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**DEVELOPMENT OF SOLIFLUCTION LOBES, KLUANE RANGE,
YUKON TERRITORY**

By

Christophe Kinnard

A Thesis Submitted to the School of Graduate Studies in Partial Fulfilment of the
Requirements for the Degree of Master of Science.

Department of Geography

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ABSTRACT

Solifluction lobe process and morphology were studied on an alpine slope (1800 m a.s.l.) above Kluane Lake, south-western Yukon Territory. Contemporary rates of surface movements, measured by theodolite survey, were found to be greatest in the first two weeks of spring thaw, and movements were consistently faster on lobe treads than on lobe risers. Precise monitoring of thaw-consolidation parameters on a lobe indicated that most thaw-settlement occurred when the soil was saturated to the surface. At the slope scale, surface rates increased downslope in response to gradients in soil moisture, while long-term rates of lobe advance, inferred from ^{14}C dating of buried organic horizons, were found to be similar among 12 dated lobes. The internal stratigraphy and age distribution of these lobes together suggest periodic advance of the fronts. Observations of lobe stratigraphy with the age distribution of buried humus points towards a conceptual model of lobe development involving the slow accumulation of soliflucted material behind a rigid riser, the progressive steepening of the riser and build-up of stress, and finally the rupture of the front and its extensive collapse on to the slope. The period for this developmental cycle was estimated to be about 600 years. This internal cycle of lobe development constitutes a serious buffer to climatic influence and must be taken into account when using solifluction lobes as paleoclimate sources.

RÉSUMÉ

La morphologie et les processus associés au mouvement des lobes de solifluxion ont été étudiés sur une pente située en milieu alpin (1800 m a.n.m.), au-dessus du Lac Kluane dans le sud-ouest du Territoire du Yukon. Les taux de déplacement en surface, mesurés avec un théodolite, indiquent que la plupart du mouvement prend place durant les deux premières semaines de dégel printanier. Le suivi continué sur un lobe des paramètres de consolidation du sol lors du dégel indique que la plupart de l'affaissement du sol a eu lieu lorsque le sol était saturé jusqu'à la surface. À l'échelle de la pente, les taux de déplacement en surface augmentent du haut vers le bas de la pente en réponse à une augmentation de l'humidité du sol, tandis que les taux à long terme d'avancée des fronts, calculés à partir de la datation au ^{14}C d'horizons organiques enfouis sous les lobes, apparaissent similaires entre 12 lobes datés. La stratigraphie interne et la distribution d'âges des lobes suggèrent une avancée périodique du front des lobes. Les observations stratigraphiques et la distribution des âges ^{14}C ont mené à l'élaboration d'un modèle conceptuel de développement des lobes où: (1) le matériel s'accumule lentement par solifluxion derrière un front comparativement plus rigide, (2) le stress se concentre au front qui se raidit progressivement, et, finalement, (3) le front se brise et s'affaisse rapidement sur la pente voisine. La période associée au développement cyclique des lobes a été estimée à environ 600 ans. Ce cycle de développement interne, qui est indépendant du climat, constitue un filtre à l'effet des changements climatiques et doit être pris en compte si l'on veut utiliser les lobes de solifluxion comme sources paléoclimatiques.

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1. INTRODUCTION

Solifluction is a geomorphic process that involves the slow flowing from higher to lower ground of masses of waste saturated with water (Andersson, 1906). It is one of the most widespread forms of soil movement in periglacial regions and occurs ubiquitously on slopes in permafrost areas and in regions with deep seasonal frost (Harris, 1981; Lewkowicz, 1988; French, 1996). Because of its importance to the evolution of slopes in cold climates, solifluction has been extensively investigated as a process (e.g. Benedict, 1970; Harris, 1972, 1977; Egginton and French, 1985; Matsuoka, 1998a, b) as have the landforms (lobes, sheets and terraces) that it produces (e.g. Price, 1974; French, 1974; Smith, 1987c; Matsuoka, 2001; Hugenholtz and Lewkowicz, 2002). However, despite recent insights gained from process-based experimental studies (Coutard et al., 1988; Calmels and Coutard, 2000; Harris et al., 1993, 1997, 2000, 2001), a clear explanation of the mechanism of active layer deformation during ground thaw is still lacking. Moreover, the mode of formation and the evolution of solifluction landforms in relation to their controlling variables, particularly climate, are still unclear.

Despite the fact that no quantitative model is currently able to predict slow mass movement induced by the combination of frost creep and gelifluction, it has been proposed that the resulting landforms, such as lobes and sheets, react to changes in climate and can therefore be used as sources of paleoclimatic information (e.g. Benedict, 1966; Alexander and Price, 1980; Gamper, 1983). Recent climatic models predict a global warming of the atmosphere, which will be two to three times greater in the Arctic and Subarctic than in more temperate regions (IPCC, 2001). The effects of a warming

climate on periglacial slope movement such as solifluction have, for the most part, been a matter of speculation. There is a need for a better understanding of the relationships between climate change and the evolution of solifluction landforms, in order to: (1) use these landforms as paleoclimatic indicators, and (2) assess their potential as geotechnical hazards (Harris et al., 2001b), since continuous creep deformation of a soil mass may lead to rapid failure and more catastrophic events (Van Ash and Van Genuchten, 1990; Harris and Lewkowicz, 2000). The general aim of this research is to investigate the relationships between environmental conditions and the evolution of solifluction lobes on an alpine slope of the Kluane Range, Yukon Territory.

2. LITERATURE REVIEW AND OBJECTIVES

2.1. SOLIFLUCTION: TERMINOLOGY

Slow mass movements on slopes in periglacial regions are often referred to as solifluction, following Anderson's (1906) original definition, "the slow flowing from higher to lower ground of masses of waste saturated with water". However, a considerable degree of confusion has resulted from the application of this general term to a range of processes (Harris, 1981). The term solifluction, as defined by Anderson (1906), does not refer to cold climate and the words "congelifluction" (Dylik, 1951, in Harris, 1981) and "gelifluction" (Washburn, 1967) have been proposed to refer to "earth flow occurring under conditions of frozen ground". Since then, the term "gelifluction" has been used by some researchers to define the slow saturated flow of soil associated with soil freezing, (Washburn, 1980; Strömquist, 1985; Gengnian et al., 1995; Kirkby, 1995;), while the term solifluction has been retained by others (Williams, 1966; Benedict, 1970; Harris, 1972, 1977, 1987; Price, 1973, 1991; Smith, 1992; Lewkowicz and Clarke, 1998).

Many studies have shown that slow mass movements in cold regions include several processes and do not necessarily require saturation. Hence, in modern usage, solifluction is used as a collective term to represent slow mass-wasting associated with freeze-thaw action (Matsuoka, 2001). Two separate processes, frost creep and gelifluction, act simultaneously to produce solifluction movement. Frost creep is the net downslope displacement of a soil layer following frost heave normal to the slope and subsequent near-vertical thaw settlement (Washburn, 1967). Gelifluction is the downslope flow of

saturated soil associated with thawing (e.g. Washburn, 1980; French, 1996). Defined in this manner, solifluction refers to the slow movement of soil occurring under cold climates where freeze-thaw action prevails, irrespective of the presence or absence of permafrost. This thesis employs this broad definition.

2.2. MECHANISMS OF SOLIFLUCTION

The precise nature of the deformation involved in solifluction is still a matter of debate. Solifluction occurs on slopes when a thawing soil mass consolidates under its own weight. When ice segregation occurs in a freezing soil, water is drawn to the freezing front because of differences in potential (Williams and Smith, 1989). Water in excess of the soil porosity freezes into ice lenses, which cause frost heaving of the soil surface. Following the melt of excess ice during spring thaw, water liberated in the thawing part of the soil may increase porewater pressures and cause a reduction in the frictional strength of the soil (Carson and Kirkby, 1972; Harris, 1981). As the effective stress on the consolidating soil is reduced, the soil mass becomes deformable and downslope movement in the form of gelifluction accompanies the vertical settlement of the soil. The *thaw consolidation theory*, introduced by Morgenstern and Nixon (1971), predicts an increase in porewater pressure when the rate of thawing is relatively rapid compared to the capacity of the soil to expel the excess water. This allows soils in periglacial regions to become unstable even on slopes with very low gradients (McRoberts and Morgenstern, 1974; McRoberts, 1978). Conventional stability analyses undertaken on solifluction slopes suggest, however, that movement takes place at stresses less than the peak strength of the soil (Williams, 1959, 1966; Harris 1977), and may therefore be regarded as a pre-

failure strain in the thawing soil (Harris et al., 1995, 1997, 2001). Thawing of ice lenses and/or snowmelt causes high moisture contents in the soil and movement often occurs at moisture contents close to or above the liquid limit of the soil (Washburn, 1967; Harris, 1972; Smith, 1988). Harris et al. (1995) concluded that soil displacement during solifluction represents post-thaw strain when soil moisture contents are high and rheological properties are intermediate between those of a viscous fluid and a plastic solid. Using centrifuge modelling of solifluction movements, Harris et al. (2000) concluded that viscous-flow processes were not dominant, and that the observed displacement of the soil was the result of time-independent frictional flow rather than time-dependent viscosity. Their conclusion was based on the fact that movements in the centrifuge model scaled with movements observed in a previous full-scale model, in regards to the time available for thaw-consolidation, whereas viscous flow should not scale appropriately.

2.3. CONTROLLING VARIABLES

Rates and processes of solifluction depend, collectively, on topography, hydrology, geology, vegetation and climate (e.g. Washburn, 1967; Harris, 1981; Lewkowicz, 1988; French, 1996; Matsuoka, 2001).

2.3.1. Slope gradient

Slope gradient exerts a topographic control on solifluction rates, as the downslope shearing force due to the soil's own weight is proportional to the sine of the gradient (Washburn, 1980). However, the relationship between measured rates and gradient is indirect, probably because it is obscured by other more important variables such as soil

moisture, freeze-thaw frequency and texture (Benedict, 1970; Harris, 1981; Matsuoka, 2001). French (1996, p.152) suggests that slope angle might be of little importance in influencing movement unless the difference between slopes is at least 10°.

2.3.2. Soil moisture

The amount and distribution of moisture on a slope is critical in determining solifluction rates. Authors such as Benedict (1970), Washburn (1967, 1999), Harris (1972, 1973, 1974), Gamper (1983), Egginton and French (1985), Smith (1988), Kirkby (1995) and Matsumoto et al. (2001) all stress that the soil moisture content, which often exceeds the liquid limit during ground thaw, is the critical controlling variable. Moisture conditions at a particular site also dictate the relative importance of frost creep and gelifluction. High moisture availability during freeze-back of the active layer in the fall promotes frost heave and a higher potential for frost creep, while the greater ice content in the soil results in increased gelifluction during spring thaw. In a full-scale experiment on solifluction, Harris et al. (1993) showed that rates of surface movement increase proportionally with the amount of frost heave. In the Colorado Front Range, Benedict (1970) found a good correspondence between high annual surface movement on a solifluction lobe, increased frost heave and high groundwater levels at the beginning of freeze-back. Field studies by Harris (1977), Matsumoto et al. (2001) and an experiment by Harris et al. (1997) suggest that thaw consolidation of the soil may induce an upward hydraulic gradient which would prevent the entry of water from external sources. These results point to the fact that high frost heave ratios, resulting from good moisture availability during autumn freeze-back, appear to be the dominant cause of increased solifluction activity. In this respect permafrost, by preventing free drainage during

ground thaw, is likely to contribute to the saturation of the soil when freeze-back occurs (Jahn, 1978; Harris, 1987). Hence, it appears that the depth and thickness of ice lenses and freeze-thaw frequency are the major control on spatial and temporal variations in solifluction rates (Matsuoka, 2001).

2.3.3. Grain-size of soil

Soil texture is the main geological variable controlling the process of solifluction. Fine-grained soils with a high silt fraction are frost-susceptible (Beskow, 1935, in Harris, 1981; Linell and Kaplar, 1959) and hence, prone to frost heave and solifluction. Harris et al. (1993) demonstrated experimentally that solifluction rates are strongly related to the silt and clay content of the soil ($< 63 \mu\text{m}$), as well as the frost heave ratio. In the Japanese Alps, Matsuoka (1998a) measured the deformation of painted lines and found that displacement of the surface debris occurred mostly in zones covered by gravely sandy loam. In his empirical model of frost creep rates, the silt and clay content of the soil increased the correlation coefficient significantly (Matsuoka, 1998b). However, Benedict (1970) found no relationship between texture and rates of mass movement in the Front Range of Colorado because, as he states, other factors such as soil moisture conditions over-ride the effect of granulometry. Washburn (1967) reached the same conclusion in his study at Mesters Vig, Greenland.

2.3.4. Vegetation

Vegetation may have a restraining effect on solifluction processes. Studies by Gamper (1983) and Matsuoka (1998a, b) have shown restricted surface movement in better-vegetated parts of slopes. The cohesion of the organic mat, combined with the anchoring effect of any plant roots, resist internal shearing, while inorganic soil deforms more

easily. Price (1991) emphasized the role played by the organic mat in inducing greater differential movement at depth. 20-year velocity profiles measured on the tread of solifluction lobes in the Ruby Range, Yukon, suggest that the thick organic turf impedes surface movement. Similarly, Benedict (1970) showed that high surface rates of solifluction are restricted to the sparsely vegetated zones on the tread of solifluction lobes.

2.3.5. Climate

Climate influences solifluction rates throughout precipitation (rain and snow) and the thermal regime of the soil (Woo et al., 1992). The climatic influence on solifluction rates is probably complex, because of the buffering effect of inter-dependent microclimatic factors (e.g. Smith and Riseborough, 1983; Williams and Smith, 1989). Such microclimatic factors, which include snow cover thickness and elevation, organic mat thickness, vegetation cover and thermal properties of the soil, may vary at different spatial scales and cause solifluction landforms to react in varying ways to climate change. For example, Gamper (1983) found that in the alpine setting of Switzerland, lower winter temperatures and a shorter period of restricted snow cover promote deeper freezing and enhanced solifluction in the spring. In Svalbard, where the climate is much colder, Åkerman (1996) found good correlations between interannual variability in solifluction rates and summer temperatures. This comparison, between a permafrost and non-permafrost environment, illustrates how different climatic forcing could result in an increase in movement rates. Also of interest is the recent re-survey by Carver et al. (2002) of the surface targets installed in 1967 by A.L. Washburn in Mesters Vig,

Greenland, which showed accelerated rates of surface movement relative to initial results, tentatively interpreted by the authors as being caused by climatic amelioration.

Although the reaction of solifluction landforms to climate change might be site-specific, a deepening of the active layer in permafrost environments, and of the maximum freeze-thaw depth in regions with seasonal frost, are generally believed to cause accelerated rates of solifluction (Benedict, 1976; Kirkby, 1995; Lewkowicz and Clarke, 1998; Matsuoka, 2001). High solifluction rates are often found in permafrost zones with thick active layers, or in regions with deep seasonal freezing (Matsuoka, 2001). Moreover, during the Holocene, long-term cycles of climatic degradation and amelioration, and of permafrost formation and disappearance, could have triggered the formation of some of the solifluction landforms observed in current periglacial environments (Worsley and Harris, 1974; Benedict, 1976; Ellis, 1979; Matsuoka, 2001).

In summary, prediction of the effects of climate change on solifluction rates and landforms requires a better understanding of: (1) the mechanism of solifluction in relation to its controlling variables; and (2) the long-term behaviour of solifluction landforms in relation to spatially-variable microclimatic and other site-specific conditions (vegetation, snow cover, frost regime, moisture conditions and soil texture).

2.3.6. Solifluction landforms

Benedict (1970) and Harris (1981) recognized three basic topographic expressions of solifluction: solifluction sheets, solifluction terraces, and solifluction lobes. Benedict (1970) first subdivided these features into sorted or unsorted sheets, and turf-banked or

stone-banked lobes and terraces. Matsuoka (2001) used the same classification but included stripes and hummocks as solifluction landforms. Turf-banked lobes are reported as having risers up to around 1 m in height, and tread lengths of between 2 m and 50 m. They develop typically on slopes of between 4° and 25° and are common below late-lying snow patches (Harris, 1981). Higher risers, up to 3-5 m, have also been reported for the plateau surrounding the study slope (Hugenholtz and Lewkowicz, 2002), in the nearby Ruby Range (Price, 1991) and in China (Gengnian et al., 1995). Vegetation is usually thickest on the fronts while the central parts of the lobe are often less vegetated. Stone-banked lobes are almost devoid of vegetation and are made of relatively coarser material.

The difference between lobes and sheets is a matter of size, and the division between the two types of landforms can be subjective. While terraces refer to more straight-fronted features, they often break-up into multi-lobate fronts which makes the distinction between terraces and lobes difficult, as one often grades into the other (Harris, 1981). A continuum probably exists between each type of landform. Lobes are more common on alpine slopes where surface conditions (vegetation, material and topography) are variable, while the larger sheets and undifferentiated solifluction (Åkerman, 1996; French, 1996; Lewkowicz and Clarke, 1998) are associated with homogenous surface conditions that dominate on polar slopes.

The initiation and evolution of lobate landforms in time and space is not well understood, but must result from variations in the rate of solifluction downslope (Harris, 1981, 1987).

Solifluction lobes, the smallest of the landform types, develop either in locally favourable conditions, such as below snow patches, or cover entire slopes. Although widely discussed in the literature, little is known with certainty about their behaviour in relation to environmental conditions and climatic forcing.

2.4. CONTEMPORARY RATES OF MOVEMENT

Numerous researchers have reported solifluction rates measured on lobes, terraces, sheets, or on undifferentiated slopes. Downslope displacement of the soil surface by solifluction has been measured mainly by resurvey of surface pebbles, boulders, or markers inserted in the ground. Surveying of targets has been done using a theodolite (e.g. Washburn, 1967, 1999) or a steel tape (e.g. Rapp, 1960; Price 1973) from natural benchmarks like bedrock outcrops and large boulders, or from stakes deeply embedded in the soil (e.g. Benedict, 1970). Modern surface rates of solifluction measured in various environments were compiled by Harris (1981). Movement rates between 0 and 213 mm per year are reported from alpine areas, and from 0 to 124 mm per year for arctic areas. In a more recent review, Matsuoka (2001) reports surface rates ranging between 0 and 1000 mm per year. This larger range is due to the inclusion of sites with shallow superficial movement, where needle-ice creep causes high surface movement but low volumetric transport rates. Solifluction movements show significant differences between sites, and variation is also evident at a local scale (Egginton and French, 1985; Washburn, 1999), especially for solifluction lobes (Price, 1973, 1991; Smith 1987b, 1988). Variability in measured rates of solifluction on the same slope has also been reported and attributed to changes in site-specific factors (e.g. Washburn, 1967, 1999).

The distribution of soil movement with depth is usually obtained by inserting columns of wooden cylinders (Rudberg, 1964; Benedict, 1970; Harris, 1972; Mackay, 1981), lengths of flexible plastic tubing (Williams, 1966, Price, 1970; Smith, 1992) or tinfoil strips (French, 1974) in the moving soil. These markers are re-excavated after a period of time to obtain the depth-integrated movement of the soil. Strain probes inserted in plastic tubing previously buried in the ground have also been used as a non-destructive method and allow for repetitive measurements of velocity profiles (Williams, 1966; Price, 1970; Smith, 1987b). More recently, specifically designed strain probes buried in the ground and connected to a data-logger were used to obtain high frequency measurements of a velocity profile (Matsuoka, 1994, 1997, 2000; Sato et al., 1997; Yamada, 1997; Yamada et al., 2000; Matsumoto and Ishikawa, 2002). Continuously recording solifluction movements have also been made with an electro-mechanical solifluction meter (Lewkowicz, 1992; Lewkowicz and Clarke, 1998).

In areas of seasonal freezing and of “warm permafrost”, velocity profiles are generally concave downslope (e.g. Williams, 1966; Price 1973; Jahn, 1978), although the shape of the observed profile can vary considerably, even at the slope scale, depending on the concentration of ice lenses (Harris, 1981) and the restraining effect of vegetation (Price, 1991). In areas of two sided-freezing, or “cold permafrost”, velocity profiles are predominantly convex downslope. This is due to concentrated shearing near the ice-rich permafrost table which results in a plug-like flow of the soil mass (Mackay, 1981; Egginton and French, 1985; Lewkowicz and Clarke, 1998). In Svalbard, Matsuoka and Hirakawa (2000) observed both types of movement at sites in the same area. The

presence of basal shearing and resulting plug-like flow at one site was explained by the fact that the whole active layer consisted of muddy sediment and had a restricted snow cover in winter. The lower air temperature caused upward freezing of the active-layer and plug-like flow on thawing. At the other site, coarser material in the lower part of the active layer rejected ice lensing, and movement at depth was restricted.

2.5. LONG-TERM RATES OF ADVANCE OF SOLIFLUCTION LANDFORMS

Long-term rates of solifluction have been obtained using radiocarbon dating of organic material buried by the advance of solifluction lobes and sheets. Sampling for buried humus at regular intervals along excavated organic horizons allows for the calculation of past rates of lobe movement. Long-term rates of lobe advance, as measured by radiocarbon dating of buried organic horizons, range from 1 to 30 mm per year (Matsuoka, 2001). The discrepancy between the measured rates of frontal advance (long-term) and tread movements (contemporary) confirms that a solifluction lobe results from local accumulation at the front of material moving from upslope.

Variations in long-term rates of lobe advance have frequently been interpreted as resulting from changes in regional climate (e.g. Benedict, 1966; Costin et al., 1967; Alexander and Price, 1980; Gamper, 1983; Reanier and Ugolini, 1983; Mathews et al., 1986; Ballantyne, 1986; Smith, 1987a; Elliott and Worsley, 1999). In addition, dating of the upper ends of organic horizons buried under solifluction lobes has been used to infer the onset of active solifluction periods (Worsley and Harris, 1974; Ellis, 1979; Nesje, 1989). Regional chronologies of solifluction activity during the Holocene have been

developed based on compilations of these results (Morin and Payette, 1988; Matthews et al., 1993; Gamper, 1993). Despite considerable regional variability, this technique suggests a peak of activity around 3000-2500 years BP for North America and Europe, with 7000 BP being the oldest date yet measured in the field, in Alaska (Reanier and Ugolini, 1983).

Conclusions reached on the response of solifluction landforms to climate change have, for the most part, been tentative because of the limited number of dates in each study, the uncertainty associated with radiocarbon dating of soils and the unknown representativeness of a particular landform for a given site. Dates obtained from solifluction lobes show that lobe advance rates have varied throughout the Holocene, but it has yet to be demonstrated that these variations represent a distinct climatic signal. For example, Dyke (1981) dated the past rates of advance for four solifluction lobes in the Canadian Arctic and found that rates varied among the lobes, probably because of differences in soil texture, slope angle, morphology and moisture availability. Moreover, changes in solifluction rates among the lobes were found to be non-concurrent in time. This led Dyke to hypothesize that these changes reflected site-specific influences such as drainage conditions instead of a climatic effect, but he suggested that a bigger sample could reveal a different trend. On the other hand, Nesje (1989) showed that the dates for the onset of solifluction were statistically indistinguishable for four solifluction lobes located in the same area, even though the lobes had different lengths, widths, gradients and aspects. He suggested that deterioration in regional climate was probably responsible for the synchronous initiation of the lobes.

Radiocarbon dating of buried soils is beset with many difficulties, which relate to the limited understanding we have of the dynamics of organic matter in soils (e.g. Sharpenseel and Schiffmann, 1977; Birkeland, 1999). Previous work on paleosol dating (Matthews and Dresser, 1983; Matthews, 1985; Yong et al., 1996) has shown that:

- (1) accumulation and decomposition of organic matter in soils reach a dynamic equilibrium at some point in time which depends on the surrounding environment;
- (2) the age of the soil, once the dynamic equilibrium is attained, is the time required for fresh organic matter to decompose completely, and is called the mean residence time (MRT). It is a minimum estimate, often far from the absolute age of the soil;
- (3) buried organic layers may display strong age/depth gradients reflecting the slow translocation of organic matter through the soil.

The age of a buried humus layer is usually seen as the sum of the time elapsed since burial and the MRT age, and this must be taken into account when dating solifluction lobes. Many authors have subtracted an apparent mean residence time (AMRT) from the dated points along the humus layer to obtain the age of burial, assuming a constant value over time. The AMRT was either estimated from dating the modern organic layer in front of the lobes, (Benedict, 1966; Alexander and Price, 1980) or derived by extrapolating the rate of lobe advance between two dated samples close to the lobe front (Dyke, 1981; Smith, 1987a). This latter technique has yielded contrasting results, with AMRT ages varying significantly between adjacent lobes. Radiocarbon dates obtained from modern soils have been shown to display some consistency among sites across Canada and Scandinavia, ranging between 140 and 465 ^{14}C BP (Alexander and Price,

1980), but a study on Baffin Island (Stuckenrath et al., 1979) gave ages ranging between 40 and 915 years.

2.5.1. Synthesis

A large number of studies have been conducted on solifluction processes and resulting landforms. Most of them, however, have been fairly descriptive in nature. With growing concern about the effect of climate change on northern environments, there is a renewed interest in slope stability, and recent studies have shown significant progress in understanding solifluction processes. Our knowledge of slow mass movement has greatly improved with recent experimental work done on solifluction, as well as with the application in the field of modern techniques of slope monitoring. It now appears possible to link the considerable number of observations made in the past by researchers working in a variety of field environments, to these recent results. In the context of climate change, the use of solifluction landforms as sources of paleoenvironmental information requires a good understanding of solifluction mechanics and a better assessment of the environmental controls on landform development.

2.6. RESEARCH QUESTIONS AND OBJECTIVES

A number of questions have arisen from the synthesis outlined above. These have been taken as guidelines throughout the research, and as the basis for the objectives of the present study. Research questions and objectives are summarized in point form below.

2.6.1. Research questions

- How do solifluction lobes initiate and develop through time?

- What is the spatial variability in the rate of advance of solifluction lobes at the slope scale, and what are the controlling factors?
- Can a climatic signal be inferred from changes in long-term rates of frontal advance, despite the potential for small-scale variations in site-specific factors ?
- Do solifluction lobes on the same slope react in the same manner to climatic forcing, allowing these landforms to be used as repositories of past climate history?
- Is it possible to model the evolution of solifluction lobes?

2.6.2. Research objectives

2.6.2.1. Primary

- To examine the spatial variability in short and long-term rates of solifluction lobes on the same slope;
- To gain knowledge on the spatial distribution of solifluction lobes and related morphological parameters, at the slope scale;
- To understand the mechanics of solifluction deformation during ground thaw, by continuously monitoring thaw consolidation processes in a solifluction lobe.

2.6.2.2. Secondary

- To propose a simple model of solifluction lobe development;
- To propose a simple model of slope evolution for the study site. The model will have to explain the spatial distribution of solifluction lobe ages, rates and morphology.

3. STUDY AREA

3.1. LOCATION

The Kluane Range, with peaks 2000-3000 m in altitude, is located in the southwestern Yukon on the northeast flank of the St. Elias Mountains, the largest group of high peaks in Cordilleran North America. The research site is situated above Kluane Lake (61°01', 138°24'), on an alpine plateau west of Outpost Mountain (alt. 2100 a.s.l.) (Figures 3.1 and 3.2a). Solifluction lobes, terraces and sheets are found in abundance on the gently sloping flanks of the plateau (Figure 3.2b, from Hugenholtz, 2002).

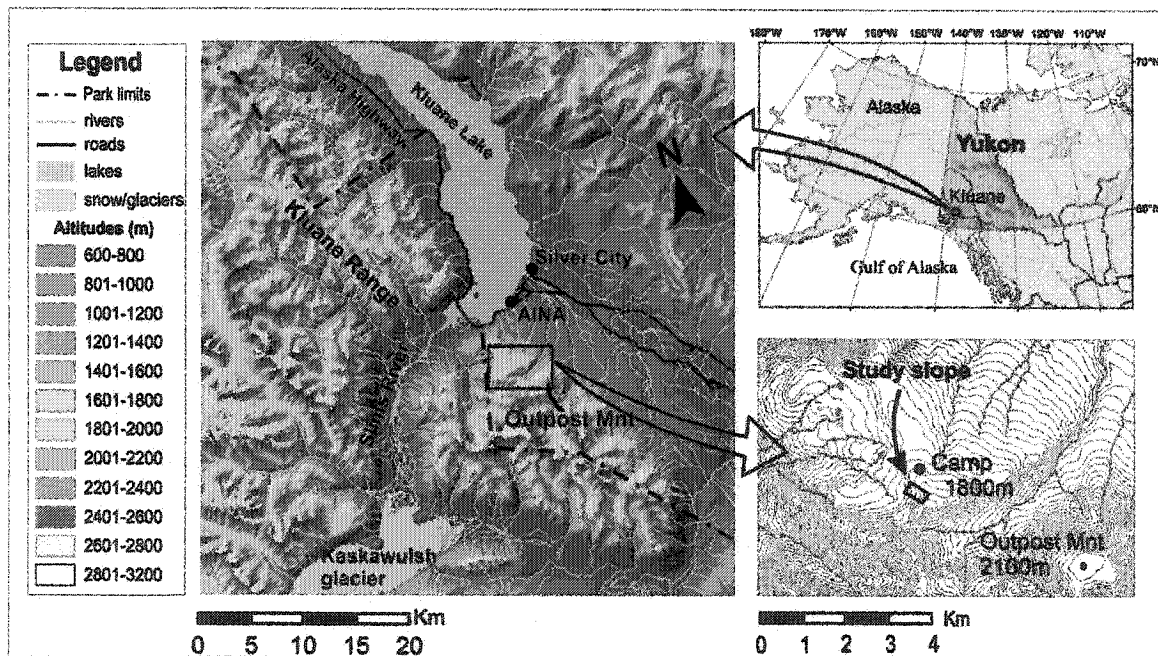


Figure 3.1: Map of study area and research site.

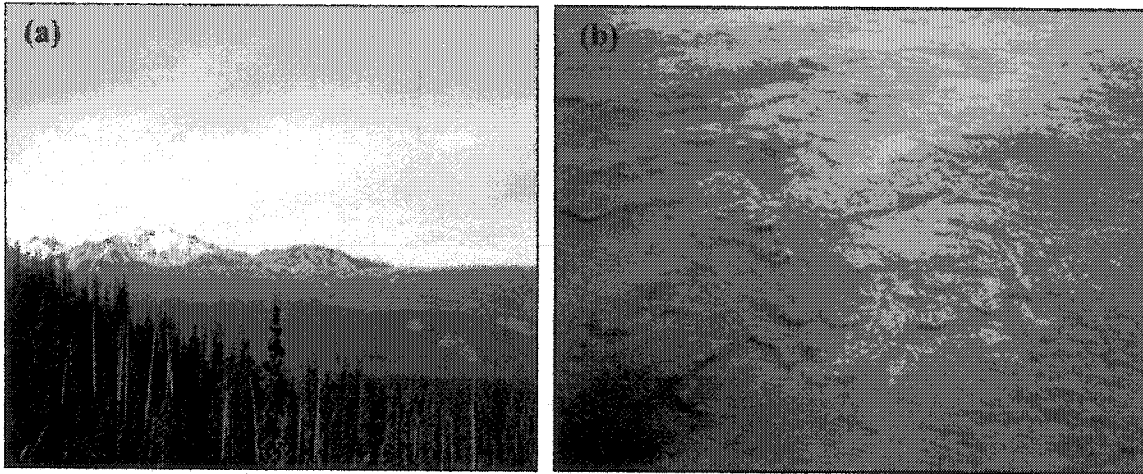


Figure 3.2: (a) View of the plateau from the north; (b) View of the North Slope from the air.

3.2. GEOLOGY AND QUATERNARY HISTORY

The physiography of the Kluane Lake region is dominated by the Shawkak Trench, a major lineament crossing the Southern Yukon Territory and forming a sharp boundary between the St. Elias Mountains and subdivisions of the Yukon Plateau to the north. The broad Shawkak Trench, whose floor is at about 700 m a.s.l., is occupied by Kluane Lake, which is fed by the Slims River flowing from the Kaskawulsh glacier 20 km to the southwest (Figure 3.1). The Ruby Range, which borders the lake to the northeast, is an exposed glaciated landscape with rounded peaks whereas the Kluane Range to the southwest constitutes a landscape strongly affected by mass-wasting. Mountains in the Kluane Range have steep sides and tend to be covered by active screes. The advance of outlet glaciers from the main icefields to the south along valleys, such as that occupied by the Slims River, have produced oversteepened slopes and deep dissection (Harris and Gustafson, 1993). The geology of the southern side of the Shawkak Trench is a mixture

of Palaeozoic slates, greywackes, limestones, dolomite, conglomerates, serpentine and greenstone (Wheeler, 1963).

The glacial history of the region has been described by Denton and Stuiver (1966, 1967). They recognized four glaciations separated by three nonglacial intervals in the northeastern St. Elias Mountains. Ice from the late Wisconsinan advance of the Kluane Glaciation entered the Shawkak Trench through valleys draining the Icefield Ranges, flowed northeast along the Trench with a thickness exceeding locally 1075 m, and terminated in the vicinity of Snag near the Yukon-Alaska boundary. In the Shawkak Valley, bedrock exposed below 1860 m has been glacially sculptured into rounded knobs, streamlined forms and blunted spurs. The alpine plateau where research was conducted is described by Denton and Stuiver (1967) as "a valley shoulder indicating particular abrasion". Ice of the Kluane Glaciation had receded to the vicinity of the mouth of Slims River valley shortly before 12,500 BP, and by 9780 BP had withdrawn behind the position presently occupied by neoglacial moraines fronting the Kaskawulsh glacier (Denton and Stuiver, 1966).

The drift deposited by the Kluane Glaciation mantles the floor of the Shawkak Trench to a thickness exceeding 90 m but thins rapidly with altitude and steepness on valley walls. The alpine plateau adjacent to Outpost Mountain is covered by glacial till up to about 1860 m a.s.l., which marks the upper limit of the Kluane Glaciation in this area. Above 1860 m an alpine morphology persists, characterized by sharp ridges, pyramidal peaks, and colluvial mantles (Figure 3.3). Following a nonglacial period (Hypsithermal) from



Figure 3.3: View from the study site of the topography above 1900 m, with tors seen in the foreground.

9780 BP to 2640 BP (Slims Nonglacial Interval), the Kaskawulsh glacier re-advanced over at least 24 km to a maximum position around 1865-1874 AD, from where it retreated 2 km back to its present position (Denton and Stuiver, 1966).

3.3. CLIMATE

The climate in the Kluane Lake area is continental with cold winters and warm summers. The mean annual air temperature (MAAT) at the Arctic Institute of North America between 1972-1980 (AINA, see Figure 3.1), at 781 m a.s.l., was -2.7°C (Environment Canada, 1982, in Harris and Gustafson, 1993).

Precipitation at AINA (468 mm a^{-1}) is primarily as rain in the summer (late May-August) and as limited snowfalls in fall or early winter (Harris and Gustafson, 1993). It is likely that precipitation is greater at the study site, located 1000 m higher than AINA. A

weather station was maintained at the research site at 1800 m a.s.l., from August 2001 to August 2002, and air temperature was recorded automatically at hourly intervals (Figure 3.4). The MAAT for that one-year period was -5.2°C . Using a theoretical lapse rate of 6°C per 1000 m, and extrapolating the MAAT from AINA (-2.7°C) to the study site at 1800 m, gives a MAAT around -9°C , a value much lower than that recorded in 2002. The mean value of -5°C may represent an unusually warm year, a much lower lapse rate, or climate warming since the 1970s. The first possibility is supported by the presence of a prolonged warm period in January 2002 (Figure 3.4), which does not appear in average values for the 1972-1980 record at Kluane Lake. It is also possible that the mean value of -2.7°C recorded at AINA may be relatively low due to the tempering effect of Kluane Lake in summer or because of temperature inversions in the valley bottom. In any case, the theoretical value of -9°C appears low for the study site and a value of -5° is probably closer to reality.

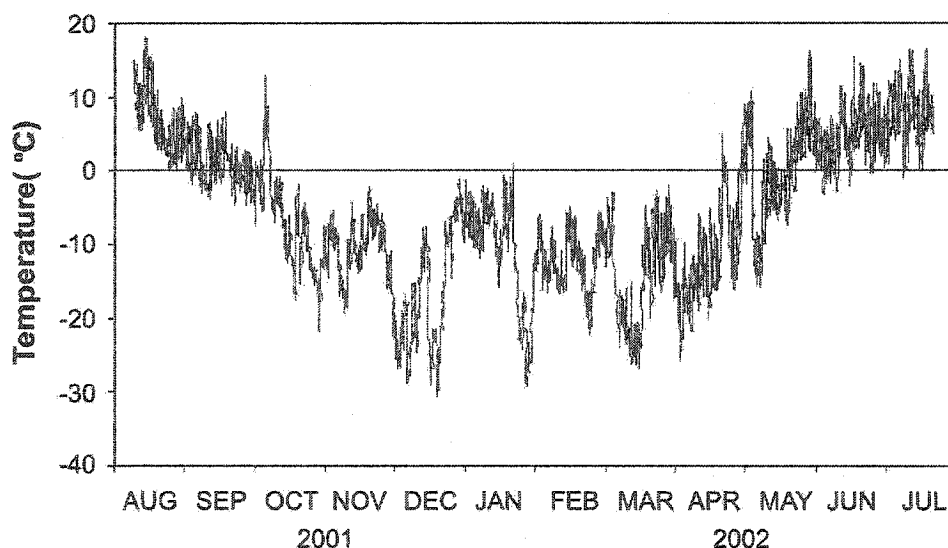


Figure 3.4: Hourly temperatures recorded at study site, 1760 m a.s.l. Note the unusually warm January.

Southern Kluane Lake is located in the zone of Sporadic Discontinuous Permafrost (Heginbottom, 1995). However, perennially frozen ground is present on all north-facing slopes at the elevation of the plateau constituting the study site (Harris, 1987; Harris and Prick, 2000), while pits excavated on south-facing slopes revealed no frozen ground within 2 m of the surface (Hugenholtz, 2002). In August 2001, a deep frost table between 1.5-2 m was encountered everywhere on the west-facing study slope.

3.4. HYDROLOGY

Observations in August 2001 and May-June 2002 on the study slope suggested that the surface hydrology of the area is controlled mainly by snowmelt in the spring, and occasional rainstorms in summer. Snowmelt in the spring of 2002 was rapid, releasing large amounts of meltwater. A shallow active layer at the beginning of snowmelt caused surface flow to occur on most of the study slope and allowed for a rapid transfer of water to the streams. Apart from occasional high-magnitude pluvial events later in the summer (R. Van Brabant, *University of Ottawa, pers. Comm.*), which can re-activate surface flow on the slopes, the active layer progressively dries throughout the summer.

3.5. VEGETATION

Boreal forest with white spruce, aspen and balsam poplar extends up to 1070-1210 m a.s.l. on the slopes of the plateau, while a birch-willow shrub zone continues to 1250 m a.s.l. (Harris and Gustafson, 1993). Above this is a zone of alpine tundra composed of grasses, mosses and small shrubs. It is in the last zone that solifluction landforms are found.

3.6. STUDY SITE

A west-facing slope, with a high density of solifluction lobes, was selected for the study. This slope was part of a broader study by Hugenholtz (2002) and Hugenholtz and Lewkowicz (2002) on the whole mountain plateau. Their work focused on the morphometric interrelationships and related environmental control on solifluction lobes to site exposure.

The profile of the study slope is broadly convexo-concave, and lobes typically increase in size downslope, evolving from small sorted steps near the slope crest to extensive complexes with multiples fronts at the foot (Figures 3.5 and 3.6). The slope is covered by colluviated glacial till and the top of the slope, at 1900 m, forms a crest which joins a small summit covered by colluvial debris that marks the upper limit of the Kluane Glaciation (Figure 3.7). Late summer probing of the thawed soil suggests that permafrost is present, with a 1.5-2 m-deep active layer. When the frost table was not determined directly by probing for resistance, a temperature profile of the soil at depth was obtained using a probe with a thermistor embedded in the tip, and extrapolation of the profile to 0°C permitted calculation of the thaw depth. The slope is covered by tundra vegetation with grasses and small shrubs on the upper slope, and includes mosses on the lower, concave part of the slope. With its high concentration of solifluction lobes, the site is ideal to conduct a detailed study on these landforms. The lobes, which vary greatly in size, cover almost the entire slope. The site thus offers a good opportunity to examine how these landforms function, both as individuals and as a population, in relation to the surrounding environment.

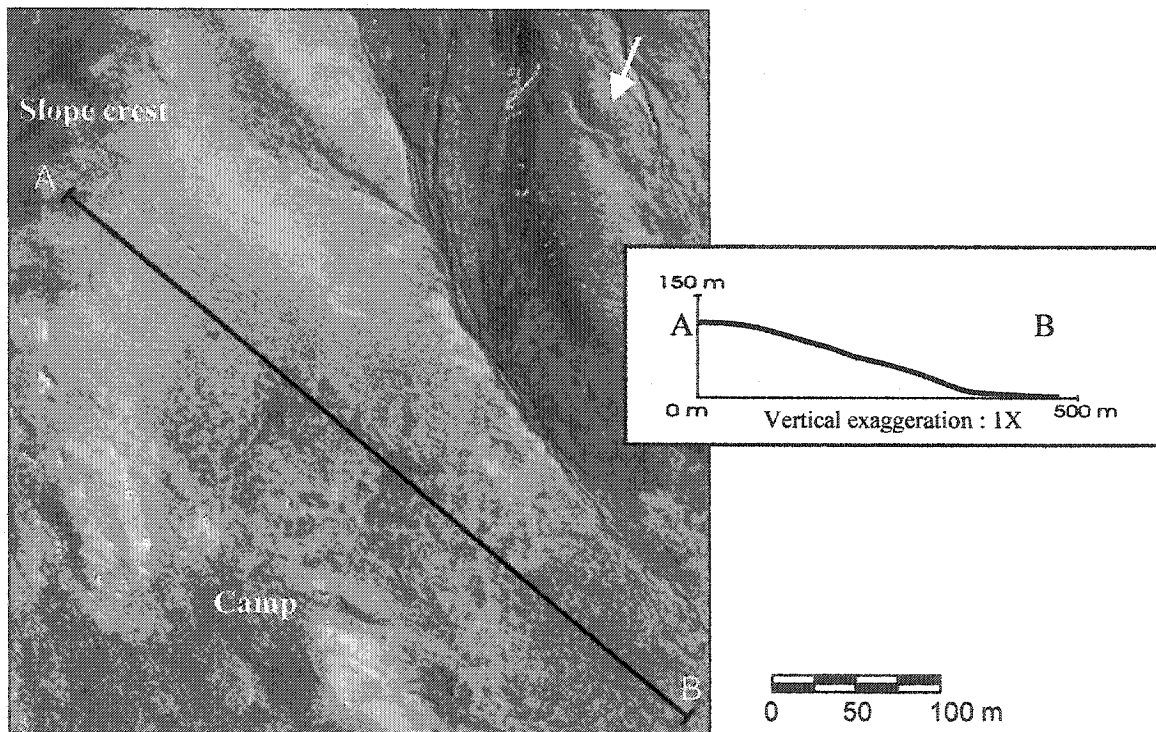


Figure 3.5: Air photo of study site, where solifluction lobes can be seen to extend from the crest to the foot of the slope (reference: National Air Photo Library A23001-68). The slope profile is shown along the line A-B.



Figure 3.6: View of the study slope, looking North. Solifluction lobes can be seen along the slope, the bigger ones being found near the footslope. Points A and B, shown in Figure 3.5, are marked.



Figure 3.7: View of the upper slope and ridge, looking South.

4. METHODS

4.1. MAPPING OF SOLIFLUCTION LOBES

Solifluction lobes on the study slope were mapped in June 2002 using a Nikon electronic digital theodolite with a resolution of 10 seconds. Readings were taken from two separate stations. The first station ("S1") was located upslope and consisted of re-bar inserted 2 meters in the ground and sticking 30 cm above it, with a 0 degree sighting point located downslope at another similar re-bar ("B1"). The second station ("S2") was installed at mid-slope on a stable bedrock outcrop, with 0 degree sighting point a few meters away on the same rock outcrop ("B2"). A *Garmin* GPS with an average accuracy of 6 meters was used to position the different benchmarks used for the study and ensure proper georeferencing of the coordinate data. The tops of two surrounding peaks were used as control points to ensure precise re-positioning of the theodolite before each re-mapping session. Readings were taken from the two stations by sighting on a stadia rod held by an assistant at each lobe location. Apart from the very small lobes and steps near the slope crest, all identified solifluction lobes on the slope were mapped by taking at least one reading at the front of each lobe, and another at each of the edges, or in the inter-lobe area.

A total of 235 lobes were mapped on the slope, and it is believed that almost the entire population of lobes was covered. A few landforms may have been missed because their outline was unclear, and some very small features near the slope crest (< 1 m long) were also omitted. A first version of the map showing the solifluction lobes, with various sampling locations, is presented in Figure 4.1.

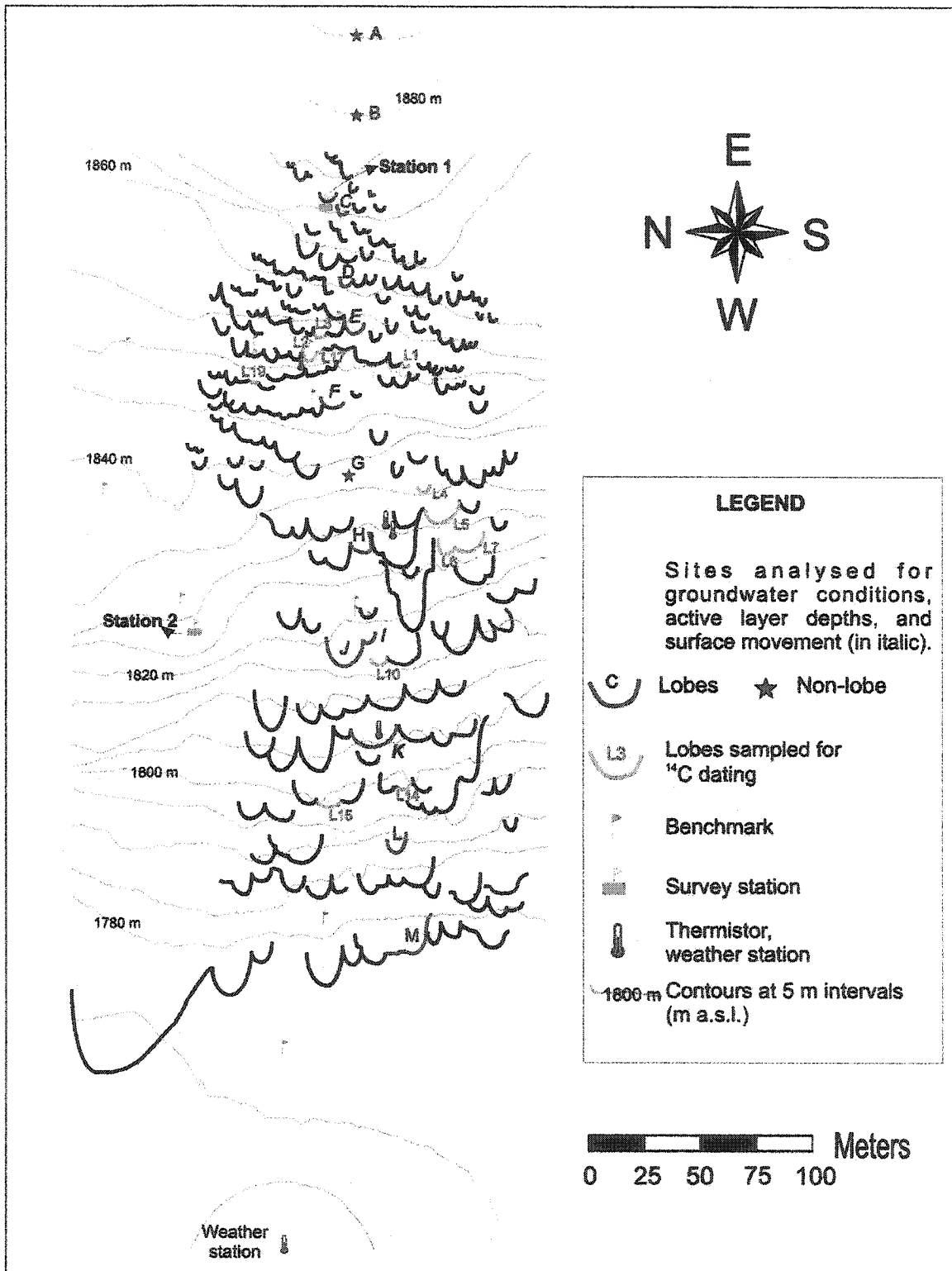


Figure 4.1: Map of the study slope showing the different measurement locations. Note: contours were interpolated at 5 m intervals from survey points and do not account for lobe effects.

Using the spatial coordinates derived from the survey data (see Appendix A), a digital elevation model (DEM) of the study slope was created by interpolating a continuous elevation surface from the point data (space coordinates). The DEM allows for a better grasp of the local configuration of the solifluction slope and can be used to obtain cross-profiles of the slope.

4.2. MORPHOMETRY OF SOLIFLUNCTION LOBES

A number of morphometric measurements were made at each of the mapped lobes. The methods used here broadly follow those used by Hugenholtz (2002), and are similar to those of Hall (1981) and Smith (1987c). The measured morphometric variables are defined below and refer to Figure 4.2.

- *Tread length* is the distance, as measured in meters by a steel tape, between the front of the lobe and its rearmost point.
- *Tread angle* is the angle in degrees obtained from a digital level placed on a 2 m-long stadia rod, positioned flat on the tread surface along the long axis of the lobe.
- *Tread width* is the distance, as measured in meters by a steel tape, between the two edges of the lobe.
- *Riser height* is the distance, as measured in cm by a steel tape, between the parent slope ahead of the riser and a stadia rod placed flat on the lobe tread along the long axis of the lobe.
- *Riser angle* is the angle in degrees obtained from a digital level positioned directly against the riser or against a stadia rod held against the riser.

From these measured variables, a number of secondary variables were computed:

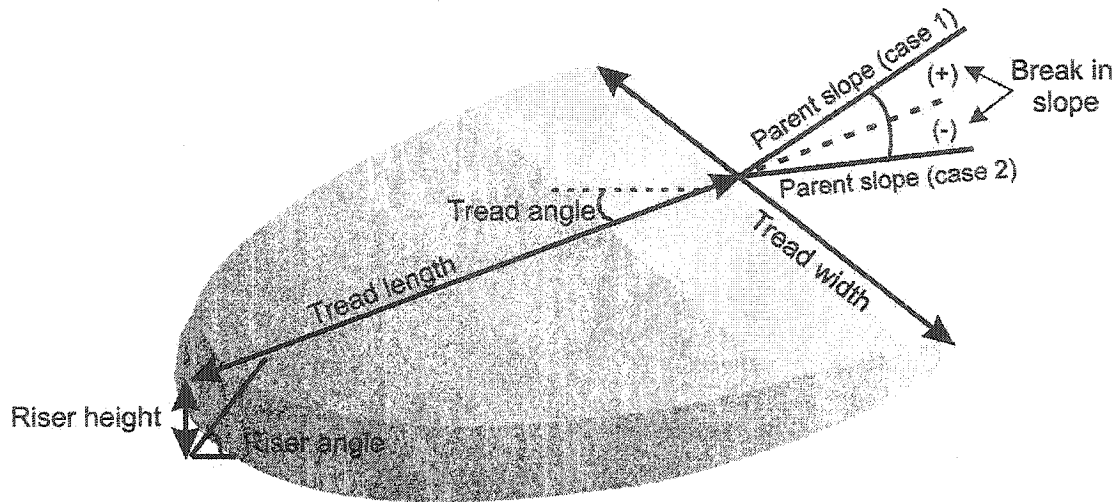


Figure 4.2: Morphometric parameters measured on solifluction lobes.

- Tread *length-to-width ratio* is the dimensionless ratio between the tread length and tread width.
- *Break-in-slope* is the difference in degrees between the angle of the parent slope and the tread angle. It gives the magnitude of the break-in-slope between the lobe tread and the parent slope. The parent slope was not measured in the field and was derived from the DEM.

4.3. RATES OF SURFACE MOVEMENT

Between May 26th and July 13th 2002, rates of surface movement were monitored on six solifluction lobes. The lobes were chosen to be as regularly spaced as possible along an upslope-downslope transect (see Figure 4.1). Between 14 and 19 surface targets were placed on each lobe's surface as soon as it was clear of snow. In some cases, snow was cleared manually to access the frozen ground. The targets consisted of small pieces of wooden dowel fixed onto the ground with a metal nail. They were set out in a cross

pattern, along the long axis of each lobe and on a transverse segment crossing the long axis near the middle of the tread (Figure 4.3). A taut line was used along the long and transverse axes of the lobe to precisely locate each target on the surface. Each target was inserted 3 cm in the ground with the nail head down, leaving the sharp end sticking 2 cm above the dowel. Most targets were fixed onto the ground surface before, or shortly after, thawing had begun.

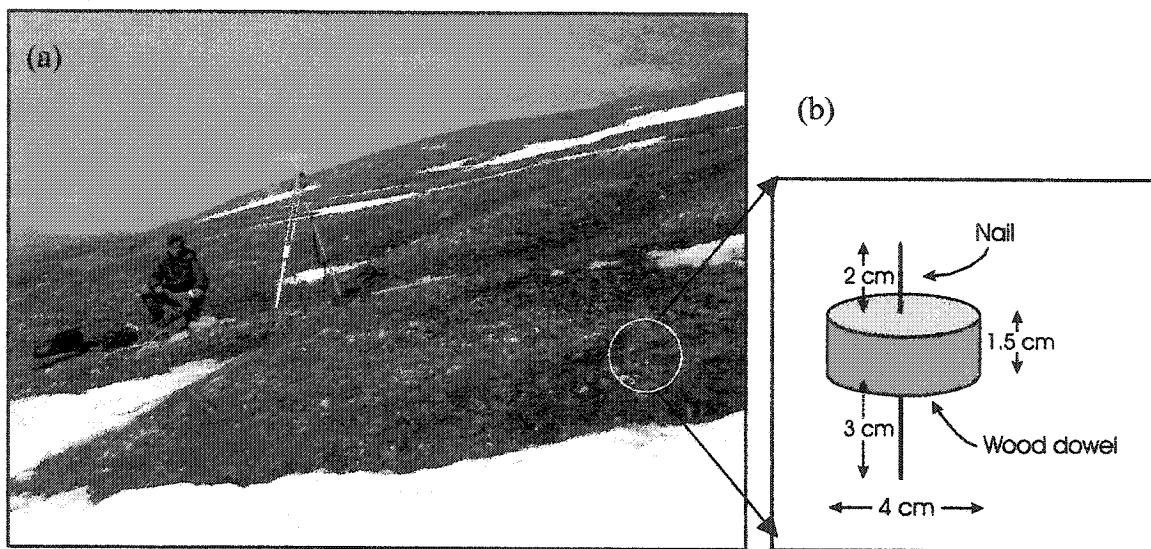


Figure 4.3: (a) Solifluction lobe with surface targets (site E); (b) Details of target.

Starting May 26th, displacement of the targets was measured using repetitive theodolite survey from benchmarks adjacent to each lobe. The benchmarks consisted of re-bar inserted as deeply as possible into frozen ground, and 0° sightings were made on the next closest benchmark, which also consisted of an embedded re-bar. For each survey station, two control points were chosen (mountain peaks and/or re-bars further away) to ensure that the station remained stable with time. At one station, it became apparent by the end of the field season that the tip of the re-bar had tilted downslope, and that site was

abandoned. Movement of the targets was recorded by the theodolite as changes in azimuth and converted into distances by trigonometric calculations. Readings of target movement were taken by sighting on the sharp end of the nail with the telescope successively pointed in both the normal and reverse position, and by averaging the results. This reduces the error due to tilting of the station after re-positioning. The mean precision on reading was 0.5 mm, which corresponds to the 10 second precision of the theodolite. The model used to derive distances from changes in azimuth assumes that movement occurs along a common downslope vector, parallel to the long axis and perpendicular to the transverse axis of the lobe. This somewhat simplified assumption permits the calculation of the downslope component of movement for each target. A full account of the computations involved in the calculation of surface movement is given in Appendix B. The targets were surveyed at each site every 2-4 days between May 26th and June 24th, and a last reading was taken on July 13th, giving a total of 48 days of cumulative displacement.

4.4. LONG-TERM RATES OF SOLIFLUCTION LOBE ADVANCE

In August 2001, 12 solifluction lobes with differing morphologies were chosen on the study slope to examine variability in long-term rates of frontal advance and to conduct stratigraphic observations. All the selected lobes had clear and well-defined fronts, and were chosen so as to cover the whole slope and provide a range of lobe sizes (see Figure 4.1, lobes in yellow). In the centre of each chosen lobe, a 2 m-long by 0.6 m-wide trench was excavated back from the front. All excavated lobes had one or more buried humus layers running continuously from the front of the lobe to the back of the trench. Samples

of buried humus were collected for radiocarbon dating at distances of 1 and 2 m behind the front, and at the base of the modern organic mat ahead of the riser (Figures 4.4 and 4.5). All lobes were sampled but not all samples were dated. Samples were collected from as close as possible to the base of the buried humus layer. Similarly, the humus collected for MRT determination was from the base of the modern organic mat. Basal sampling lessens the chances for contamination due to erosion of the top of the buried layer (Matthews and Dresser, 1983; Ballantyne, 1986).

Sampling was done by clearing the uppermost half of the buried layer and by removing the remaining half (at most) with a clean spatula, over a maximum lateral distance of 5 cm. At each sampling location, the thickness of the organic layer, the thickness of the sample collected and the lateral distance over which it was removed were noted. Each sample was sealed in double plastic bags immediately after its collection and was later allowed to air-dry in camp to avoid the formation of mould. In Ottawa, the samples were stored in a refrigerator before submission for dating. In addition to the humus samples collected for conventional radiocarbon dating, small wood fragments were extracted from three samples taken from two different lobes. These macrofossils were sent for AMS-dating in order to compare their ages with ones obtained from their respective enclosing humus samples.

