burial of snowbank ice. Thus, given that increments of debris and snowbank ice to the lateral moraine at Bukkeholsbreen have resulted in the development of a small rock glacier, it is not unreasonable to assume that similar inputs to protalus ramparts may produce downslope movement and the development of small valley-wall rock glaciers.

In conclusion, available evidence suggests that snowbank ice may become incorporated within valley-wall rock glaciers. The importance of such ice, however, remains questionable. If lenses of sufficient size become buried, it is possible that their subsequent deformation may contribute to downslope movement. Additionally, it seems possible that some incipient valley-wall rock glaciers may have evolved from protalus ramparts. It is even possible that many if not all protalus ramparts may be incipient valley-wall rock glaciers. The size of rock glaciers derived from protalus ramparts must be limited by the dimensions of firn fields. This dimensional restriction implies that the firn-field model is inadequate to explain the formation of the majority of valley-wall rock glaciers.

7.3.6 The hydrostatic pressure model

Up to this point, hypotheses for the movement of valley-wall rock glaciers have emphasised internal deformation of ice as the primary cause of movement. Giardino (1983), however, hypothesised that hydrostatic pressure from water trapped below an envelope of ice is the primary cause of downslope movement of rock glaciers on Mount Mestas in Colorado. In Giardino's model, pressure applied at the head of the rock glacier, such as an input of debris, increases shear stress and aids movement along a frozen/unfrozen interface. He proposed that this interface could be either within the rock glacier at the base of the zone of frozen debris or at the contact between the rock glacier and the underlying slope. His hypothesis stems from a study of the sedimentological characteristics of the Mount Mestas rock glaciers; specifically, that the trend of the orientation of clasts relative to the direction of surface slope indicates that catastrophic flow events have contributed to movement. His results show that the rock glacier fabric is distinct from that of clasts associated with glacial deposition. The link, however, between clast fabric and hydrostatic flow events is speculative. In addition it should be emphasised that the unusual geological conditions of the Mount Mestas field area lend themselves more readily to this model than most other geological settings. Giardino himself acknowledges that because of steep slopes and extensive shale and mudstone deposits, "the area as a whole is predisposed to failure" (p. 300).

The idea that high pore water pressures may induce rock glacier flow has also been put forward by a number of other authors. For example, both Maxwell (1979) and Shroder (1987) have noted the presence of hydrostatic pressure conditions in rock glaciers within their field areas and have suggested that such conditions may play an important role in rock glacier movement, although they offer no direct evidence to support this association. Haeberli *et al.* (1979), however, provided more persuasive evidence in support of Giardino's ideas. In their study of an active rock glacier near the Grubengletscher, they noted the widespread occurrence of subsidence (0.1 to 0.2 m yr^{-1}). From this observation, they inferred that underground ice at the site is not in thermal equilibrium with the present climate so that massive ice "simultaneously melts out and slides as a whole on a well-developed slip surface" (p. 434).

Giardino's observations are given further credence by the probable association of some rock glaciers with basal meltwater. Barsch (1975) noted a greater movement of the Murtel rock glacier in the summer months and discovered a positive relationship between high summer temperatures and relatively rapid movement. Shroder (1978) also noted a strong positive relationship between increased precipitation and increased movement in an "ice-cemented" rock glacier. These observations suggest that some rock glaciers may move *en masse* in a manner similar to some landslips: hydrostatic pressure developed beneath an impermeable frozen layer may reduce effective normal stresses and hence basal shear resistance, thus allowing movement. Hydrostatic pressure has been recognised for many years to be a prime cause of rock slide movement (Terzaghi, 1962; Whalley, 1974, 1983; Brunsden, 1979).

To a large extent, Giardino's model seems to be a reasonable hypothesis. Osborn (1985), in his discussion of the paper, stated that "it is hard to deny the presence of water under pressure in some rock glaciers" (p. 374). The crucial issue concerning Giardino's model is a question of scale. The model cannot be viewed as a *general* model of rock glacier formation, as there is no evidence to support the notion that the primary cause of valley-wall rock glacier formation is hydrostatic pressure. Indeed, the occurrence of valley-wall rock glaciers in very cold arid permafrost environments, such as Antarctica and the Arctic (eg. Humlum, 1982; Hassinger & Mayewski, 1983), cannot be explained by hydrostatic pressure as the presence of basal meltwater in such environments is not plausible. Giardino's field evidence, which itself remains questionable, is limited to one field area with unusual geologic characteristics. The widespread occurrence of "interstitial ice" that Giardino noted within the Mount Mestas rock glaciers points to the more likely conclusion that downslope creep of frozen sediments remains the primary cause of valley-wall rock glacier movement, and that hydrostatic pressure may assume no more than a secondary role at certain favourable sites. This picture conforms with the majority of field observations concerning hydrostatic pressure (eg. Barsch, 1975; Shroder, 1978; Haeberli *et al.*, 1979), which suggest that water under pressure may in certain situations lead to a localised increase in rock glacier movement. It is also worthwhile to note that a build up in hydrostatic pressure requires the development of an overlying impermeable (frozen) layer.

The study sites in Norway and Switzerland furnished no direct evidence to suggest that hydrostatic pressure played any role in rock glacier movement. In section 6.2.1 it was reported that fabric studies such as those described by Giardino (1983, 1985) could not be undertaken as component clasts at the study sites were insufficiently elongate for the measurement of preferred orientation. The presence of sub-surface water within active rock glaciers is not in question, however, as running water could be heard at many sites. In addition, the morphological characteristics of longitudinal depressions provide indirect evidence for the presence of water (cf. section 5.5.3). The depressions, which trend downslope in a sinuous manner and exhibit approximately symmetrical rectilinear sides, have probably formed as a result of melting of internal ice along paths of water flow. It is possible therefore that hydrostatic pressures may build up if spring melt is trapped under an existing impermeable frozen layer. However, given that free movement of water militates against a build up of confined hydrostatic pressure, the observed abundance of running water during the summer at some of the study sites implies that hydrostatic pressures do not occur later in the melt season.

Additional field evidence obtained from this study strongly supports the rejection of hydrostatic pressure as the primary cause of downslope movement. The morphological regularities examined in section 5.4.5, which are statistically significant for rock glaciers from a wide variety of field areas, favour slow continuous or intermittent movement rather than catastrophic sliding. For example, the occurrence of compression-type transverse ridges and the positive correlation between rock glacier length and depth indicates that rock glacier development is based on a series of thresholds, not on the build up and release of hydrostatic pressure. In addition, field and photogrammetric surveys indicate that active valley-wall rock glaciers move downslope at average speeds of 1-160cm/yr by a series of slow intermittent movement events (cf. Table 2.2; section 2.6.2). On the other hand, intermittent build-up and release of hydrostatic pressure could cause movement during which water under pressure is released.

In conclusion, although not a *general* model of formation, the hydrostatic pressure model has some relevance if it is viewed as a process model of site-specific significance.

7.3.7 The retrogressive slab failure model

The retrogressive slab failure model proposed by Maxwell (1979) is based on his hypothesis that "catastrophic processes are significant in the formation and movement of rock glaciers" (p. 11). "From the basic assumption that a talus could fail retrogressively" (p.150), Maxwell developed models to attempt to simulate rock glacier features. Unfortunately, he put forward no evidence to substantiate his basic assumption. He merely said that "it is possible that a deep-seated failure can occur in a permafrost talus slope due to the cohesive influence of small amounts of ice" (p. 35). However, evidence of large-scale cracks was not found at any of the study sites even where the talus was frozen. Similarly, an extensive search of the literature pertaining to periglacial environments also failed to produce any supporting observations for Maxwell's assumptions.

Given the very dubious background to the model, it not surprising that it is contradicted by field evidence. For example, the morphological characteristics of the study rock glaciers do not fit into the framework of the slab failure model. In particular, it was noted earlier that the geometry of the slope is related to rock glacier length; surface slope gradients are steeper for smaller valley-wall rock glaciers and more gentle for longer features (cf. section 5.4.5). This pattern, which is linked to the form of the rock glacier and, in particular, to the length of the mid section of the rock glacier where transverse ridges occur, strongly suggests a development continuum not consistent with retrogressive failure. In terms of sedimentology, the distribution of large rock glacier clasts, which generally occur down the entire length of the rock glacier, shows that their rate of input has kept pace with rock glacier movement. Furthermore, the occurrence of fall-sorting on low gradient proximal slopes of transverse ridges, indicates that these slope angles have decreased since deactivation by the melt-out of ice. In addition, the morphology and sedimentology of the frontal slopes of active rock glaciers indicate that they are experiencing differential forward motion. Finally, measurements of rock glacier movement are consistent with slow creep, not with catastrophic failure (eg. White, 1971; Hassinger & Mayewski, 1983; Haeberli, 1985)

Clearly, Maxwell's slab failure model is inappropriate as a *general* model of formation.

7.3.8 The avalanche model

In section 3.2.8, it was stated that the avalanche model could not be considered as a *general* model of formation for valley-wall rock glaciers because avalanching produces only simple landforms on which complex flow ridges are absent. The following factors provide additional support for this claim:

1) For the avalanche model to be true generally, all valley-wall rock glaciers would have to be associated with avalanche-modified talus slopes. However, in Chapter 4 it was noted that many of the study rock glaciers have formed at the base of apparently unmodified rockfall talus slopes. Similarly, many documented examples of valley-wall rock glaciers occur in association with rockfall talus rather than avalanche-modified talus (eg. Wahrhaftig & Cox, 1959; White, 1971; Shakesby *et al.*, 1987).

2) The sedimentology of avalanche-modified talus is very distinctive,

particularly in terms of the looseness of surface debris. Clast analysis undertaken at the study sites in Norway and Switzerland shows that the rock glaciers are characterised by generally well-packed debris.

3) The absence of present-day avalanche activity at the active valley-wall rock glaciers studied in Switzerland suggests that the avalanche model is inappropriate for the formation of these rock glaciers.

3) Topographic constraints on the width of avalanche-induced landforms such as avalanche boulder tongues (e.g. Rapp, 1959; Gardner, 1970) would not permit the development of wide valley-wall rock glaciers such as those at Strath Nethy and Zinal.

4) There are no documented examples of landforms that are transitional between valley-wall rock glaciers and avalanche boulder tongues.

5) The morphology of frontal slopes of valley-wall rock glaciers implies that they experience differential forward motion.

The above evidence indicates that the avalanche model is inappropriate as a *general* model of formation. Avalanching has been proposed by Maxwell (1984) to explain some of the more simple forms of valley-wall rock glaciers. However, the working definition of rock glaciers adopted at the beginning of this thesis necessitates the presence of both ice and internal deformation. Thus, small-scale tongue-like accumulations formed by avalanching alone cannot be viewed as valley-wall rock glaciers.

In effect, the avalanche model is appropriate for the formation of avalanche boulder tongues such as those described by Luckman (1978). Avalanches may contribute debris to valley-wall rock glaciers but they do not contribute to the forward movement of such debris. There is therefore no process or form continuum.

7.3.9 Summary

From the arguments presented above, only one of the models proposed in Chapter 3 emerges as a satisfactory *general* model of valley-wall rock glacier formation. Those that were rejected as general explanations are: the glacier ice-cored model, the matrix-ice model, the firn-field model, the hydrostatic pressure model, the avalanche model, and the retrogressive slab failure model. Only the segregated ice model appears to offer a valid general model of valley-wall rock glacier genesis. It is much more difficult, however, to exclude the possibility of alternative modes of formation under particular circumstances. Also, it is clear that even if ice segregation is the fundamental cause of valley-wall rock glacier formation and movement, other processes such as snow avalanching, basal sliding and deformation of snowbank or matrix ice may constitute secondary contributing mechanisms. The following section presents a comprehensive description of valley-wall rock glacier genesis based on the findings of this thesis.

7.4 Valley-wall rock glacier genesis

Perhaps the two most significant findings of section 7.3 are, first, that valley-wall rock glacier development is independent of glacier ice, and second, that ice segregation appears to play a fundamental role in the formation of valley-wall rock glaciers. Given these two basic principles, valley-wall rock glaciers must be viewed as permafrost phenomena in which the main cause of movement is creep of ground ice or frozen soil. Clearly, a number of factors and processes play a role in valley-wall rock glacier formation; understanding of these landforms is simplified if the geographic framework within which valley-wall rock glaciers develop is outlined before a study of the processes of formation is presented.

Detailed analysis of the distribution of the study rock glaciers highlighted a number of important climatic, microclimatic and non-climatic controlling factors (Chapter 4; cf. Figure 4.6). On a regional scale, the primary factors that control distribution are temperature and precipitation; active valley-wall rock glaciers appear to occur only where mean annual temperature is below *ca.* -1°C to -2°C, and average annual precipitation is less than approximately 2500mm with the precipitation threshold dropping as mean annual temperature declines (Haeberli, 1985). At the local level, altitude, aspect and topographic location appear to have a major controlling influence on microclimate and hence on valley-wall rock glacier distribution. For example, in each of the study areas, north and northwesterly aspects were found to be most common and southerly aspects least common for valley-wall rock glaciers, due to reduced insolation and hence reduced mean annual temperatures at northerly sites. Where topographic shading significantly reduces insolation, valley-wall rock glaciers with more southerly aspects may occur, as at Tytebærtind on the Lyngen Peninsula. Sites that experience the greatest amounts of snowcover, most commonly those with northeasterly aspects, tend to favour the development of ice glaciers rather than rock glaciers. This pattern is particularly evident in the Swiss field areas. Altitudinally, all of the active study valley-wall rock glaciers occur above the lower boundary of discontinuous permafrost, although several rock glaciers above this lower boundary were found to be inactive, suggesting that non-climatic factors also affect the activity status of valley-wall rock glaciers.

The development of valley-wall rock glaciers is also dependent on the debris supply characteristics of a particular site. Bedrock source-wall characteristics, such as lithology and joint density, were found to be important. Massive rocks such as gneiss, granite and gabbro, which yield abundant coarse rockfall debris on weathering appear to be the most favourable lithologies for rock glacier formation, although valley-wall rock glaciers will form on these lithologies only if rockwall height is sufficient. All valley-wall rock glaciers within the study areas are located at the foot of talus slopes, generally beneath fractured cliffwalls of massive rock. Most have formed on basal slopes inclined gently away from the foot of the talus.

The joint frequency of the bedrock source-wall was noted to have a controlling influence on rock glacier development. Several of the Scottish valley-wall rock glaciers were found to be associated with major rock-slope failures. On the granitic rocks of the Cairngorms, these failures occurred along dilation joints that are aligned parallel to the rockwall surfaces. In Scotland, therefore, the distribution of many valley-wall rock glaciers appears to correspond to sites where talus accumulated rapidly and possibly catastrophically. This pattern may be a reflection of the short period during which the Lateglacial temperature and precipitation regime in Scotland was suitable for valley-wall rock glacier formation. Widespread development of glacier ice below rockwalls during the Loch Lomond Stadial (the only period during which permafrost conditions prevailed following ice-sheet deglaciation), together with the brevity of full-stadial conditions (*ca.* 400-800yr) appear to have placed restrictions on more widespread valley-wall rock glacier development.

In terms of the processes involved in valley-wall rock glacier formation, there is a need to separate those that are active on the bedrock source-wall and talus from those active within the rock glacier, as it is important to recognise that the processes that control the supply of debris to the rock glacier and those that cause the rock glacier to move are different, though linked in the sense that a continued supply of debris appears necessary to sustain movement.

In the previous section and from sedimentological analyses presented in section 6.3, it was noted that valley-wall rock glacier formation results from the deformation and downward movement of an established ice-rich talus slope. The initial development of a valley-wall rock glacier therefore is dependent upon identifying a process or group of processes responsible for generating the debris necessary for initiation of creep in the frozen deposits. For the majority of rock glaciers, the dominant debris-supply mechanism appears to be discrete inputs of rockfall debris.

Most valley-wall rock glaciers are associated with rockfall talus slopes, and clast analyses undertaken at the Norwegian and Swiss study sites reveal a pattern of poorly-sorted and generally well-packed debris that is characteristic of rockfall talus. Indeed, at many of the active study rock glaciers, discrete rockfalls were observed throughout the summer field season. In addition, it is significant that on the study rock glaciers large surface clasts generally occur down the entire length of the landform; they are not concentrated at the downslope margin. This clast-size distribution suggests that debris input keeps pace with rock glacier movement.

Most of the rockfalls observed at the study sites consisted of individual or small numbers of fragments falling from the cliff wall and rolling or sliding down the talus slope, which at some sites was snow-covered well into the summer season. Tracks of numerous similar minor rockfalls could be traced on the snow-covered talus slopes at many of the study sites in Switzerland. No evidence of snow avalanching was found at any of the Norwegian or Swiss rock glaciers studied, although it is known that one of the main debris-supply mechanisms for protalus ramparts on the Lyngen Peninsula is snow and slush avalanching (Ballantyne, 1987). Given the close proximity of protalus ramparts and valley-wall rock glaciers on Lyngen and the fact that snow and slush avalanches are very common on the Lyngen mountains, it seems plausible that avalanche transport of debris may here constitute a secondary supply of debris.

There is some evidence to indicate that debris-supply mechanisms at some rock glacier sites may be characterised by larger but less frequent joint-controlled rockslides. For example, at Flüela-Wisshorn rock glacier in the Swiss Alps, the largest clasts on the rock glacier are much larger than those on the corresponding talus slope (cf. section 6.2.5). This suggests that large-scale rockfall events have been associated with the formation of this rock glacier.

In addition, as mentioned above, rock-slope failure scars at five of the Scottish rock glacier sites strongly suggest that the rate of debris supply at these sites probably exceeded that normally associated with discrete inputs of rockfall debris. A few attempts have been made to assess rates of rockfall activity during the Loch Lomond Stadial by calculating rock glacier volume at some of the Scottish sites. Sissons (1976) calculated 17m of stadial rockwall retreat for the massive debris accumulation at Baosbheinn in Wester Ross, which the present author interprets as a valley-wall rock glacier (cf. section 4.2.5; Plates 7.2 & 7.3). Although this figure of 17m was found to be an overestimate in a subsequent reinterpretation of the deposit by Ballantyne (1986), Ballantyne's adjusted figure of *ca.* 14.3m for average rockwall retreat during the Loch Lomond Stadial still implies a massive input of rock debris. A rockwall retreat of *ca.* 14.3m corresponds to a rockwall retreat rate of 14.3mm/yr if the stadial is assumed to have lasted 1000 years or a rate of 19mm/yr if 750 years is assumed as stadial duration. High rates of rockwall retreat for Loch Lomond Stadial valley-wall rock glaciers in Scotland were also calculated by Dawson (1977), who estimated a stadial rockwall retreat of 9.2m for a valley-wall rock glacier on Jura.