Prior to shipment for radiocarbon assay, each sample was air dried, crushed and sieved at 1 mm with the purpose of concentrating the organic material and isolating possible modern rootlets (e.g. Dyke, 1981). 5 to 10 g were taken from the < 1 mm fraction of each sample to estimate the organic content by the loss-on-ignition method (LOI)

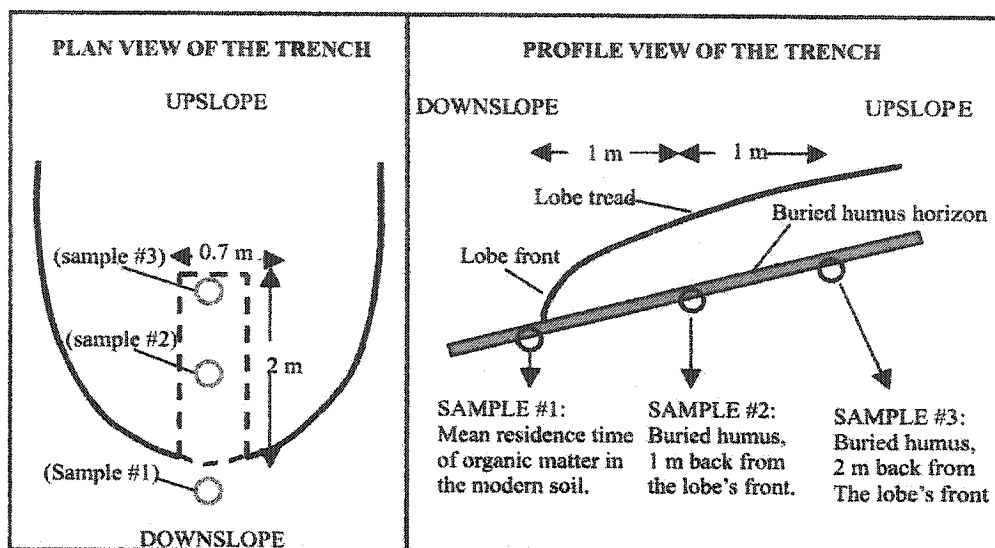


Figure 4.4: Sampling strategy for determination of long-term rates of lobe advance.

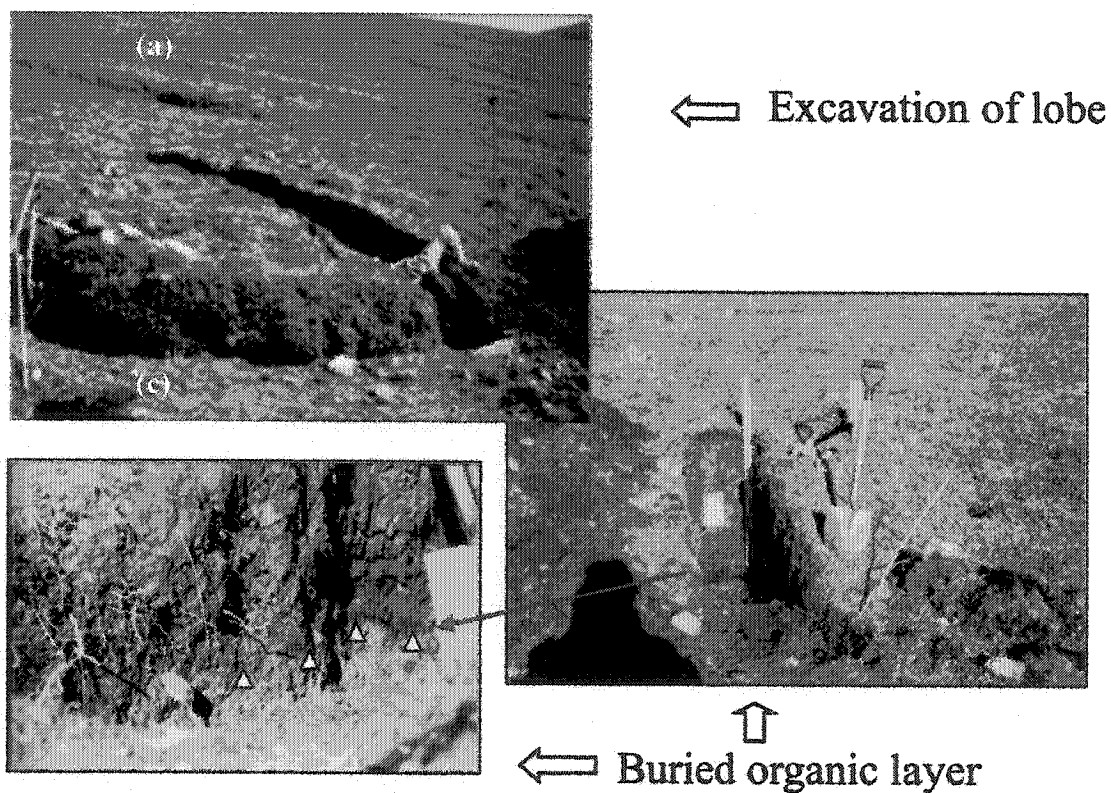


Figure 4.5: Excavation of a solifluction lobe: (a) trench in lobe 3; (b, c) buried organic horizon.

(Bengtsson and Enel, 1986). Approximately 5 g of sediment was put in cleaned weighed crucibles and placed overnight in an oven at 95°C. The sample was cooled to room temperature in a dessicator and the dry weight was measured. The crucible was then placed in a muffle furnace at 500°C for 3 hours, after which it was cooled again in a dessicator at room temperature. The ignited sample was then weighed and the difference in dry weight, expressed in percent of the total weight, yielded the percentage organic content.

The < 1 mm fraction of 25 samples were sent for conventional radiocarbon dating at the *Radiocarbon Laboratory* of the Geological Survey of Canada (3 samples), *Laboratoire de Datation au Carbone 14* de l'Université Laval (12 samples) and *Radiocarbon Laboratory* at Brock University (10 samples). Conventional radiocarbon ages were obtained by decay counting methods, which measure the activity of ^{14}C by detection of β -decays using either gas proportional counting of CO_2 or acetylene, or liquid scintillation counting of benzene (e.g. Trumbore, 2000). Age calculations are based on the Libby ^{14}C half-life of 5568 years and 0.95 of the activity of the NBS oxalic standard. Ages are quoted in radiocarbon years before present (BP), where 'present' is taken to be A.D. 1950 (Stuiver and Reimer, 1993). To account for isotopic fractionation effects, the samples sent to the GSC and at Brock University were corrected to a constant ^{13}C content of -25‰, using the measured $^{13}\text{C}/^{12}\text{C}$ ratio. Samples from Université Laval were not corrected for fractionation effects. The pre-treatment and counting techniques used by each laboratory are summarized in Table 4.1.

Table 4.1: Pre-treatment and counting technique used by each conventional ^{14}C dating laboratory.

<i>Treatment/ Lab</i>	<i>Acid leach (removal of contaminants)</i>	<i>Baseleach (removal of contaminants)</i>	<i>Rinse</i>	<i>Counting method / gas</i>	<i>Age correction</i>
GSC	yes	yes	Distilled water	b-decays / CO_2	to $\delta^{13} = -25$ ‰
Brock U.	10% HCL	no	Distilled water	Liquid scintillation / benzene	to $\delta^{13} = -25$ ‰
Laval U.	5% HCL	no	no	Liquid scintillation / benzene	no

The wood fragments collected in humus samples were sent to the *Isotracer Laboratory* at the University of Toronto for accelerator mass spectrometry dating (AMS). AMS uses a particle accelerator to achieve energies high enough to measure individual ions of ^{14}C . The results are the average of two separate analyses made on the cellulose fractions, using the Libby ^{14}C meanlife of 8033 years. The sample ages were corrected for isotopic fractionation to a base of -25‰, using the measured $^{13}\text{C}/^{12}\text{C}$ ratio.

The ^{14}C content of the atmosphere has varied with time, both because of changes in the production rate of ^{14}C (cosmic ray flux and magnetic field variations) and because of changes in the distribution of carbon among oceans, biosphere and atmospheric reservoirs (Stuiver and Reimer, 1993). More recent changes in the ^{14}C content of atmospheric CO_2 have resulted from dilution by ^{14}C -free fossil-fuel-derived carbon and by the production of ^{14}C during atmospheric testing of thermonuclear weapons (bomb ^{14}C). Correction factors to determine calendar ages (calibrated ^{14}C dates) from conventional ^{14}C ages are based on the observed ^{14}C content of independently dated tree ring cellulose (Trumbore, 2000). Calibrated dates (Stuiver and Reimer, 1993) were provided by Laval and Brock

University. The dates from the GSC were calibrated using the program CALIB 4.1, written and distributed by the Quaternary Isotope Lab, University of Washington (Stuiver and Reimer, 1993). CALIB makes the conversion from radiocarbon age to calibrated calendar years and calculates the probability distribution of the sample's true age.

In addition to sampling for buried organic layers, basic stratigraphic observations were made at each of the excavated lobes. A simple drawing of the internal stratigraphy was produced, and the depths of buried humus layers were measured. The thickness of the organic mat covering each lobe was measured with a steel tape ahead of the lobe, at the top of the riser, and on the tread at the back of the trench. The cross-profile of each lobe was measured by theodolite survey from a fixed point adjacent to the lobe.

4.5. SEDIMENT ANALYSES

4.5.1. Grain-size distribution

Two soil samples were taken two meters behind the front of each excavated lobe, at depths of 40 and 80 cm on one side of the trench. Both depths were combined and a total of 13 samples were analysed. Grain-size distributions were obtained using a combination of wet-sieving and laser particle-size analysis. The granulometry of the coarse fraction ($> 63 \mu\text{m}$) of each sample was determined using ASTM guidelines for wet sieving. 50 to 70 g of soil were taken from each sample and soaked in Sodium Hexametaphosphate ($\text{Na}_6\text{O}_{18}\text{P}_6$) for 48 hours. The mixture was then put in a milkshake mixer for approximately 1 minute and poured into a stack of sieves with openings of 4 mm, 2 mm,

1 mm, 0.5 mm, 0.25 mm, 125 μm and 63 μm . The portions left on each sieve layer were collected, oven dried, and weighed (ASTM 2001a).

The granulometry of the fine fraction ($< 63 \mu\text{m}$) was determined using a Laser Particle-Size Analyser (PSA). Organics were first removed with 30% Hydrogen peroxide (H_2O_2) (Black *et al.*, 1965). Following this, 10 to 20 g of each sample were put in 1000 mL beakers and 50 mL of 30% H_2O_2 was added. After 24 hours, another 50 mL was added and the mixture was warmed. Following the burning of organics, the samples were air-dried, placed into small jars and left to soak in 50 mL of 50g/L of $\text{Na}_6\text{O}_{18}\text{P}_6$ for 2-3 days. The samples were then taken to the Sedimentology Laboratory at the Geological Survey of Canada (GSC) to perform the PSA tests, using a LECOTRACTMLT-1 version 7.01 and LECOTRACTMASVR systems. At the GSC laboratory, the samples were first mixed with deionized water for 1 minute, and then wet sieved through a 90 μm sieve into a 1 L beaker. The 1 L beaker containing the $< 90\mu\text{m}$ fraction was placed on a bench stirrer to keep the particles in suspension. Approximately 4 mL of this solution was collected with a plastic syringe and put into the reservoir of the PSA. The sample was exposed to 30 seconds of ultrasonic waves to break up any aggregates before the test was run. Each test yielded a cumulative distribution curve for the $< 90 \mu\text{m}$ fraction. These results were combined with the ones obtained from wet sieving to construct a total cumulative grain-size distribution curve for each sample.

4.5.2. Atterberg limits

The liquid limit, plastic limit and plasticity index for each of the 12 soil samples were determined using ASTM guidelines (in Bowles, 1992). The plastic limit (w_p) is the

moisture content below which the soil is non-plastic. The liquid limit (w_L) is the moisture content below which the soil behaves as a plastic material. At this moisture content, the soil is on the verge of becoming a viscous fluid. The plasticity index (I_p) is the range of moisture contents over which the soil behaves as a plastic material, and is equal to the difference between the liquid and plastic limit ($I_p = w_L - w_P$) (Bowles, 1992).

The liquid limit of a soil is arbitrarily defined as that water content at which a pat of soil placed in a brass cup (Cassagrande device), cut with a standard groove, and then dropped from a height of 10 mm will undergo a groove closure of 12.7 mm when the cup of soil is dropped 25 times at the rate of 120 drops/minute (Bowles, 1992). The test uses a standard liquid limit device and wedge grooving tool. For each of the 12 soil samples, approximately 250 g were sub-sampled and washed through a #40 sieve (400 μ m). The residue was collected in a 1000 mL beaker, which was stored for two days to permit settlement of the soil particles. Following this, excess water was decanted manually and the beaker was left again to air-dry at room temperature. Once the beaker had lost sufficient water, the moist soil was placed on a glass plate and dried with a portable hair drier until a uniform creamy consistency was attained by mixing with a spatula. 20 to 30 g of the soil was put aside for the plastic limit test. A small portion of the soil was then placed into the brass cup of the Cassagrande device and a groove was cut using the wedge grooving tool. The test was performed by rotating the crank at approximately 120 rpm until the groove closed by 12.7 mm (visual observation). A sample of about 40 g was then taken from the brass cup and placed overnight in the oven for gravimetric water content determination. The test was repeated 5 times for each soil sample, adding water to the sample before each test. The logarithm of the number of blows (N) required to

close the groove was plotted against the corresponding water content (%) for each test and the water content at N=25 was found from linear regression.

The plastic limit has been arbitrarily defined as that water content at which a soil thread just crumbles when it is rolled down to a diameter of 3 mm (Bowles, 1992). The 20-30 g of soil set aside from the liquid limit test was broken into 5 smaller samples. Each of these samples was rolled between the fingers, using sufficient pressure to form a thread of uniform diameter. If the soil thread diameter became 3 mm and was not crumbling, the thread was broken into several smaller pieces, reformed into a ball and the process repeated, until the 3 mm thread reached the state where it is about to crumble. When this was attained, the sample was placed into a small pre-weighed container and stored under a damp paper towel. Once five soil threads were tested, the container with the threads was put in the oven overnight to determine the gravimetric water content, which represented the plastic limit of the sample. The plasticity index for each of the 12 soil samples was then calculated as the difference between the values obtained for the liquid and plastic limits.

4.5.3. Specific gravity, unit weight and void ratio

The specific gravity of any substance is defined as the density of the material divided by the density of distilled water at 4°C:

$$G_s = \rho_{\text{material}} / \rho_{\text{water at 4°C}} \quad (4.1)$$

where G_s is taken to be the average value for soil particles (Bowles, 1992). G_s can also be computed using any ratio of equal mass volume of substances to water as:

$$G_s = (M_s / V) / (M_w / V) \quad (4.2)$$

Where the volume terms V cancels. The specific gravity determination thus reduces to obtaining the volume of a known mass of soil particles and dividing this by the same volume of water. 100 to 200 g were taken from four different soil samples (four lobes), pulverized and put into 500 mL flasks, which were subsequently filled with water stabilized at 20°C. De-aeration of the soil-water mixture was accomplished by heating. The flask containing the soil-water mixture was weighed (M_{bws}). The flask was emptied and the soil put overnight in the oven for drying, and then weighed (M_s). Next, the flask was filled with stabilized water, heated for deaeration, and weighed (M_{bw}). The specific gravity of each sample (G_s) can be calculated from:

$$G_s = M_s / (M_{bw} + M_s - M_{bws}) \quad (4.3)$$

Samples from four dated solifluction lobes were collected in June 2002 to determine the density of the soil. A 470 cm³ metal canister was pushed in the side of the re-excavated trench, 2 m back from the lobe front and approximately 10-15 cm below the surface. The moist soil collected in the canister and sealed in plastic bags was weighed in the laboratory back in Ottawa. Next, the samples were oven-dried to obtain the dry weight and water content of the sample. The wet density (ρ_{wet}) and dry density (ρ_{dry}) are, respectively:

$$\rho_{wet} = \text{mass of wet soil} / \text{volume of canister} \quad (4.4)$$

$$\rho_{dry} = \rho_{wet} / 1+w \quad (4.5)$$

where w is the gravimetric water content expressed as a decimal. The unit weight (wet and dry) for each soil sample is then:

$$\gamma_{(wet/dry)} = \rho_{(wet/dry)} \cdot \gamma_w \quad (4.6)$$

where γ_w is 9.807 kN/m³. The *void ratio* (e) can then be calculated as the ratio between the value for the dry density and the one for specific gravity:

$$e = (G_s / \rho_{\text{dry}}) - 1 \quad (4.7)$$

4.6. THAW DEPTHS AND GROUNDWATER LEVELS

Starting May 29th 2002, thaw depths were measured at 13 separate sites along a downslope transect (see Figure 4.1, sites A-L). Of these, 10 were solifluction lobes and three were non-lobe sites. The six lobes used to monitor surface movement were included, and four additional lobes were also chosen along the slope. At the lobe sites, thaw depth was measured at four different points: ahead of the riser, on the crest of the riser, on the middle of the tread and at the back of the tread. At non-lobe sites, thaw depth was measured at one location only. Thaw depth was obtained by driving a steel rod into the ground until frozen soil prevented further penetration. The length of the rod section penetrating the unfrozen soil was then measured with a steel tape. These steps were repeated three times within a radius of 30 cm at each probing location, and the results averaged. Thaw depths were measured every 1-2 days at each site until June 24th, and a last reading was obtained on July 13th, 2002.

Groundwater levels were measured at the same locations as the thaw depths. Prior to ground thaw in mid-May 2002, 2.5 cm diameter holes were drilled at every sampling location with a soil auger (Figure 4.6). Perforated PVC pipes were fitted in each hole to an average depth of 75 cm, and left sticking 10-30 cm above the surface. The depth to the water table was measured using a groundwater probe. The probe consisted of a



Figure 4.6: Drilling into an emerging solifluction lobe, to install groundwater pipes.

battery-powered electronic beeper connected to two open-ended wires running along a 1.2 m-long bamboo rod. The bamboo was slowly lowered down into the groundwater pipe until the two open-ended wires touched the water surface, which induced an electrical contact and triggered the beeper. The length of the bamboo penetrating the pipe was then measured with a steel tape. All the pipes had to be re-excavated toward the end of the study period, because the thaw fronts had penetrated beyond the depth of the pipes. As with thaw depths, groundwater levels were measured every 1-2 days, starting May 29th until June 24th 2002, and a last reading was taken on July 13th 2002.

4.7. PRECISE MONITORING OF A SOLIFLUTION LOBE

Between May 25th and July 17th 2002, temperature, moisture and deformation of the thawing soil were monitored continuously on the tread of a medium-sized solifluction lobe, using a 21X Campbell Scientific data-logger and 4 x 16 relay multiplexer. Soil

temperature was measured using seven thermistors fixed along a thin bamboo rod, inserted in a hole previously drilled into the frozen ground. The thermistors were fixed at 0.5, 5, 15, 25, 35, 55 and 95 cm beneath the soil surface. The space left around the rod following its placement in the hole was backfilled with sand collected nearby on the slope. The leads from the seven thermistors were connected to the data-logger via the multiplexer relay and temperatures were sampled at 6-minute intervals. Hourly averages were computed automatically and stored in a Campbell Scientific SM716 storage module connected to the data-logger.

Adjacent to the temperature probe, the volumetric water content of the thawing soil was monitored using a Delta-T Profile Probe type PR1. The Profile Probe measures soil moisture content at different depths within the soil profile. It consists of a sealed composite rod, ~25 mm diameter, with electronic sensors arranged at fixed intervals along its length. The 110 cm-long probe used had 5 cm-long sensors placed at 10, 20, 30, 40, 60 and 100 cm depths. Readings were taken with the probe inserted into specially constructed thin-walled fibreglass tubes, which maximize the penetration of the electromagnetic field into the surrounding soil. The output from each sensor is a simple analogue DC voltage, converted into soil moisture using general soil calibration (Delta T Devices Ltd., 2001). When power is transferred from the data-logger to the profile probe, it generates a 100 MHz signal which is transmitted to pairs of stainless steel rings (the sensors). The sensors then generate an electromagnetic field which extends about 100 mm into the soil. The water content of the soil surrounding the rings determines its dielectric properties, which is a measure of a material's response to polarisation in an electromagnetic field. Since water has a dielectric constant of ~81 compared to soil ~4,

ice ~3, and air ~1, the probe is highly sensitive to changes in the unfrozen water content of the soil. A 1 m-long hole was drilled into the frozen ground on the lobe tread and an access tube was fitted tightly in the hole. Additional access tubes were also installed at the back of the tread and on the crest of the riser. The presence of blocky material at the riser prevented a complete insertion of the access tube and the tube was inserted to a depth of 50 cm beneath the surface, allowing only two sensors to be used, at 10 cm and 40 cm depths. The probe was used to monitor soil moisture continuously at the centre of the lobe tread, but was moved briefly to the other access tubes every 1-2 days to obtain near-daily measures of water content. The Profile Probe was connected to the data-logger and values of moisture content were sampled at 6-minute intervals, simultaneously with the temperature probe. The data were converted automatically into hourly averages and stored in the SM176 module.

Soil movement during ground thaw was monitored at different depths using a custom-built strain probe (Yamada and Kurashige, 1996). The strain probe consisted of a flexible spring steel strip 110 cm-long, 2 cm-wide and 0.3 mm-thick. Two pairs of *Omega* Strain gauges (Serial no: KFG-10-120-C1-11L1M2R) with dimensions of 10 x 3 mm and a nominal resistance of 120 ± 0.4 ohms were attached to both sides of the strip at 10 cm intervals, permitting strains to be recorded at 10 different depths (Figure 4.7). The strain gauges are made of a 6 μ m-thick foil measuring grid embedded into a 24 μ m-thick, 15 mm-long and 5 mm-wide Kapton carrier. Gauges were placed every 3.5 cm along a 1 m-long segment of the steel strip, using a cold-curing methyl-based cyanoacrylate adhesive. The wires connected to each strain gauge (two per gauge) were glued at several points along the strip, using the same adhesive. Next, a thin silicon coating was applied to each

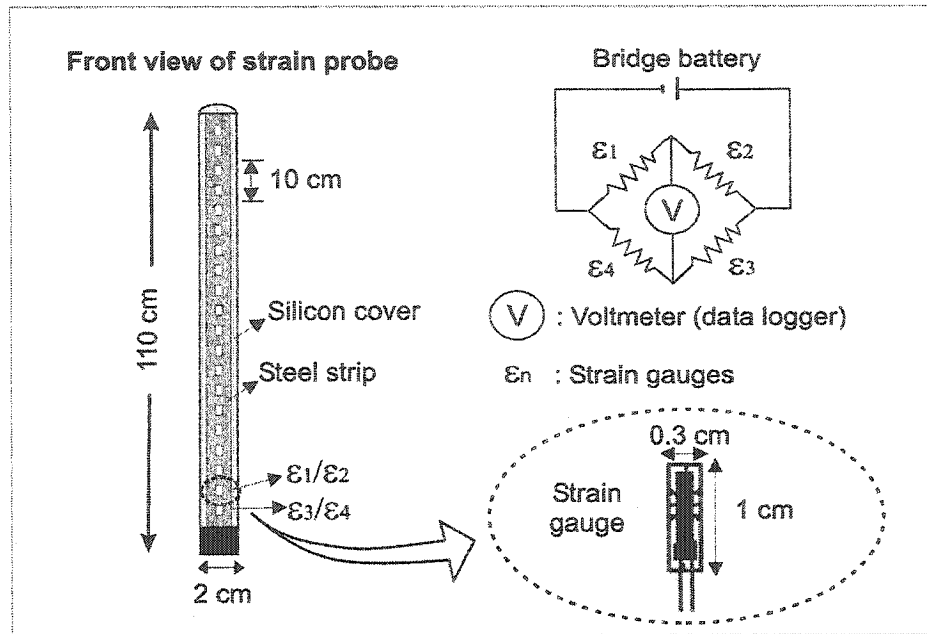


Figure 4.7: Design of the strain probe and bridge circuit. Each of the 10 sections (length 10 cm) has 4 strain gauges covering both sides of the probe and connected into a bridge circuit (after Yamada and Kurashige, 1996).

gauge separately, and then over the full length of the probe, in order to insulate and protect it from physical damage.

The silicon cover was smoothed as much as possible to give the probe a strong but flexible consistency. The two pairs of strain gauges for each 10 cm long sections were connected into a wheatstone bridge circuit. This type of circuit can compensate for thermal drift of the gauges because opposed gauges in the circuit undergo similar temperature changes. It also generates four times as much voltage compared to only one gauge per section. Each of the 10 circuits was connected to the multiplexer panel, through which the data-logger sends an excitation voltage of 0.5 V. The returned voltage for a particular section depends on the resistance offered by the corresponding strain gauge circuit, which changes when the section is bent.

The strain probe was inserted in a hole previously drilled into frozen ground. A small piece of wood dowel was fixed at the lower end of the probe to stabilize its base at the bottom of the hole. The space left around the hole was filled with wet sand. Progressive bending of the probe by solifluction induced strain variations throughout the probe, which were sampled by the data-logger at minute intervals, and recorded as hourly averages. The conversion of strain values into soil-displacement values was achieved using the method of Yamada and Kurashige (1996), the complete solution of which is presented in Appendix C. The theoretical error on each strain reading corresponds to the precision of the data-logger (0.001 mV/V), which gives an error of $\pm 0.48 \mu\epsilon$. Using the conversion method of Yamada and Kurashige (1996), this gives an error of ± 0.024 mm (see Appendix C). However, considering the effect of other variables such as the voids left around the probe when the borehole was backfilled, and variations in the thickness of the silicon coating around the probe, an error of ± 0.5 mm is a more realistic estimate.

An attempt was also made to measure the surface displacement on lobe tread, using two modified ablatometers (Lewkowicz, 1985) running from a metal beam held in place by two pieces of re-bar to a footplate fixed onto the surface of the lobe. The apparatus was inspired by the system used in a laboratory experiment by Harris et al. (1996), in which two linear voltage displacement transducers (LTVD) were used to monitor frost heave and downslope displacement. In this case, each ablatometer consists of an extending rod connected to a potentiometer. Variations in the length of the rods induced by soil movement are registered as a change in voltage by the data-logger. As thaw settlement occurred, the change in length of the ablatometers was monitored, allowing the vector of footplate movement to be calculated to an estimated accuracy of ± 0.5 mm. A tipping-

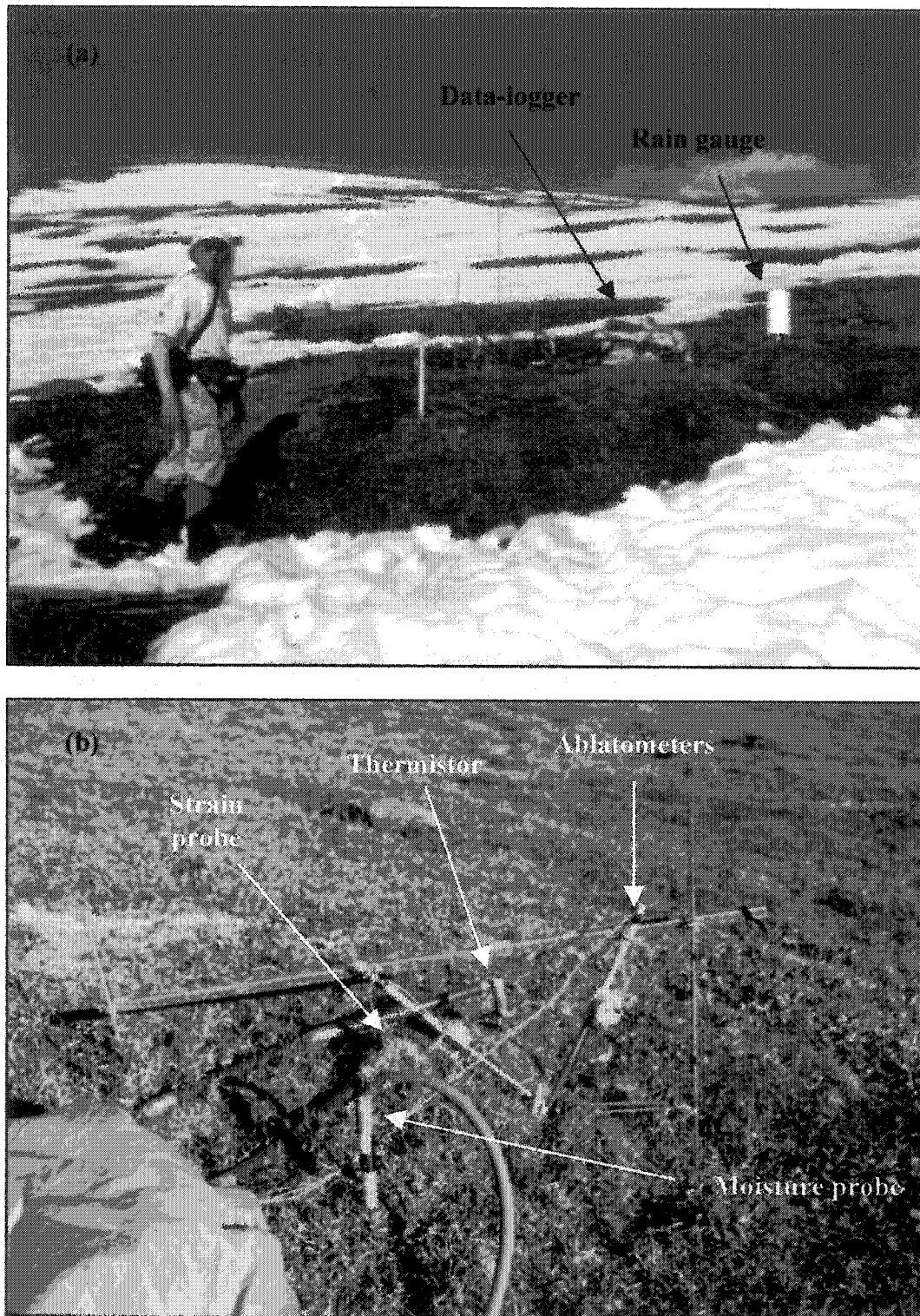


Figure 4.8: Monitoring station on solifluction lobe I. (a) view from the front; (b) close up showing the different components of the system.

bucket rain gauge was installed to measure precipitation adjacent to the monitored solifluction lobe. The rain gauge was connected to the data-logger, and readings were taken at 6-minute intervals, and stored as hourly averages in the SM176 module. The complete monitoring station is shown in Figure 4.8.

5. RESULTS

5.1. SPATIAL DISTRIBUTION OF SOLIFLUCTION LOBES AND RELATED MORPHOMETRIC VARIABLES

5.1.1. Dispersion pattern

The 235 solifluction lobes mapped during the summer 2002 were super-imposed onto the digital elevation model of the study slope (Figure 5.1). The DEM shows a slope with a broadly convexo-concave downslope profile. Solifluction lobes cover the slope almost uniformly, and appear to gradually increase in size downslope. Stripes cover the top of the slope, and these evolve into small non-sorted steps near the slope crest. The upper segment of the slope has a high concentration of small lobes, while the toe of the slope is occupied by larger lobes grading into terraces, which eventually disappear at the

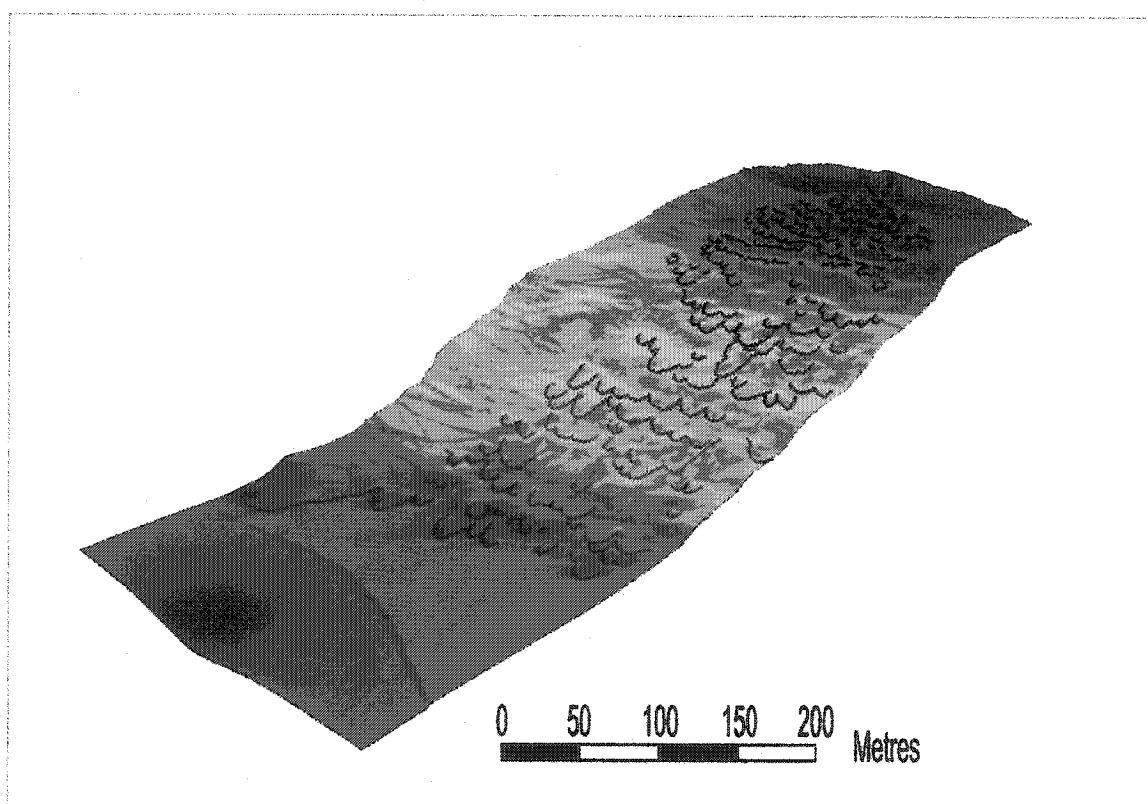


Figure 5.1: DEM of study slope showing the dispersion of solifluction lobes.

gently sloping basal segment. The detailed mapping shows that the majority of the lobes on the slope do not occur as single landforms but rather as parts of more-or-less continuous lobate segments (see also Figure 4.1). Hence, the slope appears as an organized succession of bench-like wave patterns, rather than as a random distribution of single landforms of varying sizes.

A first idea of the spatial structure of the lobe population can be obtained using techniques of point pattern analyses, such as nearest neighbour analysis (e.g. Shaw and Wheeler, 1985). The analysis involves a statistical comparison between an observed spacing of points and the spacing expected in a random pattern. The distance between each lobe centroid and its nearest neighbour (r) was calculated using an extension of ARCGIS software (Sawada, 2000). The average observed distance (r_a) was computed as:

$$r_a = \Sigma r / N \quad (5.1)$$

where N is the number of lobes on the slope. The expected average distance (r_e), whose distribution is described by the Poisson function is computed as:

$$r_e = 0.5 \sqrt{A/N} \quad (5.2)$$

where A is the area of the study slope and N the number of lobes. The area was defined by a convex hull enclosing the cloud of points. The nearest neighbour statistic (R) was derived by dividing the observed distance by the expected:

$$R = r_a / r_e \quad (5.3)$$

The R statistic shows how much more, or much less, spaced the observed distribution is compared with a random one. R values extend along a continuous scale extending from 0 (completely clustered pattern) through 1.0 (random) to 2.15 (perfectly regular) (Pinder

and Witherick, 1972). The probability associated to the R statistic is established using the standard error and z scores (e.g. Shaw and Wheeler, 1985). The results of the test are presented in Table 5.1.

Table 5.1: Results from the nearest neighbour analysis.

<i>Measured average distance (m)</i>	<i>Expected average distance (m)</i>	<i>Area (m²)</i>	<i>R statistic</i>	<i>Z value</i>	<i>Probability (%)</i>
7.36	6.64	40426	1.11	3.12	0.0019

The test yields an R value of 1.11, which was slightly, but significantly different from 1 at $p \leq 0.01$. This indicates that the location of the solifluction lobes is non-random and that there is a statistically significant degree of regularity in the spatial arrangement of solifluction lobes on the slope. At the same time, however, the R value is not far from unity, meaning that the regularity is not very strong.

Examination of the map suggests that spacing of the lobes varies downslope and this is demonstrated in Figure 5.2. This graph shows a downslope increase in lobe spacing, suggesting that environmental conditions must vary downslope to create this pattern. Hence, it is best to describe the spatial distribution as one where solifluction lobes tend to arrange somewhat regularly over an anisotropic surface, and where downslope variations in environmental conditions result in a downslope increase in lobe spacing.

5.1.2. Frequency distribution of lobes

The frequency distribution of solifluction lobes in an upslope-downslope direction is plotted in Figure 5.3. The distribution is positively skewed, with more lobes being found on the upper segments of the slope than on the footslope. Part of the decrease in lobe

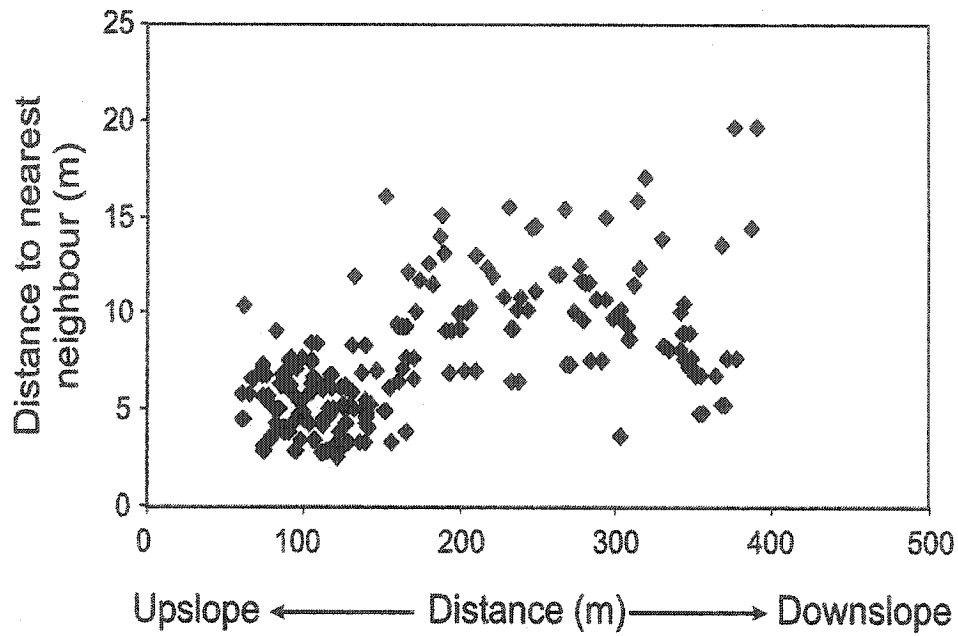


Figure 5.2: Downslope variation in nearest neighbour distance.

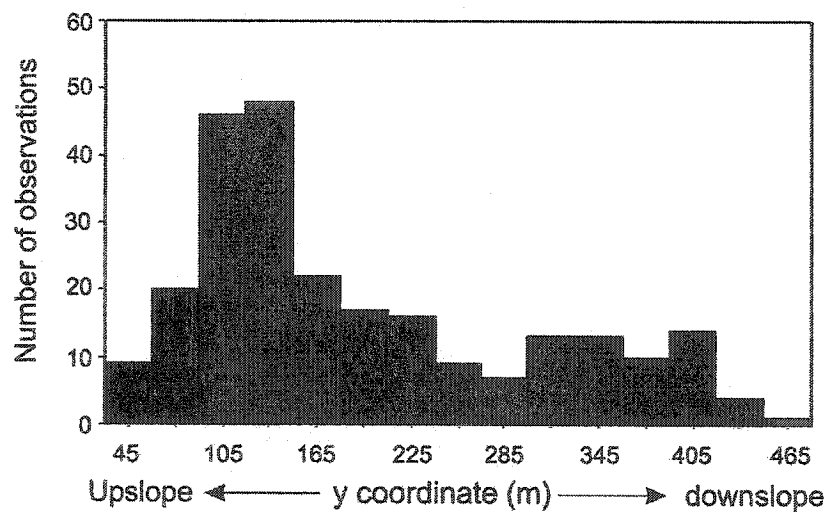


Figure 5.3: Downslope frequency distribution of solifluction lobes calculated over 30 m-long segments on the study slope (first interval starts at 30 m).

frequency at the uppermost boundary of the slope is attributed to sampling, which stopped once lobes became too small and too numerous to be mapped. The distribution is of the log-normal type, in that it becomes normal following a logarithmic transformation

(Legendre and Legendre, 1998). The downslope frequency distribution provides a picture of the structure of the lobe population, and supports the concept of an organized locational pattern which responds to spatial gradients in controlling variables.

5.1.3. Solifluction lobe morphometry

The morphometric measurements made on the 235 lobes mapped on the slope are similar to those made on the same slope by Hugenholtz (2002), who sampled 25 lobes along 5 downslope transects. The aim here is to apply the same kind of analysis to the whole population of lobes. Results are summarized in Table 5.2, and the frequency distribution of each variable is presented in Figure 5.4. Tread length, tread width, riser height and riser angle exhibit significant variability, and their distributions show pronounced positive skewness. Tread angle and length-to-width ratio are less variable and their corresponding distributions have only a slight positive skewness. A Kolomogorov-Smirnov test was used on each observed distribution to test for normality. Tread length, tread width, riser height and riser angle were found to have non-normal distributions at the 0.05 level. Normality was obtained for each of these variables following a logarithmic transformation. Tread angle and length-to-width distributions were found to be normal at $p < 0.05$.

Table 5.2: Summary statistics of morphometric variables.

<i>Parameter</i>	<i>Mean</i>	<i>Median</i>	<i>Standard deviation</i>	<i>Coefficient of variability</i>	<i>Min.</i>	<i>Max.</i>	<i>Skewness</i>
Tread length (m)	5.66	4.90	3.31	0.58	1.50	25.00	2.16
Tread width (m)	6.70	5.70	3.70	0.55	2.10	25.00	1.78
Tread angle (°)	12.4	12.3	3.1	0.25	4.1	20.0	0.18
Riser height (m)	0.35	0.32	0.17	0.49	0.05	1.30	1.46
Riser angle (°)	36.8	32.8	14.5	0.39	14.6	90.0	1.45
Length-to-width	0.86	0.85	0.19	0.22	0.23	1.54	0.14

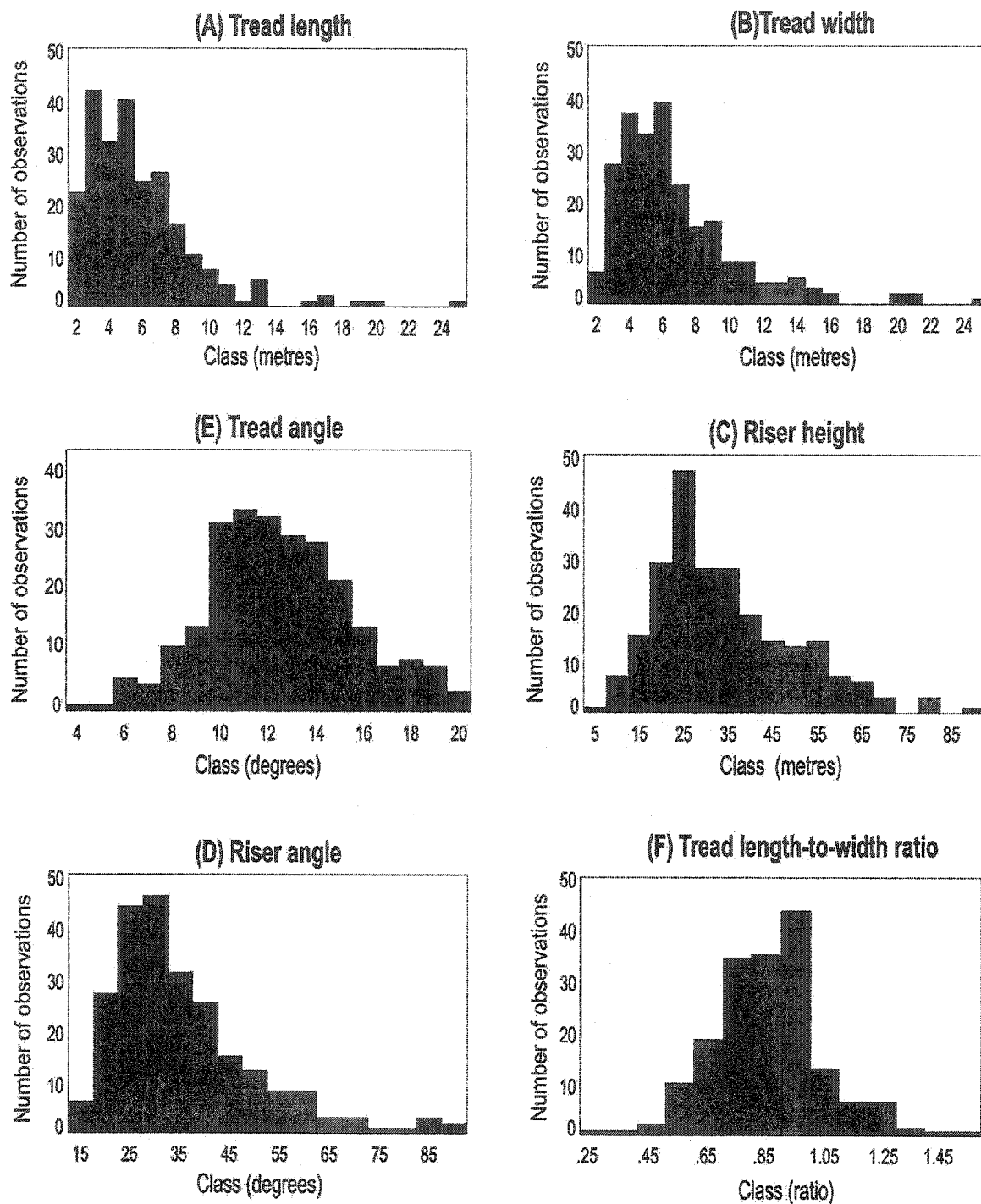


Figure 5.4: Histograms of morphometric parameters measured on solifluction lobes.

Differences arise between the summary statistics obtained in this study and those obtained by Hugenholtz (2002). Mean values for tread length (11.18 m), tread width (16.12 m), riser height (0.82 m) and riser angle (53.2°) calculated by Hugenholtz (2002) are all greater than those obtained in the present study, while his mean length-to-width ratio (0.76) was lower than the value found in this study (0.86). Only the mean tread angles were comparable (12.1° versus 12.4° in this study). It appears that the sampling technique used by Hugenholtz (downslope transects) was biased toward larger lobes having higher and steeper fronts. The upper segment of the slope, where the smallest forms are found, was probably underrepresented by the sampling transects, and because ideal, discrete landforms were favoured during sampling (Hugenholtz, 2002). On the other hand, the shape of the morphometric distributions are comparable with those of Hugenholtz (2002) for his pooled data ($n = 85$), which includes the whole mountain plateau. In both studies, log-normal distributions were found for tread length, tread width and riser height, while normal distributions were found for tread angle and length-to-width ratio.

5.1.4. Spatial relationships in lobe morphometry

Significant downslope trends were found for several morphometric variables (Figure 5.5). Tread width, tread length, riser height and riser angle all increase linearly downslope, with correlations significant at $p \leq 0.01$. Tread width exhibits the most pronounced trend, followed by tread length, while riser height and riser angle have lower R^2 values. The downslope trends identified for the main morphometric parameters confirm the qualitative statement made from inspection of the map, that lobes tend to increase in size from the slope crest to the bottom of the slope.

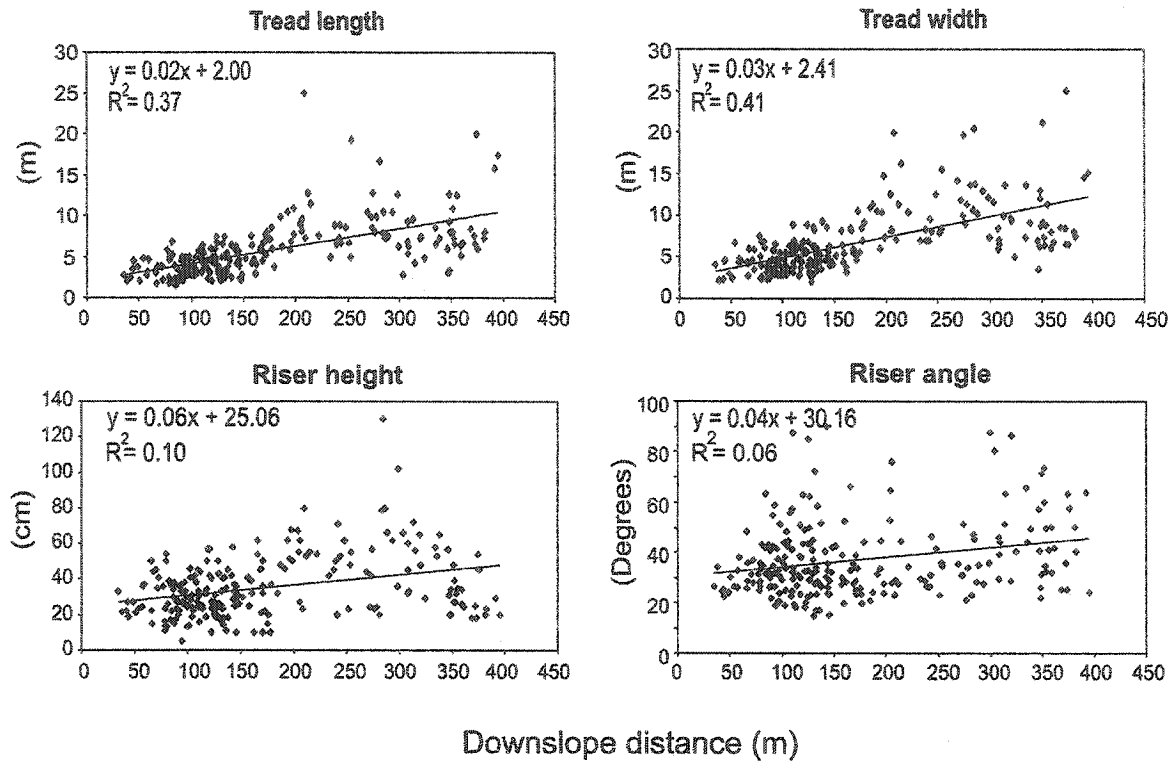


Figure 5.5: Downslope variation of morphometric variables: tread length, tread width, riser height and riser angle. Distances are in meters from the slope crest.

5.1.5. Relationships among morphometric variables

Correlation analysis was used to test for bivariate relationships between the main morphometric parameters. Because correlation analysis requires normal data, logarithmic transformations were used on tread length, tread width, riser height and riser angle to achieve normality, while tread angle and length-to-width ratios were not transformed, as they have normal distributions. Table 5.3 shows the resulting correlation matrix, with significant correlations ($p \leq 0.01$) indicated in bold. Significant positive correlations appear between tread length, tread angle and tread width; between the tread width, riser height and riser angle; and between the riser height and riser angle. Negative correlations were found between the length-to-width ratio, the riser height and

Table 5.3: Pearson-product moment correlations (r) between main morphometric parameters.

<i>parameter</i>	<i>tread length</i>	<i>tread width</i>	<i>Tread angle</i>	<i>Riser height</i>	<i>Riser angle</i>	<i>Length- to-width</i>
<i>Tread length</i>	1					
<i>Tread width</i>	0.93	1				
<i>Tread angle</i>	0.20	0.11	1			
<i>Riser height</i>	0.38	0.44	-0.01	1		
<i>Riser angle</i>	0.06	0.17	0.02	0.46	1	
<i>Length-to-width</i>	0.12	-0.12	0.15	-0.24	-0.27	1

the riser angle. A particularly strong positive correlation exists between the tread length and tread width ($r = 0.93$), indicating that these two parameters are closely related.

Among these relationships, tread length, tread width and riser height are the most correlated, indicating that lobes tend to enlarge consistently in three dimensions. The strong correlation between tread length and tread width, and the absence of correlation between these parameters and the length-to-width ratio suggests that lobes conserve a constant plan form despite their respective size. The results also indicate that steep lobe risers tend to be higher, and that lobes with widths larger than normal, tend to have steeper fronts. Consequently, lobes with low length-to-width ratios, that is lobes that are wider than longer, tend to have steeper and higher risers.

Hugenholtz and Lewkowicz (2002) reported a strong correlation between tread width and tread length ($r = 0.83$, $p < 0.001$) for a sample of 25 lobes taken on the same slope. They also showed that the riser height was best correlated with tread length ($r = 0.47$, $p < 0.05$) and tread width ($r = 0.55$, $p < 0.01$), while the riser angle was correlated with tread length ($r = 0.54$, $p < 0.01$) and tread width ($r = 0.48$, $p < 0.05$). However, the significant

correlations reported by Hugenholtz and Lewkowicz (2002) between the riser angle and lobe dimensions (tread length and width) differ from this study, which showed only a slight (but significant) positive correlation between riser angle and tread width ($r = 0.17$, $p < 0.01$). This weak positive correlation is thought to arise from the broad, concurrent increase in lobe size and riser angle downslope (see Figure 5.5). Among the lobes in general, the riser angle showed a poor correlation with other size parameters. This is probably because the morphology of the riser fluctuates in time as the lobe advances, independently of the overall size of the lobe (length and width), but with concurrent changes in the plan form of the lobe (length-to-width ratio).

5.1.6. Synthesis

The nearest neighbour analysis indicates significant regularity in the spatial arrangement of lobes. The relatively low R value (1.1), however, means that other factors counteract the regular component of the pattern. This is confirmed when looking at the downslope frequency distribution of lobes, which supports the visual impression that lobes tend to be closer-spaced on the upper slope than on the lower slope. Part of this increase in lobe spacing may be attributed to the downslope trend in lobe size. As lobes increase in size downslope, they occupy more space, and their density must decrease accordingly. The strong positive skewness in the downslope frequency distribution of lobes is thus linked, at least partly, with the strong positive skewness in lobe size parameters (tread length and tread width, see Figure 5.4).

5.1.6.1. *Spatial rhythmicity of solifluction lobes*

The question remains whether solifluction lobes align themselves along consistent and recurrent wave patterns, as suggested by inspection of the map, even though the nearest

neighbour analysis points to the presence of only a slight component of regularity in the dispersion of lobes. A simple technique was developed to investigate and quantify the rhythmic component of solifluction lobes on the slope:

- (1) Each lobe was placed in order of downslope occurrence (increasing Y coordinate), to form a one-dimensional spatial series.
- (2) Because of probable friction effects on the edges of the slope, the apparent pattern of lobe alignment has a convex-downslope appearance (see Figure 4.1), which confuses the density signal along a downslope vector. To attenuate this distortion effect, a 7-lobe running mean was applied on the Y-downslope (coordinate) spatial series. The new coordinate for each lobe was then the average of the lobe's own Y coordinate and its six nearest neighbours in a downslope direction. The effect of this was to smooth the spatial distribution of lobes and increase relative differences in lobe density along the Y axis (downslope direction).
- (3) To express the relative differences in lobe density in a downslope direction, the ratio between each lobe's new centred Y coordinate and its next closest downslope neighbour (Y ratio) was plotted against distance downslope (Figure 5.6). Closer-spaced lobes have higher Y ratios, while increased spacing of lobes results in a lower Y ratio.

The end result is a one-dimensional spatial series showing the downslope coherence in lobe coverage. For example, the ratio between each lobe coordinate and the coordinate of its next nearest neighbour for a regular point pattern would be constant downslope. A heterogeneous point pattern would show zones of increased Y ratios, which express

concentrations of lobes, whereas troughs represent breaks in lobe frequency. Figure 5.6 clearly shows such breaks in lobe frequency, represented by decreased Y ratios, and thus quantitatively demonstrates the rhythmic nature of solifluction lobe on the slope. The wavelength of troughs broadly increases downslope, reaching a maximum of about 30 m in the lower part of the slope. This confirms that lobes become more dispersed downslope, independent of trends in their size. On the other hand, the downslope increase in peak wavelength is caused by the downslope increase in lobe size (length and

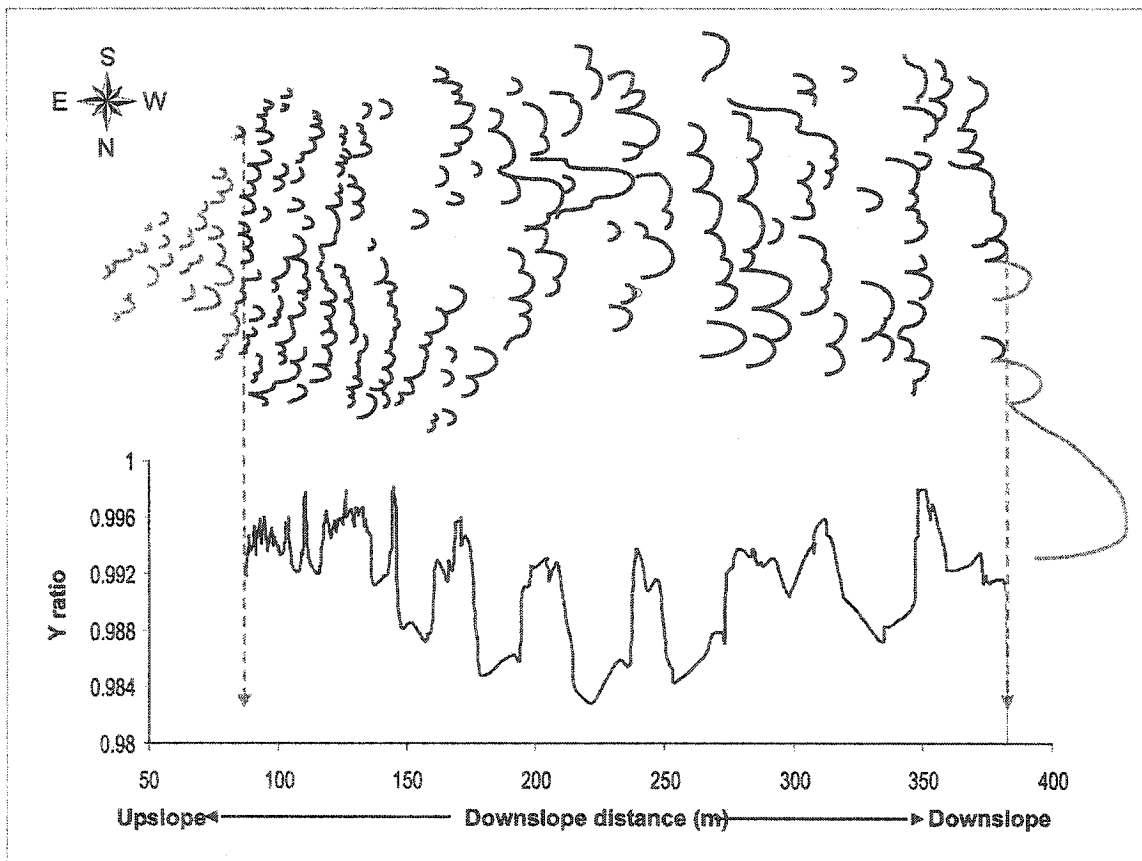


Figure 5.6: Downslope variation in lobe frequency. Some lobes from the upper slope (in grey) have been omitted from the analysis because mapping was inconsistent in that area, while three lobes on the lower slope and three lobes on the upper slope were omitted because they could not be included in the moving average.

width). Thus, the spatial structure of the lobe population on the study slope is one of decreasing lobe density, increasing lobe size and increasing dispersion downslope.

The rhythmic character is not as well defined on the upper slope as it is toward the slope bottom. This is because the resolution of the technique depends on the size of the window used for smoothing with the moving average. A smaller window would have yielded more details of the upper slope, while obscuring the larger-scale rhythmicity of the lower slope. A seven-lobe window was found to be the best compromise for obtaining a general view of the rhythmic nature of lobes on the slope.

This dispersion pattern, of increasing lobe size/decreasing lobe density in a downslope direction, suggests that the development of lobes on the slope responds to gradients in environmental conditions. This supports observations made by Hugenholtz and Lewkowicz (2002) on the same slope, where downslope trends in active layer depth, organic mat thickness and fines content were thought to cause this “solifluction lobe catena”. The authors emphasized the role of slopewash in inducing biophysical gradients down the slope, which ultimately lead to conditions favouring enhanced solifluction on lower portions of the slope and a corresponding increase in the size of the lobes. In particular, larger plan-form lobes were found to be significantly associated with a thick organic mat, a thin active layer and higher fines content. Their results, combined with the conclusions made in the present study on the spatial structure of lobe populations, form a good basis toward a more detailed explanation of solifluction lobe development on alpine slopes.

Spatial gradation of solifluction landforms was also reported in the Colorado Rockies by Benedict (1970) and in Great Britain by Ballantyne and Harris (1994). Matsuoka (2000) briefly discussed the occurrence of larger solifluction lobes at the bottom of slopes, which he described as zones of “retarded flow” caused by a reduced slope gradient. Solifluction lobe catenas were observed on other slopes in the vicinity of the research site (Figure 5.7), suggesting they may be widespread on alpine slopes in this area. Such a structure offers the possibility of building a model of slope movement. The phenomenon of a lobe catena leads to questions about lobe age and rates of advance in relation to environmental gradients on the slope, which when answered, might provide a physical basis for the spatial development of solifluction lobes on alpine slopes.

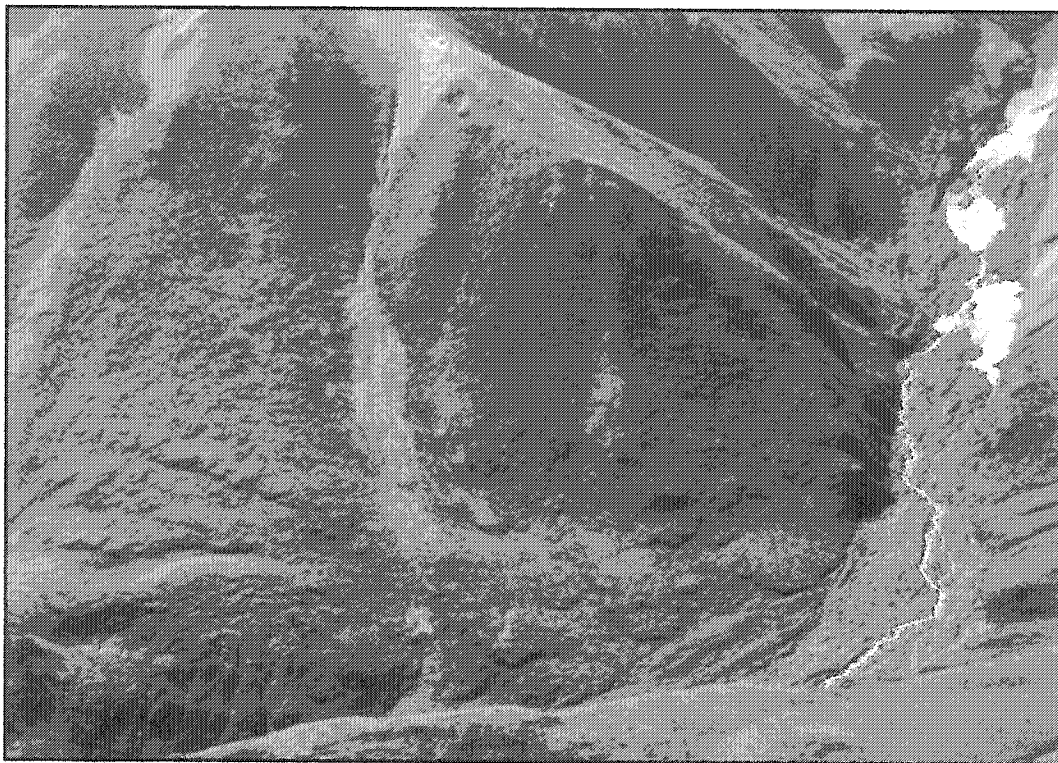


Figure 5.7: Example of solifluction lobe catena on a slope at the foot of Outpost Mountain 3 km east of the research site. Note the particularly clear contrasts in vegetation cover.

5.1.6.2. *Development of solifluction lobes*

A first idea of the development of individual solifluction lobes arises from the bivariate correlation analysis. The results point to an important fact: the morphology of lobes is consistent. The increase in the length of lobes, which occurs predominantly in a downslope direction, is accompanied by an increase in their corresponding width. Lobes conserve a constant plan form in space, which can be described by the length-to-width ratio, and the distribution of length-to-width ratios is normal, with a mean of 0.86. This value may represent an equilibrium state around which the lobe population gravitates. Variations in riser height and angle, on the other hand, are correlated with variations in the length-to-width ratio. Thus the morphology of the riser appears to indicate different stages of a recurrent cycle of lobe development.

The concept of a cyclic development of solifluction lobes was proposed by Price (1974) in the adjacent Ruby Range, Yukon, and was also discussed by Hugenholtz and Lewkowicz (2002). Price emphasized the role of snow redistribution in the lee of the riser of bigger lobes, which he thought, induced erosion of the front and collapse of the lobe. Hugenholtz and Lewkowicz (2002) proposed that continual movement downslope of the material on the tread causes a steepening of the slowly advancing riser, and this eventually leads to its instability, either resulting in a rapid advance of the front by shearing, or its collapse and destruction. In this regard, the height and steepness of the front may describe the state of stability of each lobe, where high and steep fronts represent unstable lobes about to collapse, while more gentle and subdued fronts represent collapsed forms at the beginning of an aggradational phase.

5.2. RATES OF SURFACE MOVEMENT

Rates of surface movement, measured on six solifluction lobes (see Figure 4.1), are presented and discussed here. Site J, on the middle of the slope, had to be abandoned because of obvious benchmark displacement. The surface displacements for two sites are presented in detail along with descriptions of the lobes, while data from the other sites are used to discuss spatial variability in surface rates. Since measurements were initiated after freezing of the soil occurred and just before thawing began, positive displacements represent the horizontal component of movement due to gelifluction (Figure 5.8). Negative (horizontal) displacements, on the other hand, represent retrograde movement from the vertical plane, which is the position of potential frost creep. A net negative

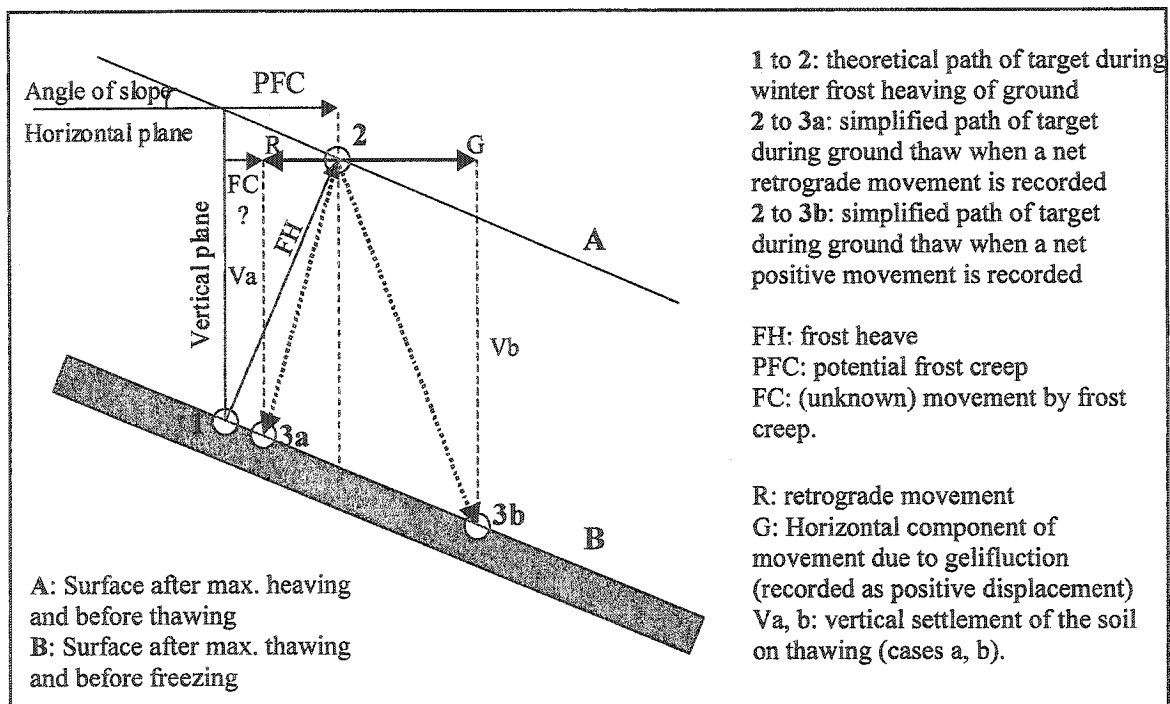


Figure 5.8: Vectors of horizontal and vertical movement measured by theodolite survey, and their relationship to frost creep, gelifluction and retrograde movement (Washburn, 1980).

displacement of the soil surface recorded by the theodolite does not necessarily mean that net forward movement by frost creep did not occur, but only that translational forward movement of the soil (gelifluction) did not occur on thawing.

5.2.1. Horizontal movement of surface targets

5.2.1.1. Site C

Site C is the 3rd site downslope along the sampling transect traversing the study slope (see Figure 4.1). It is located 60 m down from the slope crest (site A), in a relatively dry and poorly vegetated zone, with clast-rich surface material. Measurements of surface displacement were made on 15 targets along the long and transverse axes of the lobe. Lobe C is 4.7 m-long, 4.5 m-wide and has a 37 cm-high riser with an angle of 38°. The tread angle is 14.2° and the parent slope has angle of 18°. Displacement of targets on the long and transverse axes, between May 27th and July 13th 2002, are presented in Figure 5.9. The position of each target on lobe C is shown in Figure 5.10, together with its net cumulative displacement for the study period.

Figures 5.9 and 5.10 show considerable variability in rates of horizontal movement at the surface of lobe C. On the axial part of the lobe, the net cumulative movement for the study period tend to increase from the front of the lobe toward the back of the tread. On the top of the riser, and in the lee of the riser on the parent slope, net movements for the study period are negative (upslope displacement). Along the transverse axis, net downslope movements for the study period broadly increase from the right edge (looking downslope) where values are negative, to the left edge of the lobe, where target #10 indicates a particular high positive value (8.3 mm).

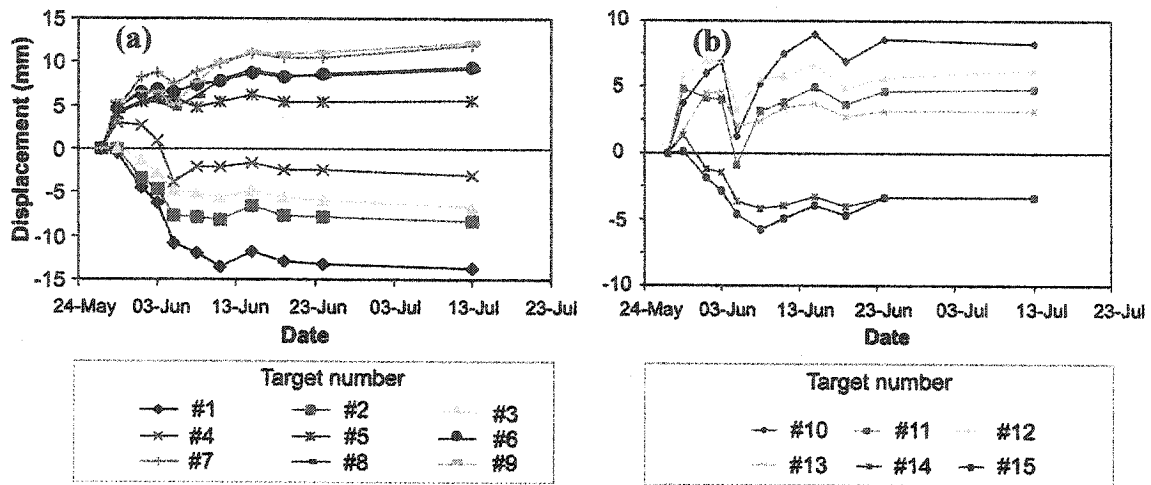


Figure 5.9: Horizontal movement of targets at the surface of lobe C: (a) Long axis; (b) Transverse axis. Error on each measurement is ± 0.5 mm.

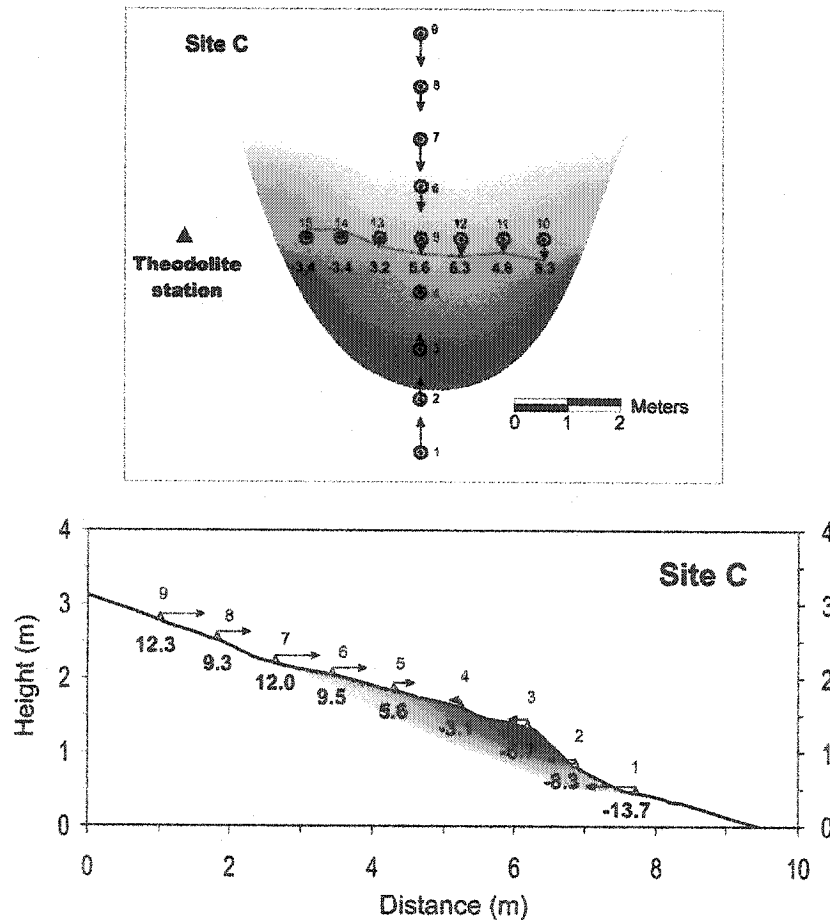


Figure 5.10: Distribution of horizontal movement on the surface of lobe C: (a) plan view; (b) profile view. Each vector of movement is 50 times the observed displacement (in mm, indicated in bold). Target number is indicated above each target.

In Figure 5.9, it can be seen that each target moved more rapidly in the first part of the monitoring period, from May 26th to mid-June, and then became more-or-less stable for the remaining period until July 13th. From June 24th to July 13th, measurements were interrupted, and so the apparent stability of the targets for that period may hide some unmeasured variations. However, the resulting net movement measured on July 13th suggests that the targets remained stable. This pattern of decreasing velocity, apparent on both the long and transverse axes, is sometimes disrupted, as for example targets #10, 11 and 12, which on June 5th displayed a strong negative inflection in surface movement. Despite this short-term variability of movement over time, and no matter the position of the target on the lobe surface or the nature of its movement (downslope or upslope), rates of surface displacement on lobe C appear to follow a broad pattern of gradual decrease in movement over time.

5.2.1.2. Site K

Site K is a large solifluction lobe located on the lower part of the slope, 266 m down from the slope crest (see Figure 4.1). This segment of the slope is wetter compared to the upper slope, and the vegetation mat is thick and well developed. Lobe K is 8 m-long, 10 m-wide and has a 70 cm-high riser with an angle of 35°. The tread angle is 14.8° and the surrounding parent slope is 16°. The riser is well-vegetated compared to the tread, which in places is bare of vegetation. The riser, overall, is bulging but at the very front, it shows signs of collapse which reduces the riser angle (Figure 5.11). Surface movement was evaluated from 19 targets placed on the surface of the lobe along the long and transverse axes. Movements of the targets between May 27th and July 13th 2002 are presented in Figure 5.12. The position of each target on lobe J is shown in Figure 5.13, along with



Figure 5.11: Lobe K, with bulging sides and collapsing riser at the front, near theodolite.

their net cumulative displacement for the study period. Lobe K exhibits a considerable variability in movement across its surface (Figures 5.11 and 5.12). At the very back of the lobe, near the break between the tread and the parent slope (target 10), net movement at the end of the monitoring period was very small (2.7 mm). From there, it increased downslope along the long axis of the lobe, and reached a maximum value of 56.2 mm in the middle of the tread, where the profile becomes convex downslope. From this point, movement decreased to 25.8 mm at the apex of the riser (target 3), and to 8.3 mm in the lee of the riser (target 1). Along the transverse axis, movement broadly decreased from the centre towards the edges, but the right edge (looking downslope) moved somewhat faster than the left. This asymmetry in movement across the lobe may be explained by

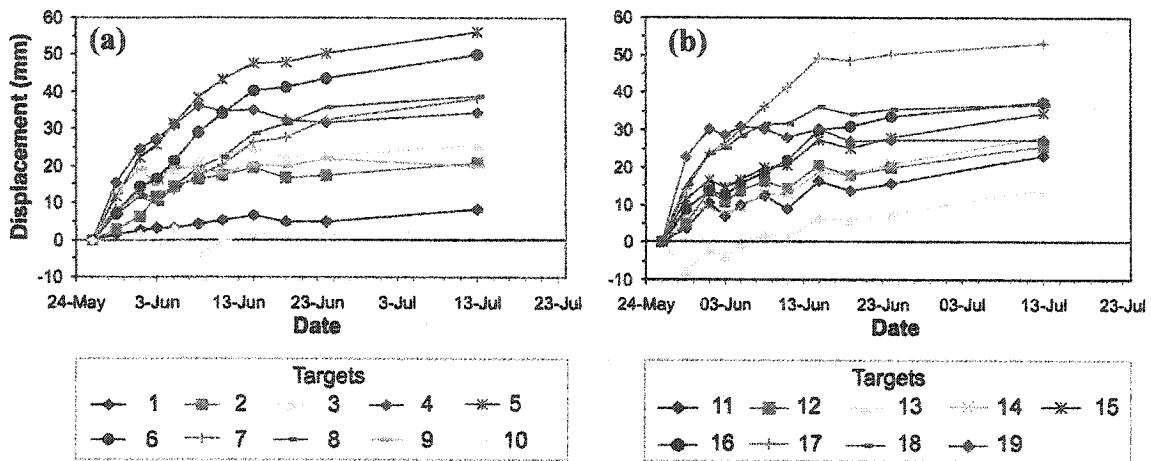


Figure 5.12: Horizontal movement of targets at the surface of lobe K: (a) long axis; (b) transverse axis. Error on each measurement is ± 0.5 mm.

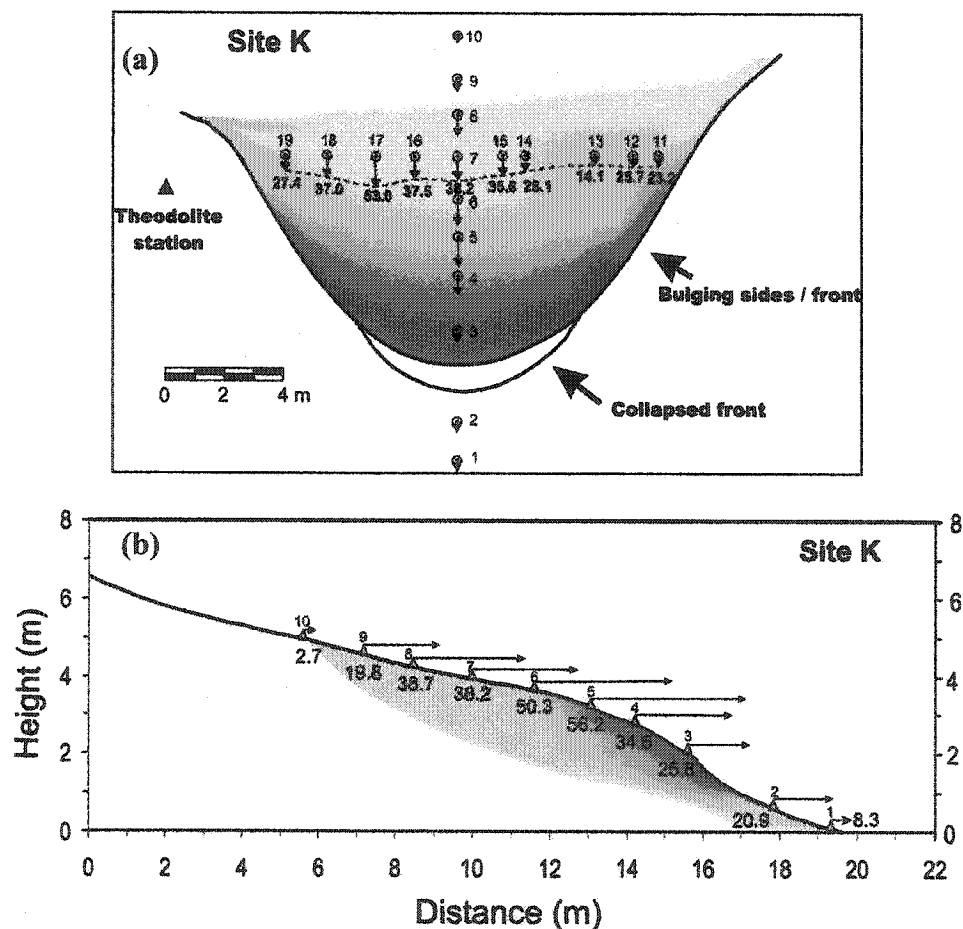


Figure 5.13: Distribution of horizontal movement on the surface of lobe K: (a) plan view; (b) profile view. Each vector of movement equals 40 times the observed displacement (in mm, indicated in bold). Target number is indicated above each target.

the fact that the left edge of the lobe (targets 11, 12 and 13) was much better vegetated than its right edge, which was almost bare of vegetation. Small non-sorted steps were present on the right-hand side of the lobe, suggesting that diurnal freeze-thaw action may have caused this zone of increase surface movement. On the left-hand side, the well-developed vegetation mat probably restrained any shallow movement of the soil caused by diurnal freeze-thaw cycles (e.g. Matsuoka, 1998a).

In regards to the timing of solifluction movement on the surface of lobe K, there is a clear pattern of a progressive deceleration of targets over time (Figure 5.12), on both the long and transverse axes. Despite marked variations in the amount of movement along and across the lobe surface, targets consistently moved faster in the first part of the study and then showed only slight movement from mid-June onward. In this latter period, some targets continued to move downslope at a much reduced rate, while others remained stable or moved slightly upslope. The pattern over time appears more coherent than for lobe C, on the upper slope.

5.2.2. Vertical movement of surface targets

The vertical displacement of targets over time was also calculated for lobes C and K, and values for the long axis of each lobe are presented in Figure 5.14. The pattern of thaw-settlement resembles that of horizontal movements, in that settlement is most pronounced in the first part of the thawing period and broadly decreases over time. For both lobes, vertical settlement of the soil was lower near the front than on the tread, where maximum values were found. Two targets (2 and 3) on the front of lobe C even showed some positive movement (heave), with a small net negative settlement by the end of the study.

Also apparent on Figure 5.14 is the large difference in the amount of vertical settlement between the two sites. The larger lobe K, located on the lower slope, exhibited net settlement values between 35 and 90 mm, while the smaller lobe C, located on the upper slope, displayed net values between 4 and 30 mm. This difference in the amount of thaw settlement between the two sites probably reflects variations in frost heave activity on the slope in the previous fall. The upper segment of the slope is probably drier than the lower segment, and ice segregation in the fall is therefore limited due to the lack of moisture in the active layer. The wetter lower slope is probably more prone to the formation of lenses of excess ice. Frost heaving of the soil surface and subsequent thaw-consolidation in the spring varies accordingly, creating the differences in soil settlement among lobes on the slope.

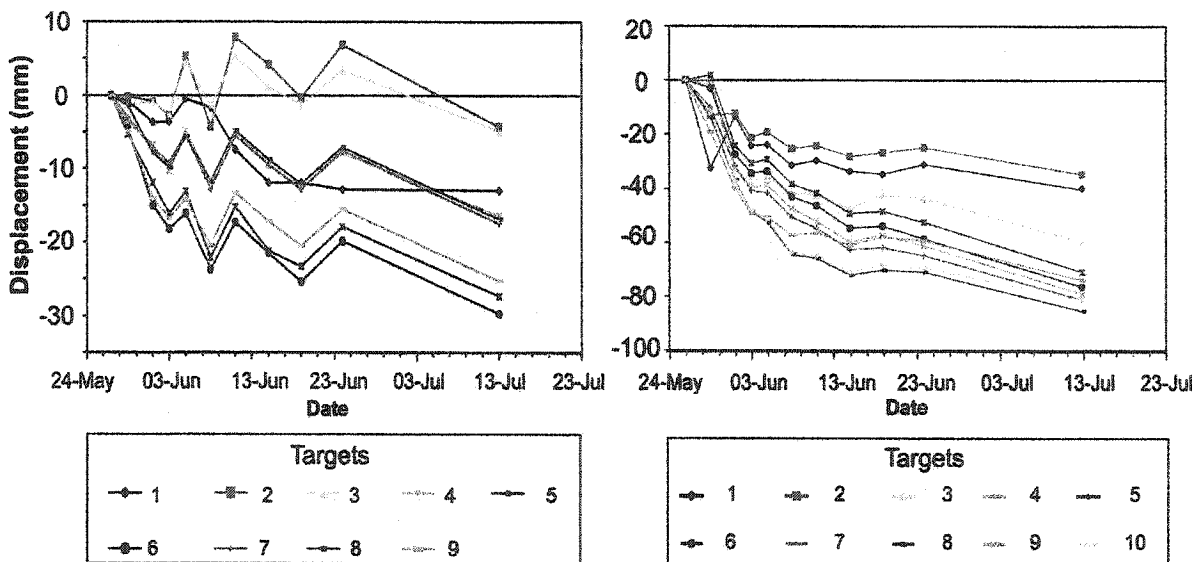


Figure 5.14: Vertical displacement of targets: (a) site C; (b) site K. Precision on each value is ± 2.5 mm.

5.2.3. Relationship between thaw settlement and solifluction movements

The apparent relationship between vertical and horizontal movements for site C and K was analysed using linear regression (Figure 5.15). A correlation coefficient of 0.84, significant at $p < 0.01$, was found between cumulative horizontal and vertical movements of targets as measured on July 13th on both lobes. Thus a significant positive linear relationship exists between the amount of horizontal movement and the amount of thaw-induced soil settlement. The slope of the regression line is significant at $p \leq 0.001$ but the y-intercept is statistically not different from 0 ($p = 0.064$). Significant variability occurs on the axial part of lobe K, where a high settlement value was obtained for target #10 on the back of the tread but only slight downslope movement was recorded. The slow rate of movement recorded for target 10 may be partly explained by the fact that the target was placed on a small patch of vegetation, which may have caused localized resistance to downslope movement. When this outlier is removed from the regression analysis,

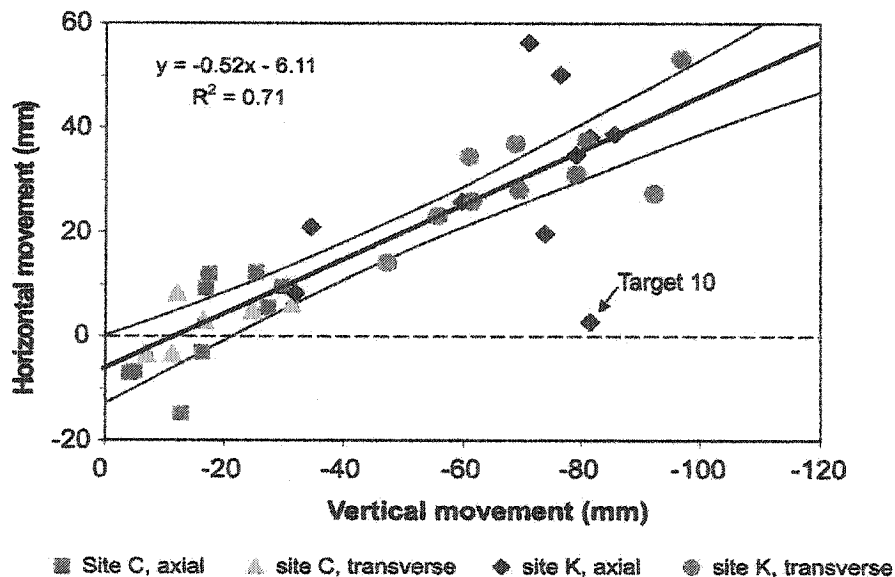


Figure 5.15: Scatterplot of net vertical movements against net horizontal movements of targets as recorded at the end of the study period on July 13th, 2002.

the R^2 value changes to 0.81, while the slope and y-intercept change to -0.56 and -7.00 , respectively, and the y-intercept becomes statistically different from 0 at $p \leq 0.05$.

A positive relationship between surface frost heave and downslope surface displacement, was previously identified by Harris et al. (1993, 1997). If thaw settlement is taken to reflect the reverse process of frost heave, the relationship presented in Figure 5.15 confirms that frost heave activity and subsequent thaw settlement have a significant influence on gelifluction movements. A higher concentration of ice in the soil must increase thaw settlement, which result in a proportional increase in downslope movement (gelifluction). Values of horizontal movement presented here, however, do not represent true gelifluction displacements, since the measured movement is not parallel to the soil surface, but they constitute a close approximation.

It is interesting to note that a threshold of thaw settlement seems to exist for gelifluction to occur. From the regression line, this threshold is approximately -12 mm. However, the maximum observed value without positive horizontal movement is -18 mm, and the maximum threshold for gelifluction to occur within the 95% confidence interval is -21 mm. With lower amounts of settlement, horizontal movement is upslope, and with greater amounts, movement is in the downslope direction. Thus a minimum amount of frost heave and subsequent thaw settlement may be necessary to disrupt the soil structure and induce gelifluction. When frost heave is limited, cohesion forces dominate on thawing and the soil tends to resettle normal to the slope. Horizontal movements are then recorded as backward movements relative to the vertical plane. This *retrograde*

movement was first identified by Washburn (1967) in Mesters Vig, Greenland. Washburn showed that apparent cohesion, due to capillary water in silts, produces a pressure normal to the slope and so causes departure from the vertical. In the present case, since only the thawing portion of movement was measured, the net downslope movement of targets resulting from a complete freeze-thaw cycle is unknown, but is situated somewhere between 0 and the value for potential frost creep (see Figure 5.8). When ice segregation causes greater amounts of frost heave, the soil has a higher potential for thaw-consolidation and related gelifluction movements, which are linearly related to the amount of settlement taking place on thawing. Variations occur, however, which must be explained in part by differences in the rate at which a given soil layer consolidates (thaws). For a given soil layer with a specific ice content, a faster rate of thaw is likely to cause faster downslope movement.

If we consider that soil settlement on thawing approximates the amount of frost heave that occurred during winter, an average heaving ratio can be obtained for the active layer at the end of the study period. Both sites had a thawed layer with a depth of around 100 cm on July 13th. The heaving ratio (e.g. Harris, 1981) is then the ratio between the amount of heave (or here, settlement) and the thickness of the frozen (or thawed) layer (100 cm). If we take -21 mm as a maximum threshold value for not triggering gelifluction, this gives a mean heaving ratio of ≈ 0.02 for the thawed layer. This value is much less than the range of 0.05 – 0.50 obtained for various soil depths in laboratory experiments by Harris et al. (1996, 1997). In their experiment, increased heaving ratios, which corresponded to zones of higher ice content, led to more consolidation and

increased downslope movement. Using data from Harris et al. (1997), it is possible to estimate the threshold heaving ratio for gelifluction to occur, by plotting the mean heave ratio against downslope surface displacement for different freeze-thaw cycles. Threshold values of 0.133 and 0.112, corresponding to zero downslope movement, are found from the regression line for their "Vire" and "Feuguerolles" soils, respectively.

Data on surface frost heave and related depth of freezing were also summarized by Benedict (1976) for a variety of locations and freezing regimes (deep/shallow, seasonal/permafrost). Despite the different freezing regimes, surface frost heave was seen to increase systematically with freezing depth. The data permitted the calculation of heaving ratios, which vary between 0.04 and 0.18 (mean = 0.09). For both lobes C and K, settlement ratios for individual measurement points are in the lower part of this range, varying between 0.004 and 0.100 (mean = 0.05). The threshold heaving ratio of 0.02 calculated empirically in this study is also lower than the minimum heaving ratio of 0.04 reported in the field by Benedict (1976). Thus the high heaving ratios and corresponding large downslope surface movements recorded in laboratory experiments by and Harris et al. (1996, 1997) differ from values measured in the field. The more intense frost heave activity and high related downslope movement probably resulted from unlimited water supply to their experimental slopes during the freezing period. In permafrost environments, water supply to the freezing front may decrease and even stop at a certain depth, which causes more pronounced frost heave near the surface, and consolidation of the lower soil because of water depletion (e.g. Smith, 1985). The experiments of Harris

et al. (1996, 1997) simulated a seasonally frozen site with a steady water supply, which probably maximized the resulting frost heave activity.

The excess water content corresponding to settlement/frost heaving ratio can be calculated from:

$$W_{ex} = \gamma_i H \quad (5.3)$$

Where W_{ex} is the excess water content, expressed as a percentage of the thickness of the frozen (or thawed) layer, γ_i is the density of ice (0.90 g/cm^3) and H is the heaving ratio. Thus a heaving ratio of 0.02 gives an excess water content of 1.8%, while the heaving ratio calculated from the maximum observed consolidation (-90 mm, lobe K) gives an excess water content of 8%. According to Figure 5.15, gelifluction movements increase from 0 to 40 mm over a range of about 6% excess volumetric water content. Below a critical excess volumetric water content of approximately 2%, cohesion forces dominate and retrograde movement occurs.

The increase in gelifluction movement in response to higher excess water content reflects a loss of frictional strength (e.g. Harris, 1995, 1997; Terzaghi et al., 1996). Between June 1st and June 24th 2002, undrained shear strength and moisture content were measured on the surface of small non-sorted steps near camp. Shear strength was measured with a *Geonor* hand shear vane, and moisture content was measured with a Theta Moisture probe which relates the dielectric signal of the soil to its volumetric water content. Moisture content had to be determined gravimetrically for some samples because of failure of the Theta probe. The plot of log yield stress (undrained shear strength) against

moisture content is presented in Figure 5.16. Despite considerable scatter, there is a trend of decreasing yield stress with increasing moisture content. The relationship is similar to that obtained by Harris et al. (1995, 1997). Part of the scatter in the data is attributable to the fact that an average value for soil density was used to convert volumetric water content into gravimetric values. Figure 5.16 indicates that an increase in water content in the thawing soil reduces the effective stress and causes a corresponding decrease in shear strength. Lower shear strength, in turn, promotes gelifluction movements in the thawing soil.

These results confirm the conclusions reached by Rein and Burrous (1980), who demonstrated experimentally that the intensity and distribution of shear strain in a thawing soil is controlled primarily by the depth and thickness of excess ice, and hence by the excess water content. Thus, as stated by Matsuoka (2001), the depth and thickness of ice lenses and freeze-thaw frequency are the major controls on spatial and temporal variations in solifluction rates.

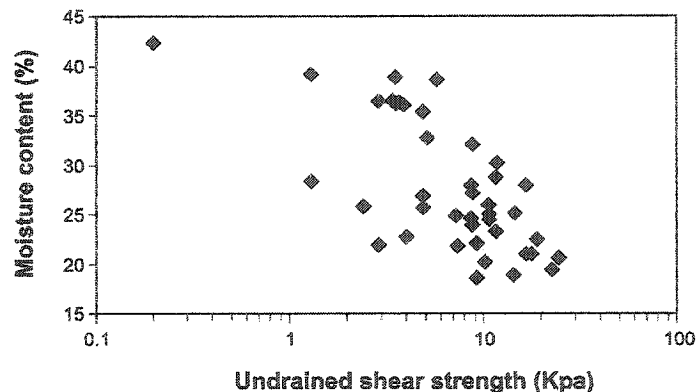


Figure 5.16: Scatterplot of undrained shear strength measured in non-sorted steps against corresponding gravimetric water content. Each value of shear strength and moisture content is an average of five different measurements.

5.2.4. Distribution of movement on the surface of solifluction lobes

The distribution of net horizontal movements on the surface of lobe K resembles that of lobe C. The broad pattern is one of reduced surface movement near the frontal part of the lobe, with a higher velocity on the tread. In both cases, the parent slope immediately downslope from the riser also shows relatively slow movement. Both lobes exhibit a break of slope between the back of tread and the parent slope, which apparently shows a zone of reduced movement. The distribution of movement on the surface of the other lobes monitored for target movement (see Figure 4.1) was similar to that of the two sites presented here. Figure 5.17 summarizes the average horizontal movement at the surface of the five studied lobes. Each lobe was divided into four axial zones and five transverse zones, and the average horizontal movement of the two or three targets (depending on the number of targets used on a particular lobe) was plotted. To allow comparison of the movements from different lobes, the percentage of the maximum recorded movement was used instead of absolute values. Although the distribution of movement varied among the lobes, on average movement increases from the front backwards and decreases from the centre to the edges.

This broad pattern of a fast-advancing tread with a relatively slower front is similar to that reported by Washburn (1967) in Greenland, by Benedict (1970) in the Colorado Front Range, by Smith (1988) in the Canadian Rockies, and more recently by Rapp and Åkerman (1993) in Northern Sweden and Gengnian et al. (1995) in China. It confirms that the frontal banks of lobes result from variations in the rate of solifluction across the

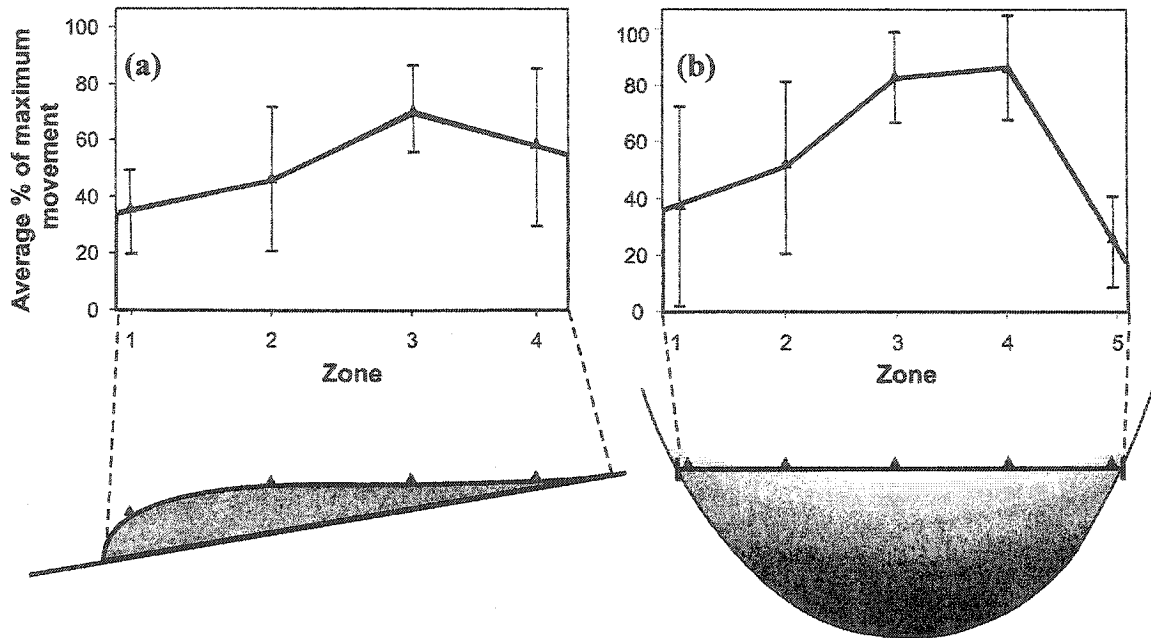


Figure 5.17: Distribution of average horizontal movements measured at the surface of five solifluction lobes: (a) along the axial part of the lobe; (b) across its surface. Movement is in percent of maximum observed movement with one standard-deviation range.

slope, material from above tending to accumulate in the zones of slower soil movement (Harris, 1981).

The distribution of vertical movements (net amounts of thaw settlement) at the surface of lobes C and K is presented for the axial part and transverse axes of both lobes in Figure 5.18. As mentioned above, total thaw settlement of soil by the end of the study period, on July 13th 2002, was low near the riser and increased gradually back along the tread (Figure 5.18a, b). Settlement also decreased momentarily near the break-in-slope at the back of the tread, but more clearly for lobe C (-6 m back) than for lobe K (-12 m back).

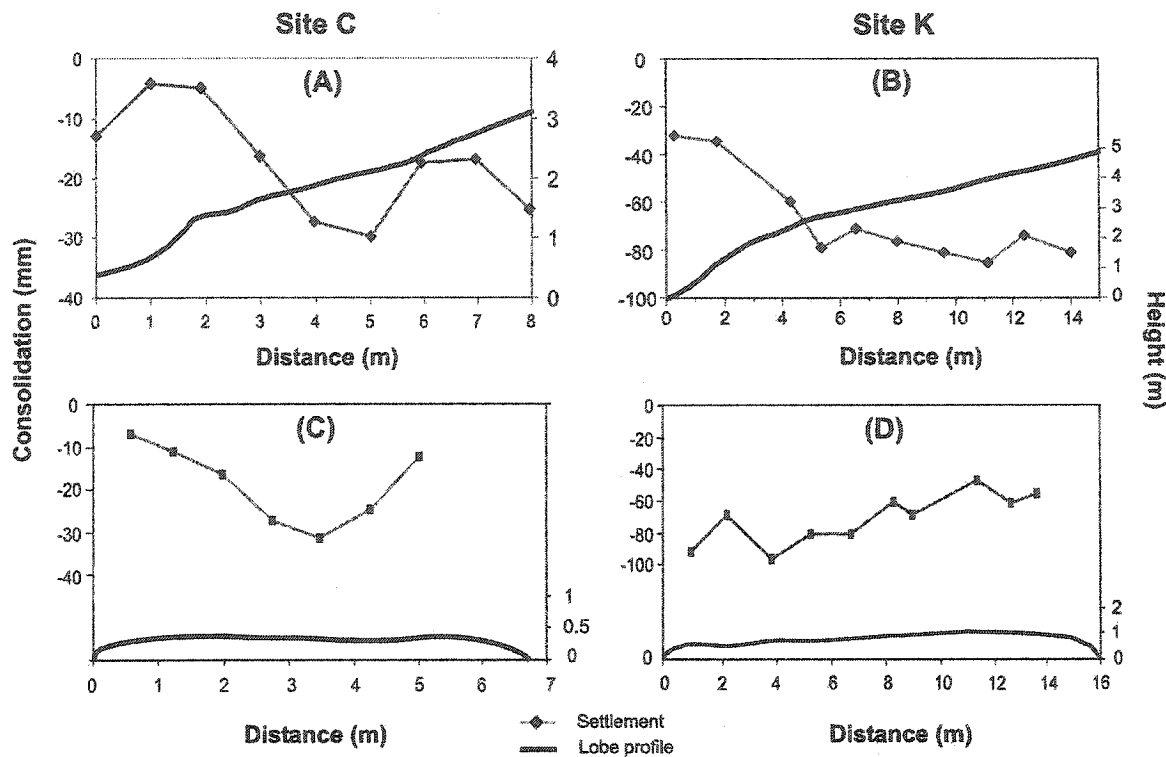


Figure 5.18: Distribution of net vertical movements at the surface of lobes C and K, as recorded on July 13th, 2002: (a) site C, long axis; (b) site K, long axis; (c) site C, transverse axis; (d) site K, transverse axis.

Across the lobe, settlement clearly decreased from the sides toward the centre of the lobe at site C (Figure 5.18c). The only trend at site K was a linear increase from the left to the right side of the lobe (Figure 5.18d). It is interesting to note that the cross-profile of lobe K is steeper on its right side, and that settlement decreased accordingly: targets in this area moved much slowly than those on the left side, which showed pronounced thaw settlement. The amount of thaw settlement, therefore, appears to vary with the steepness of the lobe profile, and steeper fronts and edges show low settlement relative to the flatter tread surface. Surface horizontal displacements follow the same pattern with reduced movements where soil settlement is less pronounced. These results confirm findings of from the Colorado Front Range, where frost heave and subsequent thaw settlement were

found to increase from the edges of a lobe toward its centre (Benedict, 1970, 1976). Accelerated downslope movement was also measured in zones of increase frost heave activity. Benedict concluded that moisture redistribution across the lobe during autumn freeze-back was responsible for these variations in vertical and horizontal movements.

5.2.5. Timing of surface movement

Horizontal and vertical displacements of targets followed a fairly consistent temporal pattern over the course of the study period. Targets tended to move faster at the beginning of ground thaw in both the horizontal and vertical planes, and then stabilized over time. The amount of surface movement, however, varied across the lobe surface, as demonstrated above.

To obtain a clearer picture of the timing of movement for the different targets, the percentage of maximum target movement was plotted against time for lobes C and K (Figure 5.19). The pattern appears more consistent for site K, especially in the vertical directions. The more variable pattern of vertical movement for lobe C may be partly explained by the fact that much smaller movements were recorded, and that some of these variations may reflect measurement errors, since the precision was only ± 2.5 mm. Diurnal freeze-thaw cycles may also be responsible for some of these variations, which appear amplified when presented as a percentage of maximum movement.

A pattern of decreasing rate of downslope movement over time was reported by Washburn (1967) in Greenland. In his study, a large number of targets were surveyed several times per year in a similar fashion to the present study. The amount of movement

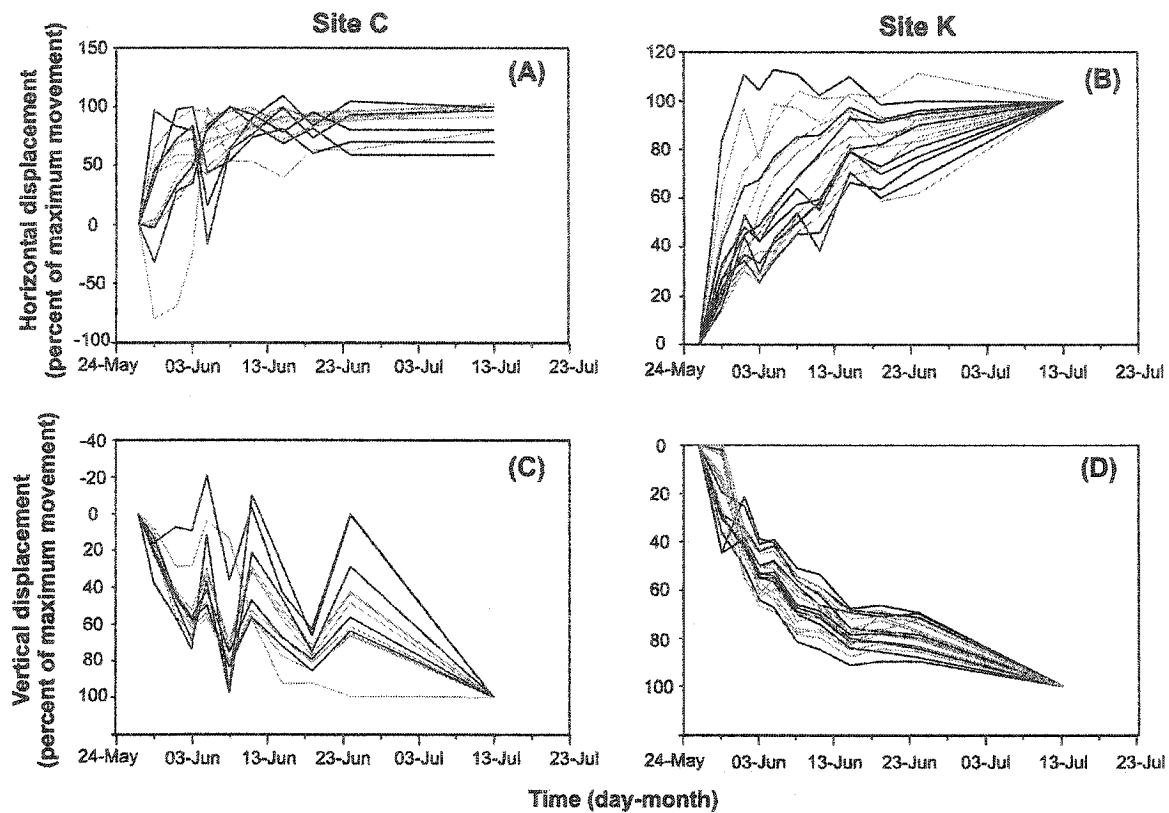


Figure 5.19: Timing of target movement at the surface of lobes C and K: (a) Site C, horizontal movement; (b) site K, horizontal; (c) site C, vertical; (d) site K, vertical. Displacement values are percent of maximum movement for each target. Grey lines represent targets on the long axis and black lines represent targets on the transverse axis.

measured varied between the targets, depending on the moisture of the surrounding soil and the slope gradient, but the targets moved rapidly at first and then slowed progressively before stabilizing. This pattern was not constant everywhere however, and some targets began to move downslope at first and then moved upslope for the remaining time.

In their experimental work on solifluction processes, Harris et al. (1993, 1996, 1997) used linear voltage displacement transformers to monitor frost heave and downslope soil displacement. During freezing of the soil, frost heave caused the soil to rise

perpendicularly to the surface. The pattern of movement was one of decreasing heaving rate with time, until freezing ceased and thawing of the ground began. Then, downslope movement parallel to the surface occurred, concurrently with thaw settlement perpendicular to the surface. The pattern of downslope movement involved a decreasing rate over time, and appeared directly correlated with thaw settlement, which also decreased over time. Repetitive measures of frost heave and thaw settlement are common in the literature. Studies by Penner (1970), Benedict (1970), Fahey (1974, 1979), Smith (1985) and Matsuoka (1998b) all present continuous measurements over time of frost heave and thaw settlement. These studies have shown how frost heave decreases over the course of the freezing period until thawing begins and settlement of the soil occurs, almost always in a decreasing fashion over time. Although temporal patterns of frost heave and thaw settlement have been reported frequently, they have rarely been discussed in relation to solifluction movements (but see Harris et al., 1996). The relation between downslope and vertical movements of the soil surface offers the possibility of linking gelifluction rates to rates of thaw-settlement, and to the thermal regime of the soil and related climatic control.

5.2.6. Spatial variability in lobe surface movement

Measurements of movement rates on different lobes allow an evaluation of their spatial variability in relation to the location of each lobe on the slope and relative to their morphology. A value for cumulative surface movement was calculated for each of the five lobes monitored, by taking the mean of all target displacements by the end of the study period. These values were plotted against tread length in order to check for relationships between solifluction activity and lobe morphology (Figure 5.20a). The

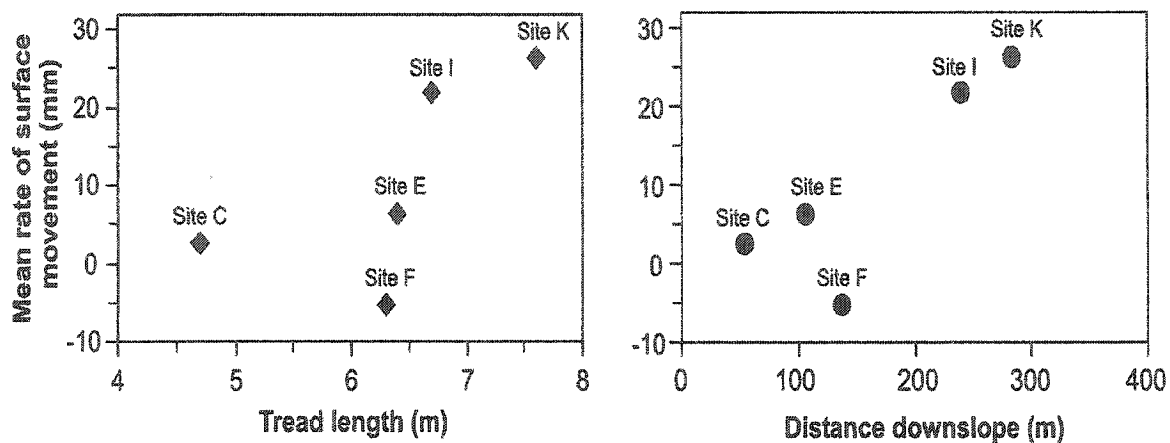


Figure 5.20: Spatial variability in cumulative rates of movement measured at the surface of five solifluction lobes: (a) scatter graph of mean rate of movement against tread length; (b) scatter graph of mean rate of movement against distance downslope.

graph shows an increase in mean surface rate with increasing tread length. Site F, in the middle of the slope, however, falls below this trend, with a mean value of about -5 mm. All but one of the targets at Site F moved upslope, and no obvious reason could be found for this, apart from the possibility of benchmark perturbation. Because of this outlier and the restricted number of points, the relationship is not statistically significant, but nonetheless suggests a general increase in surface movement at the larger lobes. Because the size of the lobes increases downslope, the mean rate of surface movement shows a corresponding increase in a downslope direction (Figure 5.20b). Once again, site F does not follow this apparent trend.

These results offer a first explanation of the spatial distribution of lobes on the study slope. Factors favouring the process of solifluction must increase downslope to cause accelerated slope movement and an increase in the corresponding size of the lobes.

A downslope increase in solifluction rates was also reported by Rapp (1960) in northern Scandinavia. The author measured the deformation of painted lines of boulders and of pegs fixed onto the soil surface. Average annual movements were seen to increase downslope from 1 to 22 cm per year, despite the reduction in gradient downslope. On a larger scale, Smith (1987b) also found a negative correlation between surface movements and altitude in the Canadian Rockies. In more recent laboratory experiments by Coutard et al. (1988) and Calmels and Coutard (2000), movement of material by solifluction was found to increase down an experimental slope, in response to increased ice segregation and greater frost heave activity. Coutard et al. (1988) concluded that the mechanism of frost heave can differ depending on temperature gradients and drainage conditions along the slope. Because the water supply was effective longer in the lower part of their experimental slope, ice segregation went deeper, and the total heave measured was greater in the middle and lower parts of the slope. Downslope movement recorded in these locations was more intense, both at depth and at the surface.

The distribution of movement measured at the surface of solifluction lobes on the study slope in Kluane is similar to the results described above. The relationship put forward between thaw settlement and resulting surface movement suggests that a control mechanism similar to that identified by Coutard et al. (1988) may be responsible for the downslope increase in solifluction rates and concurrent lobe size. Moisture distribution along the slope appears as the most probable explanation for the observed gradation in rates and landform size downslope. It is likely that the active layer drains progressively toward the bottom of the slope throughout the summer, inducing a spatial gradient in soil

moisture. The upper slope then becomes increasingly depleted of water relative to the lower slope where the water-table is typically closer to the surface (Woo and Steer, 1983; Hinzman et al., 1993). This in turn induces greater frost heave downslope, larger thaw-consolidation and enhanced downslope movement.

5.2.7. Synthesis

Movement of targets placed at the surface of five solifluction lobes along a transect running down the study slope showed a consistent pattern over time in both the vertical and horizontal plane. Whether translational movement is in the downslope direction (*gelifluction*) or upslope from the vertical (*retrograde*), it was more rapid during the initial thawing phase and decreased over time. The pattern is similar in the vertical plane (settlement), with targets tending to settle more rapidly early in the thawing period, and a subsequent decrease of movement over time. Despite intra- and inter-lobes variation in absolute displacements, the targets exhibit a consistent behaviour over time when the displacement data is transformed into percentages of maximum observed movement. This general pattern of a decreasing rate of movement over time was reported in a field study by Washburn (1967) in Greenland, but it is only recently that more sophisticated techniques have permitted continuous monitoring of intra-annual slope movement. Most of these studies, in the field or in the laboratory, have shown similar patterns of thaw settlement and surface movement over time. The relationship identified between thaw-settlement and the nature and amount of resulting downslope movement appears promising for modelling purposes. Gelifluction movements may be triggered by a threshold in ice segregation, and may increase linearly with thaw-settlement. Groundwater distribution along the slope appears as the most probable cause for inducing

downslope variations in ice segregation, which controls the amplitude of the annual cycle of surface heave and settlement. This could explain the observed increase in surface rates downslope, and the corresponding enlargement of solifluction lobes, as well as confirming the experimental results by Coutard et al. (1988). However, before drawing further conclusions in this direction, additional information is needed on environmental characteristics on the slope.

In regard to the development of solifluction lobes, target movement was shown to increase systematically backward from the front, and from the edges toward the centre. This radial pattern of surface velocity has been reported previously in the literature, and confirms that the existence of lobes on the slope must be directly related to these variations in movement across the slope. The distribution of solifluction movements on the surface of lobes suggests that these forms are active and will become unstable if the movement distribution is maintained over time. The faster advancing tread must inevitably result in a thickening of the riser, which moves at a much slower rate. The questions of landform evolution over time then arises. Solifluction lobes with steep overhanging risers were observed in the field, as well as others with more subdued or collapsed fronts. Despite the fact that lobes conserve a more-or-less consistent plan form, variations occur in their length-to-width ratios, which are correlated with changes in the height and steepness of the lobe front. Thus the distribution of surface rates observed on lobes must lead to a state of maximum instability, with stress being concentrated in the frontal part of the lobe. This critical state appears to correspond to lobes with steep fronts which have become larger than normal compared to their length (low length-to-width

ratio). Beyond that critical evolution stage, the front of the lobe likely deforms more rapidly to relieve the accumulated stress.

Thus it is inferred here that the distribution of solifluction movements at the surface of lobes reflects the key mechanism in their formation and development over time and space. Variations in solifluction rates on the slope induce zones of retarded movement, behind which material slowly accumulates to form lobate outcrops. As the movement becomes increasingly differential across this zone, a solifluction lobe grows toward a state of instability. What form this instability manifests remains an open question: a slow caterpillar-like advance of the front (Benedict, 1970; Elliot, 1996b), rapid advance by shearing (Alexander and Price, 1980; Matthews et al., 1986), or complete collapse and final destruction of the lobe (Price, 1974). Although some completely collapsed lobes were observed on the slope, they were rare. On the other hand, the variation in frontal morphology together with the stability of the plan form points towards a cyclic mechanism of lobe advance.

5.3. LONG-TERM RATES OF SOLIFLUCTION LOBE MOVEMENT

Radiocarbon dates were obtained for 12 of the 13 solifluction lobes excavated on the study slope. Figure 5.21 presents the surface profiles of the lobes, ordered by position from the top to the base of the slope (see Figure 4.1), and the 2 m-long sections that were excavated back from their fronts. The internal stratigraphy of each lobe, conventional ^{14}C dates for humus samples collected from buried organic horizons and the main morphometric parameters for each lobe are summarized. The presentation below

examines stratigraphic observations made on each excavated lobe. The sedimentological properties of lobe sediments are then presented and discussed, and finally, radiocarbon date are interpreted in terms of rates and timing of solifluction lobe movement.