Dawson stressed that the value of 9.2m was offered as a maximal value for rockwall retreat as post-glacial talus aggradation may have occurred.

Plate 7.2 Lateral view of a relict valley-wall rock glacier at Baosbheinn, Scotland, showing travel distance away from talus slope.

Plate 7.3 View of the 31m high frontal slope at Baosbheinn. In the background, the source area can be seen. Note that the most extensive rockwall areas overlook the flanks of the ridge, not the centre, yet the rock glacier is thickest in the centre.

Certainly, the unknown contribution of *pre*-stadial debris to valley-wall rock glaciers places some doubt on calculations of Loch Lomond Stadial rockwall retreat rates from the volumes of such features. However, the calculated rockwall retreat rates for the valley-wall rock glaciers at Baosbheinn and on Jura are approximately an order of magnitude greater than estimates of present-day rockwall retreat in high alpine environments (eg. Barsch, 1977). It is significant also that the calculated stadial rates for Baosbheinn and Jura greatly exceed those recorded for contemporaneous protalus ramparts in upland Britain (Ballantyne & Kirkbride, 1987). Ballantyne & Kirkbride proposed stadial rockwall retreats of between 1.14-1.61m on the basis of the volume of debris in eight protalus ramparts. The contrast between the calculated rates of rockwall retreat implied by the volumes of protalus ramparts and valley-wall rock glaciers strongly suggests that intermittent rockfall alone could not have produced large valley-wall rock glaciers during the Loch Lomond Stadial and lends further credence to the notion of major rockslides contributing to the development of at least some of the Scottish valley-wall rock glaciers.

Moreover, evidence obtained from aerial photographs of each of the study sites also emphasises the likelihood that a higher proportion of debris in Scottish rock glaciers has been derived from large-scale rock-slope failures rather than discrete rockfall events. A positive relationship was observed between rock glacier length and the height of the corresponding talus for the samples of Norwegian and Swiss study rock glaciers, but relict talus slopes above valley-wall rock glaciers in Scotland are much smaller than expected in terms of this relationship. The most pronounced example of this anomaly is at Strath Nethy where rock glacier length exceeds 350m, yet the talus slope is less than 100m long. In contrast at Fornesdalen East in Lyngen, where rock glacier lengths are also 350m, talus slopes are over 200m long.

In sum, evidence obtained from the study areas is generally consistent with the hypothesis that rockfall has been the main debris-supply mechanism for valley-wall rock glaciers. At some sites, however, discrete rockfalls appear to have been supplemented by snow avalanching and larger-scale rock-slope failures; such a pattern seems most likely for many of the Scottish valley-wall rock glaciers. Given sufficient input of debris, it is envisaged that the overburden load increased to a level at which the shear strength of the ice-rich talus was exceeded and movement occurred. The relationship between debris-supply mechanisms and movement mechanisms may be viewed in terms of a series of thresholds, and not as a continuum of process.

Before turning our attention to the processes active within the ice-rich talus, it is important to review our understanding of the climatic conditions present at the time of rock glacier formation. Of the models of formation that were proposed in Chapter 3, only the segregation ice model satisfies all available evidence as a *general* model of formation. This model assumes that a layer or several layers of segregated ice within the lower layers of talus are overlain by interstitially frozen sediments and an unfrozen debris mantle. The origin of such segregated ice is well understood. The process involves the migration of water in fine sediments to the freezing plane at temperatures below 0°C as a result of differences in the free energy levels of water and ice; gradients of potential between the higher free energy of the water and the lower free energy of the ice enable migration of available water to the freezing plane. The amount of ice segregation depends greatly on soil type, rates of freezing and groundwater conditions, and there is great variation in the form of such ice. Most segregated ice is in layers or lenses that are aligned horizontally. In addition to this segregated ice, significant amounts of pore or interstitial ice may form due to the refreezing of percolating melt and rainwater in seasonally-frozen debris.

The presence of ice modifies the properties and behaviour of the talus mass. It is known that the strength of a frozen soil increases with an increase in ice content up to the point at which the soil pores become completely filled with ice (Nickling & Bennett, 1984). This is due to the cementing effect of pore ice; cohesion increases many times due to the bonding between ice crystals and soil particles. With further increases in ice content, however, the strength of frozen ground decreases and becomes subject to time-dependent creep behaviour similar to that of ice. The role and extent of subsequent deformation depends upon the magnitude of load, the temperature, soil composition and ice content and distribution (McRoberts, 1975).

It is likely that overburden loading under accumulating debris will promote deformation in basal segregated ice layers first, as these layers will require the smallest additional shear stress to enter creep. Although the precise behaviour of ice-rich talus under stress is not understood, it is envisaged that if loading of rockfall debris is maintained for a considerable time, significant downslope deformation of the basal ice-rich sediments will occur. The amount of creep will undoubtedly vary even over short distances because the composition of ice-rich talus is not uniform. Thus the rock glaciers on the Lyngen Peninsula that record two distinct phases of movement may reflect unequal distribution and growth of segregation ice or possibly variations in debris supply. Moreover, if hydrostatic pressures develop as a result of impeded drainage, downslope movement rates may be temporarily enhanced within part of the rock glacier.

The morphological characteristics of the frontal slopes of the study valley-wall rock glaciers strongly suggest that the upper layers of active rock glaciers move downslope at a velocity greater than that of the rest of the feature. This pattern of differential forward motion is consistent with the hypothesis that creep is the principal movement mechanism, as differential strain is maximised at the base of the rock glacier where shear stress is greatest. Moreover, the morphological regularities examined in section 5.4.5, which are statistically significant for rock glaciers from a wide variety of field areas, imply process regularities such as those produced by creep, rather than some more catastrophic form of movement.

On many of the study rock glaciers a series of inner transverse ridges occurs immediately upslope of the frontal ridge. In section 5.5.3, it was noted that these transverse ridges are convex downslope, suggesting that their formation may be related to the movement of the rock glacier. Given the form of these inner transverse ridges and the observed increase in number of ridges for progressively larger rock glaciers, it seems reasonable to suggest that their development is in some way related to debris-induced thresholds. The presence of marked cross-sectional asymmetry on these ridges (steep distal slopes and much gentler proximal slopes), further suggests that they may develop due to internal shearing as a result of compressive flow. Evidence from the Arctic has shown that upward shearing of massive segregated ice bodies can occur even where overburden depths exceed 30m (Williams, 1968). Overburden depths on the study rock glaciers were estimated to be between 3 and 7.5m (cf. section 5.5.2). Although the segregated ice layers are probably thinner than the ice bodies observed by Williams, it seems plausible that internal shearing along these layers may deform the overburden forming a transverse ridge.

Cessation of rock glacier movement appears to be a response to at least one of three factors: first, a decrease in the rate of debris supply; second, a change in ground temperature regime causing melt of ground ice; and third, reduction in shear stress as a consequence of movement.

7.5 Conclusions

The aims of this chapter were to define theoretical boundary conditions for rock glacier movement, and to evaluate the feasibility of seven models of valley-wall rock glacier formation using both empirical and theoretical evidence.

Theoretical evidence for valley-wall rock glacier formation was obtained by applying a simple laminar flow equation to a number of different density models that reflect different distribution of internal ice. The calculations, which incorporated field measurements of rock glacier thickness and average surface gradient, demonstrated that the most feasible model of valley-wall rock glacier formation is one in which a layer of segregated ice is overlain by interstitially frozen sediments and an unfrozen debris mantle. Basal shear stress values calculated for each of the study rock glaciers under the assumptions of this model are higher than those obtained in any of the alternative models, and for the most part exceed an assumed critical threshold of 1 bar.

Interestingly, calculated basal shear stress values for a glacier ice-cored model all lay below the 1 bar critical threshold for movement, suggesting that it is unlikely that the study valley-wall rock glaciers contained large glacially-derived ice-cores when active. Furthermore, if the study rock glaciers when active were composed of ice-rich frozen sediments and an unfrozen debris mantle, maximum basal shear stresses would have been sufficient for movement to occur at only half of the sites on the assumptions of a 1 bar threshold. Some evidence was found in support of the notion that basal sliding contributes to the movement of valley-wall rock glaciers.

In the second half of the chapter, attention was focused on a detailed consideration of each model of valley-wall rock glacier formation outlined at the beginning of this thesis. Only one model, the segregated ice model of formation, was found to offer a valid general explanation of valley-wall rock glacier genesis. However, other processes such as snow avalanching and basal sliding under conditions of high hydrostatic pressure were shown to constitute possible secondary contributing mechanisms of formation and movement.

The chapter concludes with a summary discussion of genesis, in which valley-wall rock glaciers are presented as permafrost phenomena where the main cause of movement is creep of ground ice or frozen soil. A discussion of the processes involved in their formation was centred on the important distinction between those processes that control the supply of debris to the rock glacier and those that cause the rock glacier movement.