5.3.1. Morphology and stratigraphy of excavated lobes

5.3.1.1. Lobe profiles

Each lobe studied becomes steeper towards the front, but the angles of the front and of the tread surface vary among the lobes. All lobes begin at their upslope end with a break-in-slope which marks the upper limit of the tread. The magnitude of this break-in-slope varies, with some lobes having a tread angle steeper than the parent slope (e.g. Figures 5.21b, c: lobes #4 and 5), while others exhibit a marked decrease in gradient from parent slope to tread (e.g. Figures 5.21b: lobes #1 and 19). In some cases, a small hollow is present on the tread just upslope of where the profile steepens at the front (e.g. Figure 5.21d: lobe #15).

The morphology of the front also varies: certain lobes have low-angled risers (e.g. Figures 5.21b, d: lobes #4 and 10; Figure 5.22b) while others have steep and well-defined risers (e.g. Figures 5.21a, b, c: lobes #1, 2, 7 and 19; Figure 5.22a). The boundary between the parent slope and the riser is often difficult to outline for risers with lower angles, while steeper risers represent an obvious line between the lobe and the surrounding slope. Between these two extremes, many variations exist: some lobes with steep risers display signs of collapse near the base of the lobe at the intersection with the parent slope (e.g. Figure 5.21a, b: lobes #1, 17 and 4; Figure 5.22d), where material seems to have moved more rapidly. Other lobes possess multiple-fronted risers, which

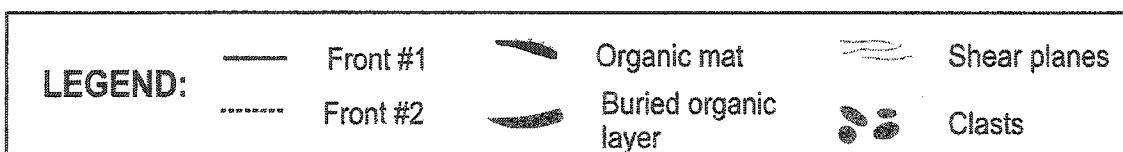
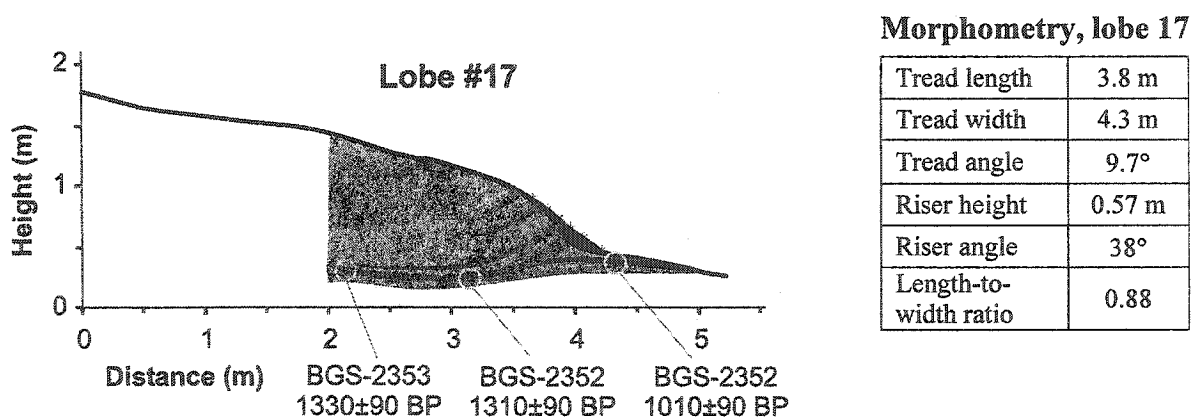
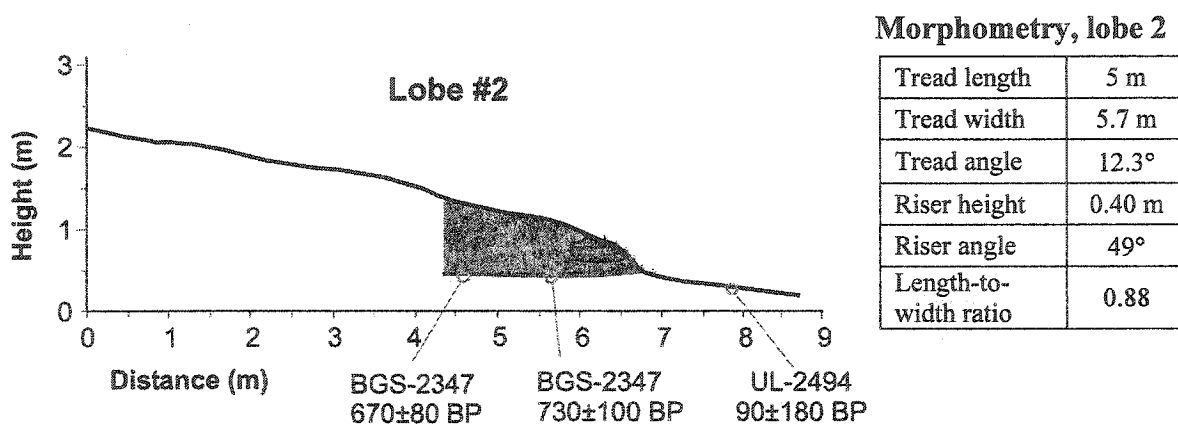
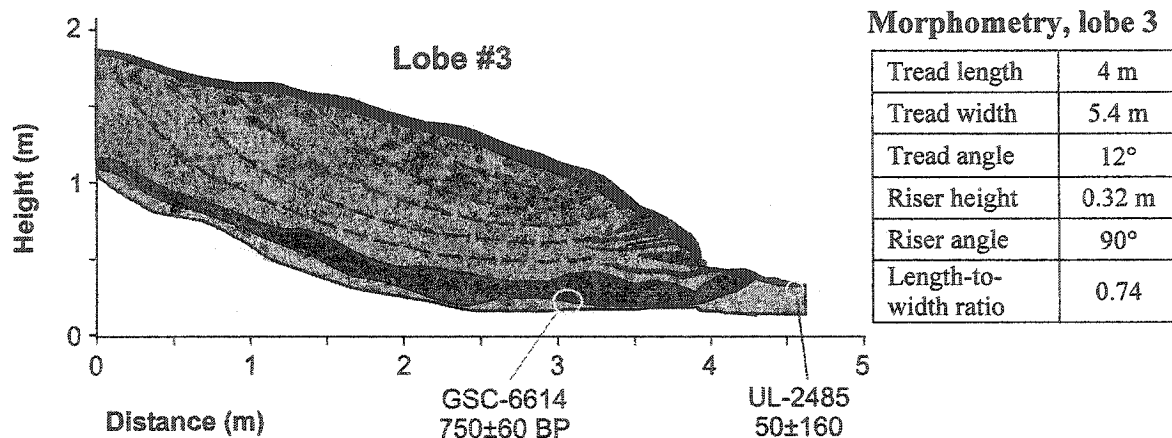
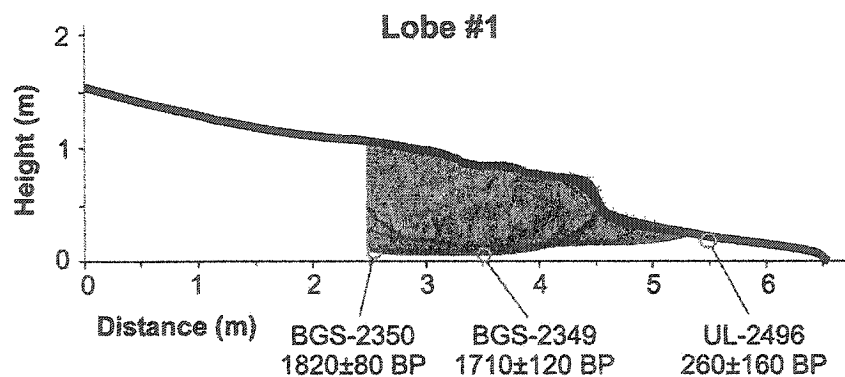
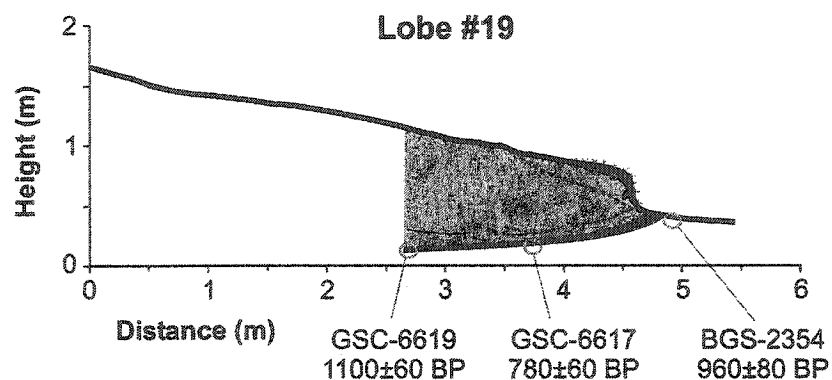


Figure 5.21a: Diagrams of dated lobes. The shaded areas represent the excavated sections. Note: all lobes have a 1X vertical exaggeration, but scales vary.



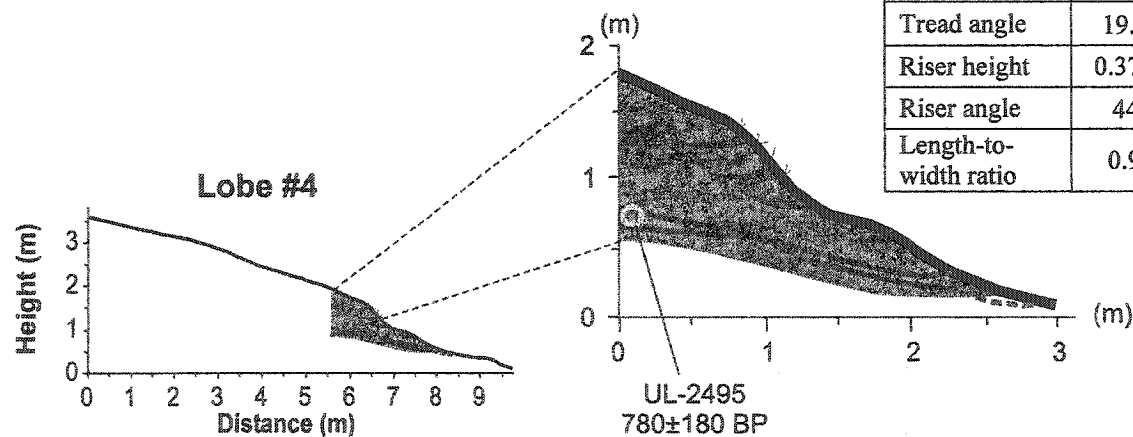
Morphometry, lobe 1

Tread length	3.3 m
Tread width	4 m
Tread angle	9.7°
Riser height	0.27 m
Riser angle	44°
Length-to-width ratio	0.83



Morphometry, lobe 19

Tread length	3.3 m
Tread width	4.6 m
Tread angle	8.3°
Riser height	0.33 m
Riser angle	72°
Length-to-width ratio	0.72



Morphometry, lobe 4

Tread length	8.0 m
Tread width	9.2 m
Tread angle	19.9°
Riser height	0.37 m
Riser angle	44°
Length-to-width ratio	0.92

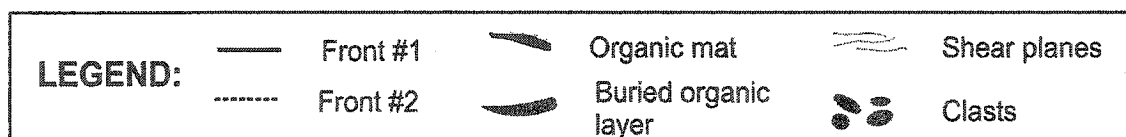


Figure 5.21b: Diagrams of dated lobes. The shaded areas represent the excavated sections. Note: all lobes have a 1x vertical exaggeration, but scales vary.

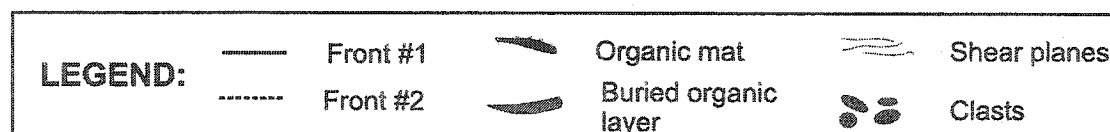
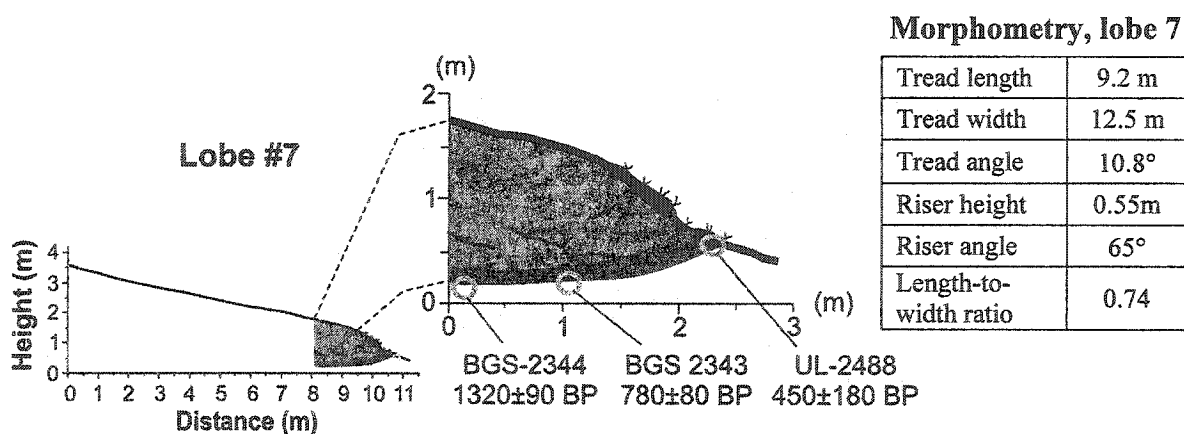
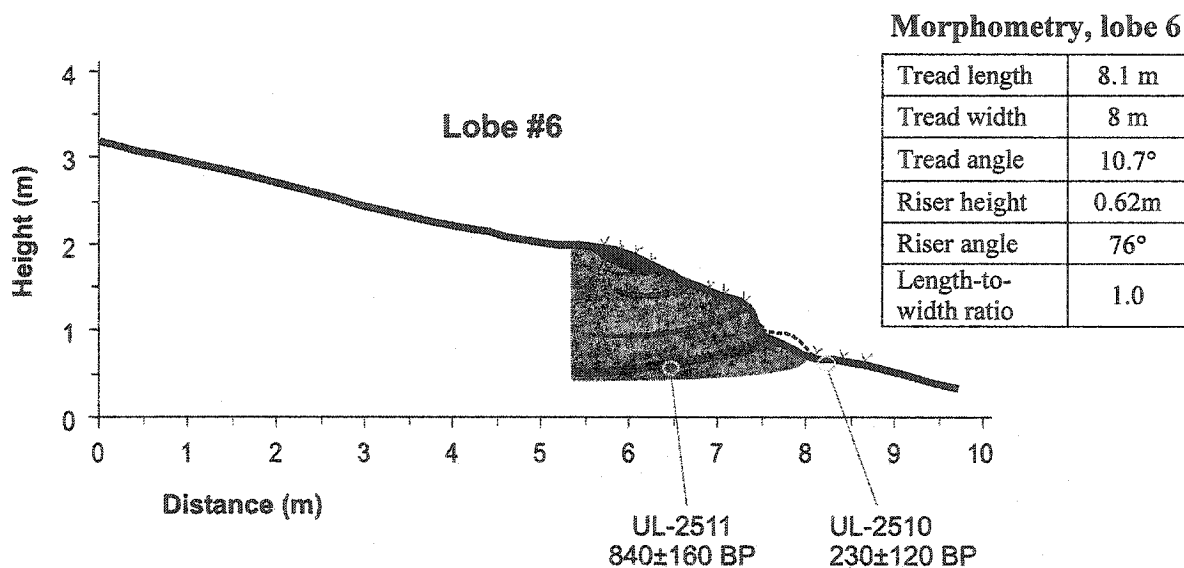
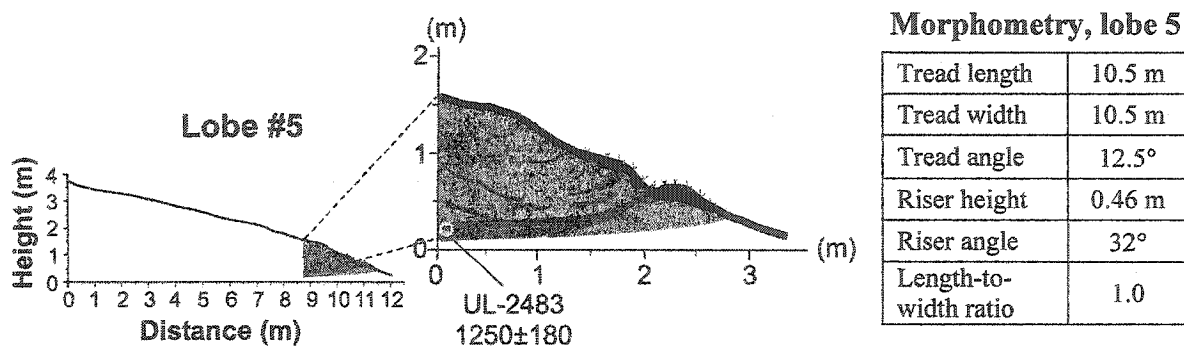


Figure 5.21c: Diagrams of dated lobes. The shaded areas represent the excavated sections. Note: all lobes have a 1x vertical exaggeration, but scales vary.

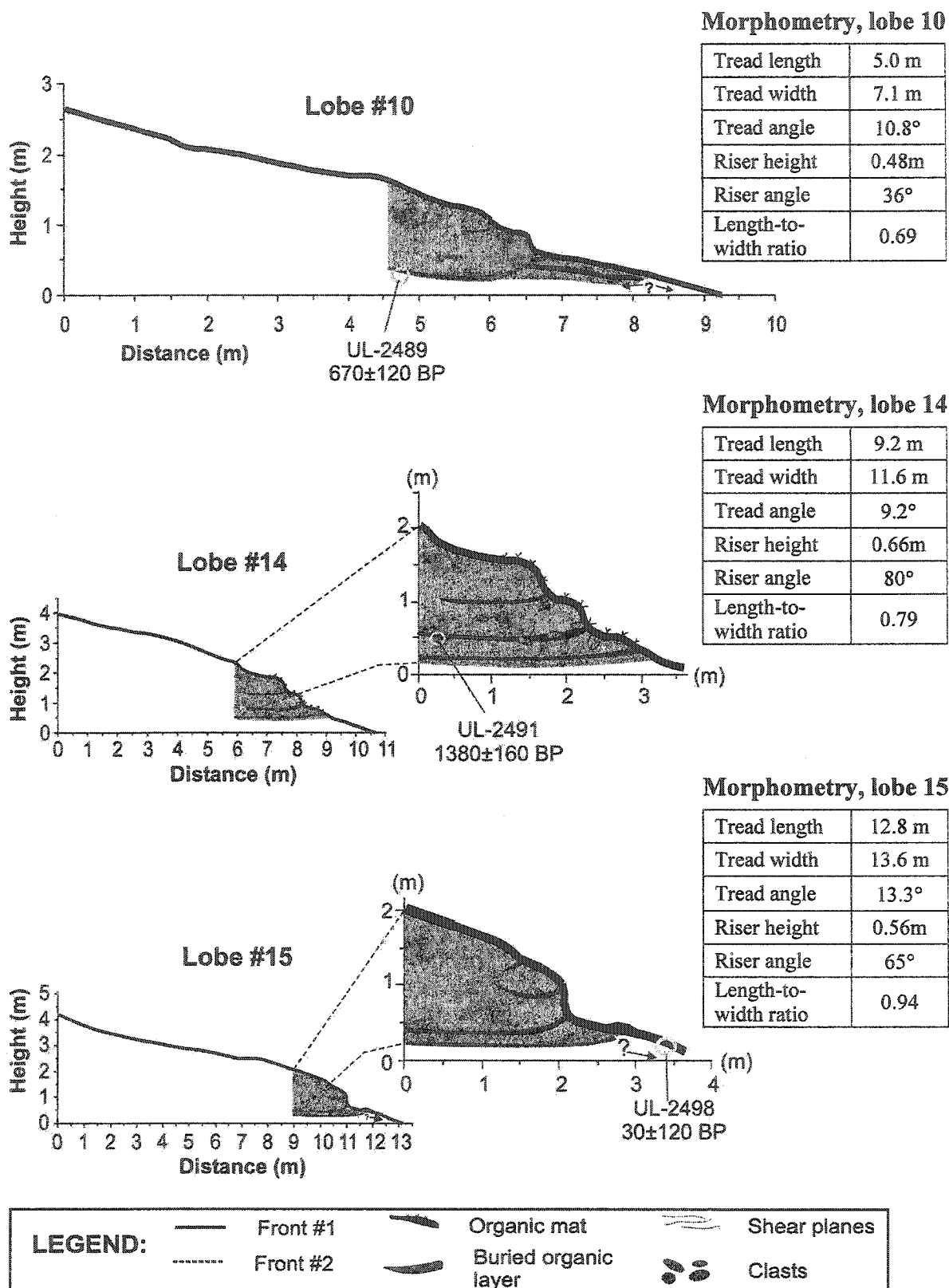


Figure 5.21d: Diagrams of dated lobes. The shaded areas represent the excavated sections. Note: all lobes have a 1x vertical exaggeration, but scales vary.

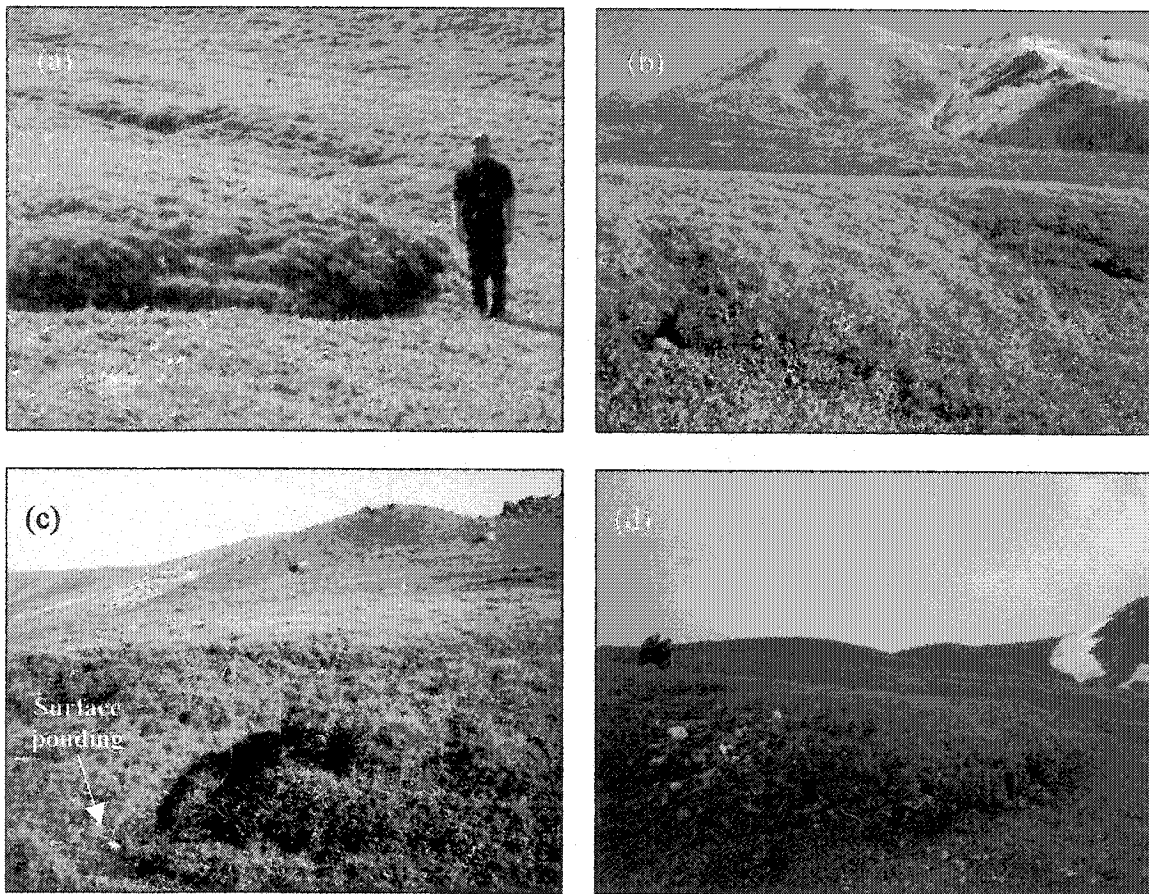


Figure 5.22: Morphology of solifluction lobe risers: (a) steep, well-defined overhanging riser; (b) low-angle riser grading into the surrounding parent slope; (c) multiple-fronted risers with external rim and surface ponding ahead of the front; (d) steep front showing signs of collapse near the base.

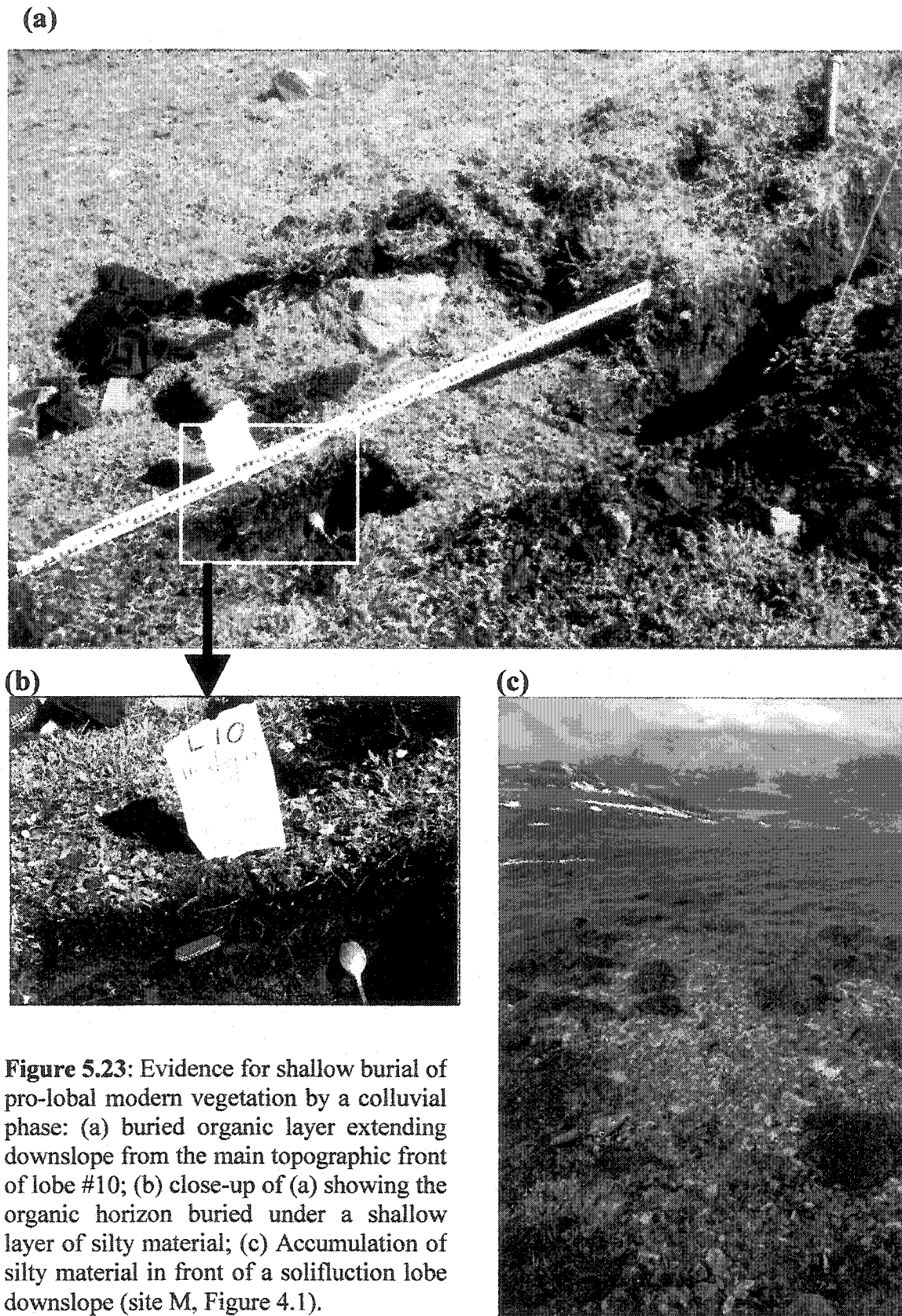
gives them a “step-like” appearance (e.g. Figure 5.21d: lobes #10 and 14 and Figure 5.22 c). A small secondary ridge was also observed downslope of many lobe risers, with tension cracks paralleling the ridge and the riser, suggesting that a zone of differential movement exists ahead of the main lobe front (e.g. Figures 5.21c, d: lobes #5 and 15).

5.3.1.2. *Internal stratigraphy of lobes*

Each excavated lobe revealed one or more buried organic horizons running back from the front. A main basal organic horizon, 3 to 15 cm-thick, was always present. Each of the smaller lobes on the upper slope (e.g. Figures 5.21a, b: lobes #1, 2, 3, 17 and 19), had a

relatively simple stratigraphy, with a main basal organic horizon and some thin humus layers running backwards from the front, but with the latter extending no further than the length of the trench (2 m). Bigger lobes on the middle and lower slope (e.g. Figures 5.21b, c, d: lobes #4, 5, 10, 14 and 15) revealed two buried organic horizons running continuously from the front to the back of the trench. The basal layers were closely spaced in the vertical plane, separated by only 3-10 cm of sediment. The average organic content for buried humus, obtained by loss-on-ignition of small sub-samples, was 14.6% at 1 m ($n=20$, $CV = 0.37$), 11.6% at 2 m behind the front ($n=17$, $CV = 0.30$), and 23.0% for samples collected ahead of the lobe fronts ($n=13$, $CV = 0.29$). The mean organic contents at the three locations differ statistically from each other at $p \leq 0.01$.

For some lobes, the lowermost buried organic horizon extended downslope of the identified lobe front, and sometimes the intersection of this buried horizon with the modern organic mat could not be defined with certainty (e.g. Figures 5.21b, d: lobes #1, 4, 10 and 19). These buried layers were covered by a thin layer of soil (Figure 5.23a). Grey silty horizons were found in some places to cover the lowermost buried horizon in front of the riser (e.g., Figure 5.21b: lobe #19; Figure 5.23b). Accumulations of silty material also covered the modern organic mat ahead of the riser of some lobes (Figure 5.23c), particularly on the lower segment of the slope where silt deposits were also found in inter-hummock zones, on lobe surfaces. These observations suggest that shallow burial of the modern organic mat on the parent slope may occur due to the out-bursting of finer material following the collapse of the front, or simply by extensive collapse of the front over the parent slope. Surface ponding in the depression between the riser



and the rim-like accumulation of material immediately downslope may also cause sedimentation of fine material transported by surface wash processes during the snowmelt period, resulting in shallow burial of vegetation ahead of the front. The fact that some lobes have two closely-spaced basal organic horizons running continuously behind the front over a distance of at least 2 m suggests that shallow burial by colluvial phases and by slumping of the front may represent an integral component of the mechanism of lobe movement. Organic material, buried by colluvial phases ahead of the lobe risers, was reported by Matthews et al. (1986) in Norway. Alexander and Price (1980) described the presence of two main buried organic horizons at the base of a 12 m-long section excavated in a solifluction lobe in the Ruby Range, only 30 km from the Kluane study site, which they interpreted as being planes of differential frontal movement (Figure 5.24a).

Multiple, thin buried organic horizons at the front of lobes have been frequently reported in the literature (Figure 5.24b, c). Worsley and Harris (1974) suggested “that these minor organic horizons represent former poorly developed soils which had formed during periods of relative stability”. Ellis (1979) related the distorted organic horizons at the front to high energy flow of material from the lobe over the parent slope following the breaching of the riser, but argued that the main buried horizon was undisturbed because the cover of *Salix herbacea* formed a continuous surface mat highly resistant to erosion by high energy flow. Alexander and Price (1980), Matthews et al. (1986) and Elliott and Worsley (1999) all emphasized that smaller organic inclusions at the front represent planes of differential movement either by flow or shearing. On the study slope, shear

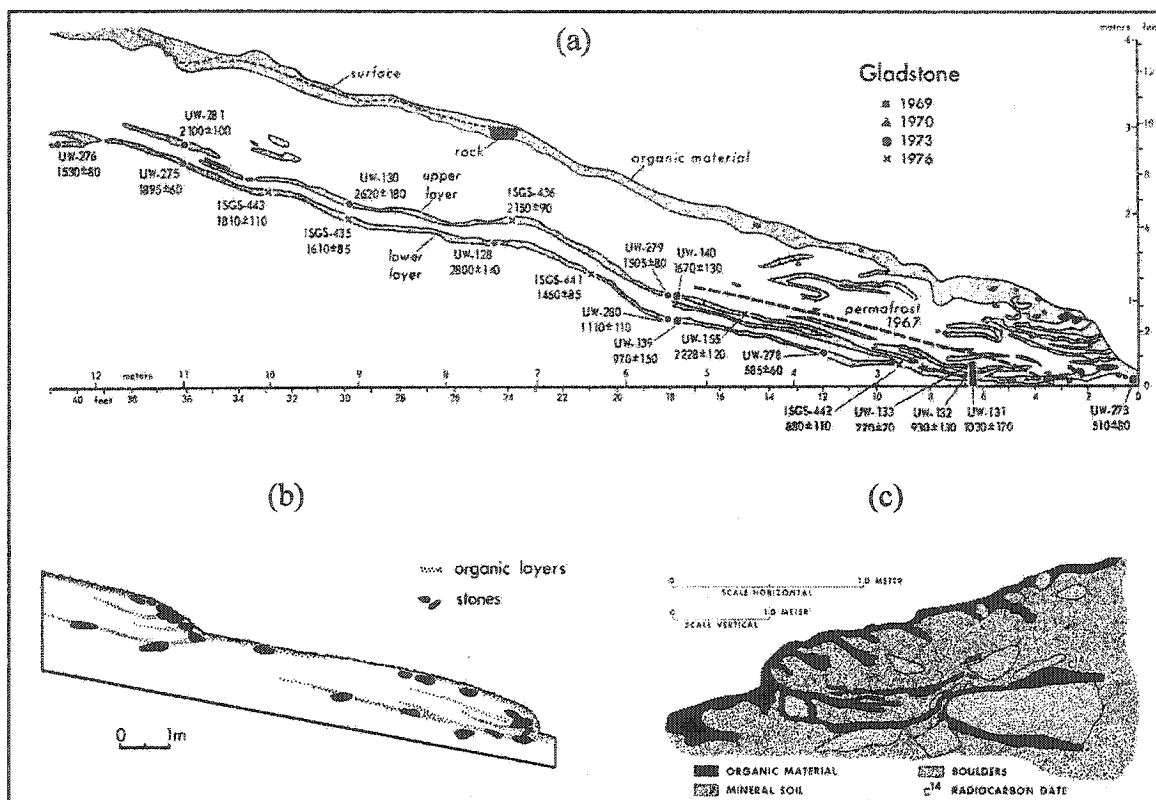


Figure 5.24: Internal stratigraphy of solifluction lobes showing two basal buried horizons and organic inclusions near the front: (a) Alexander and Price (1980); (b) Worsley and Harris (1974); (c) Everett (1967).

planes were observed in many of the excavated lobes. They followed a consistent pattern of a downward inclination at the back of the trench with a sudden inflection upward at the front (Figure 5.25a). This pattern is not unlike the plunging and emergent shear planes observed in glaciers, which reflect, respectively, zones of extending and compressive flow.

Thus the stratigraphy of lobes at the study site suggests that frontal movement may occur by shearing, and that movement may be episodic in nature, with events of rapid flow of fine material resulting from repeated temporally spaced breaches of the front. On the

other hand, excavated sections also show that frontal movement can occur by “rolling over” (Figure 5.25b). A hypothesis that can explain both sets of observations is that lobes on the study slope build up by continuous accumulation of material behind the slowly creeping front, until a threshold of stability is reached and rapid movement by shearing and/or saturated flow occurs. Boulder accumulation at the surface and behind the front, observed on almost every excavated lobe, together with the thicker organic mat on the riser, enhance the shear strength of the front and restrain lobe movement until the strength of the turf layer and underlying soil is overcome. As Ellis (1978, in Harris, 1981) inferred from his measurement of stone dips in a solifluction lobe in Norway, frontal advance can take place spasmodically as a result of rupture of the retaining turf bank through which rapid viscous flow of saturated sediment occurs, depositing a thin sheet over the surface immediately downslope.

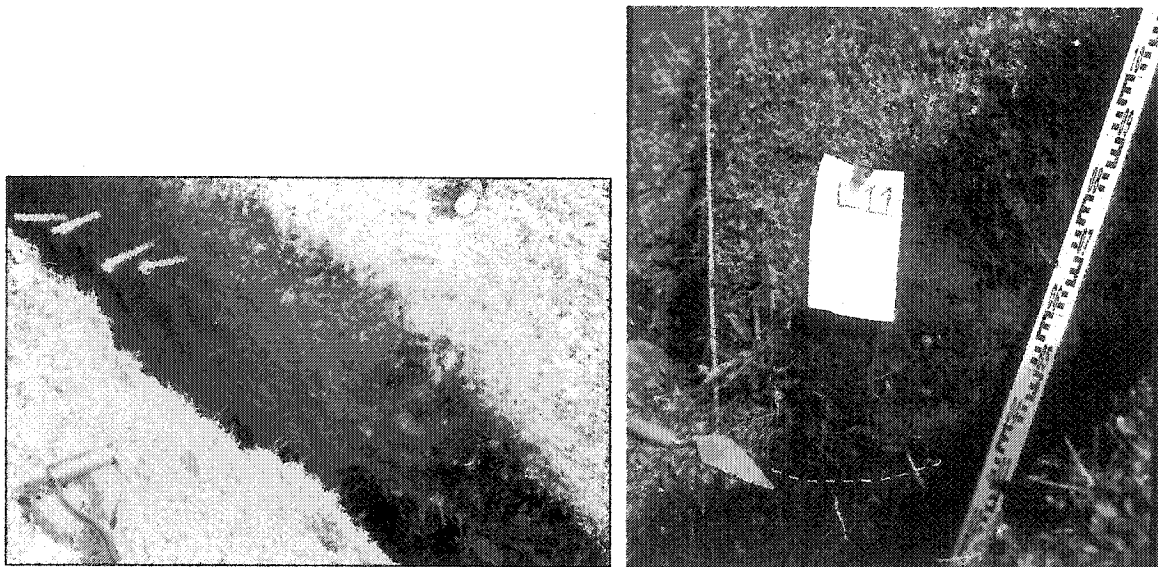


Figure 5.25: (a) Example of shear planes observed inside a solifluction lobe. Note the emergence of the shear planes at the front where they become filled with organic matter; (b) example of a riser “folding over itself”, with the corresponding inverted soil profile.

5.3.2. Bio-physical characteristics of excavated lobes

5.3.2.1. Grain-size characteristics

The grain-size characteristics of each excavated lobe are presented in Table 5.4, and are summarized in Figure 5.21. Lobe sediments are dominantly sands (mean = 48.8%) with almost equal secondary fractions of gravels (mean = 25.8%) and silts (mean = 25.5%), very little clays (mean = 1.0%) and a low organic content (mean = 4.0%). Grain-size distributions are all positively skewed (fine-skewed) (Folk, 1974). All lobes have heterogeneous grain-size distributions, with standard deviation (SD) values ranging between 3.45 and 3.84 (Folk, 1974). This very-poor sorting is characteristic of glacial till and of colluvial sediments derived from till. All samples contain a significant amount of fine material that make the sediments frost-susceptible (e.g. Harris, 1981). Grain-size distributions are similar among the different lobes giving a relatively well constrained textural envelope (Figure 5.26a), and the textural properties of the matrix component

Table 5.4: Grain-size characteristics for 13 excavated lobes.

Lobe#	% Gravel	% Sand	% Silts	% Clays	Mean ϕ	SD	Skewness	% organic
1	20.2	48.5	31.3	1.3	1.8	3.84	0.08	3.9
2	32.2	40.5	27.3	0.8	0.9	3.78	0.26	4.1
3	28.2	49.1	22.7	0.8	0.8	3.64	0.26	3.3
4	34.0	41.1	24.9	1.1	0.6	3.80	0.35	3.8
5	26.3	49.4	24.4	1.1	1.0	3.72	0.25	4.0
6	22.1	50.5	27.4	0.9	1.2	3.61	0.16	3.6
7	31.9	50.8	17.3	0.7	0.3	3.45	0.34	3.9
10	18.7	51.7	29.6	1.4	1.8	3.78	0.05	3.7
11	30.8	51.5	17.8	0.7	0.4	3.48	0.32	5.6
14	20.6	50.8	28.6	1.4	1.6	3.82	0.17	3.9
15	22.9	49.7	27.4	1.1	1.3	3.73	0.13	3.7
17	22.8	51.3	25.9	1.2	1.2	3.70	0.17	3.7
19	24.2	48.9	27.0	0.7	1.2	3.65	0.16	4.1
mean	25.8	48.8	25.5	1.0	1.1	3.7	0.2	4.0
CV	0.20	0.08	0.16	0.25	0.45	0.03	0.47	0.14

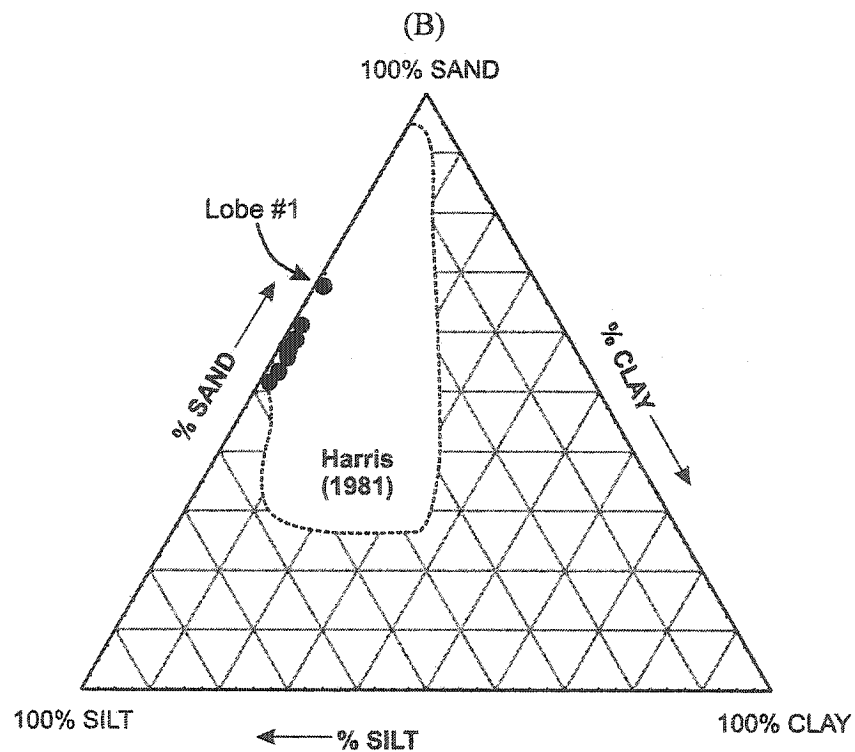
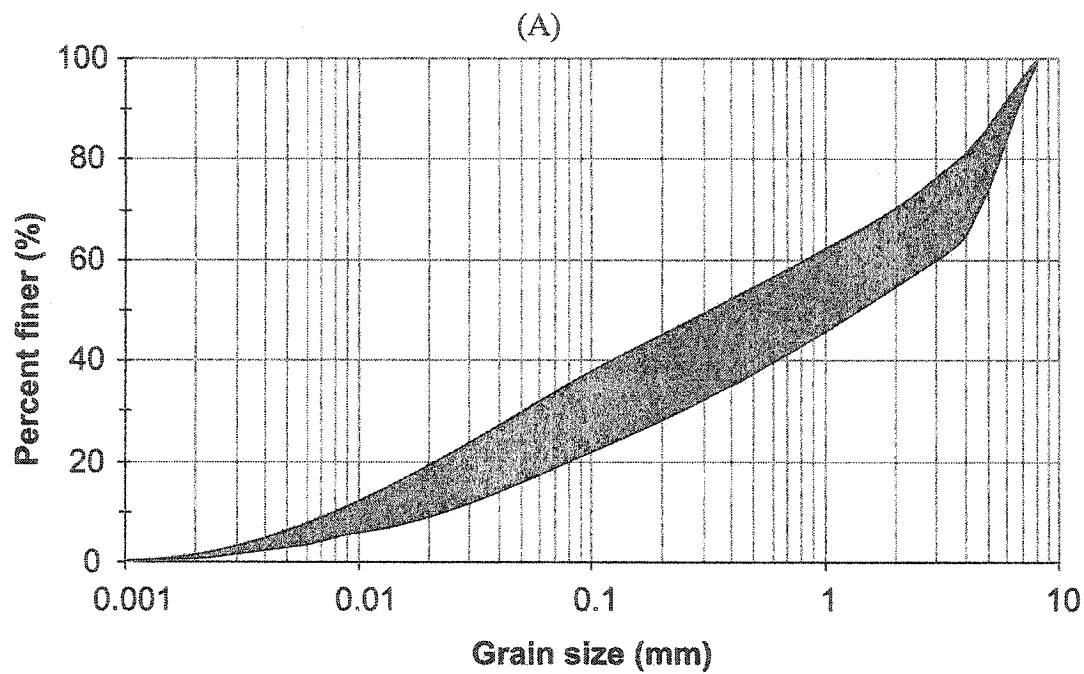


Figure 5.26: Grain size distribution of lobe sediments: (a) grain-size envelope of lobe sediments (<8 mm fraction); (b) ternary plot showing textural properties for the matrix component of lobe sediments (< 2 mm), compared to Harris' (1981) textural envelope for active solifluction.

(sand, silt, clay) also show little scatter (Figure 5.26b). One sample, from lobe #1, stands out from the rest (Figure 5.26b). It has the finest mean grain size (1.8ϕ), the highest content of fines ($32.6\% < 0.063 \text{ mm}$) and the most heterogeneous granulometry ($SD = 3.84$).

5.3.2.2. *Atterberg limits*

Atterberg limits for lobe sediments are presented in Table 5.5 and Figure 5.27. All samples fall below or close to the A-line, in the area corresponding to inorganic silts of low or intermediate plasticity. Lobe #1 again differs from the rest of the samples, with a higher liquid limit and plasticity index. Although the variability in soil consistency between samples is not very high, the apparent scatter of points along a line roughly parallel to the A-line may be interpreted as changes in the dry strength of lobe sediment as the plasticity index and liquid limit increases (Terzaghi et al., 1996). All samples fall at the lower end of the A-line, in the range for solifluction soils identified by Harris (1987).

5.3.2.3. *Structure of lobe sediments*

Parameters describing the structure of lobe sediments are presented in Table 5.6. Soil density and corresponding unit weight, measured 2 m back from the front of the lobes and at a depth of 10 cm in late June 2002, show rather low values. High void ratios, associated with low near-surface densities, give porosity values that are much greater than the typical value of 0.20 reported by Terzaghi et al. (1996) for very mixed-grained glacial till. These low densities/high void ratios probably reflect frost action within lobe sediments. Downward freezing of the soil results in suction forces, and water flow along suction gradients promotes the formation of segregation ice and frost heave pressures.

Table 5.5: Atterberg limits for lobe sediments.

Lobe #	LL (%)	PL (%)	PI
1	46.9	36.4	10.5
2	30.8	23.0	7.7
3	33.8	25.3	8.5
4	35.2	28.9	6.4
5	28.6	22.1	6.6
6	31.9	24.5	7.4
7	33.3	25.1	8.2
10	30.7	22.2	8.5
11	31.3	28.3	3.0
14	38.4	28.4	10.0
15	30.5	25.0	5.5
17	32.5	25.1	7.3
19	36.8	27.8	9.1
mean	33.9	26.3	7.6
CV	0.14	0.14	0.26

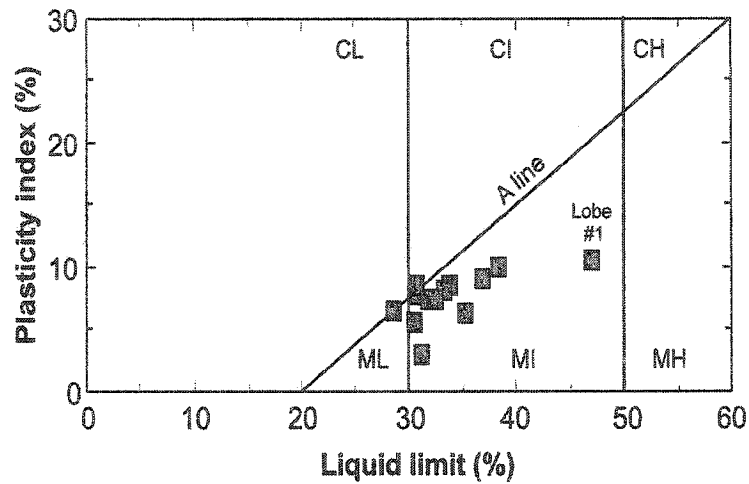


Figure 5.27: Plasticity chart of lobe sediments.

Table 5.6: Porosity, void ratio, natural water content (20 cm depth), dry/wet density and unit weight measured on June 24th 2002 on six solifluction lobes.

Lobe #	Porosity <i>n</i> (%)	Void ratio <i>e</i>	% water content (24-06-02)	% water content (15-08-01)	Bulk density (Mg/m ³)		Unit weight (kN/m ³)	
					ρ_d	ρ_{wet}	γ_d	γ_{wet}
1	0.58	1.37	15.9	13.4	1.15	1.33	11.3	13.1
5	0.56	1.26	12.8	16.5	1.20	1.36	11.8	13.3
7	0.60	1.47	13.6	14.8	1.11	1.26	10.9	12.4
10	0.60	1.51	15.0	14.7	1.09	1.26	10.7	12.4
11	0.54	1.19	12.1	13.3	1.25	1.40	12.3	13.8
19	0.62	1.63	16.0	18.0	1.06	1.23	10.4	12.1
Mean	0.58	1.40	14.2	15.1	1.15	1.31	11.2	12.8
CV	0.05	0.12	0.12	0.12	0.06	0.05	0.06	0.05

While suction in the unfrozen soil causes contraction and a consolidation of the soil aggregates, resulting in a typical platy micro-structure, soil pores and other voids are enlarged by the segregation of ice (Van Vliet-Lanoë, 1985). Overall, the density of soils are generally lower following ground thaw. Average dry unit weights reported by Harris (1972a) in North Norway varied between 14.8 and 16.3 kN/m³, while Washburn (1967)

reported values between 14.4 and 17.2 kN/m³ in Greenland. These values are higher than the mean bulk unit weight of 11.2 kN/m³ reported here for near-surface soil.

Williams (1959) reported increases in unit weight from 12.7-13.4 kN/m³ in the spring to 14.3-14.9 kN/m³ in late summer. This change in void ratio reflects the process of thaw-consolidation, which increases the density of the soil. In this study, samples were collected on June 24th, 2002, more than 10 days after lobe surfaces had begun to thaw. The relatively low gravimetric water contents measured (mean = 14.2%) were similar to those measured in late-summer the previous year (mean = 15.1%), and give a mean saturation level of 30.0%. This suggests that consolidation at the 10 cm depth had been completed, as excess water had been expelled from the soil voids. Thus the high void ratios encountered near the surface of the lobes may result from restricted thaw consolidation due to the low overburden pressures (Van Vliet-Lanoë, 1985), as well as from the high rate of surface thawing which helps to preserve macropores (Dyke and Egginton, 1990). The loose structure of near-surface lobe sediments, in general, reflects the effectiveness of frost activity and the likelihood for associated solifluction movements.

5.3.2.4. *Organic mat*

The organic mat, measured on each excavated lobe, was nearly always thicker on the apex of the riser than at 2 m back from the front (Figure 5.28). This is consistent with previous field observations by Benedict (1970) and Price (1970). Movement has been shown to be minimal near the lobe front, and plant communities take advantage of this more stable microhabitat (Price, 1971). Plant colonisation must play a significant role in

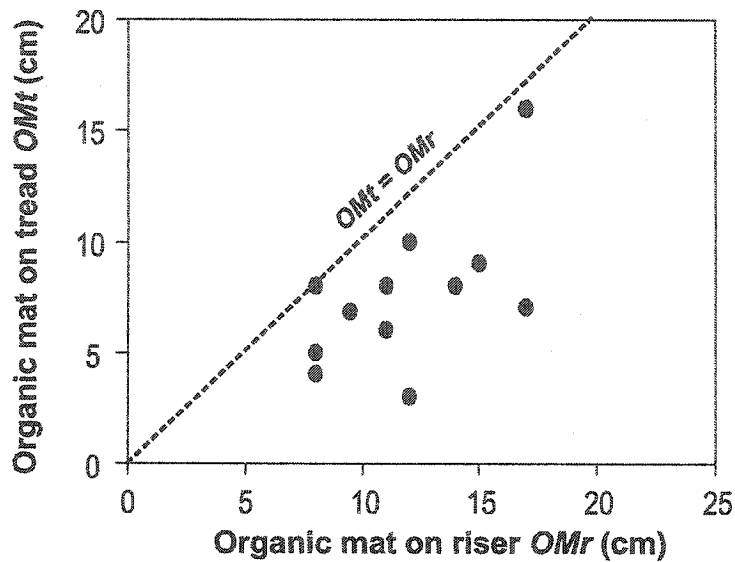


Figure 5.28: Comparison between organic mat thickness on the lobe front and at 2 m back from the front.

the development of lobes. The more stable and drier soil of lobe risers promotes the colonisation by shrub plant communities. Their deeper root system augments the stability of the front and increases the differential movement, allowing the slow build-up of material behind the front until the strength of the turf layer and underlying soil is overcome.

5.3.3. Age of buried organic horizons

Conventional radiocarbon ages obtained on humus samples are summarized in Table 5.7, with 95% confidence intervals reported for both ^{14}C ages and calibrated ages. The location of each sample relative to the front is indicated, at -1 or -2 m behind the front for samples collected under the lobes, and ahead of the front for MRT determination of the modern sod (“pro-lobal samples”).

Considerable variability exists among ages obtained from both the buried organic layers and the pro-lobal samples. The ^{14}C ages at -1 m behind the front vary between 750 ± 60

Table 5.7: Conventional ^{14}C age and calibrated age for humus samples collected from organic horizons buried under solifluction lobes. BP ages refer to years before 1950.

Lobe number	Lab number	Location relative to lobe front (m)	^{14}C age (yrs BP $\pm 2\sigma$)	Calibrated age (2 σ cal. yrs BP) Max. (cal age) min.
1	UL-2496	0.17	260 $\pm$ 180	500 (300) 0
1	BGS-2349	-1.00	1800 $\pm$ 120	1870 (1710) 1560
1	BGS-2350	-2.00	1860 $\pm$ 80	1880 (1820) 1630
2	UL-2494	1.20	90 $\pm$ 180	300 (40) 0
2	BGS-2347	-1.00	840 $\pm$ 100	910 (730) 670
2	BGS-2348	-2.00	735 $\pm$ 80	730 (670) 570
3	UL-2495	0.70	50 $\pm$ 160	290 (0) 0
3	GSC-6614	-1.00	750 $\pm$ 60	730 (670) 660
4	UL-2490	-2.00	780 $\pm$ 180	920 (690) 560
5	UL-2483	-2.00	1250 $\pm$ 180	1310 (1170) 970
6	UL-2510	0.30	230 $\pm$ 120	460 (290) 0
6	UL-2511	-1.00	840 $\pm$ 160	930 (730) 650
7	UL-2488	0.30	450 $\pm$ 180	650 (510) 300
7	BGS-2343	-1.00	880 $\pm$ 80	920 (780) 690
7	BGS-2344	-2.00	1440 $\pm$ 90	1410 (1320) 1290
10	UL-2489	-2.00	670 $\pm$ 120	690 (650) 540
14	UL-2491	-2.00	1380 $\pm$ 180	1480 (1290) 1090
15	UL-2492	1.35	30 $\pm$ 120	270 (0) 0
17	UL-2484	0.10	1010 $\pm$ 90	1170 (930) 730
17	BGS-2352	-1.00	1430 $\pm$ 45	1410 (1310) 1280
17	BGS-2353	-2.00	1460 $\pm$ 90	1420 (1330) 1290
19	BGS-2354	0.25	960 $\pm$ 80	950 (920) 760
19	GSC-6617	-1.00	780 $\pm$ 60	740 (690) 660
19	GSC-6619	-2.00	1100 $\pm$ 60	1060 (1040) 930

BP and 1800 ± 120 BP, with a median age of 840 ± 100 BP. At -2 m behind the front, ages vary between 735 ± 80 BP and 1860 ± 80 BP, with a median age of 1250 ± 120 BP. Ages obtained for pro-lobal samples, collected near the identified junction with the main buried layer, vary between 30 ± 120 BP and 1010 ± 180 BP, with a median of 245 ± 150 BP. Median ages ($\pm$ average 2σ error) are used here because age distributions at each location show pronounced skewness. The use of summary statistics allows a general

picture of the long-term rates of frontal advance for the twelve lobes sampled on the slope: the median age increases from ahead of the lobes to the back of the trench at -2 m behind the front. However, significant variability exists among the different lobes, and the range of ^{14}C ages (min, max), when grouped by location, overlap each other. To compare the ages obtained from the different lobes, ^{14}C dates were plotted for each lobe according to their location relative to the front (Figure 5.29). Three important observations can be made:

- (1) Three lobes with samples dated at both the -1 and -2 m position (#1, 2, 17) have similar ages at both positions (overlapping 95% confidence intervals);

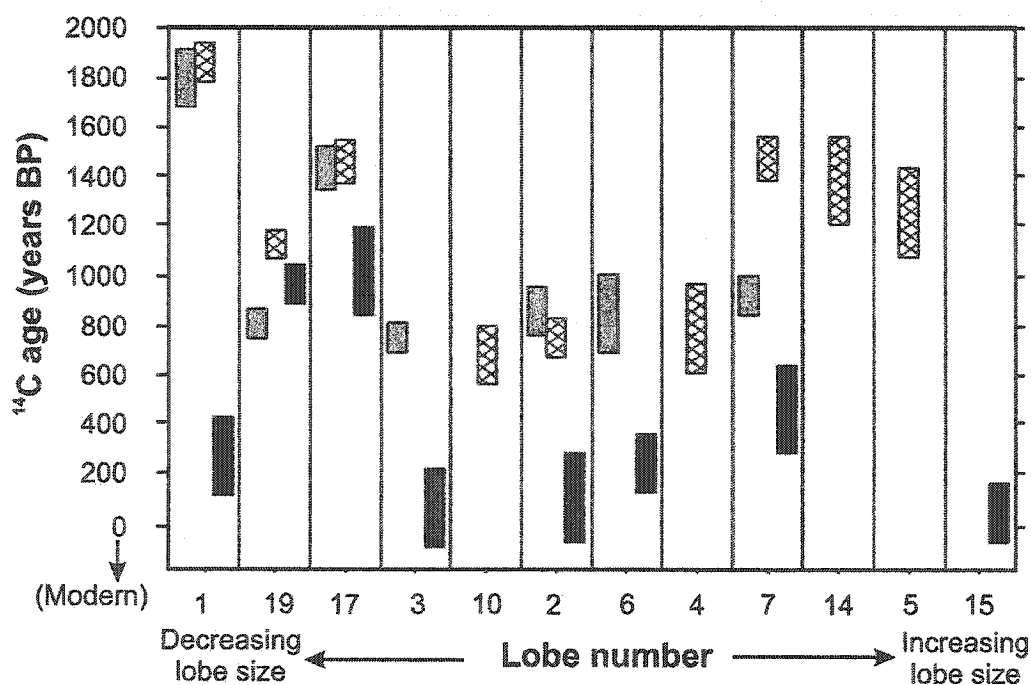


Figure 5.29: Plot of ^{14}C ages for each lobe according to their position relative to the lobe front. Bars represent the 95% confidence intervals around ^{14}C ages. Black bars represent pro-lobal samples, grey bars represent samples at 1 m behind the front and cross-hatched bars represent samples at 2 m behind the front. Lobes are sorted by size.

- (2) Lobes 19 and 17 have particularly high ages ahead of their fronts, while lobe 1 is notably older than other lobes at the -1 and -2 m positions;
- (3) Despite this apparent scatter, some consistency appears between inter-lobe dates obtained at the same location relative to the lobe front.

5.3.3.1. Timing of lobe movement

The fact that some lobes exhibit a similar age over a 1 m distance appears, at first, to be inconsistent with the common representation of solifluction lobes as being slow-moving landforms. Following Ballantyne (1986), it is argued here that it is improbable that contamination, either by modern rootlets or by percolating acids, has affected the age of buried humus at -1 and -2 m so as to produce this pattern of uniformity in three different lobes. Contamination of samples of different ages most likely would have produced younger, but still different ages. Instead, the similar ages found over a distance of 1 m under three excavated solifluction lobes in this study are taken as evidence for a rapid movement of lobe fronts resulting in the burial of vegetation over a minimum distance of 1 m. This interpretation is strengthened by the stratigraphic observations previously discussed, which suggest that frontal movement proceeds by shearing, with episodic bursting of fine material over the parent slope. The ^{14}C dates at -1 and -2 m behind the front of lobes #1, 2 and 17 support this hypothesis.

If we accept the idea of a mechanism of lobe movement where periodic rupturing of the front leads to slumping/bursting of soil over the parent slope, the fact that some lobes (lobes #2, 4 and 10) have ages at -2 m falling into the median age range obtained at the -1 m position may be explained. Despite the fact that dates are not available at -1 m for

lobes #4 and #10, it may be hypothesized that these would be similar to the ages found at -2m. It is certainly the case for lobe #1 and #17, whose ages at -1 m fall out of the median age range for that position, but into the -2 m median age range. Similarly, the age of lobe #2 at -2 m falls out of the median age range for that position because it is the same as its age at -1 m. Thus age differences between 2 successive positions under a lobe, such as here between 1 and 2 m behind the front, may give a false impression of slow continuous movement over time, whereas in reality frontal movement is episodic, and a continuous series of dated humus samples buried by lobe advance would in fact display “step-like” changes over time.

5.3.3.2. Age of pro-lobal samples and mean residence time of soil

If we turn our attention to the ages obtained ahead of the fronts, Figure 5.29 suggests a clustering of dates between 0 – 400 years BP, with two particularly old ages of 1010 ± 90 and 960 ± 80 for lobes #17 and #19, respectively. Apart from these two lobes, and lobe #7 with a date of 450 ± 180 , all dates ($\pm 1\sigma$) obtained from the base of the organic mat in front of the lobes are not statistically different from an age of 220 ^{14}C years BP. Because of a decrease in the ^{14}C content of the atmosphere between 1650 and 1950 AD (0-300 calendar years before present), it is nearly impossible to use ^{14}C in this time range (e.g. Trumbore, 2000). Samples formed during that period generally have conventional radiocarbon ages of 80-220 years BP, and are indistinguishable without “wiggles-matching” (e.g. Van Geel and Mook, 1989). Consequently, the ^{14}C age of pro-lobal samples for lobes #1, 2, 3, 6, and 15, once calibrated, are not statistically different from 0, meaning that their ages may be close to being modern. In all cases the ages of these pro-lobal samples must be less than 300 years old, and this represents a maximum value for

the apparent mean residence (AMRT) time of the soil, since samples were taken at the base of the organic mat.

The abnormally high ages obtained for lobe 17 and 19 are more surprising. A high AMRT value for lobe 17 could explain why the ages at -1 and -2 m behind are slightly higher than surrounding lobes of similar size, but in the case of lobe #19, an AMRT of 960 ± 80 years BP, if it was to be subtracted from the ages at -1 m and -2m, would yield a true age of burial of 0 years BP at -1 m and 140 ± 140 years BP at -2 m. This is improbable since lobe #19 showed no signs of major disturbance which might suggest such a rapid and recent advance. On the contrary, lobe #19, with its steep and well-defined riser, appears as if it is growing toward instability, and not as if it had collapsed relatively recently.

In late June 2002, the front of lobe #19 was re-excavated, with more emphasis being put on observation of the stratigraphy downslope from the front. The stratigraphy proved to be complex, showing signs of both shearing and shallow burial, but a reasonably well-defined silty horizon was observed disrupting an organic layer that had previously been interpreted as being one coherent layer of modern organic material. Therefore, the age of 960 ± 80 years BP is interpreted as being part of the lobe, in that it was buried by a shallow layer of fine material which probably burst out of the front. The other possibility is that this silty layer accumulated by sedimentation within an area of surface drainage in front of the lobe, as was sometimes observed for lobes with an external rim. This could explain why the age is a bit greater than the age at -1 m (780 ± 60 years BP). Lobe #17

was also re-excavated in late June 2002. Although no evidence was found for shallow burial by a silty layer, the basal buried organic horizon was seen to extend further downslope from the front and the exact location of its intersection with the modern organic mat was difficult to pinpoint. Consequently, it is highly probable that the age of 1010 ± 90 years BP also reflects shallow burial in front of the lobe.

These observations demonstrate the difficulty of identifying the true fronts of solifluction lobes. The question is whether the front of a lobe is defined by the most prominent topographic bulge, which is easy to identify, or by the intersection of the lowermost buried organic horizon with the modern organic mat, which may be difficult to locate. The complexity of the burial process at the front, however, makes it difficult to outline the limit of a particular lobe, and hinders the inter-lobe comparison of the age of samples taken "at a fixed distance" behind the front.

In an effort to objectively discriminate the ages relating to the modern sod from the ages inherited from shallow burial at the front, all ^{14}C ages obtained for pro-lobe samples were plotted against their respective distance from the main topographic front (Figure 5.30). The graph shows that samples collected further downslope from the front yielded the youngest ages. Between 0 and 40 cm ahead of the front, older values are found. The calibrated ages for lobes #1, 2, 3, 6 and 15 are all statistically indistinguishable from each other and range between 0 and 300 cal. years BP. It is believed that these ages represent the AMRT of modern sod. Lobes 7, 17 and 19, on the other hand, are statistically different from the age of 220 ^{14}C years BP, which is the limit of applicability of ^{14}C

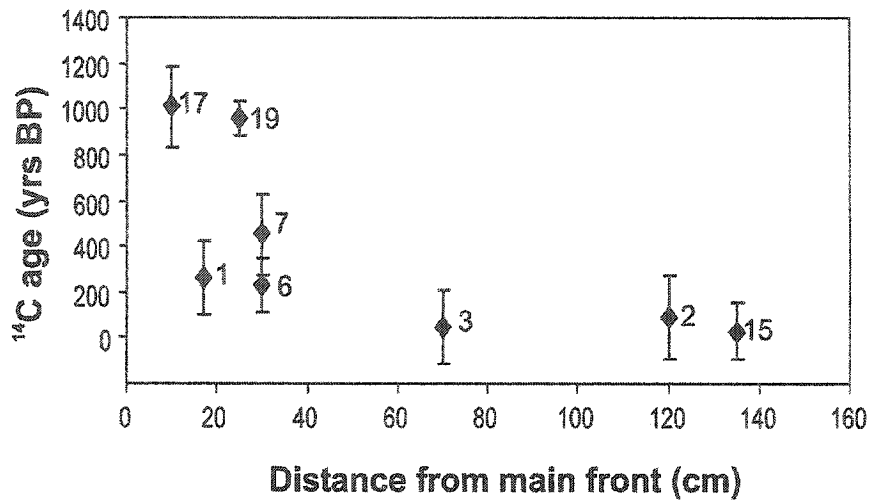


Figure 5.30: Scatter plot of ^{14}C age for samples from ahead of lobes, with 95% confidence intervals, against corresponding distance from main topographic front.

dating. Therefore it is thought that these humus samples were taken from the main organic horizon, which had been buried by a thin sheet of burst/slumped material, or by sedimentation in a surface ponding area proximal to the lobe front. Since three of the five dates interpreted as representing modern sod AMRT are less than 100 years BP, and the two others are 260 ± 180 and 230 ± 120 years BP, it is tempting to conclude that the AMRT is insignificant in the interpretation of ages for buried humus. However, because of the limits of radiocarbon dating in the 0-300 cal. years BP range, the best estimate for the AMRT on the study slope remains < 300 ^{14}C years BP. As already stated, this is a maximum estimate since samples were taken from the base of the organic mat.

Similar AMRT values were reported for alpine regions by Benedict (1966) in Colorado (355 ± 155 ^{14}C years BP), Ellis (1979) in North Norway (50 ± 100 ^{14}C years BP) and Alexander and Price (1980) in the nearby Ruby Range (340 ± 100 ^{14}C years BP). Greater AMRT ages have been reported in the Central Canadian Arctic (800-950 ^{14}C years BP;

Dyke, 1981), and in the Canadian Rockies (638 ± 100 ^{14}C years BP; Smith, 1987b). These ages, however, were obtained by extrapolating the rate of lobe advance between the two closest dated samples to the front of the lobe. This technique appears dubious, considering the episodic nature of lobe movement, and these dates therefore are not considered reliable.

5.3.3.3. *Inter-lobe comparison*

There is no apparent relationship between current lobe size and long-term rate of advance. Although this might be the result of the small sample size and/or the lack of dates for bigger lobes, especially at the -1 m position, the present distribution of dates nonetheless suggests that lobes have moved at a more-or-less similar rate over time. This contradicts the short-term field data showing an increase in surface movement rates with increasing lobe size. The resolution of this contradiction lies in a model of solifluction lobe development that includes the slow accumulation of material behind a rigid front, lobe enlargement proceeding backwards and sideways from this front and increasing the pressure until break-up happens. Finally, a rapid advance of the front occurs over a distance of up to 1 m or more.

5.3.3.4. *Cycle of lobe development*

The strategy of comparing the age of samples taken at a fixed distance behind the front of different lobes loses its validity if the true front of a lobe, that is the intersection point between the main buried humus horizon and the modern sod, lies some distance downslope from the topographic front. Because shallow burial may occur over distances up to 1 m downslope from the front, it can be hypothesized that samples have been in fact taken at various locations within a *minimum* distance of 2 m from their "true front".

Moreover, if frontal advance is an intermittent process as proposed, there may be no direct relationship between the age of a sample and its position behind the current front. In this case, ^{14}C ages must be interpreted as discrete burial events resulting from the intermittent and rapid advance of the fronts.

From this perspective, it may be possible to approximate the cyclic period for lobe development by looking at the relative concentration of ^{14}C dates over time, independent of their sampling location. If frontal advance is intermittent and if the period of movement is similar between lobes, burial events should be clustered for lobes with similar frontal morphology. A probability density function for all ^{14}C dates which takes into account the laboratory error has been constructed using a Kernel estimator (e.g. Silverman, 1986). The density function is equal to :

$$f(x) = 1/nh \sum_{i=1}^n K \left(\frac{x - X_i}{h} \right) \quad (5.4)$$

where n is the number of observations X_i , and h is the window width, also called the smoothing parameter or bandwidth. K is a Kernel weight function which satisfies the condition :

$$\int_{-\infty}^{\infty} K(x) dx = 1 \quad (5.5)$$

That is, a symmetric density function with a cumulative probability of 1. Various smoothing functions can be used. The Kernel estimate is constructed by placing a 'box' of width $2h$ on each observation and summing to obtain the estimate. This is a standard practice for smoothing histograms to obtain a density curve (e.g. Jacoby, 1997). Since

radiocarbon dates represent mean ages with a 2-sigma laboratory error ($\pm 2\sigma$), a normal distribution is used for the kernel function.

A Kernel density function is constructed by summing the normal distributions centred on each ^{14}C date and with a base $2h = 240$ yrs representing $\pm 2 \times$ the average standard deviation for all the dates (95% probability). A similar technique was used by Gagnon and Payette (1981) to infer Holocene fluctuations of the tree-line in Northern Québec. The authors incorporated the mean laboratory error for all their ^{14}C dates into a histogram showing variations in the concentration of macrofossils from peat environments. Morin and Payette (1988) made further use of the method to infer Holocene variations in solifluction activity on the coast of Hudson Bay in Québec, which they compared with a North American compilation of solifluction activity.

The resulting density plot (Figure 5.31) shows the age probability over a minimum distance of 2 metres behind the fronts of the 12 lobes studied. Interpretation of the density curve is as follows:

- (1) The first concentration in ages, centred around 100-150 ^{14}C years BP, is associated with the apparent mean residence time (AMRT) of the modern soil, downslope from the lobes;
- (2) The second and third concentrations, centred at about 800 ^{14}C years BP and 1450 ^{14}C years BP, represent two distinct periods of burial by frontal movement;
- (3) The smaller concentration, centred at about 1800 ^{14}C years BP, is considered to be an outlier. It is interesting to note that this outlier, lobe #1, is the only lobe with

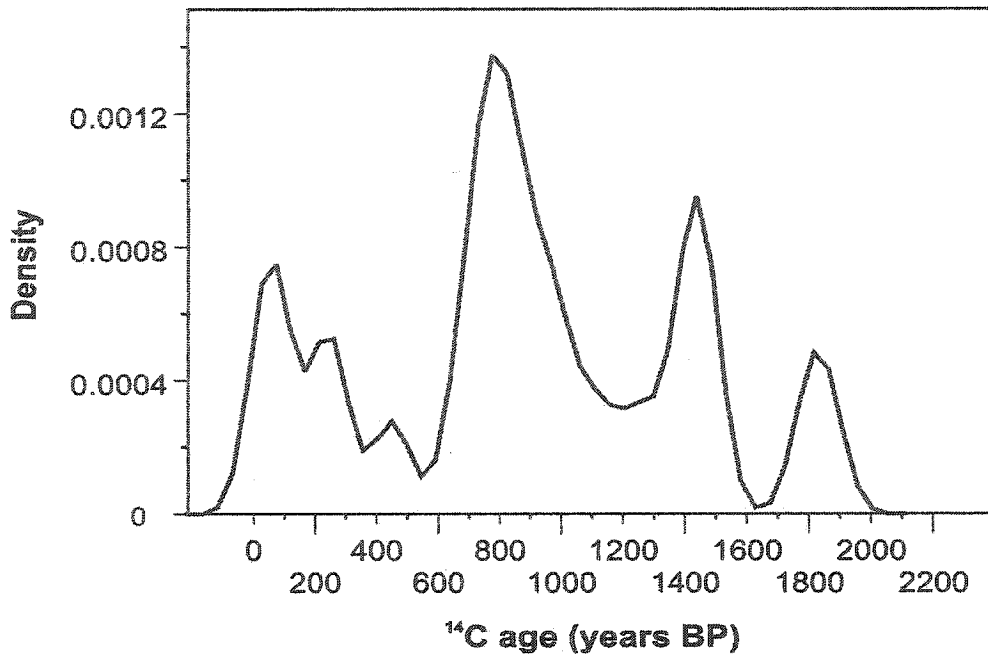


Figure 5.31: Density plot of ^{14}C dates obtained from 12 solifluction lobes.

significantly different sedimentological properties. Lobe 1 is finer grained and has the most heterogeneous grain-size distribution. It has the highest liquid limit and plasticity index, which makes it more resistant to shearing forces (see Figures 5.26 and 5.27).

No attempt was made to test the statistical significance of modal peaks. Although there are existing methods to test the multimodal character of probability density functions, these require complicated computations and none appear entirely satisfactory (see review by Henderson et al., 2002). It is believed that the three identified modes stand out clearly enough to be significant, but no proof is presented.

From a traditional perspective of slow continuous movement by the lobes, the two distinct age concentrations at ≈ 800 and ≈ 1450 ^{14}C years BP could be interpreted as the average position of the lobes fronts at ≈ -1 and -2 m from present. Considering a mean AMRT age of 150 ± 150 ^{14}C years, this would yield an average rate of frontal advance of 0.15 cm/year over 2 meters, a value which falls in the lower part of the range 0.1-1.0 cm/year reported by various authors and compiled by Matsuoka (2001). In the nearby Ruby Range, Alexander and Price (1980) reported more rapid mean rates of 0.6-1.0 cm/year, over a time range of 2500 years. However, in the light of both stratigraphic observations and radiocarbon dates reported in this study, which provide the evidence for a mechanism of rapid and episodic lobe advance, these mean rates are of significance only for comparing long-term evolution of solifluction landforms between sites. At the Kluane site, they must be analysed in term of cyclic deformation of the front.

The peak concentrations of ages (≈ 800 and ≈ 1450 ^{14}C years BP) can be interpreted as the means of two distinct periods of rapid burial by frontal advance. A general estimate of the time taken by the lobes to rebuild following break-up of their front can thus be obtained from the difference between these two peaks. From the density curve, the time-difference is 650 ± 350 ^{14}C years. Moreover, if we assume that the present position of the lobes results from slow build-up since the last burial event, we can subtract the mean age at present time, which is the AMRT of the modern soil (taken as 150 ± 150 ^{14}C years) from the mean age of the last event of burial (800 ± 200 years BP), which gives the same time range of 650 ± 350 ^{14}C years. This suggests a constant period for the cyclic

development of lobes. Since lobes on the slope have different frontal and planar morphologies, lobes must be at different stages of this break-up cycle across the slope.

Two lobes, which have dates obtained from three positions, can be used to test the hypothesis about the time period of cyclic lobe development. Lobe #17 has similar ages of 1330 ± 90 and 1310 ± 90 ^{14}C years BP at -2 and -1m, respectively, while an age of 1010 ± 90 ^{14}C years BP was found at 0.10 m ahead of the front. This latter age was first considered in term as the AMRT of the modern soil, but re-excavation in late-June 2002 revealed that it was part of a thin layer of buried humus running downslope from the main front for approximately 1 m (see cross section, Figure 5.21a). This humus layer, buried under a thin sheet of material ahead of the main front, must have resulted from the last break-up of the front. Its intersection with the main buried organic horizon, 20-30 cm behind the limit of the main front, therefore marks the ancient position of the lobe front prior to its rupture and collapse. If we consider that an age of $[1310-1330] \pm 90$ (mean = 1320 ± 100) years BP represents the previous event of rapid burial by collapse of the front, the time difference between these two events is $1320 \pm 100 - 1010 \pm 90 = 310 \pm 190$ ^{14}C years. This value is less than the average time period estimated from the difference between the two peak concentrations (650 ± 350 ^{14}C years). On the other hand, if the last collapse event occurred at 1010 ± 90 ^{14}C years BP, and considering an AMRT age of 150 ± 150 ^{14}C years, lobe 17 would have taken 860 ± 140 ^{14}C years to rebuild to its present form and position.

Although both periods are within the rather large estimate of 650 ± 350 ^{14}C years, their means vary considerably from each other. In all respects, it appears that lobe 17 has developed more slowly during its last development cycle. Lobe 17 is a small lobe adjacent to the bigger lobe 2 (see Figure 4.1), and it is possible that an extensive collapse of the front and sides of lobe 2 may have triggered the formation of the small secondary outcrop which is lobe 17. The last cycle of frontal movement, slower than the previous one, may be explained by the fact that lobe 2 appears to be growing at the expense of lobe 17, which is gradually becoming incorporated into the expanding sides of lobe 2.

The second lobe examined is lobe #7, a 9 m-long lobe located midslope. Ages of 1320 ± 90 and 780 ± 80 ^{14}C years BP were found at -2 m and -1 m from the front, respectively, while a date of 450 ± 180 ^{14}C years BP was found 0.30 m ahead of the front. This age was obtained at the bottom of a confusing intersection of the main buried organic layer and the modern organic mat, and was interpreted as being part of the main buried layer. Lobe 7 had a remarkably thick buried organic horizon. A closer examination revealed that it was actually composed of numerous discontinuous organic layers showing obvious signs of shearing (see Figure 5.21c). The lobe seems to have advanced by basal shearing with successive periods of restricted colluvial outbursts at the front. Compared to other lobes, which have two closely-spaced basal organic horizons, suggesting extensive shallow burial ahead of the front, lobe 7 appears to have advanced by more restricted collapse of its front. The age of 450 ± 180 ^{14}C years BP must then represent the last burial event, as it was found just ahead of the main topographic front. Assuming that the previous burial period occurred around 780 ± 80 ^{14}C years BP (the age at -1 m), the time

difference yields an estimate of 330 ± 260 ^{14}C years for a full cycle of lobe development. Taking an AMRT of 150 ± 150 ^{14}C years for the modern soil, and that lobe 7 is actually rebuilding from its previous collapse event at 450 ± 180 ^{14}C years BP, Lobe 7 has been growing for approximately 300 ± 300 ^{14}C years. With its rather steep front, which shows no sign of recent disturbance, lobe 7 may be approaching its critical state of stability.

In summary, lobes 7 and 17 provided some insights into the timing of cyclic lobe development. Periods of 310 ± 190 , 330 ± 260 and 300 ± 300 ^{14}C years BP, inferred from the stratigraphy of these lobes, tend to suggest that the time for the developmental cycle of lobes is closer to the lower end of the broad estimate of 650 ± 350 ^{14}C years calculated from the density curve. Lobe 17, on the other hand, suggests that it has been rebuilding for the past 860 ± 140 ^{14}C years, a number higher than all previous estimates. Whether this lobe has become inactive, or is being incorporated by its larger neighbour (lobe 2) is unknown, and for now this higher value can only be interpreted as a possible natural variability in the cyclic period. It is thus concluded here that the best estimate of the time period required for the completion of a full cycle of development remains 650 ± 350 ^{14}C years, over which a lobe will slowly grow steeper by accumulation of soliflucted material behind its resistant front until it collapses rapidly onto the parent slope.

5.3.3.5. *AMS dating of macrofossils*

Ages obtained from AMS dating of small wood fragments are presented in Table 5.8, together with the conventional radiocarbon dates obtained from their respective enclosing humus samples. In most previous studies, ages for organic horizons buried under

Table 5.8: Conventional and AMS radiocarbon ages for three humus samples buried under solifluction lobes #3 and #19. Ages are presented with their 95% confidence intervals.

Lobe #	Lab #	Distance to front	Conventional ^{14}C age ($\pm 2\sigma$)		AMS age ($\pm 2\sigma$)	
			<i>corrected</i>	<i>calibrated</i>	<i>corrected</i>	<i>Calibrated</i>
19	GSC-6619	2 m	1100 $\pm$ 60	1060 (1040) 930	1560 $\pm$ 160	1620 (1420) 1300
19	GSC-6617	1 m	780 $\pm$ 60	730 (670) 660	910 $\pm$ 100	940 (840) 750
3	GSC-6614	1 m	750 $\pm$ 60	740 (690) 660	1090 $\pm$ 100	1090 (980) 930

solifluction lobes were obtained from conventional radiocarbon dating of bulk samples, or thin slices of soil humus. Costin et al. (1967), however used conventional radiocarbon dating of wood remnants buried under solifluction terraces in Australia. This is the first time we know of that ^{14}C dates of both humus and macrofossils are available for direct comparison.

All AMS dates were found to be older than the conventional ^{14}C ages of their corresponding enclosing humus samples. The age difference, however, varied among the three lobes. It was 460 ± 220 ^{14}C years at -2 m behind the front of lobe #19, 130 ± 160 ^{14}C years at -1 m behind the front of lobe #19 and 340 ± 160 ^{14}C years at -1 m behind the front of lobe #3. At first glance, it may appear surprising that wood fragments date older than their enclosing humus. Well-preserved remnants of vegetation are often thought to approximate the true age of burial, while the organic matter, which consists of a mixture of material in various states of decomposition and humification, is of relatively older age (Matthews, 1984). On the other hand, Goh and Molloy (1978) found that removing charcoal fragments from a podzol buried by Late Pleistocene solifluction deposits in South Island, New Zealand, reduced its age by 150 ± 40 ^{14}C years. The authors

interpreted the charcoal remnants as being a very resistant fraction, which can yield older, but also variable ages.

Thus it is possible that wood remnants, if covered under an advancing lobe, have an age which approximates burial. On the other hand, small wood remnants may also represent resistant fractions present in the soil before burial occurred. In the present study, since ages obtained from wood remnants were consistently older than the enclosing organic matter, it is concluded that wood was already present (dead) in the sample prior to burial. The age difference between the AMS date and the conventional date thus reflects the time which a particular wood fragment was present in the soil prior to burial, but is not an approximation of the apparent mean residence time of the soil. The AMRT age is younger, since it represents an average of all fractions present in the organic matter, including any small wood fragments present.

The ^{14}C age of the oldest uncontaminated organic fraction in a surface (or buried) soil provides the closest approximation of the date at which a soil formation was initiated at the site (Matthews, 1984), but Wang et al. (1996) later showed that this age is at best only about one-tenth the age of the soil. On the other hand, the age of the youngest uncontaminated fraction in a buried soil provides the best estimate of the time elapsed since burial (Matthews, 1984). Because discrete wood samples may represent resistant fractions of various and older age, and also because bulk humus samples yielded the youngest ages, the latter are preferred to the former for the interpretation of burial events by solifluction lobe advance.

There is a paucity of knowledge of the dynamics of soil organic matter. Despite some attempts to develop unified methods of fractionation to discriminate the age of the various components of soil organic matter, consistent results have yet to emerge (Goh and Molloy, 1978; Birkeland, 1999). Therefore, our best estimate of the age of a particular soil, especially for thin, poorly developed soils, remains an average value for the whole organic horizon, which is the apparent mean residence time.

5.3.4. Synthesis

Observations of the profiles, internal stratigraphy and sediment characteristics of 13 solifluction lobes, along with radiocarbon dating of their buried organic horizons, have led to several conclusions:

- (1) Lobe development occurs by solifluction of material behind a rigid front. Enlargement of lobes proceeds backwards and sideways from this front, with higher solifluction rates leading to the build-up of larger lobes downslope.
- (2) Frontal advance occurs when accumulation of soliflucted material reaches a threshold of stability. Breaching of the front with colluvial outburst, and slumping of the front onto the parent slope, are the two mechanisms of lobe advance which result in the rapid burial of organic material.
- (4) The time required for completion of this developmental cycle is estimated to be between 300 and 1000 years. The morphology of the front, and the overall profile of the lobe indicate the stage of development of a particular lobe. Since lobe profiles vary across the slope, lobes are probably at different stages of this cycle, and the age

at a particular location behind the front may differ by up to 300-1000 years between two adjacent lobes.

An episodic mechanism of rapid lobe advance has been discussed by numerous authors. Price (1969), Ellis (1979) and Washburn (1979) reported on the breaching of lobes fronts with subsequent washing out of the fines, while Price (1974) proposed that snow accumulation in the lee of steep risers causes erosion of the front and destruction of the lobe by complete collapse. Price concluded that lobes grow to a critical point where snow accumulation starts eroding the front so that "...such lobes carry within them the seeds of their own destruction" (Price, 1974). Other authors, such as Williams (1957), Price (1970), Alexander and Price (1980), Matthews et al., (1986) and Elliott and Worsley (1999), concluded from observations made of the internal stratigraphy of lobes that multiple organic layers near the front indicate at least some episodic movement.

Although the episodic nature of lobe movement has been discussed at times, it has never been considered in studies of long-term rates of movement and their relationship with paleoclimate. In studies where multiple dates were obtained from a single buried organic horizon, near contemporaneous or overlapping ^{14}C dates were interpreted as periods, not episodes, of accelerated lobe movement. For example, Benedict (1966), Alexander and Price (1980), Smith (1987) and Elliott and Worsley (1999) all reported ^{14}C dates that are statistically indistinguishable from each other over distances commonly between 1-2 m. Gamper (1983) reported four similar dates over a distance of 5.2 m from a lobe in Switzerland, while Ballantyne (1986) found that the ages of five samples along a 2.8 m -

long segment of buried organic horizon were indistinguishable from an age obtained ahead of the front, which he interpreted as being the AMRT of the modern soil. All these studies interpreted the closeness in age as evidence for increased solifluction activity and faster but continuous lobe movement, and further inferences were made of the climatic significance of these increased rates of movement.

Only the detailed stratigraphic study by Matthews et al. (1986) considered lobe collapse and colluviation as a dominant processes in the formation of solifluction lobes, as proposed in an early and more descriptive study by Price (1969). The authors observed washed sandy layers alternating between the various organic bands and over the main buried horizon. They believed that this colluvial material originated from two sources: from the erosion of fines by saturated overland flow during snowmelt, and by less frequent but greater magnitude events associated with the collapse of other lobes farther upslope. Because of this, they re-iterated the proposition of Price (1974) that the slope as a whole exhibits lobes that can be interpreted as representing different stages in a developmental cycle which is intimately linked to colluvial processes and lobe collapse. Apart from the current study, only Matthews et al. (1986) warned against the use of average rates of lobe advance, which are “hardly realistic if the underlying process is intermittent” (Matthews et al., 1986).

The interpretation of ^{14}C dates in relation to the perspective of a more “rigid” and episodic mechanism of lobe advance has led to an estimation of the magnitude of the time period required to complete a full developmental cycle. Unlike Price (1974), it is

believed that this internal cycle of development is recurrent in time, and that lobe collapse does not terminate the development of solifluction lobes. While snow accumulation may help to precipitate the collapse of the front, it is not believed to be the dominant factor (see also Hugenholtz and Lewkowicz, 2002). Deeper and earlier thaw-consolidation of the front, a thicker organic mat on the riser and sorting of coarse materials toward the edges of the lobe are inter-related factors that are part of a positive feedback system, which increases the stability of the front and concentrates the stress that will ultimately lead to the collapse of the lobe and its rapid advance onto the surrounding slope. Before the climatic significance of lobe movement may be discussed, therefore, additional information is needed on the other factors controlling the movement of lobes on the study slope.

5.4. ACTIVE LAYER THICKNESS AND GROUNDWATER LEVELS

Frost table and groundwater-table depths, measured between May 29th - July 13th 2002 at three different sites along the slope, are presented in Figure 5.32. The xeric site C, which was also monitored for surface movement, is located upslope in a dry and poorly vegetated area. The mesic site K, also monitored for surface movement, is located about two-thirds of the way down the slope and has abundant shrub and grass vegetation. The hydric site M is located in the lower, concave part of the slope, in a moist area with moss and shrub vegetation (see Figure 4.1 for precise location of sites). These three sites were chosen out of the thirteen monitored sites to illustrate the variations in frost and groundwater-table depths within solifluction lobes and across the whole slope.

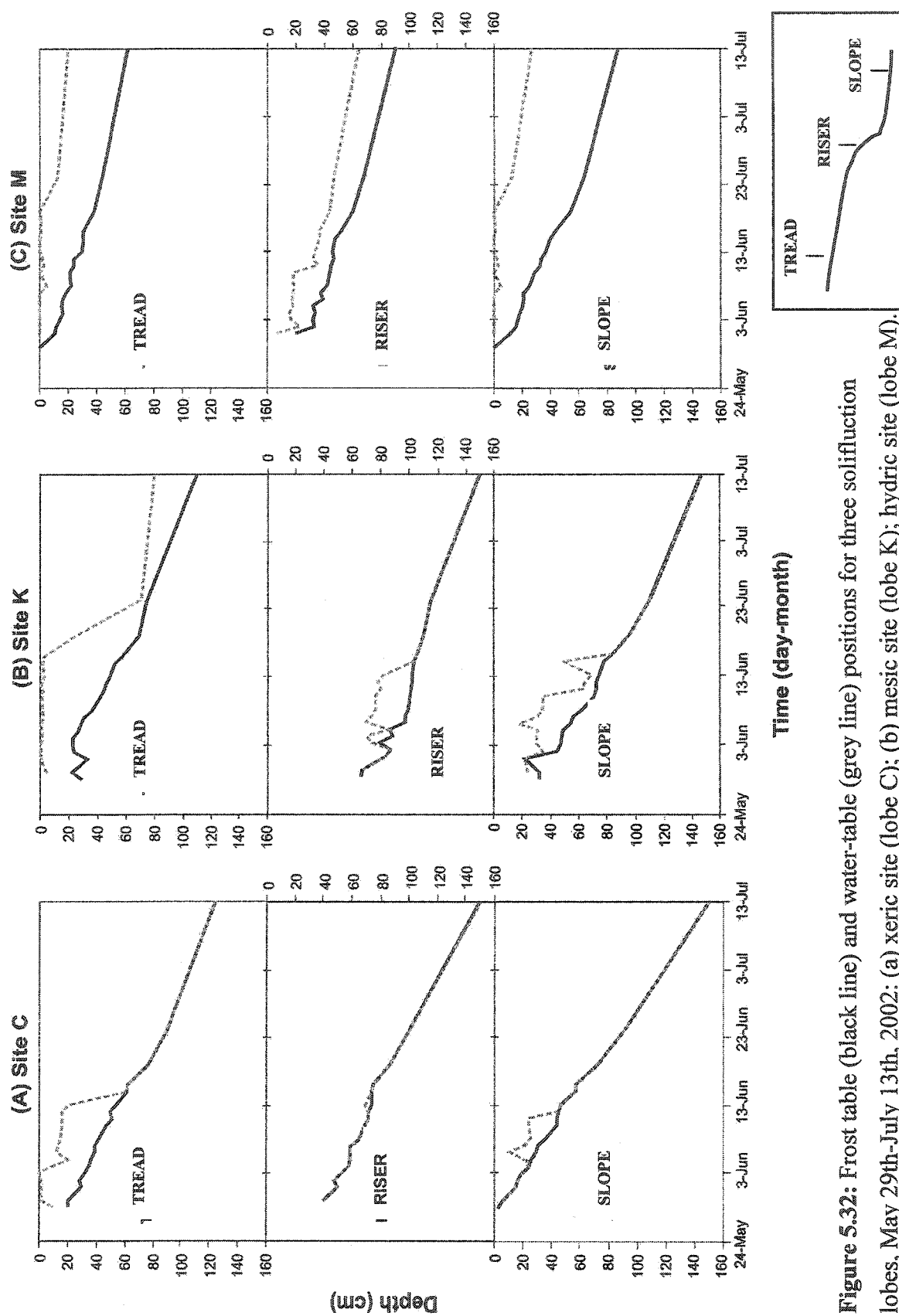


Figure 5.32: Frost table (black line) and water-table (grey line) positions for three solifluction lobes, May 29th-July 13th, 2002: (a) xeric site (lobe C); (b) mesic site (lobe K); hydric site (lobe M).

5.4.1. Intra-lobe variations

The depth to the frost table was consistently greater on the riser of lobes than on the tread over the course of the monitoring period. For most sites, the frost table depth downslope from the riser (slope) was similar to that on the riser, but in some cases where a snow-bank persisted for some time in the lee of the riser, thawing of the ground began later, and the slope had a shallower frost table. The water-table was high when ground thawing began, and it was consistently higher on the tread than on the riser. For some lobes (e.g., site M, Figure 5.32), the water-table rose again downslope from the riser. As thawing of the ground progressed, the thawed layer progressively drained and groundwater levels fell. While the fronts of many lobes dried out rapidly, water levels stayed relatively high on the treads.

5.4.2. Inter-lobe variations

5.4.2.1. Groundwater levels

When thawing of the ground began, the water-table was near the surface everywhere and saturated overland flow occurred throughout the slope in response to intense snowmelt. As thawing of the ground progressed and the active layer deepened, groundwater progressively drained down the slope, and the upper slope became increasingly depleted of water relative to the lower slope (Figure 5.33). Saturated overland flow continued downslope for some time after snowmelt where the water-table intersected the ground surface (Figures 5.33 and 5.34). The progressive downslope drainage of the active layer caused lobes on the lower segment of the slope to remain wetter for a longer period of time compared to lobes upslope. This is demonstrated by a significant downslope increase in the mean saturation level (the ratio between the thickness of the saturated

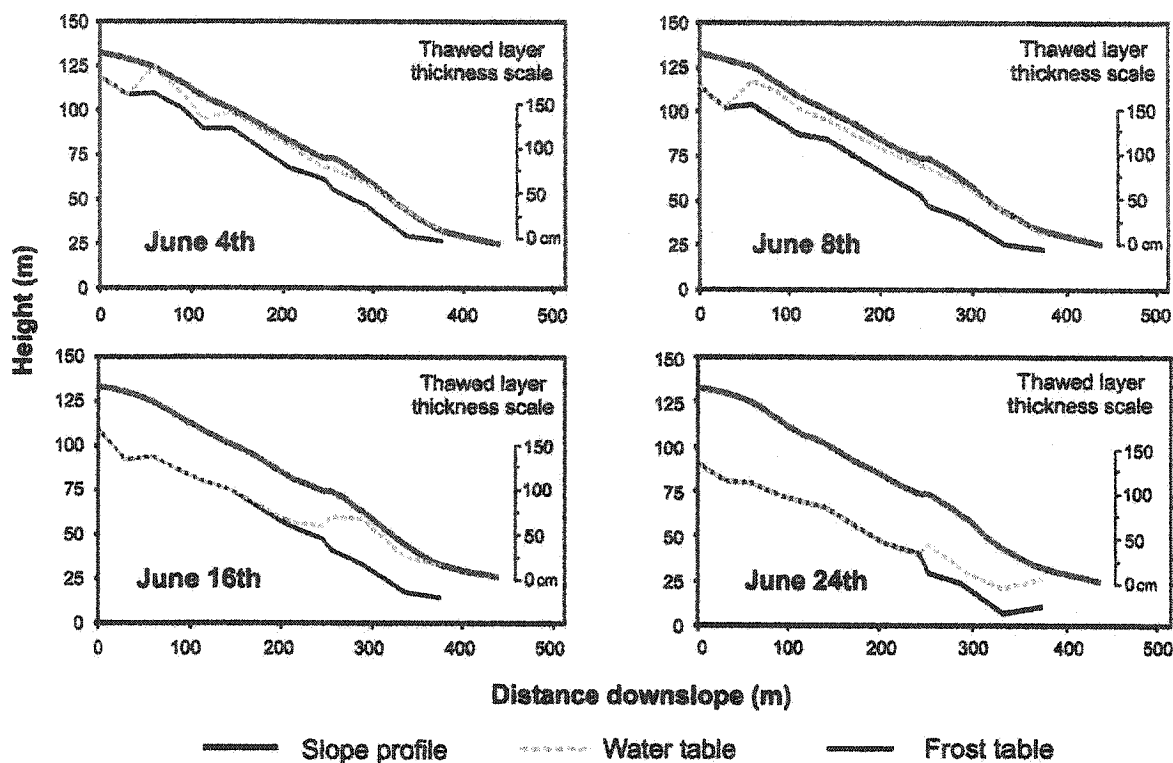


Figure 5.33: Profiles of the study slope showing the location of the frost table and water-table measured on the treads of 13 lobes, from June 4th to June 24th, 2002.



Figure 5.34: Example of saturated overland flow on the emerging surface of lobe J, on the lower slope.

layer and the thickness of the thawed layer, expressed as a percentage) for the monitoring period between June 2nd - June 24th (when data was available for all sites) (Figure 5.35). The graph also illustrates the difference in saturation within lobes, saturation being highest on the treads and lowest on the risers.

5.4.2.2. Active layer thickness

Although clear variations in thawed-layer thickness occurred within lobes, with the front being thawed more deeply than the tread, no downslope trends appeared in the thawed-layer thickness, unlike Hugenholtz and Lewkowicz (2002) who suggested that the late-summer thawed layer was shallower downslope. Although no significant downslope variation was found in thawed layer thickness between June 2nd - June 24th, the lowermost sites on the bottom slope did display shallower frost tables (e.g. site M, Figure 5.32). Since not all sites were probed for frost table depth on July 13th, it is impossible to assess with confidence the downslope variation in thawed layer thickness for that date.

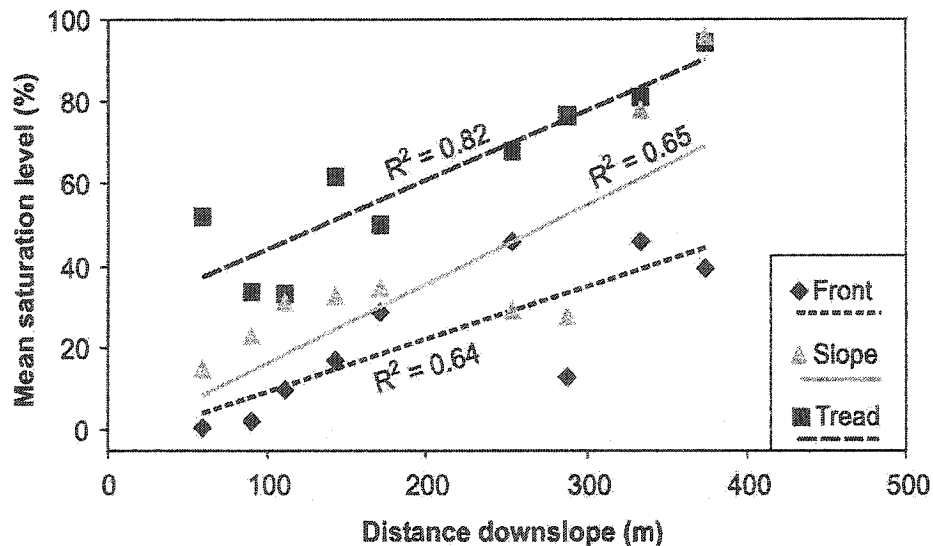


Figure 5.35: Scatterplot showing the average percentage of thawed-layer saturation against distance downslope.

Thermistors installed in August 2001 on the treads of 3 different lobes and near the weather station at the bottom of the slope continuously recorded soil temperatures at different depths (Figure 5.36), permitting the calculation of maximum thaw depth. The soil temperature shows progressive downward freezing beginning around September 15th with prolonged zero curtain periods at depths. Thawing of the ground began in May. Lobe K, which is the biggest of the three, was the first to thaw (May 5th) followed by the medium-sized lobe 8 (May 10th) and then by the smaller lobe 2 (May 23rd). This concurs with field observations, which showed that the treads of bigger lobes emerged first from the snowpack (Figure 5.37).

Soil temperatures at each successive depth display a progressive attenuation of surface fluctuations. The CALM site differs from the lobe sites, having generally less steep temperature gradients at depth, and considerably less variation in surface temperature. This is explained by an extensive moss cover, which acts as an insulating layer. The soil, in consequence, remains cooler in that part of the slope. The bigger lobes, #8 and K, located on the mid- and lower slope, have more pronounced variations in surface temperature, compared to the smaller lobe 2. This is probably explained by a thicker snow cover on smaller lobes, whereas bigger lobes, more exposed to wind, are likely to have a thinner cover of snow on their surfaces. For example, lobe 8 was probably virtually clear of snow throughout February and March 2002, based on the large surface temperature variations at this time. Hugenholtz and Lewkowicz (2002) reported a mean snow depth of 74 cm with a large coefficient of variation of 1.07 (n=205) for the study

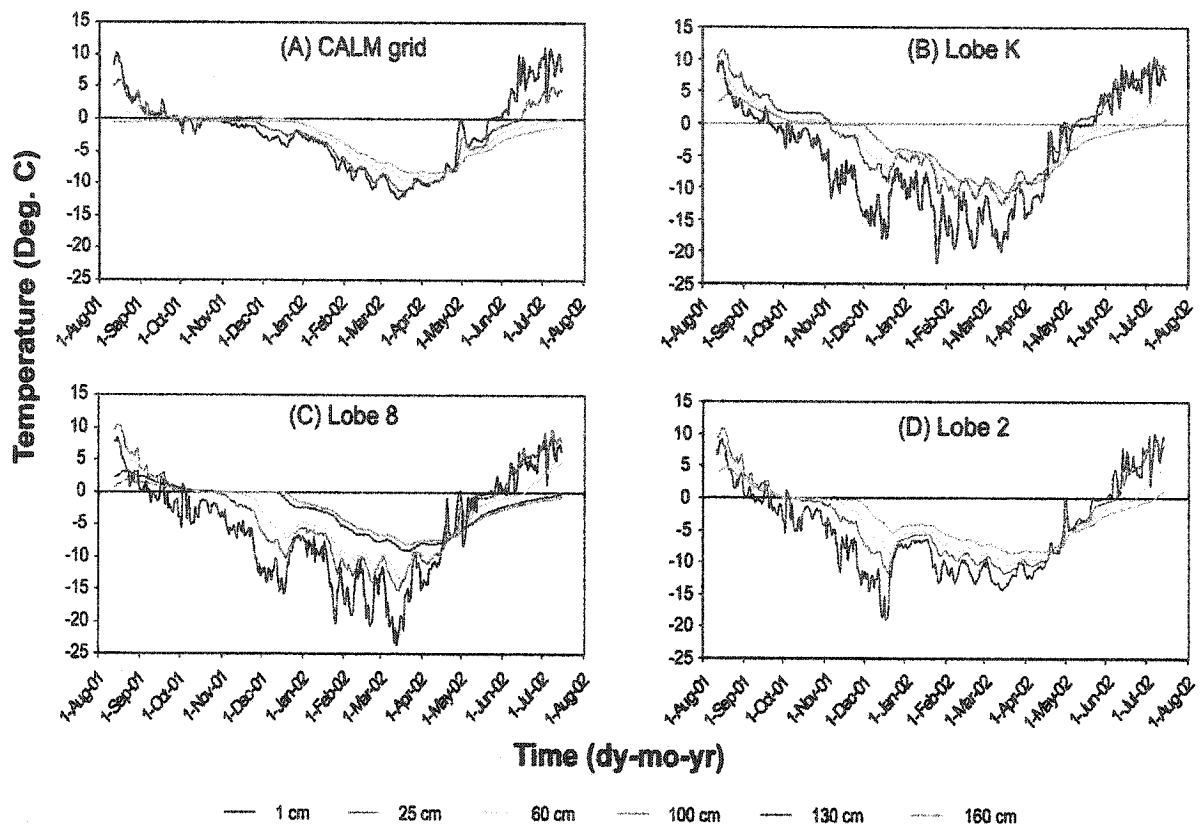


Figure 5.36: Daily averages of soil temperatures at various depths: (a) CALM grid site (lowermost site); (b) lobe K (lower slope); (c) lobe 8 (mid-slope); (d) Lobe 2 (upper slope).



Figure 5.37: Example of a medium-sized lobe emerging from the snowpack in the spring.

slope. This high variability in snow depth may be explained by the variations in lobe morphology, which influence the redistribution of snow on the slope.

Soil temperature was plotted against depth for the four monitored sites on August 15th, 2001 (Figure 5.38). This date was chosen because soil temperatures at depths were the warmest around that time. The depth to the frost table was obtained by extrapolating the temperature profile to 0°C using a second order polynomial equation. The three lobe sites display a strong negative inflection in their profile near the surface, which reflects the beginning of surface cooling. In these cases, the surface point was excluded from the trend analysis. The active layer thickness is seen to decrease downslope between the four different sites (Figure 5.39). The difference is most marked toward the bottom of the slope, where clear changes in vegetation type and moisture content occur. The moss cover insulates the soil from summer heat while the higher moisture content slows the thawing rate by raising the latent heat of the soil (Harris, 1981). On the upper slope,

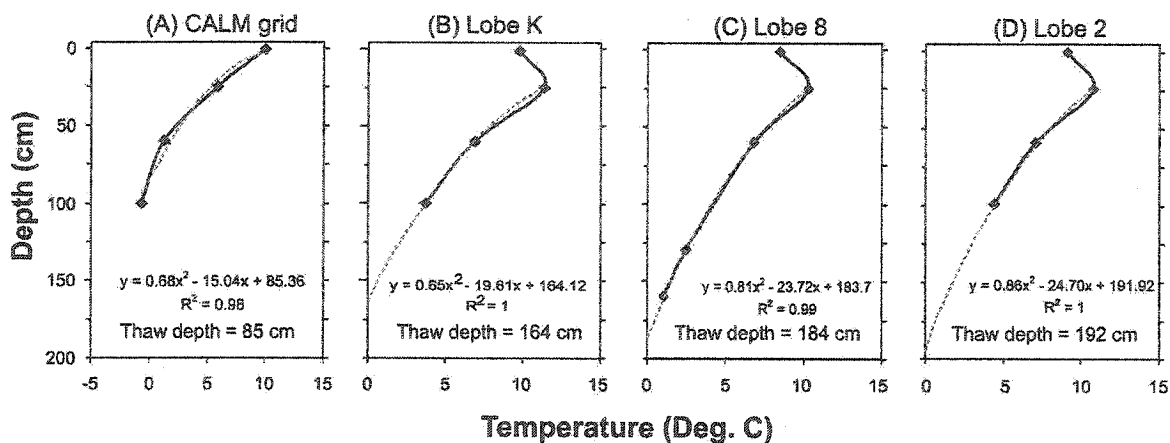


Figure 5.38: Vertical temperature profiles below the treads of three solifluction lobes and at the CALM site at the bottom slope, with corresponding active layer thicknesses calculated from extrapolation of each profile to 0°C (dotted line).

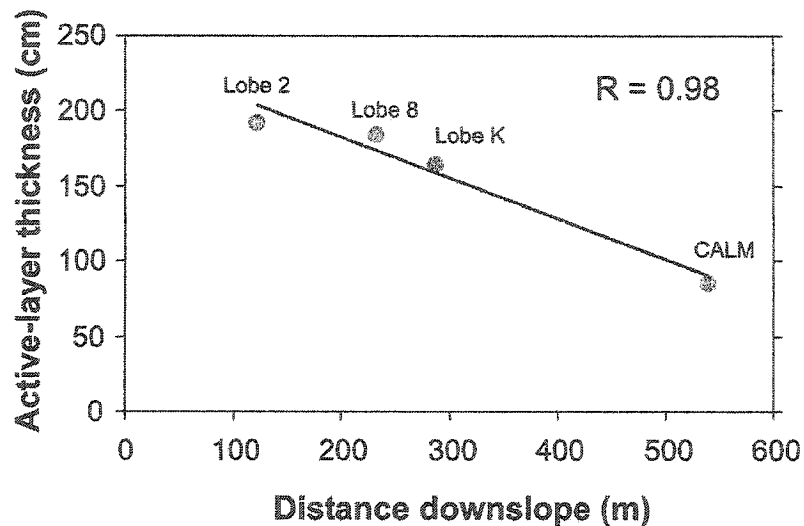


Figure 5.39: Scatterplot of active layer thickness at three lobe sites and the CALM site against distance downslope.

where the organic mat is the thinnest and the soil dry, thawing penetrates deeply into the ground.

5.4.3. Synthesis

Significant differences were found between thaw depths on the riser and on the tread of solifluction lobes, the riser having a deeper frost table than the tread. Price (1971) reported similar results in the nearby Ruby Range, where variations in active layer depth were found to be a response largely to the presence of vegetation and microtopography. Two reasons were advanced by Price to explain the deeper active layer beneath lobe fronts: the first was the higher angle of incidence of solar radiation and resulting warmer temperatures; the second was the vegetation cover, which consisted of low herbaceous species with little mosses to insulate the soil. Price also argued that the deeper active layer beneath the front leads to a better drained environment not conducive to mosses, while the shallower and wetter active layer beneath the tread is more favourable to the accumulation of mosses.

The results presented here certainly confirm the findings of Price (1971). It is argued, however, that it is probably because the lobe front is more compacted and consolidated that the water-table is kept high on the tread. The front starts to thaw earlier, as it often emerges first from the snow, and the higher angle of incidence of solar radiation leads to a more rapid thaw. As a consequence, the front is already consolidated when thawing of the tread occurs at a given depth. The hydraulic conductivity of the consolidated front is reduced, and the water-table can stay closer to the tread surface behind the front, which acts as a sort of dam.

This marked variation in groundwater levels within solifluction lobes was demonstrated by Benedict (1970) in the Colorado Rockies, where deep water-tables were found in standpipes on the riser and sides of a turf-banked lobe, and shallow water-tables were encountered in standpipes in the central part of the lobe. Benedict (1976) proposed that the progressive thickening of the soil by solifluction modifies the hydrologic conditions across the lobe. He further emphasized the relationship between high water-table depths in the fall and more intense winter frost heave and spring thaw-consolidation. Thus a positive feedback system exists where the lobe develops by progressive steepening of the front, which leads to increasing differences in thaw-depths and water-table depths between the front and the tread. More rapid and deeper thawing beneath the front leads to the accentuation of relative consolidation and segregation of water behind the rigid, consolidated front. Differences in water levels between the front and tread further promote relative frost heave activity and related downslope movement, until lobe development reaches the critical state of stability and collapse occurs.

At the slope scale, evidence was brought for a broad downslope decrease in active layer depth, and of a general downslope increase in active layer saturation throughout the study period. Downslope thinning of the active layer with a corresponding increase in groundwater levels was reported by Woo and Steer (1983) in the High Canadian Arctic, Hinzman et al. (1993) in Alaska and Carey and Woo (2001) in the southern Yukon. Moisture redistribution causes catenary effects on slopes, where fine sediments are removed from the upper slope and accumulate in the concave, basal part of the slope (e.g. McCaig, 1984). Moisture redistribution on slopes underlain by permafrost controls vegetation colonisation, and both factors act to control the thermal regime of the soil and the resulting active layer thickness. Woo and Steer (1986) recognized three general slope conditions in continuous permafrost areas: (1) interfluvies and convex slope segments where the winter snow is thin, surface flow is rare and only a part of the soil is saturated for a short time; (2) mid-slope areas where overland flow and rill flow occur during snowmelt, but which dry out when snowbanks are depleted; and (3) vegetated strips and patches of standing water downslope which indicate a high water-table during most parts of the thawed period.

It is thought here that moisture distribution in the thawing active layer is the major control on spatial variations in solifluction rates and lobe morphology. The general model for sub-surface flow on a slope underlain by an impermeable layer predicts that during long wet periods, the discharge in the soil layer tends towards a steady state in which discharge is proportional to slope length (from divide), and so increases downslope (Whipkey and Kirkby, 1978). It has been shown here that groundwater drains

progressively from the top towards the bottom of the slope, which remains wetter for a longer period of time. It is proposed that moisture distribution downslope has two dominant effects on solifluction:

(1) *During autumn freeze-back:* The spatial distribution of groundwater in the active layer determines the amount of frost heave activity on the slope, and the location of ice lenses within a specific soil profile. The precipitation regime during freeze-back of the active layer thus represents a climatic control on the intensity and spatial variability of solifluction rates. In dry years, it is possible that the upper slope remains inactive, while frost heave activity allows the wetter basal slope to move by solifluction.

(2) *During spring ground-thaw:* constant moisture supply from upslope must retard the expulsion of water from the pores of the consolidating soils downslope. Saturated sub-surface flow will maintain the saturation of the newly-thawed soil pores, thereby maintaining the high pore-water pressures in the thawing soil. Experimental work by Harris et al. (1995) and field observations by Harris (1977) and Matsumoto et al. (2001) have demonstrated that infiltration of meltwater from the surface of the thawing soil is unlikely to occur because of the presence of an upward hydraulic gradient in the soil. Saturated sub-surface flow, on the other hand, can sustain the high water content and related pore-water pressures, thereby augmenting the chance for solifluction movement to occur.

5.5. CONTINUOUS MONITORING OF A SOLIFLUCTION LOBE

5.5.1. Time series of meteorological parameters, soil temperature, soil deformation and moisture content

5.5.1.1. *Air and soil temperatures*

Results from continuous recording of soil movement, temperature and water content on Lobe I, and of air temperature and precipitation for the entire study slope between May 25th – July 17th 2002 are presented in Figure 5.40. Air temperature, recorded at hourly intervals, exhibited clear diurnal cycles with amplitudes between 5-15°C. The monitoring system, operational on May 25th, began recording soil movement and related parameters when daily air temperatures had already risen above 0°C and thawing of the surface had been initiated a few days before. Consequently, temperatures recorded at the soil surface at the beginning of the monitoring period were close to, or slightly above 0°C.

The time series of soil temperatures at depths shows the progressive downward penetration of the thawing front, which passed the final thermistor, at 95 cm-depth, around July 3rd. Diurnal variations in air temperature and incoming radiation are reflected at the soil surface, but progressively attenuate at depth, being present at 55 cm depth but absent at 95 cm. These daily cycles were such that the temperature frequently went below 0°C overnight throughout the study period, and thus the soil surface was affected by diurnal freeze-thaw cycles in its uppermost 5 cm. These surficial freeze-thaw cycles persisted from May 25th until mid-June, after which the temperature at the soil surface rose progressively toward a maximum average daily value of 10°C on July 17th.