Chapter 8

Conclusions

8.1 Introduction

In Chapter 1, it was noted that the rock glacier literature shows a clear bias towards the study of *valley-floor rock glaciers,* and published work on the formation of *valley-wall rock glaciers* exhibits little consensus. The main objective of the research reported in this thesis has been to test several hypotheses of valley-wall rock glacier formation in order to determine which, if any, are feasible. Evidence pertinent to testing such hypotheses has been obtained both from field data and theoretical considerations. In this final chapter, the major findings of the research are summarised before avenues for future work within this field are explored.

8.2 Major findings

8.2.1 Hypotheses of formation

Review of the literature suggests that seven major models of formation constitute the crux of current debate on the origin of valley-wall rock glaciers. The primary conclusion of the research reported here is that only one of these models, the segregation ice model, emerges as a satisfactory *general* model of valley-wall rock glacier genesis. This model assumes that a thin layer or several thin layers of segregated ice near the base of the rock glacier are overlain by interstitially frozen sediments and an unfrozen mantle of coarse debris. It is envisaged that overburden loading under accumulating debris promotes deformation in the basal segregated ice layers first, as these layers will require the smallest additional shear stress to enter creep. If loading of rockfall debris is maintained, downslope deformation of the basal ice-rich sediments will occur. The assumptions of this model indicate that valley-wall rock glaciers are permafrost phenomena that have developed independent of glacier ice.

A wide range of empirical and theoretical findings have been shown to be consistent with the assumptions of the segregation ice model. Indeed, not one line of evidence could be found to contradict the proposition that the model represents a potential *general* model of valley-wall rock glacier formation. Several lines of field evidence collected from four major study areas support the segregation ice model of formation, as follows.

1) In terms of rock glacier distribution, the segregation ice model neither restricts the distribution of valley-wall rock glaciers to those areas that have experienced glaciation nor does it preclude their occurrence in such glaciated areas, and thus is consistent with observed distributions. In addition, field-validated temperature and precipitation boundary conditions for active valley-wall rock glaciers are consistent with requirements for ice segregation.

2) In terms of sedimentology, it was noted that the structure of relict valley-wall rock glaciers largely resembles that of talus slopes, in which a coarse boulder surface layer overlies much finer deposits. The

segregation ice model conforms to the concept that rock glacier movement involves the deformation of a relatively well established talus slope.

3) Valley-wall rock glaciers on the Lyngen Peninsula that record two distinct phases of movement can be readily explained under the assumptions of the model as ice segregation may vary both spatially and temporally. Unequal distribution and growth of segregation ice may produce varying amounts of creep even over short distances.

4) Only small differences were recorded in the thickness of active and relict valley-wall rock glaciers. As the volume of segregated ice is assumed to be small, limited volumetric loss on deactivation is consistent with the segregated ice model.

5) Particle size distributions for several samples of fine sediments obtained from relatively shallow depths within valley-wall rock glaciers on the Lyngen Peninsula show that these samples lie within the limits of frost susceptibility.

6) The results of ice isotopic analyses strongly support the theory of a non-glacial origin. Although discrete ice lenses were absent from the rock glaciers sampled (possibly because samples were obtained only from near-surface layers), the results are consistent with the presence of near-surface non-glacial frozen sediments that have formed largely through refreezing of percolating snow and meltwater. Deuterium concentrations in ice taken from these near-surface layers in active rock glaciers are most similar to those of ground ice within talus.

Evidence in support of the segregation ice model was also found in the results of theoretical analyses. Theoretical evidence was obtained by applying a simple laminar flow equation to a number of different density models that reflect different distribution of internal ice. The calculations, which incorporated field measurements of rock glacier thickness and average surface gradient, demonstrated that the ice segregation model is the most feasible model of valley-wall rock glacier formation. Basal shear stress values calculated for each of the study rock glaciers under the assumptions of this model are higher than those obtained in any of the alternative models, and for the most part exceed an assumed critical threshold of 1 bar.

Six hypotheses of valley-wall rock glacier formation were rejected as *general* models on the basis of field and theoretical work. They are: the glacier ice-cored model, the matrix-ice model, the firn-field model, the hydrostatic pressure model, the avalanche model, and the retrogressive slab failure model. However, although these models do not appear to be valid at a general level, the possibility cannot be excluded of alternative modes of valley-wall rock glacier formation or movement under particular circumstances. Snow avalanching, deformation of snowbank or matrix ice, and basal sliding under conditions of high hydrostatic pressure all constitute possible contributing mechanisms of formation and movement in particular cases.

8.2.2 Additional significant findings

In addition to providing data for testing the models of valley-wall rock glacier formation, the study sites in Norway, Switzerland and Scotland furnished several new findings worthy of note. These include the following.

1) The interdependence of climatic, microclimatic and non-climatic factors in controlling valley-wall rock glacier distribution at the local level was demonstrated at many of the study sites. Although temperature and precipitation have long been recognised as the primary factors that control rock glacier distribution on a regional scale, factors such as lithology, source rockwall height, aspect, altitude, and topographic location were shown to play an important role in determining local distribution. For example, although all of the active valley-wall rock glaciers studied in Switzerland were found to occur above the lower boundary of discontinuous permafrost, several rock glaciers above this lower boundary were found to be inactive where debris supply had decreased.

2) Despite considerable variability in the form of long profiles surveyed on the study rock glaciers, bivariate statistical analyses revealed a number of morphometric regularities. In particular, strong positive relationships were observed between rock glacier length and maximum thickness, and between rock glacier length and slope geometry. As length/thickness ratios were found to be similar for active, inactive and relict valley-wall rock glaciers, thickness appears to remain a relatively constant fraction of rock glacier length from initial formation to deactivation.

3) Detailed examination of a large number of rock glaciers has

demonstrated that the frontal ridges of valley-wall rock glaciers possess certain common morphological and sedimentological characteristics. Four characteristic units were identified. These are: 1) a frontal ridge crest comprising a steep distal and gentle proximal slope characterised by a concentration of very large clasts; 2) an upper rectilinear slope between 3m and 7.5m long that constitutes the steepest gradient on any part of the rock glacier; 3) a lower rectilinear slope underlain by much finer deposits and characterised by lower gradients than the upper slope; and 4) a basal talus apron comprising fall-sorted clasts that have accumulated mainly by rolling and sliding down the frontal slope. Morphological differences between the frontal slopes of active and inactive rock glaciers reflect the presence or absence of differential forward movement and may usefully be described within the framework of this four-unit subdivision. On deactivation, maximum facet angles are not maintained and lower average gradients appear to be a response to slumping, melt of internal ice and settling.

4) Unlike rockfall talus cones, valley-wall rock glaciers do not exhibit a general downslope increase in clast-size. Clast-size measurements at a number of study sites revealed that large clasts generally occur down the entire length of valley-wall rock glaciers; large clasts are not concentrated only at the downslope margins. This pattern suggests that the deposition of large clasts at the foot of talus slopes during rock glacier formation must keep pace with rock glacier movement.

5) Observations in each of the major field areas are consistent with the hypothesis that rockfall is the main debris-supply mechanism for valley-wall rock glaciers. At some sites, however, discrete rockfall was

shown to be supplemented by snow avalanching and larger-scale rock-slope failures. Rock-slope failure scars at five of the Scottish study sites suggest that the rate of debris supply at these sites probably greatly exceeded that normally associated with discrete inputs of debris. In addition, at some study sites, the largest rock glacier clasts appeared to be greater than the largest clasts on adjacent talus slopes, which suggests that large-scale rockfall events are likely to have been associated with the formation of these rock glaciers. Clast-size measurements at one of the Swiss study sites provided statistical support for this observation.

6) Chemical and granulometric analyses of fine rock glacier sediments sampled from valley-wall rock glaciers in Norway indicate that the majority of fine material within rock glaciers appears to have been inherited from pre-existing talus deposits. A combination of field and theoretical evidence also suggests that lateral moraine sediments may have become incorporated within some of the study rock glacier during formation.

7) Ice samples obtained from a number of genetically distinct ice groups in Switzerland provided a number of important findings. Isotopically, ice samples obtained from an active valley-wall rock glacier proved significantly different from both neighbouring snowbank and glacier ice. Physical and isotopic analyses of a range of ice samples obtained from near-surface locations suggest that such ice forms largely through the refreezing of percolating rain and meltwater. These results also strongly support a non-glacial origin for the ice in valley-wall rock glaciers.

8.3 Future directions

A number of possible pathways for further investigations can be identified from the research reported in this thesis. Two methodologies, new to rock glacier research, have proved to be of value in providing empirical and theoretical data. First, ice isotopic studies were shown to be useful in determining the origin of rock glacier ice, and second, simple laminar flow modelling provides theoretical boundary conditions for rock glacier movement and formation. Further experimentation and refinement to each approach would undoubtedly be of value.

The isotopic measurements presented in this thesis constitute the only known comparative measurements to have been made on rock glacier ice, glacier ice, ice from perennial firn fields, ground ice and lake water. The partial separation of the ice groups that was achieved on the basis of deuterium concentrations alone demonstrates the potential for an isotopic approach when comparing genetically distinct ice groups. Measurement of both deuterium and oxygen isotopes, however, would provide even better discrimination between ice groups. In particular, this procedure could be most fruitfully applied in locations where rock glacier ice is exposed at depth.