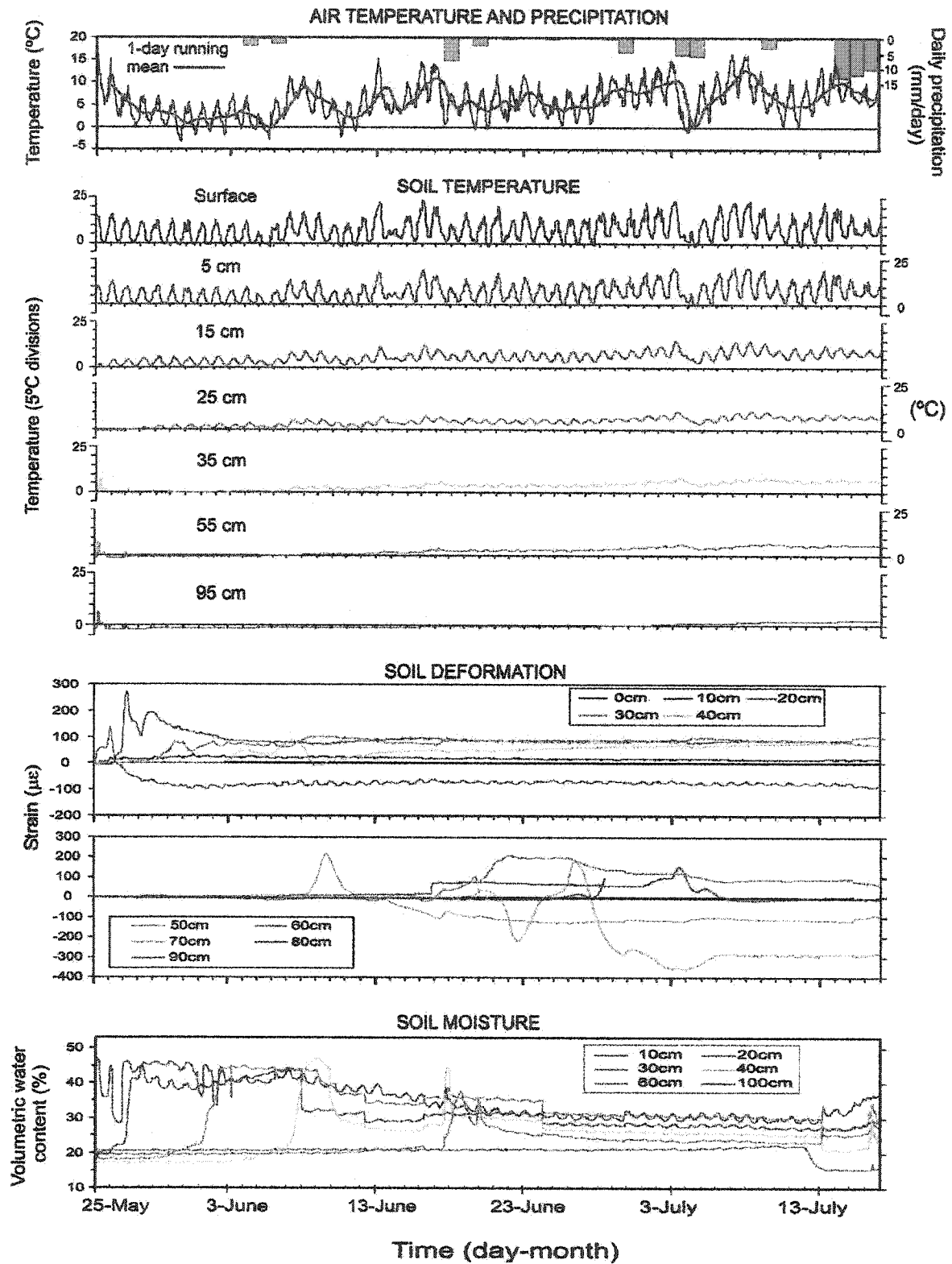


Figure 5.40: Time series of hourly air temperatures with daily precipitation at the study site, and hourly deformation, temperature and water content of soil at several depths beneath the surface of lobe I.

5.5.1.2. Soil moisture

The unfrozen volumetric water content was initially low (~20%) for all the sensors, but rose rapidly to a maximum of 45 % in response to the passage of the thaw front. At first the water content rose slowly and gradually, and then reached an inflection point from which it increased abruptly to its maximum value (Figure 5.40). The rise was most pronounced for the four uppermost sensors (10, 20, 30 and 40 cm) and was less at 60 cm depth, while at 100 cm, the water content dropped suddenly on July 12th. This behaviour is the result of a change in the dielectric constant of the soil (ϵ) as it changes from frozen to thawed state. The dielectric value of a three-phase system comprised of aggregate particles, water, and air depends upon the volumetric percentage and dielectric values of each constituent (e.g. Guthrie and Scullion, 1999). In its frozen state, the various components of the soil have similar dielectric constants (air ~1, soil ~3-7, ice ~3), and the probe is not able to discriminate water (ice) from soil and air. As the thaw front approaches a sensor, the temperature rises and the unfrozen water content increases, causing a slow and gradual rise in water content recorded by the probe. Once a soil layer is completely thawed, free water ($\epsilon \sim 81$) originating from thawing ice lenses and/or from the overlying thawed soil, raises the dielectric constant of the soil. At this point, the probe records an abrupt increase in (liquid) water content.

The correspondence between the timing of soil water release and the passage of the 0°C isotherm was not perfect because the string of thermistors was placed 1 m away from the moisture probe, and 50 cm away from the strain probe, on a slightly higher surface. On the other hand, the depth of the thaw line obtained by repetitive probing of the frost table

around the moisture probe confirmed that the recorded increase in soil moisture directly coincided in time with soil thaw at a given depth (compare Figures 5.40 and 5.41).

From May 25th to around June 8th, water contents remained high after passage of the thaw line, particularly in the near-surface layers (Figure 5.40). This suggests that during this period, thawing of the soil occurred in a saturated environment. This is confirmed when looking at the groundwater levels measured on the tread of adjacent lobes J and H, which show that the thawed layer was saturated to the surface until about June 7th, after which it started to drain progressively over time (Figure 5.41). Groundwater levels from lobes J and H were used for comparison because no measurements were made on lobe I, to avoid disturbing the monitoring station. Thawing of the soil at 40 cm depth on lobe I caused an abrupt rise in (liquid) water content on June 7th (Figure 5.40). However after three days, this water drained rapidly, possibly downward to a newly thawed soil layer.

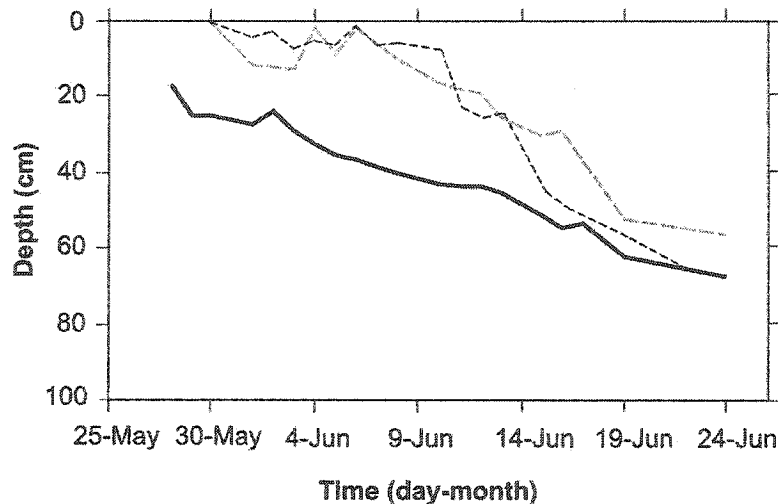


Figure 5.41: Thawed layer depth measured by probing on the tread of lobe I (black line), and groundwater levels measured on the treads of lobes H (black dotted line) and J (light dotted line), located 40 m upslope and 10 m downslope from lobe I, respectively.

Indeed, Figure 5.41 shows an abrupt decrease in water level beneath lobe H around June 10th, suggesting that at this time, lobe I may have become unsaturated at 40 cm depth.

Three rain storms clearly affected the soil water content. The first one saw 2.8 mm of rain falling on June 4th, followed by 2 mm on June 26th. These two rain events caused the water level to rise to the surface of lobes H and J (Figure 5.41), but because the soil was saturated at 10 cm depth, the probe did not respond to these events. The second storm occurred between June 18th and June 20th : 8.4 mm of rain fell on June 18th and 2.5 mm on June 20th. The first event caused an abrupt but temporary rise in moisture at 40 cm depth, immediately followed by an abrupt and more sustained rise in moisture at 60 cm depth (Figure 5.40). Small, instantaneous rises were also observed closer to the surface (10-30 cm). At 60 cm depth, following the first rain event, the moisture content started to drop but rose again in response to the second rain event on June 20th, before dropping again progressively over time. It is suggested that the June 18-20th rain event caused the temporary saturation of the soil at 40 cm depth, and precipitated the thawing of the soil at 60 cm by rapidly transferring heat to the frozen soil. Accordingly, Figure 5.41 shows that the thaw depth on June 18th was indeed around 60 cm depth, while the water table was around 40 cm depth on lobe J, and a bit lower on lobe H. The third rain storm, which occurred between July 14th and July 16th, was the most important of the record, with a total of 35.8 mm falling on the slope in three days. Although increases in moisture were observed at most sensors, especially on July 16th, the rises were not substantial, and the moisture content even dropped slightly at 40 cm depth (Figure 5.40). The thaw front at

that time was below 100 cm and it is suspected that the over-consolidated soil at that depth may have been drawing moisture from above.

The 0°C isotherm was around 95 cm depth on July 3rd. Apart from a small and gradual increase in water content, the probe responded to thaw with an abrupt decrease in moisture at 100 cm around July 12th. Two explanations are proposed for this behaviour:

- (1) the soil was overconsolidated and desiccated, and on thawing may have swelled slightly, thereby augmenting the relative proportion of air in the soil and reducing its dielectric value (and corresponding moisture content);
- (2) some water was present in a free state within the frozen soil, and on thawing quickly became adsorbed to the over-consolidated soil, thereby reducing the dielectric value of the soil.

Whatever explanation is favoured, it is likely that ice content around 100 cm depth was very low, and that the soil was desiccated and over-consolidated because of upward migration of moisture during downward freezing of the soil the previous fall (Nixon and McRoberts, 1973; Smith, 1985). Hence, if moisture release in the top soil (0-40 cm) was greatest and sustained, favouring solifluction movements, thawing at depth did not result in significant moisture releases, as moisture contents were probably initially low and drainage occurred rapidly in an unsaturated environment.

Another conspicuous feature in the soil moisture time series is the presence of distinct diurnal variations (Figure 5.40). These diurnal cycles are most developed near the surface and attenuate at depth, being closely associated with diurnal cycles of soil

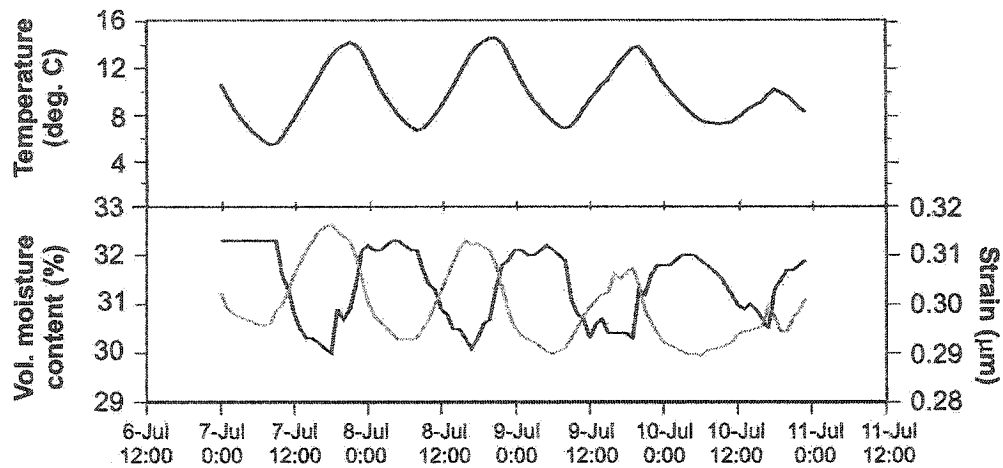


Figure 5.42: Examples of diurnal cycles in soil temperature at 15 cm depth (black line), soil moisture (light grey) and soil strain (dark grey) at 10 cm beneath the surface of lobe I, July 6th- 11th, 2002.

temperature and inversely related to diurnal cycles in soil strain (Figure 5.42). While some of these diurnal variations in moisture and soil strain may be related to near surface diurnal freeze-thaw cycles, it is evident that they also occurred when temperature fluctuations remained well above 0°C. As temperature decreased during the night so did the moisture content, but strain increased as positive movement was recorded by the strain probe. Matsuoka (1996) found a nocturnal decrease and a daytime increase in matric potential corresponding to fluctuations in near-surface temperatures. He thought that such variations reflected water movement resulting from temperature gradients in the unfrozen soil. This explanation is rejected here, because diurnal variations in moisture can be observed throughout the soil profile while reversal of the temperature gradient occurred only in the uppermost 15 cm of the soil; secondly, because these moisture variations were almost synchronous between depths, which would not occur if water was drawn from one layer to another. In an earlier study of soil creep in Western Alberta, S. Harris (1973) reported concurrent diurnal fluctuations between positive temperatures at

the soil surface and rates of creep. Retrograde movement occurred at night while positive movements occurred during the day. Simple thermal expansion of the soil was discarded as the cause for these cyclic variations and the phenomenon was left unexplained.

In recent experiments on the effect of temperature on soil bulk dielectric permittivity measured by time domain reflectometry, Wraith and Or (1999) and Or and Wraith (1999) found that temperature has competing effects on the dielectric response of a soil-water mixture. A rise in temperature lowers the dielectric constant of bulk water in a soil, but it increases that of bound water. Hence the relative proportions of bulk versus bound water controls the aggregate dielectric response to temperature for a given combination of soil and wetness (Or and Wraith 1999). Therefore finer soils/lower water contents favour an increase in the dielectric constant, with a corresponding increase in the derived volumetric water content.

In the present case, the relationship between temperature and moisture fluctuations is positive. This must mean that the fine fraction of the soil holds significant amounts of bound water, part of which is liberated as temperature increases during the day (Wraith and Or, 1999). In return, this release of bound water lowers the cohesion of the soil, which swells in response to lower adsorbing forces (e.g. Selby, 1999). The response of the strain probe to these diurnal swell/shrink cycles depends on its angle normal to the slope (Sato et al., 1997). In this case, the probe had an angle of about 7° down from the axis normal to the slope. Consequently, daily swelling of the soil resulted in negative

displacements of the probe (upslope) while shrinking at night caused the probe to move downslope. However, there was little or no net movement.

5.5.1.3. Soil deformation

The progressive downward penetration of the thaw front resulted in deformation recorded by the strain probe at successive levels (Figure 5.40). In the upper 40 cm of the soil, the probe responded in a similar way to the passage of the thaw front, with an initial abrupt bending downslope (positive strain) followed by a partial and progressive upslope recovery (upslope movement). Net negative movements occurred at the surface of the soil. This may be explained in part by the fact that some of the positive movement was missed, since thawing of the surface had already begun before measurement was initiated.

Deeper in the soil (50-80cm), the response of the probe to thaw was generally similar, with positive strains generated by the passage of the thaw front, followed by a more gradual upslope recovery which in some cases, curiously ended in significant negative net displacements (e.g. at 50 and 70 cm depths, Figure 5.40). Strain values however, were high and more variable. The signal at 80 cm depth was interrupted on June 28th, because strain became higher than permitted by the data-logger configuration. While differential movement may be locally high at depth, particularly in areas where two-sided freezing leads to higher ice concentrations near the bottom of the active layer, increasing strain with depth is contradictory to previous field and experimental studies. The onset of deformation still coincided with the passage of the thawing front, but the transformation of strain into real displacement values, using the method of Yamada and Kurashige

(1996), yielded strange results (Figure 5.43a). The time-series of probe movements shows fluctuations in surface movement of almost 10 cm over a one day period. This is very unlikely, since surface markers do not corroborate the movements. Two explanations for this abnormal behaviour are possible:

- (1) Macro-shearing of the lobe;
- (2) Subsidence effects on an already bent probe.

The first explanation is unlikely because there is no reason for the negative recovery of the probe following a shearing event at the lobe scale. Instead, the second option appears more suitable to explain these high amplitude fluctuations. Strain values obtained immediately after placing the probe inside its borehole revealed that it was already bent in its lower sections despite using a bamboo stick fixed to the lower end of the probe to push the probe straight in the hole. In addition, incomplete backfilling of the space between the probe and the surrounding hole may have left voids around the probe, particularly in the lower section of the hole where the wet slurry was difficult to compact. As a result, thaw-induced stress may have concentrated in void zones resulting in abnormal bending. Moreover, because of the initial probe configuration, which showed significant initial strain in its lower section (60-90 cm depth), subsidence of the soil by thaw-consolidation was likely responsible for much of the movement observed. This problem was first identified by Matsuoka (1994) who emphasized that a strain probe will bend not only due to the downslope soil movement but also through contraction (subsidence) and expansion of the soil. The higher the angle between the probe and the axis perpendicular to the soil surface, the more sensitive is the probe to subsidence effects.

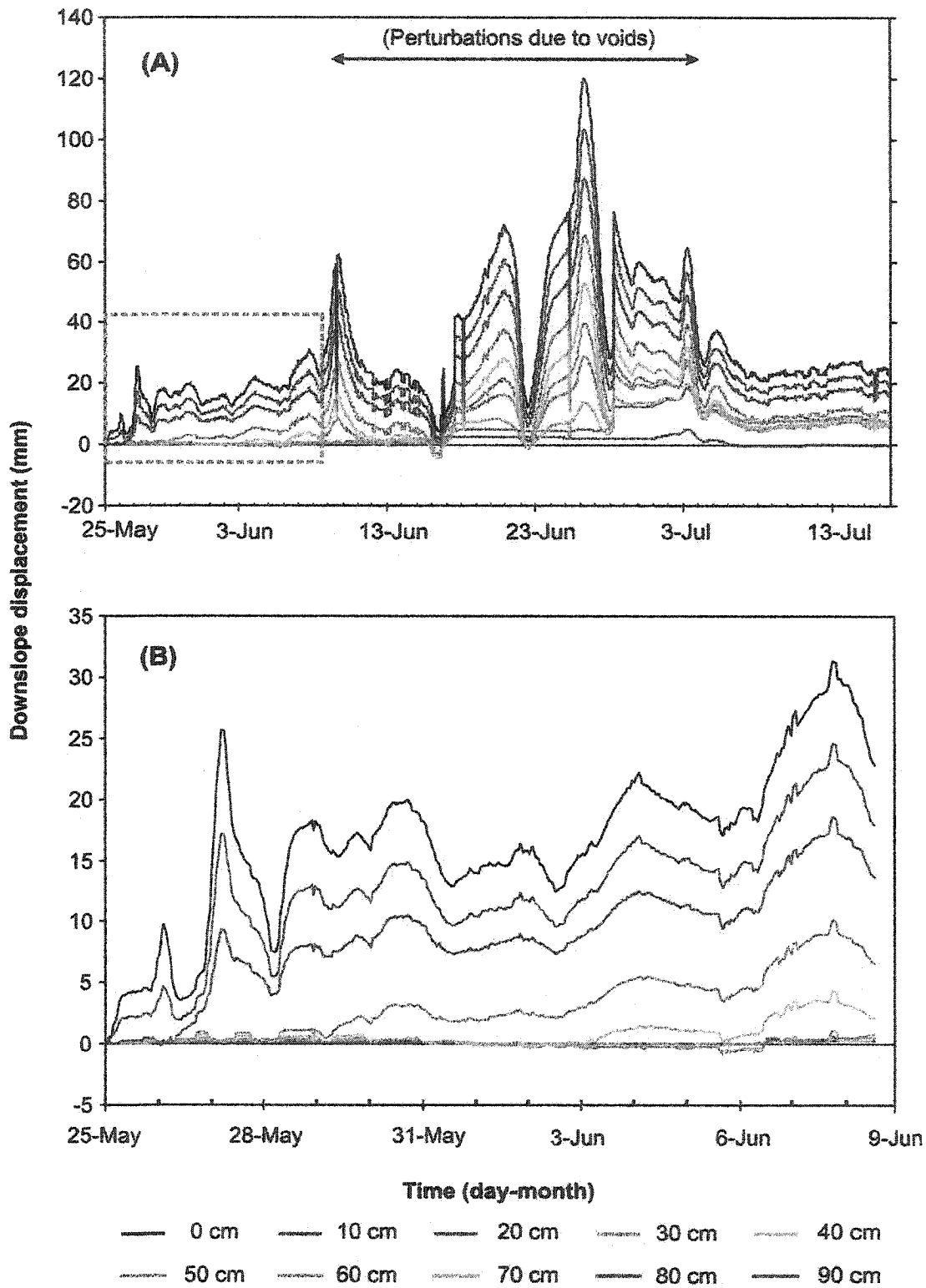


Figure 5.43: Hourly displacements of the probe at various depths, calculated from strain variations: (a) complete series, May-25th - July 17th, 2002 (b) inset, May-25th - July 17th, 2002.

Since the upper probe was backfilled properly and was set straighter, more confidence is placed in its results. Strain variations in the lower section (50-90 cm), can be used only to indicate the timing of movements. However, after July 8th, once thawing had proceeded below the base of the probe at 100 cm depth, calculated displacements between 50-90 cm depths went back to more normal values, suggesting that the abnormal strains developed at depth were indeed caused by warping of the probe inside the poorly-backfilled borehole. Displacements observed around July 8th, after thawing occurred at 100 cm, are only slightly higher than those recorded around June 9th, before the apparent perturbations occurred. Consequently, the time series of probe displacements was truncated at the time when thawing began at 50 cm and abnormal strains developed (Figure 5.43b).

Abrupt, positive changes in strain at the thaw front during seasonal ground thaw were reported by Matsuoka (1994, 1998b) and Sato et al. (1997) in Japan, and by Matsuoka et al. (1997) in the Swiss Alps. Yamada et al. (2000) recorded abrupt negative changes in strain on thawing in northern Japan, and the partial reversal of probe displacements was reported previously by them and Auzet et al. (1986). This probably reflects the elastic character of the probe which tends to regain its original form, and because the probes used here are not entirely flexible, it is possible that strain may be transferred between different depths. It is thus concluded that part of the recovery is due to the elastic behaviour of the probe in a low stress environment (viscous conditions) but also by the abrupt downslope movement of the next section down, which transfers negative strain to the overlying sections of the probe.

Following the passage of the thaw front, the gravimetric water content in the upper 40 cm of the soil ($\approx 0.45 \text{ cm}^3 \cdot \text{cm}^{-3} / 1.31 \text{ g} \cdot \text{cm}^{-3} = \approx 34\%$ mass of soil) corresponded to the mean value of liquid limit calculated from 13 lobes (34%). Thus following thaw, the upper soil was at or close to a viscous state. Despite this, there is no indication that strain increased with time, as would be expected from a viscous body under constant stress. Instead the probe data indicate rapid downslope movements, restricted in time, and coincided with thawing events, with almost no further movements afterwards other than slow negative recovery or a progressive decrease in positive strain. Two mechanisms may have caused the rapid movement:

- (1) Viscous flow, following the release of water from melting ice lenses which raised the water content above the liquid limit of the soil.

Because strain decreased or reversed while water content remained constant (and above the liquid limit of the soil), a change in viscosity cannot explain the changes in probe movement. One explanation for not having sustained viscous movement is that the viscosity of the soil decreases toward the lobe front, which at some point is rigid enough to hold back the flowing soil behind. Hence an initial phase of viscous flow may occur on thawing until the soil cannot flow further downslope and the soil consolidates. In the field, very wet and soft soil was often encountered on the treads of lobes while lobe risers were rigid enough to stand on without disturbing the soil.

- (2) Micro-shearing near the interface between the soil and thawing ice-lenses.

From this perspective, the effective stress is reduced by thawing of excess ice along discrete and irregular shear planes and rapid, but localized translational movement follows (Carson and Kirkby, 1972; Williams and Smith, 1989). Again, if sustained high

water contents are taken to reflect conditions of low effective stress, some resistance must be encountered downslope to explain why the rapid movement ceases. Movement of the soil under these conditions represents the integration of several small-scale displacements where shear strength is locally overcome by stress, a mechanism apparent in creep movements in temperate regions (e.g. Kirkby, 1967; Carson and Kirkby, 1972).

The strain rate behaviour is not unlike the creeping behaviour of soils, in a rheological sense, where the soil deforms at low stress (as a viscous fluid) but the strain rate decreases with time (Singh and Mitchell, 1968; Van Asch et al., 1989). Although this behaviour is more like that of a “pseudo-plastic material” (e.g. Allen, 1997), a Bingham plastic model, in which the material deforms as a Newtonian fluid once the residual shear stress of the soil has been overcome, has been used to model the slow creep of soils (Van Asch et al., 1989; Van Asch and Van Genuchten, 1990). However, experiments by Harris et al. (1995, 1997) showed that movement occurs predominately below the residual yield strength of the soil. As recognized by Van Asch (1989), there is probably no yield stress for soil creep and movement occurring below the residual yield strength is especially important to the analysis of long-term geomorphological processes.

The best general model for soil creep (including solifluction) is probably that of Allen (1985), which assumes an apparent viscosity which decreases with depth, and a shear stress which increases with depth. This model gives S-shape velocity profiles, as described by Kirkby (1967) and reported by various authors for periglacial regions (e.g. Benedict, 1970; Smith, 1987; Price, 1973, 1991).

5.5.2. Profiles of soil movement and moisture content

Profiles of soil displacement and of soil moisture have been calculated from the time series and are presented in Figure 5.44 for six different dates. Five profiles were calculated for the thawing period between May 25th – June 8th, when most of the surface movement was recorded by the surface markers (A-E, Figure 5.44). The period when thaw depths of 50-90 cm occurred was avoided because results would have had no significance due to the probe problems described above. A last profile was calculated for July 8th, once thawing had proceeded below the depth of the probe and strains released during the problem period had stabilized. This profile is considered to represent the depth-integrated displacement over a depth of 100 cm below the surface for the thawing period between May 25th - July 8th. Figure 5.44 also shows soil temperatures and moisture content summarized by isolines for the whole monitoring period.

The profiles generated are predominantly rectilinear with slight convexities near the surface and slight concavities at depth (Figure 5.44a). The profile on July 8th displays some irregularities in its lower segments, which may reflect warping of the probe during the preceding perturbation period (F, Figure 5.44a). The high differential movement at depth may reflect a local increase in movement, but since the thaw front was near the bottom of the probe, it may also reflect the transfer of stress from the section immediately above downward to the tip of the probe. This is also suggested by the quasi-limited moisture at 100 cm depth.

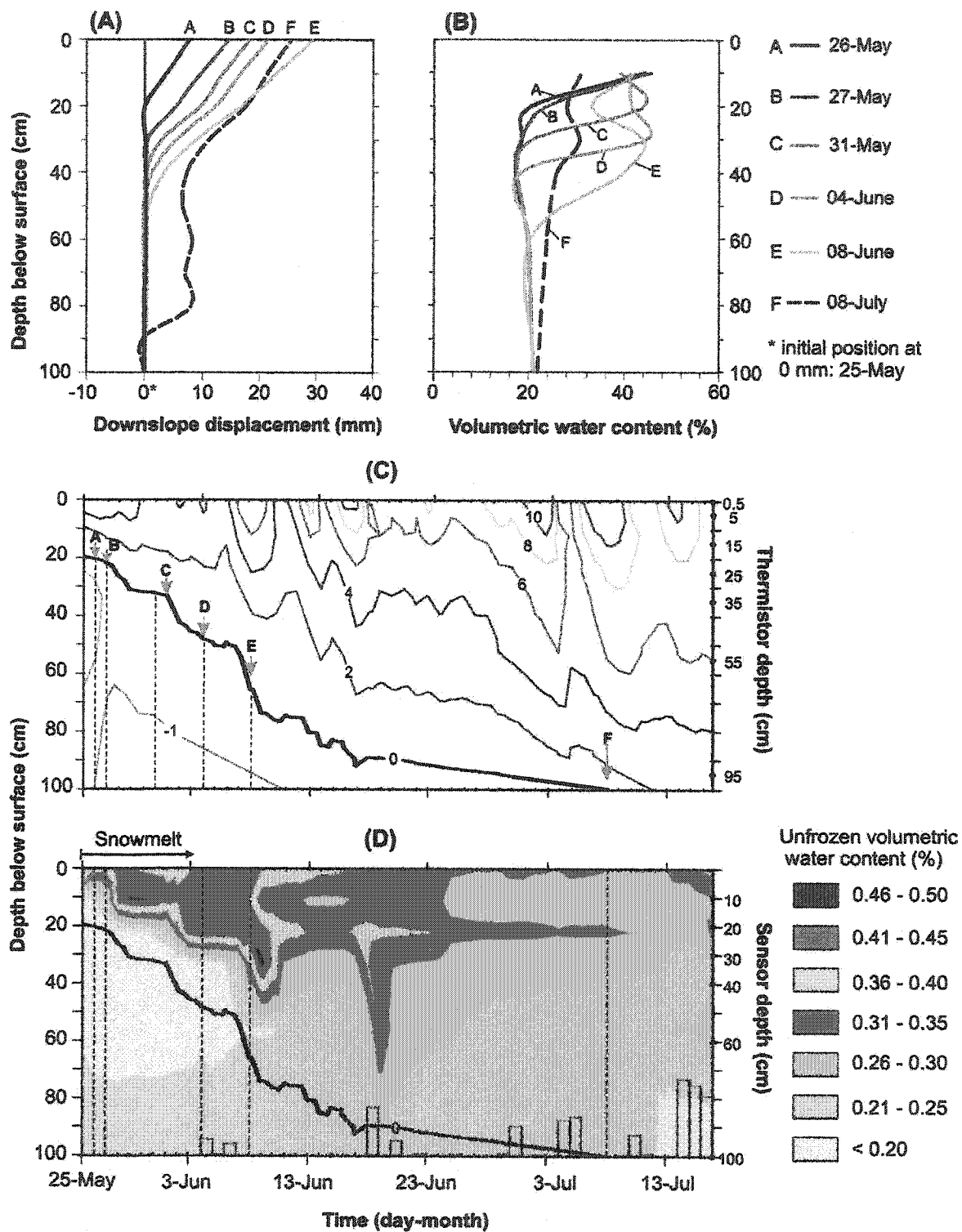


Figure 5.44: Profiles of (a) soil movement and (b) moisture content on six different dates; (c) subsurface isotherms and (d) contours of volumetric water content for the thaw period between May 25th – July 17th, 2002. The 0°C isotherm is shown in (d) along with bars representing precipitation events.

In a general way, movements decreased with depth and increased with time, with a net cumulative surface movement of 2.8 cm. Profiles of moisture content at depth, presented in Figure 5.44b, show the increase in moisture associated with the downward passage of the thaw front (Figure 5.44c), with values changing from $\approx 20\%$ (frozen) to $>40\%$ of soil volume once thawed. High moisture contents (darker colours) broadly follow the thaw line as defined by the 0°C isotherm (Figure 5.44d), and pockets of higher moisture sometimes develop above it, giving way to momentary upward hydraulic gradients. In general, moisture content broadly decreases with increasing thaw depth/time, as does solifluction movement.

5.5.3. Synthesis

Analysis of the time series of soil movements, temperatures and moisture contents during the thaw period between May 25th – July 17th has led to several conclusions:

- (1) The passage of the thaw front through successive depths causes abrupt releases of water from the melting soil, and synchronous abrupt downslope movement.
- (2) Soil movement occurs mainly at the thaw line, with little downslope movement afterwards. Upslope recovery may be attributed to the redistribution of stress from the section of the probe being thawed to the overlying section already thawed. Downslope movement thus appears directly related to phase-change consolidation where rapid but limited movement occurs, either by viscous flow or localized shearing along ice-lens boundaries.
- (3) Soil freezing is thought to be mainly responsible for conditioning the active layer to thaw-induced downslope movements. The rate of freezing and moisture supply determine ice concentrations and the intensity of ice segregation at depth.

Groundwater levels (and thus position on the slope) are thus critical in controlling moisture supply to the freezing front and resulting ice distribution throughout the soil. During ground thaw, increasing effective stress with depth leads to a general decrease in void ratio and moisture content, with a resulting increase in shear strength. The combination of lower ice content and larger shear strength at depth leads to reduced consolidation and reduced downslope movement on thawing. This results in a general pattern of decreasing rate of surface movement over time/depth.

6. DISCUSSION AND CONCLUSION

6.1. ENVIRONMENTAL CONTROLS ON SOLIFLUCTION RATES AND LANDFORMS

6.1.1. Relationships between environmental variables

Relationships between environmental variables measured on 13 excavated solifluction lobes were tested using bivariate correlation. Because the number of lobes sampled was small, no attempt was made to normalize the data. Consequently, only the highest correlations of interest are presented. The variables examined are summarized in Table 6.1. Downslope distance was included to check for any spatial dependence of environmental variables. Despite the small sample, some interesting relationships emerge between grain-size parameters. As the mean grain-size decreases (increasing mean ϕ), so does the standard deviation, while the skewness of the distribution decreases (Figure 6.1a, b). Although this might be the result of random variations in grain-size distribution, the redistribution of fines along the slope by surface wash and sub-surface flow could be the cause of these changes. From this perspective, it appears that the fines are added to an existing "background" distribution leading to a less skewed distribution with a larger variance.

Table 6.1: Environmental variables tested for bivariate correlation.

Charac- teristic	Grain-size		Atterberg limits	Organic mat thickness	Water content	Downslope Distance
Variables	% gravel	Mean ϕ	Liquid limit (LL)	Riser	w (%) (24-Aug-01)	Y (m)
	% sand	St_dev. (Sd)	Plastic limit (PL)	Tread		
	% fines	Skewness	Plasticity index (Pi)			

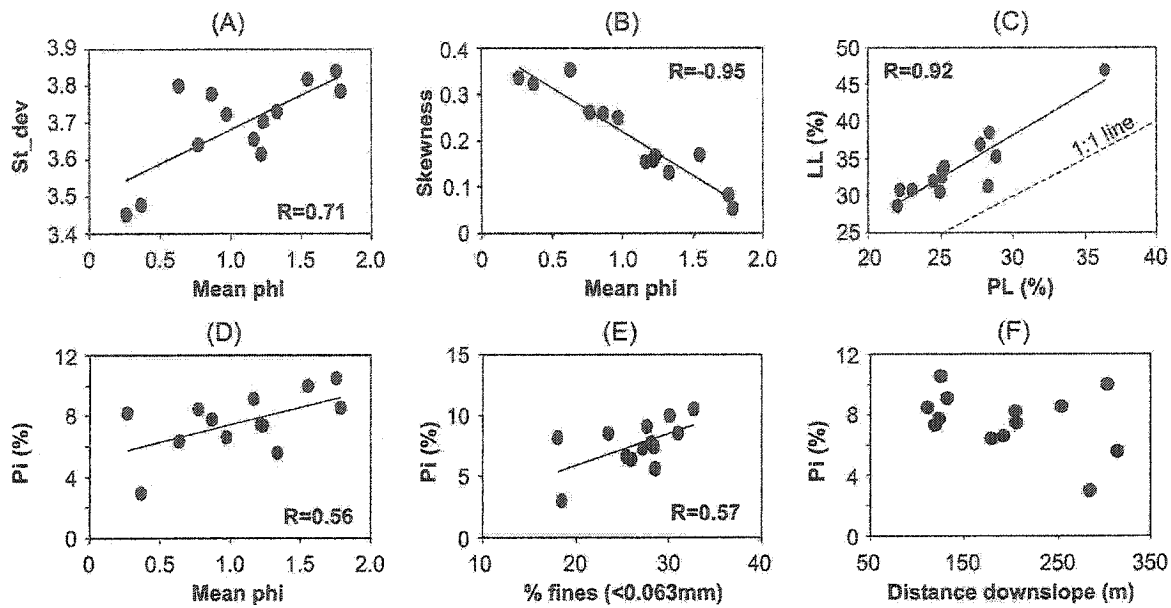


Figure 6.1: Scatterplots of environmental variables showing bivariate correlations with large r values.

In turn, finer soil is associated with an increase in the plasticity index (Pi) (Figure 6.1d, e). Since the liquid limit increases linearly with the plastic limit (Figure 6.1c), soil with a higher liquid limit has a higher plasticity index. Positive relationships between mean grain size (mean ϕ) and plasticity index have been reported by Shilts (1974) for sediments subject to periglacial processes in N.W.T, Canada. A positive relationship was also reported by Skempton (1957) between plasticity index and undrained shear strength (in Harris, 1981). The results presented here indicate that solifluction lobes with a higher proportion of fines have a higher plasticity index and hence, tend to be more resistant to shear stress (Terzaghi et al., 1996).

No relationship was observed among the mean phi (ϕ) or soil fines content and downslope distance ($r = 0.09$ and 0.08 , respectively) as found by Hugenholtz and

using a slightly bigger sample ($n=19$). This implies that although a soil catena may be present on the slope, redistribution of fines is likely not a dominant environmental control on the distribution of solifluction rates and resulting lobe morphometry. Instead, the weak positive relationship between downslope distance and plasticity index (Figure 6.1f) suggests that among the lobes sampled, bigger lobes downslope had less fines and hence, lower plasticity indices. Downslope and random variations in soil consistency may then have a secondary influence on rates of lobe advance and resulting lobe morphology. This appears to have been the case for Lobe #1, which yielded considerably higher ^{14}C ages for buried humus, suggesting that the relative stability of this lobe may be attributed to its anomalously high fine content, its heterogeneous character and high Atterberg limits (see Figures 5.26 and 5.27).

No relationships were found between the other environmental parameters. It was surprising not to observe a downslope increase in the thickness of the organic mat, since it was clear from visual observations that vegetation changes occur in a downslope direction, and that the organic mat is best developed on the lower segments of the slope. This may be explained in part by the small size of the sample used, and its lack of spatial homogeneity.

6.1.2. Relationships between environmental factors and lobe morphology

6.1.2.1. Grain-size, soil consistency and vegetation

The environmental factors listed in Table 6.1 were tested against parameters of lobe morphology as defined in section 4.2 (see Figure 4.2). Scatterplots showing relationships with large r values are presented in Figure 6.2. The results show that the plasticity index

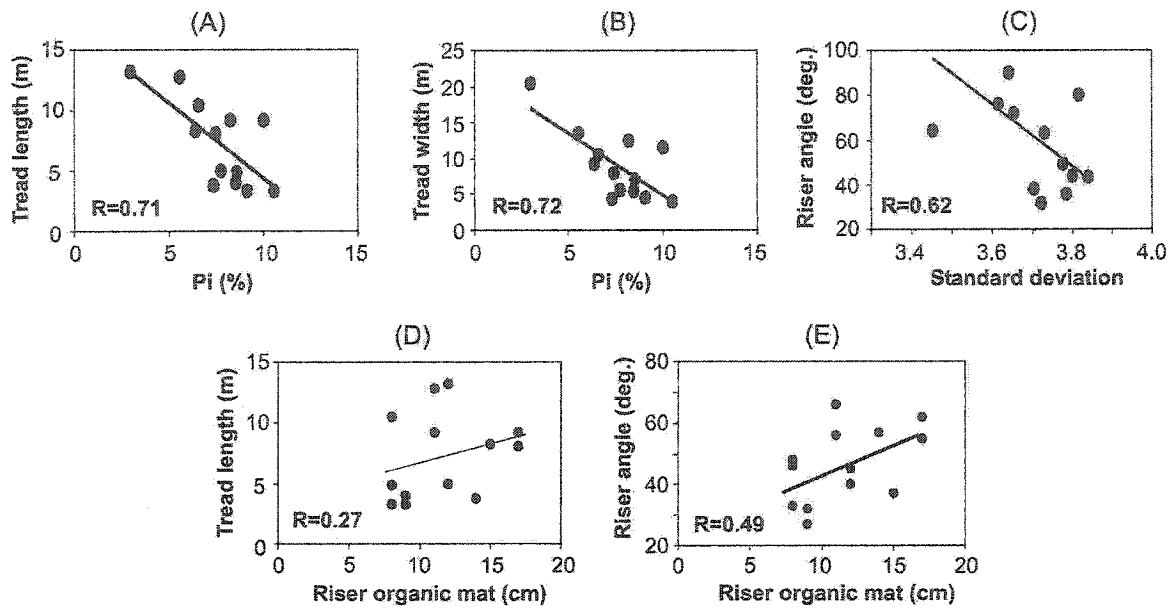


Figure 6.2: Scatterplots showing relationships with large r values between environmental variables and morphometric parameters.

of soil is negatively correlated with lobe size as defined by the length and width of the tread (Figure 6.2a, b). Hence, as suggested in the previous section, the consistency of lobe sediment may partly influence its resistance to movement. The lower plasticity of bigger lobes suggests that this factor contributes to reduce the shear strength at the front, eventually allowing the riser to advance more easily. On the other hand, the degree of sorting may also increase the stability of lobes, as higher riser angles were associated with more poorly sorted lobes (Figure 6.2c).

The correlation between tread length and the thickness of the organic mat on the riser was low but the minimum thickness of the organic mat appears to increase ($r = 0.27$) with increasing lobe size (Figure 6.2d). Organic mat thickness on the riser is also weakly associated with the riser angle (Figure 6.2e). These results nonetheless suggest that the organic mat has an influence on the form of the lobes. In the field, the organic mat was

thickest on steeper and larger lobes and is believed to be an important factor of lobe stability, as the thicker turf cover increases the rigidity of the front and allows lobes to grow for a longer time before the front collapses. One reason for the poor relationships observed between size parameters and the thickness of the organic mat may be that it is the rooting depth which influences the stability of the soil, while only the A horizon thickness was measured. Two big lobes, which were excavated in 2001 and not backfilled entirely, showed in 2002 that the soil had collapsed everywhere in the trench, leaving only a “roof” of soil supported by plant roots. The turf layer, including the rooting depth, appeared to be holding the lobe back from complete collapse.

6.1.2.2. Slope gradient

Although slope gradients were not directly measured in the field, values for each of the 235 lobes mapped on the slope were derived from the DEM (Figure 6.3a). No correlations were found between any of the lobe size parameters and slope gradient. However, inspection of the map showing lobes super-imposed on slope gradients clearly indicates that lobes are present only on a restricted range of slope gradients, represented by blue cells on Figure 6.3a. The resulting histogram showing the distribution of gradients for the entire slope indicates that gradients vary between 0 - 46°, with a mode of 13°. The distribution of gradients measured in front of lobes is normal and homogeneous, varying between 2 - 22° with a mean of 12° and a mode of 11°. Hence, if the local slope gradient does not appear to control the size of solifluction lobes, it apparently controls their presence on the slope. Slope segments with gradients in excess of 22° were devoid of lobes, as water probably drains too fast on these steeper slopes, and shear stresses are too great for significant accumulation of soliflucting material to occur.

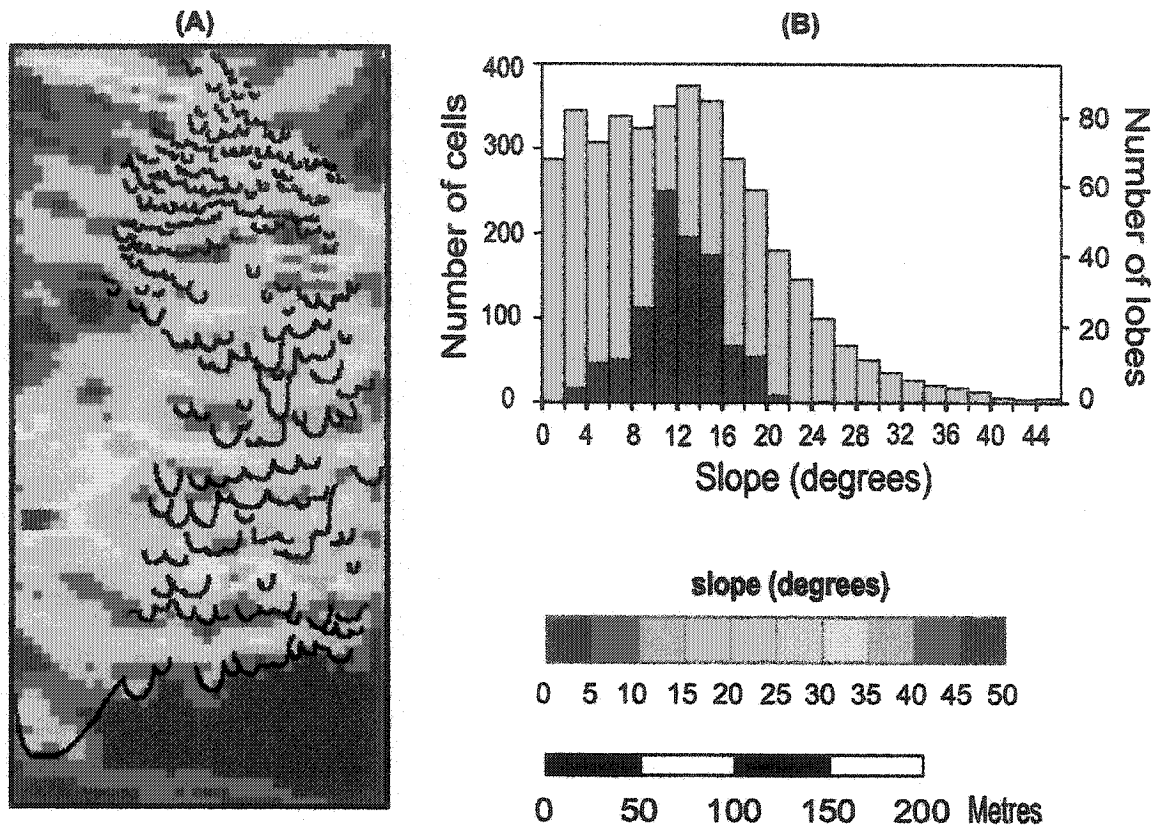


Figure 6.3: Distribution of lobes on the slope in relation to slope gradient: (a) map of lobes showing classes of slope gradient (1 cell = 5 m x 5 m); (b) histograms showing the distribution of gradients on the entire slope (in grey) and the distribution of gradients in front of lobes (in red).

Very gentle slopes have the opposite effect: they are concentrated at the bottom of the slope and are often saturated, but the low shear stress acting on these slopes does not cause differential movement and accumulation of material does not result in the formation of lobate outcrops. Solifluction is probably active, but it is of the undifferentiated type. This resembles the mechanism of turbulent flow, in which a critical velocity is necessary to initiate zones of retarded flow and resulting turbulent structures (e.g. Allen, 1997, p.167).

In summary, the results showed that the fining of lobe sediment, either due to catenary or random variations, increases the plasticity index of the soil and decreases its level of sorting. The higher plasticity index and lower sorting combine to increase the shear strength of the soil, by increasing the cohesion and interlocking forces within the soil. The frontal stability of smaller lobes upslope, which have less-well vegetated fronts, is probably more controlled by the shear strength of the soil (and hence by P_i and sorting), while this effect is probably offset for bigger lobes downslope, where stability is controlled primarily by the better-developed turf cover on the riser. Once the turf layer is disrupted, a low plasticity index favours colluvial outbursts and rapid burial under a thin sheet of sediment, as seen in the stratigraphy of bigger lobes (see Figure 5.21). Hugenholtz and Lewkowicz (2002) reported an increase in lobe size (tread length and width) with increasing organic mat thickness. Hence, a thicker turf cover may better retain the soliflucting soil behind the riser and causes the lobe to enlarge backwards and sideways. In general, relationships between grain-size parameters and size parameters are not strong. Nonetheless, these factors are integral constituents of the environmental gradient responsible for the downslope variation in lobe development.

6.1.3. Relationship between surface rates and groundwater levels

It was shown in Section 5.2.6 that larger lobes are associated with larger surface movements. In Section 5.4.2, it was demonstrated that water in the active layer progressively drains toward the bottom of the slope as the thaw plane advances and snowmelt ceases. This pattern of slow downslope drainage of active layer groundwater, previously reported by Woo and Steer (1983) and Woo and Xia (1995), leads to increased levels of saturation downslope (see Figure 5.33). As stated in the summary of Section

5.4, the distribution of groundwater is critical during autumn freeze-back as it determines the depth and concentration of segregated ice within the active layer. During spring thaw, sub-surface flow of water from upslope helps to maintain high water content and low effective stress in the thawing soil, thereby augmenting the chance for longer solifluction movements. Thus solifluction rates must depend on the level and duration of soil saturation during thawing of the ground.

To test this hypothesis, accumulated percentages of thawed-layer saturation (APTS) were calculated for the treads of the five lobes monitored for groundwater levels and surface movement, and were plotted against the cumulative surface movement for each lobe. APTS was calculated by multiplying the saturation level (water table thickness/thawed layer thickness x 100) at a particular date by the number of days before the next measurement was made, and adding the results to obtain an integration over time of the saturation level (in %·days). APTS was calculated for the period May 25th- June 24th, when data were available for all sites. Thawed-layer saturation for site I was calculated from adjacent lobe J since no groundwater well was installed on the tread of lobe I to avoid disturbing the monitoring station. A positive, and possibly non-linear relationship appeared between the APTS and the maximum surface displacement observed on July 13th (Figure 6.4). These results confirm the importance of sustained high water levels in the thawed layer in enhancing solifluction rates. As the mean surface movement was related to lobe size and downslope distance (see Figure 5.20b), the amount of moisture received by a particular lobe appears to be a determinant in accelerating solifluction rate and corresponding lobe enlargement. As lobes downslope receive more moisture for a

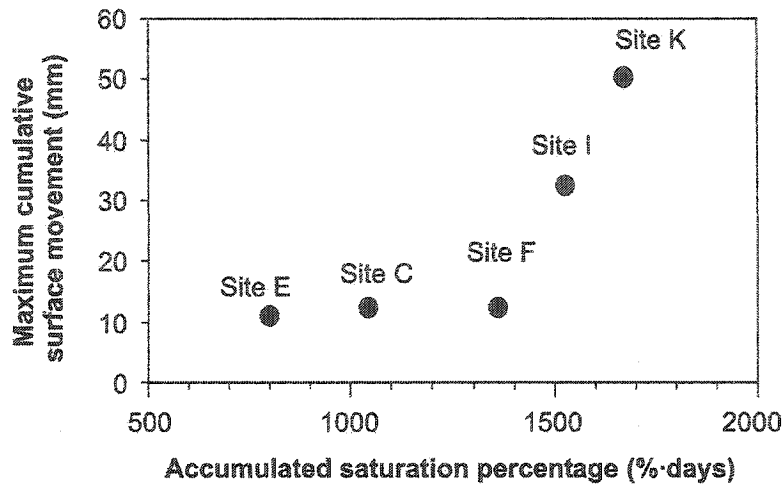


Figure 6.4: Scatterplot of accumulated percentage of thawed-layer saturation (APTS) against maximum observed surface displacement for five solifluction lobes, for the period May-25th-June 24th, 2002.

longer period of time over the thawing season, they are more active and hence, can grow larger.

6.1.4. Synthesis

Correlation analyses between environmental variables and lobe size parameters, and between groundwater levels and rates of surface movement confirm observations made in the previous section regarding the spatial variation of lobe movement and resulting dimensions. The progressive downslope drainage of water within the thawed layer appears to be the dominant reason for the initiation of a multi-component positive feedback system responsible for the existence of a solifluction lobe catena on the slope. While solifluction rates appear directly related to groundwater flow, drainage of the slope also has secondary effects. Fine material is redistributed downslope, which promotes ice segregation but also increase the shear strength of the soil, and the downslope increase in moisture leads to a more extensive vegetation cover on the lower slope. In turn, the thicker vegetation cover on the lower slope insulates the soil from solar radiation and a

shallower active layer develops, thereby causing the water-table to rise closer to the surface. The moister and finer soil downslope enhances solifluction rates, but the thicker turf cover on lobes and the higher plasticity of soil leads to the formation of lobe fronts more resistant to movement, explaining why lobes are larger downslope, but also generally steeper.

The effect of slope gradient on the rates of lobe advance is indirect: basal slope concavities generally correspond to zones of groundwater resurgence and enhanced solifluction. For an equal supply of moisture, solifluction rates and lobe size are probably directly related to shear stress and therefore slope gradient, but very steep slopes cause water to drain too rapidly for significant movement to occur. An experimental centrifuge-based study by Harris et al. (2001) showed that surface movement rates on an 18° slope were on average 3.5 times higher than on a 12° slope, but volumetric transport rates were only 2.2 times greater; and on a 24° slope, thawing resulted in failure by mudflow. They concluded that the threshold for the transition between solifluction and rapid soil failure for their test soil was around 18° . This is of the same order as the maximum slope of 22° for solifluction lobes to occur on the study slope. The limiting angle of 22° is also close to the upper limit of 25° reported by Harris (1981) for solifluction lobes to occur. The transformation of lobes into mudflows was observed by Price (1969) in the Ruby Range, where lobes located on a valley shoulder collapsed into mudflows as they approached the steeper slope of the valley side. Similarly, on the study slope, certain lobes located at the contact of the steep slope running down to the nearby

stream appear to have failed completely and slid downslope as more-or-less coherent masses (Figure 6.5).

Such forms of rapid solifluction have been described and analysed by Vtyurina et al. (1990) and are termed “splivs” in the Russian literature. Although they resemble active layer detachment slides, their initiation necessitates the flow of soil under a turf cover, then the breaking of the turf cover and finally the passive movement of the turf with flowing soil down the slope (Vtyurina et al., 1990). The difference between this process and the frontal collapse of lobes on more gentle slopes is the complete failure of the lobe body and its sliding some distance downslope, in the same manner as a detachment slide (e.g. Lewkowicz, 1990).



Figure 6.5: Example of complete lobe failure. The scar zone is indicated by the thin line while the still intact lobe is delineated by the thicker line.

6.2. A MODEL FOR SURFACE RATES OF SOLIFLUCTION

6.2.1. Relationship between thaw settlement and solifluction

Surface displacements measured on five different lobes revealed a pattern of decreasing movement rates over time (see Section 5.2.5). Vertical movement of the surface of lobes C and K also exhibit a pattern of decreasing settlement rates over time. Soil settlement was also recorded by one of the ablatometer systems installed on the surface of lobe I. The rate of settlement clearly decreases over time (Figure 6.6). A decreasing rate of soil settlement was reported by Tsytovich et al. (1965) and further confirmed in thaw-consolidation tests by Morgenstern and Smith (1973). In Section 5.2.3, a positive relationship was identified between net soil settlement and concurrent downslope movement, which suggested that a critical amount of thaw consolidation (and hence prior frost heave) is needed to induce gelifluction (see Figure 5.15). This threshold of thaw-consolidation must correspond to a critical ice content which if not attained, leads to the thawing soil resettling to its pre-freezing position, resulting in retrograde movement of

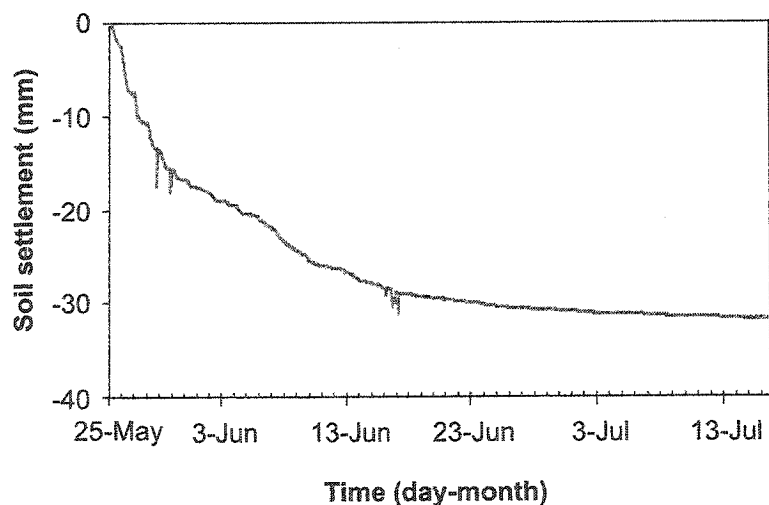


Figure 6.6: Hourly averages of surface settlement on the tread of lobe I.

the surface. Cumulative rates of soil settlement and of downslope movement thus appear closely interrelated between lobes, as demonstrated by Figure 5.15, but also in time, as suggested by the similar temporal pattern observed between vertical and horizontal movements (see Figure 5.19).

To confirm this, time-accumulated vertical displacements were plotted against corresponding cumulative downslope movements for lobes C and K (Figure 6.7). The data set includes 33 different targets, each of them surveyed 10 times during 45 days. ($N = 33 \times 10 = 330$). One outlier was removed, from lobe K (target 10), because it showed anomalously low movement due to the local restraining effect of vegetation (see Section 5.2.3). A clear positive relationship ($p < 0.01$), appears between cumulative soil settlement and corresponding downslope movement. The relationship is similar to that identified in Section 5.2.3, using only net movements as measured on July 13th. A slope of -0.52 with an intercept of -6.11 was found for the former, while values of -0.51 and -2.03 are found from Figure 6.7. This suggests that the ratio of vertical settlement to downslope movement, as recorded by surface targets, is constant over time.

The value for vertical settlement corresponding to zero horizontal movement was previously interpreted as being the threshold of thaw settlement, and hence of ice segregation or excess moisture content, necessary to disrupt the soil and induce positive horizontal movement (gelifluction). A theoretical threshold of -20 mm was found, giving a thaw-consolidation ratio of about 0.02 for a 1 m-thick thawed-layer. In Figure 6.7, the regression line suggests a settlement value of -4 mm for zero downslope movement.

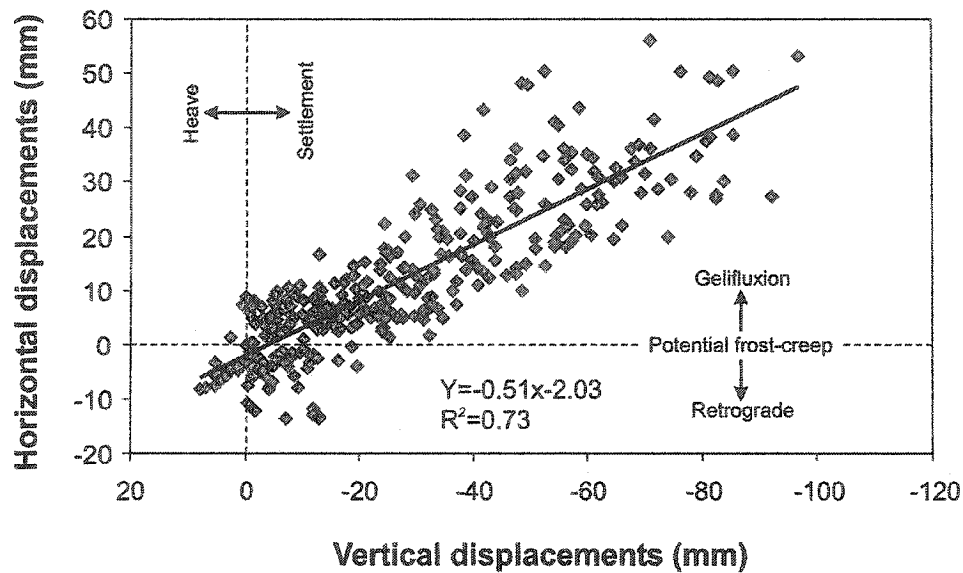


Figure 6.7: Scatterplot of vertical displacements (settlement) against horizontal movement (downslope) measured on lobes C and K.

This value is very close to 0, and because of the significant scatter around the regression line, it is difficult to ascertain a threshold value with precision. However, it can be seen that all points below the line of zero downslope movement (= PFC), that is, in the zone of retrograde movement, correspond to a range of vertical movements between -20 and 10 mm. Thus the maximum threshold of thaw settlement for only retrograde movement to occur remains around 0.02 settlement ratio. It is also interesting to note that virtually all points lying to the left of the line of zero vertical movement, that is, in the zone of heave, exhibit retrograde movement. Therefore, it is safe to say that downslope movement in the spring only occurs when the soil is consolidating, and that between 0-20 mm of consolidation, corresponding to about 0-1.8% excess water content for a 1 m-thick thawed layer, are needed to induce gelifluxion movement. The concept of threshold was explored by Chamberlain (1981), who found that a critical amount of moisture,

corresponding to the plastic limit of the soil, was necessary for thaw strain to occur. Above this limit, thaw-consolidation increased linearly with increasing moisture content.

Heaving of the soil, on the other hand, is generally a sign of retrograde movement. This phenomenon confirms theoretical findings by Morgenstern and Smith (1973), who stated that the overconsolidated elements of a soil may swell almost instantaneously by absorbing excess water from thawing segregated ice. Overconsolidation of the soil occurs on freezing when negative pore water pressures build up in fine grained soils, and add up to the frost heave pressures generated by growing ice lenses (e.g. Williams and Smith, 1989). Thus the soil, once frozen, may possess significant residual stress. Nixon and Morgenstern (1973) found a strong correlation between the thawed void ratio and the logarithm of the residual stress, and since the residual effective stress increased with soil depth, the authors concluded that the undrained shear strength and volumetric strain would also decrease with depth for a given loading condition.

An increase of undrained shear strength with thaw depth was reported by Harris et al. (1997). Harris and Davies (2000) stated that since the effective stress acting on the soil during thawing is a large proportion of the self weight stress, the void ratio and moisture content must decrease with the depth of thaw. Since the logarithm of undrained shear strength varies inversely with moisture content (see Figure 5.16, and Harris et al., 1997), it is expected that the moisture content immediately behind the thawing front will decrease with increasing depth, while the undrained shear strength will increase. The potential for gelifluction is thus likely to decrease with depth (Harris and Davies, 2000).

Measurements made in this study of soil moisture, and of soil displacement at depth and at the surface certainly confirm these statements. If the increase in overburden stress with depth results in a reduction in void ratio and moisture content, a reduction of water supply to the downward freezing front in the fall may mark the limit of gelifluction at depth. Unless there is an unlimited supply of water, water migration to the advancing freezing front will invariably cause some desiccation at depth and overconsolidation of the soil (e.g. Smith, 1985). This residual stress will generate a finite frictional resistance and the thawed soil will not behave like a viscous fluid (Nixon and Morgenstern, 1973). Groundwater supply to the freezing front, and thus position on the slope, is therefore of critical importance during autumn freeze-back, but also during spring thaw, where a high water-table reduces the effective stress on the consolidating soil and helps to maintain high pore-water pressures in the thawing soil.

6.2.2. Thaw settlement functions

The apparent relationship between solifluction movements (including here retrograde, potential frost creep and gelifluction movements) and thaw settlement may be promising for modelling purposes. Although the relationship is only empirical and the precise mechanism of translational movement has not been demonstrated in this study or in the literature, a considerable amount of experimental work on frost heave and thaw-consolidation has made models available for the prediction of such movements. The linear relationship between solifluction and thaw settlement allows such models to be used for predicting surface rates of solifluction.

Thaw-consolidation theory (Morgenstern and Nixon 1971) states that:

$$S_t / S_{max} = \text{constant} = S \quad (6.1)$$

Where: S_t denotes consolidation settlement that has occurred at time t , S_{max} denotes consolidation settlement that would occur if thawing were stopped at time t . S denotes the average degree of consolidation of the thawed soil.

This relationship, first identified by Tsytovich (1965), states that the average degree of consolidation is constant and independent of time. The total consolidation settlement that would occur if thawing were stopped at time t is found from:

$$S_{max} = m_v (P_0 X + \gamma' X^2 / 2) \quad (6.2)$$

(Morgenstern and Smith, 1973). Where P_0 denotes the stress applied to the surface, X denotes the depth of thawing, γ' is the submerged unit weight of soil and m_v is the coefficient of compressibility of the soil skeleton, and:

$$m_v := \frac{\Delta V}{V \cdot \Delta \sigma'} \quad (6.3)$$

where $\Delta V/V$ is the change in volume of soil resulting from an increase in effective stress $\Delta \sigma'$ (e.g. Terzaghi et al., 1996). For a soil consolidating under its own weight, $P_0 = 0$ and equation 6.2 reduces to:

$$S_{max} = m_v (\gamma' X^2 / 2) \quad (6.4)$$

The depth to the thaw line is evaluated from:

$$X = \alpha \sqrt{t} \quad (6.5)$$

or in its modified form:

$$X = b \sqrt{CDDT} \quad (6.6)$$

Where α and b constants and CDDT denotes the cumulative thawing degree days (e.g. Hinkel and Nicholas, 1995). Hence consolidation can be expressed as a function of time,

combining equations 6.1, 6.4 and 6.5, and adding a term for consolidation due to phase-change of ice to water (Morgenstern and Smith, 1973):

$$S_t = (S \cdot S_{\max}) = (S \cdot m_v \cdot P_0 \cdot \alpha) \sqrt{t} + (0.09n \alpha) \sqrt{t} \quad (6.7)$$

for a soil consolidating under an external load P_0 and where self-weight stress is comparatively small and can be neglected. For a soil consolidating solely under its own weight:

$$S_t = (S \cdot m_v \cdot \gamma' \cdot \alpha^2 / 2) t + (0.09n \alpha) \sqrt{t} \quad (6.8)$$

where the far right term in equations 6.7 and 6.8 represents the phase change consolidation of ice, with n being the bulk porosity of the soil. It can be seen that when a soil consolidates under an external load, settlement is directly proportional to the depth of thaw and since this advances proportionally to the square root of time, settlement proceeds accordingly. This behaviour was reproduced by Tsytovich (1965) and Morgenstern and Smith (1973) in thaw-consolidation tests, using soils under constant external loads. For a soil consolidating under its own weight, settlement by filtration consolidation is linearly related to time, as the thaw-consolidation theory predicts an increase in consolidation with depth. A slight decrease in settlement rate is predicted by the model because of phase change consolidation, which decreases over time. For a soil consolidating under its own weight, the average degree of thaw-consolidations is equal to:

$$S = 1 - \frac{1}{1 + \frac{1}{2R^2}} \quad (6.9)$$

where: $R = \frac{\alpha}{2\sqrt{c_v}}$, and is the thaw-consolidation ratio (Morgenstern and Nixon, 1971). R

is the ratio between the rate of thaw and the capacity of the soil to expel excess water under loading conditions. Therefore, the average degree of consolidation decreases when the thaw-consolidation parameters R increases.

The relationship between thaw depth, settlement and time was tested for different lobes on the slope. First, the relationship between thaw depth and time, and between thaw depth and cumulative thawing degree days was tested, using contour interpolation of soil temperature measured on the treads of lobes #2, lobe I and lobe K. The depth to the thaw plane, as defined by the 0°C isotherm, was calculated and plotted against the square root of time since the start of thaw, and against corresponding cumulative degree days (CTDD) of the air (Figure 6.8). For lobe I, because the thermistors began recording temperatures some time after ground thaw had begun, thaw-depth could not be calculated near the surface. Instead the time when thawing began was evaluated from adjacent lobe 8 where year-round temperatures were available. For lobe K, thaw depths for the first 8 days of thaw were eliminated because they were followed by refreezing at the surface, while the soil was still thawed 10 cm deeper. A heavy snowfall occurred around May 7th which covered the lobe and provided insulation from solar radiation. The soil at the surface remained cool, giving an unusually gentle temperature profile, the soil being close to 0°C over a depth of 20 cm.

The depth to the thaw plane broadly increases with the square root of time, and rates of thaw (α) are 0.20, 0.19 and 0.14 for lobes #2, I and K, respectively. Although these values are similar, especially between lobes #2 and I, the rate of thaw slightly decreases

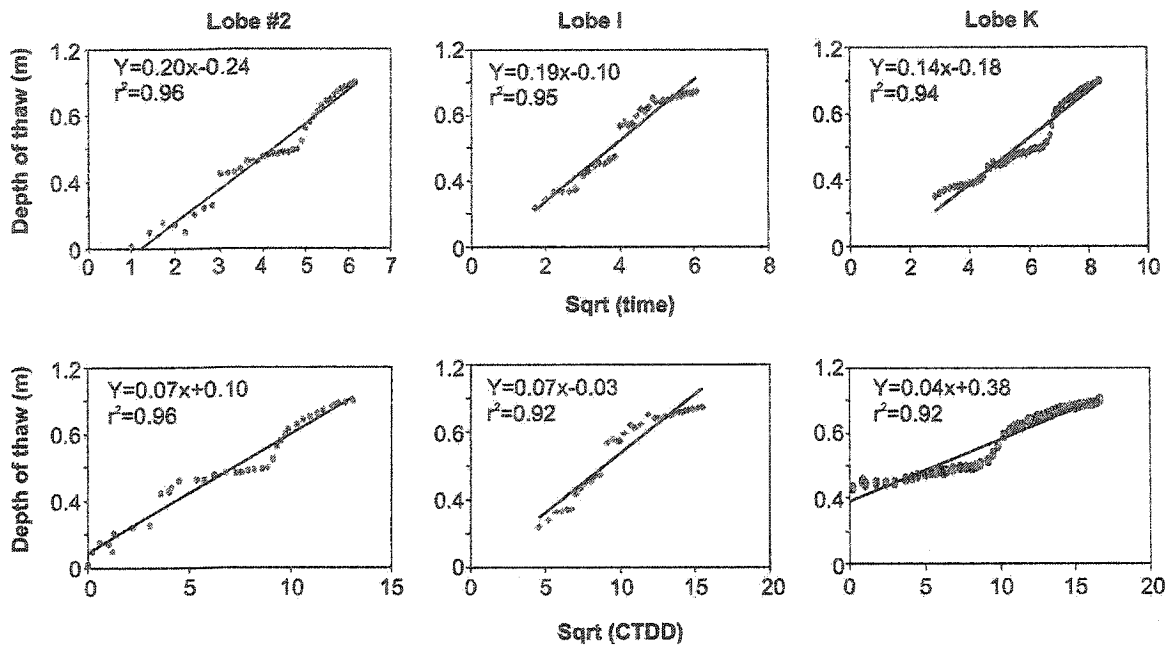


Figure 6.8: Scatterplots of thaw depth against square root of time, and cumulative thawing degree days (CTDD) for three solifluction lobes.

from the smaller lobe #2 (upslope) to the bigger lobe K (downslope). The reduced rate of thaw may be attributed to the higher water content found in the bigger lobe, which controls the thermal properties of the thawing soil and tends to lower the rate of thaw (Nixon and McRoberts, 1973). In addition, the thicker organic mat on lobes downslope reduces the thaw rate by restricting surface penetration of heat. For example, a thaw rate of $0.12 \text{ m} \cdot \text{day}^{1/2}$ was found at the CALM site at the base of the slope, where a thick moss cover is present.

Cumulative soil settlement, as recorded by the ablatometer on the surface of lobe I, broadly increased with the square root of time, as predicted by thaw-consolidation theory for soil with an external load (Figure 6.9a). Soil settlement broadly increases linearly with thaw depth (Figure 6.9b), and so departures from the linear pattern $x = \alpha\sqrt{t}$

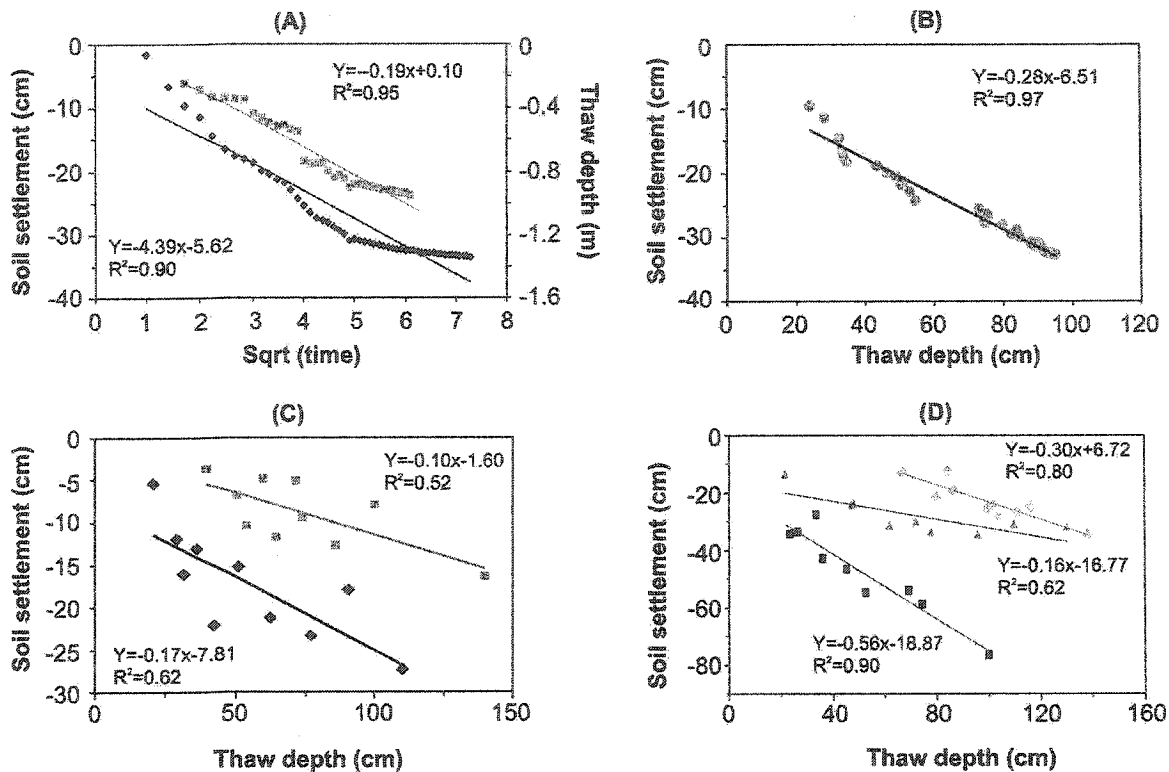


Figure 6.9: Scatterplots of: (a) Soil settlement and corresponding thaw depth against square root of time, lobe I (black dots are settlement values and grey dots are thaw depths); (b) soil settlement against corresponding thaw depth, lobe I; (c) soil settlement against thaw depth, lobe C; (d) soil settlement against thaw depth, lobe K. Black dots are points on tread, medium grey are points on front and light grey dots are points on the slope downslope of the riser.

are reflected in the settlement rates (Figure 6.9a). Thaw settlement, as measured from surface targets, also appears to be broadly linearly related to the depth of thaw (Figure 6.7c, d), but with an increasing scatter of points. This may be due to the lower resolution of the measurement technique, compared to the more sensitive ablatometer on lobe I. Although a linear model appears as a reasonable fit between thaw depth and soil settlement, Figure 6.9 suggests that the relationship may be curvilinear, with increasing slope near the soil surface.

The ratio between soil settlement and thaw depth differs between and within lobes. It is lowest on smaller lobe C, located upslope (0.10 – 0.17), larger on medium-sized lobe I, located mid-slope (0.28), and largest on the biggest lobe K, located downslope (0.16 – 0.56). The fronts of the lobes (riser and slope ahead of the riser) have lower ratios than the treads, for both lobes C and K. In regards to equations 6.7 and 6.8, the difference in the ratio of thaw depth to thaw settlement can be related to four controlling parameters: the rate of thaw α , the porosity of the soil n , the coefficient of compressibility m_v and the hydraulic conductivity k , which enters with m_v in the coefficient of thaw-consolidation R . The parameter n , relates to the amount of ice in the soil. According to equations 6.7 and 6.8 and from field observations, it is hypothesized that lobe fronts become compacted due to the slow build-up of pressure resulting from the accumulation of soliflucted material upslope. Combined with the deeper water-table, lobe fronts will have a lower coefficient of compressibility (m_v) and less segregated ice. Water depletion near the front is also likely to cause over-consolidation of the soil and so a reduction of thaw settlement and an increase in shear strength. Higher values of undrained shear strength were found from shear vane tests made at three solifluction lobes during thaw. The mean shear strength for treads was 38 ± 14 kPa ($n=15$), compared with 64 ± 25 kPa for lobe fronts ($n=11$). In summary, the more compacted lobe fronts and their deeper water-tables restrict ice segregation, increase over-consolidation and as a result, thaw settlement and solifluction movements are small.

The observed behaviour of thaw settlement in lobes C, I and K is incompatible with equation 6.4 for self-weight thaw settlement, which predicts an exponential increase of

settlement with increasing thaw depth, and an almost linear increase of settlement with time. Departure from the linear trend is controlled by phase-change consolidation, which decreases with the square root of time (Equation 6.8). The settlement behaviour is more consistent with equation 6.7 (a soil with an external load), which predicts a linear relationship between thaw depth and settlement and an increase of settlement with square root of time. Even when plotted against the square root of time, the settlement curve still shows a slight decrease in settlement rate over time (Figure 6.9a). This suggests that some assumptions made in the equations for thaw settlement may be inappropriate for the field site. A re-interpretation of the settlement equation for a soil consolidating under its own weight is therefore presented next.

6.2.3. Settlement function for self-weight consolidation

When a soil freezes downward from the surface, strong negative pore-pressures develop at the freezing front, and if the overall void ratio increases from the segregation of ice lenses, the soil skeleton becomes overconsolidated. This is especially true for deeper soil horizons where water supply to the freezing front is likely to decrease with time. Nixon and Morgenstern (1973) found that this residual stress arising from soil freezing typically increases with depth. As they state, “unless the residual stress can be shown to have a significant value compared with the overburden stress, serious deformation may still occur at the thaw plane” (Nixon and Morgenstern, 1973). In engineering practices, an overconsolidated soil with low excess ice content reduces the chance for pronounced perturbations by thaw settlement. Three assumptions were made in the first formulation of the Thaw-consolidation theory (Morgenstern and Nixon, 1971):

- (1) The residual stress σ'_0 is nil;

- (2) The soil remains saturated;
- (3) The soil is homogeneous, with a constant porosity n .

Morgenstern and Smith (1973) recognized that the first assumption was not valid, but that low residual stresses were to be expected for ice-rich soils or soils with high void ratios when thawed. They emphasized that when the stress and thermal histories are such as to reduce the void ratio of the soil, a significant residual stress may be present in the soil upon thawing. The second assumption, although possible on flat terrain, is fairly unlikely on moderate slopes. As shown in the present study, once the thaw plane reaches a certain depth and snowmelt ceases, groundwater progressively drains down the slope so that only a concave basal slope segment remains near saturation into mid-summer. Saturated conditions facilitate the formulation of the consolidation theory by allowing the use of Darcian flow for pore water, and a moving water-table would introduce additional complications in the formulation of the theory. This assumption is maintained for now for simplicity in the computations, but general conclusions may still be drawn on the effect of water-table variations. The third assumption is linked to the first. If residual stress is considered to increase with depth, the void ratio and associated porosity must decrease accordingly.

Starting from equation 6.1, the effect of residual stress may be introduced in the settlement equation for self-weight consolidation as:

$$S_{max} = m_v \int_0^x (\sigma' - \sigma'_0) dx \quad (6.10)$$

Where the total settlement S_{max} equals the change in effective stress at the thaw plane times the coefficient of compressibility m_v . The change in effective stress at X (the depth of the thaw plane) is the depth integration of the overburden effective stress σ' minus the residual stress σ'_0 (Morgenstern and Nixon, 1971). If the residual stress is a function of depth, we may begin by assuming a linear increase in σ'_0 with depth such that:

$$\sigma'_0 = ax + b \quad (6.11)$$

where a is the slope and b is the initial residual stress at the surface (Morgenstern and Smith, 1973). Equation 6.10 then becomes:

$$S_{max} = m_v \int_0^x (\gamma'X - aX - b) dx \quad (6.12)$$

where X is the depth to the thaw plane and γ' is the submerged unit weight of soil (Morgenstern and Smith, 1973). Integrating and adding the volume change from ice-water transformation gives:

$$S_{max} = m_v (\gamma'X^2/2 - aX^2/2 - bX) + 0.09nX \quad (6.13)$$

Where the first term on the right hand side denotes settlement by filtration consolidation and the second term denotes phase-change consolidation. The porosity of the soil n is equal to:

$$n = e / 1 + e \quad (6.14)$$

where e is the void ratio (e.g. Terzaghi et al., 1996). Since there is a linear relationship between the residual stress of the soil and the logarithm of the void ratio (Morgenstern and Smith, 1973), we can write:

$$e = -j \log (aX + b) + j \quad (6.15)$$

where i is the slope and j is the void ratio at the soil surface. This equation states that as the residual stress increases linearly with depth, the void ratio increases with the logarithm of the depth function of σ'_0 (linear in this case). The depth function for soil porosity then becomes:

$$n = \frac{1}{1 + \frac{1}{(-i \log (aX+b) + j)}} \quad (6.16)$$

and the settlement equation 6.13 with regards to depth becomes:

$$S_{\max} = AX^2 - BX \quad (6.17)$$

where: $A = 0.5 S m_v (\gamma' - a)$

$B = (b + 0.09n)$ and n is defined in equation 6.16

With respect to time, considering that $X = \alpha \sqrt{t}$:

$$S_t = C t - D t^{0.5} \quad (6.18)$$

where: $C = 0.5 S m_v \alpha^2 (\gamma' - a)$

$D = (b + 0.09n)$

$$\text{and: } n = \frac{1}{1 + \frac{1}{(-i \log (a\alpha\sqrt{t}+b) + j)}} \quad (6.19)$$

Equations 6.17 and 6.18 introduce the influence of residual stress on thaw settlement. The importance of the effective stress variation with depth, here taken as a simple linear function, is determined by the constants a and b . If $\gamma' \approx a$, that is when the effective stress σ' due to overburden weight increases with depth at the same rate as the residual

stress σ'_0 (recall that $\sigma' = \gamma'X$ and $\sigma'_0 = aX+b$), the increase of effective stress is determined by bX , and is linear with depth, and linear with $\sqrt{t}$. If $\gamma' \approx a$ and residual stress is nil at the surface ($b = 0$), settlement by filtration consolidation is nil and phase change consolidation dominates surface settlement. Settlement by phase-change consolidation decreases with the logarithm of depth, where the slope i differs between soil types (Morgenstern and Smith, 1973), and the intercept j is the void ratio at the soil surface.

Some constants appearing in equations 6.17 and 6.18 are difficult to evaluate, especially in the field. The constant S , which is the ratio between settlement at time t and total settlement if thawing were stopped at time t , depends on the relative contribution of settlement by filtration consolidation and by phase change consolidation. If enough pore-water has developed, soil settlement will continue once thawing has stopped, as excess water is expelled from the soil along pressure gradients. In other cases, water will drain rapidly from the soil or will be partly or fully absorbed by the overconsolidated soil matrix, and most of the observed settlement will be due to phase change consolidation, giving a high S ratio. The most difficult coefficients to evaluate are a and b , which determine the relationship between the residual stress and depth. These two parameters are critical in equations 6.19 and 6.20, as they determine the importance of filtration consolidation versus phase change consolidation. The higher the values of a and b , the closer the settlement curve approaches to a linear relationship with $\sqrt{t}$. A linear increase in residual stress may be an over-simplification, as it is likely that the residual stress increases abruptly somewhere at a depth where water supply had decreased during

freezing. The decrease in phase-change consolidation with depth, as predicted from the decrease in the bulk porosity of the soil, will also accentuate the exponential character of the settlement curve. The constants i and j , which determine the relationship between the logarithm of void ratio and residual stress, may be determined in the laboratory (e.g. Morgenstern and Smith, 1973), but this is not an easy task. The thaw rate α can be evaluated from knowledge of thaw plane depth over time (Figure 6.6). The coefficient of compressibility m_v , could be evaluated after thawing is finished in late-summer, by $S_{\max}/X^2\gamma'$, but this would be an over estimation since the residual stress is unknown.

Therefore, although equations 6.17 and 6.18 provide the physical background for one-dimensional self-weight thaw-consolidation of a soil frozen from the surface downward, some generalizations must be made in order to allow their use in field studies such as the present one. Equations 6.17 and 6.18 can be re-organized in a more applicable form if the depth function for soil porosity is simplified. Equation 6.16, which requires many unknown parameters, broadly describes an exponentially decreasing soil porosity. The best way to apply equations 6.17 and 6.18 is to evaluate the variation in soil porosity with depth. Although direct measurements of soil void ratio in this study were made only in the top soil of five solifluction lobes, soil porosity at depth can be evaluated from the volumetric water content of lobe I, if it is assumed that immediately after thawing the soil is saturated. The volumetric water content of a soil layer then represents its bulk porosity at that time. The volumetric water content measured immediately after the passage of the thawing front was plotted against the corresponding square of depth, and a good linear relationship was found (Figure 6.10).

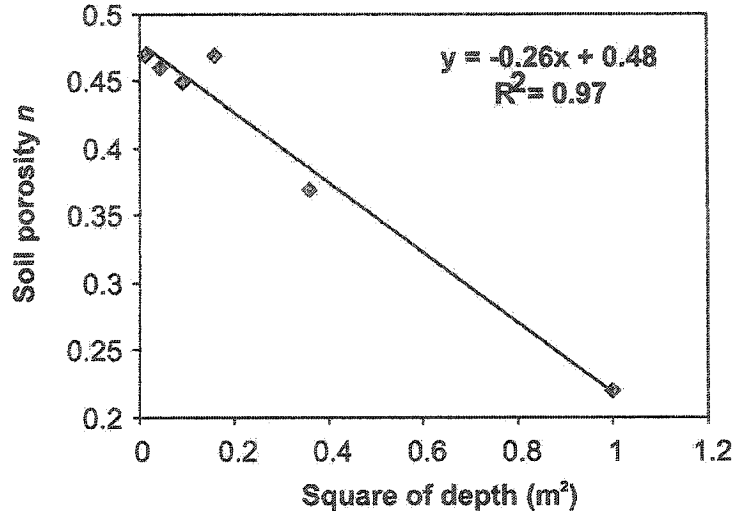


Figure 6.10: Variation of soil porosity (as estimated from volumetric water content immediately after thawing) with the square of depth at lobe I.

This suggests that soil porosity decreases more slowly near the surface than deeper in the soil. The depth function for soil porosity is thus empirically established by the exponential equation:

$$n = -0.26 X^2 + 0.48 \quad (6.20)$$

In turn, this empirical equation can be included in equations 6.17 and 6.18 to give:

$$S_x = TX^3 - AX^2 + UX \quad (6.21)$$

where: T is the coefficient of phase-change consolidation, in this case, equal to 0.0234;

A denotes the previously defined coefficient of filtration consolidation;

$U = b + 0.041$ and b is the residual stress at the surface, as previously defined.

This equation takes the form of a third order polynomial, where curvature of the settlement curve is controlled firstly by depth-variations in bulk soil porosity, and thus by phase change-consolidation, and secondly by filtration consolidation, the amplitude of which is controlled by depth variations in residual stress (constants a and b) and the coefficient m_v . Because most parameters underlying equation 6.21 are unknown, it is

impossible to physically validate the model. However the cubic form of equation 6.21 makes it possible to measure its goodness of fit on observed settlement values and by this to evaluate the constants of equation 6.21. This was done for lobe I and the following equations were obtained:

$$S_x = 0.0244 X^3 - 0.066 X^2 + 0.0794 X - 0.0048 \quad (R^2 = 0.98; n = 30) \quad (6.22)$$

$$S_x = 0.0068 X^3 - 0.029 X^2 + 0.0551 X \quad (R^2 = 0.98; n = 30) \quad (6.23)$$

Equation 6.22 is the result of a cubic fitting with an intercept equal to -0.0048 m. Since monitoring began when some thawing had already taken place, some data is missing for the upper 20 cm of the soil, and settlement values may be underestimated for a particular thaw depth. Because of this, the curve is offset down and a negative intercept appears. Equation 6.22 assumes that 4.8 mm of soil settlement were missed before monitoring began on May 25th. In order to model the settlement curve over time, it is assumed that no settlement occurs at $X = 0$ and the intercept of equation 6.22 was forced through 0. The resulting equation 6.23 can be transformed in respect to time so that:

$$S_t = 0.0068 \alpha^3 t^{1.5} - 0.029 \alpha^2 t + 0.0551 \alpha t^{0.5} \quad (6.24)$$

and with respect to the time-temperature integral CTDD:

$$S_t = 0.0068 b^3 (CTDD)^{1.5} - 0.029 b^2 (CTDD) + 0.0551 b (CTDD)^{0.5} \quad (6.25)$$

Where α and b are found from Figure 6.7 and are equal to 0.19 and 0.07, respectively.

Settlement curves calculated from equations 6.24 and 6.25 are presented in Figure 6.11. The fit between the model and the observed settlement curve over time is generally good. However, the model slightly overestimates settlement over time (Figure 6.11a), while the fit improves when the effect of surface temperature is included (Figure 6.11b).

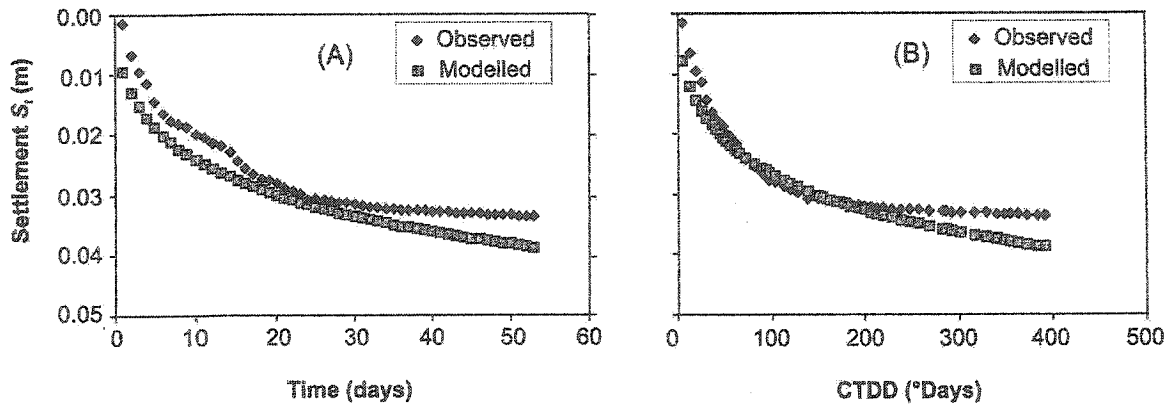


Figure 6.11: Scatterplots of observed versus modelled soil settlement at lobe I: (a) against time; (b) against cumulative thawing degree days (CTDD).

In both cases however, observed settlement diverges from the theoretical curve in the last part of the study period, with actual settlement values being less than predicted. This suggests that some assumptions in the model may be inappropriate. Because the variation in soil porosity was determined empirically, it is the assumption of residual stress which is more suspect. Little is known of the theoretical distribution of residual stress with depth, and a linear increase was used here as the simplest case. The divergence between observed and modelled values deeper in the soil suggests that the coefficient a may not be constant, meaning that the increase in residual stress with depth is non-linear. Different depth functions could be substituted for the linear relation to obtain a settlement function different to that of equation 6.23. In general, however, the cubic model resulting from the integration of the thaw-settlement function of Morgenstern and Nixon (1971), a linear increase in residual stress and a second power decrease in soil porosity, has yielded good results.

Its now possible to relate surface displacements to the model of thaw settlement. Considering from Figure 6.7 that:

$$R_t = 0.51S_t - 0.002 \quad (6.26)$$

where R_t is the cumulative surface displacement in meters and S_t the cumulative surface settlement in meters, then:

$$R_x = 0.51 TX^3 - 0.51 AX^2 + 0.51 UX - 0.002 \quad (6.27)$$

$$R_t = 0.51 T\alpha^3 t^{1.5} - 0.51 A\alpha^2 t + 0.51 \alpha t^{0.5} - 0.002 \quad (6.28)$$

Where R_x and R_t denote cumulative displacement of the surface relative to depth of thaw and relative to time, respectively. The effect of changing surface temperature can then be included in equation 6.28 by substituting $b\sqrt{CTDD}$ for X .

Because surface movements were not recorded by the ablatometers, it is not possible to test this model on lobe I. The model was tested on two targets closest to the location of active layer probing, on the tread of lobe K (Figure 6.12). The fit between observed and calculated displacement values is good overall, but the model slightly overestimates displacement at shallower depths while it underestimates displacements at greater depths. This arises because settlement was underestimated at depth and over-estimated near the surface.

6.2.4. Synthesis

The model presented here for time-displacement of the lobe surface results from the integration of a linear increase in residual stress and an exponential decrease in void ratio with depth into the settlement equations developed by Nixon and Morgenstern (1971) and Morgenstern and Smith (1973) for one-dimensional, self-weight thaw-consolidation of

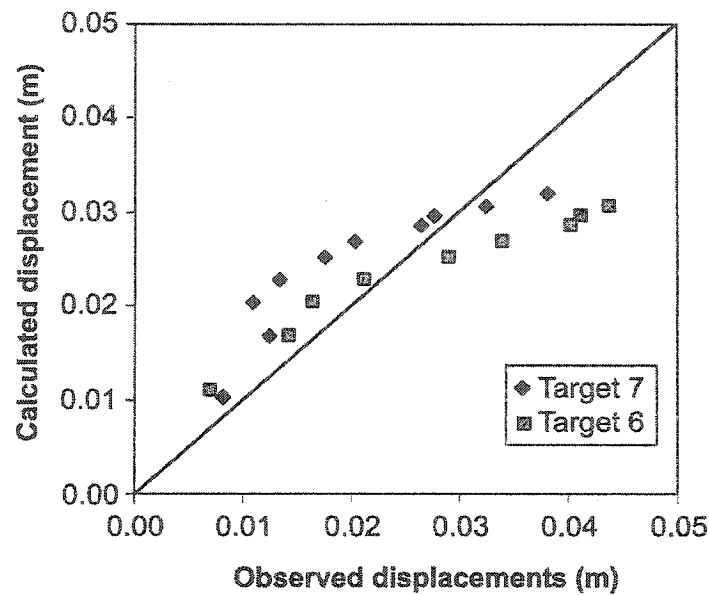


Figure 6.12: Observed versus predicted surface displacements on the tread of lobe K.

soils. Although many parameters of the resulting equation are difficult to evaluate outside the laboratory, knowledge of the thaw rate and of soil settlement may be sufficient to evaluate the resulting surface displacements. The model is the first attempt to relate solifluction rates to better-known processes such as frost heave and resulting thaw-consolidation. One major question is whether the use of a cubic equation is appropriate to model self-weight thaw settlement and resulting downslope movement, since this kind of equation will invariably lead to a reversal of the trend at some time (or depth). Prolongation of the modelled series of thaw settlement until August 30th, however, did not produce inflections in the curve. It is thus safe to assume that within the time range of ground thaw, the cubic curve maintains its exponential character. Alternatively, it was found that the predicted settlement curve can be almost perfectly approximated by a logarithmic function ($R^2 = 0.99$). The use of a logarithmic function may be a more convenient tool for predicting cumulative settlement over time/depth. The

reasonable agreement between the modelled and observed values of settlement and downslope movement leads to three important conclusions:

- (1) Translational movements occur in direct response to soil consolidation at the thaw line and surface movements represent the incremental response of movement deeper in the soil;
- (2) Increases in residual stress and a decrease in void ratio with depth leads to restricted consolidation by both filtration and phase-change processes, and to an increase in shear strength.
- (3) Phase-change consolidation appears as the dominant process controlling thaw settlement, as frost heave ratios measured in the field are often lower than the 9% expansion of the pore fraction of the soil.

Harris et al. (1997) obtained good reconstructions of velocity profiles by integrating surface displacements over thaw depth/time. However, despite the generally good agreement between calculated and observed profiles, displacements were slightly underestimated near the surface and slightly overestimated at depth. This was the result, according to the authors, of an upward redistribution of moisture, which sustained viscous flow of soil in the near-surface zone. It was concluded earlier that soil movement by viscous flow may not be the dominant mechanism for gelifluction. Because probe deformation occurred so suddenly following the passage of the thawing front, and also because of the direct relationship between thaw settlement and downslope movements, it is felt that translational movement of soil occurs mainly during phase-change

consolidation. Slow, viscous deformation of the soil may arise after thawing if sufficient excess moisture is available, but appears to be of secondary importance.

The model of self-weight thaw settlement proposed here suggests that phase-change consolidation is the main factor controlling the behaviour of settlement over time. The bulk porosity of the soil decreases with depth because of increases in overburden and residual stress with depth. This gives a lower water content, and an increase in shear strength and in overconsolidation of the soil at depth. As a result, settlement by phase-change consolidation decreases with depth/time, as does settlement from filtration consolidation. Indeed, as thawing proceeds to greater depths, it is likely that excess water liberated by thawing ice lenses is quickly absorbed by the overconsolidated soil matrix (e.g., Morgenstern and Smith, 1973; Terzaghi et al., 1996). This seems to have occurred in lobe I, where thawing of the soil around 100 cm depth released almost no water. It can be concluded that a combination of low ice content and high overconsolidation of the soil did not produce excess water. Hence settlement recorded at the surface was minimal and positive pore-water pressures did not develop. The mechanism of soil movement probably results from micro-shearing of soil at ice lens boundaries (e.g., Carlson and Kirkby, 1972; Van Vliet Lanoë, 1985; Williams and Smith, 1989, Harris et al., 1995).

As discussed in Section 5.2.3, frost heave ratios reported from field studies are often quite low, but solifluction movements are still recorded. On slopes, water supply to a downward freezing front will cease at some point, and ice segregation in the upper soil will cause this layer to heave at the expense of the deeper soil. The depth to water

depletion probably controls the depth of gelifluction, and this will vary with the position on the slope. When freezing begins, each position on the slope represent a different state between a closed system (slope crest) and an open system (concave slope bottom). Thus the depth of movement is more likely to correspond to the active layer depth downslope than upslope. This may explain the poor relationship obtained by Matsuoka (2001) between active layer depth and depth of movement for a variety of field locations.

6.3. A MODEL OF SOLIFLUCTION LOBE DEVELOPMENT

Results from process measurements and morphometric analysis suggest that solifluction lobes constitute semi-rigid landforms which develop in zones of retarded flow: these zones give rise to lobate outcrops (the risers) behind which material accumulates by solifluction. The recurrent pattern of surface velocity (fast-moving tread and slow-moving riser) identified in this study and in previous work (e.g. Benedict, 1970; Smith, 1987a), points to the long-term unstable nature of lobes. However, their constant planar morphology suggests that this instability does not lead to permanent and infinite deformation of lobes, as would happen in a viscous body. Instead, it can be hypothesized that solifluction lobes are pulsating landforms which possess an internal cycle of development.

Modelling of viscous currents on a slope (Huppert, 1982) has shown that such currents eventually break up into a series of waves of ever-increasing amplitude and with a wavelength depending on surface tension. Such models present many similarities with the spatial pattern of solifluction lobes. In particular, the wavelength of the more-or-less continuous lobate segments increases downslope, and this may be in response to the

increasing resistance of the surface to flow. In the field, this resistance is a function of the thickness of the rooting depth, while the plasticity index and level of sorting control the cohesion and interlocking force of the soil and hence its resistance to shear. The main difference is that deformation of lobes does not result in extending fingers of accelerated flow, as predicted by the lubrication theory once surface tension has been overcome (e.g. Batchelor, 1967). This may happen however, under particular and localized conditions, and complete collapse of a few lobes on the slope has produced elongated and very subdued forms. In cases of complete viscous collapse of lobes, deformation is probably permanent and the lobe is destroyed, as in Price's (1974) concept of a lobe cycle.

Whether or not the process of solifluction itself includes a viscous component, solifluction lobes can probably be considered as plastic landforms, which possess a finite yield stress. The analogy with an ideal plastic is supported by the long-term behaviour of lobes under stress, which slowly creep forward until a critical shear strength is attained and rapid collapse occurred. Thus a perfect plastic model analogous to the one used for glaciers (e.g. Paterson, 1981) may be suitable to describe long-term deformation of solifluction lobes. In the perfect plastic model for glacier flow, the basal shear stress is equal to:

$$\tau_b = \rho g h \sin \alpha \quad (6.29)$$

Where τ_b denotes the basal shear stress, ρ is the ice (soil) density, g is the gravitational acceleration (9.8 m/s^2), h is the thickness of ice (soil) and α is the surface slope. It has been shown that provided the slope is gentle, the shear stress is determined by the surface

slope of the ice. For a glacier in equilibrium (mass-balance = 0), basal shear stress is constant and equal to the yield stress for ice, and the equilibrium profile is a parabola.

Solifluction lobe profiles resemble glacier profiles, but the equilibrium model cannot be used to describe the evolution of solifluction lobes, which result from a progressive increase in shear stress near the front until the yield stress for rapid deformation is attained. If equation 6.29 does not help to reconstitute lobe profiles, it may be useful to calculate stress conditions near the front. This was done for the study slope using riser height for h , riser angle for α , and setting ρ to 1.40 g/cm^3 , i.e. the maximum soil density measured on a lobe tread. This latter value is probably an under-estimate of soil density near the front, which is highly compacted. Use of equation 6.29 serves to broadly estimate the maximum stress condition for every lobe at the time of study (Figure 6.13). The plot of frontal shear stress against distance downslope shows significant scatter, expressing the various states of lobe development on the slope. The maximum shear stress, however broadly increases downslope, and can be taken as an approximation of the minimum yield strength for rapid lobe movement. The mean calculated shear strength ($\pm 1\sigma$) is equal to $30 \pm 20 \text{ kPa}$, a value lower than the value of $64 \pm 25 \text{ kPa}$ measured with a shear vane at the front of three lobes. However, the distribution of calculated shear stresses is highly skewed, so some lobes will be greater than the measured average shear stress. The higher measured value probably arises from the fact that steeper lobes were chosen in the field and because the density value employed (1.40 g/cm^3) most likely underestimates soil density in the more compacted lobe front.

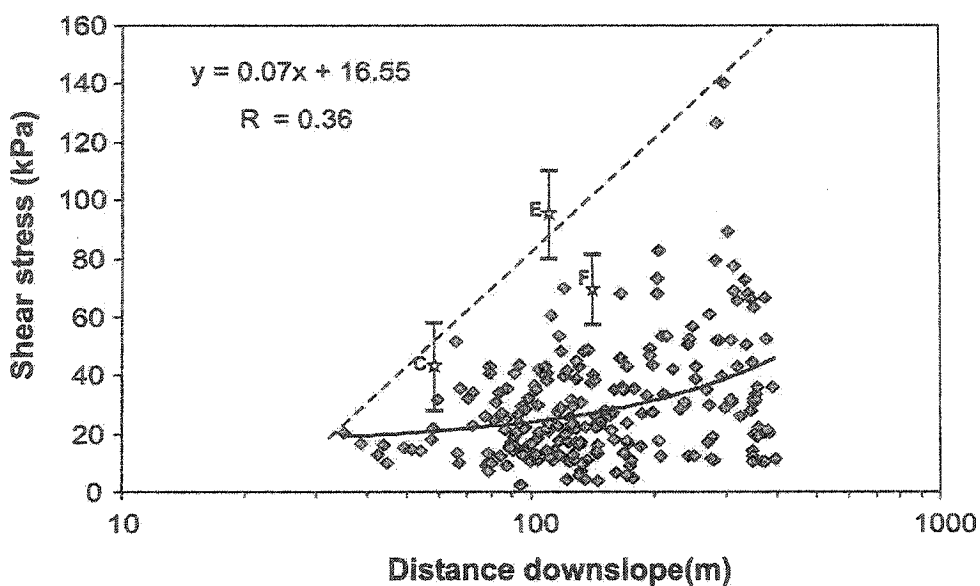


Figure 6.13: Scatterplot of calculated frontal shear stress against the logarithm of distance downslope. Black line is a linear regression and dotted line is a hand-traced line delineating the maximum shear stress. Stars are mean shear stress values, with one standard deviation bars, measured with a hand shear vane in lobes C, E and F.

The progressive build-up of material by solifluction leads to the thickening and steepening of the front, until the critical yield strength is attained in the zone of maximum thickness/steepness. Before this occurs, the riser grows by slow plastic deformation. This is possible because the thawing soil is not incompressible, and so progressive compaction of the front occurs within the elastic zone of plastic flow (e.g. Allen, 1997). Lobes with larger plasticity indices and/or with a thick frontal turf cover are likely to creep for a longer time before the yield stress is attained, and thus must have a longer cycle of development, for equal rates of surface movement.

The time for the cyclic development of lobes was broadly estimated from ^{14}C dates as being 650 ± 350 ^{14}C years. It is possible to test this estimate by extrapolating surface rates over time in order to model the evolution of a lobe profile in two dimensions. This

technique has severe limitations, since it does not take into account variations in soil density over time, and flow toward the sides of the lobe. However, Benedict (1970) has shown that near the centreline of lobes, movement is predominantly downslope. Lobe K, which is presently in the process of breaking-up, was chosen to test the empirical model. The foremost part of the riser has already collapsed and flattened out, but most of the remaining part is bulging and showing signs of active collapse (Figure 6.14). Time-displacements of the surface are calculated using the principle of conservation of mass: any longitudinal variations in transport rates must result in an increase or decrease of the lobe surface over time. No attempt was made to incorporate the effect of changing soil density, which surely occurs in zones of increased stress. Transport rates ($\text{m}^3/\text{m}\cdot\text{yr}$) were evaluated as:

$$Q = V_s \cdot d/2 \quad (6.30)$$

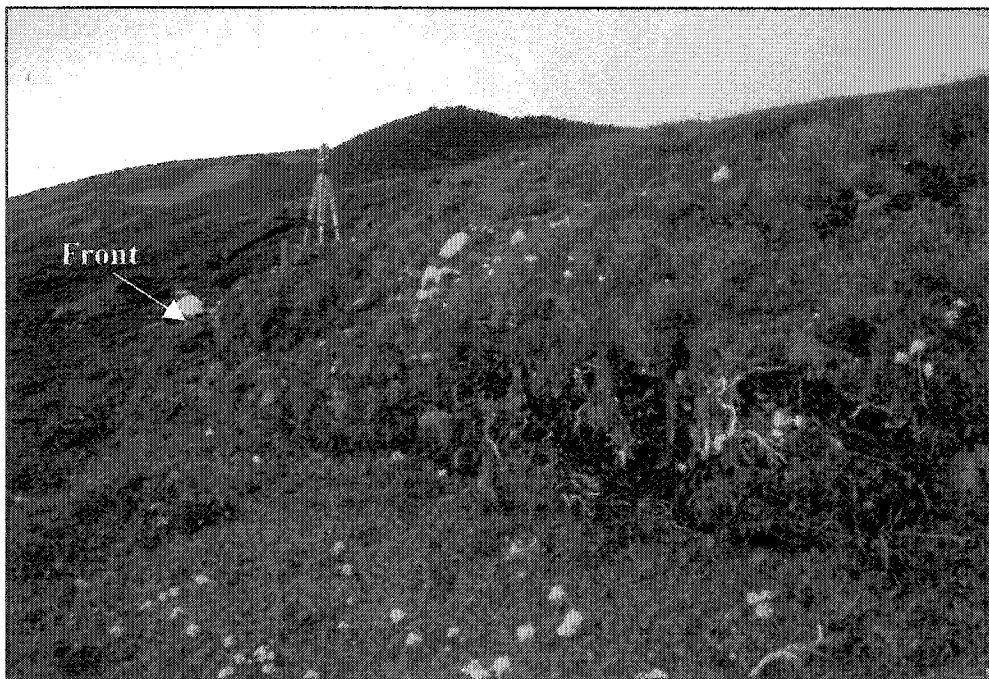


Figure 6.14: View of the collapsing sides of lobe K.

Where V_s is the annual rate of surface movement, and d is the depth of movement, which was set at a conservative value of 1.0 m. This formulation assumes a linear decrease in movement with depth and may tend to overestimate transport rates, but is adequate for the present exercise. Transport rates were smoothed using a third order polynomial to avoid the local amplification of small-scale variations in rates (Figure 6.15). The model follows the equation for continuity of mass:

$$\partial Q / \partial x = -\partial z / \partial t \quad (6.31)$$

(e.g. Wilson and Kirkby, 1978). Q represents the transport rate, x represents the horizontal distance downslope, z represents the vertical distance (normal to x) and t represents time. The rate function Q has to be defined. For now, it was evaluated empirically from the third order polynomial relationship between transport rates and distance downslope from the back of the lobe (Figure 6.15). The cubic equation was derived in x and integrated in t to give:

$$z_t = z_0 - t (0.00005x^2 - 0.002x + 0.013) \quad (6.31)$$

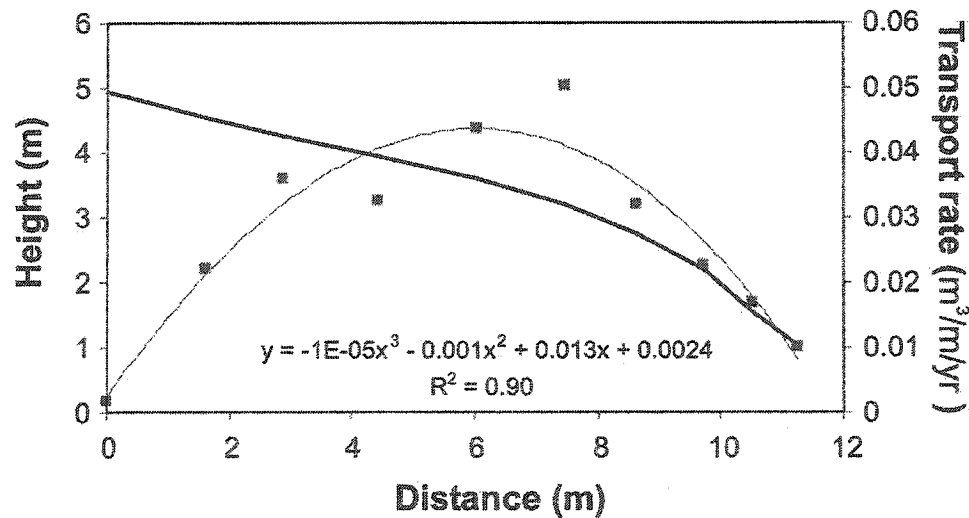


Figure 6.15: Interpolation of surface rates on lobe K using a third-order polynomial.

where z_0 is the elevation of the initial profile. Results from the model are presented in Figure 6.16. The model shows calculated profiles for 150 years before present (black line) and from present to 350 years in the future, at 50 year intervals. The model indicates that 150 years ago, the tread angle was steeper and the front was lower, with an apparent external rim in front of the riser. Forward in time from the present, material slowly moves from the tread toward the riser, with an equilibrium point where change in transport rates is nil ($dQ/dx = 0$) at about 6 m). At +250 years, a slight bulge develops on the apex of the riser. Assuming that the profile is not unstable at this time, and if this slight bulge is smoothed and the profile extended to around 350 years, the riser angle reaches the maximum angle of 90° measured on the slope.

Calculated profiles resemble those measured in the field (see profiles of excavated lobes, Figure 5.21), which vary between lobes of low convexity to steeper profiles, which

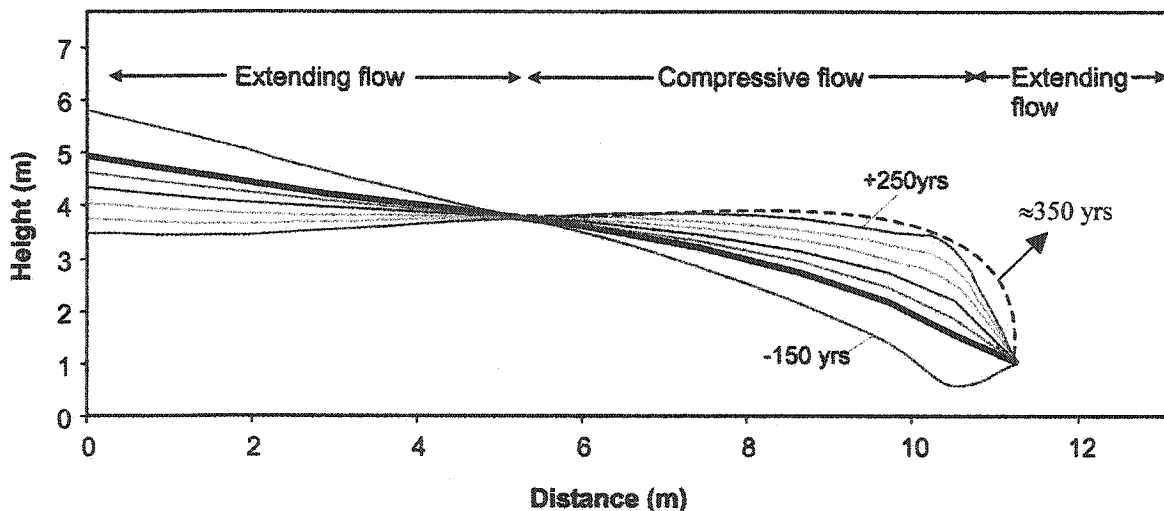


Figure 6.16: Model of lobe development based on the continuity of mass, showing extrapolation of present lobe profile back to -150 years and to +350 years, at 50 year intervals.

sometimes display a small hollow on the tread. This small hollow reflects the progressive transfer of soliflucted material from the tread to the riser. This feature was also reported by Yamada et al. (2002), who presented a qualitative model of lobe movement similar to that developed here. They related lobes with a small depression at the rear to the state of frontal advance. It is concluded here that this type of morphology is typical of lobes about to collapse or having collapsed recently. The variation in transport rates on the lobe creates zones of differential movement: behind the “equilibrium line” of zero mass transfer, the flow is extending and ahead of it, flow is compressive. Extending flow also occurs immediately downslope of the riser as movement again increases, and transverse tension cracks are often observed in this zone. The plunging and emergent shear planes observed in excavated lobes confirm the existence of extending and compressive regimes of flow, and provide evidence for lobe movement by macro-shearing.

The time for cyclic development calculated from the model is around 500 years and corresponds to the time range previously identified from ^{14}C dating of lobes. This may be regarded as a minimum estimate, as transport rates were most likely overestimated and changes in soil density were not taken into account. On the other hand, using transport rates dependent on the changing lobe morphology over time could reduce this time estimate. Thus the mean estimate of 650 years calculated from ^{14}C dates is probably a reasonable assumption for the time required to complete a full cycle of development. Smaller lobes, which have slower surface rates and probably even slower transport rates,

will take more time to build-up, but may deform under lower shear stresses than bigger lobes downslope, which have a stronger retaining turf cover.

The models proposed here presents two problems. The first one is that surface rates measured on lobe K do not represent a full year of movement. If annual movements are larger than measured, the model would yield a reduced time for the cyclic development of lobe K. Conversely, if the annual surface rates are less than measured, a larger time period would elapse for the modelled cycle completion. Since surface movements on lobe K were found to decrease over time and to be stable between June 24th and July 13th, it is thought that the net displacements recorded on July 13th probably represent most annual movement. The second problem is that surface rates were assumed to be independent of surface morphology. In fact, as the lobe profile steepens over time, it is likely that conditions responsible for the differentiation of movement between the tread and the riser amplify, causing the tread to accelerate behind the slowing riser. If this feedback mechanism exists, then the time for cyclic development of lobe K may be less than calculated. This points to the necessity of finding a function relating Q to lobe profile over time, instead of using from an empirical function based on present-day conditions.

The model confirms that the frontal morphology is a good indicator of the stage of development of a particular lobe. While the plan form (length-to-width ratio) remains broadly constant in time and space, the tread angle and the riser angle change in a consistent manner as the lobe develops. As a lobe grows toward instability, the tread

angle decreases relative to the parent slope while the riser angle increases. After its collapse, a lobe will have a steeper tread angle relative to the surrounding slope and a reduced riser angle. It is thus possible to use these two parameters to describe the development stage of every lobe on the slope. A relative tread angle was computed by subtracting the actual tread angle from the angle of the surrounding parent slope. This relative tread angle was plotted against the riser angle in order to characterize the level of development of each lobe (Figure 6.17). A logarithmic scale was used for the non-normal riser angle distribution. The bivariate distribution can be divided in four quadrats, representing four stages of lobe development. Although the mean of adjusted tread angle was 1.7° , the dividing limit was set to 0° degrees, as this represents a change in the stage of development. The dividing limit for the riser angle was set to the median, 32° , since the data is non-normal.

Stage #1 corresponds to the critical stage, where the tread angle is less than the parent slope behind, and the riser angle is steep. The second stage is the break-up stage, where the lobe front is abruptly disrupted and the riser angle declines in response to frontal slumping. In the third stage, the rigid front has collapsed and this allows material from the rear of the lobe to advance forward, and the long-profile becomes smooth. From there, the lobe slowly rebuilds as material starts to accumulate near the front again, which becomes increasingly steep (stage #4), until the lobe profile becomes convex again and the lobes grows towards instability. The time taken to accomplish this internal cycle is around 600 years, as evaluated from the previous model and ^{14}C dates. Classification of lobe morphology according to Figure 6.17 is an objective way to describe the state of

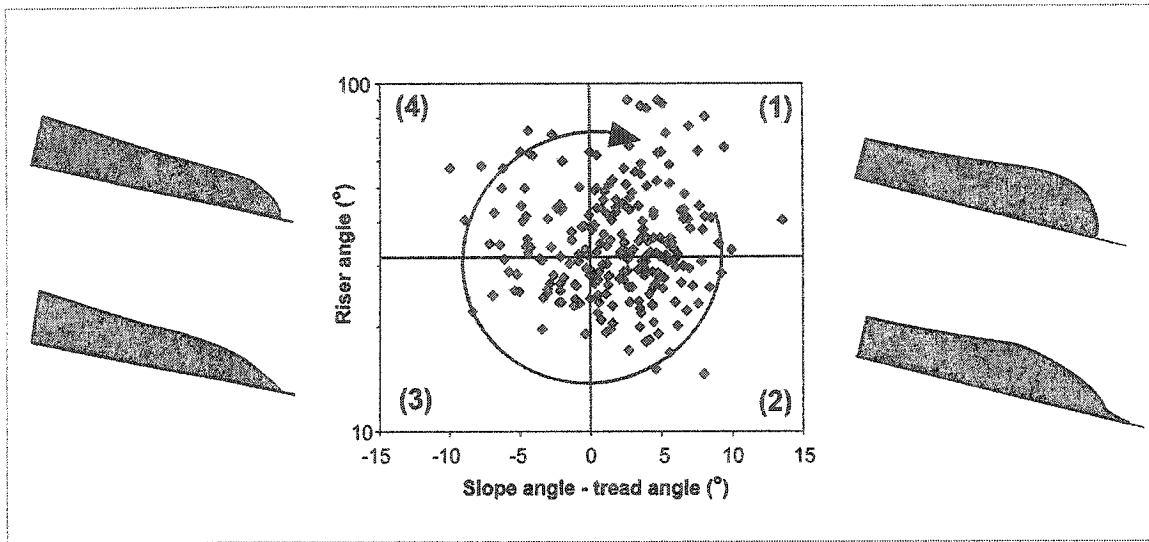


Figure 6.17: Scatterplot of relative tread angle against riser angle, showing the distribution of lobes in different development stages.

lobes on the slope. This classification may be useful in the future for inter-site comparisons.