The modelling work presented in Chapter 7 is based upon numerous simplifying assumptions (eg. steady longitudinal flow, uniformly sloping bed, and a constant overburden thickness) that are unlikely ever to be matched by actual field conditions. However, in spite of these shortcomings, the approach provides a theoretical basis for comparing the validity of different models of rock glacier movement. The general model can be readily applied to other valley-wall rock glaciers as field measurements are required only for average thickness and average surface gradient. It would be useful, therefore, to use a wider data set to test the generality of the boundary conditions determined by this study. In addition, the basic model could be developed to examine the effect of transverse spreading and variations in debris input, both of which may help to explain the development of microrelief features such as transverse ridges and longitudinal depressions.

A third area of future research concerns the potential for using rock glacier volume to assess rates of debris supply. This methodology is valid only for those areas where the duration of rock glacier activity is known. In Scotland, for example, valley-wall rock glacier development presumably corresponds closely to the duration of the Loch Lomond Stadial as this is the only period during which the Lateglacial temperature and precipitation regime was suitable for the development of permafrost. Regional comparisons of rockwall retreat rates would help to determine the importance of major rockslides in contributing to the development of valley-wall rock glaciers.

Essentially, this thesis confirms the views of Haeberli (1985), Barsch (1987), and others, who regard valley-wall rock glaciers as permafrost phenomena. Perhaps the most promising avenue of future research that leads on from the conclusion that permafrost and <u>not</u> glacial conditions are required for formation, is the use of valley-wall rock glaciers for interpreting past climates. The palaeoclimatic information that may be obtained from valley-wall rock glaciers is at present limited by a lack of detailed knowledge on the precise climatic constraints for initial formation and development of individual valley-wall rock glaciers. Thus, further research should concentrate on the climatic boundary conditions, and in particular the precipitation requirements, for actively developing valley-wall rock glaciers. Given the worldwide distribution of these landforms, valley-wall rock glaciers may provide a valuable record to past climatic changes in many mountain environments.

References

Alley, R.B., Blankenship, D.D., Bentley, C.R., and Rooney, S.T., 1986. Deformation of till beneath ice stream B, West Antarctica. *Nature,* v.32.

Ambach, W., Eisner, H., Elsässer, M., Löschhorn, U., Moser, H., Rauert, W. and Stichler, W., 1976. Deuterium, tritium and gross-beta activity investigations on Alpine glaciers (Oetztal Alps). *Journal of Glaciology,* v. 17, p. 383-400.

Andrews, J.T., 1975. Glacial Systems. An approach to glaciers and their environments, *Duxberry Press.*

Ballantyne, C.K., 1984. The Late Devensian Periglaciation of Upland Scotland. *Quaternary Science Reviews,* v. 3, p. 311-343.

Ballantyne, C.K., 1986. Protalus rampart development and the limits of former glaciers in the vicinity of Baosbheinn, Wester Ross. *Scottish Journal of Geology,* v.22, n.1, p. 13-25.

Ballantyne, C.K., 1987a. Some observations on the morphology and sedimentology of two active protalus ramparts, Lyngen, northern Norway. *Arctic and Alpine Research,* v. 19, n. 2, p. 167-174.

Ballantyne, C.K., 1987b. The Beinn Alligin 'Rock Glacier', in, *Ballantyne, C.K. & Sutherland D.G., eds.,* Wester Ross Field Guide. Quaternary Research Association, Cambridge.

Ballantyne, C.K. and Kirkbride, M.P., 1986. The characteristics and significance of some Lateglacial protalus ramparts in upland Britain. *Earth Surface Processes and Landforms,* v. 11, p. 659-671.

Ballantyne, C.K. and Kirkbride, M.P., 1987. Rockfall activity in upland Britain during the Loch Lomond Stadial. *The Geographical Journal,* v.153, p. 86-92.

Barry, R.G., 1981. Mountain Weather and Climate, *Methuen, London.*

Barsch, D., 1969. Studien und Messungen an Blockgletschern in Macun, Unterengadin. *Zeitschrift fur Geomorphologie, Supplementband,* v.8, p. 11-30.

Barsch, D., 1971. Rock glaciers and ice-cored moraines. *Geografiska Annaler,* v. 53A, p. 203-206.

Barsch, D., 1977a. An estimation of talus production around and transport by active rock glaciers in the Swiss Alps. *Zeitschrift fur Geomorphologie,* v. 28, p. 148-160.

Barsch, D., 1977b. Nature and importance of mass-wasting by rock glaciers in alpine permafrost environments. *Earth Surface Processes,* v. 2, p. 231-245.

Barsch, D., 1977c. Alpiner Permafrost-Ein Beitrag zu seiner Verbreitung, zur seiner Charakteristik und seiner okologie am Beispiel der Schweizer Alpen. *Abhandl.d. Akadamied. Wissenschaft. Gottingen, Math-Physik.*, Kl. 3, 3, Folge, Nr. 31, p. 118-141.

Barsch, D., 1978. Active rock glaciers as indicators for discontinuous alpine permafrost: an example from the Swiss Alps. *3rd International Conference on Permafrost, Edmonton, 1978*, Proc. 1, p.348-353.

Barsch, D., 1987a. Rock glaciers: an approach to their systematics, in, *Giardino, J.R., Shroder, J.F., Jr., & Vitek, J.D., eds.*, Rock Glaciers. Allen and Unwin.

Barsch, D., 1987b. The problem of the ice-cored rock glacier, in, *Giardino, J.R., Shroder, J.F., Jr., & Vitek, J.D., eds.*, Rock Glaciers. Allen and Unwin.

Barsch, D., Fierz, H., and Haeberli, W., 1979. Shallow core drilling and bore-hole measurements in the permafrost of an active rock glacier near the Grubengletscher, Wallis, Swiss Alps. *Arctic and Alpine Research*, v. 11, n. 2, p. 215-228.

Barsch, D. and Hell, G, 1975. Photogrammetrische beweguns - messungen am blockgletscher Murtèl 1, Oberengadin, Schweizer Alpen. *Zeit. für Gletscherkunde und Glazialgeol.*, Bd. XI, Heft. 2, p.111-142.

Barsch, D. and King, L., 1975. An attempt to date fossil rock glaciers in Grison, Swiss Alps. *Questiones Geographicae* (Poznan), v. 2, p. 5-14.

Benedict, J.B., 1973. Origin of rock glaciers. *Journal of Glaciology*, v. 12, p. 520-523.

Benedict, J.B., Benedict, R.J., and Sanville, D., 1986. Arapaho Rock Glacier, Front Range, Colorado, U.S.A.: A 25-Year Resurvey. *Arctic and Alpine Research*, v.18, n.3, p.349-352.

Beskow, G., 1935. Tjäbildningen och tjällyftningen med särskild hänsyn till vägar och jarnägar. *Sveriges Geologiska Undersökning*, Arsbok 26, Ser. C., No. 375.

Birnie, R.V., 1978. Rock debris transport and deposition by valley glaciers in South Georgia. *Unpub. Ph.D. thesis*, University of Aberdeen.

Birnie, R.V. and Thom, G., 1982. Preliminary observations on two rock glaciers in South Georgia, Falkland Islands dependencies. *Journal of Glaciology*, v.28, p.377-386.

Bishop, W.W. and Coope, G.R., 1977. Stratigraphical and faunal evidence for Lateglacial and Early Flandrian environments in Scotland, in, *Gray, J.M. & Lowe, J.J., eds.*, Studies in the Scottish Lateglacial environment. Pergamon, Oxford.

Blagbrough, J.W. and Breed, W.J., 1967. Protalus ramparts on Navajo Mountain, Southern Utah. *American Journal of Science,* v.265, p.759-772.

Blagbrough, J.W. and Farkas, S.E., 1968. Rock glaciers in the San Mateo Mountains, south-central New Mexico. *American Journal of Science,* v. 266, p. 812-823.

Bouët, M., 1972. Climat et Météorologie de la Suise Romande. *Lausanne, Payot.*

Boulton, G.S., 1986. Isotopic fractionation at the base of polar and sub-polar glaciers. *Journal of Glaciology,* v. 32, n. 112, p. 475-485.

Boulton, G.S. and Eyles, N., 1979. Sedimentation by valley glaciers: a model and genetic classification, in, *Schlüchter, C. ed.,* Moraines and Varves: Origin/Genesis/Classification. Balkema, Rotterdam.

Boulton G.S. and Jones, A.S., 1979. Stability of temperate ice caps and ice sheets resting on beds of deformable sediment. *Journal of Glaciology,* v. 24, n. 90, p. 29-43.

Briggs, D., 1977. Sediments and methods in geography. *Butterworths. London.*

Brunsden, D., 1979. Mass movements, in, *Embelton, C. and Thornes, J., eds.,* Process in Geomorphology. Arnold, London.

Bryant, B., 1971. Movement measurements on two rock glaciers in the eastern Elk Moutains, Colorado. *U.S. Geological Survey Professional Paper* , n.750-B, p.B108-116.

Brown, W.H., 1925. A probable fossil glacier. *Journal of Geology,* v. 33, p.464-466.

Calkin, P.E., Haworth, L.A., and Ellis, J.M., 1987. Rock glaciers of central Brooks Range, Alaska, U.S.A., in, *Giardino, J.R., Shroder, J.F., Jr., & Vitek, J.D., eds.,* Rock Glaciers. Allen and Unwin.

Capps, S.R., 1910. Rock glaciers in Alaska. *Journal of Geology,* v. 18, p.359-375.

Carrara, P.E. and Andrews, J.T., 1975. Holocene glacial/periglacial record: northern San Juan Mountains, southwestern Colorado. *Zeit. f. Gletscherkunde und Glazialgeol.,* v. 11, p. 155-174.

Casagrande, A., 1932. Discussion, in, *Benkelman, A.C. & Olmstead, E.R.,* A new theory of frost heaving. Natl. Research Council Highway Research Board, Proc. 11th Ann. Mtg.

Chaix, A., 1923. Les coulees de blocs du Parc National Suisse d'Engadine. *Le Globe (Organe de la Societe de Geographie de Geneve; Journal Geographique),* v. 62, (Memoires), 34pp.

Chandler, R.J., 1973. The inclination of talus, Arctic talus terraces and other slopes composed of granular materials. *Journal of Geology*, v.81, n. 1, p.1-14.

Chattopadhyay, G.P., 1984. A Fossil Valley-Rock Glacier in the Cairngorm Mountains. *Scottish Journal of Geology*, v. 20, n. 1, p.121-125.

Church, M., Stock, R.F., and Ryder, J.M., 1979. Contemporary Sedimentary Environments on Baffin Island, N.W.T. Canada: Debris Slope Accumulations. *Arctic and Alpine Research*, v. 11, n.4, p.371-402.

Corner, G.D., 1980. Preboreal deglaciation chronology and marine limits of the Lyngen-Storfjord area, Troms, North Norway. *Boreas*, v. 9, p.239-249.

Corte, A.E., 1976. Rock Glaciers. *Buil. Peryglac.*, v. 26, p. 175-97.

Corte, A.E., 1978. Rock glaciers as permafrost bodies with a debris cover as an active layer. A hydrological approach. Andes of Mendoza, Argentina. *Proc. 3rd Int. Conf. on Permafrost* Vol. 1, Nat. Res. Council Canada.

Corte, A.E., 1987. Rock glacier taxonomy, in, *Giardino, J.R., Shroder, J.F., Jr., & Vitek, J.D., eds.*, Rock Glaciers. Allen and Unwin.

Cross, W. and Howe, E., 1905. Geography and general geology of the quadrangle in Silverton Folio. *U.S. Geol. Survey, Geologic Folio*, n.120 (Silverton), p. 1-25.

Dawson, A.G., 1977. A Fossil Lobate Rock Glacier in Jura. *Scottish Journal of Geology*, v. 13, n. 1, p. 37-42.

Domaradzki, J., 1951. Blockstrome in Kanton Graubunden. *Ergebnisse der Wissenschaff Untersuchungen der Schweiz.* Nationalparks, N.S.3, p. 173-235.

Drewry, D.J., 1987. Glacial Geologic processes, *Edward Arnold.*

Dumbell, E., 1984. The Characteristics, Origin and Genesis of Talus Foot Rock Glaciers in Lyngen, Norway. *Unpubl. B.Sc. dissertation*, University of St. Andrews, 67p.

Eisbacher, G.H., 1979. Cliff collapse and rock avalanches (stürzstroms) in the Mackenzie Mountains, northwestern Canada. *Canadian Geotechnical Journal*, v. 16, p. 309-334.

Ellis, J.M. and Calkin, P.E., 1979. Nature and distribution of glaciers, Neoglacial moraines, and rock glaciers, east-central Brooks Range, Alaska. *Arctic and Alpine Research*, v. 11, p. 403-420.

Evin, M., 1985. Contribution a l'etude de la macrofissuration dans les parois situees a l'amont des glaciers roucheux des Alpes du Sud (France, Italie). *Revue de Geomorphologie Dynamique*, v. 34, n. 1, p.17-30.

Evin, M., 1987. Lithology and fracturing control of rock glaciers in southwestern Alps of France and Italy, in, *Giardino, J.R., Shroder, J.F., Jr., & Vitek, J.D., eds.,* Rock Glaciers. Allen and Unwin.

Finlay, D.P., 1987. The role of late-lying snow in promoting chemical weathering in the Arolla valley, southern Switzerland. *Unpubl. B.Sc. dissertation,* University of St. Andrews, 56p.

Fisch, W.,Sr., Fisch, W.,Jr., and Haeberli, W., 1977. Electrical D.C. resistivity soundings with long profiles on rock glaciers and moraines in the Alps of Switzerland. *Z. Gletscherkd. Glazialgeol.,* v. 13, p.239-260.

Gardner, J., 1970. Geomorphic significance of avalanches in the Lake Louise District, U.S.A. *Arctic and Alpine Research,* v. 2, n. 2, p.135-144.

Giardino, J.R., 1983. Movement of ice-cemented rock glaciers by hydrostatic pressure: an example from Mount Mestas, Colorado. *Zeitschrift fur Geomorphologie,* v. 27, n. 3, p. 297-310.

Giardino, J.R., Shroder, J.F., Jr., and Lawson, M.P., 1984. Tree-ring analysis of movement of a rock glacier complex on Mt. Mestas, Colorado, U.S.A. *Arctic and Alpine Research,* v.16, n.3, p.299-309.

Giardino, J.R., Shroder, J.F., Jr., and Vitek, J.D., *eds.,* 1987. Rock Glaciers, Allen and Unwin.

Giardino, J.R. and Vitek, J.D., 1985. A statistical interpretation of the fabric of a rock glacier. *Arctic and Alpine Research,* v. 17, n. 2, p165-177.

Giardino J.R. and Vitek, J.D., 1988. The significance of rock glaciers in the glacial-periglacial landscape continuum. *Journal of Quaternary Science,* v. 3, n.1, p. 97-103.

Gorbunov, A.P., 1983. Rock glaciers of the mountains of middle Asia. *Proc. 4th Int. Conf. on Permafrost* Vol. 1, p. 359-362. National Academy Press, Washington.

Grabczak, J., Niewodniczanski, J., and Rózanski, K., 1983. Isotope stratification in high mountain glaciers: examples from the Peruvian Andes and Himalaya. *Journal of Glaciology,* v. 29, n. 103, p. 417-424.

Gray, J.T., 1972. Processes and rates of development of talus slopes and protalus rock glaciers in the Ogilvie and Wernecke Mountains, Yukon Territory. *Unpubl. Ph.D. thesis,* McGill University, Montreal.

Griffey, N.J. and Whalley, W.B., 1979. A rock glacier and moraine - ridge complex, Lyngen Peninsula, north Norway. *Norsk geogr. Tidschr.,* v.33, p.117-124.

Griffiths, T.M., 1958. A theory of rockstream morphology. *Annals of the Assocaition of American Geographers (Abstract),* v. 48, p. 265-266.

Grout, M.A., 1982. Sourcewall relief chiefly influences the current supply of rockfall debris to rock glaciers, northern Sangre de Cristo Mountains, Colorado. *Geological Society of America (Abstracts with programs)*, v. 14, n. 6, p.313.

Haeberli, W., 1975. Untersuchungen zur Verbreitung von Permafrost zwischen Fluelpass und Piz Grialetsch (Graubunden). *Mitteilung der Versuchsanstalt fur Wasserbau, Hydrologie und Glaziologie*, ETH Zurich, Nr. 17, 221p.

Haeberli, W., 1982. Klimarekonstuktiomen mit gletscher-permafrost-bezeihungen. *Materialien zur Physiogeographie*, v. 4, p. 9-17.

Haeberli, W., 1985. Creep of Mountain Permafrost: Internal Structure and Flow of Alpine Rock Glaciers. *Mitteilungen der Versuchsanstalt fur Wasserbau, Hydrologie und Glazialogie*, Nr. 77.

Haeberli, W., King, L. and Flotron, A., 1979. Surface movement and lichen cover studies at the active rock glacier near the Grubengletscher, Wallis, Swiss Alps. *Arctic and Alpine Research*, v.11, n. 4, p. 421-441.

Harris, S.A., 1981. Distribution of active glaciers and rock glaciers compared to the distribution of permafrost landforms, based on freezing and thawing indices. *Canadian Journal of Earth Sciences*, v.18, n.2, p. 376-381.

Hassinger, J.M. and Mayewski, P.A., 1983. Morphology and dynamics of the rock glaciers in southern Victoria Land, Antarctica. *Arctic and Alpine Research*, v. 15, n. 3, p. 351-368.

Hodge, S.M., 1974. Variations in the sliding of temperate glaciers. *Journal of Glaciology*, v. 13, p. 349-367.

Hollermann, P., 1983. Rock glaciers as meso-scale landforms of the periglacial belt studies in high mountains of Europe and N. America. *Bonner Geographische Abhandlungen*, v. 67, 1983, 83p.

Holmes, G., 1984. Rock slope failure in parts of the Scottish Highlands. University of Edinburgh, *unpubl. Ph.D. thesis.*

Howarth, P.J. and Bones J.G., 1972. Relationships between process and gemetrical form on High Arctic debris slopes, south-west Devon Island, Canada. *Inst. Br. Geogr. Spec. Pubn.*, n.4, p. 139-153.

Howe, E., 1909. Landslides in the San Juan Mountains, Colorado. *U.S. Geological Survey Professional Paper*, n. 67.