The next step to improve this model would be to properly define the function for transport rates within a lobe. The spatial organisation of lobes suggests that surface rates at the slope scale follow a sinusoidal function of increasing wavelength in a downslope direction (see Figure 5.6). Lobe locations appear to result from regular, three-dimensional perturbations in the velocity field of the active layer. As in many fluid substances, once a critical velocity threshold is reached, strong transfer of momentum occurs in the flow field. The organisation of lobes on slopes is very similar to the segregation of flow in the near-wall region of boundary layers of river beds. Turbulent conditions give rise to zones of retarded flow across the bed, known as successions of low and high-speed streaks, which periodically erupt upward as rapid bursting structures

(e.g. Allen, 1997). Solifluction lobes appear to follow this same pulsating behaviour, where momentum accumulates in a zone of retarded flow before erupting faster than the surrounding flow.

Further modelling efforts should concentrate on the proper definition of the spatial function for transport rates, and its inclusion within the equation of mass continuity. Wave theories appear particularly promising to model landforms resulting from sinusoidal variations in transport rates. Simple theories for small-amplitude waves are probably not applicable to solifluction lobes, however, since lobes involve a strong mass transport in the direction of wave propagation, and higher-order theories should be sought (Allen, 1997).

In summary, the use of a simple model of two-dimensional development of solifluction lobes provides a clearer picture of the morphological evolution of lobes over time as well as a time scale for the completion of a full cycle of development. Unlike the traditional representation of lobes as being slow, continuously-moving landforms, it is proposed that solifluction lobes represent plastic landforms, which deform rapidly by shearing once the critical yield stress is overcome.

6.4. CLIMATIC INFLUENCES ON SOLIFLUCTION LOBE DEVELOPMENT

The conclusions reached in this study on the existence of an internal cycle of lobe development have important implications for climatic interpretations of lobe movement. Several previous studies have used more-or-less continuous series of ^{14}C dates, collected along organic horizons buried under solifluction lobes, to infer changes in regional

climate (e.g. Benedict, 1966; Gamper, 1983; Elliot and Worsley, 1999). Variations in long-term rates were correlated, often tentatively, with climatic events. In such models, solifluction lobes were interpreted as slow, continuously-moving landforms whose rates depend directly on solifluction processes. This study has emphasized that while solifluction processes constitute a mechanism of slow, continuous movement which may be directly correlated to climatic parameters (e.g. Akerman, 1996), lobes as a whole are plastic in nature and so possess a response time to climatic control. This lag time, as demonstrated earlier, may be at least 600 years. Direct correlation of long-term rates of lobe advance as inferred from radiocarbon dating of buried organics and paleoclimate is thus problematic for two main reasons:

- (1) The inherent episodic nature of frontal advance, which may lead to erroneous interpretations of accelerated (slow) movement.
- (2) The lag time between climate-change events and the response of solifluction lobes, which reduces the resolution of time-series of lobe movement.

Therefore, it is suggested that the models of Morin and Payette (1988), Gamper and Sutter (1982) and Gamper (1993) are the most suitable to infer long-term trends in solifluction lobe activity in response to climatic change. These models used large numbers of dates to identify periods of solifluction activity. The distribution curve used in this study and another developed by Morin and Payette (1988) are particularly useful in attributing statistical weight to a series of radiocarbon ages and to infer peak periods of lobe activity. These peak periods are interpreted as burial events, most likely by abrupt and rapid downslope advance of lobe fronts. Therefore, the time difference between these peaks of lobe activity (lobe advance) can be taken as changes in the time response

of lobes to controlling factors such as climate. As for dating of individual lobes, interpretation of lobe movement is hindered by the problem of mean residence time in soils, and the fact that changes in ages are step-like in nature. Closely-spaced sampling is thus required to be able to identify sharp changes in lobe age, which correspond to episodes of frontal advance. Sampling at regular interval behind the front of lobes may be the best way to critically discriminate changes in response time of a specific lobe. Correlations of a time series of lobe movement with other paleoclimate proxy indicators, however, must consider the lag time of lobe development in order to precisely identify the cause of accelerated cyclic lobe movement.

The response of solifluction lobes to climate may vary in different environments (Matsuoka, 2001). On a long time scale (hundreds to thousands of years), the development of permafrost or deep seasonal freezing on slopes is likely to be the main triggering cause for solifluction. Although shallow frost creep, and seasonal soil creep may operate on temperate slopes, the onset of deep or permanent soil freezing will result in an impermeable soil boundary which impedes drainage of groundwater. As suggested in this study and by many previous authors (e.g. Benedict, 1970; Smith, 1988; Matsumoto et al., 2001; Matsuoka, 2001), soil moisture is the dominant control mechanism of solifluction. Therefore, although regional climate controls the thermal regime of the soil and its freezing regime, moisture supply always remains a necessary conditions for significant solifluction to occur. Differential movement occurs under accelerated conditions of slope movement, and solifluction lobes result from periodic

flow perturbations in the active layer. The intensity of solifluction on a particular slope is controlled by the combination of slope gradient and moisture regime.

It is probable that a period of general climatic cooling during hypsithermal and early Neoglacial times induced widespread or continuous permafrost conditions at the Kluane study site, while the nearby glaciers were advancing, around 3000 ^{14}C years BP (Denton and Stuiver, 1966). There is much evidence in the literature for enhanced solifluction activity during that period (see the compilation by Morin and Payette, 1988). The onset of permafrost conditions would have resulted, first, in enhanced overland flow and near-surface throughflow on slopes. Moisture redistribution would have created an alpine catena in which soil material, vegetation and active layer depth become increasingly affected by distance downslope. The periodic segregation of flow within the active layer would have resulted in the development of a solifluction lobe catena, in which lobes increase in size and spacing downslope in response to enhanced groundwater flow and slope resistance (sediment and vegetation type).

Annual variations in solifluction intensity on the slope are mainly controlled by moisture supply, especially in the fall. Lobes downslope have a higher chance of becoming saturated during autumn freeze-back, but the upper slope segment may remain dry and almost inactive in some years. For unlimited moisture conditions (fully open system), such as in the lowermost, concave part of the slope, annual changes in solifluction rates may be directly correlated with annual temperatures. Warmer summer temperatures will lead to a deepening of the active layer and additional slope movement. This is more

likely in the lower zone where the active layer is thinner and corresponds more to the depth of movement, than upslope where the depth of movement is related to the supply of water to the freezing front, which if limited may cause a significant increase in the effective stress of the soil. Solifluction lobes located in the lowermost segment on the slope, where moisture supply is more constant, therefore, may be the best candidates for studies of long-term rates of advance in relation to changing temperatures. Otherwise, it is likely that annual variations in transport rates on the slope represent annual variations in the moisture regime of the active layer.

6.5. CONCLUSIONS

The following conclusions can be drawn from this study:

- (1) Solifluction lobes are spatially-organized landforms, tending toward a regular pattern of multi-lobate segments of increasing amplitude and increasing wavelength downslope.
- (2) This pattern results from transverse and longitudinal variations in solifluction transport rates, which give rise to zones of retarded flow behind which soliflucted material accumulates in lobate masses.
- (3) Gradients in environmental conditions are responsible for the increase in lobe size downslope. Sub-surface flow within the active layer is the primary controlling mechanism of solifluction, as it determines the potential for ice segregation and resulting frost-heave activity. Downslope increases in vegetation thickness augments the resistance of lobes to shear, while the shallower active layer and higher water-

table downslope enhance solifluction rates. Hence lobes downslope grow bigger behind a more resistant front.

- (4) Measurements of surface displacements on lobes of different sizes have shown that the pattern of movement is similar between lobes. Rates are initially high at the beginning of ground thaw and decrease over time. Most of the movement occurs in a 15-20 day period, after which targets remain stable or move slightly upslope. Surface rates of movement increase with lobe size and distance downslope.
- (5) Excavation of 12 solifluction lobes revealed the presence of plunging and emergent shear planes inside the lobes as well as thin layers of fine material ahead of the risers. ^{14}C dating of 24 humus samples collected within two metres of the lobe fronts or ahead of it revealed consistent age concentrations between lobes at a specific distance from the front. Overlapping ages were sometimes found over a 1 m distance, and these were taken as evidence of rapid burial by frontal slumping and colluviation. Despite the difference in lobe size and in sampling distance relative to the true modern organic mat, ages form three main concentrations representing the MRT of modern soil and two distinct burial events by rapid lobe advance.
- (6) Continuous monitoring of thaw-consolidation processes in a solifluction lobe showed that the release of moisture and soil strain occurred concurrently with the passage of the thawing front. Rapid downslope movement of the soil was recorded by a strain probe immediately after thaw. Rates of movement either decreased afterward, or

reversed partly as retrograde displacement. This behaviour suggests that phase-change consolidation is important in inducing downslope movement, with movement occurring by micro-shearing at ice lens boundaries. The influence of post-thaw viscous flow on total displacements cannot be ruled out, however, because of the possibility of stress redistribution across the strain probe.

- (7) Downslope movement can be linearly related to thaw settlement and active layer depth. A minimum amount of ice segregation is necessary to induce gelifluction movement in a thawing soil. This threshold corresponds to a critical excess moisture content of about 2%, which may be linked to the plastic limit of a soil (Chamberlain, 1981). Below this threshold, cohesion forces dominate and movement is retrograde. Equations for one-dimensional thaw-consolidation of thawing soils can be used to model thaw settlement of soil and the resulting downslope movement in relation to air temperature.
- (8) Solifluction lobes represent plastic landforms with yield stresses varying between 20-100 kPa (minimum estimates), depending on the plasticity of the soil, the degree of sorting and the strength of the retaining turf layer. Stress becomes concentrated near the bulging front as material accumulates on the riser, until break-up occurs and rapid advance of the lobe takes place by slumping and colluviation. This cycle of lobe development was evaluated from ^{14}C dates to be around 600 years.

- (9) The evolution of lobe profiles can be modelled in two-dimensions using an equation of mass continuity. Downslope variations in transport rates result in a progressive transfer of material from the back of the tread toward the riser, which thickens over time until a critical state of stability is attained. The model shows that at least 600 years are necessary to accomplish this cycle.
- (10) The response of solifluction lobes to climate change will be lagged by the time required for cyclic lobe development. Therefore, extensive and regular sampling along buried organic layers is required to be able to identify the step-like changes in lobe position over time. Under pronounced climate changes, the response time will vary and thus differences between age concentrations may be correlated with long-term changes in climate.

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APPENDIX A: DETERMINATION OF COORDINATES FROM SURVEY DATA

First, coordinates for all sighted points are calculated from benchmark #1 (BMP1), located on the bedrock outcrop, to which coordinates are arbitrarily assigned (see Figure 1). (x,y) coordinates are set to (1000,1000) to avoid negative coordinates for other points, while z is equal to the altitude given by the GPS.

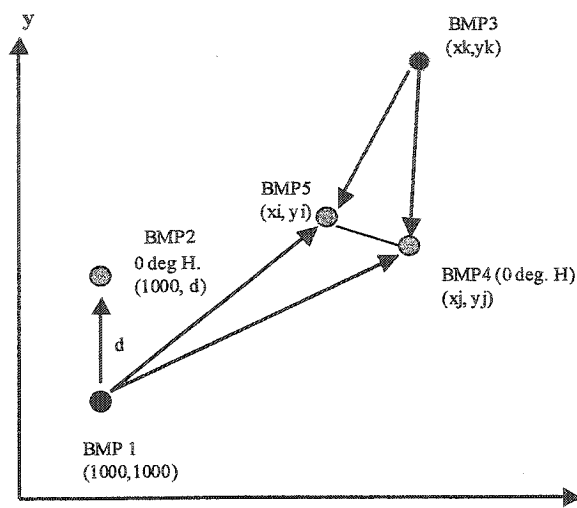


Figure 1: map showing the two stations used in the survey and their 0° sighting points.

All points sighted from BMP1 have the coordinates:

$$X_2 = d_2 * \sin H^\circ_2 \quad (1)$$

$$Y_s = d_s * \cos H^\circ_s \quad (2)$$

$$Z_s = z_s + d_s \cdot \tan V^{\circ}_s + (h_s - h_i) \quad (3)$$

Where:

X_a = x coordinate of point a (m)

$$Y_a = y \text{ coordinate of point a (m)}$$
$$Z_a = \text{elevation of point a (m)}$$
$$d_s = \text{distance between BMP1 and point 1 (m)}$$
 z_s = elevation of station BMP1 (from GPS, m)
$$H^{\circ}_a = \text{azimuth of point a}$$

V°_s = vertical angle of point a

$$h_s = \text{height of station above ground (m)}$$

h_a = height of staedia above point a (m)

Second, the coordinates of the second station upslope (BMP3) are calculated as:

$$X_k = X_i + d_{ik} * \sin \alpha_{ik} \quad (4)$$

$$Y_k = Y_i + d_{ik} * \cos \alpha_{ik} \quad (5)$$

$$Z_k = z_i + d_i * \tan V^\circ_i + (h_s - h_i) \quad (6)$$

Where:

$$\alpha_{ik} = \alpha_{ji} + \beta \quad (7)$$

$$\beta = [2 \arctan (d_{ki} - d_{kj} / d_{ki} + d_{kj}) * \tan 1/2(180 - H^\circ_i) + (180 - H^\circ_j)] / 2 \quad (8)$$

$$\alpha_{ij} = \arctan (x_i - x_j) / (y_i - y_j) \quad (9)$$

and:

$$X_{ijk} = x \text{ coordinate of point } i,j,k \text{ (m)}$$
$$Y_{ijk} = y \text{ coordinate of point } i,j,k \text{ (m)}$$
$$Z_{ijk} = \text{elevation of point } i,j,k \text{ (m)}$$

d_{ik} = distance between BMP3 and point i

d_{jk} = distance between BMP3 and point j

 α_{ik} = azimuth of vector ik (deg) α_{ij} = azimuth of vector ij (deg.)

β = angle between vector ik and vector ij

all other points sighted from BMP3 with 0° degree sighting at BMP4 have the coordinates:

$$X_h = X_k + d_h * (\sin H^\circ_h + \alpha_{ki})$$

$$Y_h = Y_k + d_h * (\cos H^\circ_{h+} \alpha_{kj})$$

$$Z_b = z_b + d_b * \tan V_b^o + (h_s - h_b), \text{ using the same terminology than for equations 1,2 and 3.}$$

(from Milton and Wong, 1985)

APPENDIX B: COMPUTATION OF RATES OF SURFACE MOVEMENT

Targets were placed on the surface of the lobe, along two axis: axial (line B1-B2) and transverse (line B3-B4) (Figure 1). The displacement of each target along the line B1-B2 is calculated from the changes in azimuth over time. At each time t , each side (at_x , bt_x , ct_x) and angle (At_x , Bt_x , Ct_x) of the new oblique triangle resulting from a change in azimuth αt_x for target i are calculated. The movement of target i at time from $t-1$ to t is the segment of the triangle ct_x opposed to the azimuth αt_x .

HORIZONTAL MOVEMENTS

Calculation of movement for targets along line B1-B2 (axial).

The first step is to compute the initial position of each target relative to B1 (y) and perpendicular to B1-B2 (x , or deflection from taut line)

There are two possible conditions in step 1:

$$\left\{ \begin{array}{l} x < 0 \text{ (negative deflection)} \\ x > 0 \text{ (positive deflection)} \end{array} \right\}$$

Initial conditions ($t=0$) for target i are:

$$\alpha Ct_0 = [\alpha B1 - \alpha it_0] \quad (1)$$

$$EXCEL = ABS(\alpha B1 - \alpha it_0)$$

x = measured deflection from taut line ($<$ or $>$ 0)

y = measured distance from point B1

at_0 = measured distance from station to B1

αit_0 = azimuth of point i at $t=0$

Calculated parameters are:

$$\alpha At_0 = 180 - \alpha Bt_0 - \alpha Ct_0 \quad (2)$$

$$EXCEL = 180 - \alpha At_0 - \alpha Ct_0$$

$$\alpha Bt_0 : \text{if } x < 0, \alpha Bt_0 = B_0 + \varepsilon \quad (3)$$

$$\text{if } x > 0, \alpha Bt_0 = B_0 - \varepsilon$$

$$EXCEL = IF(x < 0, B_0 + \varepsilon, B_0 - \varepsilon) \quad (4)$$

$$\text{where: } \varepsilon = \sin^{-1}(x/y)$$

$$EXCEL = ASIN(x/y * 180/PI())$$

$$\text{and: } (\alpha A_0 - \alpha B_0) = 2 \sin^{-1}(a_0 - b_0 / a_0 + b_0 * \tan 1/2(180 - \alpha C_0)) \quad (5)$$

$$EXCEL = (180 - \alpha C_0) - 2 * ATAN((a-b)/(a+b) * TAN(0.5 * (180 - \alpha C) * PI() / 180)) * 180 / PI()$$

*where αA_0 , αB_0 , αC_0 , a_0 , b_0 , c_0 refer to trigonometric parameters of triangle S-B1-B2

$$bt_0 = c * \sin Bt_0 / \sin Ct_0 \quad (6)$$

$$\begin{aligned}
EXCEL &= ct_0 * \sin(\alpha Bt_0 * \pi / 180) / \sin(\alpha Ct_0 * \pi / 180) \\
\alpha t_0 &= \sqrt{(x^2 + y^2)} \\
EXCEL &= \text{POWER}((\text{POWER}(x, 2) + \text{POWER}(y, 2)), 0.5)
\end{aligned} \tag{7}$$

The second step is to calculate the movement of the target at $t = 1$ (t_0 to t_1) from the initial position calculated in the first step.

There are 4 possible conditions in step 2:

$$\left\{ \begin{array}{l} \alpha t_1 > \alpha t_0 \text{ and } x < 0 \text{ (downslope movement)} \\ \alpha t_1 > \alpha t_0 \text{ and } x > 0 \\ \alpha t_1 < \alpha t_0 \text{ and } x < 0 \text{ (upslope movement)} \\ \alpha t_1 < \alpha t_0 \text{ and } x > 0 \end{array} \right\}$$

$$\begin{aligned}
\alpha At_1 : & \text{ if } \alpha t_1 > \alpha t_0 \text{ and } x < 0, \alpha At_1 = \alpha At_0 + \varepsilon \\
& \text{ if } \alpha t_1 > \alpha t_0 \text{ and } x > 0, \alpha At_1 = \alpha At_0 - \varepsilon \\
& \text{ (else) if } \alpha t_1 < \alpha t_0 \text{ and } x < 0 \text{ (or) } \alpha t_1 < \alpha t_0 \text{ and } x > 0, \alpha At_1 = 180 - \alpha Bt_1 - Ct_1 \\
EXCEL &= \text{IF}(\text{AND}(\alpha t_1 > \alpha t_0, x < 0), \alpha At_0 - \varepsilon, \text{IF}(\text{AND}(\alpha t_1 > \alpha t_0, x > 0), \alpha At_0 + \varepsilon, 180 - \alpha Bt_1 - Ct_1)))
\end{aligned} \tag{8}$$

$$\begin{aligned}
\alpha Bt_1 : & \text{ if } \alpha t_1 < \alpha t_0 \text{ and } x < 0, \alpha Bt_1 = 180 - \alpha At_0 + \varepsilon \\
& \text{ if } \alpha t_1 < \alpha t_0 \text{ and } x > 0, \alpha Bt_1 = 180 - \alpha At_0 - \varepsilon \\
& \text{ (else) if } \alpha t_1 > \alpha t_0 \text{ and } x < 0 \text{ (or) } \alpha t_1 > \alpha t_0 \text{ and } x > 0, \alpha Bt_1 = 180 - \alpha At_1 - Ct_1 \\
EXCEL &= \text{IF}(\text{AND}(\alpha t_1 < \alpha t_0 \text{ and } x < 0), 180 - \alpha At_0 + \varepsilon, \text{IF}(\text{AND}(\alpha t_1 < \alpha t_0 \text{ and } x > 0), 180 - \alpha At_0 - \varepsilon, 180 - \alpha At_1 - Ct_1)))
\end{aligned} \tag{9}$$

$$\begin{aligned}
\alpha Ct_1 &= [\alpha it_1 - \alpha it_0] \\
EXCEL &= \text{ABS}(\alpha t_1 - \alpha t_0)
\end{aligned} \tag{10}$$

$$\begin{aligned}
at_1 : & \text{ if } \alpha t_1 > \alpha t_0, at_1 = bt_1 \sin(Bt_1 + Ct_1) / \sin Bt_1 \\
& \text{ (else) if } \alpha t_1 < \alpha t_0, at_1 = bt_0 \\
EXCEL &= \text{IF}(\alpha t_1 > \alpha t_0, bt_1 * \sin((Bt_1 * \pi / 180) + (Ct_1 * \pi / 180)) / \sin(Bt_1 * \pi / 180), bt_0)
\end{aligned} \tag{11}$$

$$\begin{aligned}
bt_1 : & \text{ if } \alpha t_1 > \alpha t_0, bt_1 = bt_0 \\
& \text{ (else) if } \alpha t_1 < \alpha t_0, bt_1 = at_1 \sin Bt_1 / \sin At_1 \\
EXCEL &= \text{IF}(\alpha t_1 > \alpha t_0, b_0, at_1 * \sin((Bt_1 * \pi / 180)) / \sin(At_1 * \pi / 180))
\end{aligned} \tag{12}$$

$$\begin{aligned}
ct_1 \text{ (displacement)} &= bt_1 \sin Ct_1 / \sin(Bt_1) \\
EXCEL &= bt_1 * \sin(Ct_1 * \pi / 180) / \sin(Bt_1 * \pi / 180)
\end{aligned} \tag{13}$$

The third step is to calculate the movement of the target at t from the previous position at $t-1$, calculated in the second step.

The 4 possible conditions in step 3 are (example is for t_2):

$$\left\{ \begin{array}{l} \alpha t_2 > \alpha t_1 \text{ and } \alpha t_1 > \alpha t_0 \text{ (downslope movement)} \\ \alpha t_2 > \alpha t_1 \text{ and } \alpha t_1 < \alpha t_0 \\ \alpha t_2 < \alpha t_1 \text{ and } \alpha t_1 > \alpha t_0 \text{ (upslope movement)} \\ \alpha t_2 < \alpha t_1 \text{ and } \alpha t_1 < \alpha t_0 \end{array} \right\}$$

$$\mathbf{At}_2 : \text{if } \alpha t_1 > \alpha t_0 \text{ and } \alpha t_1 > \alpha t_0, \mathbf{At}_2 = 180 - \mathbf{Bt}_1 \quad (14)$$

$$\text{if } \alpha t_1 > \alpha t_0 \text{ and } \alpha t_1 < \alpha t_0, \mathbf{At}_2 = \mathbf{At}_1$$

$$\text{(else) if } \alpha t_1 < \alpha t_0 \text{ and } \alpha t_1 < \alpha t_0 \text{ (or) } \alpha t_1 < \alpha t_0 \text{ and } \alpha t_1 > \alpha t_0, \mathbf{At}_2 = 180 - \mathbf{Bt}_2 - \mathbf{Ct}_2$$

$$\text{EXCEL} = \text{IF}(\text{AND}(\alpha t_2 > \alpha t_1, \alpha t_1 > \alpha t_0), 180 - \mathbf{Bt}_1, \text{IF}(\text{AND}(\alpha t_2 > \alpha t_1, \alpha t_1 < \alpha t_0), \mathbf{At}_1, 180 - \mathbf{Bt}_2 - \mathbf{Ct}_2)).$$

$$\mathbf{Bt}_2 : \text{if } \alpha t_1 < \alpha t_0 \text{ and } \alpha t_1 < \alpha t_0, \mathbf{Bt}_2 = 180 - \mathbf{At}_1 \quad (15)$$

$$\text{if } \alpha t_1 \text{ if } \alpha t_1 < \alpha t_0 \text{ and } \alpha t_1 > \alpha t_0, \mathbf{Bt}_2 = \mathbf{Bt}_1$$

$$\text{(else) if } \alpha t_1 > \alpha t_0 \text{ and } \alpha t_1 < \alpha t_0 \text{ (or) } \alpha t_1 > \alpha t_0 \text{ and } \alpha t_1 > \alpha t_0, \mathbf{Bt}_2 = 180 - \mathbf{At}_2 - \mathbf{Ct}_2$$

$$\text{EXCEL} = (\text{IF}(\text{AND}(\alpha t_1 < \alpha t_0, \alpha t_1 < \alpha t_0), 180 - \mathbf{At}_1, \text{IF}(\text{AND}(\alpha t_1 < \alpha t_0, \alpha t_1 > \alpha t_0), \mathbf{Bt}_1, 180 - \mathbf{At}_2 - \mathbf{Ct}_2)))$$

$$\mathbf{Ct}_2 = [\alpha it_2 - \alpha it_1] \quad (16)$$

$$\text{EXCEL} = \text{ABS}(\alpha it_2 - \alpha it_1)$$

$$\mathbf{at}_2 : \text{if } \alpha t_1 < \alpha t_0 \text{ and } \alpha t_1 < \alpha t_0, \mathbf{at}_2 = \mathbf{bt}_1 \quad (17)$$

$$\text{if } \alpha t_1 < \alpha t_0 \text{ and } \alpha t_1 > \alpha t_0, \mathbf{at}_2 = \mathbf{at}_1$$

$$\text{(else) if } \alpha t_1 > \alpha t_0 \text{ and } \alpha t_1 < \alpha t_0 \text{ (or) } \alpha t_1 > \alpha t_0 \text{ and } \alpha t_1 > \alpha t_0, \mathbf{at}_2 = \sin \mathbf{At}_2 * \mathbf{bt}_2 / \sin \mathbf{Bt}_2$$

$$\text{EXCEL} = (\text{IF}(\text{AND}(\alpha t_1 < \alpha t_0, \alpha t_1 < \alpha t_0), \mathbf{bt}_1, \text{IF}(\text{AND}(\alpha t_1 < \alpha t_0, \alpha t_1 > \alpha t_0), \mathbf{at}_1, \text{SIN}(\mathbf{At}_2 * \text{PI}() / 180) * \mathbf{bt}_2 / \text{SIN}(\mathbf{Bt}_2 * \text{PI}() / 180))))$$

$$\mathbf{bt}_2 : \text{if } \alpha t_1 > \alpha t_0 \text{ and } \alpha t_1 > \alpha t_0, \mathbf{bt}_2 = \mathbf{at}_1 \quad (18)$$

$$\text{if } \alpha t_1 < \alpha t_0 \text{ and } \alpha t_1 > \alpha t_0, \mathbf{bt}_2 = \sin \mathbf{Bt}_2 * \mathbf{at}_2 / \sin \mathbf{At}_2$$

$$\text{(else) if } \alpha t_1 > \alpha t_0 \text{ and } \alpha t_1 < \alpha t_0 \text{ (or) } \alpha t_1 < \alpha t_0 \text{ and } \alpha t_1 < \alpha t_0, \mathbf{bt}_2 = \sin \mathbf{Bt}_2 * \mathbf{at}_2 / \sin \mathbf{At}_2$$

$$\text{EXCEL} = \text{IF}(\text{AND}(\alpha t_1 > \alpha t_0, \alpha t_1 > \alpha t_0), \mathbf{at}_1, \text{IF}(\text{AND}((\alpha t_1 > \alpha t_0, \alpha t_1 < \alpha t_0), \mathbf{bt}_1, \text{SIN}(\mathbf{Bt}_2 * \text{PI}() / 180) * \mathbf{at}_2 / \text{SIN}(\mathbf{At}_2 * \text{PI}() / 180))))$$

$$\mathbf{ct}_2 \text{ (displacement)} = \mathbf{at}_2 \sin(\mathbf{At}_2 + \mathbf{Bt}_2) / \sin \mathbf{At}_2 \quad (19)$$

$$\text{EXCEL} = \mathbf{at}_2 * \text{SIN}((\mathbf{At}_2 * \text{PI}() / 180) + (\mathbf{Bt}_2 * \text{PI}() / 180)) / \text{SIN}(\mathbf{At}_2 * \text{PI}() / 180)$$

Calculation of movement for targets along transverse axis B3-B4

The first step is to compute the initial position of each target relative to B3 (y) and perpendicular to B3-B4 (x, or deflection from taut line)

Initial possible conditions for step 1 are the same as for long axis: $x < 0$ or $x > 0$

Initial conditions are the same as for long axis but B1 is replaced by B3

At₀ : use equation (2)

Bt₀ : use equation (3), (4) and (5), but A₀, B₀, C₀, a₀, b₀ and c₀ refer to oblique triangle S-B3-B4 instead.

bt₀ : use equation (6)

ct₀ (displacement): use equation (7)

The second step is the movement from t₀ to t₁

The 4 possible conditions are the same as for long axis:

$$\left\{ \begin{array}{l} \alpha t_1 > \alpha t_0 \text{ and } x < 0 \text{ (downslope movement)} \\ \alpha t_1 > \alpha t_0 \text{ and } x > 0 \\ \alpha t_1 < \alpha t_0 \text{ and } x < 0 \text{ (upslope movement)} \\ \alpha t_1 < \alpha t_0 \text{ and } x > 0 \end{array} \right\}$$

At₁ : if $\alpha t_1 > \alpha t_0$ and $x < 0$, $At_1 = At_0 - 90 + \varepsilon$ (20)

if $\alpha t_1 > \alpha t_0$ and $x > 0$, $At_1 = At_0 - 90 - \varepsilon$

(else) if $\alpha t_1 < \alpha t_0$ and $x < 0$ (or) $\alpha t_1 < \alpha t_0$ and $x > 0$, $At_1 = 180 - Bt_1 - Ct_1$

EXCEL = IF(AND($\alpha t_1 > \alpha t_0$ and $x < 0$), $At_0 - 90 + \varepsilon$, IF(AND($\alpha t_1 > \alpha t_0$ and $x > 0$), $At_0 - 90 - \varepsilon$, 180 - $Bt_1 - Ct_1$))

Bt₁ : if $\alpha t_1 < \alpha t_0$ and $x < 0$, $Bt_1 = 270 - At_0 - \varepsilon$ (21)

if $\alpha t_1 < \alpha t_0$ and $x > 0$, $Bt_1 = 270 - At_0 + \varepsilon$

(else) if $\alpha t_1 > \alpha t_0$ and $x < 0$ (or) $\alpha t_1 > \alpha t_0$ and $x > 0$, $Bt_1 = 180 - At_1 - Ct_1$

EXCEL = IF(AND($\alpha t_1 < \alpha t_0$ and $x < 0$), $270 - At_0 - \varepsilon$, IF(AND($\alpha t_1 < \alpha t_0$ and $x > 0$), $270 - At_0 + \varepsilon$, 180 - $At_1 - Ct_1$))

Ct₁ = use equation (10)

at₁ = use equation (11)

bt₁ = use equation (12)

ct₁ = use equation (13)

For the third step (movement from t-1 to t): use equations (14) to (19) for the long axis.

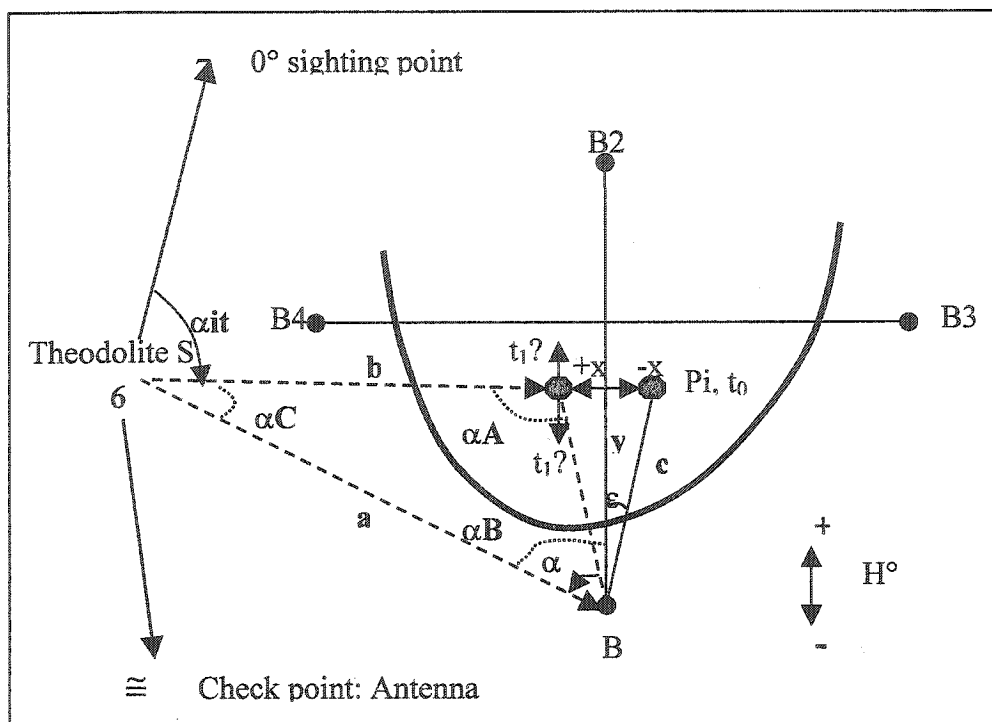


Fig. B.1: Possible initial positions of targets along long axis B1-B2 at time t_0 , with trigonometric parameters shown for position $+x$.

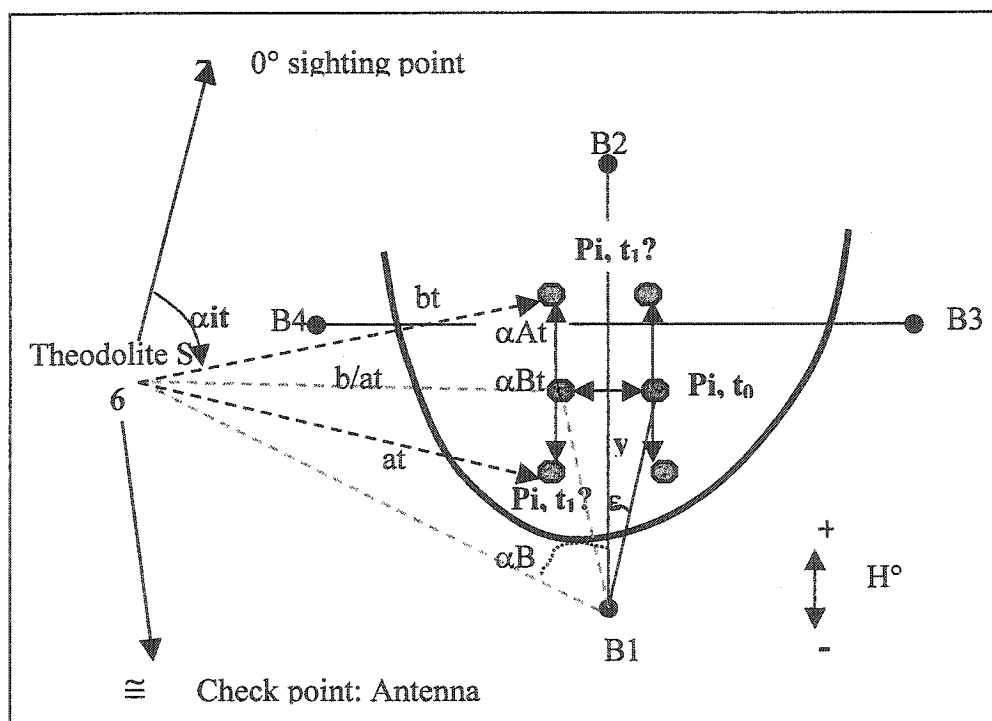


Fig. B.2: Possible positions of targets along long axis B1-B2 at time t_1 , with trigonometric parameters shown for position $+x$, $-y$ (upslope movement)

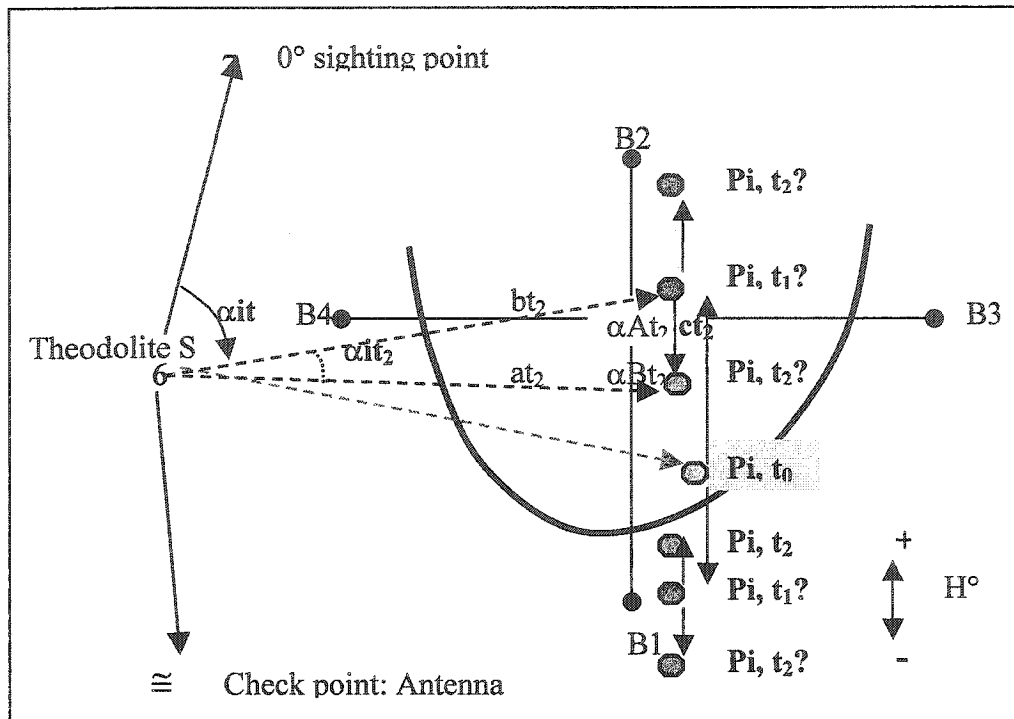


Fig. B.3: Possible positions of targets along long axis B1-B2 at time t_2 , with trigonometric parameters shown for position $-x, +y$ (downslope movement from t_1 to t_2).

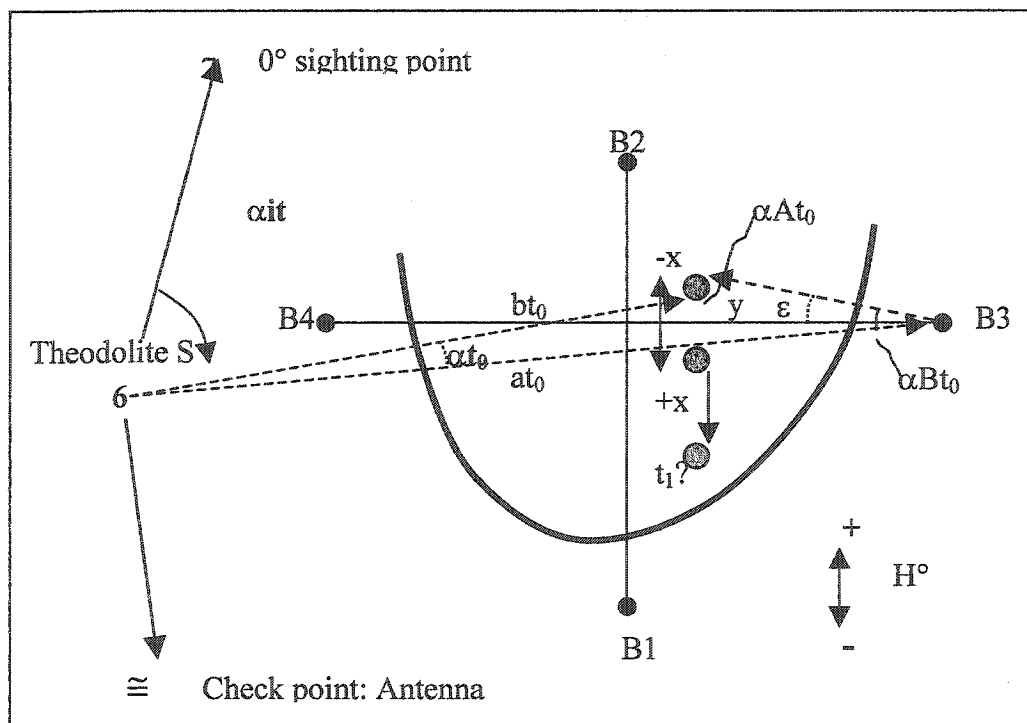


Fig. B.4: Possible initial positions of targets along transverse axis B3-B4 at time t_0 , with trigonometric parameters shown for position $-x$.

VERTICAL MOVEMENTS

The vertical displacement of each target was evaluated as follows:

The first step is to calculate, for every time (t), the height of the theodolite station relative to a fixed point. The height of the theodolite relative to its positioning benchmark (re-bar) was measured every time with a steel tape. But, data is missing for some days, and precision is not better than 5 mm. Thus, the height of the station is calculated for each time (t) using the vertical angle when sighting on the 0° benchmark (Vb). The lateral distance between the theodolite benchmark and sighting benchmark is known (D), so the relative height (H_t) is:

$$H_t = D \cdot \tan V_b \quad (22)$$

Changes in theodolite height at each time (t) is calculated as:

$$dH_t = H_{t_0} - H_t \quad (23)$$

The lateral distance d_t between the theodolite and each target at each time (t) is needed and is equal to:

$$\begin{aligned} \text{If } \alpha_t > \alpha_{t-1}: d_t &= a_t \text{ (calculated from equations 11 and 17)} \\ \alpha_t < \alpha_{t-1}: d_t &= b_t \text{ (calculated from equations 6,12 and 18)} \\ \alpha_t &= \alpha_{t-1}: d_t = d_{t-1} \end{aligned} \quad (24)$$

The vertical distance z_t between each target and the theodolite at time (t) is:

$$h_t = d_t \cdot \tan (90 - V_t) \quad (25)$$

where V_t is the vertical angle when sighting on the target. Changes in vertical distance dh between each time t and the first reading h_{t₀} is:

$$dh_t = h_t - h_{t_0} \quad (26)$$

and subtracting from the changes in theodolite height (H_t) gives the displacement dz at each time (t):

$$(dz)_t = dh_t - dH_t \quad (27)$$

Finally, the coordinate z_t, relative to 0 at time=0, is:

$$z_t = dz_t - dz_{t_0} \quad (28)$$

The error for each reading is the addition of the precision in theodolite height (equation 23) and the precision of the vertical angle of reading on each target (10''):

$$\text{Error} = D \text{ (mm)} \cdot \tan 10''/3600 + \tan 10''/3600 \cdot d_t = 1.98\text{mm} + \approx 0.5\text{mm} = \approx \pm 2.5 \text{ mm}$$

APPENDIX C: CONVERSION OF STRAIN VALUES INTO SOIL DISPLACEMENT VALUES

Strain is defined as the ratio of the change in length along a given axis of a physical member under stress to the unstressed length, and may be expressed as:

$$\varepsilon = \Delta L / L \quad (1)$$

where ε = strain (dimensionless)
and L = length (mm)

The *strain gauge* is a transducer which converts force to an electrical signal. The underlying principle of the strain gauge is that a stressed metallic conductor undergoes a change in electrical resistance directly proportional to a change in length. The *gauge factor* is a dimensionless quantity defined as the ratio of the fractional change in resistance to the fractional change in length along the axis of the strain gauge. The larger the value, the more sensitive the gauge. One of the most accurate ways to measure strain gauges is to use a *wheatstone bridge* configuration. In a bridge measurement system, a precisely regulated voltage (excitation voltage, V_{IN}) is applied across the bridge (see Figure 4.7). The voltage across the two center taps is then measured. As the resistance of the stressed strain gauge changes, the voltage across the center taps will also change. This measured change in voltage is proportional to the change in resistance of the strain gauge which in turn is proportional to the strain experienced by the strain gauge. For the strain probe, 4 strain gauges were connected into a full-bridge circuit. The result of the datalogger's full bridge measurement is:

$$1000 \cdot V_{OUT} / V_{IN} = 1000 \cdot [R_4 / (R_3 + R_4) - R_2 / (R_1 + R_2)] \quad (2)$$

where V_{IN} = the voltage supplied by the 21X datalogger to power the bridge (5V)

V_{OUT} = the return voltage measured by the 21X data logger

R_1, R_2, R_3, R_4 = Strain gauges numbers

For a balanced full-bridge, the strain is equal to:

$$\varepsilon = -V_r / G_f \quad (3)$$

where $V_r = [(V_{OUT}/V_{IN})_{strained} - (V_{OUT}/V_{IN})_{unstrained}]$ (mV/V)

G_f = gauge factor (here = 2)

Strain values obtained for each of the ten 10 cm-long sections of the probe may be used to calculate soil displacement values. The method outlined by Yamada and Kurashige (1996) was used for this purposed. The method assumes that each 10 cm-long section of the probe deforms as an arc with a curvature function related to the measured strain (see for example, Williams (1957, 1962) and Auzet et al. (1986). For a complete outline of the constitutive equations, the reader is referred to Yamada and Kurashige (1996). These equations were programmed using the software MATHCAD 2001 Professional. The program is reported below.

i := 1..1273

k := 1..10

INPUT VALUES: EXCEL CHART OF STRAIN VALUES
(1273,10) matrix

$$R_{i,k} := \begin{cases} 0 \cdot \text{mm} & \text{if } \varepsilon_{i,k} = 0 \\ \frac{0.1 \cdot \text{mm}}{|\varepsilon_{i,k}|} & \text{otherwise} \end{cases}$$

$$\theta_{i,k} := |\varepsilon_{i,k}| \cdot \frac{100 \cdot \text{mm}}{0.1 \cdot \text{mm}}$$

t :=	10	20	30	40	50	60	70	80	90	100
1	0	0	0	0	0	0	0	0	0	0
2	0	0	0	0	4.76E-07	4.76E-07	4.76E-07	0	6.67E-06	-4.8E-07
3	0	0	0	0	4.76E-07	9.52E-07	0	0	1.52E-05	0
4	0	0	-4.8E-07	0	0	4.76E-07	-4.8E-07	-4.8E-07	2.33E-05	-1.9E-06
5	0	0	-4.8E-07	0	0	9.52E-07	0	-4.8E-07	3.33E-05	-4.8E-06
	0	0	-4.8E-07	0	0	9.52E-07	0	-1.4E-05	4.19E-05	-4.3E-06

biere_{i,k} := if $\varepsilon_{i,1} > 0$

$$\alpha_{i,1} \leftarrow 0 \cdot \text{cm} + R_{i,1}$$

$$\beta_{i,1} \leftarrow 0 \cdot \text{cm}$$

$$\phi_{i,1} \leftarrow \text{asin} \left(\frac{-\beta_{i,1}}{R_{i,1}} \right)$$

$$x_{B_{i,1}} \leftarrow \alpha_{i,1} - R_{i,1} \cdot \cos(\phi_{i,1} + \theta_{i,1})$$

$$z_{B_{i,1}} \leftarrow \beta_{i,1} + R_{i,1} \cdot \sin(\phi_{i,1} + \theta_{i,1})$$

if $\varepsilon_{i,1} < 0$

$$\alpha_{i,1} \leftarrow 0 \cdot \text{cm} - R_{i,1}$$

$$\beta_{i,1} \leftarrow 0 \cdot \text{cm}$$

$$\phi_{i,1} \leftarrow \text{asin} \left(\frac{\beta_{i,1}}{R_{i,1}} \right)$$

$$x_{B_{i,1}} \leftarrow \alpha_{i,1} + R_{i,1} \cdot \cos(\phi_{i,1} - \theta_{i,1})$$

$$z_{B_{i,1}} \leftarrow \beta_{i,1} - R_{i,1} \cdot \sin(\phi_{i,1} - \theta_{i,1})$$

otherwise

$$x_{B_{i,1}} \leftarrow 0 \cdot \text{cm}$$

$$z_{B_{i,1}} \leftarrow 10 \cdot \text{cm}$$

$j \leftarrow 2$

while $j \leq 10$

$$a_{i,j} \leftarrow \frac{z_{B_{i,j-1}} - z_{B_{i,j-2}}}{x_{B_{i,j-1}} - x_{B_{i,j-2}}} \text{ if } j \geq 3 \wedge \varepsilon_{i,j} \neq 0 \wedge x_{B_{i,j-1}} - x_{B_{i,j-2}} \neq 0$$

$$a_{i,j} \leftarrow \frac{z_{B_{i,j-1}}}{x_{B_{i,j-1}}} \text{ if } j < 3 \wedge \varepsilon_{i,j} \neq 0 \wedge x_{B_{i,j-1}} \neq 0$$

$a_{i,j} \leftarrow 0$ otherwise

if $\varepsilon_{i,j} > 0 \wedge a_{i,j} > 0$

$$\alpha_{i,j} \leftarrow x_{B_{i,j-1}} + a_{i,j} \sqrt{\frac{(R_{i,j})^2}{1 + (a_{i,j})^2}}$$

$$\beta_{i,j} \leftarrow z_{B_{i,j-1}} - \sqrt{\frac{(R_{i,j})^2}{1 + (a_{i,j})^2}}$$

$$\phi_{i,j} \leftarrow \arcsin\left(\frac{z_{B_{i,j-1}} - \beta_{i,j}}{R_{i,j}}\right)$$

$$x_{B_{i,j}} \leftarrow \alpha_{i,j} - R_{i,j} \cdot \cos(\phi_{i,j} + \theta_{i,j})$$

$$z_{B_{i,j}} \leftarrow \beta_{i,j} + R_{i,j} \cdot \sin(\phi_{i,j} + \theta_{i,j})$$

if $\varepsilon_{i,j} > 0 \wedge a_{i,j} < 0$

$$\alpha_{i,j} \leftarrow x_{B_{i,j-1}} - a_{i,j} \sqrt{\frac{(R_{i,j})^2}{1 + (a_{i,j})^2}}$$

$$\beta_{i,j} \leftarrow z_{B_{i,j-1}} + \sqrt{\frac{(R_{i,j})^2}{1 + (a_{i,j})^2}}$$

$$\phi_{i,j} \leftarrow \arcsin\left(\frac{-z_{B_{i,j-1}} + \beta_{i,j}}{R_{i,j}}\right)$$

$$x_{B_{i,j}} \leftarrow \alpha_{i,j} - R_{i,j} \cdot \cos(\phi_{i,j} - \theta_{i,j})$$

$$z_{B_{i,j}} \leftarrow \beta_{i,j} - R_{i,j} \cdot \sin(\phi_{i,j} - \theta_{i,j})$$

if $\varepsilon_{i,j} < 0 \wedge a_{i,j} > 0$

if $\varepsilon_{i,j} < 0 \wedge a_{i,j} > 0$

$$\alpha_{i,j} \leftarrow x_{B_{i,j-1}} - a_{i,j} \cdot \frac{(R_{i,j})^2}{\sqrt{1 + (a_{i,j})^2}}$$

$$\beta_{i,j} \leftarrow z_{B_{i,j-1}} + \frac{(R_{i,j})^2}{\sqrt{1 + (a_{i,j})^2}}$$

$$\phi_{i,j} \leftarrow \text{asin}\left(\frac{-z_{B_{i,j-1}} + \beta_{i,j}}{R_{i,j}}\right)$$

$$x_{B_{i,j}} \leftarrow \alpha_{i,j} + R_{i,j} \cdot \cos(\phi_{i,j} - \theta_{i,j})$$

$$z_{B_{i,j}} \leftarrow \beta_{i,j} - R_{i,j} \cdot \sin(\phi_{i,j} - \theta_{i,j})$$

if $\varepsilon_{i,j} < 0 \wedge a_{i,j} < 0$

$$\alpha_{i,j} \leftarrow x_{B_{i,j-1}} + a_{i,j} \cdot \frac{(R_{i,j})^2}{\sqrt{1 + (a_{i,j})^2}}$$

$$\beta_{i,j} \leftarrow z_{B_{i,j-1}} - \frac{(R_{i,j})^2}{\sqrt{1 + (a_{i,j})^2}}$$

$$\phi_{i,j} \leftarrow \text{asin}\left(\frac{z_{B_{i,j-1}} - \beta_{i,j}}{R_{i,j}}\right)$$

$$x_{B_{i,j}} \leftarrow \alpha_{i,j} + R_{i,j} \cdot \cos(\phi_{i,j} + \theta_{i,j})$$

$$z_{B_{i,j}} \leftarrow \beta_{i,j} + R_{i,j} \cdot \sin(\phi_{i,j} + \theta_{i,j})$$

if $\varepsilon_{i,j} > 0 \wedge a_{i,j} = 0$

$$\alpha_{i,j} \leftarrow x_{B_{i,j-1}} + R_{i,j}$$

$$\beta_{i,j} \leftarrow z_{B_{i,j-1}}$$

$$\phi_{i,j} \leftarrow \text{asin}\left(\frac{z_{B_{i,j-1}} - \beta_{i,j}}{R_{i,j}}\right)$$

$$x_{B_{i,j}} \leftarrow \alpha_{i,j} - R_{i,j} \cdot \cos(\phi_{i,j} + \theta_{i,j})$$

$$z_{B_{i,j}} \leftarrow \beta_{i,j} + R_{i,j} \cdot \sin(\phi_{i,j} + \theta_{i,j})$$

```

if  $\varepsilon_{i,j} < 0 \wedge a_{i,j} = 0$ 
     $\alpha_{i,j} \leftarrow x_{B_{i,j-1}} - R_{i,j}$ 
     $\beta_{i,j} \leftarrow z_{B_{i,j-1}}$ 
     $\phi_{i,j} \leftarrow \arcsin\left(\frac{\beta_{i,j} - z_{B_{i,j-1}}}{R_{i,j}}\right)$ 
     $x_{B_{i,j}} \leftarrow \alpha_{i,j} + R_{i,j} \cdot \cos(\phi_{i,j} - \theta_{i,j})$ 
     $z_{B_{i,j}} \leftarrow \beta_{i,j} - R_{i,j} \cdot \sin(\phi_{i,j} - \theta_{i,j})$ 
otherwise
     $x_{B_{i,j}} \leftarrow x_{B_{i,j-1}}$ 
     $z_{B_{i,j}} \leftarrow z_{B_{i,j-1}} + 10 \cdot \text{cm}$ 
j  $\leftarrow$  j + 1
 $z_{B_{i,k}}$ 

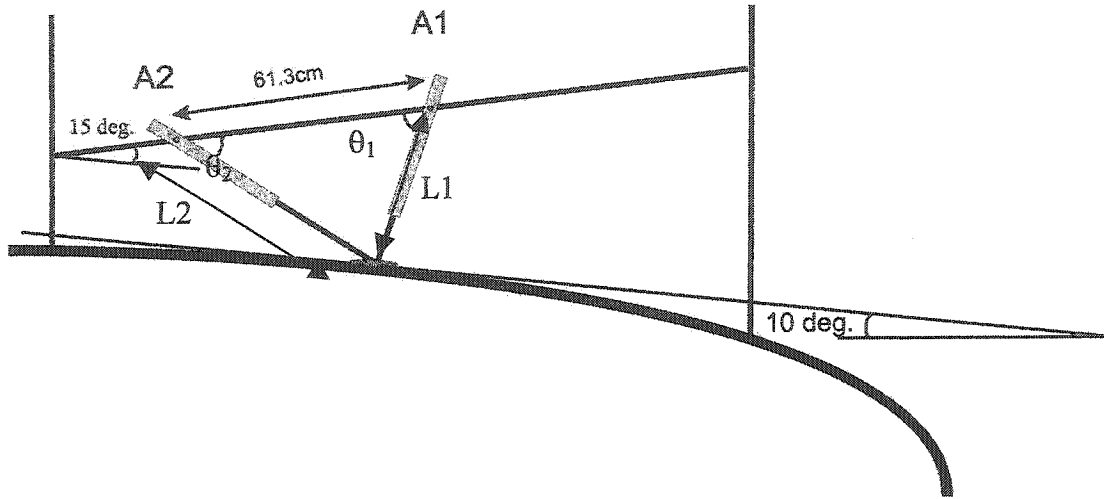
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OUTPUT VALUES: EXCEL CHART OF $X_{Bi,j}$, $Z_{Bi,j}$ VALUES (mm)

	1	2	3	4	5	6	7	8	9	10
1	0	0	0	0	0	0	0	0	0	0
2	0	0	0	0	0.02381	0.03968	0.07937	0.07937	0.33333	0.22487
3	0	0	0	0	0.02381	0.06349	0.06349	0.06349	0.76189	0.76189
4	0	0	-0.02381	-0.02381	-0.02381	0.02381	-0.02381	-0.09524	0.99995	1.99994
5	0	0	-0.02381	-0.02381	-0.02381	0.04762	0.04762	-0.02381	1.71412	3.21406
6	0	0	-0.02381	-0.02381	-0.02381	0.04762	0.04762	-0.07143	2.14251	4.14238
7	0.02381	0.02381	-0.02381	-0.02381	0.02381	0.04762	0.09524	0.04762	2.4757	4.78502
8	0.02381	0.07143	0.07143	0.07143	0.02381	0.13492	0.31746	0.38095	3.11045	5.86407
9	0.02381	0.07143	0.07143	-0.02381	0.04762	0.21429	0.5	0.59524	3.35646	6.26082
10	0.02381	0.07143	0.07143	-0.02381	0.04762	0.21429	0.5	0.59524	3.47542	6.52256
11	0.04762	0.11905	0.11905	-0.02381	0.07143	0.2619	0.57143	0.66667	3.61822	6.73676
12	0.04762	0.11905	0.11905	-0.02381	0.07143	0.2619	0.57143	0.66667	3.64201	6.78434
13	0.04762	0.11905	0.11905	-0.02381	0.07143	0.2619	0.57143	0.66667	3.66579	6.83193

The general precision of each measurement refers to the precision of the data-logger (0.001mV/V). Thus, a positive change of 0.001 mV/V corresponds to a strain of 0.476 micro-strain ($\mu\epsilon$). Using the program for soil displacement calculation, this corresponds to a precision of 0.024 mm.

APPENDIX D: CALCULATION OF ABLATOMETER DISPLACEMENTS



Calculation of surface displacement refers to the Figure above, where the position of each ablatometer arm is shown. We need to calculate the displacement parallel to the lobe surface, dx , and the displacement perpendicular to the lobe surface, dz .

Knowing the fixed distance between both ablatometers on the transverse beam (613 mm), we can calculate the changes in angle θ_1 and θ_2 .

$$\theta_1 = \sin^{-1} (\sin \theta_2 \cdot L2/L1) \quad (1)$$

$$\theta_2 = 2\sin^{-1} \sqrt{[(s-L2) \cdot (s-613\text{mm})] / (L2 \cdot 613\text{mm})} \quad (2)$$

and $s = 0.5(613\text{mm} + L1 + L2)$

where $L1$ and $L2$ are the length of the ablatometers in mm, and θ_1 and θ_2 their angle with the transverse beam.

The parallel (dx) and perpendicular (dz) displacement of the footplate in mm is given by:

$$dz_t = \cos(90-15-\theta_1) \cdot L1 \quad (3)$$

$$dx_t = \cos(\theta_2-15) \cdot L2 \quad (4)$$

The length of each ablatometer is calculated for each time (t) from voltage returned by the 21X data logger. Each ablatometer was calibrated prior to the study, and linear relationships between the extension of each ablatometer and the corresponding voltage signal was found:

$$(A1): dz_t = 419\text{mm} + 0.5981 \cdot V \quad (5)$$

$$(A2): dx_t = 390\text{mm} + 0.6067 \cdot V \quad (6)$$

Where V is the return voltage

The error is equal to the precision of the logger: $0.1\text{mV} \cdot 0.61 = 0.061\text{mm} \approx 0.1 \text{ mm}$