Humlum, O., 1984. Altitudinal trends of talus-derived lobate rock glaciers on Disko, west central Greenland. *Geografisk Tidsskrift*, v. 84, p. 35-39.

Humlum, O., 1988. Rock glacier appearance level and rock glacier iniatition line altitude: a methodological approach to the study of rock glaciers. *Arctic and Alpine Research*, v. 20, n. 2, p. 160-178.

Ives, R.L., 1940. Rock glaciers in the Colorado Front Range. *Bulletin of the Geological Society of America*, v. 51, n. 9, p. 1271-1294.

Jackson, L.E. and MacDonald, G.M., 1980. Movement of an ice-cored rock glacier, Tungsten, N.W.T., Canada, 1963-1980. *Arctic*, v.33, n.4, p.842-847.

Johnson, J.P., 1974. Some problems in the study of rock glaciers, in *Fahey, B.D. and Thompson, R.D.*, eds., *Research in Polar and Alpine Geomorphology*, p. 84-91.

Johnson, J.P. and Nickling, W.G., 1979. Englacial temperature and deformation of a rock glacier in the Kluane Range, Yukon Territory, Canada. *Canadian Journal of Earth Science*, v.16, p. 2275-2283.

Johnson, P.G., 1974. Mass movement of ablation complexes and their relationship to rock glaciers. *Geografiska Annaler*, v. 56A, p. 93-101.

Johnson, P.G., 1975. Mass movement processes in Metaline Creek, southwest Yukon Territory. *Arctic*, v. 28, p. 130-139.

Johnson, P.G., 1978. Rock glacier types and their drainage systems, Grizzly Creek, Yukon Territory. *Canadian Journal of Earth Sciences*, v. 15, n.9, p. 1496-1507.

Johnson, P.G., 1979. Rock glaciers: glacial and non-glacial origins. *International Assc. of Hydrological Sciences Publ. 126.* Proc. of the Workshop at Riederalp, Switzerland, 1978, organised by the temporary secretariat for the World Glacier Inventory, p. 285-293.

Johnson, P.G., 1980. Glacier-Rock Glacier Transition in the Southwest Yukon Territory, Canada. *Arctic and Alpine Research*, v. 12, n. 2, p.195-204.

Johnson, P.G., 1981. The structure of a talus-derived rock glacier deduced from its hydrology. *Canadian Journal of Earth Science*, v. 18, p.1422-1430.

Johnson, P.G., 1983. Rock glaciers. A case for a change in nomenclature. *Geografiska Annaler*, v. 65A, p. 1-2.

Johnson, P.G., 1984. Rock Glacier Formation by High-Magnitude Low - Frequency Slope Processes in the Southwest Yukon. *Annals of the Association of American Geographers* , v. 74, n. 3, p. 408- 419.

Johnson, P.G., 1984. Paraglacial conditions of instability and mass movement. A discussion. *Zeitschrift fur Geomorphologie*, v. 28, n.2, p. 235- 250.

Johnson, P.G., 1987. Rock glacier: glacier debris systems or high-magnitude low-frequency flows?, in, *Giardino, J.R., Shroder, J.F., Jr., & Vitek, J.D., eds.*, Rock Glaciers. Allen and Unwin.

Johnson, R.B., 1967. Rock streams of Mount Mestas, Sangre de Cristo Mountains, southern Colorado. *U.S. Geological Survey Professional Paper,* 575-D, p. D217-D220.

Kendall, R.F., 1891. On a Moraine-Like Mound near Snowdon. *The Glacialists Magazine,* p.68-70.

Kerschner, H., 1978. Paleoclimatic inferences from late Wurm rock glaciers, Eastern Central Alps, Western Tyrol, Austria. *Arctic and Alpine Research,* v. 10, p. 635-644.

Kesseli, J.E., 1941. Rock streams in the Sierra Nevada, California. *Geographical Review,* v.31, p.203-227.

Liestøl, O., 1962. Talus terraces in Arctic regions. *Norsk. Polarinstitutt Arbok,* 1961 Yearbook, p. 102-105.

Lindner, L. and Marks, L., 1985. Types of debris slope accumulations and rock glaciers in South Spitsbergen. *Boreas,* v. 14, p. 139-153.

Lliboutry, L., 1977. Movement of a covered glacier embedded within a rock glacier. *Journal of Glaciology,* v.18, p.255-273.

Lorrain, R.D. and Demeur, P., 1985. Isotopic evidence for relict Pleistocene glacier ice on Victoria Island, Canadian Arctic Archipelago. *Arctic adn Alpine Research,* v. 17, n. 1, p. 89-98.

Luckman, B.H., 1978. Geomorphic work of snow avalanches in the Canadian Rocky Mountains. *Arctic and Alpine Research,* v. 10, n.2, p.261-276.

Luckman, B.H. and Crockett, K.J., 1978. Distribution and characteristics of rock glaciers in the southern part of Jasper National Park, Alberta. *Canandian Journal of Earth Sciences,* v. 15, p. 540-550.

Mackay, J.R., 1971. The origin of massive icy beds in permafrost, western Arctic coast, Canada. *Canadian Journal of Earth Sciences,* v.8, n.4, p. 397-422.

Mahoney, W.C., 1980. Late Quaternary rock glaciers, Mount Kenya, East Africa. *Journal of Glaciology,* v. 25, p. 492-497.

Matthews, J.A., 1981. Quantitative and statistical approaches to Geography. *Pergamon, Oxford.*

Matthews, J.A. and Petch, J.R., 1982. Within-valley asymmetry and related problems of Neoglacial lateral moraine development at certain Jotunheimen glaciers, southern Norway. *Boreas,* v. 11, p.225-247.

Maxwell, M., 1979. The role of catastrophic processes in the evolution of talus derived rock glaciers. *Unpubl. M.A. Thesis,* Dept. of Geography, Univ. Ottawa, 197p.

Means, W.D., 1976. Stress and strain. Basic concepts of continuum mechanics for geologists. *Springer-Verlag.*

Messerli, B. and Zurbuchen, M., 1968. Blockgletscher im Weissmies und Aletsch und ihre photogrammetrische Kartierung. *Die Alpen SAC*, v. 3, p. 139-152.

Morris, S.E., 1981. Topoclimatic factors and the development of rock glacier facies, Sangre de Cristo Mountains, Southern Colorado. *Arctic and Alpine Research*, v. 13, n. 3, p. 329-338.

Nickling, W.G. and Bennett, L., 1984. The shear strength characteristics of frozen coarse granular debris. *Journal of Glaciology*, v.30, n.106, p. 348-357.

O'Neil, J.R., 1968. Hydrogen and oxygen isotope fractionation between ice and water. *Journal of Physical Chemistry*, v. 72, n. 10, p. 3683-3684.

Osborn, G.D., 1975. Advancing rock glaciers in the Lake Louise Area, Banff National Park, Alberta. *Canadian Journal of Earth Sciences*, v.12, p.1060-1062.

Ostrem, G., 1971. Rock glaciers and ice-cored moraines, a reply to D. Barsch. *Geografiska Annaler*, v. 53, p. 207-213.

Ostrem, G. and Arnold, K., 1970. Ice-cored moraines in southern British Columbia and Alberta, Canada. *Geografiska Annaler*, v. 52, p.120-128.

Outcalt, S.I., and Benedict, J.B., 1965. Photointerpretation of two types of rock glacier in the Colorado Front Range, U.S.A. *Journal of Glaciology*, v. 5, p. 849-856.

Papertezian, V.C., 1973. A study of rock glaciers. *B.Sc. thesis*, Queen's University, Kingston, Ontario.

Paterson,W.S.B., 1981. The physics of glaciers (2nd ed.), *Oxford, Pergamon.*

Patton, H.B., 1910. Rock streams of Veta Peak, Colorado. *Bulletin of the Geological Society of America*, v.21, p.663-676.

Pillewizer, W., 1957. Untersuchungen an Blockströmen der Ötztaler Alpen. *Abhandl. Geographischen Institutes der F.U. Berlin*, v.5, p.37-50.

Potter, N., 1972. Ice-cored rock glacier, Galena Creek, Northern Alsaroka Mtns., Wyoming. *Bulletin of the Geological Society of America*, v.83, p. 3025-3058.

Randall, S., 1971. The igneous rocks of the Lyngen Peninsula, Troms, Norway. *Nor. Geol. Unders.*, v. 269, p. 134-146.

Rapp, A., 1959. Avalanche boulder tongues in Lappland, Sweden. *Geografiska Annaler*, v. 41, n. 1, p. 34-48.

Raymond, C.F., 1980. Temperate valley glaciers, in, *Colbeck, S.C., ed.*, Dynamics of snow and ice masses. Academic Press, N.Y.

Reeh, N., Clausen, H.B., Dansgaard, W., Gundestrup, N., Hammer, C.U. and Johnsen, S.J., 1978. Secular trends of accumulation rates at three Greenland stations. *Journal of Glaciology*, v. 20, p. 27-30.

Richmond, G.M., 1962. Quaternary stratigraphy of the La Sal Mountains, Utah. *U.S. Geological Survey Professional Paper*, n. 324.

Rohn, O., 1900. A Reconnaissance of the Chitna River and Skolia Mountains, Alaska. *U.S. Geological Survey 21st Annual Report*, Part 2, p. 399-440.

Rothlisberger, H., 1972. Seismic exploration in cold regions. *Cold Regions Science and Engineering (CREEL) Monograph* II-A, 2a, Hanover, New Hampshire.

Rutten, M.G., 1969. The Geology of Western Europe. *Elsevier Publishing Company, Amsterdam.*

Shakesby, R.A., Dawson, A.G., and Matthews, J.A., 1987. Rock glaciers, protalus ramparts and related phenomena, Rondane, Norway: a continuum of large-scale talus-derived landforms. *Boreas.*, v.16, p.305-317.

Shaw, J. and Healy, T.R., 1977. Rectilinear slope formation in Antarctica. *Annals of the Association of American Geographers*, v. 67, p.46-54.

Shroder, J.F., Jr., 1973. Movement of boulder deposits, Table Cliffs Plateau, Utah. *Geological Society of America (Abstr with Programs)*, v. 5, p.511-512.

Shroder, J.F., Jr., 1978. Dendrogeomorphic analysis of mass movemnt of Table Cliffs Plateau, Utah. *Quaternary Research*, v.9. p.168-185.

Shroder, J.F., Jr., 1987. Rock glaciers and slope failures: High Plateaus and La Sal Mountains, Colorado Plateau, Utah, U.S.A., in, *Giardino, J.R., Shroder, J.F., Jr., & Vitek, J.D., eds.,* Rock Glaciers. Allen and Unwin.

Shroder, J.F., Jr. and Giardino, J.R., 1987. Analysis of rock glaciers in Utah and Colorado, U.S.A. using dendrogeomorphological techniques, in, *Giardino, J.R., Shroder, J.F., Jr., & Vitek, J.D., eds.,* Rock Glaciers. Allen and Unwin.

Sissons, J.B., 1975. A Fossil Rock Glacier in Westor Ross. *Scott. J. Geol.*, v.II, n.1, p. 83-86.

Sissons, J.B., 1976. A remarkable protalus rampart complex in Wester Ross. *Scottish Geographical Magazine*, v.92, p.182-190.

Sissons, J.B., 1977. The Loch Lomond Readvance in the northern mainland of Scotland, in, *Gray, J.M. & Lowe, J.J., eds.,* Studies in the Scottish Lateglacial environment. Pergamon Press, Oxford.

Sissons, J.B., 1979. The Loch Lomond Advance in the Cairngorm Mountains. *Scott. Geogr. Mag.*, v. 95, p. 66-82.

Sissons, J.B., 1980a. The Loch Lomond Advance in the Lake District, northern England. *Trans. R. Soc. Edinb., Earth Sci.*, v.71, p.13-27.

Sissons, J.B., 1980b. Palaeoclimatic inferences from Loch Lomond Advance glaciers, in, *Lowe J.J., Gray, J.M. & Robinson, J.E., eds.*, Studies in the Lateglacial of N.W. Europe. Pergamon Press, Oxford.

Smith, H.T.U., 1973. Photogeologic study of periglacial talus glaciers in Northwestern Canada. *Geografiska Annaler*, v. 55A, p. 69-84.

Souchez R.A. and De Groote, J.M., 1985. δD - $\delta^{18}O$ Relationships in ice formed by subglacial freezing: palaeoclimatic implications. *Journal of Glaciology*, v. 31, n. 109, p. 229-232.

Souchez R.A. and Jouzel, A., 1984. On the isotpic composition of δD and $\delta^{18}O$ of water and ice during freezing. *Journal of Glaciology*, v. 30, n.106, p. 369-372.

Spencer, A.C., 1900. A peculiar form of talus. *Science*, abs., v. 11, p. 188.

Statham, I., 1976. A scree slope rockfall model. *Earth Surface Processes*, v.1, n.1, p. 43-62.

Statham, I., 1977. Angle of repose, angles of shearing resistance and angles of talus slopes; a reply. *Earth Surface Processes*, v.2, n.4, p.437-440.

Stichler, W., 1987. Snowcover and Snowmelt Processes studied by means of Environmental Isotopes, in, *Jones, H.G., & Orville-Thomas W.J., eds.*, Seasonal Snowcovers: Physics, Chemistry, Hydrology. NATO ASI Series, v.211, Reidel Publishing Company.

Sugden, D.E., 1968. The Selectivity of Glacial Erosion in the Cairngorm Mountains, Scotland. *Inst. of Brit. Geogr. Trans.*, v.45, p.79-92.

Swett, K., Hambrey, M.J. and Johnson, D.M., 1978. Lobate rock glaciers and related talus terraces in northern Spitsbergen. *Geological Society of America (Abstr. with Programs)*, v. 10, n. 7, p. 502.

Terzaghi, K., 1952. Permafrost. *Boston Soc. Civil Engineers Journal*, v.39, p. 1-50.

Thompson, W.F., 1962. Preliminary notes on the nature and distribution of rock glaciers relative to true glaciers and other effects of climate on the ground in North America. *Int. Assoc. Sci. Hydrol. Publ.*, n. 58, Symp. Obergurgl, p. 160-165.

Tyrrell, J.B., 1910. Rock glaciers or 'Chrystocrenes'. *Journal of Geology*, v.18, p.548.

Vere, D.M. and Matthews, J.A., 1985. Rock glacier formation from a lateral moraine at Bukkeholsbreen, Jotunheimen, Norway: a sedimentological approach. *Zeutschrift fur Geomorphologie*, v. 29, n. 4, p. 397-415.

Vick, S.G., 1981. Morphology and the role of landsliding in the formation of some rock glaciers in the Mosquito Range, Colorado. *Bulletin of the Geological Society of America*, part 1, v. 92, p. 75-84.

Vick, S.G., 1987. Significance of landslides in rock glacier formation and movement, in, *Giardino, J.R., Shroder, J.F., Jr., & Vitek, J.D., eds.*, Rock Glaciers. Allen and Unwin.

Vitek, J.D. and Giardino, J.R., 1987. Rock glaciers: a review of the knowledge base, in, *Giardino, J.R., Shroder, J.F., Jr., & Vitek, J.D., eds.*, Rock Glaciers. Allen and Unwin.

Wahrhaftig, C. and Cox, A., 1959. Rock glaciers in the Alaska Range. *Bulletin of the Geological Society of America*, v. 70, p. 383- 436.

Wahrhaftig, C., 1987. Foreward, in, *Giardino, J.R., Shroder, J.F., Jr., & Vitek, J.D., eds.*, Rock Glaciers. Allen and Unwin.

Wallén, C.C., 1977. Climates of central and southern Europe, (World Survey of Climatology, v.6). *Elsevier Scientifc Publishing Company, Amsterdam.*

Washburn, A.L., 1979. Geocryology: a survey of periglacial processes and environments. *Edward Arnold, London.*

Wayne, W.J., 1981. Ice segregation as an origin for lenses of non-glacial ice in 'ice-cemented' rock glaciers. *Journal of Glaciology*, v. 27, n.97, p. 506-510.

Wells, N. and Smidt, R.E., 1978. Methods for mineral and element analysis. *New Zealand Soil Bureau Scientific Report*, 10D, Dept. of Scientific and Industrial Research.

Whalley, W.B., 1974. Rock glaciers and their formation (as part of a glacier debris-transport system). *Geographical papers, n. 24, Dept. of Geography, University of Reading, England*, 64p.

Whalley, W.B., 1983. Rock glaciers - permafrost features or glacial relics? *Procs. of the Fourth Int. Conf. on Permafrost*, Fairbanks, Alaska, p. 1396-1401.

Whalley, W.B., Douglas, G.R. and Jonsson, A., 1983. The magnitude and frequency of large rockslides in Iceland in the postglacial. *Geogr. Annlr.*, v.65A, n.1-2, p. 99-110.

White, P.G., 1979. Rock glacier morphometry, San Juan Mountains, Colorado. *Bulletin of the Geologocal Society of America*, v. 90, p. 1515-1518.

White, S.E., 1971. Rock Glacier Studies in the Colorado Front Range 1961-1968. *Arctic and Alpine Research*, v. 3, n. 1, p. 43-64.

White, S.E., 1976. Rock Glaciers and Block Fields: Review and new data. *Quaternary Research*, v. 6, p. 77-97.

White, S.E., 1981. Alpine mass movement forms (non catastrophic): classification, description and significance. *Arctic and Alpine Research*, v. 13, n. 2, p. 127-137.

White, S.E., 1987. Differential movement across transverse ridges on Arapaho rock glacier, Colorado Front Range, U.S.A., in, *Giardino, J.R., Shroder, J.F., Jr., & Vitek, J.D., eds.*, Rock Glaciers. Allen and Unwin.

Whittow, J.B., 1977. Geology and Scenery in Scotland. *Penguin Books.*

Williams, P.J., 1968. Ice distribution in permafrost profiles. *Canadian Journal of Earth Sciences*, v.5, p. 1381-1386.

Williams, P.J. and Smith, M.W., 1989. The Frozen Earth. *Cambridge University Press.*

Wilhelmy, H., 1958. Klimamorphologie der Massengesteine. *Georg Westermann Verlag, Braunschweig, Germany.*

Yarnal, B.M., 1982. Surface sediment analysis of a rock glacier. *Journal of the International Association for Mathematical Geolologists*, v. 14, n. 4, p.371-386.